Paleozoic Geodynamic Domains
and their Alpidic Evolution in the Tethys
IGCP Project No. 276
NEWSLETTER Vol. 5 *special issue*

CONTRIBUTIONS TO THE GEOLOGY OF ITALY

WITH SPECIAL REGARD TO THE PALEOZOIC BASEMENTS

A VOLUME DEDICATED TO TOMMASO COCOZZA

Edited by
L. CARMIGNANI & F.P. SASSI
with the collaboration of
P.L. Fantozzi, M. Meccheri & R. Spiess

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FOREWORD

The present volume is related to the Meeting held in Siena (21-22 March1991), organized in memory of Tommaso Cocozza by the Department of Earth Sciences of Siena, with the sponsorship of the IGCP Project No. 276.

Its content reflects Tommaso's wide range of scientific interests, from stratigraphy and paleontology to regional geology, petrography and tectonics. However, the problems related to the crystalline basements and the Paleozoic geodynamic domains certainly prevailed among the many topics studied by him.

This prevalence explains the inclusion of the present volume in the Newsletter of IGCP Project No. 276. It is the best way to acknowledge the strong support Tommaso gave to the activities of expired IGCP Project No. 5 ("Pre-Variscan and Variscan Events in the Alpine-Mediterranean Belts") and new IGCP Project No. 276 ("Paleozoic geodynamic domains and their Alpidic evolution in the Tethys"), and to record his enthusiasm in promoting international scientific collaboration.

L. CARMIGNANI, F.P. SASSI
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Project Leader IGCP No. 276

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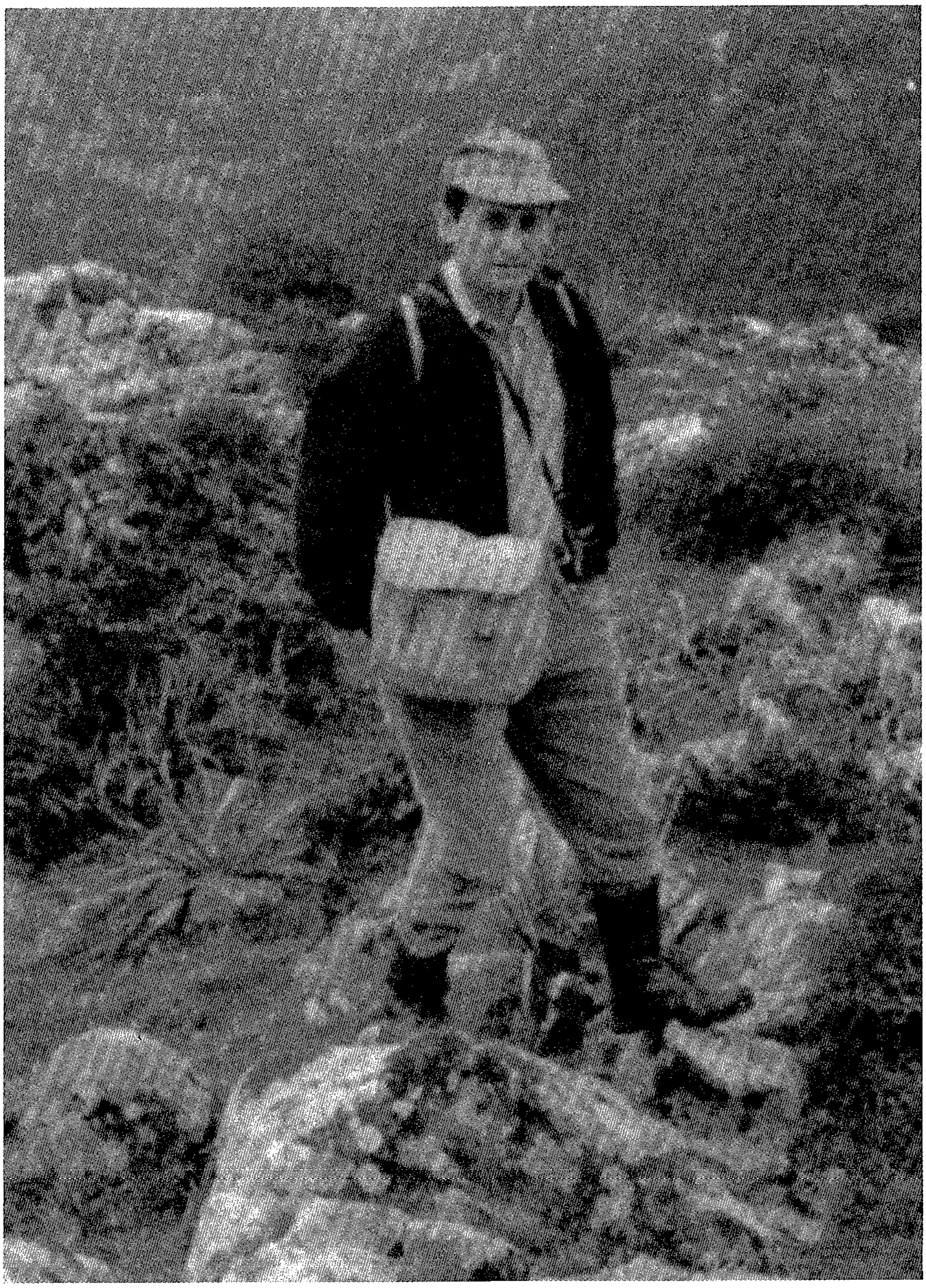
In ricordo dell'amico Tommaso



DEDICATED TO THE MEMORY OF TOMMASO COCOZZA

Tommaso Cocozza's working life was spent almost equally between the Universities of Rome, Cagliari and Siena. That is why we chose to report the remembrances of his friends from these three Universities.

From Rome

When i was newly graduated, i got a job mapping for the second edition of the Geological Map of Italy, and found myself working with Tommaso Cocozza and Giuseppe Sirna. Although i knew Cocozza only by sight, since he had graduated from the same University, La Sapienza, a few years before me, and Sirna not at all, a friendship between the three of us was struck up right away and never failed.

Our friendship was reinforced by the fact that we shared not only our daily work with its scientific puzzles and the practical difficulties of mapping in hard-to-reach areas, but also the same run down boarding house, which was all we could afford. Every evening on returning from our field trips we would talk over ours problems, results and the questions that had arisen during the day's work. We often went out together to have another look, check and compare results.

This extended period of steady contact allowed me get to know, appreciate and esteem Tommaso Cocozza. His opennes to others was something that he always maintained, even when life was most difficult for him, so that each of us knew that we could always count on his perceptive and impartial opinion, his thorough knowledge of geology, and his scientific intuition. He soon became our reference-point for our field problems and guided our scientific discussions, not just because of his greater experience but also because of those traits of his that made him so ready to involve himself in his friends' problems.

As time went by, my esteem increased along with the friendship that he inspired by his straightforwardness, his unique style of talking and gesticulating, the spirit of co-operation that he managed to instill in those around him and his ability to blend perceptive scientific commentary whith his ironic and easygoing personal opinions.

During that early period he refined his skills as a field geologist and laid the groundwork for his brilliant academic career. When he moved to Cagliari as "assistente ordinario", we realized how important he had been for us and how much we would miss his warm openness.

Altough our scientific pathways separated, we kept in touch and distance seemed strangely to strengthen our friendship. Meeting again many years later in Somalia, we both had the pleasant opportunity to renew our exchange of friendship and ideas, as in the past.

In this new milieu too, he was immediately valued for his profound knowledge of geological matters, togeter wit and easy going good nature, and he made substantial contributions, both as a teacher and in planning local geological research.

Having known Tommaso Cocozza at such different times and in such different environments, I am aware that his competence and personality were felt wherever he was, so that his memory remains very much alive for all of us who knew him and shared ideas and emotions with him.

A. Angelucci , Università "La Sapienza", Roma.

From Cagliari

In the early 1960s Tommaso Cocozza was appointed "assistente ordinario" and "docente" of physical geography at the University of Cagliari. He therefore moved there from Rome and married.

At the University of Cagliari he enthusiastically threw himself into work on the various problems of Sardinian geology , in particular those relating to the Paleozoic. He soon acquired a broad knowledge of the abundant but not recent literature on the subject and began methodical and accurate field work especially in Iglesiente and Sulcis. He was a natural in Geology and soon published various papers and was able to get his "libera docenza" in Geology in 1968.

During his stay in Cagliari, which lasted about ten years, he showed himself to be an aimable extrovert, always ready to see the best and most human side of others. He soon became an active and capable organizer of numerous field trips for research and teaching. Some of the field trips abroad (Morocco, Spain, France) were memorable for their well-thought-out scientific organization and their exceptional educational value. In fact, his strengths in matters both scientific and personal were exhibited to their fullest extenct on field trips.

A very modest person, able to adapt easily to various situations, he always made people feel at ease. People also enjoyed his lively and interesting conversation, in which geological discussions were enriched by his broad interests and general culture, and his concise speech was enlivened by his expressive miming.

Tommaso Cocozza loved telling anecdotes about his life and his friends in Rome and also often spoke with affection of his birthplace, Molise.

Naturally inclined to friendship, Tommaso made many friends in Sardinia, both inside and outside the scientific world, and even after his move to Siena, right up to his death, he maintained close ties with Sardinia, because of his affection for many people there as much as his scientific interest.

During his brief temporary return to Cagliari in 1975 as "professore straordinario" teaching Stratigraphical Geology, he was a decisive influence in the creation of new research projects which he followed and developed unfailingly despite grave and increasing worries due to illness in the family.

Even after he had been definitely transferred to the University of Siena, he continued an intensive research program in Sardinia, organizing a large and active group of researchers into an efficent and cohesive body.

His contribution to the analysis and interpretation of the Paleozoic was essential to the development of our present day understanding of the Paleozoic basement and its related structural model.

It is with deep feeling that at this congress in memory of Tommaso Cocozza we, his numerous friends, colleagues and students from the University of Cagliari, renew and express our feelings of sincere esteem, affectionate friendship, and deep and grateful indebtedness towards him.

S. Barca and V. Palmerini, Università degli Studi di Cagliari.

From Siena

Tommaso Cocozza has been full professor of geology at the University of Siena.

His death on July 26 1989, left a void that will be difficult to fill in the scientific world. Many of you know what a large and significant body of scientific research he produced, dealing mainly but not exclusively with the stratigraphy, sedimentology, paleontology and tectonics of the Sardinian and Tuscan Paleozoic. His broad culture and intellectual and organizational gifts made him a frequent participant, often as co-ordinator, in national and international research projects, where he earned the esteem and friendship of his colleagues.

Those who knew him personally can attest to his great capacity to give and receive friendship and warmth and especially his ability to instil optimism in those around him, even in the most difficult and trying periods of his life. Another remarkable quality of Tommaso was his readiness to encourage, guide, and financially help young researchers. The numerous letters of admiration and condolence which reached us after his premature death led us to organize a congress covering the same research fields that Tommaso himself worked on during his productive scientific life. We give below a brief summary of his work.

After a brief engagement as a field geologist, with SOMICEM (ENI Group) and the Monte Amiata Mining Company, he started mapping for the Geological Map of Italy in Sicily, the Lepini Mountains, Monticiano-Roccastrada Ridge and Marsigliana area (Orbetello).

His stratigraphical and sedimentological work on the Paleozoic formations of Southern Tuscany were particularly noteworthy, inasmuch as he was the first to define the Carboniferous stratigraphy in the Farma Valley and the the unconformity between it and the overlying Triassic Verrucano. His monography on the Paleozoic outcrops in the Farma Valley is a model of scientific synthesis especially when compared with the verbosity of some of his predecessors who had covered reams of paper on the same subject.

In December 1962 he began his university career at the University of Cagliari, where he remained for over a decade teaching Physical Geography and doing intensive research on Sardinian geology. In particular, after making an important contribution to certain problems of the Mesozoic and Tertiary, he concentrated on the Sardinian Paleozoic which at that time, was largely unknown, apart from sporadic data coming from mining activities. This research was of such depth, breadth and quality as to make it a point of reference for any subsequent study on Sardinian geology. Thus, his stratigraphical and sedimentological research on the Cambrian-Ordovician of Sardinia led to the first modern understanding of this sequence and of its tectono-sedimentary evolution.

Within the Paleozoic Working Group, in wich his exceptional intellectual and organizing qualities led to his election as national co-ordinator, he actively worked and stimulated interest in research on the Sardinian Hercynian Chain, which brought to light that Sardinia, only slightly affected by the Alpine orogeny, underwent a complex polyphasic history during the Hercynian Cycle. The recognition of an important, polyphasic tangential tectonics changed the view on the structural pattern of the Sardinian Paleozoic and led to new interpretations of important stratigraphical problems as well as of the overall characters of the Sardinian Hercynian Orogen, now considered to be a collisional chain. It is thus possible, from southwest to northeast to trace an External Zone, a slightly deformed foreland (Iglesiente-Sulcis), a Nappe Zone of green-schist metamorphic facies (central-eastern Sardinia) and an Axial Zone of amphibolitic metamorphic grade (Northern Sardinia).

In the early 1970s, without giving up his Sardinian research, Tommaso Cocozza resumed his work in Southern Tuscany together with his colleagues at the Universities of Siena, Bologna and Modena. His knowledge on the Sardinian Paleozoic led to new research on the Tuscan one, which bore certain resemblances to that of Sardinia. New Paleozoic outcrops were identified and mapped and this work was thus the the starting-point of a research project wich is still under way and yielding substantial results.

During the 1980s, he also collaborated on several occasions with the Italian cooperation programs in Somalia, as teacher at the National Somali University.

Tommaso's arrival in Siena brought renewed vitality and a more international outlook to research here. He had already been collaborating with German and American geologists for some time and thus he was one of the first to tackle the problem of the geodynamics of the Mediterranean and the rotation of the Sardo-Corsican Massif.

In fact, in this period his research on the Paleozoic of Sardinia and Tuscany gradually evolved from an analysis of detail to an analysis of the overall regional pattern. This process emerged in the various meetings and congresses of IGCP Project No. 5 and the Italian Paleozoic Working Group, of which Tommaso was national co-ordinator until his death.

Among these Meetings were:

— the scientific meeting on the Paleozoic and Italian basement, held in Siena, May 18-19, 1979;

— the field trip on the Sardinian Paleozoic, celebrating the first centenary of the "Società Geologica Italiana", May 28-31 1982;

— the scientific meeting held in Siena, December, 13-14, 1985, on the stratigraphic, tectonic, metamorphic and magmatic evolution of the Italian Paleozoic;

— the Final Meeting of IGCP Project No. 5 held in Cagliari, May 26-31 1985.

The Congress held in Siena on the 21-22 March 1991 has been organized with the intent of carrying on the work that our friend engaged with such enthusiasm and care. The large number of partecipants and the high quality of the scientific communications, collected as papers in the present volume, made the Congress an event worthy of commemorating him.

His friends from the Dipartimento di Scienze della Terra, Università degli Studi di Siena.

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CHARACTERS OF THE CAENOZOIC SEDIMENTARY AND VOLCANIC SUCCESSION OF WESTERN SULCIS (SW SARDINIA)

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Recent investigations of the Sulcis Tertiary basin have been carried out mostly on cores drilled by Carbosulcis Spa. The results of the volcanological, stratigraphical and structural analysis of surface and sub-surface sequences are here described. From bottom to top, the Tertiary basin was filled by Palaeogene sediments, a complex of Oligo-Miocene calc-alkaline volcanics and Neogene fluvial and fluvio-lacustrine deposits (Fig.3).

The pre-Eocene substratum consists of Permian-Triassic sediments with intercalations of volcanics and volcanoclastics, or of the Palaeozoic basement.

The Palaeogene group, from bottom to top is composed of: Early Eocene (Cappetta & Thaler, 1974; Agus & Pecorini, 1978; Cherchi, 1979; Pittau Demelia, 1979) mainly calcareous, lignite-bearing sediments deposited in marine-lagoonal to lacustrine-marshy environments; Middle Eocene-?Early Oligocene mainly reddish and coarse, clastic deposits (Cixerri Formation: Pecorini & Pomesano Cherchi, 1969; Barca & Palmerini, 1973) of fluvial and lacustrine environments, predating the volcanic activity of 29-27 Ma

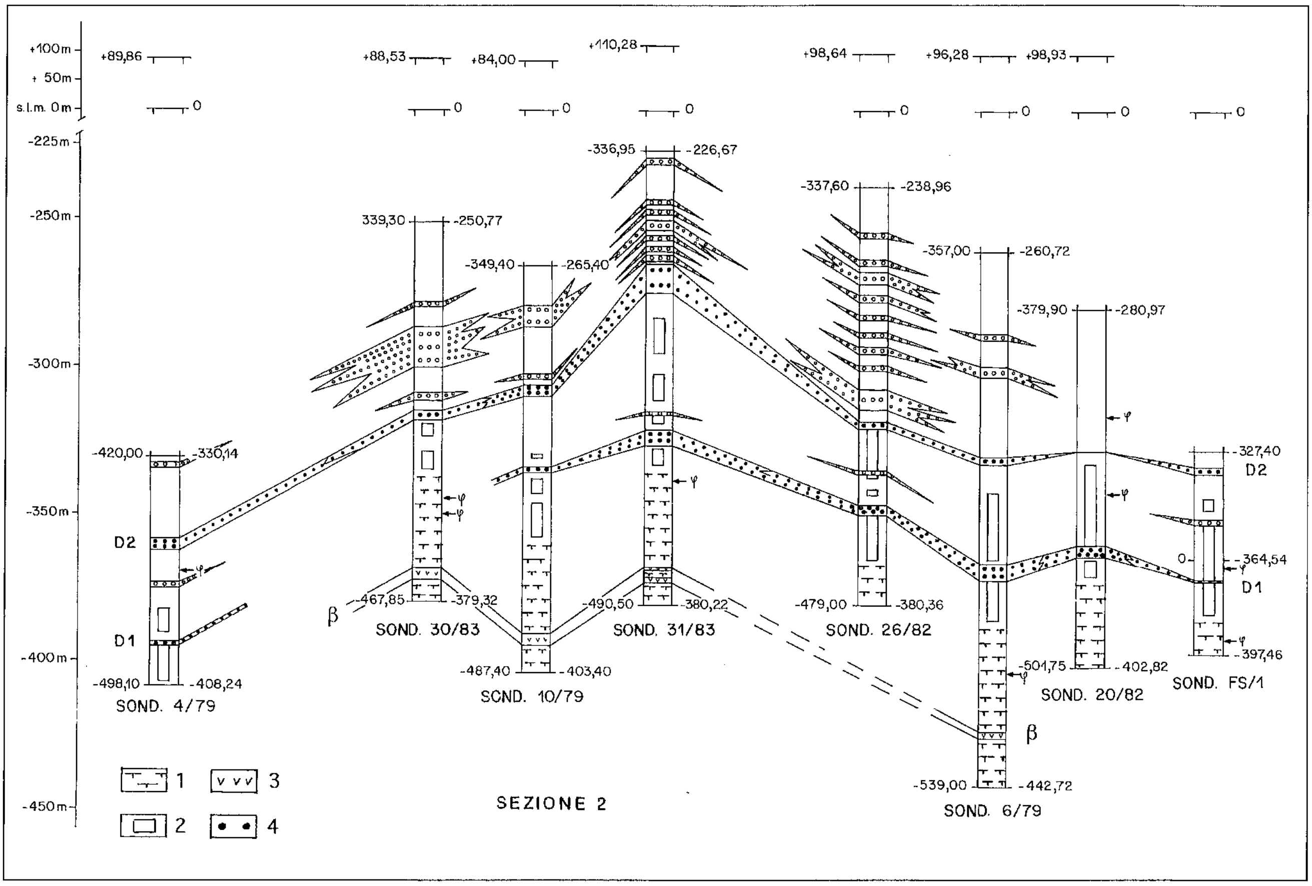

Fig. 1 - Bore-hole logs along cross-section 2 (see Fig.2 A) correlated on the base of ß, D1 and D2 marker beds.
Legend of symbols: 1 - "Calcare a Miliolidi"; 2 - Lignite bearing beds; 3 - ß marker bed (basalt); 4 - D1 and D2 marker beds (microconglomerate);5 - y faults.

(Bellon, 1976; Savelli et al., 1979; Beccaluva et al., 1985). This group can be divided into four units (Fig.1) which, in ascending order, are:

— A polygenic conglomerate unit, made up of well sorted gravels of limestone (partly of Cretaceous age) and less commonly quartz, Palaeozoic chert and schists. The irregular areal distribution of the conglomerate bodies suggests that they represent the distal part of an alluvial fan system.

— The "Calcare a Miliolidi", characterized by thick layers of pure limestone (Miliolid packstone) and thin intercalations of marly limestone (mudstone/wackestone with paralic organisms). The sedimentological features and the faunal content of the different facies indicate a coastal depositional system under warm climatic conditions, characterized by adjacent hyperhaline and mesohaline lagoons. The sequence represents a complete transgressive-regressive cycle, the end of which is marked by a layer of rather altered alkali-basalt (ß marker bed; Fig. 1) which in its upper part, shows evidence of pedogenetic processes.

— The mixed carbonate-siliciclastic, lignite-bearing unit ("formazione argilloso-calcareo arenacea con lignite"), characterized by a variety of lithologies including gray limestone (wackestone/packstone with Miliolids and paralic organisms), pink to white marly limestone (mudstone/ wackestone with lacustrine and terrestrial organisms), gray to black clay, siltstone, sandstone, lignite beds, and microconglomerates. Even if these lithologies recur irregularly and with variable thickness in the sequence, it is possible to distinguish three main depositional cycles composed of shallowing-upward low-order cycles that often end with pedogenetic features.The upper cycle (C3) is marked at the base and top by microconglomerate marker beds (Fig. 1). The oldest of them (D1) is characterized by the occurrence of rhyolitic clasts whereas the yunger one (D2) contains also basaltic-andesitic clasts.

— The Cixerri Formation, consisting from bottom to top, of: sandstone and polygenic conglomerates, siltstone and shale and local lenses of lacustrine limestone, with gastropods and *Chara* remains.Its basal part corresponds to the C4 sequence (Fig.1,3). All these sediments are characteristic of a fan and an alluvial-plain depositional system, fed from a source area located in the west.

The Tertiary volcanics encountered in Bore-hole 37 (South East of Nuraxi Figus) are approximately 450 m thick (Fig.2) and, from bottom to top, include the following units:

1) a 10 m thick layer of andesitic basalts constituted by Pl + Opx + Cpx + 01, that are interbedded within a continental terrigenous sequence;

2) the Corona Maria unit, consisting of generally welded ignimbrites with dacitic composition and phenocrysts of Pl + Opx + Cpx + 01; the average thickness is 40 m;

3) the Lenzu unit, consisting of welded, glassy ignimbrites rich in Pl - Kf + Opx + Cpx phenocrysts, with a thickness of 7 m;

4) the Acqua Sa Canna unit, consisting of several layers of dacitic ash and pumice flow, constituted by Pl + Bt; the thickness range up to 15 m;

5) the Seruci unit, consisting of welded, vesiculated ignimbrites with rhyolitic composition and porphyritic texture due to Pl + Kf + Ol phenocrysts; thickness is 30 m;

6) the Conca Is Angius unit, consisting of weakly welded ignimbrites constituted by vesciculated fragments embedded in a cineritic matrix rich in Pl, Kf and weathered pyroxenes; thickness is 6 m;

7) the Nuraxi unit, consisting of weathered, highly welded rhyolitic ignimbrites, with porphyritic texture due to Pl + Kf + Px phenocrysts. These ignimbrites probably form the base of the ignimbrite sequence of San Pietro Island (Garbarino & Maccioni, 1970; Garbarino et al., 1985);

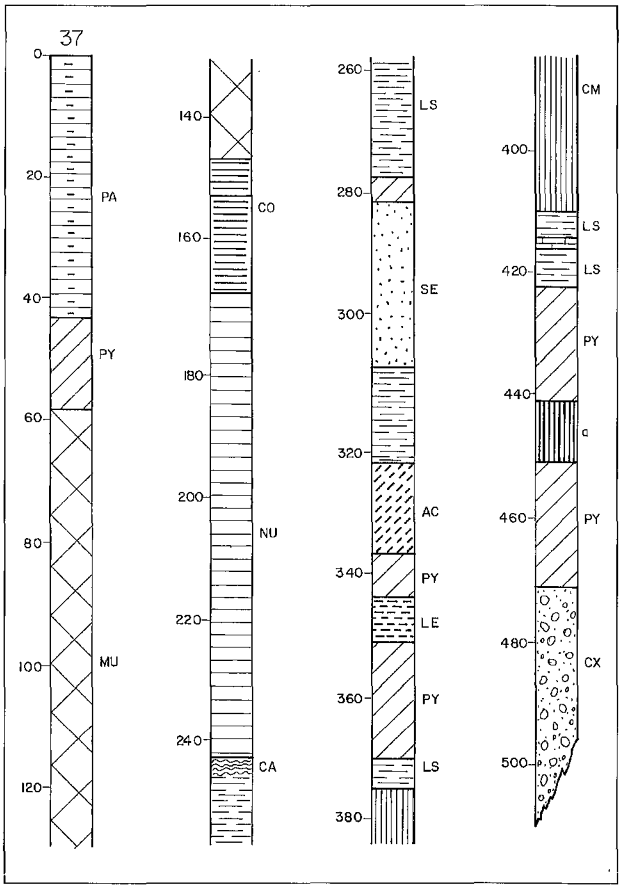

Fig. 2 - Stratigraphy of Bore-hole 37 (southeast of Nuraxi Figus). PA- Paringianu igninbrites; PY- pyroclastic and epiclastic layers; MU- Monte Ulmus igninbrites; CO- Comenditic igninbrites; NU- Nuraxi igninbrites; CA- Conca is Angius igninbrites; LS- sedimentary layers and paleosols; SE- Seruci igninbrites; AC- Acqua sa Canna igninbrites; LE- Lenzu igninbrites; CM- Corona Maria igninbrites; a- andesitic basalts; CX- Cixerri Formation.

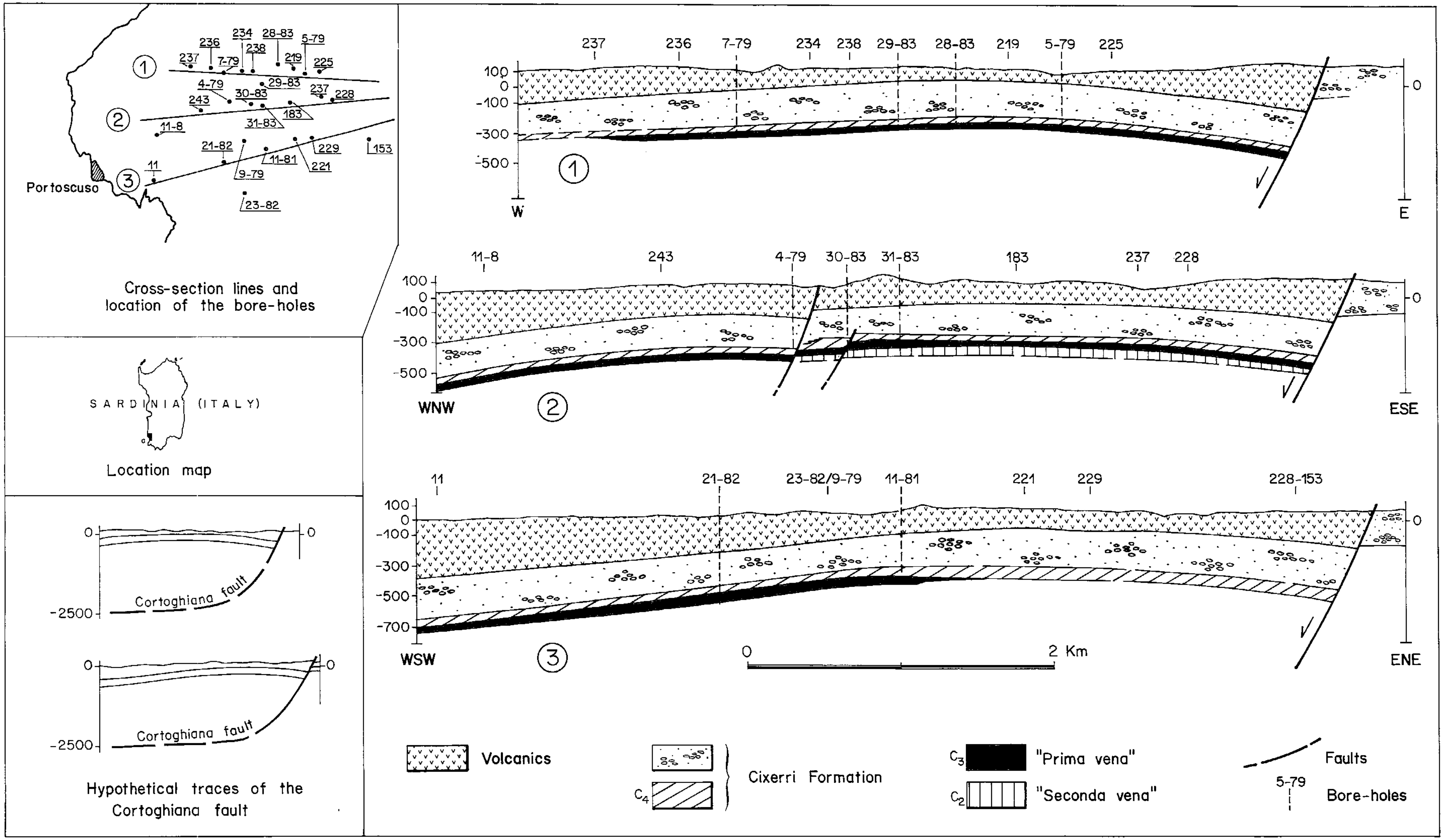

Fig. 3 - Cross sections in the investigated area.

8) a comenditic complex, characterized by several flow units, either welded or loose, composed by Oz + Kf + Aegirine + Arf; thickness is 22 m;

9) the Monte Ulmus unit, consisting of welded, vesiculated, poorly porphyritic (mostly Kf) ignimbrites, with alkali-rhyolitic composition; thickness is 80 m;

10) the Paringianu unit, consisting of a poorly welded ash flow of rhyolitic composition, characterized by Pl + Kf phenocrysts.

The Tertiary Sulcis Basin began to develop from the beginning Early Eocene as a consequence of extensional events that later deformed also the sedimentary formations and volcanic cover.

Based on the results of deep drilling analysis, Eocene normal growth faults have been recognized (Fig.3). Their activity produced "semigraben" structures and thickness variation of coeval sedimentary deposits. The faults that cross-cut the Miocene volcanics caused rotation of the sedimentary and volcanic layers as documented by E-dipping hanging-wall volcanics along the W-dipping Cortoghiana normal fault. The geometry of the hanging-wall, reconstructed using bore-hole data, shows that these volcanics belong to the eastern limb of the Seruci anticline which developed as a consequence of the rotational movements along the Cortoghiana fault. It is here suggested that the geometry of the Cortoghiana fault is listric and that the Seruci anticline and the Cortoghiana fault both result from the same extensional event which followed the deposition of the Oligo-Miocene volcanics.

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UPPER OLIGOCENE-LOWER MIOCENE SEQUENCES OF THE ARBUS-FUNTANAZZA COAST (SOUTH-WESTERN SARDINIA, ITALY)

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Abstract

The complete Upper Oligocene-Lower Miocene marine succession cropping out along the Arbus coast, in the Tertiary Funtanazza graben, has been analyzed from a stratigraphic point of view for the first time. The succession is structurally connected with a monoclinal dipping NNW (around 10 degrees) intersected by andesitic-basaltic dykes without important fault throws.

The four selected sequences constituting the whole succession are located, respectively:

1) in the Gutturu Flumini area (A-A'); 2) on top of the Miocenic basaltic laccolith breaking the sequences continuity (B-B'); 3) in the Cala Campu Sali area (C-C'); 4) close to the Colonia Sartori (D-D').

These sequences represent the most distal marine facies outcropping in the whole Funtanazza area.

Evidences of transgressive (three) and regressive (two) episodes, different in importance, exist in such sequences.

Given the stratigraphic relationship and the paleontological data, the first transgressive event could be ascribed to the Upper Oligocene, while the other two would be Aquitanian in age. Moreover, the most ancient regressive phase should be placed at the Oligocene-Aquitanian boundary, while the successive regression separates the two Aquitanian transgressive events.

In the sedimentary basin the two sea level regressive variations are responsible for the presence of lacustrine episodes in the proximal areas, located eastward to the coast. In the Funtanazza marine succession,however, they would be documented by the presence of two conglomeratic levels that are well exposed in the Calada Bianca section (Fig. 1).

The Funtanazza stratigraphic succession shows a sedimentary evolution thast is considerably influenced by a coeval volcano-tectonic activity related to the Oligo-Miocene rift system of the western Mediterranean area. From a lithostratigraphic point of view, five units have been distinguished and described from bottom to top as follows:

UNIT A (level 1-8 in the sequence D-D') - It consists of continental and transitional deposits, characterized by paleosoils and fluvial conglomerates; pumice-and-ash pyroclastic flows; fetid limestones with *Planorbis sp.* and vegetal fragments; thin carbonaceous levels.

UNIT B (level 9 in the sequence C-C') - Littoral and infralittoral deposits, constituted by fine-grained biocalcarenite .

UNIT C (level 1 in the sequence A-A'; 10-12 in the sequence B-B') - Ash pyroclasic deposits of both submarine and transitional environment.

UNIT D (level 2 -23 in the sequence A-A' and 1-18 in the sequence D-D') - Infra and circalittoral deposits consisting of a massive level of fine-grained biocalcarenite and silty- clayey tuffite.

UNIT E (levels 18-53 in the sequence D-D') - Littoral, infralittoral and circalittoral deposits represented by alternating ash pyroclastite and pumiceous tuffite, organogenic limestone with *Lithothamnium sp.*, coarse and fine-grained biocalcarenite, tuffaceous sandstone and microconglomerate.

Key words: *Stratigraphy, Paleoecology, Upper Oligocene-Lower Miocene, Sardinia, Italy.*

Riassunto

E' stata studiata la successione lito- biostratigrafica affiorante nella marina di Arbus-Funtanazza. La successione costituisce una monoclinale che si immerge a N-NW di circa 10 gradi ed è attraversata da filoni andesitico- basaltici che solo raramente determinano rigetti apprezzabili. I quattro segmenti di successione campionati sono esposti rispettivamente in località Gutturu Flumini (A- A'), a tetto del laccolite basaltico intramiocenico che interrompe la continuità della sequenza (B-B'), a Cala Campu Sali (C-C') e, infine, in prossimità della "Casa al mare Sartori" (D-D').

Le sequenze studiate rappresentano le facies marine più distali del bacino corrispondente alla fossa tettonica oligo- miocenica di Funtanazza.

Nella presente nota vengono segnalate, in seno alla successione di Funtanazza, variazioni relative del livello del mare, sia trasgressive (tre) che regressive (due), di diverso grado di importanza. Sulla base dei rapporti stratigrafici e dei dati paleontologici sinora acquisiti, il

primo evento trasgressivo apparterrebbe all'Oligocene superiore, mentre gli altri due sarebbero di età intraquitaniana. La fase regressiva più antica andrebbe confinata al limite Oligocene-Aquitaniano, mentre la successiva regressione separa i due eventi trasgressivi intraquitaniani.

Le due variazioni regressive sono responsabili nei settori prossimali del bacino di sedimentazione, situati più ad est, della presenza di episodi lacustri (ASSORGIA et al., 1988). Nella sezione a mare di Funtanazza esse sarebbero invece documentate da due livelli conglomeratici ben esposti nella falesia di Calada Bianca.

La successione stratigrafica esaminata mostra un'evoluzione sedimentaria strettamente influenzata dalla concomitante attività vulcano-tettonica riconducibile al rifting oligo- miocenico dell'area mediterranea occidentale (CHERCHI & MONTADERT, 1982).

Dal basso verso l'alto vi sono state distinte cinque differenti unità litostratigrafiche:

UNITA' A - Depositi continentali e di transizione, caratterizzati da paleosuoli commisti a conglomerati fluviali; piroclastiti cineritiche e pomicee; calcari fetidi a *Planorbis sp.* e frustoli vegetali; letti carboniosi.

UNITA' B - Depositi marini litorali e infralitorali, costituiti da biocalcareniti fini.

UNITA' C- Depositi piroclastici cineritici, marini e di transizione.

UNITA' D - Depositi marini infra e circalitorali, composti da banchi di calcareniti fini a Turritellidae; calcareniti fini e tufiti siltoso-argillose.

UNITA' E - Depositi marini litorali, infralitorali e circalitorali rappresentati da piroclastiti cineritiche alternate a tufiti pomicee; calcari organogeni a *Lithothamnium sp.*; biocalcareniti grossolane e talora fini; arenarie e microconglomerati tufacei.

PAROLE CHIAVE: *Stratigrafia; Paleoecologia; Oligo-Miocene; Sardegna; Italia*

1. INTRODUCTION

The present work concerns to the detailed lithostratigraphic definition of the Funtanazza Lower Miocenic succession, to the identifications of the existing thanatocoenosis and to the interpretation of the paleobathymetry of the different units they are included in.

The researches are based on the physical (granulometries, petrographic thin sections, etc.) and chemical (carbonates value) rock classification and on the study of fossil associations, sampled with quantitative and/ or semiquantitative methods.

These researches are parts of a project concerning the ecobiostratigraphy and the paleogeographical and paleobiogeographical evolution of the Sardinian Tertiary sedimentary basins in relation to coeval volcanic and tectonic events.

2. VOLCANIC AND SEDIMENTARY RELANTIONSHIPS WITHIN THE FUNTANAZZA AREA.

The detailed study of litho-biostratigraphic sequences permitted to define the relationships between the marine and continental Upper Oligocene (?) - Lower Miocene sedimentation(N4-N6,N7(?);BLOW,1969) (CHERCHI,1974; SMIT,1974) and the volcanic activity related to the Sardinian Cenozoic tectonic phase within the Funtanazza area.

The volcanic events that took place in that area can be considered as parts of the calcoalkaline volcanic cycle developed in Sardinia between 33 and 11 Ma (COULON, 1977; SAVELLI et al., 1979; BECCALUVA et al., 1985). Also, they can be related to the Sardo-Corsican drift with a oceanic crust subduction NW dipping (BECCALUVA et al., 1987 and references). Absolute dating, performed with the K/Ar method (ASSORGIA et al., 1985), permitted to recognize, in the area included between Funtanazza and Monte Arcuentu, a succession of volcanic events whose age is included between 30 and 16 Ma. The first events are essentially lavic and show prevailingly dome structures whose emplacement took place before the first deposition of marine sediments paleontologically referable to the Upper? Oligocene-Lower Aquitanian.

Also, rhyolitic-rhyodacitic pyroclastic products, related to the rifting that generated the large Fossa Sarda (VARBABASSO, 1962; CHERCHI & MONTADERT, 1984), are found. They precede and are in part concomitant with the Miocene marine sedimentation. This important Oligo-Miocene extensional phase in the Funtanazza graben is also supported by the presence of fluvial-lacustrine continental deposits that precede the marine ingression (BARCA, 1973) and that are associated, in part at least, with subaerial acid pyroclastic products.

In the Funtanazza graben, large banks of acid pyroclastic rocks croup out both over and under the first Miocene marine sediments, represented by fine-grained biocalcarenites with *Pereiraia gervaisi* Vezian.

To these first explosive volcanic events, a submarine basic lava activity follows, documented by a sedimentary-volcanic succession made of fossiliferous tuffites in which pillows flows and hyaloclastic breccias with a total thickness of around 70 m are intercalated.

Moreover, such a volcano-sedimentary succession was interested by the repeated emplacement of numerous Aquitanian-Burdigalian basaltic or andesitic-basaltic lava dikes that can be ascribed to the submarine lavic activity. A terminal effusive activity, Upper-Burdigalian in age, took place afterwards (ASSORGIA et al., 1985).

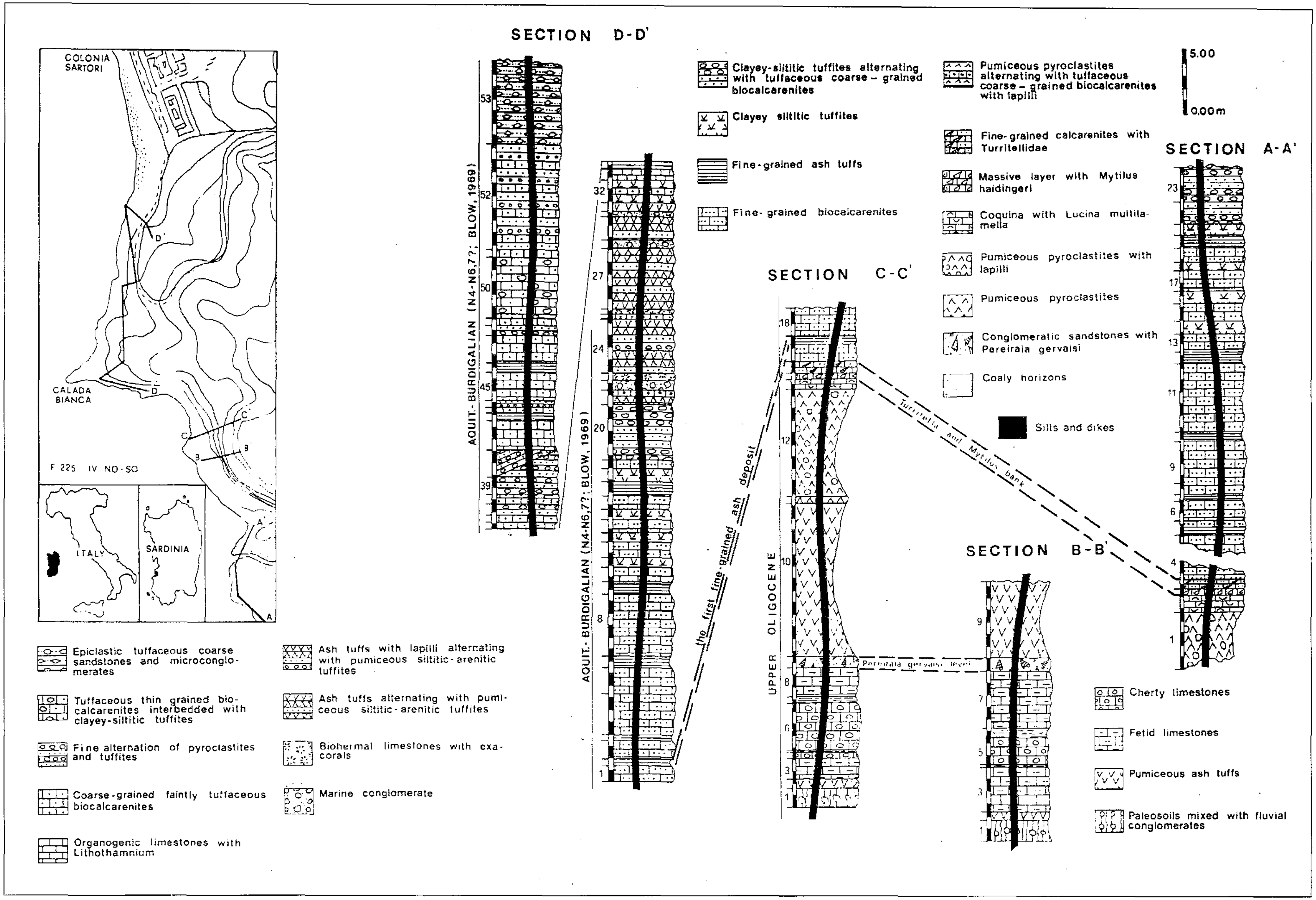

Fig. 1 - Location and lithostratigraphic correlation of the studied sequences

3. Funtanazza Sea-Side Section

The examined Funtanazza sequence is exposed along the Sardinian western coastline (Fig. 1); it (sequence c-c') begins in the geographical coordinates point 3 59' 10 '' W Monte Mario longitude and 39 36' 20" N latitude, and ends nearby the Sartori's beach House in the longitude point 3 59' 12" W Monte Mario and the latitude point 39 36' 56" N(sections C-C).

Different toponyms exist in literature, such as: "Baia di Funtanazza", "Calada Bianca", "Cala Campu Sali", "Is Argiolas Mannas", "Zona costiera fra Calada Bianca e Cala Campu Sali" (COMASCHI CARIA, 1963; SECCHI, 1982). They indicate partial sections of the surveyed sequence. The whole sequence had already been proposed in its essential lines by SPANO (1987).

The section B-B' of fig. 1 is almost totally included in the C-C' section of the same figure. The section A-A' instead is totally included in the upper part of the section C-C' and in the part that is stratigraphically lower than the D- D' one. The data concerning the section B-B' are not quoted for brevity reasons.

All the overlapping segments of the sections show lithostratigraphic and paleoenvironmental parallel trends, as shown in the tables 1,2 and 3 hereafter.

Five lithostratigraphic units are here distinguished: A- B- C- D-E.

3.1. *Unit A*

It includes the levels 1-7 of the section B-B' and 1-8 of the section C-C' (Tab.2). For a thickness of about 11 metres it is characterized by alternations of fetid limestones with *Planorbis* and plant remains, carbonaceous horizons and nodular limestones. The fossil associations and the lithotypes are typical of the transitional and lacustrine environments.

3.2. Unit B

It is present in the sections B-B' and C-C' in correspondence of the respective levels 8 and 9 (Tab. 2).

The fine-grained biocalcarenites it is made of, provided a rich "mixed assemblage". Here, among others, the following species prevail: *Pereiraia gervaisi* VEZIAN, *Ostrea edulis lamellosa* BROCCHI, *Cubitostrea frondosa* (DE SERRES), *Ancillaria glandiformis* LAMARCK, *Lucina multilamella* DESHAYES, *Glycymeris inflata* (BROCCHI), *Antale fossile* SCHROETER, *Protoma cathedralis* (BRONGNIART), *Tympanotonos margaritaceous* BROCCHI, *Pleurotoma rotata* (BROCCHI) and *Turritella terebralis* LAMARCK.

The thanatocoenosis consists of a mixture of infra-circalittoral species (*Lucina multilamella, Glycymeris inflata, Cubitostrea frondosa, Ancillaria glandiformis, Antale fossile*) and of more littoral forms (*Pereiraia gervaisi, Ostrea edulis lamellosa, Tympanotonos margaritaceous*). They are related to the presence of important river mouth (Turritellids in large quantities).

The allochthony of the most littoral forms is well documented by the presence, in the overlying ash pyroclastic deposits, of copious *Pereiraia* individuals removed from the littoral area during the volcanic events.

The *Pereiraia* (Strombacea) individuals reworking is suggested by their "regular" and rich presence in the underlying level and by the desordened disposition withing the pyroclasyic level.

The paleobathymetry concerning the deepest species is estimated to be around 20-30 metres (Tab. 2).

3.3. *Unit C*

This unit is characterized by ash pyroclastic deposits corresponding to the levels 1; 9; 10-12 of the sequences A-A'; B-B'; C-C' respectively (Fig. 1; Tab. 1,2).

They are massive, yellowish-grey or sometimes greenish-grey in colour. The sandy cineritic matrix includes polygenic centimetric lithics (Paleozoic rocks; basaltic-andesitic rocks), copious free crystals and clear pumices of centimetric dimension and chipped contour rare black scoriae.

From a microscopical point of view, the rock shows a vitroclastic structure marked by the presence of cuspidate glassed septii almost always cementing fragmented crystals. Among these, it is possible to recognize different minerals such as quartz, which shous a prismatic-bipyramidate contour sometimes openly reabsorbed, and plagioclase (An content of about 30%), almost always geminated according to the Albite Law. The K-feldspar is the last sialic component; generally it comes in single openly zoned crystals, rarely geminated according to the "Karlsbad Law". The only femic phase is represented by the biotite with an intense pleochroism that can vary from dark-brown to straw- yellow. The crystal content in the pyroclastic rock varies and it results concentrated in the flow basal layer.

The examined stratigraphic relationships in the Funtanazza graben permit to suppose the deposition of this pyroclastic flow within a shallow water marine basin corresponding with an internal platform, as shown by the Strombidae thanatocoenosis quoted in the enclosed tables. The volcanic material volume deposited was so large to fill it up almost totally. The pyroclastic flow reworked during the explosion event, mixing the most littoral benthic macrofaunas them to the deepest infralittoral ones. Later, the joined action of the relative sea level changes and the subsidence of the area was so important to permit initially the marine sedimentation of fine-grained calcarenite facies and platform marls, and in more littoral coarse calcarenite

UNITS	SEMPLES	THANATOCOENOSIS	LITHOLOGY	SEDIMENTARY STRUCTURES	RELATIVE CHANGE OF SEA LEVEL	BATHYMETRIC ZONES	ESTIMATED PALEOBATHYMETRY (m)
D	23	Entale badense, Anadara diluvii, Venus multilamella, Chlamys northmptoni, Entale dentale, Cardium arcella, Astraca carinata, Corbula gibba, Megaxinus sp., Myrtea spinifera, Arcar tauroclathrata, Dentalium bonei geniculatum, Ceratrotrochus duodecimcostatus, Dentalium mutabile, Arca turonica	Siltitic-clayey tuffite alternating with tufaceous coarse biocalcarenite.			CIRCALITTORAL	50 - 80
	21 18 16 14	Schizaster lovisatoi, Spatangus marmorae, Chlamys northamptoni Amusium destefanii.	Siltitic-clayey tuffite				
	15 13 11 6 4	Chlamys nortamptoni, Amusium destefanii, Cardium turonicum, Meretrix rudis, Cardium vidali, Arca turonica, Arca umbonaria Dentilucina borealis, Lucina fragilis, Tapes eremita, Corbula gibba curta, Cardita crassa, Myrtea spinifera, Venus multilamella, Dentalium bonei, Columbella nassoides, Ormastralium carinatum, Flabellipecten burdigalensis, Schizaster lovisatoi, Schizaster desori	Fine-grained biocalcarenite				
	3	Turritella terebralis, Archimediella archimedis dertoniator, protoma cathedralis, Turritella turris, Turritella tricarinata, Haustator vermicularis, Lucina multilamella, Haustator desmaretinus, Ostrea edulis lamellosa, Chlamys northamptoni, Ormastralium carinatum.	Fine-grained biocalcarenite ("Turritella bank")				
	2bis	Mytilus haidingeri	Fine-grained biocalcarenite "Mytilus haidingeri bank"			INFRALITTORAL	2 - 5
	2	"Coquina" with Lucina multilamellata	Coarse-grained biocalcarenite	Massive		INFRALITTORAL	10
C	1		Ash pyroclastic deposits			CIRCALITTORAL	see unit C of section C - C'

LEGEND: = clean lamination; = indistinct lamination; = warvy lamination; = large scale cross stratification; = cross bedding lamination; = nodular structure; = flat clean limit; = erosional surface; = normal graded bedding; = reverse graded bedding; = enclosed pelitic fragments; = reworked enclosed block; = bioherm; = vertical track; = burrows of "Lithodomus"; = carbonized vegetal fragments; = gas pipe; = sea level positive variation; = sea level negative variation; = standing sea level.

Tab. 1 - Dominant thanatocoenosis and paleobathymetric meaning of the section A-A'.

facies later on. In these sediments, an acid volcanic component is always present (K-feldspar and biotite) probably because of the introduction, in the sedimentary basin, of the pyroclastic-flow erosion products that must have covered a large area of the inland under the lee of the marine Funtanazza rift.

3.4. *Unit D*

The following levels belong to this unit: 2 bis -23 of the section A-A'; 13-18 of the section C-C'; 1-17 of the section D-D' (Fig. 1). The total thickness is of about 31 m.

It begins in the section A-A' and C-C' with a "*Lucina*

UNITS	SAMPLES	THANATOCOENOSIS	LITHOLOGY	SEDIMENTARY STRUCTURES	RELATIVE CHANGE OF SEA LEVEL	BATHYMETRIC ZONES	ESTIMATED PALEOBATHYMETRY (m)
D	15	Turritella terebralis, Archimediella archimedis dertoniator, Protoma cathedralis, Turritella turris, Turritella tricarinata, Haustator vermicularis, Lucina multilamella, Haustator desmasretinus, Ostrea edulis lamellosa, Chlamys northamptoni, Astraea carinata	Fine-grained biocalcarenite: ("Turritelle Bank")			CIRCALITTORAL	50 - 80
	14	Mytilus haidingeri	Fine-grained biocalcarenite ("Mytilus haidingeri Bank")			INFRALITTORAL	2 - 5
	13	"Coquina" with Lucina multilamella	Coarse-grained biocalcarenite	massive		INFRALITTORAL	10
C	10-12		Ash pyroclastic deposits			CIRCALITTORAL	20 - 30
B	9	Pereiraia gervaisi, Ostrea edulis lamellosa, Cubitostrea frondosa, Ancillaria glandiformis, Lucina multilamella, Glycymeris insubrica, Chlamys haueri, Dentale fossile, Archimediella archimedis dertoniator, Protoma cathedralis, Tympanotonos margaritaceus, Pleurotoma rotata, Turritella vermicularis lineatocincta, Turritella terebralis	Fine-grained biocalcarenite			CIRCALITTORAL	20 - 30
A	8	Planorbis sp., and plants remains	Fetid limestone remains		LACUSTRINE	TRANSITION	
	7	Lignite	carbonaceous horizon				
	6	Planorbis sp. and plants remains.	Nodular limestone				
	5	Lignite	carbonaceous horizon				
	4	Planorbis sp. and plants remains	Nodular limestone				
	3	Planorbis sp. and plants remains	Fetid limestone				

Tab. 2 - Dominant thanatocoenosis and paleobatymetric meaning of the section C-C'

multilamella coquina", which is followed by the "Mytilus haidingeri bank" and the "T*urritella bank*"; and than by an alternation of fine-grained ash tuffs, fine-grained biocalcarenite and tuffites.

The fossiliferous content in the "*Mytilus haidingeri bank*" and "*Lucina multilamella* coquina" are represented by oligotypical malacofaunas; while in the "*Turritella bank*" there is an extraordinary number of individuals belonging to a few species of *Turritella*, among which, in abundance order: *Turritella terebralis* LAMARCK, *Archimediella archimedis* dertoniator (SACCO), *Protoma cathedralis* (BRONGNIART), *Turritella turris* (BASTEROT), *Turritella tricarinata* (BROCCHI), *Haustator vermicularis* (BROCCHI) and *Haustator desmarestinus* (BASTEROT). Associated to them there are above all: *Lucina multilamella* DESHAYES, *Chlamys northamptoni* MICHELOTTI, *Ostrea edulis lamellosa* BROCCHI and *Astraea carinata* (BORSON). Some individuals of *Vaginella austriaca* KITTL come from this unit too.

The other fossiliferous levels regard the fine-grained biocalcarenites in which the thanatocoenosis is dominated by Pectinides with *Chlamys northamptoni* MICHELOTTI, whose individuals constitute more than the 50% of the whole microfauna, *Chlamys haneri* MICHELOTTI, *Amusium destefani* (UGOLINI), *Chlamys burdigalensis* LAMARCK associated, among others, with a large number of specimens: *Cardium arcella* DUJARDIN, *Meretrix rudis* (POLI), *Arca turonica* DUJARDIN, *Myrtea spinifera* (MONTAGU), *Venus multilamella* (LAMARCK),

UNITS	SAMPLES	THANATOCOENOSIS	LITHOLOGY	SEDIMENTARY STRUCTURES	RELATIVE CHANGE OF SEA LEVEL	BATHYMETRIC ZONES	ESTIMATED PALEOBATHYMETRY (m)
	53	Mytilus haidingeri and other reworker faunas	Microconglomeratic sandstone and tufaceous-arenaceous microconglomerate			MESOLITTORAL-INFRALITTORAL	0 - 5
	52	As in the coarse-grained horizons of level 36 As in the fine-grained horizons of level 37	Alternating coarse grained biocalcarenite and tuffite			INFRALITTORAL CIRCALITTORAL	15 - 50
	51	As in the level 36	Coarse-grained biocalcarenite				
	50	As in the level 10 except the Bryozoa	Tuffite sometimes gravelly			CIRCALITTORAL	
	49	As in the level 36	Coarse-grained biocalcarenite				
	48	As in the level 37	Tuffite sometimes gravelly				
E	47	As in the level 36	Coarse-grained biocalcarenite				
	46	Sterile	Fine-grained ash tuffs				
	45	As in the level 36	Coarse-grained biocalcarenite				
	44	As in the level 37	Gravelly tuffite				50 - 70
	43	As in the level 36	Coarse-grained biocalcarenite				
	42	Sterile	Fine-grained ash tuffs				
	41	As in the level 36	Coarse-grained biocalcarenite				
	40	Reworked faunas	Tuffite				
	39	As in the level 37	Pyroclastic and tuffitic alternances				

Tab. 3 - Dominant thanatocoenosis and paleobathymetric meaning of section D - D'

Schizaster lovisatoi COTTEAU, *Schizaster desori* WRIGHT, *Cheilostomata, Echinolampas sardiniensis* COTTEAU. The tuffittic levels essentially show the same faunas as the ones seen in the previous lithotype. The Bryozoa, existing in allmost unit D, become particularly important in these levels thanks to the abundance of individuals and to shape variety.

Almost exclusively reworked paleofaunas have been found in the fine-grained ash tuffs.

A paleobathymetry included between 2-5 m, which can be referred to the "*Mytilus haidingeri* bank", and 50-80 m, referable to the Pectinides paleofaunas, can be attributed to those deposits.

3.5. *Unit E*

This unit is only exposed in the section D-D' starting from level 18. Its thickness is of about 64 m.

It consists of numerous lithotypes among which coarse-

	38	As in the level 36	Coarse-grained biocalcarenite				
	37	In the tuffite horizons: Bryozoa and other reworked faunas	Pyroclastic and tuffite thin alternances				
	36	Cyclostomata, Cheilostomata, Pecten corsicanus, Ostrea forskalii Ostrea edulis lamellosa, Cubitostrea frondosa, Porifera, Chlamys northamptoni, Echinoidea, Chlamys scabriuscula, Chlamys multistriata	Coarse-grained biocalcarenite				
	33	Gryphaea gryphoides, Nerita fumata, Ostrea gingensis, Ostrea edulis boblayei, Clypeaster sp., Lithotamnium sp.	Lithotamnium limestone				
	32	In the tuffitic horizons: Macropneustes saheliensis, Spatangus equidilatatus, Chlamys northamptoni, Cuspidaria cf. cuspidata Glycymeris insubrica, Pecten revolutus, Drillia crebicosta Mitra fusiformis. In the coarse-grained biocalcarenite horizons: As in the coarse-grained horizons of the level 20 plus Pecten corsicanus and Pecten revolutus	Alternating tuffite and biocalcarenite			INFRALITTORAL /CIRCALITTORAL	15 - 50
E	31	Reworked faunas	Fine-grained pyroclastic deposits				
	30	As in the level 28	Conglomerate				
	29	Reworked faunas	Fine-grained ash tuffs				
	28	Chlamys northamptoni, Pecten corsicanus, Pecten revolutus, Ostrea edulis lamellosa, Chlamys tauroperstriata, Ostrea forskalii, Flabellipecten burdigalensis, Anomia ephippium, Turritella turris, Amusium destefanii, Conus sp., Ficula sp. Cubitostrea frondosa, Chlamys scabriuscula, Chlamys spinulosa, Balanus tintinnambulum	Conglomerate	massive			
	27	Reworked faunas	Pyroclastic and tuffite thin alternances	massive			
	26	Bioherm with hermatypic corals: as in the level 23 but completely eroded and reworked	Bioherms				5 - 15
	25	As in the coarse-grained horizons of the level 20	Coarse-grained biocalcarenite			INFRALITTORALE	
	24	Level with single Corals and Dentalium In the tuffite horizons: Balanophyllia rovasendai, Ceratotrochus decussatus, Ceratrotrochus duodecimcostatus, Flabellum geniculatum, Dentalium badensis, Dentalium mutabile, Amusium destefanii, Chlamys scabriuscola	Pyroclastic and tuffitic alternances			CIRCALITTORAL	50 - 80

Tab. 3 - (continued)

grained biocalcarenites, tuffites and pyroclastic and tuffitic alternances prevail (Tab. 3). Conglomeratic , microconglomeratic and breccioid facies follow, the latter in alternance with coarse-grained biocalcarenites. In the stratigraphically lower part of the unit two bioherme levels of hermatypic Corals are present, while a red Algae limestone exists, corresponding to level 33.

The coarse-grained biocalcarenites contain fossil associations with a prevalence, in order, of *Bryozoa Cyclostomata* and *Cheilostomata*, followed by Mollusks with *Chlamys northamptoni* MICHELOTTI, *Ostrea edulis boblayer* DESHAYES, *Ostrea edulis lamellosa* BROCCHI, *Cubitostrea frondosa* (DE SERRES), *Chlamys scabriuscula* MATHERON, *Ancillaria glandiformis* LAMARCK, *Chama gryphina* LINNEO, *Amusium destefanii* (UGOLINI) and *Cardita crassa* LAMARCK.

	23	Bioherms with hermatypic Corals: Favites crenulata, Favites diversiformis, Heliastrea defrancei Heliastrea ellissiana, Heliastrea rosacea, Solenastrea turonensis. In the coarse-grained arenaceous matrix: Cyclostomata, Cheilostomata, Pecten corsicanus, Pecten revolutus Chlamys northamptoni, Glycymeris insubrica, Cuspidaria cf. cuspidata, Chlamys spinulosa.	Bioherms				
	22	Ostrea edulis lamellosa, Ostrea gingensis, Chlamys northamptoni Chlamys scabriscula, Pecten revolutus, Pecten corsicanus, Flabellipecten burdigalensis, Cardita crassa, Lucina borealis, Glycymeris bimaculata, Venus multilamella, Petaloconchus intortus, Cubitostrea frondosa.	Conglomerate				
E	21	In the fine-grained horizons: Natica avitensis, Glycymeris pilosa, Glycymeris insubrica, Protoma quadriplicata, Limopsis dumasi. In the caorse-grained horizons: Charonia apenninica, Ancillaria glandiformis, Amusium destefanii Chama gryphina, Natica avitensis, Chlamys northamptoni, Limopsis aurita, Drillia costae, Astraea carinata, Turritella quadriplicata, Cardita crassa, Chama gryphina.	Alternating pyroclastic deposits and caorse-grained biocalcarenite			INFRALITTORAL/ CIRCALITTORAL	10 - 50
	20	In the fine-grained horizons: Glycymeris insubrica, Glycymeris pilosa In the coarse-grained horizons: Cheilostomata, Cyclostomata, Chlamys northamptoni, Cardita crassa, Glycymeris bimaculata, Ostrea boblayei, Ostrea edulis lamellosa, Ostrea gingensis, Spatangus marmorae, Spatangus aequidilatatus, Echinolampas sardiniensis, Porifera, Echinoidea, Astraea carinata.	Tufaceous alternances				
	19	Cheilostomata, Cyclostomata, Chlamys northamptoni, Ostrea edulis lamellosa, Ostrea gingensis, Astraea carinata, Porifera, Echinoidea.	Coarse-grained biocalcarenite				
	18	As in the level except the Bryozoa	Tuffite	massive			
	17	Reworked faunas	Fine-grained ash tuffs				
	16	Chlamys northamptoni, Chlamys haueri, Echinolampas sardiniensis, Echinolampas angulatus, Macropneustes saheliensis	Fine-grained biocalcarenite				
D	15	As in the level 10 except the Bryozoa	Tuffite			CIRCALITTORAL	50 - 80
	14	Chlamys northamptoni, Chlamys destefanii, Clypeaster zanoni, Echinolampas angulatus, Echinolampas sardiniensis, Cheilostomata.	Fine-grained biocalcarenite				
	13	As in the level 10 except the Bryozoa	Tuffite				
	12	Chlamys nortamptoni, Chlamys haueri, Echinolampas sardiniensis,	Fine-grained				

Tab. 3 - (continued)

In the alternating pyroclastic deposits and coarse-grained biocalcarenites the same associations occur within the coarsest lithotypes and within the fine-grained horizons, among the other species, there are: *Natica avitensis* COSSMANN & PEYROT, *Glycymeris insubrica* (BROCCHI), *Protoma quadruplicata* BASTEROT and *Limopsis aurita* (BROCCHI) .

The conglomeratic facies are characterized (with decreasing abundance) by rich thanatocoenosis of *Ostrea edulis lamellosa* BROCCHI, *Chlamys northamptoni* MICHELOTTI, *Pecten revolutus* MICHELOTTI, *Pecten corsicanus* DEPERET & ROMAN, *Flabellipecten burdigalensis* LAMARCK, *Cardita crassa* LAMARCK, *Glycymeris bimaculata* (POLI), *Cubitostrea frondosa* (DE SERRES), *Petaloconchus intortus* (LAMARCK) and other species.

The bioherms, on the contrary, are represented by thin deposits whose maximum thickness is of about 1 m. Its

		Macropneustes saheliensis, Amusium destefanii, Cheilostomata	biocalcarenite				
	11	As in the level 10 except the Bryozoa	Tuffite				
	10	Chlamys northamptoni, Echinolampas angulatus, Echinolampas sardiniensis, Chlamys haueri, Macropneustes saheliensis, Amusium destefanii	Fine-grained biocalcarenite				
	9	Reworked faunas	Fine-grained ash tuffs				
D	8	Chlamys northamptoni, Chlamys haueri, Amusium destefanii, Clypeaster zanoni, Echinolampas angulatus, Echinolampas sardiniensis, Macropneustes saheliensis, Cheilostomata.	Fine-greined biocalcarenite			CIRCALITTORAL	50 - 80
	7	Reworked faunas	Fine-grained ash tuffs				
	6	Chlamys northamptoni, Chlamys haueri, Echinolampas angulatus, Porifera.	Fine-greined biocalcarenite				
	5	Chlamys northamptoni, Chlamys haueri, Clypeaster zanoni, Echinolampas, angulatus, Porifera.	Fine-grained biocalcarenite				
	4	Sterile	Fine-grained ash tuffs				

Tab.3 - (continued)

organogenic component is given by an extraordinary number of hermatypic Corals belonging to a dozen of different species.

In the coarse-grained arenaceous matrix of the bioherms, among other species, the following in particular are present: *Cyclostomata, Cheilostomata, Pecten corsicanus* DEPERET & ROMAN, Pecten revolutus MICHELOTTI and *Chlamys northamptoni* MICHELOTTI.

Unit E, which was deposited in a deeper sedimentary basin, seems to be interested by different relative oscillations of the water column. The basin maximum depth is here estimated in 80 m because of the extremely abundant Pectinides paleofaunas. The minimum bathymetric limit occurs in the stratigraphically highest part of the unit and, therefore, of this whole Funtanazza sequence, concomitant of the last regressive phase in the area and of the basin infilling.

4. Relative Change of Sea Level

The Funtanazza sequence begins with lacustrine deposits documented by the unit A in the sections B-B' and C-C'(Tab. 2). They are well represented also in the inland of the studied area (ASSORGIA et al; 1982). The marine sedimentation begins with the unit B (section C-C') at a bathymetry typical of the Upper Circalittoral. The sedimentation is likely to be related to a sin-rift tectonics.

A first regressive pulsing is documented in corrispondence of the levels 2, 2bis and 13, 14 respectively of the section B-B' and C-C'. They are constituted by the " Coquina a *Lucina multilamella*" and by the " *Mytilus haidingeri* bank".

The wather column starts again to increase from level 3 of the section A-A' and from level of the section D-D' and takes place within a deeper circalittoral zone (50-80 m). This trend is also documented within the level 15 (section C-C'). From level 19 up to level 23 of the section D-D' there is a new regressive pulsing that determines infra-circalittoral facies (10-50 m).

A subsequent transgressive pulsing (level 24 of the section D-D'), with a 50-80 m bathymetry, is immediately followed (level 25 and 26) by a further negative change in sea level. It must not have been higher than 5-15 m, since those levels are constituted by significant bioherms.

The layers 27-52 of the section D-D' show, despite the internal or minor pulsing, the last prolonged transgressive phase of the Funtanazza area.

Such oscillations can be ascribed both to a local and to a more general context.

The pulsing that have been recognized within the units A- B-C-D seem to be related to local volcanic-tectonic events. The regressive phase concerning the levels 19-23 of the section D-D', instead, can be related to some geodynamic events generalized to the western Mediterranean area (CHERCHI & MONTADERT, 1984; BECCALUVA et al, 1987). Those levels are coeval with the Miocenic continental deposits of the Funtanazza inland (Assorgia et al., 1988) and reffered to the Aquitanian themselves.

5. Conclusion

Five lithostratigraphic units (A-B-C-D-E) have been recognized within the Funtanazza sequence.

The paleoenvironmental conditions of the sedimentary basin, inferred from the thanatocoenosis and from the lithostratigraphic-sedimentological characters of the investigated units,evolve from transitional in the stratigraphically lower part of the sequence (unit A) to circalittoral (units B and C) and than to infralittoral-circalittoral-middlelittoral (unit D) of the top.

The relative changes of the sea level, inferred from the evaluation of the water column thickness through the paleobathymetric meanig of the thanatocoenosis and associated lithotypes, paleobathimetry , are here partially related to local volcano-tectonic conditions and partially to more general geodynamic events.

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RELATIONSHIPS BETWEEN FOREDEEP DEPOSITS AND HERCYNIAN NAPPE BUILDING IN SOUTHEASTERN SARDINIA (ITALY)

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Abstract

Both the succession comprising the lower unit of the Rio Mulargia area, as well as the flyschoid deposits of eastern Sarrabus display lithological and sedimentological similarities with the Lower Carboniferous Pala Manna Formation. These successions are characterized by the presence of resedimented material, formed by olistoliths, olistostromes, lidite breccias and conglomerates. That representing syntectonic foredeep deposits (Culm facies) accumulated at the advancing nappe front of Hercynian chain.The flysch of Rio Mulargia area appears to have been deeply involved in the Hercynian deformation while in the Sarrabus area it is situated at the front of the allochthonous units. The nappes, in their movement toward the SW, initially involved the inner flysch basin and subsequently overroded it, thereby incorporating the flysch deep within the stack of nappes. Presently the flysch forms the structurally lowest unit (Rio Mulargia area).

The presence of lower carboniferous flysch among the Hercynian nappes (Rio Mulargia area) and on the nappe front (eastern Sarrabus) implies that, the external nappes cropping out in the central Sardinia, float tectonically on a substratum of Culm-type flysch.

Key words: *Hercynian orogeny, Sardinian basement, Culm type flysch, Hercynian tectonic.*

1. Introduction

The presence of synorogenic deposits in the Hercynian successions of Sardinia has, for some time, been the subject of discussion.

Vai and Cocozza (1974) attributed a Lower Carboniferous and/or Upper Devonian age to vast outcrops of metasandstone ("Postgotlandiano" Auct.) present in central Sardinia and in the southwestern part of the island. Later, Barca et al. (1982) assigned many of these sandstone formations to the Middle Cambrian-Lower Ordovician period based on micropalaeontological (Acritarchs) research .

The only exposures of terrigenous sediments, attribuited, by indirect evidence, to Lower Carboniferous age, were reported in the Gerrei and Sarrabus areas (Teichmuller, 1931; Barca, 1981; Spalletta & Vai, 1982; Barca & Spalletta, 1985). These were, however, limited and isolated outcrops insufficient to justify the presence of an important synorogenic sedimentation in the Sardinian Hercynian chain. Normally this type of sediment always comprises extensive outcrops at the front of the chain. Naud (1984) has challenged the existence of these deposits in Sardinian Palaeozoic successions, suggesting, rather, that Sardinia has been submerged for much of the Lower Carboniferous.

The Sardinian Hercynian chain therefore presented an anomalous situation with respect to other coeval circummediterranean chains (Pyrénées, Balearic Islands, Montagne Noire and Carnian Alps: Boyer et al., 1968; Feist & Schonlaub, 1974; Engel et al., 1978; Bourrouilh et al., 1980; Pelhâte & Mirouse, 1980; Cantelli et al., 1982; Engel & Franke, 1983; Engel & Raymond, 1983; Engel, 1984) which were characterized by an important synorogenic sedimentation.

Maxia (1984), on the basis of lithostratigraphic correlations, reported the presence of thick marine Carboniferous successions in the eastern Sulcis (Monte Calcinaio), southern Sarrabus (S of Punta Serpeddì) and Gerrei (S of Ballao) areas. Subsequently, the identification of significantly different ages of various carbonate blocks found in a terrigenous sequence exposed in southern Sarrabus (area around Pala Manna) allowed Barca (1991) and Barca and Olivieri (1991) to identify these blocks as olistoliths. Furthermore, these authors pointed out that the sedimentologic characteristics of the host terrigenous deposits englobing the blocks are typical of flyschoid deposits in Culm facies of the southern European Hercynian orogen.

These deposits have now been recognized throughout central-southern Sardinia over a wider area; synorogenic deposits have actually been identified over almost all of southern Sarrabus, eastern Sulcis and in the Lago di Mulargia area (Eltrudis, 1991).

In this paper the Culm type outcrops of eastern Sarrabus (A in Fig.1) are compared and correlated with those of the Lago di Mulargia area (B in Fig.1). This is followed with a discussion of their location and significance with respect to the Hercynian structure in these two sectors of Sardinian orogen.

2. General Geological Setting

The Sardinian Hercynian chain displays a clear-cut structural and metamorphic polarity from the SW External Zone, to the Central Nappe Zone, to the NE Axial Zone (Carmignani et al., 1980; 1982a; 1986a).

The Nappe Zone is characterized by nappes and low grade metamorphism (Fig. 1). Lithostratigraphic sequences of the various tectonic units in this zone are composed of a Middle Cambrian-Lower Ordovician pelitic-arenaceous substrate that was affected by the Sardinian Phase (Stille,

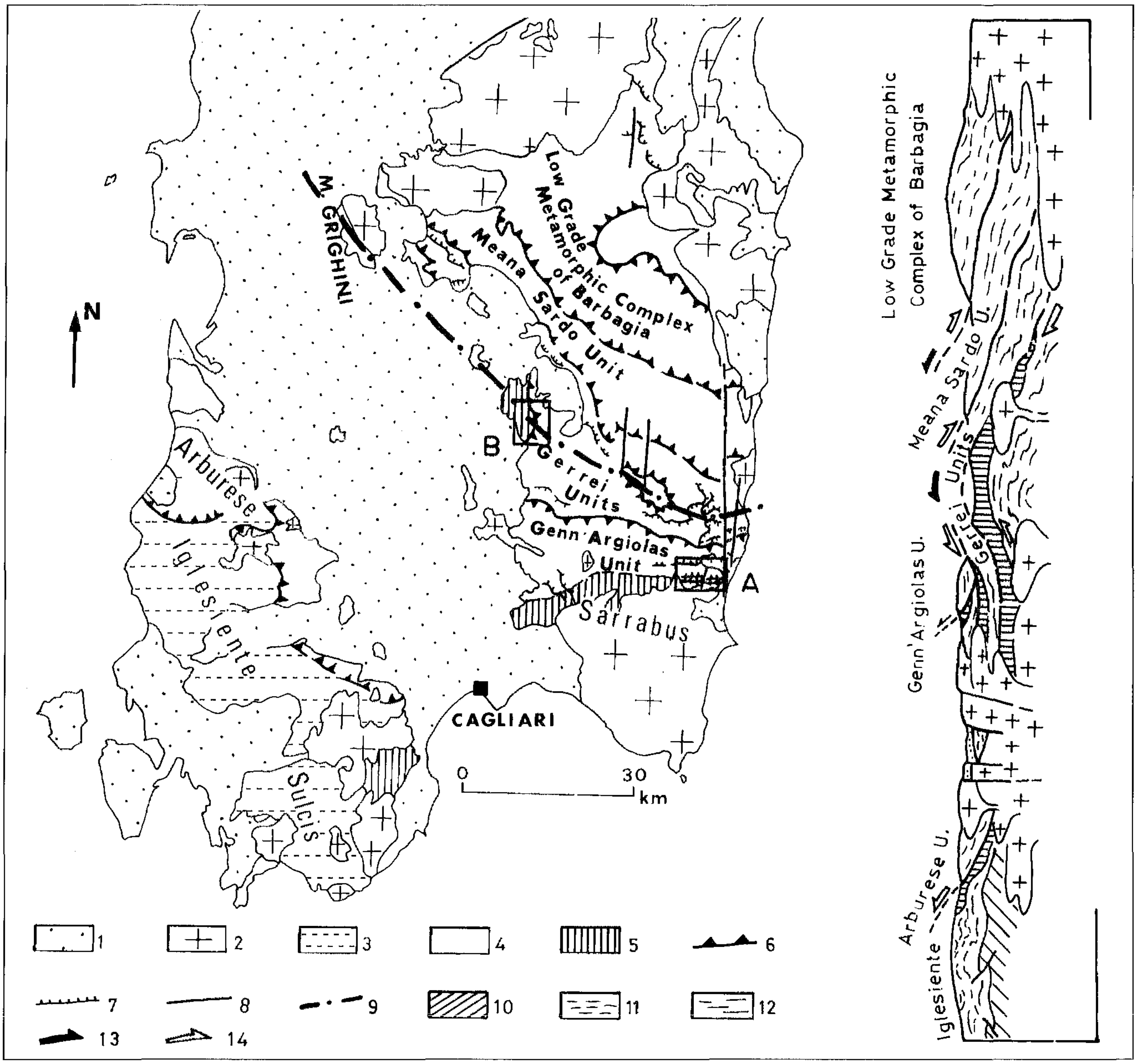


Fig.1 - Schematic structural map of south-central Sardinia. 1: Post-Hercynian sediments and volcanics; 2: Granitoids; 3: External Zone (Low to Very low grade metamorphics); 4: Nappe Zone (Low grade metamorphics); 5: Carboniferous flysch; 6: Major overthrusts; 7: Minor overthrusts; 8: Faults; 9: Axes of Flumendosa Antiform; A: Location map of Fig. 2; B: Location map of Fig. 4.
In the geological cross section: 10: Pre-Hercynian basement; 11: External Zone (Low to Very low grade metamorphics); 12: Nappe Zone (Low grade metamorphics); 13: D1 Phase movement; 14: D2 Phase movement.

1939). During the Middle-Upper Ordovician period a subaerial volcanic complex formed and is principally recorded by intermediate-acid metavolcanics with subalkaline affinity. Upper Ordovician deposits are transgressive on the volcanic complex and are represented by terrigenous and, subordinately, carbonate metasediments. Subsequent deposition is recorded by Lower-Middle Silurian pelitic neritic metasediments and by Upper Silurian-Devonian pelagic platform carbonate deposits.

The Hercynian deformational history in the Nappe Zone involved essentially two phases.

The first phase (D1), was produced during collision-related shortening and is characterized by synmetamorphic, SW-vergent isoclinal folds with subhorizontal axial planes and major tectonic overthrusts with NE to SW transport.

In the Flumendosa Valley crop out a wide antiformal stack (Carmignani et al., this volume) with associated isoclinal folds, thrusts and synmetamorphic schistosity (Fig. 1). This antiform trends N 60W and shows the most complete superposition of the tectonic units in the Hercynian chain of central Sardinia (Carmignani et al., 1982c).

The deepest tectonic units of the antiformal stack crop out in the axial culminations (Castello di Quirra, Riu Gruppa, Rio Mulargia area, Castello Medusa, M. Trempu-M. Grighini) and are tectonically covered by Gerrei type units (Carmignani et al., 1985).

The latter are overthrust, on the northern side of the antiform, by the Meana Sardo Unit and by the Low Grade Metamorphic Complex of the Barbagia area; while on the southern side they are overthrust by the Genn'Argiolas Unit, which comprises the entire metamorphic complex of the Sarrabus area (Fig.1) (Carmignani et al., 1982c; 1986b; 1986c; Barca et al., 1986).

The second deformational event (D2) appears largely extensional in character as previously thickened crust was raised and denuded resulting in the exposure of the deepest tectonic units in the axial culminations of the Flumendosa Antiform, ("metamorphic core complexes") (Carmignani et al., this volume).

During this second event, pre-existing thrust planes were reactivated as low angle detachments and asymmetric folds developed. Both structures show movement away from (i.e. north-eastwards and south-westwards) the culminations of the metamorphic core complexes.

3. Eastern Sarrabus Area

Eastern Sarrabus (A in Fig.1) is composed of metamorphosed and deformed terrains associated with the Hercynian orogeny and carboniferous granitoids (Fig. 2).

The low grade Hercynian regional metamorphism is linked to the first phase of deformation. A subsequent contact metamorphism that produced extensive recrystallization and silicification of Palaeozoic sequences in the central-southern outcrops is tied to the intrusion of Hercynian granitoids.

3.1. *Lithostratigraphic Characteristics*

3.1.1. *The Cambrian-Silurian Succession*

Exposures in the Eastern Sarrabus area (Fig. 2) consist of the southernmost part of the Genn'Argiolas Unit whose Cambrian-Silurian succession is in tectonic contact with the Carboniferous foredeep deposits.

The Genn'Argiolas Unit consists of Middle Cambrian-Lower Ordovician metasediments composed of alternating arenaceous-pelitic layers ("Arenarie di S. Vito" Fm: Calvino, 1961; 1967).

This sequence is unconformably covered ("Discordanza Sarrabese" Auct.) by a Middle-Upper Ordovician metarhyolitic-metarhyodacitic magmatic complex with subalkaline affinity magmatic complex ("Porfidi Bianchi" and "Porfidi Grigi": Calvino, 1956; 1961; 1967). The pelitic-arenaceous metasediments and silicified metalimestones of the Caradocian-Ashgillian (Punta Serpeddì Formation and Tuviois Formation: Barca & Di Gregorio, 1980) are generally transgressive on the volcanic complex and are overlain by the carbonaceous metargillites and black lidites containing Lower Silurian Graptolites. Sporadic outcrops of metalimestones, probably of Upper Silurian age, are also present.

3.1.2. *The Eocarboniferous Foredeep Deposits (Pala Manna Formation)*

In this paper we maintain the name "Pala Manna Formation" that was previously proposed by Barca (1991) for analogous deposits cropping out in western Sarrabus.

The Hercynian flysch in eastern Sarrabus comprises a clastic succession over 300 metres thick.

These Eocarboniferous synorogenic metasediments are composed of alternating layers of arenaceous-pelitic and siltite material that include intercalated quartzites, breccias and polygenic conglomerates. Locally, acid and basic metavolcanics are interstratified. These basic magmatic products range in composition from typical alkalic WPB to continental tholeiites (Di Pisa et al., this volume). The pelitic-arenaceous sediments exhibit frequent $T_{a\text{-}e}$ Bouma sequences and slumps.

The succession also contains olistostromes and olistoliths (Bruncu su Tidori, Riu Canale sa Figu, Baccu S'Ollastu and Riu Molas) composed of Silurian lidites, Upper Ordovician silicified limestones, Precaradocian acid volcanics and Silurian-Devonian dark limestones.

The ruditic intercalations, interpreted as debris flow, consist of horizons and pockets of polygenic metabreccias

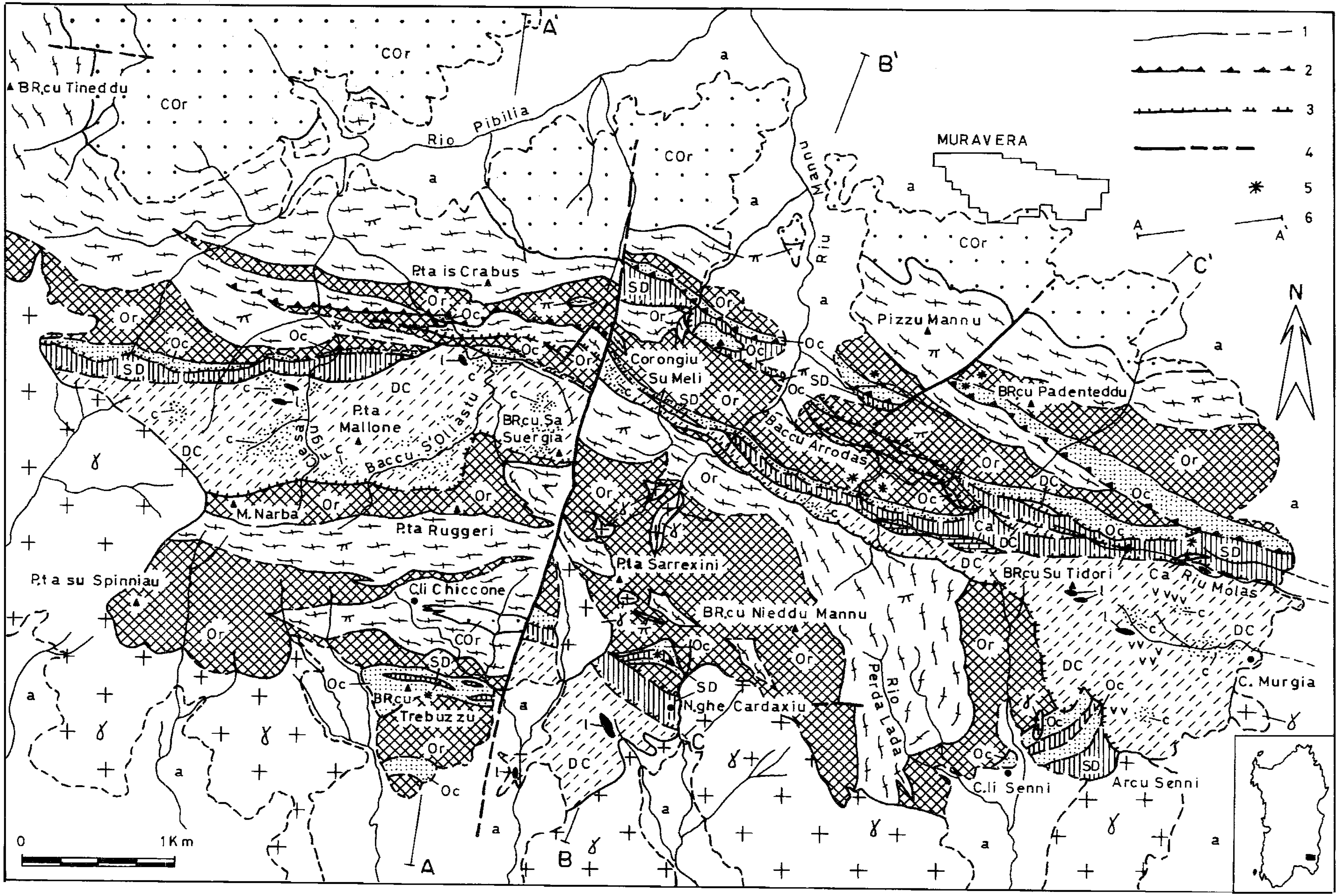


Fig. 2 - Geological map of the Muravera Area. a: Quaternary cover. γ : Late Hercynian granitoids. DC: Metasandstones and metasiltstones with: Metabreccias and poligenic metaconglomerates (c), basic metavolcanics (v), olistoliths of the cambrian-devonian sequence (l) (Pala Manna Formation); ?Devonian - Early Carboniferous). Ca: Nodular metalimestone (Devonian). SD: Black shales and lidites (Silurian). Oc: Silicified metalimestone (Tuviois Formation; Ashgill). Or: Metaconglomerates metasandestones and metasiltstones (P.ta Serpeddì Formation; Caradoc). π: Metarhyolitic and metarhyodacitic ("Porfidi Bianchi" and "Porfidi Grigi" Auct.; Ordovician). COr: Quartzites, metasandstones and metapelites ("Arenarie di S. vito" Formation; Middle Cambrian- Early Ordovician). 1: Stratigraphic boundary;2: Major overthrusts; 3: Minor overthrusts; 4: Faults; 5: Fossils; 6: Geological sections traces.

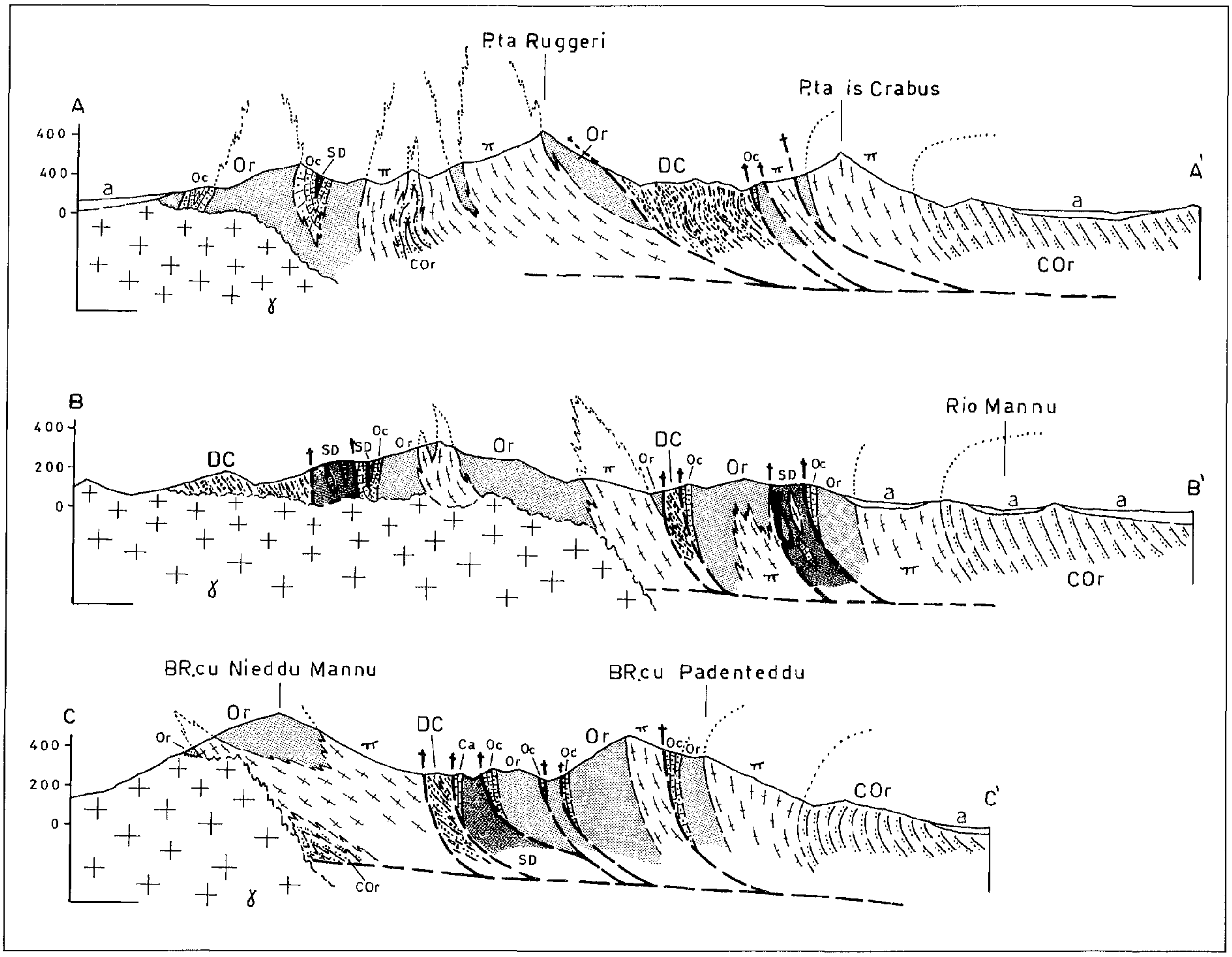


Fig. 3 - Schematic geological cross sections of the Muravera Area. Symbols as figure 2.

and metaconglomerates, composed mainly of lidite clasts and subordinately of silicified limestones, quartzites, quartz, sandstones and volcanics.

The clasts, which are relatively smooth and range from mm to dm in size, are both matrix supported and clast supported. In some places, the conglomerate layers show erosive contacts at their base. In the outcrops at Case Murgia the metaconglomerates exhibit both direct as well as inverse graded structures.

The levels of metabreccias and metaconglomerates display thicknesses that vary from a few decimetres to over 30-40 metres; the latter values being observed at Rio Baccu s'Ollastu.

3.2. *Tectonics*

The most prominent structures in eastern Sarrabus are thrust faults marked by cataclasites and mylonites. These planes trending WNW, dip strongly north, and separate the classic Middle Cambrian-Silurian succession of the Genn'Argiolas Unit from its clastic Culm facies cover.

The sub-units are composed of Genn'Argiolas Unit sequences and synorogenic deposits, which show well expressed schistosity and the effects of low-grade synkinematic metamorphism.

The structures produced during the first Hercynian phase are evident in the Middle Cambrian-Silurian formations. These include ESE-trending isoclinal anticlines and synclines and minor fold axes trending N 90-110 E with west dips. These folds are especially evident in the southernmost sub-unit; from the structurally highest to lowest the following structures can be identified:

— the "Porphyroid"-nucleus anticline of Punta Ruggeri;

— the Punta Su Spinnau-Bruncu Nieddu Mannu nucleus syncline, which consists of Upper Ordovician metasandstones;

— the Cuili Chiccone-nucleus anticline, in Cambrian metasandstones, which continues eastwards in the

"Porphyroid" nuclei of Punta S'Arexini and Riu Perda Lada;

— the Bruncu Trebuzzu-nucleus syncline, in Silurian metapelite, which continues eastwards in the outcrops north of Nuraghe Cardaxiu and Arcu Senni. This structure contains a series of minor folds involving the Upper Ordovician-Silurian deposits.

The thrust faults cut at low angles the axial planes of first phase folds. This is especially evident along the thrust fault between Monte Narba and Arcu Senni: in the Monte Narba area this tectonic contact cuts the southern sub-unit at the level of the upright limb of the structurally higher Punta Ruggeri anticline; in the Arcu Senni area the same tectonic contact cuts the structurally lower syncline of Bruncu Trebuzzu.

The cataclasites and mylonites defining the thrusts are often developed within Silurian metapelites. The cataclastic-mylonitic rocks crop out with thicknesses of up to hundreds of metres, enveloping tectonic wedges of Ordovician-Silurian formations and synorogenic sediments.

The well known Sarrabus mineralizations with sulphides ("Argentifero" Auct.) are located along this tectonic contact.

These have been interpreted either as stratabound mineralizations (Valera, 1975 and Bakos et al., 1989); or as material originating from the recirculation of mineralizing fluids linked to the late Hercynian granitoids (Checchi & Duchi; 1983).

In the area of Bruncu Padenteddu - Pizzu Mannu - Punta Is Crabus the southern sector of the northernmost sub-unit of the Genn'Argiolas Unit is folded and overturned toward the south (Fig.3).

This fold refolded the D1-schistosity. This metamorphic surface has a shallow dip in the Bruncu Tineddu area where the Ordovician "Porphyroids" lie in a straight series on the Cambrian metasandstones. The same D1-schistosity surface in the Punta Is Crabus area is subvertical. In this area the Ordovician "Porphyroids" lie below the Cambrian metasandstones and part of the overturned limb of the structure is exposed.

This megastructure has an axial direction of N 60 W, parallel, therefore, to the large late phase structures of the nappe zone ("Flumendosa Antiform").

4. Rio Mulargia Area

In the Rio Mulargia area, (B in Fig.1) at an axial culmination of the Flumendosa Antiform, three tectonic units crop out: the Meana Sardo Unit, the Gerrei Unit and the Rio Mulargia Unit (Fig. 4) (Eltrudis,1991).

4.1. *Lithostratigraphic Characteristics*

The structurally higher unit crops out in the NE part of the area (Fig.4) and is correlated with the Meana Sardo Unit (Carmignani et al., 1982c,b) on the basis of lithostratigraphic affinities. From the stratigraphic bottom to top, it comprises a complex of metavolcanics composed of acid as well as intermediate-basic rocks with subalkaline affinity and metavolcanoclastics.

This complex is correlated with the Monte Corte Cerbos, Manixeddu and Serra Tonnai Formations of the Middle or Middle-Upper Ordovician period (Carmignani et al., 1982c) which characterize the Palaeozoic successions of the Sarcidano-Barbagia area.

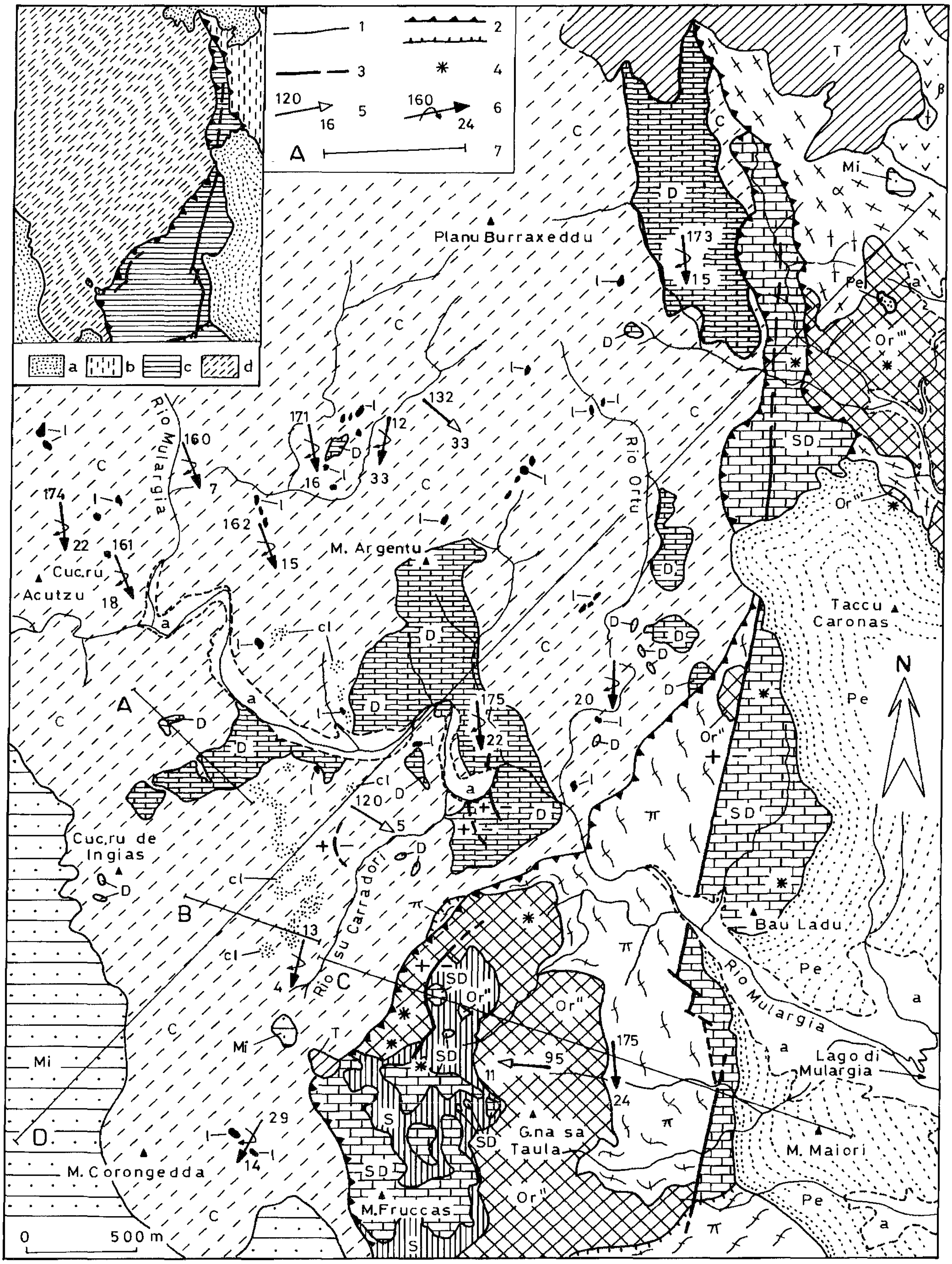
The volcanic complex is covered by metarkoses, metasandstones, green-brown metasiltites and Upper Ordovician (Caradocian-Ashgillian) shales.

The intermediate unit which crops out in the SE part of the area (Fig. 4) is a succession that is common throughout the Gerrei area (Gerrei Unit: Carmignani et al., 1985). Upper Ordovician (Caradocian-Ashgillian) metarkoses, metasandstones and fossiliferous metasiltites (brachiopods and crinoids) with intercalated carbonates are transgressive ("Caradocian Transgression" Auct.) on a substrate of Middle Ordovician "Porphiroid". These metasediments are overlain by Lower-Middle Silurian black shales and, subsequently, by Upper Silurian-Devonian (Alberti, 1963) fossiliferous, at times nodular, metalimestones (crinoids, orthoceratides, tentaculites).

4.1.1. *The Foredeep Deposits (Pala Manna Formation)*

The lower unit, the Rio Mulargia Unit (Eltrudis, 1991), crops out extensively in the central-eastern part of the area (Fig.4). It is composed mainly of a siliciclastic succession that is several hundreds of metres thick and is

Fig. 4 - Geological map of the Rio Mulargia area. Late-Hercynian sediments and volcanics: a: Quaternary cover; Mi: Conglomerates, sandstones and marls (Miocene); T: Limestones, marly limestones and dolomitic limestones (Triassic); Pe: Conglomerates, reddish clays and sandstones,acidic volcanics (Permian).Meana Sardo Unit - Or"': fossiliferous metasiltites and metasandstones (Upper Ordovician); β: Intermediate to basic metavolcanics and metamorphic products of reworked volcanics (Manixeddu and Serra Tonnai Formations: Middle Ordovician); α: Acidic metavolcanics (M. Corte Cerbos Formation: Middle Ordovician). Gerrei Unit: SD: Metalimestones(Silurian-Devonian); S: Black shales and quartzites (Silurian); Or": Metarkoses, metasandestones, fossiliferous metasiltites (Upper Ordovician); π: "Porphyroids" (Middle Ordovicia).Rio Mulargia Unit - C: Metasandstones and metasiltstones with: Metabreccias and poligenic metaconglomerates (cl),olistoliths of the cambrian-devonian sequence (l); D: Marble; (?Devonian-Early Carboniferous). 1: Stratigraphic boundary; 2: Major and minor overthrusts; 3: Faults; 4: Fossils; 5: D1 minor fold axes; 6: D2 minor fold axes with vergence; 7: Geological sections traces. In the tectonic scheme - a: Post-Hercynian sediments and volcanics; b: Meana Sardo Unit; c: Gerrei Unit; d: Rio Mulargia Unit.



a
b
c
d
1
2
3
4
5
6
7
A
N
Planu Burraxeddu
M. Argentu
Rio Mulargia
Rio Ortu
Cuc.ru Acutzu
Cuc.ru de Ingias
Rio su Carradori
Tacco Caronas
Bau Ladu
Lago di Mulargia
M. Maiori
G.na sa Taula
M. Fruccas
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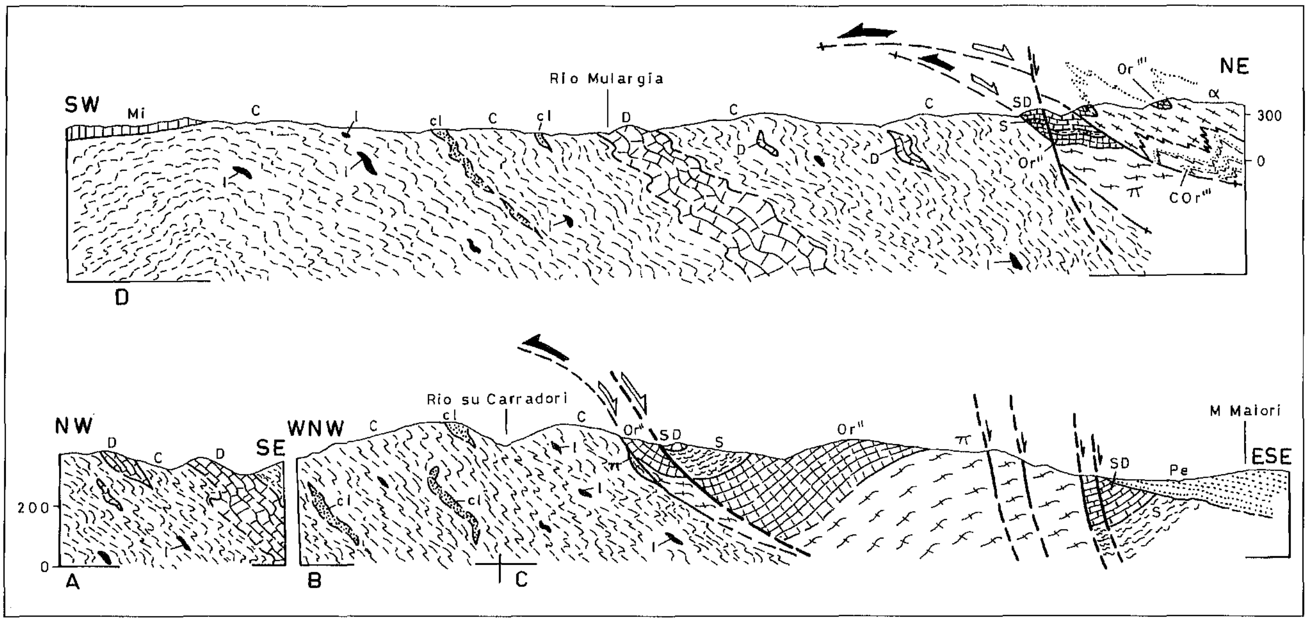

Fig. 5 - Geological cross sections of the Rio Mulargia area. Symbols as figure 4. COr''': Quartzites, metasandstones and metapelites ("Arenarie di Solanas" Formation: Middle Cambrian Early Ordovician); Black arrows: D1 Phase movement; white arrows: D2 Phase movement.

correlated with the Culm type deposits based on lithological and sedimentological characteristics. It contains monotonously alternating layers of metasandstones, metasiltites and greyish to blackish phyllites that host thick, quartzite lenses (i.e. Monte Corongedda, along the Rio Mulargia) interpreted as palaeochannels. The succession also contains sequences of metaconglomerates up to 80 metres thick (i.e. W of Riu Su Carradori, N of Rio Mulargia, about 1 km South of Monte Argentu), including paraconglomerates with fairly well rounded, centimeter-sized pebbles of lidites and silicified limestones; subordinate pebbles of sandstones and siltites in dark scanty pelitic matrix are also present. The clasts are considerably deformed and elongated representing an extension lineation on the D1 phase schistosity.

Numerous resedimentated blocks of lidites and limestones, characteristic of the Silurian-Devonian successions of SE Sardinia, can be observed intercalated in the arenaceous-pelitic metasediments and are interpreted as olistoliths. These blocks can even reach tens of metres in size.

The hectometric outcrops of marble and calcschists cropping out at Monte Argentu and E of Planu Burraxeddu could also be considered megaolistoliths. Until now, however, no evidence has been presented to affirm that they represent anticlinal nuclei of D1 phase isoclinal folds of the underlying Upper Silurian-Devonian limestones which crop out extensively in other axial culminations of the Flumendosa Antiform (i.e. Castello Medusa, Riu Gruppa, Castello di Quirra).

4.2. *Tectonics*

The trend of the tectonic contacts between the three units in Fig.4 and the characteristics of the late tectonic structures (D2 phase) indicate that the Mulargia area forms part of the southeastern sector of an axial culmination of the Flumendosa Antiform.

The deformational and metamorphic character of the lower tectonic unit are analogous to those of the deeper units of the Flumendosa Antiform (Castello Medusa and Castello di Quirra units Carmignani et al., 1982c). The Hercynian deformations that have affected the synorogenic deposits are divided into two phases:

—The first deformational phase (D1) produced a pervasive synmetamorphic axial Plane schistosity. Associated extensional lineations and schistosity/ stratification intersection lineations, trend from N10-30E to N90-140E, respectively.

The monotonous series, has not been subdivided and has so far failed to permit the identification of large D1 structures.

The subsequent extensional phase (D2), shows an evolution from ductile to brittle behaviour. The ductile deformations are represented mainly by variably sized, variably developed intrafolial folds of the D1-schistosity. The folds have a wide range of sizes and are variably developed from open to isoclinal. Axes generally trend N-S and folds systematically verge toward the east giving a down to the east sense of movement (Fig. 5).

These folds are overprinted by brittle structures which include both high and low angle normal faults. Parts of some listric faults drop blocks down to the East, causing their simultaneous tilting toward the west.

During this extensional phase, shear zones and normal faults utilized the pre-existing structural discontinuities that formed during the earlier phase of nappe building.

These movements caused drastic thinning in the Gerrei Unit (intermediate unit) such that the structurally higher unit became directly superimposed on the lower one (Fig.4).

5. Discussion

Both the succession comprising the lower unit of the Rio Mulargia area, as well as the flyschoid deposits of eastern Sarrabus display lithological and sedimentological similarities with the Lower Carboniferous Pala Manna Formation (Barca,1991; Barca & Olivieri, 1991).

Both flyschoid successions described here are characterized by the presence of resedimented material, forming olistoliths and olistostromes, lidite breccias and conglomerates typical of the deposits in Culm facies of other sectors of the southern European Hercynian chain (Spalletta, 1982; Maxia, 1984). These resedimented units can therefore represent syntectonic foredeep deposits that accumulated at the advancing nappe front. The flyschoid sedimentation occurred in the foreland of the developing chain where material was shed from the SSW-advancing nappes as well as from the foreland. The foreland was probably deformed by transcurrent tectonics that would explain the presence of contemporaneous bimodal volcanism. These Eocarboniferous volcanic products are analogous to those present in the Dinantian-Namurian flysch of the Carnian Alps (Rossi & Vai, 1986; Spalletta & Venturini, 1988). The transcurrent tectonics probably also produced a series of raised blocks in the foreland where the Hercynian succession was profoundly eroded. These high structures became sources for the flysch which were subsequently incorporated into the nappe building. With time, the orogenic front involved progressively more external areas of the developing chain.

The intimate relationships between tectonics and flysch deposition are confirmed by the present position of the synorogenic deposits with respect to the Nappe Zone and to the External Zone of the Hercynian chain in Sardinia (Fig. 1). Indeed, the Rio Mulargia unit appears to have been deeply involved in the Nappe Zone deformation and thus in Hercynian tectonics. On the other hand, the outcrops of the Sarrabus area are situated at the front of the allochthonous units, while in the eastern Sulcis area these deposits crop out extensively and cover the Silurian-Devonian succession of the External Zone of the chain (Maxia, 1984).

The nappes, in their movement toward the SW, initially involved the inner flysch basin and subsequently overroded it, thereby incorporating the flysch deep within the stack of nappes, where it presently forms the structurally lowest unit (Rio Mulargia area). Simultaneously, the synorogenic resedimentation continued at the front of the nappes and in the External Zone.

If we correlate the syntectonic deposits of the Rio Mulargia and Eastern Sarrabus areas with those of the Montagne Noire, the latter chacterized by an increase, stratigraphically upwards, of the quantity and size of the olistoliths (Franke & Engel, 1986), then we can interpret the placing of the large olistoliths of Rio Mulargia as an orogenic landslide that occurred immediately before the interruption of the synorogenic sedimentation and was followed by the arrival of the Gerrei-type nappes. In this context, the metric-sized olistoliths in the flyschoid succession of eastern Sarrabus would represent distal deposits whose terminal parts would be absent.

In the eastern Sarrabus area the tectonic relationships between sub-units composed of the Cambrian-Silurian succession and sub-units containing synorogenic deposits indicate that the latter should have been the stratigraphic cover of the classic succession of the Genn'Argiolas Unit. This flyschoid sedimentation occurred before, during and probably after the emplacement (late D1 phase) of the Genn'Argiolas Unit.

On the basis of the relationships between tectonic deformation and flysch sedimentation, we can assert that:

— the age of synorogenic deposits of the Genn'Argiolas Unit remains undefined, but could be Precarboniferous. This unit actually lacks the Devonian carbonate platform deposits, whose sedimentation would have been prohibited by the arrival of synorogenic clastic material. These precocious flyschoid deposits would subsequently have been passively transferred externally together with the underlying Cambrian-Silurian succession;

— the flysch of the more originally external areas (Gerrei, Mulargia and Sulcis) dates back to at least post Lower Tournaisian. Indeed, in the Gerrei units, the Devonian-Tournaisian carbonate platform deposits are overlain by synorogenic sediments affected by Hercynian deformation (metaconglomerates of Villasalto: Teichmuller, 1931; Spalletta & Vai, 1982; Barca & Spalletta, 1985; Barca, 1985);

— the flysch of the Rio Mulargia area was deposited before and during the southwest translation of the innermost units (Gerrei and Genn'Argiolas Units). These innermost units invaded the flysch basin of the Mulargia area, interrupted its synorogenic sedimentation and produced the antiformal stack of the Flumendosa Valley. The presence of the Devonian Gerrei carbonate platform probably played some role in the antiformal stack development;

— the flyschoid sedimentation continued at the chain front where the Genn'Argiolas Unit, already having been translated south-westward, was synchronously covered by the younger part of these deposits. Meanwhile, the inner zones of the chain experienced uplift and served as a source area for the clastic deposits;

— consequently the flyschoid deposits are progressively younger from the internal to the external parts of the chain. In such a structural framework, we can suggest a young age for part of the synorogenic deposits of the southern Sarrabus area based on the presence of metaconglomerates whose pebbles show effects of earlier deformation and metamorphism (Barca, 1991). Such pebbles originated from the innermost areas that had been uplifted during the extensional (D2) event, thereby providing the clastic material for the contemporaneous synorogenic deposits.

The presence of flyschoid deposits in the nucleus of the Flumendosa Antiform and at Sarrabus nappe front implies that the Gerrei and Genn'Argiolas Units float tectonically on a substratum of Culm-type flysch. A similar relationship is present in the Saxothuringian Zone in which the nappes overran the flysch basin incorporating the proximal flysch to the inside of the nappe stack, where it now forms the lowest structural unit (Franke & Engel, 1986).

In the Sardinian Hercynian chain, during the final stages of the SW progradation of the orogenic front, gravity sliding, linked to the late-stage extensional tectonics and chain uplift, probably also occurred.

The tectonic unroofing of the Flumendosa Antiform would thus have provoked a mechanical shifting of material towards the External Zone of the chain and subsequent SSW-translation of the Genn'Argiolas Unit. In the Muravera area the formation of a main footwall ramp caused the overthrusting of the internal part of Genn'Argiolas Unit onto the external one that was covered by synorogenic deposits. As a result, the newly overthrust sub-unit formed a large concentric ramp fold overturned toward the south. In the footwall a series of thrust faults formed, producing a trailing imbricate fan type structure that also involved the synorogenic deposits of the chain.

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LITHOSTRATIGRAPHY AND MICROFLORISTIC ANALYSIS OF THE FLUVIAL - LACUSTRINE AUTUNIAN BASIN IN THE SULCIS AREA (SOUTHWESTERN SARDINIA, ITALY)

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Abstract

Lithostratigraphical aspects and microfloristic content of the sedimentary-volcanic succession of the Guardia Pisano section (Sulcis area, SW Sardinia) are described. The lithofacies analysis and microfloristic data make it possible for the first time to assigne an Autunian (Early Permian) age to the exposed lower part of this fluvial-lacustrine succession. For this reason, the Guardia Pisano Autunian succession in the Sulcis area, together with the well-known Westphalian D (?)-Stephanian sequence of the San Giorgio lacustrine basin in the Iglesiente area, represent the only evidence of post - Hercynian continental molasses in Southwestern Sardinia.

Keywords: *Lithology, Biostratigraphy, Palynology, Continental molasse, Permian, Hercynian orogen, Mediterranean Area.*

1. Introduction

A sedimentary succession with interbedded acid volcanic products cropping out at Guardia Pisano (Sulcis area, SW Sardinia) (Fig. 1) is described in this note. The Guardia Pisano outcrop was previously quoted as Corona Maria toponomy (a site that actually lies 1 km SW of Guardia Pisano) and ascribed to the "Permo-Trias" without palaeontological evidence (Barberi & Cherchi, 1980). Moreover, the purplish-red clastic sediments at the top of the Guardia Pisano succession, the only ones cropping out before the recent cutting of the panoramic road from Gonnesa to Portoscuso, was ascribed to the overlying Eocene formation (R.Ufficio Geologico , 1919).

Finally, it is interesting to point out that a thick (at least 250 m) sedimentary and volcanic succession drilled near Nuraxi Figus (some km S of Guardia Pisano) has been referred to an indetermined and not proved Permian-Triassic interval only on the basis of facies analogy with other successions cropping out in North Sardinia (Salvadori, 1980).

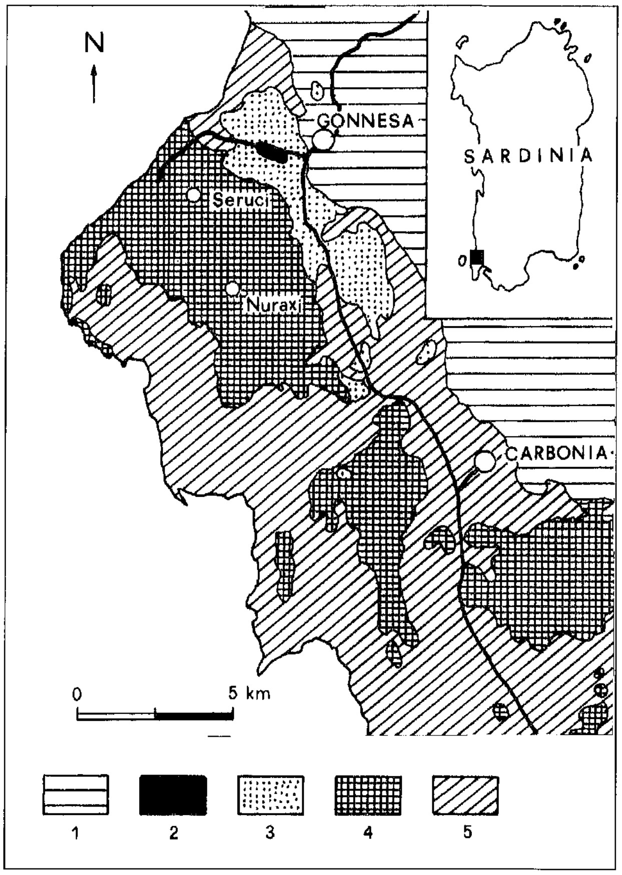

Fig. 1 - Geological sketch-map of the Sulcis area : 1 - Palaeozoic basement; 2 - Permian studied outcrop; 3 - Palaeogenic sediments; 4 - Oligo-Miocenic volcanites; 5 - Quaternary sediments.

1.1 *Description of the section*

The section studied is clearly exposed along the northern slope of Guardia Pisano hills, some 2 Km SW of Gonnesa (Fig.1), with an overall thickness of at least 100 m, although the base is not visible. From bottom to top, it consists of (Fig. 2) :

A) Dark grey shales more or less carbonaceous, thinly stratified, with rare sandstone lenticles rich in carbonaceous plants remains. The visible thickness is 6-7 m. This level

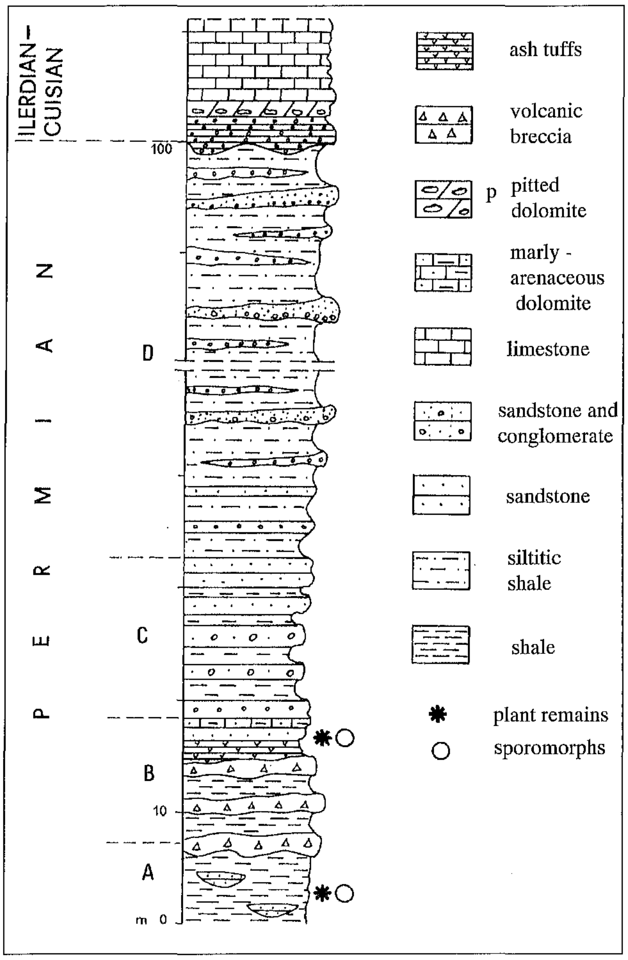

Fig. 2 - Stratigraphical column of Guardia Pisano section.

has provided the majority of the interesting microflora studied (samples N° 2189, 2191, 2194 and 2205).

B) At least three levels of volcanic breccias in thick irregular beds, alternated with carbonaceous shales and tuffs.

This unit is followed by at least 2 m of whitish ash tuffs and decimetric layers of dolomite and coarse-grained sandstones containing thin carbonaceous levels with fragments of Conifers, probably including branches of *Walchia* sp. Furthermore the palynological analysis carried out on the sandstone, sample N° 2192, provided only scarse specimens of *Potonieisporites* sp. On the contrary, sample N° 2193 of intercalated tuffitic shales in the volcanoclastites has provided no palynological results whatsoever. The volcanic breccia comprises heterometric fragments of rhyolitic and rhyodacitic lava of up to 30-40 cm in size showing markedly porphyritic texture with quartz phenocrysts, K-feldspar, plagioclase mainly transformed into calcite and partly chloritized biotite. A very small outcrop of this lava is found in the field on the other side of the road section. The overall thickness of B is approximately 10-12 m.

C) Coarse-grained greyish sandstone, with small conglomerate levels and dark shales. The sandstone shows a cross-bedding structure and is very rich in biotite, altered feldspars and quartz, clearly derived from erosion of a volcanic building close by. The thin conglomerate levels also consist largely of pebbles and angular fragments of porphyritic volcanite and of Palaeozoic metamorphic rocks. The total thickness is about 10 -15 m.

D) Purplish-red siltitic shales, with repeated intercalations of lenticular layers and thick beds of siltitic sandstones, at times with cross-bedding structures and burrows. The base of the largest sandstone bank is conglomeratic with well-elaborated Palaeozoic pebbles some cm in size. These lithotypes are wholely organized in a fining-upwards megasequence and indicate high density deposit mechanism. In this unit there is no sign of volcanic intercalations or fossils. The thickness is around 60 - 70 m.

The sedimentation environment of the studied succession range from predominantly lacustrine (Units A, B) and fluvial-lacustrine (Unit C) with associated volcanic activity, to an alluvial plain (Unit D) system.

Towards the top of the Guardia Pisano hills, the studied succession is covered by the Ilerdian-Cuisian (Pittau, 1977; Cherchi, 1979) sediments of the"serie lignitifera" Auct, through an angular unconformity. These sediments consist of a basal conglomerate, containing pebbles of quartz and Palaeozoic schistes , as well as of lacustrine-lagoon yellow pitted dolomites and whitish limestones.

2. Palynology

Units A and B have proved to be fossiliferous for pollen and spores. In Unit A we note the presence of three distinct levels reflecting consistent variations in the spore/ pollen spectrum proportion and in the composition of the Pteridophytic community.

This is certainly linked to climatic changes which occurred during the deposition of this unit and, moreover, reflect different points of deposition in the lacustrine basin.

The list of sporomorphs recognized is given below, with the exception of the new taxa which number many spores and pollens , and will be described elsewhere.

List of sporomorphs

Calamospora breviradiata Schopf, Wilson & Bentall 1944

Calamospor liquida Kosanke 1950

Leiotriletes sp.

Cadiospora magna Kosanke 1950

Cyclogranisporites cf. *minutus* Bhardwaj1957

Granulatisporites elegans Peppers 1964

Anaplanisporites globulus (Butterworth & Williams) Smith & Butterworth1967

Apiculatisporis abditus (Loose) Potoniè & Kremp1955

Lophotriletes gibbosus (Ibrahim) Potonè & Kremp 1954

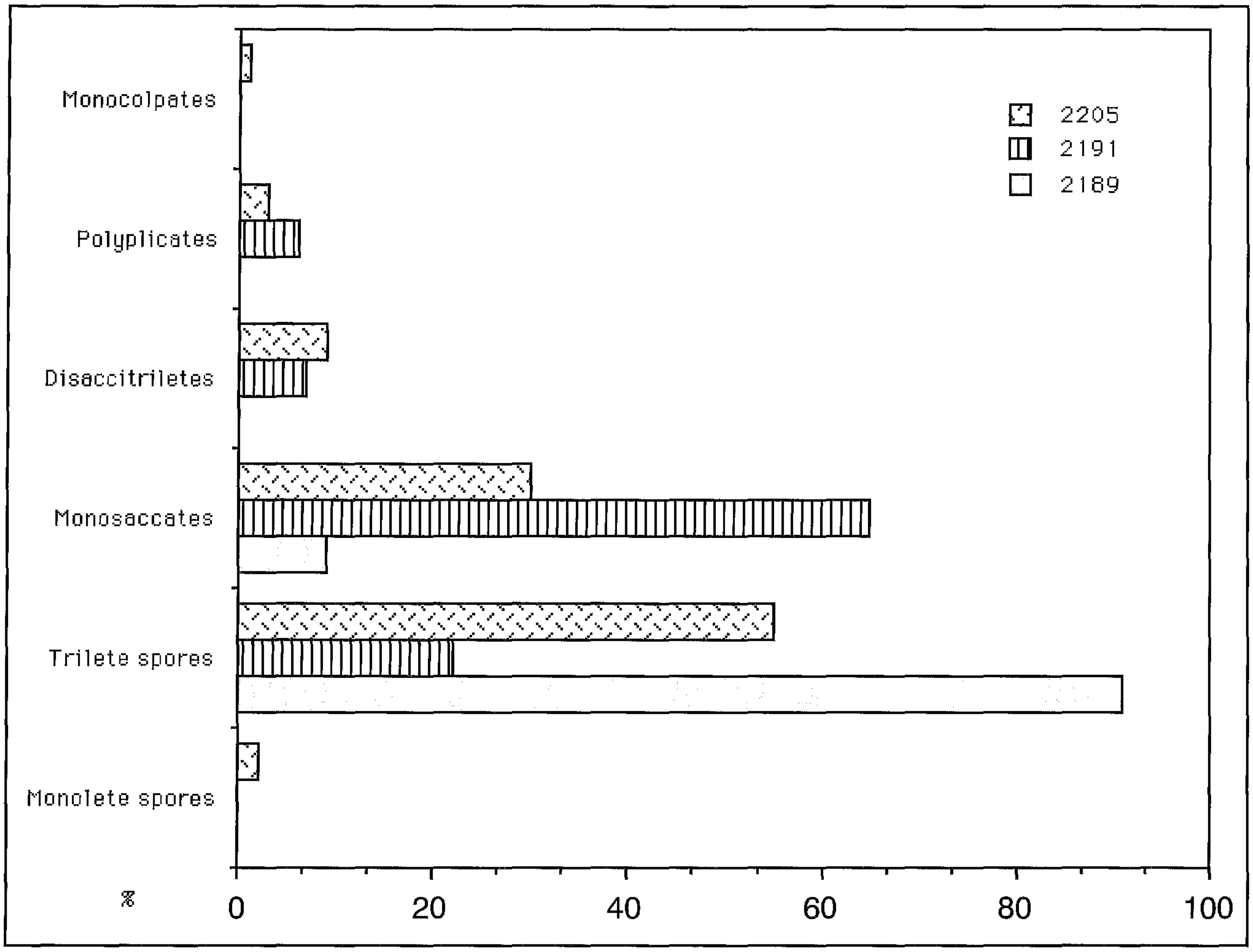

Fig. 3 - Spectrum reflecting the proportion of morphographic groups in Unit A

Knoxisporites glomus Schwartsman 1976
Densosporites sp.
Lundbladispora simonii Peppers 1964
Lycospora orbicula (Potoniè & Kremp) Smith & Butterworth 1967
Rugaletes sp.
Candidispora candida Venkatachala 1963
Cordaitina sp.
Florinites antiquus Schopf 1944
Florinites florinii Imgrund 1960
Florinites similis Kosanke 1950
Latensina trileta Alpern 1958
Plicatipollenites gondwanensis (Balme & Hennelly) Lele 1964
Potonieisporites bharadwaj Remy & Remy 1961
Potonieisporites grandis Tschudy & Kosanke 1966
Potonieisporites novicus Bhardwaj 1954
Hamiapollenites tractiferinus (Samoilovich) Jansonius 1962
Illinites unicus Kosanke 1950
Kosankeisporites sp.
Limitisporites sp.
Schopfipollenites ellispoides Potoniè & Kremp 1954
Vittatina costabilis Wilson 1962
Vittatina subsaccata Samoilovich 1953

Conipherophyte pollens are constantly present in Units A and B reaching in some levels the total amount of the microflora (as in unit B) or 78% in level 2191. In any case the presence of *Potonieisporites novicus* Bhardwaj 1954 alone ranges from 5 % up to 35 % .

Pteridophyte representatives range from 0, in some levels of Unit B, to 91% in certain levels of Unit A.

Comparing this basin with the Autunian basin of Massif Central in France (Doubinger, 1974; Bouroz & Doubinger, 1974; Broutin et al., 1986; Chateauneuf & Pacaud, 1991) we note that spores of *Pecopteris* (monolete) are completely missing and spores of Sphenophyllales (*Laevigatosporites)* are sporadically present. Small ferns, on the other hand, are largely represented by species of the genus *Punctatisporites*, *Apiculatisporites* and *Cyclogranisporites*. This can be linked to the morphology of the lacustrine basin, which probably lacked one extensive swamp; moreover we suppose that the lake was surrounded by a small fern prairie . Their conspicuous presence in different levels is related to climatic change, and their

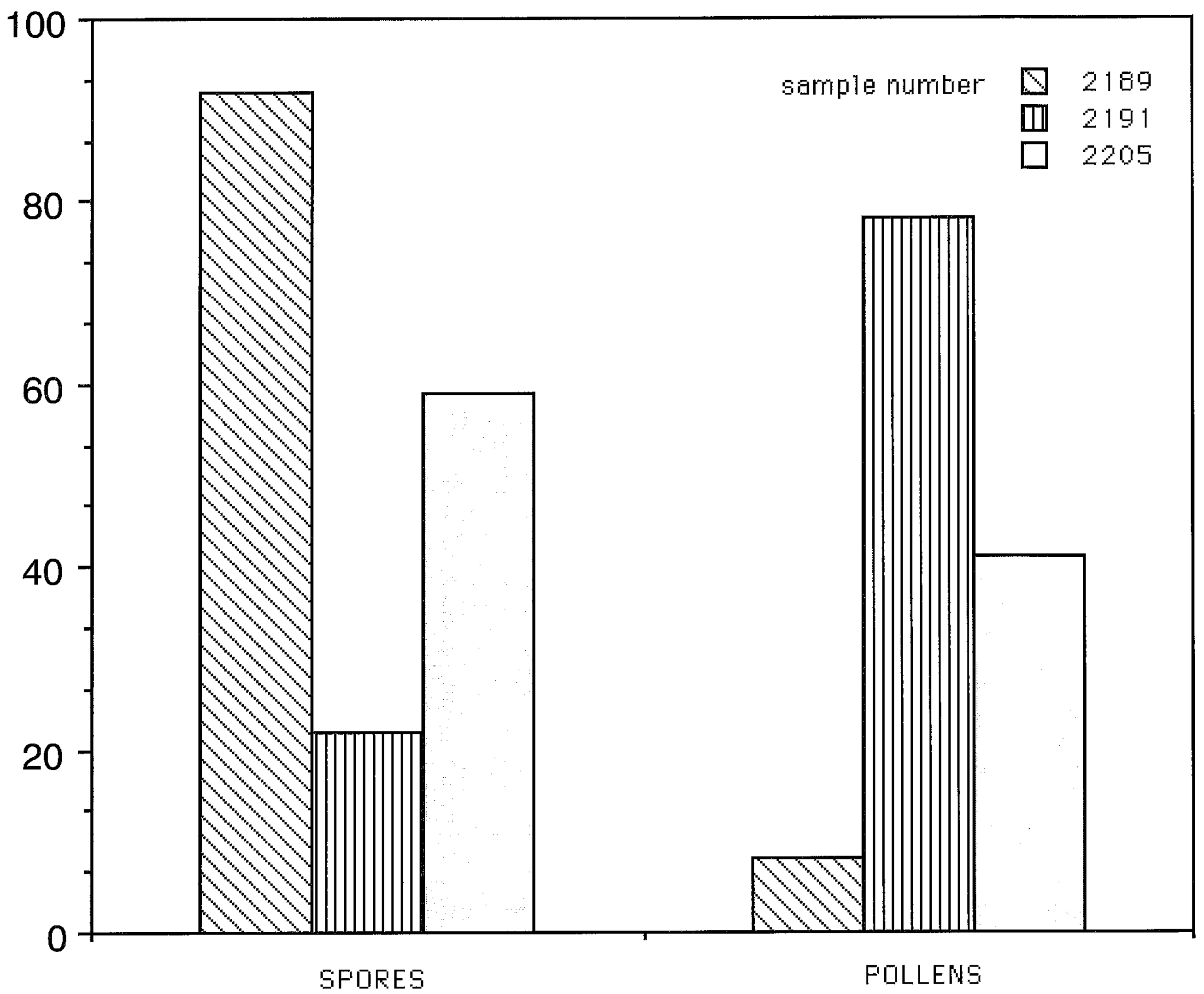

Fig. 4 - Bar chart showing the proportion of spore-pollen content in Unit A.

biostratigraphic value is presently unproved.

From a chrono-correlative point of view we have taken into consideration the following:

— the conspicuous presence of Monosaccate pollens of which *Potonieisporites novicus* Bhardwaj 1954 and *Potonieisporites bharadwaj* Remy & Remy 1961 represent the major components and, together with *Latensina trileta* Alpern 1958, *Cordaitina* sp., *Candidispora candida* Venkatachala 1963 and *Florinites* (different species), allows direct correlation with the Autunian;

— the moderate presence of Polyplicate pollen grains (*Vittatina*);

— the very limited presence of Disaccate Striatites;

— the absence of Monolete spores like *Thymospora* and *Spinosporites*, which are largely represented in the Stephanian and Lower Autunian of the Autun Basin.

Disaccate striatites pollen seem not to be present in our association except for *Kosankeisporites* sp. and rare specimens of *Hamiapollenites tractiferinus* (Samoilovich) Jansonius 1962, whose presence dates from Virgilian (Upper Carboniferous) and Wolfcampian (Lower Permian) in the U.S.A. (Jizba, 1962) and from the "Kazanian to the Upper Permian" in Russia (Samoilovich, 1953). Also the presence of moderately abundant *Limitisporites* sp. may have a correlative value with the Autunian of Autun .

Assemblages of Unit A may be referred to the Early Autunian of French palynologists, but a detailed correlation with French zonation is difficult to establish. Based on the pollen content, relations may be established with Zone 2 corresponding to the Igornay strata, and Zone 3 of the Lally strata (Chateauneuf & Pacaud, 1991) even if we note that pollens of *Gardenasporites*, present in Zone 3, are absent from our assemblages, and *Hamiapollenites* appears in the Muse strata belonging to Zone 4.

3. Conclusions

Based on the lithofacies analysis and the palynological study of the Guardia Pisano succession, it is possible to formulate a number of conclusive considerations of stratigraphic and palaeoenvironmental order.

Units A, B, C of the succession studied are assigned to the Lower Autunian and are referred to lacustrine or occasional fluvial-lacustrine environments under conditions of a hot climate with alternating wet and dry periods.

Consequently volcanic products present in these units are also closely datable; they can be more or less correlated with the acid volcanism present in the Autunian basin of Seui in the Barbagia area (central Sardinia), whose radiometric age K/Ar is set at between 250 and 256 Ma (Cozzupoli et al., 1971). This volcanism is linked to a post-Hercynian extensional tectonics (Gasperi & Gelmeni, 1977; Fontana et al., 1982).

The reddish detrital Unit (D), which lies with erosional contact on the preceding units, in the absence of fossils can for the time being only be generically allottable to the Permian; it would mark the transition to an alluvial plain environment under semiarid hot climate.

Finally the Permian succession of Guardia Pisano, represents so far the only Permian outcrop dated palaeontologically in the whole Iglesiente-Sulcis area; indeed the well-known continental deposits of the San Giorgio basin (Iglesias) are ascribed to the Westphalian D(?)-Stephanian (Cocozza, 1967; Del Rio, 1973; Vai & Cocozza, 1974; Fondi, 1980). They all comprise the only evidence of post- Hercynian continental molassic deposits cropping out in Southwestern Sardinia. Their presence, together with that of the thick sedimentary and volcanic "Permo-Triassic" succession reported in the subsurface of the Sulcis area (Salvadori,1980), make it possible to assume the existence of a late Palaeozoic continental basin in this part of Sardinia as well, at least comparable in size with those of the Nurra (Northwestern Sardinia) and Barbagia (central Eastern Sardinia) areas (Maxia, 1938; Pecorini, 1974; Francavilla et al.,1977).

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MINOR TECTONIC UNITS WITHIN THE HERCYNIAN ARBURESE NAPPE IN SOUTHWESTERN SARDINIA. NEW STRUCTURAL AND STRATIGRAPHIC EVIDENCES

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ABSTRACT

Three different tectonic sub-units have been recognized within the Hercynian Arburese Tectonic Unit according to works recently completed in Southwestern Sardinia. In addition, the detailed stratigraphic study of each sub-unit, mainly with conodonts, resulted in the report, for the first time in this area, of some Silurian and Lower Devonian series and conodont biozones.

KEY WORDS: *Hercynian Tectonics, Stratigraphy, Conodont biostratigraphy, Palaeozoic, SW Sardinia.*

1. INTRODUCTION

The Hercynian Arburese Tectonic Unit covers the all allocthonous Palaeozoic rocks exposed in Southwestern Sardinia, thrusted from NNE to SSW over the autochtonous Palaeozoic successions of the Iglesiente-Sulcis foreland (Barca et al., 1982). This main unit is structurally located in the nappe front of the Hercynian Sardic Orogen (Carmignani et al., 1982). The Arburese Unit is predominantly made up of sedimentary and volcanic low grade metamorphic rocks which age is so far considered ranging from Cambrian-Early Ordovician to a not well defined Silurian.

New structural and stratigraphic data concerning the Arburese Unit cropping out in the S. Antonio di Santadi area in the northern Arburese region (Fig. 1), are here reported.

2. NEW STRUCTURAL AND STRATIGRAPHIC EVIDENCES

Recent intensive studies in SW Sardinia have led to recognize within the Arburese Unit at least three new tectonic sub-units, separated each other by important overthrusting surfaces marked by cataclasites. From the base to the top of the structure, we have distinguished: the Donigala sub-unit, the S.Antonio di Santadi sub-unit and the Monte Fonnesu sub-unit (Figs 2, 3, 5).

A complete and detailed stratigraphic analysis of each tectonic sub-unit is here given. Particularly, some stages and biozones of the Silurian as well as the presence of the Early Devonian is documented in this area for the first time on the basis of conodont faunas found in the Donigala sub-unit.

2.1 *Stratigraphy*

The three tectonic sub-units are synthetically described according to their stratigraphic successions (Fig. 3).

2.1.1 *The Donigala Sub-Unit*

The Donigala sub-unit consists of different lithologies which, in stratigraphic order, are the followings:

O'v: Rhyolitic-rhyodacitic volcanites and volcanoclastites with interbedded epiclastic products. This volcano-sedimentary complex may be ascribed to the pre-Caradocian sub-alkaline to calc-alkaline magmatism of Southeastern Sardinia (Di Simplicio et al., 1974; Memmi et al., 1982; Carmignani et al., 1991);

O'sq, O's: Coarse quartzites and quartzitic-feldspatic sandstones, at times microconglomeratic, light to dark-grey in colour (O'sq). Dark-grey fine sandstones and ash-grey siltstones follow the basal coarse sediments. A very rich benthic association of brachiopods, bryozoans, crinoids, cystoids, trilobites, corals and cornulites, occurring inside the siltstones, document the Caradoc-Ashgill (Barca & Salvadori, 1974; Giovannoni & Zanfrà, 1979). These sediments, typical of an inner shelf environment, mark the "Caradocian transgression" Auct. on the underlying volcano-sedimentary complex;

SD: Black shales with lenses of black limestone,

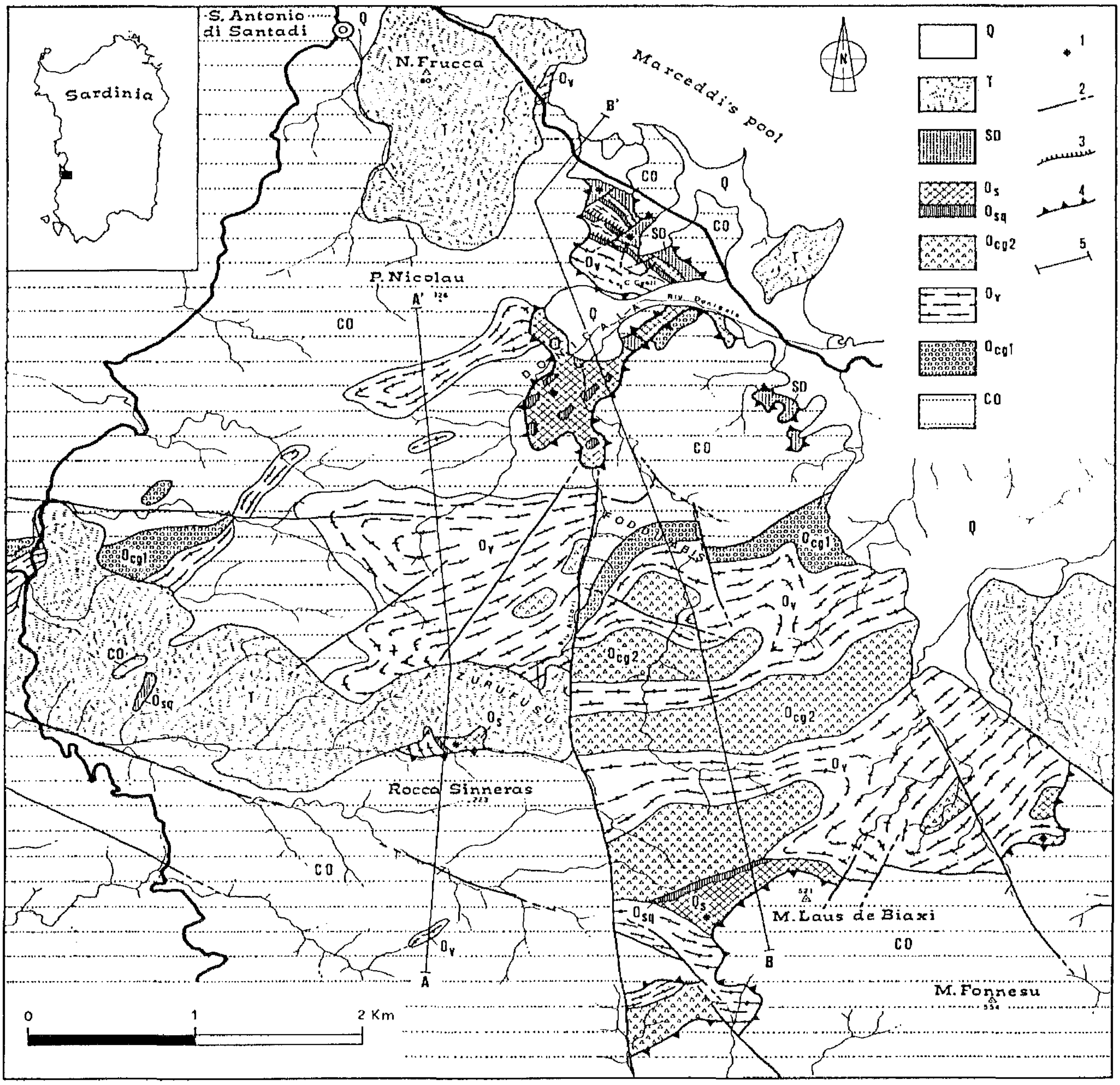

Fig.1 - Geological sketch-map of the S. Antonio di Santadi area.
CO= Cambrian- Lower Ordovician sandstones, siltstones and mudstones. Ocg1= Ordovician (pre-Caradoc) feldspatic sandstones with siltitic intercalations and conglomerates. Ov= Ordovician (pre-Caradoc) acid volcanites and volcanoclastites. Ocg2= Ordovician (pre-Caradoc) coarse clastic sediments. Osq= Upper Ordovician (Caradoc-Ashgill) coarse and mainly quartzitic detritic rocks. Os= Upper Ordovician (Caradoc-Ashgill) fossiliferous siltstones and sandstones. SD= Silurian- Lower Devonian black fossiliferous limestones and grey nodular limestones. T= Sedimentary and volcanic Tertiary rocks. Q= Quaternary deposits. 1= fossiliferous outcrops. 2=faults. 3= minor overthrusts. 4= main overthrusts. 5=location of sections.

marly limestone and micritic limestone enriched in nautiloids, bivalves (cardiolids), graptolites, conodonts, ostracods, rare gastropods and brachiopods. The biostratigraphic analysis based on conodonts permits to assigne these limestones to the Latest Llandovery (italic zone)- Middle Ludlow (italic zone) interval (Fig. 4). The index-species *Pterospathodus amorphognathoides, Polygnathoides siluricus* and *Ozarkodina bohemica bohemica* are reported for the first time in Sardinia. The frequent nautiloid infillings give rise to wonderful geopetal fabrics that can be used to determine the polarity of parts of the sequences. All these black Silurian limestones show close similarities with the coeval *"Orthoceras tondo"* levels of the Fluminimaggiore Fm. in other areas of SW Sardinia (Gnoli et al., 1990).

Grey nodular limestones alternate with dark siltstones and shales at the top of the sequence. The fauna is composed of "tentaculitids", locally abundant, rare nautiloids, fragments of crinoidal stems, gastropods, bivalves and trilobites. As regards conodonts, although these limestones

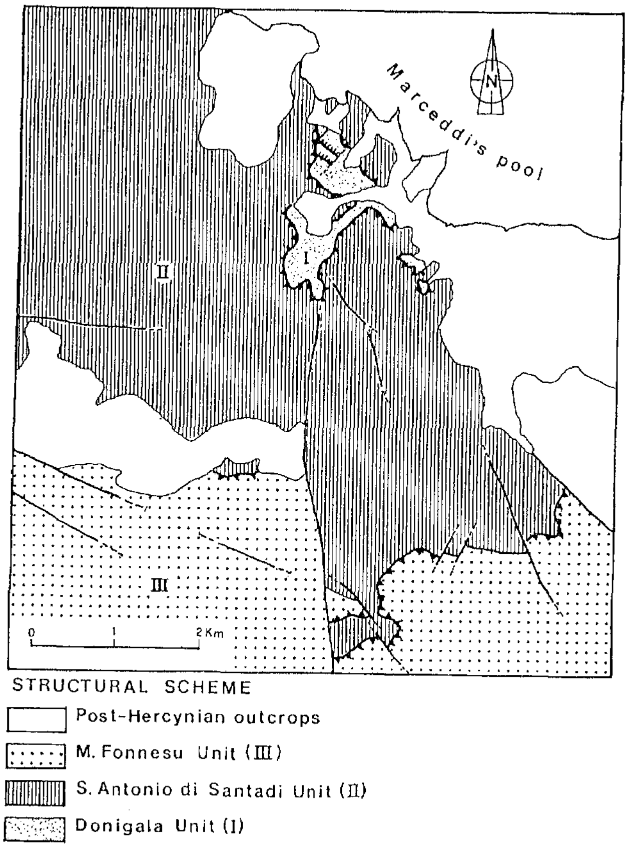

Fig. 2 - Structural scheme of the S. Antonio di Santadi area.

yielded only poor unrecognizable fragments or very few elements with no stratigraphic value, we consider these sediments as Early Devonian in age according to their close lithologic affinity with the well studied nodular limestones of the Mason Porcus Fm. exposed in SW Sardinia (Gnoli, 1985).

The total thickness of the sub-unit is only some tens metres, probably due to a strong tectonic thinning or to intensive elision effects.

2.1.2 *The S.Antonio di Santadi Sub-Unit*

It consists of:

— CO": Flyschoid sequence characterized by a rhytmic alternance of grey-greenish, black or sometimes violet micaceous sandstones, quartzites, siltstones and mudstones. Current structures are quite frequent at the base and within the beds. The total thickness is more than 600 metres. The age of these sediments is regarded as Cambrian- Early Ordovician on the basis of acritarchs found in the upper part of the sequence (Barca et al., 1982; Pittau, 1985). Also the analogies with the San Vito Sandstone and Solanas Formation (Southeastern Sardinia) and with the Cabitza Shales (Southwestern Sardinia) strongly support this interpretation (Barca et al., 1987, 1988);

— O"cg1, O"v, O"cg2: Continental volcano-sedimentary complex consisting of arkosic sandstones and polygenic conglomerates (O"cg1) at the base of the sequence. Volcanoclastic products and rhyolitic-rhyodacitic volcanites (O"v) occur in the upper part, where coarse clastic sediments (O"cg2), rich in volcanic clasts, are also frequently interbedded.

This complex is the equivalent of the pre-Caradocian volcano-sedimentary sequence (O'v) in the Donigala sub-unit and attains a maximum thickness of some hundred metres;

— O"sq, O"s: Transgressive fossiliferous sediments that correspond to the Caradocian-Ashgillian sequence already described in the Donigala sub-unit. Several unfossiliferous small lenses of light-brown marly limestone interbedded with fossiliferous siltstones rich in bryozoans and brachiopods are exposed in the new fossiliferous Zurufusu locality belonging to the S. Antonio di Santadi sub-unit. This sequence is some ten metres thick.

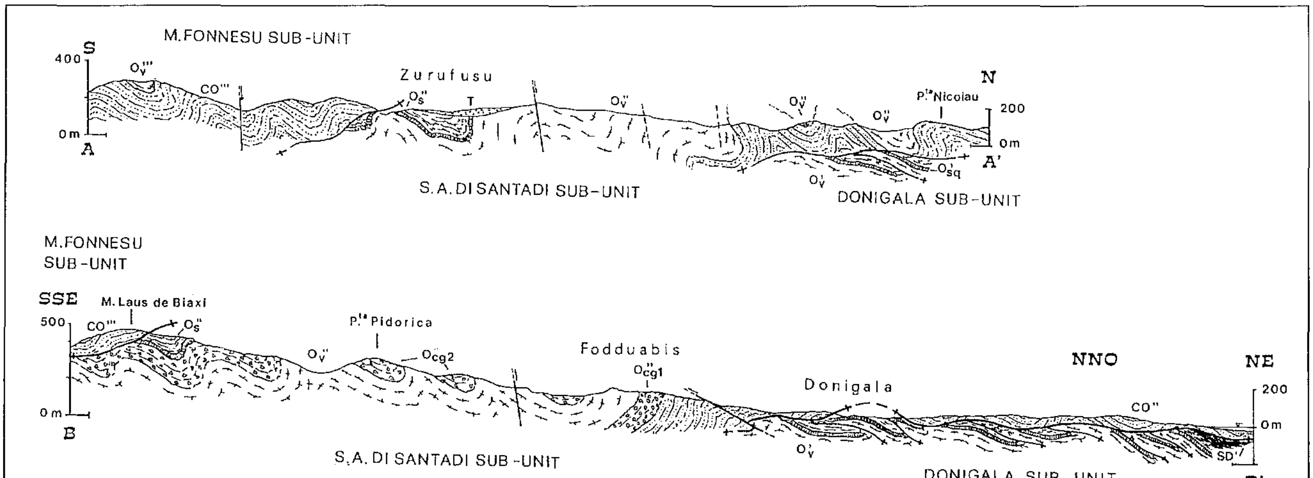

Fig. 3- Geological sections of the S. Antonio di Santadi area.

	SERIES	CONODONT BIOZONES
SILURIAN		I. w. woschmidti
	PRIDOLI	O. r. eosteinhornensis
	LUDLOW	O. crispa
		O. snajdri
		P. siluricus
		A. ploeckensis
	WENLOCK	O. bohemica bohemica
		O. sagitta sagitta
		O. sagitta rhenana
	LLANDOVERY	P. amorphognathoides
		P. celloni
		D. staurognathoides
		D. kentuckyensis

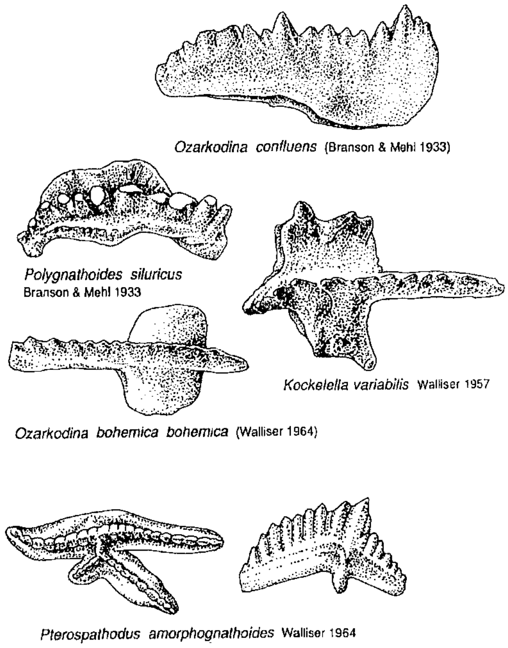

Fig. 4- Silurian conodont biozones and some remarkable index-conodonts found in the studied area.

2.1.3. *The Monte Fonnesu Sub-Unit*

It is mainly represented by the thick siliciclastic flyschoid sediments of Cambrian-Early Ordovician age already described in the lower part of the S. Antonio di Santadi sub-unit. Volcanoclastic products and acid volcanites referrable to the pre-Caradocian calc-alkalinic magmatism are also locally present.

2.2. *Tectonics*

Pre-Hercynian, Hercynian and Alpidic tectonics are well expressed in the whole studied region.

The Arburese area is affected by a polyphasic Hercynian tectonics with a strong NNE to SSW transport such as the whole nappe zone of central-eastern Sardinia. The structural analysis reveals infact that the main folding phase is represented in this area by southwards verging isoclinal folds of hectometric-kilometric size. A slaty cleavage structure is well preserved especially in the incompetent lithotypes. The first-phase folds evolved in large overthrusts creating the pile of the minor tectonic units, each one composed of more or less incomplete sequences due to tectonic thinning or elision. This structure

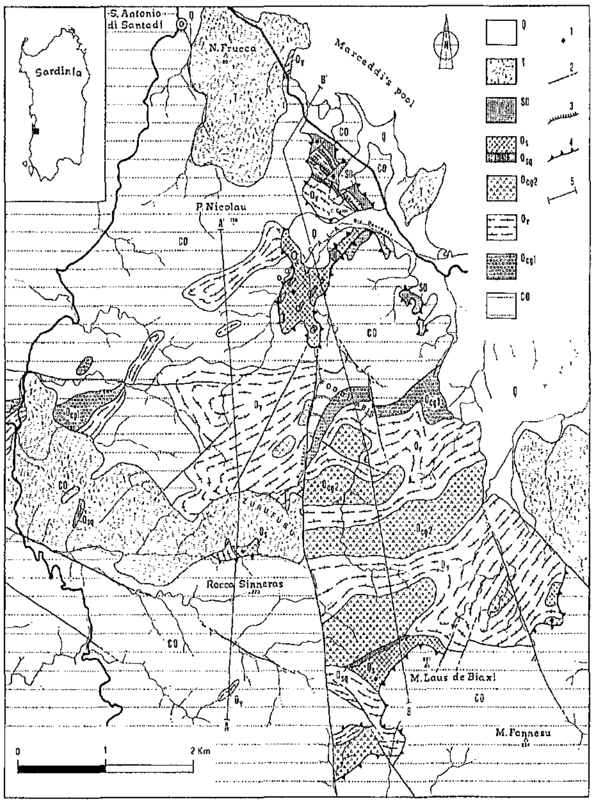

Fig. 5 - Stratigraphic columns and tectonic relation of the three tectonic sub-units recognized within the Hercynian Arburese Tectonic Unit. I= Donigala sub-unit, II= S. Antonio di Santadi sub-unit and III= M. Fonnesu sub-unit.

is well documented in the studied area by the three described tectonic sub-units: the Donigala sub-unit, the S. Antonio di Santadi sub-unit and the Monte Fonnesu sub-unit (Figs 3 and 5). These first-phase folding structures were subsequently deformed by late folds whose axial directions scattered from NW-SE to NE-SW and that produced a fracture or crenulation cleavage without notable metamorphic effects.

Together with the Hercynian tectonic events, also certain clues of pre-Hercynian tectonics are present in this area. The latter are in fact testified by the important stratigraphic unconformity separating the Cambrian- Lower Ordovician sedimentary sequence from the overlying pre-Caradocian volcano-sedimentary complex. This "Eocaledonian" phase can be related to the coeval Sardic Phase (Stille, 1939) of the Iglesiente-Sulcis area and the Sarrabese Phase (Calvino, 1961) of the Sarrabus area. As in these last areas, also in the Arburese no significant metamorphic or deformative effects were produced by these older tectonic events.

Finally, the Alpidic fault systems mapped in the area, which occur with N-S, E-W and NNW-SSE directions (Fig. 1) may be related to the Oligo-Miocenic and Plio-Pleistocenic, mainly extensional phases, which characterized the late geodynamic evolution of Sardinia (Cherchi & Montardet, 1984).

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A REAPPRAISAL OF THE STRATABOUND ORES AT THE MID - ORDOVICIAN UNCONFORMITY IN SW SARDINIA

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Ba-Pb-Zn-Fe stratabound ores related to strongly silicified lithotypes (the so-called "Quarziti") have been recorded in the Ordovician of SW Sardinia (Fig.1). They consist mostly of barite with minor galena, in places containing some sphalerite and Cu-sulphosalts. As-rich pyrite, in form of concretions and framboids, is locally very abundant. These deposits are today economically unimportant (few million tons of ore combined), but were partially exploited in the past. The most popular hypothesis regarding the origin of these ores (Benz, 1963; Brusca and Dessau, 1968; Padalino et al., 1973; 1989; Boni, 1979; 1985; Marcello et al.,1983) is that of a supergenic metal deposition in paleokarstic cavities, shortly followed by the formation of a siliceous "crust", fossilizing both the solution cavities and their contained ores.

The prevailing host rocks for the mineralization are fanconglomerates and breccias of Upper Ordovician age, possibly laterally equivalent to part of the Monte Argentu

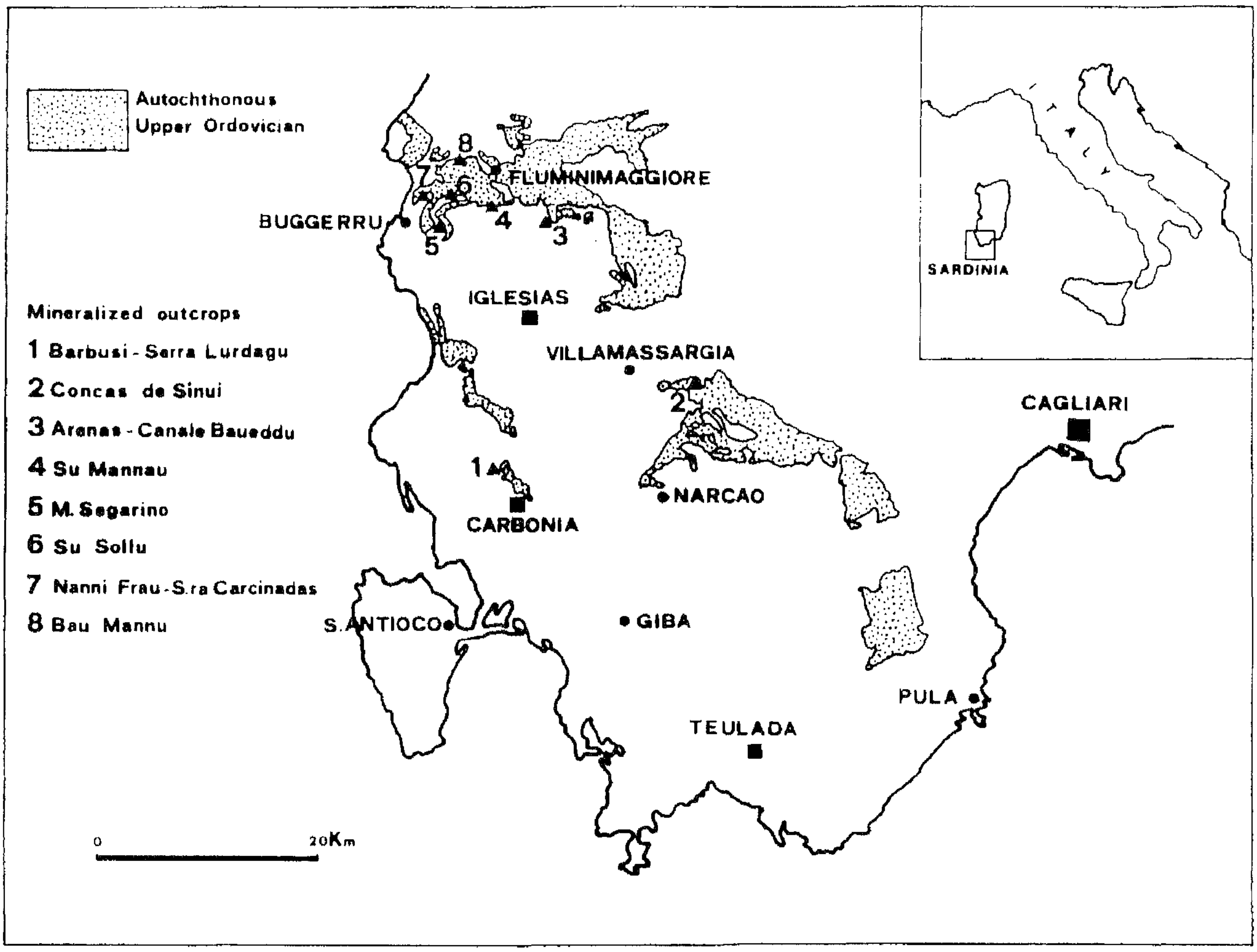

Fig. 1 - Geographic distribution of the mineralized sites

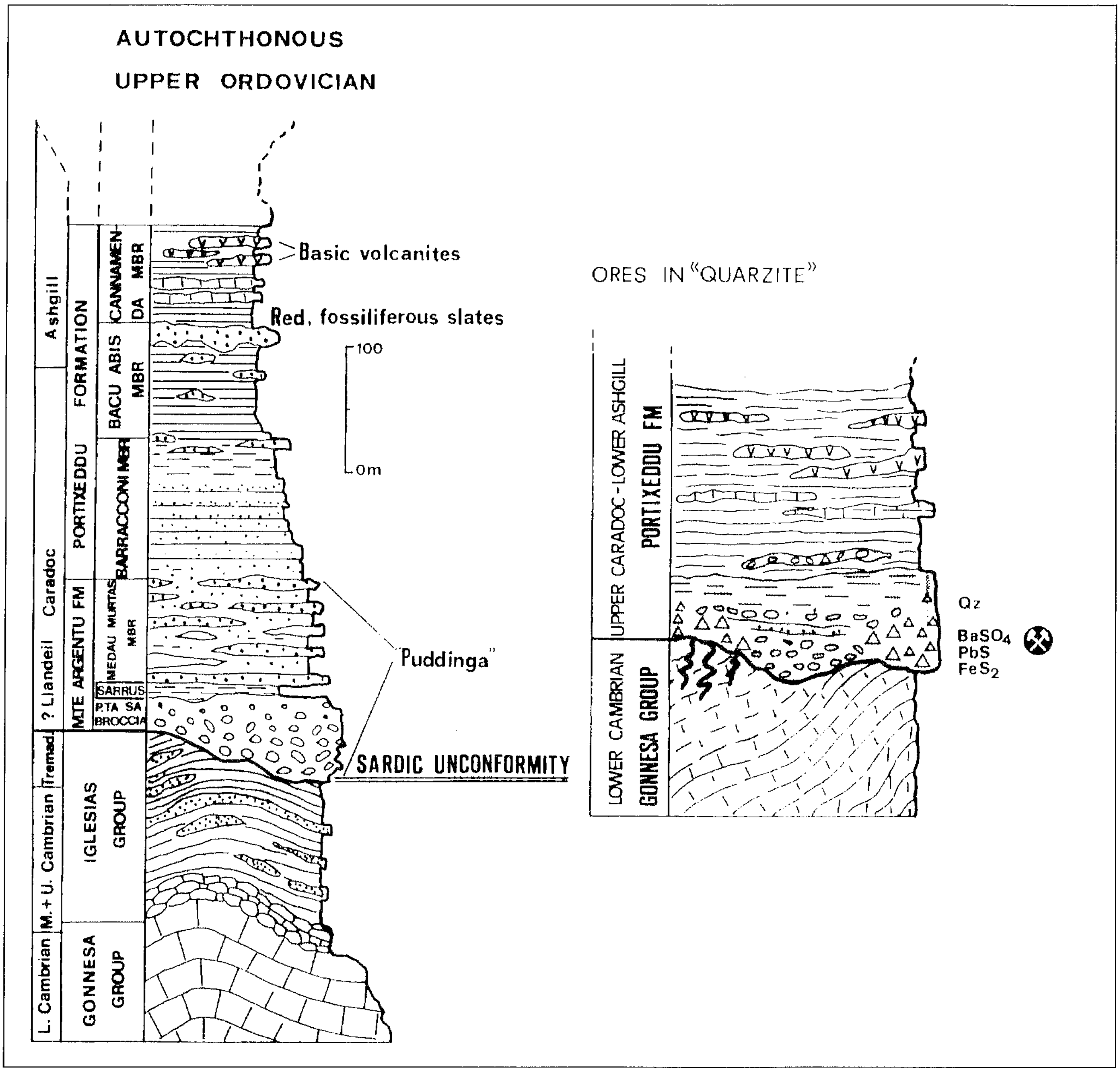

Fig. 2 - Stratigraphic column of the Middle-Upper Ordovician with the position of the mineralizations in the "Quarzite"

Formation, then evolving to the slaty Portixeddu Formation (Dümmen, 1990) (Fig.2). These sediments have variable thicknesses and are pervasively silicified when in angular unconformity on the Cambrian carbonates. Also the latter are often both silicified and mineralized along the unconformity. The Cambrian carbonates might have played as structural highs where the erosion (locally reaching down to the Dolomie Rigate lithotype), was particularly intense, before the Upper Ordovician transgression, possibly following a pattern of active tectonic lines.

The mineralized conglomerates and breccias underlie directly the Portixeddu Fm (Dümmen,1990), containing the well known Caradocian-Ashgillian fauna. Thin horizons of nodular barite, as well as monomictic barite breccias with often pyritic cement have been recorded in the upperlying slates.

In South-West Sardinia, even if hardly visible in the sequences containing the mineralization, magmatic products, consisting of metabasalts and metagabbros of alkaline affinity, are intercalated in the upper parts of the Portixeddu Fm. This magmatism has been compared to the basic occurrences emplaced in a tensional environment within lithospheric plates, possibly related, in this case, to the initial rifting of the paleo-European continental margin (Beccaluva et al., 1981). The magmatic activity in the area may continue at least until the Silurian (San Marco Fm, Dümmen, 1990).

A systematic analysis of several type-sections of the

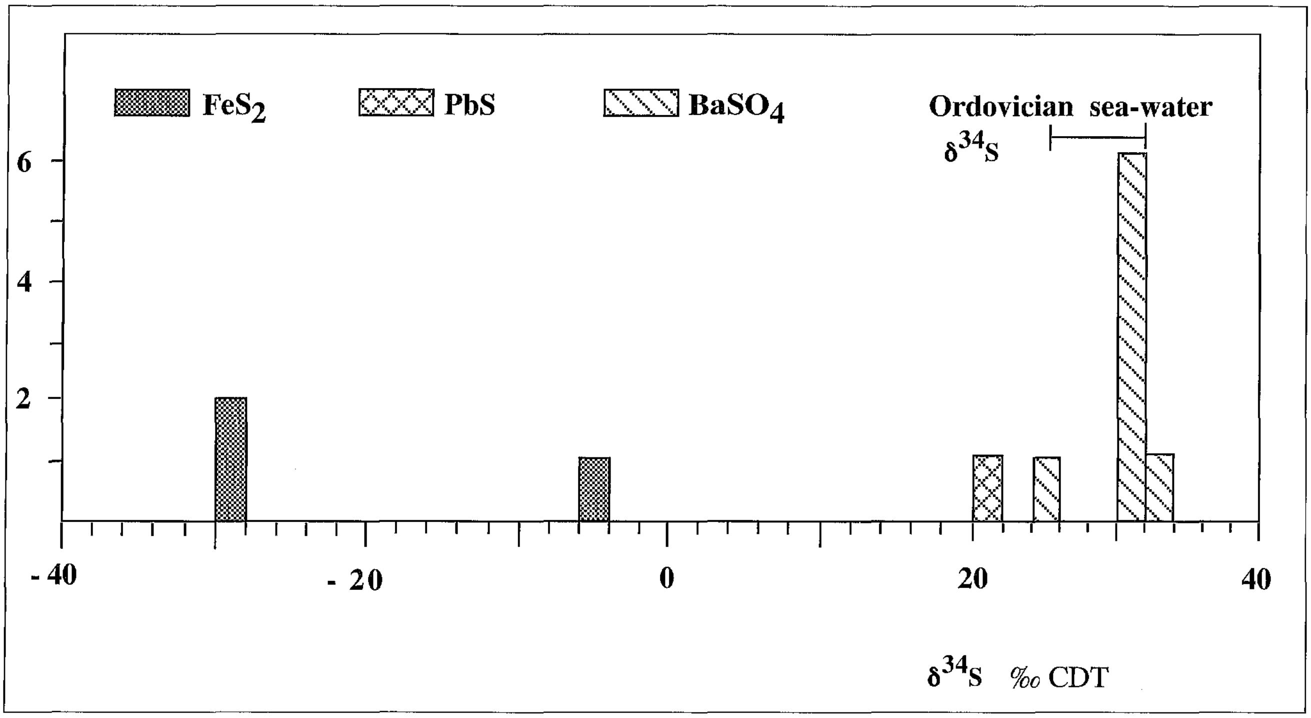

Fig. 3 - Sulfur isotopes in barite and galena of the Ordovician deposits

Ordovician ores in the Iglesiente-Sulcis area (Russo, 1990) revealed two fairly continuous metal-hosting belts with major local concentrations: the first one in the Northern Iglesiente and Fluminese, extending from the Buggerru coast in the West to the first outcrops of the Linas granite in the East, the second starting immediately north of the town of Carbonia (Sulcis) and reaching the easternmost outskirts of the Villamassargia village. In all the mineralized sites a strong pervasive silicification is always present, locally corresponding to massive horizons, 30-40 m thick, made-up of micro- to macro-crystalline quartz and chalcedony, whose precursors are often extremely difficult to recognize. This silicification is a necessary, but not the only evidence to look for in search of Ordovician minimal deposits. In fact, there are areas, like around the small town of Carbonia or east of the Mount Marganai group, where the "Quarziti" are extremely thick (more than 30 m of silicified breccia) and almost devoid of mineralization.

When the footwall of the unconformity consists of Cambrian carbonates, especially of the Ceroide limestone facies, a paleokarstic surface is sometimes developed, lined by small deposits of iron oxides.

S-isotopic data point to an origin from Ordovician sea water for the stratabound barites (mean $\delta^{34}S$ +31‰), whereas related galenas (mean $\delta^{34}S$ +20-21,5‰) (Fig.3) suggest sulphate reduction in a closed system.

The former economic mineralization is generally metasomatic and follows at least a first silicification phase. For both processes could be held responsible hydrothermal fluids circulating under an impermeable cover of slates, and discharging in the more porous and leachable lithotypes along the unconformity. It is possible, instead, that the stratiform barites in the slates and part of the pyrites might be syngenetic or early diagenetic. Owing to the characteristic position of the "Quarziti" as marker horizon in the Ordovician stratigraphy and to the S-isotopic ratios, we are inclined to consider the hydrothermal phenomena responsible for the mineralization as still of lower Paleozoic age (Boni et al.,1991). This genetic hypothesis contradicts the former assumption of a purely supergenic origin of the ores (Pretti et al., 1978; Padalino et al., 1973; 1989). Against a supergenic origin speak: the thickness of the accompanying "quarzites" (often higher than 40m), the dimensions of both quartz and barite crystals (from cm to dm), the irregular silicification reaching from the Cambrian substrate up to the Ashgillian slates and, last, but not least, the presence of the mineralization itself.

The geotectonic situation in the Upper Ordovician of Sardinia, with its intraplate volcanism, could have been favorable to the rise of the geothermal gradient and to the circulation of hydrothermal fluids.

The "Quarziti" of the Iglesiente-Sulcis might be also compared with their homologous of similar age (Upper Ordovician) in the eastern areas of the island, even if the latter are never mineralized.

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A TENTATIVE GEODYNAMIC MODEL FOR THE HERCYNIAN BASEMENT OF SARDINIA

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Abstract

Recent studies have confirmed the collisional nature of the European Hercynian belt and paleogeographic reconstructions place an Early Paleozoic oceanic zone between Armorica and the Gondwana plates in the southern region of the Armorican Massif. The suture zone of this ocean is now located in the Massif Central, in the External Crystalline Massifs of the Western Alps, and probably in the Maures and Northern Sardinia basement ("Posada-Asinara Line Suture Zone"). Analysis of the closure of this ocean allows us to place all of the most relevant Hercynian events that are recorded by the Sardinian basement into a new geodynamic model which implies a complete Wilson cycle. The most important stages of the model are as follows:

During Precambrian (?) - Lower Ordovician, several evidences support the existence of an old passive continental margin trending sub-parallel to the future Hercynian Chain. The sedimentary sequences of this margin, constituted by carbonate and terrigenous platform facies, crop out in southern Sardinia.

Over all of central and southeastern Sardinia ("Nappe Zone"), Middle Cambrian - Early Ordovician metasedimentary sequences are constantly overlain by a metavolcanic complex which occurs between the Arenigian and Caradocian times. This calc-alkaline magmatic complex was derived from an Andean-type continental arc.

At the end of the Ordovician subduction-related processes ceased and continental rifting began. Rifting is indicated by Late Ordovician alkaline to transitional WP magmatism and by the metasedimentary sequences of SW, central and SE Sardinia, which show a vertical facies transition from a syn- to a post-rift depositional setting. Therefore, in the Silurian, the Northern Gondwana margin was completely restored to a new passive margin.

During Silurian-Devonian times, there was thermal subsidence of the passive margin and pelagic sedimentation in the deeping basin as a result of worldwide climatic Silurian transgression. The increased carbonate sedimentation is probably related to the shifting of the Gondwana margin towards warmer paleo-latitudes.

From the Late Devonian to the Early Carboniferous continental collision began. The Armorican continental crust overthrusted the opposing Gondwanian passive continental margin and remnants of eclogitic oceanic crust were trapped along the Posada-Asinara Line Suture Zone. The collision was followed by crust-mantle detachment and crustal stacking with migration of deformation and metamorphism from the suture zone towards the foreland. Collision was accompanied by strike-slip in the suture zone.

The phases of convergence and collision are represented in the Gondwanian margin (LMGMC and Posada-Asinara Line belt) by two tectono-metamorphic events: an early Eclogite facies metamorphism and a subsequent prograde, medium pressure Barrovian metamorphism dated at about 350 Ma. In the overriding plate (HGMC), composed of a Precambrian(?)-Paleozoic polymetamorphic continental basement, an Hercynian essentially decompressional metamorphic history is detectable.

During the Early Carboniferous, sediments eroded from the chain, accumulated in the foreland, were strongly deformed (Culm type flysch).

In Middle and Late Carboniferous, the orogenic wedge collapsed causing ductile extension at mid- to lower- crustal levels, tectonic denudation and uplift of the chain. Low-angle normal faults and synmetamorphic shear zones are parallel to compression-related thrust surfaces and to the limbs of antiformal stacks. LP-HT metamorphism, which overprints earlier Barrovian metamorphism, accompanied extension (locally linked to the formation of metamorphic core complexes) and the emplacement of synkinematic granites which, in the axial zone, show a peraluminous anatectic nature and yield emplacement ages of about 300 Ma. The calc-alkaline plutonism, which formed the Sardinian-Corsic batholith, ended by about 275 Ma and is believed to have been emplaced during post-orogenic crustal extension at least partially concomitant with the development of the Late Carboniferous - Early Permian molasse basins.

KEY WORDS: *European Hercynian belt, Hercynian orogeny, Southern Hercynian Suture, Palaeozioic palaeogeography, Sardinian basement.*

1. INTRODUCTION

More then ten years ago it was discovered that the Sardinian basement is characterized by SW verging nappes interposed between the migmatitic complex of northern Sardinia and a folded external zone that crops out in the SW corner of the island (Carmignani et al., 1982 with references). Although a collisional tectono-metamorphic zonation was recognized, the uncertainty of the existence of oceanic crust in the orogene resulted in a long-standing interpretation of a completely ensialic evolution: "stacking" of thickened continental crust (Carmignani et al., 1980) or evolution linked to large scale transforms that were active repeatedly from the Upper Cambrian to Carboniferous (Vai & Cocozza, 1986).

Recently, however, many structural studies have confirmed the collisional nature and denied the ensialic model of the European Hercynian belt (Cogné, 1977; Mattauer & Etchecopar, 1976; Burg & Matte, 1978; Autran & Cogné, 1980; Bard et al., 1980; Matte & Burg, 1981; Behr et al., 1984; Matte, 1986) further stimulating geochemical and geochronological studies of metabasites with high pressure metamorphism relics that are scattered along the axial zone of the Hercynian Chain from southern Spain to Bohemia.

Many of these metabasites display geochemical ocean-floor basalt affinities (Bodinier et al., 1986; Bouchardon et al., 1989; Pin, 1990 with references) and several deeply metamorphosed and strongly tectonized mafic-ultramafic massifs can be regarded as metaophiolites (Cornwall: Kirby, 1979; Poland: Pin et al., 1988; Galicia: Bernard-Griffiths et al., 1985; Armorican Massif: Hammer, 1977, Paquette et al., 1985; Massif Central: Dubuisson et al., 1988; Western Alps: Ménot et al., 1988; Bohemia: Misar, 1984). Many geochemical and isotopic data of the metabasites support a depleted mantle source (Cabanis et al., 1982; Floyd, 1984; Bernard-Griffiths et al., 1985; Pin & Carme, 1987). The most frequent radiometric age of the protoliths is Early Paleozoic (Pin, 1990 with references) and more rarely Precambrian (Peucat et al., 1982; Paquette et al., 1985).

The most common ages of a HP-HT metamorphic event (HP granulitic/eclogitic facies) that overprinted the metabasites, range from Silurian to Lower Devonian (Peucat et al., 1982; Peucat, 1986; Pin & Lancelot, 1982; Paquette, 1987; Paquette et al., 1987; Quadt & Gebauer, 1988). These data can be easily explained only in terms of subduction of oceanic lithosphere before the late Devonian (?) - Early Carboniferous continental collision.

The matter regarding how and when closure of the ocean occurred, is still open (Matte, 1986; Paquette, 1987; Bodinier et al., 1986, Pin, 1990). However, many authors currently believe that remnants of oceanic crust are involved

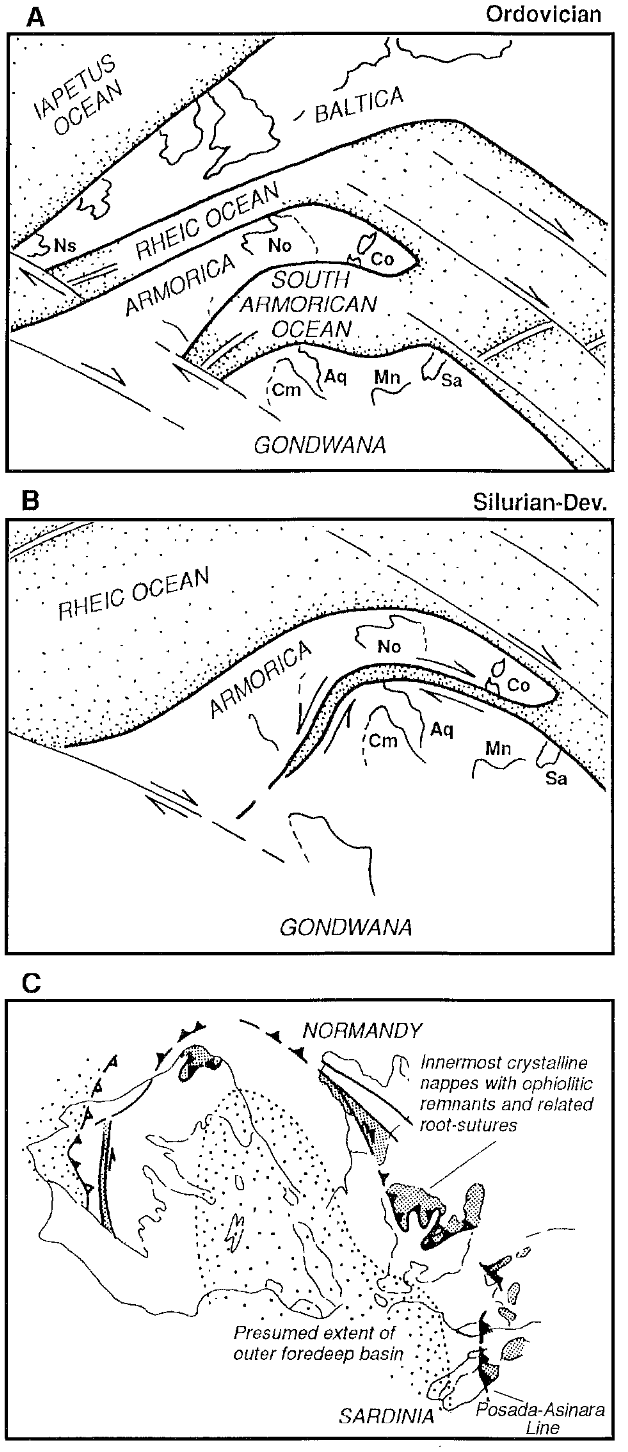


Fig. 1 - A and B - Ordovician and Silurian-Devonian paleogeography. Stippled: Oceanic crust; White: continental crust; Sa: Central and Southern Sardinia (LMGMC); Co: Northern Sardinia and Corsica (HGMC); No: Normandy; Cm: Cantabrian Mountains; Aq: Aquitaine; Mn: Montagne Noire; Co: Cornwall; Ns: Nova Scotia (From Vai & Cocozza, 1986; Paris & Robardet, 1990, Vai, 1991; modified).C - South European Hercynian Chain (From Matte, 1986 and Franke, 1989; modified).

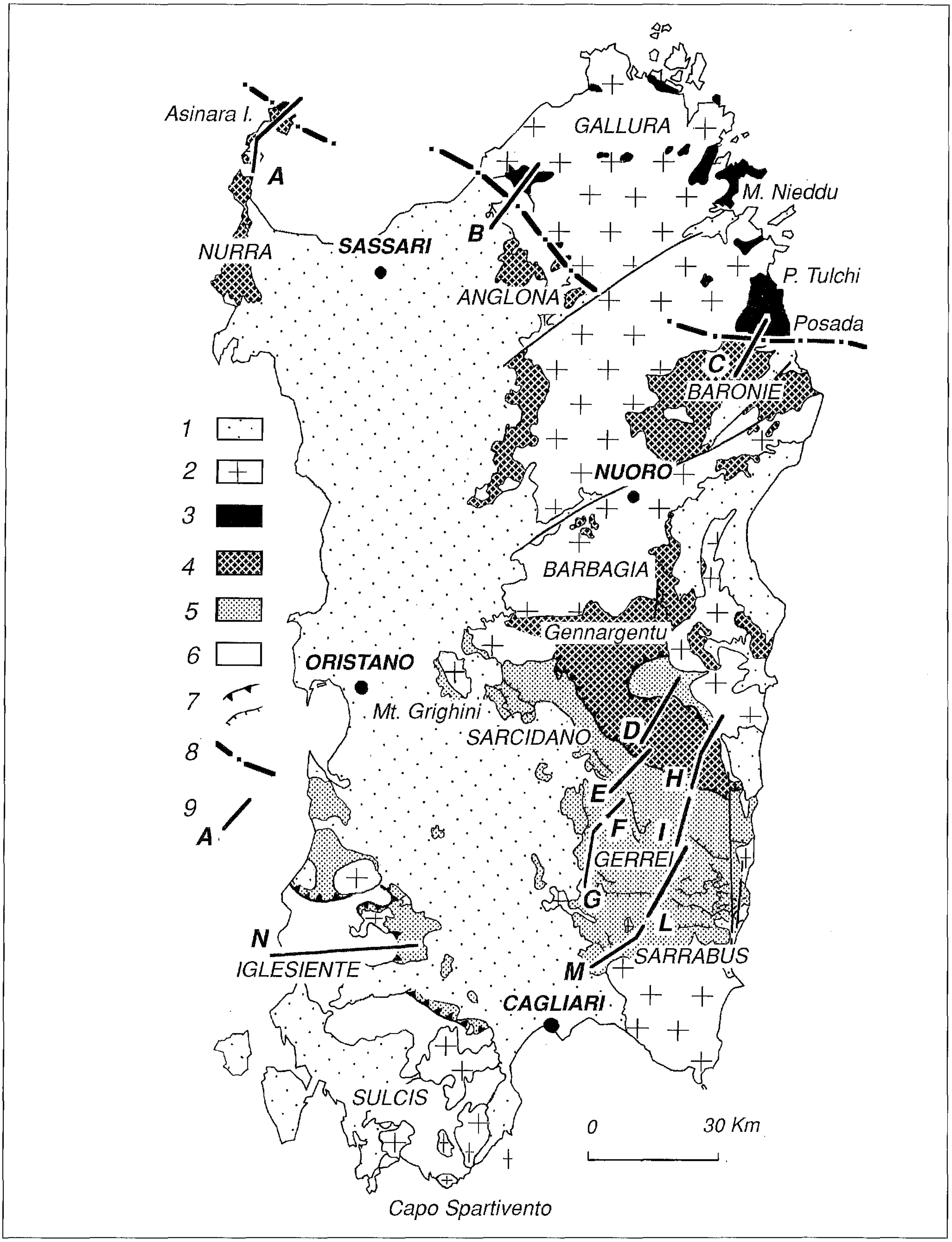


Fig. 2 - Main structural elements of the Sardinian Basement. 1: Post-Hercynian cover; 2: Hercynian batholith; 3: High Grade Metamorphic Complex (HGMC); 4: Internal Nappes; 5: External Nappes; 6: External Zone; 7: Major and minor thrusts; 8: Posada-Asinara Line; 9: Cross section traces of Fig. 3 and 7.

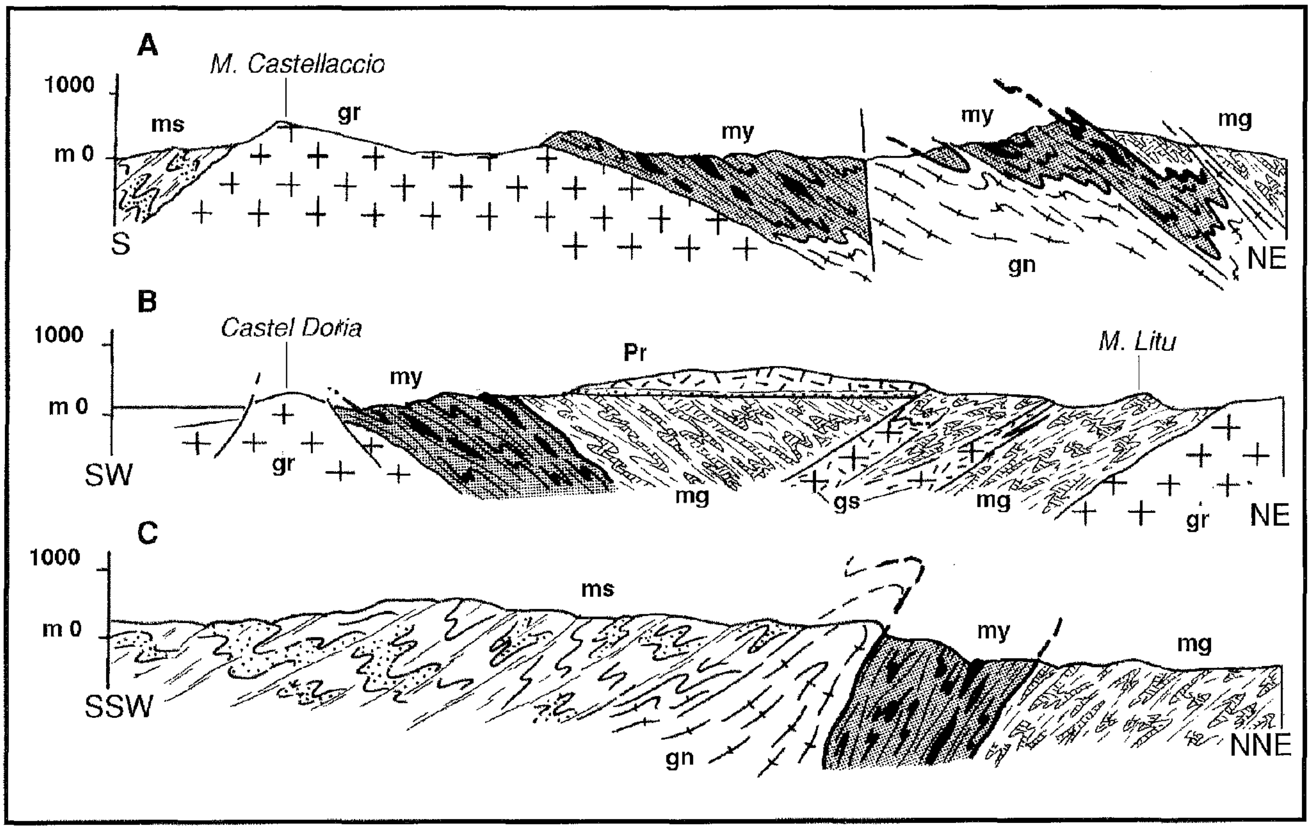


Fig 3 - Geological sections across the Posada-Asinara Line (traces in Fig.2). A: Geological cross section through Asinara Island . B: Geological cross section through Southern Gallura. C: Geological cross section through Posada Valley.
Pr: Permian volcanic and sedimentary deposits. gr: Granite. gs: synkinematic peraluminous granite. <u>High Grade Metamorphic Complex (HGMC)</u> - mg: migmatite of sillimanite + K-feldspar zone. <u>Posada-Asinara belt</u> - my: micaschist, paragneiss and quartzites of staurolite + biotite and kyanite + biotite zones and amphibolites with relics of granulite and eclogite. <u>Low to Medium Grade Metamorphic Complex (LMGMC)</u> - ms: micaschist and paragneiss of garnet + oligoclase, garnet + albite zones and phyllites and metasandstone of biotite zone; gn: orthogneiss.

in the Hercynian orogenesis (Bodinier et al., 1986; 1988 with references), and that paleomagnetic (Van der Voo et al., 1980; Perroud et al., 1984; Bonhommet & Perroud, 1986) and paleogeographic (Behr et al., 1984; Paris, 1990; Paris & Robardet, 1990) data supports closure of the oceanic basin during the Hercynian orogeny. Some paleogeographic reconstructions place an Early Paleozoic oceanic zone in the southern region of the Armorican Massif (e.g. Massif Central Ocean: Matte, 1986; South Armorican Ocean: Paris & Robardet, 1990) between the Armorica and Gondwana plates (Fig. 1a), but palaeobiogeographical data no support the hypothesis of a very wide Palaeozoic ocean (Vai, 1991). The suture zone of this ocean is now located in the Massif Central, in the External Crystalline Massifs of the Eastern Alps, and probably in the Maures ("Southern Hercynian Suture": Bodinier et al., 1986; Matte, 1986).

Structural evidence and lithologic and stratigraphic affinities show that the Sardinian and Corsican basement was linked to the Provencial Basement before the Miocene (Cherchi & Montadert, 1982; Cherchi & Trémolierès, 1984 with references) and that the Carboniferous suture between the Armorica and Gondwana plates cuts northern Sardinia (Fig. 1C) along the "Posada-Asinara Line" which separates the High Grade Metamorphic Complex (HGMC) from the Medium to Low Grade Metamorphic Complex (LMGMC) (Fig. 2 and 3).

The Posada-Asinara Line is a belt characterized by the presence of small bodies of amphibolite with granulite relics, eclogite relics (Miller et al., 1976; Ghezzo et al., 1982) and relict mylonitic textures that developed under high grade metamorphic conditions (Carosi & Elter, 1989). Geochemical and geochronological data studies indicate a MORB origin of the amphibolite protolith and an age of about 950 Ma (Cappelli et al., 1991). This relatively old age could imply: a) a long lasting oceanic basin between the Gondwana and Armorica plates (Perroud & Bonhommet, 1981) that formed during the Precambrian, and subducted underneath N-Gondwana margin since the Lower Paleozoic. After an interruption in the Upper Ordovician, this convergence continued causing subduction in opposite direction underneath Armorica plate; b) alternatively, the old oceanic crust might have been incorporated into a Precambrian orogen, possibly by obduction (Bernard -

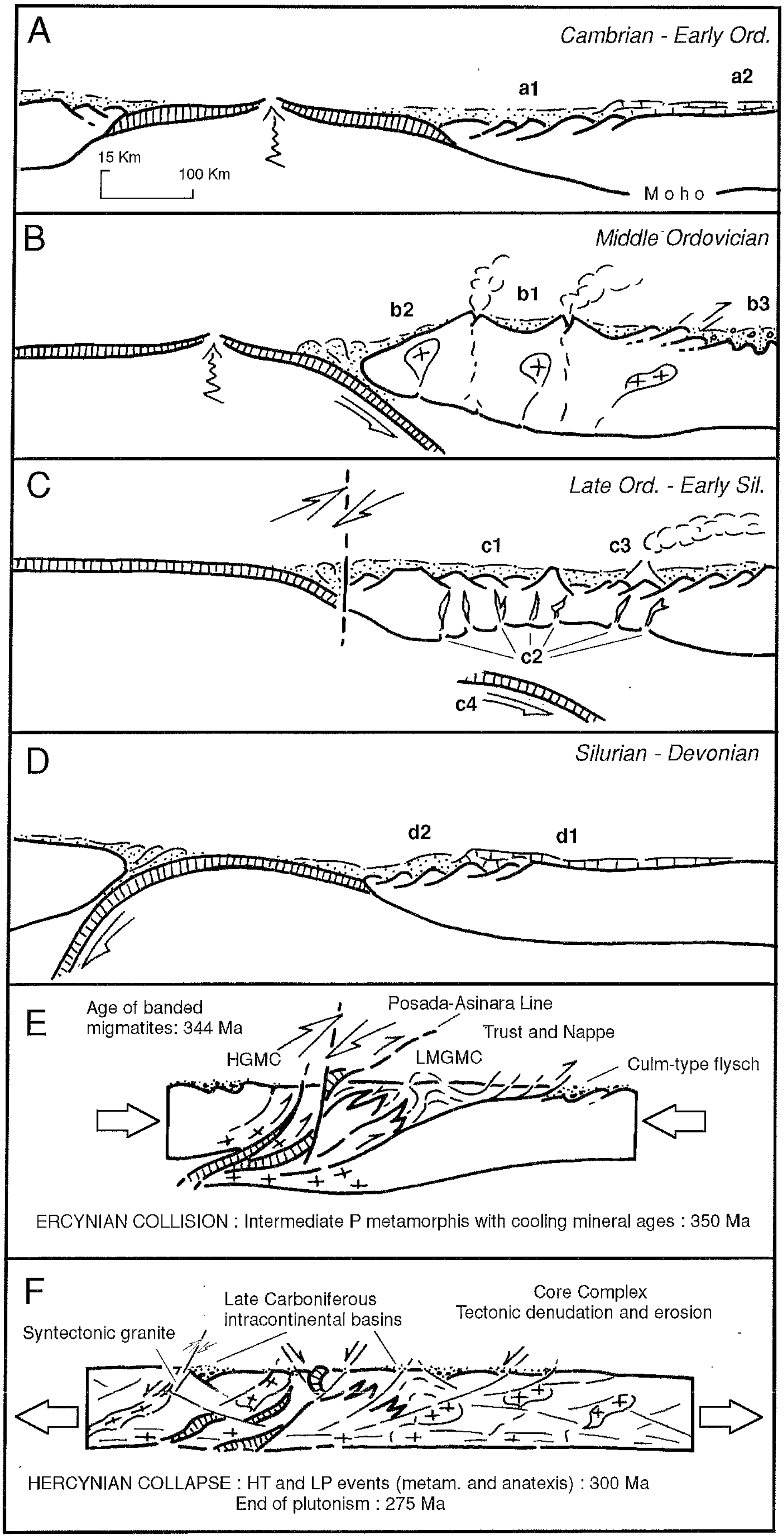


Fig. 4 - Skecth of the geodynamic evolution of the Hercynian Basement of Sardinia. See explications in the text.

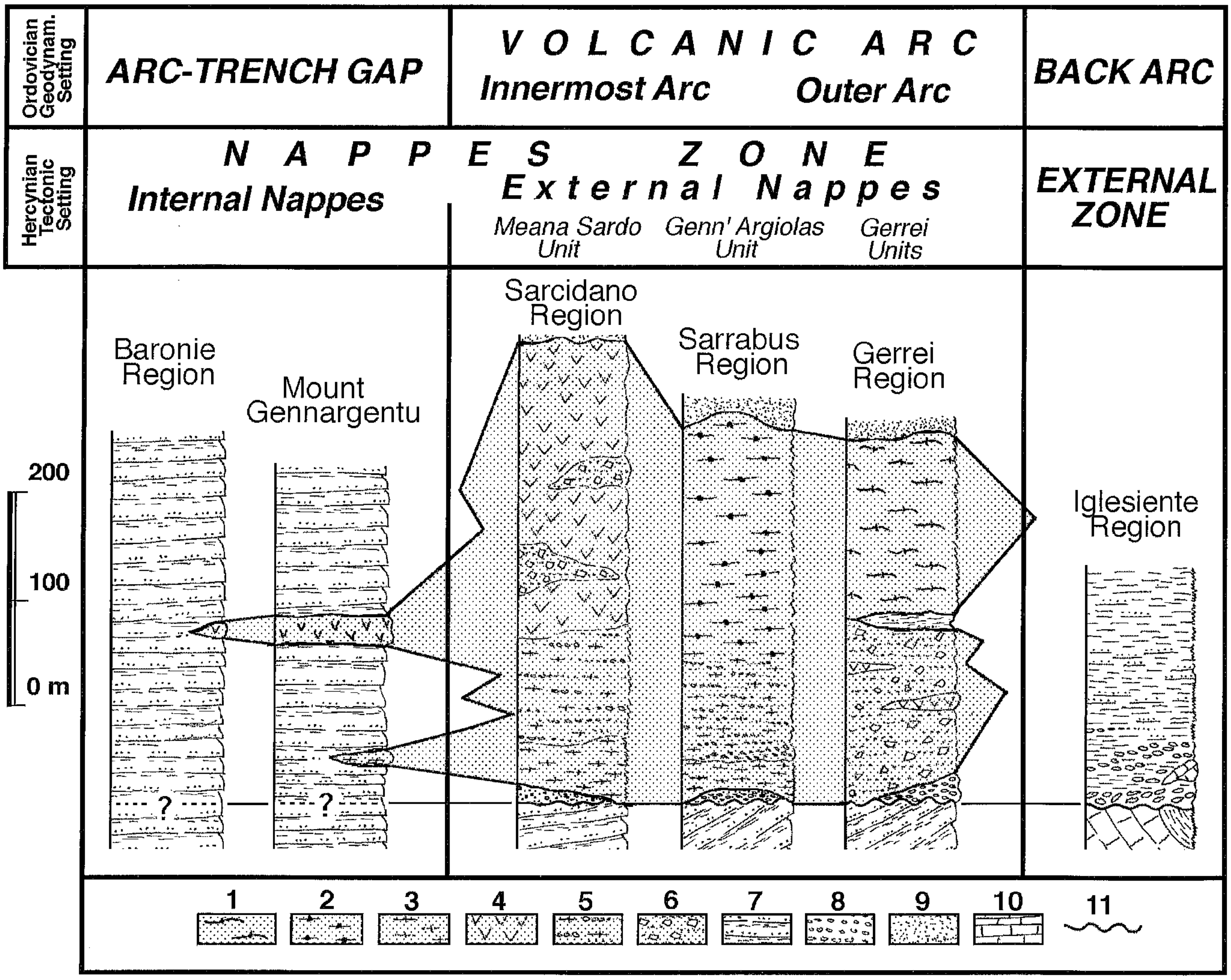


Fig. 5 - Middle Ordovician volcanic sequences (restored original paleogeographic setting).
1: Metarhyolites and metarhyodacites with "augen texture" ("Porfiroidi" Fmt.). 2: Metadacitic to metarhyodacitic volcanics ("Porfidi Grigi" Fmt.). 3: Metarhyolites ("Monte Corte Cerbos" and "Porfidi Bianchi" Fmts). 4: Metandesites and metadacites lavas ("Serra Tonnai" Fmt.). 5: Metamorphosed reworked acid volcanics (Manixeddu Fmt.). 6: Metamorphosed reworked intermediate volcanics. 7: Metasandestones and shales. 8: Metaconclomerates. 9: Metarkoses. 10: Cambrian sequence of Iglesiente-Sulcis region. 11: Sardic Unconformity.

Griffith et al., 1985; Paquette et al., 1985), and was later metamorphosed under eclogitic conditions during the Hercynian Orogeny.

The terranes north and south of Posada-Asinara Line have different metamorphic histories. In fact the line divides a migmatitic complex from a medium to low grade zone and is interpreted to represent an oceanic paleosuture that was squeezed between the crystalline basement of an overthrusting hinterland and the cover rocks of an underthrusting continental margin.

In this work we propose a new geodynamic model of Sardinian basement that implies: (a) B-type subduction followed by (b) continent-continent collision, (c) stacking of Gondwanian continental margin and finally, at the end of convergence, (d) gravitational collapse of the Hercynian orogenic wedge.

2. Pre-Collisional Evolution

2.1 *The Cambrian Passive Margin (Fig. 4 A)*

MORB type metabasites testify to an oceanic hiatus between Northern and Central-Southern Sardinia. The lack of both calc-alkaline volcanism and deformation in the Lower Cambrian (or Precambrian ?) - Lower Ordovician epicontinental sequences supports the existence of an old passive continental margin. The sedimentary succession of this margin crop out in southwestern Sardinia. It is a (?)Precambrian - Lower Ordovician succession constituted by only carbonate and terrigenous platform facies.

The oldest deposits consist of mainly terrigenous metasediments represented by feldspatic metasandstones, quartzites, metaconglomerates and thin dolomitic intercalations that grade upwards into shales, metasiltites and metasandstones (Bithia Formation) (Junker & Schneider, 1983). In the lower part of the sequence believed to be Upper Precambrian (Cocozza, 1980; Minzoni, 1981; Gandin, 1987), basic and intermediate metavolcanics occur (Tucci, 1983) possibly due to a rifting stage .

The Bithia Formation represents a terrigenous shelf deposit, perhaps transgressive on the Gondwanian continental margin. The same sedimentary features found in part of the Bithia Formation also characterize the lower part of the overlying Nebida Formation (Matoppa Member). Basic spilitic metavolcanics are interlayered within this mainly terrigenous deposit (Maccioni, 1967), which could represent an intracratonic rift that began in the Precambrian. The Nebida Formation is represented by prevalent terrigenous metasediment with minor intercalations of oolitic limestones bearing Lower Cambrian Archeocyats, Trilobites and algal stromatolites. This succession is believed to represent an ancient continental shelf environment with eastward prograding deltaic systems (Matoppa Member) that evolved in an oolitic lagoonal environment. The mainly terrigenous Nebida Formation grades upward into a thick carbonate succession consisting of dolostones and limestones (Gonnesa Formation) that represented an arid tidal flat system in which deeper basins formed intermittently due to extensional phases (Boni & Cocozza, 1978; Boni & Gandin, 1980; Fanni et al., 1982; Vai, 1982). The "drowing" of this carbonate platform is seen by nodular limestones ("Calcescisti" Auct.) rich in Middle Cambrian trilobites, echinoderms and brachiopods. The overlying deeper environment deposit consist of a 400 m thick neritic terrigenous succession in which the younger levels contain achritarcs and graptolites (*Dictyonema flabelliforme*) (Barca et al.,1987) of Tremadocian age ("Argilloscisti di Cabitza", Auct.).

These lagoonal and epicontinental carbonate and terrigenous deposits, correspond to thick siliciclastic sequences in the Nappe Zone (S.Vito and Solanas Fmts.). Their facies represent slope conoids (a1 in Fig. 4A) hence a more distal facies with respect to the mainly carbonate Sulcis - Iglesiente facies (a2 in Fig. 4A). The composition of these siliciclastic deposits suggests a provenance from an old crystalline basement located to the southwest. The lower part of this sequence aside from the lack of carbonate intercalations, appears sedimentologically similar to the Matoppa Member of Nebida Formation. The partial correlation of the Sulcis - Iglesiente Cambrian succession and the more distal detrital succession (San Vito and Solanas Fmts.) is also supported by paleontological data (Barca et al., 1981a,b; Naud & Pittau, 1985; Tongiorgi et al., 1984; Albani et al., 1985).

The transition from distal slope deposits to a terrigenous - carbonate continental shelf and a emerged source area to the southwest points to a passive Gondwanian continental margin since Infracambrian (?) - Cambrian time.

2.2 *Subduction Under Gondwana Continental Crust Causing an Andean Type Volcanic Arc (Fig. 4 B).*

Over all of central and southeastern Sardinia (Nappe Zone), Middle Cambrian - Early Ordovician metasedimentary sequences are overlain by a metavolcanic complex (Fig. 5). This volcanic cycle occurred between the Arenigian and Caradocian and consists of a great number of effusive episodes with intrusions into the pre-volcanic basement. The Middle Ordovician ("pre-Caradocian") magmatic activity constitutes a complete "suite" ranging in composition from basaltic-andesitic to rhyolitic, clearly of sub-alkaline affinity, with acid (rhyolite-rhyodacite) compositions more abundant than intermediate and basic (andesite-basalt) compositions (Memmi et al., 1982; 1983). This suite is also characterized by the presence of interlayered continental metasediments and by the calc-alkaline nature of the basic rocks. Such characteristics have been interpreted as typical of volcanic arc activity (Garbarino et al., 1981) or late- to post-orogenetic magmatism (Memmi et al., 1982; 1983; Carmignani et al., 1986).

The suite of metamorphosed igneous rocks of calc-alkaline affinity is similar, both in age and geochemical character, to other igneous products occurring practically everywhere in the Hercynian massifs of the Mediterranean area (Spanish Basement, Pyrenees, Massif Central, Maures, Alps, etc.).

In the restored, original paleogeographic setting of the stacked Hercynian tectonic units, the Middle-Ordovician volcanic activity shows a compositional polarity and probably younging direction from NE to SW (Fig. 5). The Mount Gennargentu and Baronie region are characterized by the scarcity of Ordovician volcanic rocks. In the Sarcidano region, where the thickness of Arenigian to Caradocian metavolcanic-sedimentary complex can reach 400-500 m, volcanic activity is characterized by abundant intermediate rocks with andesites and dacites ("Serra Tonnai Formation") and minor acid rocks (metarhyolites of the "Monte Corte Cerbos Formation").

In the Sarrabus region (Genn'Argiolas Unit) the metavolcanic complex is composed of originally dacitic to rhyodacitic lava domes, lava flows, ignimbrites or tuffs. The lower part of the sequence consists of minor rhyolitic flows ("Porfidi Bianchi") and their reworking ("Riu Ceraxa Conglomerate"). Most of the volcanics are dacitic to rhyodacitic in composition ("Porfidi Grigi"). In the Gerrei units, the base of the metavolcanic complex is made of reworked andesitic volcanics and rare andesitic flows. Above this basal sequence metarhyodacites and dominant metarhyolites up to 200 m thick, with "augen texture" and sometimes with "large phenocrysts" of K-feldspar ("Porfiroidi"), crop out.

A probable SW direction of younging of Ordovician volcanics can be inferred by the presence of thick terrigenous transgressive Caradocian sediments in the Sarcidano-Barbagia successions (about 400 m in the Sarcidano region: Bruncu su Pizzu Formation; up to 150 m in the Sarrabus

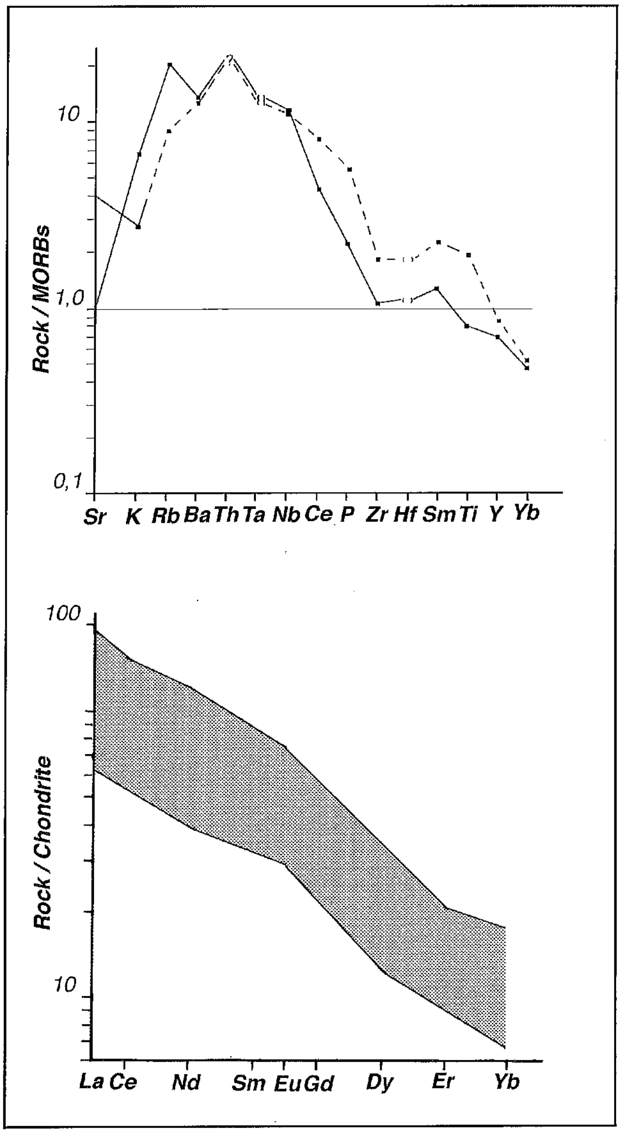


Fig. 6 - a) Spider diagram for upper Ordovician WP metabasalts of Sardinia. Values normalized to MORBs (after Pearce, 1983). Ta and Hf calculated after Jochum et al. (1986); solid line: External Zone and External Nappe; dashed line: Internal Zone and Internal Nappe.
b) Range of Chondrite REE normalized patterns (after Evensen et al., 1978) of Upper Ordovician WP metabasalts of Sardinia

region: Punta Serpeddì Formation) compared to the few meters of siliciclastic sediments (sometimes missing) immediately overlain by the typical Ashgillian limestones in the Gerrei region.

Plutonic rocks emplaced in the pre-Hercynian cover show geochemical features and ages (Beccaluva et al., 1985) comparable to those of the calc-alkaline volcanics.

The subalkaline character of the magmatic activity combined with the prevalence of acidic effusives and the occurrence of large amounts of pyroclastic flows are characteristic of an orogenic suite involving continental crust.

The hypothesis of a continental crust-supported arc connected to a SW subduction of oceanic crust, as shown in Fig. 4B, is suggested by the following pieces of evidence (Fig. 5) : (a) progressive increase of silica content in the volcanics towards the SW (from Sarcidano to Gerrei); (b) probably younging of the magmatic activity in the same direction; (c) the great volume of andesite-derived clastics at the base of the volcanic sequence in the outer arc (Gerrei Unit) probably formed from the pre-existing innermost arc volcanic domains; (d) occurrence of shallow depth plutonic rocks.

These data also suggest that the volcanic arc (b1 in Fig. 4B) migrated towards the hinterland in a time spanning between Arenig to Caradoc and that the melting involved progressively more and more continental crust.

The arc-trench gap (b2 in Fig. 4B) was incorporated in the innermost Hercynian nappes (Mount Gennargentu, Baronie region) which are characterized by the shortage of Ordovician magmatic rocks (Fig. 5).

The back-arc basin (b3 in Fig. 4B) in the Iglesiente region (fig. 5) is devoid of calcalkaline magmatism and underwent a pre-Hercynian compressional event ("Sardinian Phase").

This post-Tremadoc and pre-Caradoc phase of deformation, known in many parts of Europe, is very evident here, especially in the Iglesiente region, where the Cambrian-Lower Ordovician sequences was folded and eroded before the Caradocian age.

The products of this erosion can reach several hundreds of meters in thickness ("Puddinga" Auct.). This angular unconformity has also been reported in the Sarrabus (Calvino 1961; 1967; Naud, 1981) and Gerrei regions and in central Sardinia the Cambrian - Lower Ordovician successions are often separated from the Middle to Upper Ordovician volcanics and sedimentary rocks by conglomerates derived over all from the volcanic arc.

However :

— in central Sardinia pre-Hercynian compressive structures, like those of the Iglesiente region, have never been described;

— in southwestern Sardinia and particularly in the Iglesiente region, the post-unconconformity conglomerate ("Puddinga "Auct.) is much thicker and is often composed of coarse grained syntectonic deposits with Cambrian limestone olistoliths (Oggiano et al., 1986; Martini et al., 1991).

— in the same region the conglomerates unconformably cover formations very different in age: from Lower Cambrian to Tremadocian; on the contrary in southeastern and central Sardinia a large number of biostratigraphic studies have not revealed an important stratigraphic hiatus between the Middle Ordovician volcanics and the Cambrian-Lower Ordovician metasedimentary sequence.

These observations suggest that pre-Hercynian compressive tectonics decreased from the back-arc to the volcanic arc.

According to Coney (1973) behind the continental margin volcanic arc, a back-arc thrust belt and associated syntectonic compressive basin could have been present. This process is described in several active continental margins, such as in Chile, were landward migration of the trench causes compression and strong intraplate coupling and back-thrusting (Uyeda, 1981).

We believe that the "Sardinian Uncorformity" was caused by back-arc compressive tectonics.

2.3 *The End of Ordovician Subduction and Volcanic Arc Collapse (Fig. 4C)*

Both the continental clastics of the External Zone (b3 in Fig. 4B) ("Puddinga" Auct.) and Middle-Ordovician metavolcanics of central - southeastern Sardinia (b1 in Fig. 4B) are capped by terrigenous continental to littoral sediments with late Ordovician alkalibasaltic intercalations that show a large variability in thickness and facies ("Caradocian transgression") (c1, c3 in Fig. 4C). This succession grades upward to neritic argillitic and carbonate deposits ("Ashgillian limestones") followed by uniform deposits of Silurian black shales and black cherts. These vertical facies transitions to be the result of a change of depositional setting from syn-rift to post-rift (Vai, 1982; 1991), the former due to magmatic arc collapse and the latter to thermal subsidence and silurian positive eustatic variation.

In this way, the North Gondwana margin was completely restored to a new passive margin, in the Silurian.

The cessation of any magmatic activity after Lower Silurian is supposed by the absence of volcanic rocks in the successions until the occurrence of lower Carboniferous Within Plate (WP) volcanism. This volcanism might have been related to wrench dynamics linked to the ensialic stages of the Variscan collision (Vai, 1982; Vai & Cocozza, 1986; Di Pisa et al., 1991; Vai, 1991).

The magmatic quiescence for about 100 Ma, combined with the stratigraphic sequences pointing to a pelagic type sedimentation (see above), indicates that, at least in this time span, the Gondawanian continental edge behaved as a subsiding passive continental margin. This interpretation appears to be consistent with a change from subduction-related event to a subduction-stopping one.

The cause of this dramatic tectonic change and the manner in which subduction ceased are two foundamental problems.

An understanding of the geodynamic evolution that could have led to subduction-stopping can be gained by the geochemical character of the Caradoc-Ashgillian magmatism that is represented by dike swarms and effusive volcanics (c2, c3 in Fig. 4c).

Even though metamorphism may have modified the original content of some discriminant chemical elements, these basic rocks show WP character (low Zr/Nb and strongly differentiated REE patterns and generally medium-high Ti/Y and Nb/Y ratios, fig. 6) that generally chatacterizes magmas derived from partial melting of the mantle during extension (as already proposed for Sardinia by Ricci & Sabatini (1978) and Memmi et al., (1983). The widespread distribution of these WP metabasalts along the dismembered Middle-Ordovician calc-alkaline volcanic arc and along the back-arc hinterland may represent the magmatic manifestation of this kind of environment that occurred after the subduction related volcanism.

Nevertheless slight differences have been observed between upper Ordovician metabasalts coming from the inter-arc region (Internal Nappe Zone) and coming from the back-arc region (External Nappe Zone and External Zone) which appear to reflect some differences in the geochemical nature of the respective source regions in the mantle. The former possibly deriving by the intersection of incompatible elements enriched magmas (OIB-Type component) with a mantle which previously suffered a subduction zone related modification near the Ordovician trench; the latter possibly deriving from sub-continental mantle volumes where the subduction component is strongly dilute in a thicker mantle wedge above the descending slab (i.e. far from the trench) so that the HFSE enriched component predominates (Di Pisa et al., this volume).

A similar picture is possible if the OIB-type component, whatever the mechanism, proceed from sub-lithospheric sources (convecting upper mantle? deep mantle reservoirs ?) and if no physical barrier exists to prevent its interaction with the mantle wedge above the subducting slab, that had previously been depleted by arc magmatism (Saunders & Tarney, 1984).

Detachment of the subducting oceanic lithosphere and its sinking into the sub-continental astenosphere must have occurred to produce a sufficiently wide lithospheric window through which an enrichment component could have risen up from the mantle.

Among the possible mechanism that could have modified pristine converging framework, slab roll back, variation in dip angle of subduction, ridge collision, the latter seems the most likely to explain the observed character of the alkaline-to-transitional Late Ordovician metabasic rocks in Sardinia (Nelson & Forsythe, 1989; Ormerod et al., 1988). The former possibilities, depend upon variations in the speed of convergence, may only account for variations in the spatial distribution of arc magmatism.

Alternatively, a ridge-trench collision could have caused the cessation of arc volcanism in a wide area ("no-slab area" after Dickinson & Snyder, 1979) and the release of the compressive stress field produced by the subduction zone (Scholz et al., 1971). The regional extension may have allowed inter-arc spreading that enabled the emplacement of sub-lithospheric magmas thereby interacting with a subduction component. A similar picture has been described by Ormerod et al. (1988) and Fitton et al. (1988) in the tectonic and magmatic transition from a supra-subduction zone to a within plate environmentin the Western Great Basin (USA).

2.4 *The Silurian-Devonian Passive Margin (Fig. 4D)*

On the new restored North Gondwanian margin the pervasive Silurian black shales pass continuosly into Lower and Middle Devonian pelagic marly shales and nodular Tentaculite-bearing limestone. From Middle Devonian up to Lower Carboniferous (Tournaisian) thick pelagic limestones are common, particularly in the Gerrei region (d1 in Fig. 4D). These limestones are replaced by thick terrigenous sequences in Nurra (Di Pisa & Oggiano, 1984) and in other internal areas (Baronie region) (d2 in Fig.4D). Increasing carbonate sedimentation, also documented in other Southern European areas, could be due to the shifting of Gondwanian margin towards warmer paleo-latitudes (Vai, 1976; 1982; Babin et al., 1980).

The discontinuity of the Upper Devonian carbonate platforms may be attributed to a period of strike slip tectonic activity leading to the fragmentation of the continental shelf to form pull-apart basins (Vai & Cocozza, 1986, Vai, 1991). These basins are associated with basic alkaline volcanism that has been attributed to the early stage of collision between the Gondwana and Armorica plates (Di Pisa et al., 1991).

In the externalmost platforms the carbonate sedimentation was suddenly interrupted and sedimentation changed to Culm type syn-orogenic deposits (Spalletta & Vai, 1982; Maxia, 1984; Barca & Spalletta, 1985; Barca, 1991).

3. Collisional Structure

The Hercynian collisional event is well preserved in the Sardinian basement: the overthrusting continental margin is represented by the "High Grade Metamorphic Complex" (HGMC) (Fig. 4E) that crops out in Northern Sardinia and Corsica; the overthrust continental margin is represented by the "Low to Medium Grade Metamorphic Complex" (LMGMC) which crops out in Central and Southern Sardinia. The two complexes are separated by the "Posada-Asinara Line" suture zone (Fig. 2 and 3).

The convergence and collision are recorded by two tectono-metamorphic events: an early eclogite facies metamorphism and a subsequent medium pressure Barrovian type metamorphism.

—*Early Eclogite facies metamorphism.* This event is preserved in the HGMC and along the "Posada-Asinara Line" as relics in amphibolite facies metabasites that formed part of an oceanic crust.

In Sardinia this metamorphic event has not been dated, but eclogite facies metamorphism in the Massif Central and Armorican Massif has been dated at between 430 to 380 Ma (Peucat & Cogné, 1977; Peucat et al., 1982; Ducrot et al., 1983; Postaire, 1983; Paquette, 1987; Paquette et al., 1987; Quadt & Gebauer, 1988) and has been interpreted as a subduction stage (Matte, 1986; Bodinier et al., 1986; Paquette, 1987; Pin, 1990).

—*Barrovian metamorphism.* The rocks belonging to the LMGMC and the Posada-Asinara Line Belt have experienced Barrovian metamorphism caused by collision and crustal stacking.

Metamorphism was prograde and varies from greenschist to amphibolite facies moving from the external towards the internal nappes and Posada-Asinara Line belt.

In the Internal Nappes Rb/Sr and Ar/Ar metamorphic cooling ages of muscovite and amphibole (Del Moro et al., 1991) within the Barrovian metamorphic assemblages, are approximately 350 Ma. This age together with structural and textural evidence, is interpreted to represent the minimum age of intracontinental collision. A similar (Rb/Sr) age from a banded migmatite within the HGMC (344 Ma; Ferrara et al., 1978) suggests that each system passed through the respective blocking temperature for strontium at the same time.

The trondjemite leucosomes point to a metamorphic differentiation through pressure solution mechanisms (Lindh & Whalgren, 1985; Sawyer & Barnes, 1988) during the collisional phase. The age of 344 Ma proposed by Ferrara et al. (1978), which represents the closure with respect to the radiogenic Sr excange between the migmatite bands can be ascribed to the late stages of Hercynian collision or early uplift.

3.1 *The High Grade Metmorphic Complex(HGMC)*

The innermost part of the chain is composed of intermediate pressure amphibolite facies migmatite and migmatitic gneiss (Sillimanite and K-feldspar zone, Franceschelli et al., 1982). These rocks generally occur as small roof pendants in the late-orogenic calcalkaline batholith. The Hercynian age of the amphibolite facies metamorphism is based on radiometric ages (Beccaluva et al., 1985).

The migmatite complex protoliths are probably Precambrian because they are intruded by Ordovician orthogneisses (Orthogneiss of Tanaunella, 458 ± 31 Ma; Di Simplicio et al., 1974) and at Argentella (NW Corsica) they are unconformably overlain by a very low grade metamorphic sequence containing Early Paleozoic sediments (Baudelot et al., 1977; Menot & Orsini, 1990).

As in the Massif Central and in the Crystalline Massifs of the Eastern Alps, the Sardinian HGMC contains mafic and ultramafic rocks, some of which show a polymetamorphic evolution with an early HP stage (granulitic or eclogitic) and a superimposed Barrovian type retrograde metamorphism (e.g. Belgodere (Corsica): Palagi et al., 1985; Solenzara-Fautéa (Corsica): Libourel & Vielzeuf, 1988; Punta de li Tulchi (Sardinia): Miller et al., 1976; M.giu Nieddu (Sardinia): Ghezzo et al., 1980).

Compositional, textural and geochemical evidence, however, points to a different origin of these basic rocks. The ultramafic rocks containing granulitic relics (eg.. M.giu Nieddu, NE Sardinia) are regarded as fragments of layered basic bodies that intruded the lower crust (Ghezzo et al., 1980), whereas other amphibolites were probably derived from oceanic or continental tholeiites (Ricci & Sabatini, 1978). Sometimes these amphibolites are closely associated with acidic ortho-derived and, according to Arthaud & Matte (1977a) and Ricci & Sabatini (1978), they could represent an Early Paleozoic continental rifting phase.

We interprete the HGMC as a crustal nappe that was overthrust on the medium grade metamorphics of Posada-Asinara Line Belt during the intracontinental shortening.

3.2 *The "Posada-Asinara Line" Belt*

All of the internalmost zone of the "South European Hercynian Chain" is characterized by high grade polymetamorphic Precambrian basement with eclogitic and granulitic relics ("Innermost crystalline nappes": Matte, 1983; 1986) which has been overthrust on lower grade metamorphic units (Burg & Matte, 1978). At the base of the overlying plate, squeezed ophiolitic remnants often crop out (e.g. North Portugal - Galician nappes: Ribeiro et al., 1964; Bayer & Matte, 1979; Iglesias et al., 1983).

In Sardinia the HGMC and LMGMC are separated by the Posada-Asinara Line Belt, a belt of amphibolite facies micaschist, paragneiss and quartzites that crops out at Asinara Island, in southern Gallura and at Posada Valley (Fig. 3) and reach a thickness of up to several kilometres. The original relationships between the two complexes are sometimes obscured by late Hercynian uplift-related deformation. Posada-Asinara Line Belt is characterized by the widespread presence of small bodies of amphibolites which at Posada Valley (Elter, 1987; Elter et al., 1990) show an important retrograde greenschist facies metamorphism related to late- to post-Hercynian dextral transcurrent shearing. Amphibolite bodies contain granulite facies relics (salitic pyroxene + grossular-rich almandine) (Ghezzo et al., 1982) and eclogite facies relics (Cappelli et al., 1991) as well as relics of mylonitic textures that developed under high grade metamorphic conditions (Carosi & Elter, 1989). In terms of major element chemistry, all amphibolites show compositions strongly resembling those reported for evolved oceanic tholeiites and concentrations of high field strength elements (HFSE) also indicate a MORB origin of the amphibolite protolith. The observed REE patterns are interpreted as the result of partial melting of a depleted MORB-like source followed by fractionation. Sm/Nd whole rock isochron yields a protolith age of about 950 Ma and an initial $^{143}Nd/^{144}Nd$ ratio of 0.51167 (Cappelli et al., 1991). The obtained epsilon value of the initial ratio (5.3) closely maches that expected for a depleted mantle source, yet at the same time (5.8), and supports the original oceanic signature of the Posada-Asinara Line amphibolites.

A strong magnetic anomaly (Cassano et al., 1979) along the Posada-Asinara Line is attributable to important mafic bodies at depth and suggests that the lineament can be considered a crustal scale structure (Vai, 1988 and Vai, this volume).

The Posada-Asinara Line can therefore be considered a segment of the "South European Hercynian Suture Zone".

3.3 *The Low to Medium Grade Metamorphic Complex (LMGMC)*

Cambrian to Early Carboniferous sedimentary cover rocks of the underlying continental margin are stacked in a pile of nappes (Fig. 4E and 7) that crops out between the "Posada-Asinara Line" and the "External Zone" of the chain (SW Sardinia), where slightly deformed anchimetamorphic autochthonous rocks crop out (Arthaud, 1963; 1970; Poll, 1966; Poll & Zwart, 1964; Palmerini et al., 1980).

On the basis of stratigraphic and structural character this allochthonous complex can be divided into "Internal Nappes" and "External Nappes" (Fig. 2 and 7).

3.3.1 *The "Internal Nappes"*

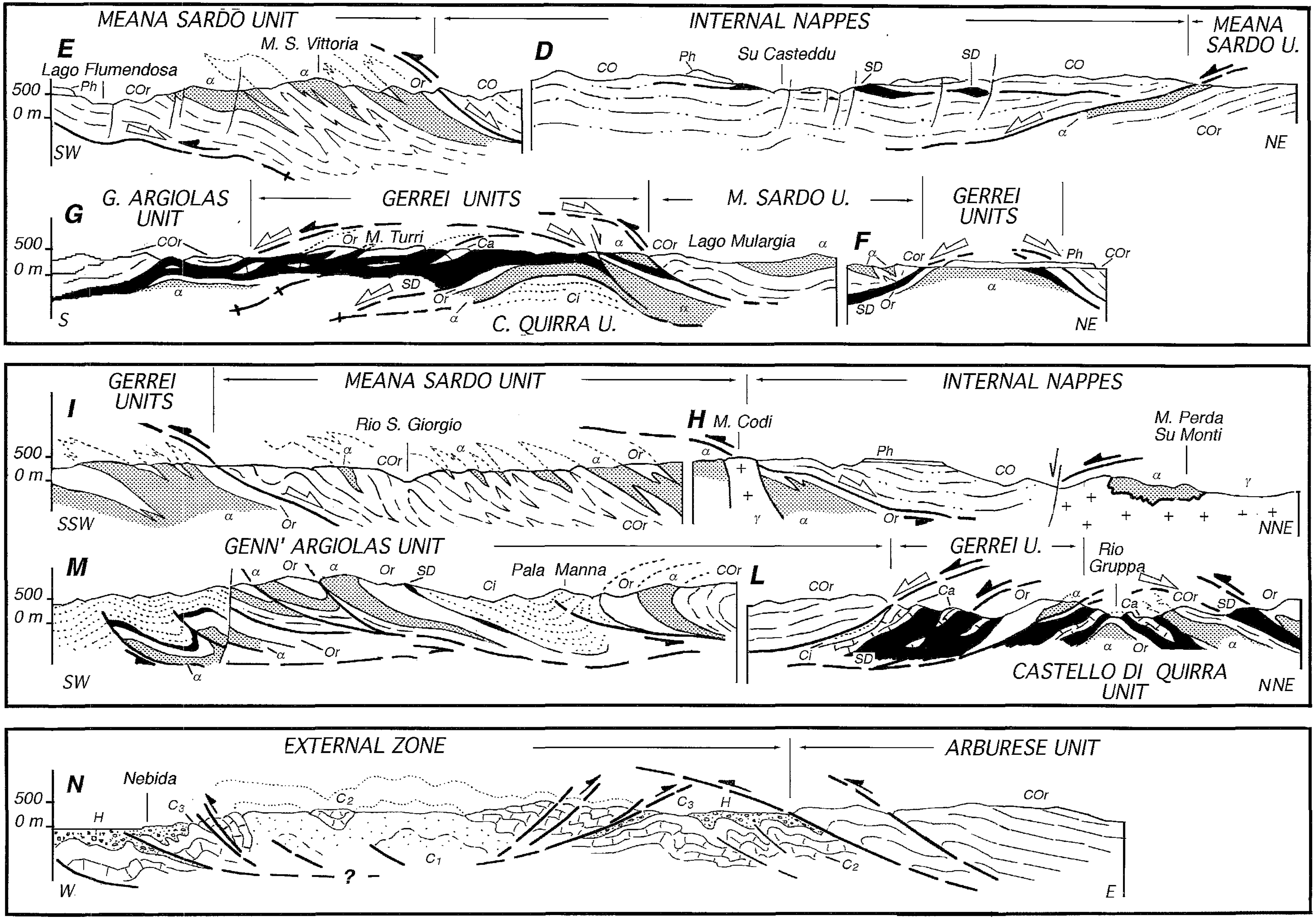
The "Internal Nappes" crop out between the "Posada-Asinara Line" and the overthrusted External Nappes. In these areas, as a result of the higher metamorphic grade and lack of geochronologic data the age of the protoliths is still uncertain.

Probably all the metamorphic rocks in the "Internal Nappes" were derived from a Paleozoic succession equivalent to the one that crops out in Central-Southern Sardinia. The differences between these two successions, including the scarceness of Ordovician metavolcanics and Silurian - Devonian limestones in the "Internal Nappes", are caused by the different locations of the Paleozoic basins on the Gondwanian continental margin.

The deformation occurred under generally greenschist facies conditions, that increased to amphibolite facies not far from the "Posada-Asinara Line".

Structures within the Internal Nappes include a well developed synmetamorphic foliation, strong stretching lineations, ductile thrusts and isoclinal folds.

Major recumbent folds in the nappes trend parallel to the chain and are described by Carmignani et al. (1980b) in northwestern Sardinia. Nappes without major folds are developed in the monotonous succession of Mount Gennargentu and in the more internal part below the HGMC, in which progressive deformation has created a composite schistosity (sensu Williams & Compagnoni, 1983 and Tobisch & Paterson, 1988) and complex fold interference


MEANA SARDO UNIT
INTERNAL NAPPES
MEANA SARDO U.
E
M. S. Vittoria
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GERREI UNITS
M. SARDO U.
GERREI UNITS
G
M. Turri
Lago Mulargia
F
C. QUIRRA U.
S
NE
GERREI UNITS
MEANA SARDO UNIT
INTERNAL NAPPES
I
Rio S. Giorgio
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M. Codi
M. Perda Su Monti
SSW
NNE
GENN' ARGIOLAS UNIT
GERREI U.
M
Pala Manna
L
Rio Gruppa
CASTELLO DI QUIRRA UNIT
SW
NNE
EXTERNAL ZONE
ARBURESE UNIT
N
Nebida
W
E

pattern (Carosi & Pertusati, 1990; Carosi et al., 1991).

In these areas (Gennargentu, Baronie, Goceano, etc.) stretching lineations are generally parallel to the chain and minor coeval folds trend normal to the chain. The fold axes are usually obliquely cut by their schistosity ("Transected Fold"). This type of fold (Borradaile, 1978; Treagus & Treagus, 1981) has been observed in the Appalachian/Caledonian chain where it has been used as evidence for major strike-slip motion that accompanied regional shortening (Stringer, 1975; Sanderson et al., 1980; Stringer & Treagus, 1980; Blewett & Pickering, 1988; Soper & Hutton, 1984; Woodcock et al., 1988; Lafrance et al., 1989). Analogously, these structures probably represent a transpressional component of the Hercynian intracontinental deformation in Sardinia. Transcurrent tectonics are well known throughout the internal zone of the Ibero-Armorican arc (Matte & Ribeiro, 1975; Brun & Burg, 1982) and have been explained as the result of a rigid-plastic indentation with lateral extrusion of wedge-shaped lithospheric blocks (Tapponier et al., 1982; Matte, 1986; Tapponier & Molnar, 1977).

Important dextral transcurrent movements along the Posada-Asinara Line (Elter, 1985; 1987; Elter et al., 1990) have also been related to this intracontinental evolution (Elter & Sarria,1989). An oblique convergence between the Armorica and Gondwana plates could also explain the shortage of oceanic crust reworked in the orogen.

3.3.2 *The "External Nappes"*

The "External Nappes" crop out between the Iglesiente and Barbagia regions (Fig. 2) and are composed of Cambrian to Lower Carboniferous metasedimentary cover. They represent the more external part of the allochthon emplaced in the fore deep basin of the chain (Fig. 4E and 7) during Namurian-Westfalian time. The nappes are detached along less competent stratigraphic horizons. The main décollement surfaces are located at three levels: at the bottom and at the top of the Middle Ordovician metavolcanic complex and at the bottom of Cambrian metasandstones.

Nappe movement was from NE towards SW, normal to the chain trend.

The synkinematic metamorphism ranges from very low- to low-grade (Chlorite Zone). Metamorphism and deformation increase slightly from the upper units to the lower ones.

Higher grade metamorphic rocks (from upper greenschist facies to amphibolite facies) crop out only in core complexes linked to post-collisional extensional tectonics (Castello Medusa, Castello di Quirra and M. Grighini-Mt. Trempu Units).

In these deepest metamorphic core the more evident surface is a crenulation cleavage (S2) sub-horizontal and paraconcordant with the main overthrusts. This cleavage is axial plane of shear folds that deforms a previous metamorphic layering (S1). P. Pertusati belives that this S2 cleavage is related to the final evolution of the first deformation phase and he attributes to this deformation the thrusts in shallow structural level as well as the important internal deformation in deeper units.

The dippest units are overlain by two allochthonous complexes:

a) the Gerrei Units, b) and a very large alloctonous complex called Meana Sardo Unit, Genn'Argiolas Unit and Arburese Unit in the Sarcidano, Sarrabus and Arburese regions respectively (Fig. 7). The Gerrei Units are detached at the base of the Ordovician metavolcanics which are mainly composed of very competent metarhyolites ("Porfiroidi"). This complex is divided into many minor units that are stacked in duplexes at the base of the overlying allochthonous complex.

The emplacement of the Gerrei Units is associated with a well developed cleavage that developed under low grade metamorphic conditions (Chlorite Zone).

The Meana Sardo - Genn'Argiolas - Arburese Units (Fig. 7) constitute the main allochthonous complex of the External Nappes. They are detached at the base of Cambrian metasandstones and have overridden the Gerrei Units to constitute the frontal allochtonous part of the chain. The complex was affected by synkinematic metamorphism that varies from low- to very low-grade.

The end of the thrusting is recorded by the Visean-Namurian flysch (Spalletta & Vai, 1982; Maxia, 1984; Barca, 1991; Barca & Spalletta, 1985; Barca & Olivieri, this volume) which is deformed by the last nappe movements (Barca et al., this volume) (Fig. 7 M). This Culm-type terrigenous synorogenic deposit crops out mainly in front of the Hercynian nappes: in South Sarrabus and in the Iglesiente-Sulcis region. It contains olistostromes of portions of the Silurian sequence and olistoliths of the Ordovician-Silurian, and Devonian formations. This flysch also crops out between the Genn'Argiolas and Gerrei Units (Spalletta & Vai, 1982; Barca & Spalletta, 1985) (Fig. 7 L) and at the bottom of Gerrei Units (Maxia, 1984; Eltrudis, 1991) (Fig. 7 G). All the External Nappes probably rest on late-orogenic fore deep deposits.

Fig. 7 - Geological sections across the Central and SW Sardinia (traces in Fig. 2). Solid arrows indicate the early compressional event; open arrows indicate the late extensional event. Ph: Post-Hercynian sediments and volcanics. g: granitoids; Carboniferous.
Internal Nappes - CO: Metasandstones and phyllites ("Postgotlandiano" Auct.); ?Cambrian - ?Ordovician.
External Nappes - Ci: Culm syntectonic deposits; Early Carboniferous. Ca: Marbles and metalimestones; Devonian - Early Carboniferous. SD: Shales, black shales and cherts; Silurian - Devonian. Or: Metarkoses, metasandstones and metalimestones with intercalations of basic metavolcanics; Late Ordovician. α: Metavolcanics and metamorphosed reworked volcanics; Middle Ordovician. COr: Metasandstones and shales; Middle-Upper Cambrian to Early Ordovician.
External Zone - H: Metaconglomerates, metasandstones and slates; Ordovician-Silurian. C_3: Slates and metasandstones; Middle Cambrian-Early Ordovician. C_2: Limestones and dolostones; Early Cambrian. C_1: Metasandstones and shales; Early Cambrian.

3.4 *The External Zone*

The "External Zone", cropping out in the Iglesiente-Sulcis region, is a classic fold-thrust belt characterized by medium to steeply dipping thrusts, fold axial planes and cleavage and very low- grade Hercynian metamorphism. The main structure is due to interference from Middle Ordovician E-W trending folds ("Sardic Phase") and N-S trending folds of the main Hercynian phase (Poll & Zwart, 1964; Poll, 1966; Brusca & Dessau, 1968; Arthaud, 1963; 1970). décollement probably exists at the base of the folded Cambrian sequences, along which all the thrusts of external zone might be linked (Fig. 7 N) (Carosi et al. this volume). Similar structures have been shown by seismic profiles of the frontal regions of other segment of the Hercynian chain (Cazes et al., 1986).

4. Post-Collisional Evolution

An important extensional event must have developed also in the Sardinian Variscides as response to gravitational re-equilibration within the collisional structure (Fig. 4 F).

Although available data on post collisional tectonics are few compared to those on the pre- and syn-collisional evolution, our recent and published data support the occurrence of inversion tectonics, from compression to extension. Extension is characterised at different structural levels by thrust reactivation by low angle normal faults, LP-HT metamorphic overprint, coeval synkinematic intrusions and by molasse basins.

Evidence of Late Hercynian extension at lower structural levels is confined to the axial zone within wide basement pendants resting on the northern portion of the Sardic batholith and at Asinara Island. In the Anglona region a late LP-HT (andalusite-sillimanite facies) metamorphic assemblage overprints a previous Barrovian amphibolite facies assemblage (Bt + Grt $\pm$ Ky $\pm$ Sta). The former assemblage developed during late deformations (Di Pisa & Oggiano, 1987b) that are characterized by normal ductile shear belts and extensional strain-slip cleavages that often evolve into a composite cleavage. In the migmatitic complex the post-collisional evolution leads to widespread anatexis (essentially through dehydration melting reactions in an active decompressional regime) and the emplacement of several small bodies of syntectonic parautochthonous peraluminous granites (Di Pisa & Oggiano, 1985; 1987a; Oggiano e Di Pisa, 1988; Macera et al., 1989) that became strongly foliated under high temperature solid state conditions (Paterson et al.,1989).

Typical extensional structures are "metamorphic core complexes" which domes are of strongly deformed metamorphics rocks (lower plate) in sharp tectonic contact, across normal mylonitic shear zones, with less deformed sedimentary or weakly metamorphosed covers (upper plate). During core complex formation, mid crustal metamorphic rocks were exhumed at Capo Spartivento in the southwesternmost part of the Island, and at Mt. Grighini and Mt. Trempu in the Nappe Zone, in the major tectonic culminations of the Flumendosa Antiform.

At Capo Spartivento (fig. 2), in the core of a roughly N-S trending antiform, in the External Zone, Middle Ordovician orthogneiss (Mt. Filau Orthogneiss; different radiometric methods; Cocozza et al., 1977; Scharbert, 1978; Delaperriere & Lancelot, 1989) which shows medium to high grade Hercynian metamorphism And $\pm$ Sil $\pm$ Grt (Mazzoli & Visonà, 1991), crops out. This gneissic nucleus is capped by a mylonitic shear zone which, in turn, underlies a weakly metamorphosed Lower Cambrian, Upper Pre-Cambrian pelitic-arenaceous sequence with minor carbonate horizons (Bithia Formation).

Throughout the Nappe Zone compressional-related structures are overprinted by extensional "post-nappe deformations". Minor drag folds with opposite facing with respect to the culmination of regional domes have been reported by Dessau et al. (1983) for the Gennargentu antiform.

More recently, a widespread extensional phase has been discovered in the Flumendosa Antiform (Cappelli, 1989), and resulted in different deformational styles and different structures at different depths. At shallow structural levels extensional kinematics are largely controlled by the geometry of the collisional antiformal stack: an important inversion takes place in which thrust planes were reactivated as low angle detachments whose sense of movement was away from the culmination of the previous structures (i.e. northeastward and southwestward). The main regional thrust surface in particular (Villasalto Thrust) suffered such reactivation by normal faults with a strike-slip component. The centrifugal movement away from the NW trending crest of the Flumendosa Antiform led to unroofing along the crest and rapid uplift of mid-crustal levels causing their juxtaposition with lower grade metamorphics along mylonitic shear zones. This was accompained by thermal input favored by a combination of pressure reduction and synchronous intrusion of magmatic bodies (Cappelli, 1991). Mid-crustal level rocks crop out both in the Mt. Trempu and Mt. Grighini tectonic windows that are alligned along the axial culmination of the Flumendosa Antiform.

In detail, Mt. Grighini (fig. 2) consists of a crystalline core composed of amphibolite facies micaschists and gneisses that were intruded by Hercynian granitic bodies which show an S-C mylonitic texture that is particularly well developed in leucogranites. This crystalline core is overlain by Paleozoic cover (Gerrei Units) rocks, that suffered greenschist facies metamorphism (Carmignani et al., 1985; Cherchi & Musumeci, 1987). The contact between the core and Gerrei Units in the southwestern part of the window is a NW-trending, steeply SW dipping dextral strike-slip shear zone (Elter et al., 1990; Musumeci, 1991). In northeastern zone this contact is a normal ductile shear zone. At Mt. Trempu, along the SE axial prolongation of the Mt. Grighini structure, And $\pm$ Sil $\pm$ Crd $\pm$ Pl bearing paragneisses and micaschists crop out, and are intruded by sheet like granitic bodies that show prothomylonitic to S-C mylonitic fabrics. The planar fabrics dip consistently northeastward and might represent the prolongation of the northeastern flank of the Mt. Grighini structure. Based on these features Mt. Trempu and Mt. Grighini can be considered a core complex exhumed through a normal slip shear zones on the NE boundary and a dextral strike-slip shear zone with normal component on the SW boundary.

These outcrops, together with Capo Spartivento, correspond to the deepest crustal levels seen in the External Nappe and External Zone of the chain. In these areas they are the only outcrops showing medium grade metamorphism and clearly disagre with the Hercynian collisional metamorphic zoneography.

The chronological development of the post-collisional extension in Sardinia can be documented by sedimentary, paleontological and radiometric data. The extensional evolution is constrained to a time interval extending from the end of collision to the emplacement of the widespread calcalkaline plutonism of the Sardinian-Corsic batholith and to the development of the largely coeval Stephano-Autunian basins.

In the External Nappes the end of Hercynian shortening is signaled by a wild flysch of Visean-(?)Namurian age. It consists of off-shore deposits with large olistoliths, that crop out extensively at the nappe fronts and that suffered the deformation related to the tail end of thrusting (Fig. 7 M) (Barca et al. this volume).

The widespread extension postdates the 344 and 350 Ma collision-related ages as inferred by the age of the LP-HT metamorphism in the Internal Nappes as well as by the ages of some synkinematic granites emplaced into the HGMC.

In the Internal Nappes, syn-extensional LP-HT metamorphism which overprints the Barrovian phase (Di Pisa & Oggiano 1987b) yields a muscovite Rb/Sr blocking age of 303 $\pm$ 6 Ma (Del Moro et al., 1991).

In the HGMC, as well as throughout the Sardinian basement, synkinematic granites affected by non-coaxial deformation, crop out. The cooling age of magmatic muscovite from several syntectonic anatectic granitic bodies that crop out in the Bassa Gallura region, are between 308$\pm$9 Ma and 298$\pm$9 Ma and the emplacement age of a comagmatic suite among them is 300$\pm$7 Ma (Rb/Sr w.r. isochron, initial $^{87}Sr/^{86}Sr$: O, 7134, mswd 0.19; Macera et al., 1989).

At the Mt. Grighini core complex, the ages (Rb/Sr method) of some synkinematic granitoids are also close to the Westphalian-Stephanian boundary (303 $\pm$ 18 and 305 $\pm$ 16 Ma: Carmignani et al., 1985; Musumeci, 1991). Finally at the Capo Spartivento core complex the available data on synkinematic granites points to a similar age (299$\pm$21 Ma w.r. Rb/Sr isochron; Scharbert, 1978; Beccaluva et al., 1985).All the synkinematic intrusions suffered a brittle-ductile deformation that overprints a magmatic foliation. Hence, the Westphalian-Stephanian boundary age (303 Ma: Harland et al., 1989), of the synkinematic granitoids represents a step in the extensional evolution rather than its beginning or end.

The overlap between above reported ages and the ages of the so called "post-Hercynian basins" is noteworthy. Remnants of these basins crop out everywhere: in the external zone a ?Westphalian D-Stephanian (S. Giorgio basin; Fondi, 1980) or Stephanian-Autunian (Guardia Pisano; Barca et al., 1991) age is attributed to these basins , in the Nappe Zone a Stephanian age is proposed for the Seui antracitiferous basin (Deplano, 1985). The Lower Permian ages proposed for other basins in the Nappe Zone (Perdasdefogu, Mulargia, Nurra; Pecorini, 1974) could be schifted down to the late Carboniferous on the basis of paleomagnetic evidence in the calcalkaline volcanics associated with the clastic infillings of some basins (Edel et al., 1981) as well as on paleofloristic considerations (Filigheddu, 1985).

Such an age is also likely for the continental clastic deposits that overlie the HGMC in Southern Gallura. In fact an age of 284$\pm$16 Ma (K/Ar "total isochron"; Cozzupoli et al., 1984) is the most reliable for the rhyolitic volcanics of the Baronie region .

The development of Permo-Carboniferous basins that are in part, contemporaneous with the ductile extension at lower crustal levels is typical of other segments of the Hercynian Chain that underwent crustal thinning, ductile extention, heating, unroofing and exhumation of deeper crustal levels (Echtler & Malavieille, 1990; Malavieille et al., 1990). The extensional evolution inferable from the available data on Sardinian Variscides is similar to a model proposed also for other chains (Malavieille 1987; Polino et al., 1990), in which the collapse of the previously thickened crust follows the first underplating related extension (Platt, 1986) and it is probably coeval with trascurrent tectonic transecting the Hercinian orogen during the latest Carboniferous and Permian (Arthaud & Matte, 1977b; Ziegler, 1982; 1884; Vai, 1991).

In Sardinia the extension could also have started in the internal zones of the Chain by underplating and have propagated throughout the basement through conjugate ductile normal shear zones that grade into normal faults at shallow crustal levels, similar to that reported in the Basin and Range province (Kligfield et al., 1984; Hamilton, 1987 and references).

In summary the regions from which the normal shear zones diverged underwent tectonic denudation, unloading and offsetting with respect to the regions from which the normal shear zones converged; the latter, relatively subsiding, experienced clastic sedimentation and volcanism of Upper Carboniferous-Permian age.

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TERTIARY TRANSPRESSIONAL TECTONICS IN NE SARDINIA, ITALY

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Abstract

Tertiary deformation in NE Sardinia is mainly represented by two left-lateral strike-slip fault systems: the most important one strikes NE (Nuoro, Tavolara Island Faults etc.), the other strikes EW (Cedrino and Posada Faults). The sinistral "Nuoro Fault" along the Mount Albo massif shows an important compressional component.

The transpressional event is characterized by NW-striking overthrusts that either place basement rocks on cover rocks or affect only the Jurassic-Cretaceous carbonate cover.

Transpression resulted from convergence between the Nuoro and Cedrino left-lateral stike-slip faults which bound a compressed uplifted wedge. Mount Albo thrusts resulted from wedge movement towards the west with respect to the basement of the Cedrino Fault.

A similar kinematic scenario is also possible for the wedge existing between Posada and Tavolara Faults.

The most important strike-slip system (NE-striking) shows a shortening direction of about NNE. The EW-striking strike-slip faults are coeval and inherited their orientation from the Hercynian basement structure.

Key words: *Sardinia, Tertiary tectonics, Strike-slip, Thrusts.*

1. Introduction

Mount Albo is a 30 km long, NE-trending Meso-Cenozoic carbonate massif in northeastern Sardinia (Fig. 1). Overthrusts, recognized for some time in this zone, affect the metamorphic basement rocks and their Mesozoic carbonate cover.

On the basis of data collected during geological surveys of this area (Funedda, 1990; Pasci, 1990; Disperati, 1992;), we suggest a new structural interpretation of the M ount Albo massif.

1.1 *Previous Work*

The first important geological study in the Mount Albo area concerned the Mesozoic of eastern Sardinia (Vardabasso, 1948) and consisted of a series of geological sections in which Mount Albo was considered an open syncline cut by important subvertical faults along the southeastern flank of the structure. These faults, in some places (Dispensa Gulletti), thrust Paleozoic metamorphic rocks onto Mesozoic limestones.

Later, Mount Albo was interpreted as a SE-dipping monoclinal structure made up of Dogger dolomites and sandstones, transgressive on the Hercynian basement and Malm limestones (Calvino et al., 1967, sheet 195, Geological Map of Italy at 1:100.000). In southern Mount Albo, the same Dogger dolomitic rocks tectonically overlie Malm

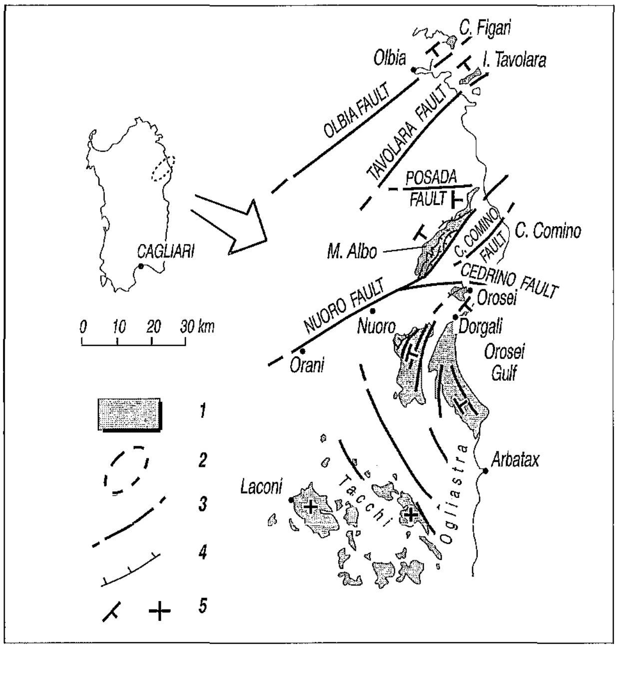


Fig. 1 - Mesozoic outcrops and main "alpine" structures in eastern Sardinia; 1) Mesozoic outcrops; 2) studied area; 3) faults; 4) overthrusts; 5) bedding strike and dip; horizontal beds.

limestones; this contact was previously considered stratigraphic by Calvino et al. (1967) who had assumed a facies heteropy between both formations.

Dieni & Massari (1970) pointed out that the upper dolomites and limestones are lithostratigraphically related to those below, refusing the facies heteropy hypothesis and showing that the upper part of the Mesozoic Mount Albo section would have been laterally transported. The authors proposed gravity sliding of blocks as mechanism of emplacement which involved sliding of an allochthonous complex off of the incompetent phyllitic basement onto the autochthon during the Oligocene. The source of the allochthon had been a high to the northwest of Mount Albo, now completely eroded.

Chabrier (1970), on the contrary, considered Mount Albo as a series of monoclinal tectonic slices of basement with their sedimentary cover overthrust toward the NW. In this scenario he explained the doubling in the sedimentary sequence and thrusting of the basement on its cover. According to Chabrier the deformation was after Upper Cretaceous on the basis of the age of a polygenic breccia involved in the thrusting.

Alvarez & Cocozza (1974) re-proposed the gravity sliding mechanism of Dieni & Massari (1970) to explain Mount Albo structures, but they assumed the highs southeast of Mount Albo (M. 'e Senes area) as the source for the allochthon, in accordance with Chabrier (1970). Moreover Alvarez & Cocozza identified strike-slip tectonics: they asserted that faults bounding southeastern Mount Albo, originally normal faults, were reactivated as strike-slip faults, probably with left-lateral movement.

Recently Dieni et al. (1987) have proposed the possibility that Mount Albo is a "flower structure" connected with probable transpression along the strike-slip fault at the southeastern margin of the massif. This is the hypothesis we examine in this paper.

2. Stratigraphic Outlines of The Mount Albo

2.1. *Hercynian Basement*

Mount Albo area is included in the so called "Axial Zone" of the Hercynian belt of Sardinia (Carmignani et al., 1986). To the west of Mount Albo a sequence of low to medium grade schists crops out and metamorphism increases northward (Franceschelli et al., 1982).

The main structure is an EW-trending, E-plunging kilometric antiform (Calvino et al., 1967) with a core of granodioritic orthogneisses (cf. Fig. 2).

East of Mount Albo low grade metamorphic rocks (phyllites and micaschists), two mica granites (M. 'e Senes Granite), tonalites and tonalitic granodiorites and leucogranites (Capo Comino Leucogranites) crop out.

2. 2. *Jurassic-Cretaceous Sequence*

Mount Albo is made up of essentially a rather thick Jurassic dolomitic-calcareous sequence. Otherwise, Cretaceous and Tertiary rocks constitute only small outcrops, often with tectonic or uncertain relationships with Paleozoic and Jurassic lithologies.

Stratigraphic knowledge about the Mount Albo sedimentary sequence come mainly from Azema et al. (1977), Massari & Dieni (1983), Dieni & Massari (1985 a,b) and Dieni et al. (1987). We refer chiefly to these authors for stratigraphic data about the Mount Albo sequence.

— DORGALI DOLOSTONE (Bathonian-Late Kimmeridgian) ("G1" in Fig. 2): it is transgressive and unconformable on the Hercynian basement. At the bottom this formation consists of continental conglomerates and sandstones, often with iron deposits at the contact with the basement. Upward, brown massive or poorly stratified dolomites (beds 40-90 cm thick) make up most of the formation. In the northern Mount Albo region, at the top of this formation, lenses of oolitic limestones and dolomitic limestones are interbedded. Dieni & Massari (1970) suggest a 15-600 m thickness for this formation.

— S'ADDE LIMESTONE (Callovian-Late Kimmeridgian) and BARDIA LIMESTONE (Portlandian-Berriasian) ("G2" in Fig. 2): the first formation is heteropic with the Dorgali Dolostone (G1) and crops out in the northern area of Mount Albo, whereas to the west it is gradually replaced by the upper part of the Dorgali Dolostone.

It consists of fine-grained, brown, stratified limestones (beds over 50 cm thick), oncolitic limestones with Ammonites, packstones and fine-grained grainstones with micritic ooids, sometimes bearing chert nodules. This formation is about 0-120 m thick.

The second formation, lying on the S'Adde Limestone in the north and on the Dorgali Dolostone in the south, is comprised of massive or poorly stratified grainstones, rudstones and algal bindstones. Maximum thickness is greater than 600 m (Dieni & Massari, 1970).

— SCHIRIDDE' LIMESTONE and MARLS (Early Valanginian-basal Hauterivian) ("C1" in Fig. 2): these rocks are only known in little outcrops south of Siniscola and near M. Pizzinnu (Fig. 2). Marly limestones passing into regularly stratified cherty limestones (beds 20-40 cm thick) with interbedded marls .

The Schiriddè Limestone pass into grey marls and dark marls with pyrite, interbedded with dark grey marly limestones. According to Dieni et al. (1987) the Schiriddè Limestone is about 15 - 20 m thick an theMarl are 50 - 60 m thick, but the occurrence of the tectonic contacts precludes the evaluation of the true thickness.

2. 3. *Cuccuru 'e Flores Conglomerate (Post Early Lutetian) ("cg" in Fig. 2)*

This formation crops out discontinously along the southeastern border of the Mount Albo massif (Fig. 2). South of Siniscola it consists of polygenic breccias, conglomerates and sandstones with heterometric elements from metamorphic Paleozoic basement and sedimentary Jurassic-Paleocene cover (Dieni et al., 1987).

These deposits can be chaotic or regularly stratified with local clayey beds and unconformably cover the M. Bardia Limestone in the Dispensa Guletti zone(Fig. 2). North of Siniscola this formation lies unconformably on the Paleozoic basement. It is represented mainly by calcareous conglomerates and breccias, with ooids, bioclasts, chert and clasts from the metamorphic basement and sedimentary Jurassic-Cretaceous cover.

Stratified calcilutites with chert nodules (beds maximum 30 cm thick) are interbedded. In the M. Longu zone (Fig. 2) limestones and marly limestones crop out; they are in places stratified and contain heterometric metamorphic and sedimentary clasts and blocks.

Even though various authors have studied these deposits, their age and significance are still uncertain. Vardabasso (1948, 1959) and Calvino et al. (1958) mistook tectonic and stratigraphic relationships between the conglomerates and Jurassic limestones in the Dispensa Guletti zone and assigned the former to Malm- Cretaceous age range.

Dieni & Massari (1965) correlated the conglomerates with similar rocks from near Orosei (Fig. 1) and Lanaitto, which contain reworked Cuisian faunas.

On the basis of regional considerations pertaining to Sardinia, they related all these deposits to a Late Eocene or more probably Oligocene tectonic event.

Chabrier (1970) attributed the conglomerates cropping out north of Siniscola (with interbedded *Globotruncanas*-bearing cherty limestones) and near Dispensa Guletti to the Late Cretaceous.

In addition, Calvino et al. (1972) attributed marls and cherty limestones cropping out north of Siniscola to the Late Cretaceous. However, they considered these rocks as reworked sediments in the (? Oligocene) conglomerate.

Lastly, Dieni et al. (1987) have discovered Paleocene clasts in the conglomerates along the southeastern margin of Mount Albo, thereby excluding the possibility that this deposit has a Cretaceous age.

According to Dieni & Massari (1985c) and Busulini et al. (1987) the conglomerates (now called Cuccuru 'e Flores Conglomerate) have a post Early Lutetian age in the Orosei area.

Therefore, accepting the correlation of Dieni & Massari (1965), we assume the same age for the Mount Albo area conglomerates.

3. Alpine Tectonic Lineaments of Central-Eastern and Northern Sardinia

3.1 *Northeastern Domain*

In this domain, North of the Orosei Gulf, alpine tectonics is represented by two regional fault systems (Fig. 1):

— NE-trending system characterized, from north to south, by the Olbia Fault, the Tavolara Fault, the Nuoro Fault and the C. Comino Fault. The Olbia and Nuoro Faults reach from the Tyrrhenian coast to the Oligo-Miocene Sardinian basin. All alpine carbonate outcrops of this domain (Capo Figari, Tavolara Island and Mount Albo) are joined and aligned with these lineaments and their stratification dips SE.

— E-W system: characterized by the Posada Fault and Cedrino Fault. Along the latter, the M. Tuttavista, another alpine carbonate massif, crops out. This system inherited its orientation from the Hercynian basement structure such as can be seen by the E-W Hercynian blastomylonitic shear zone along the Posada Fault (Elter, 1985).

3.2 *Orosei Gulf Domain*

This domain is characterized by an arcuate fault system trending from NE in the north (M. Tuttavista) to NW in the south (Ogliastra region). Alpine carbonate massifs of Supramonte and Orosei Gulf show the same trend. In the Supramonte area the bedding is nearly horizontal in the western - southwestern zone, whereas it is strongly inclined in the eastern - northeastern one, near the principal faults.

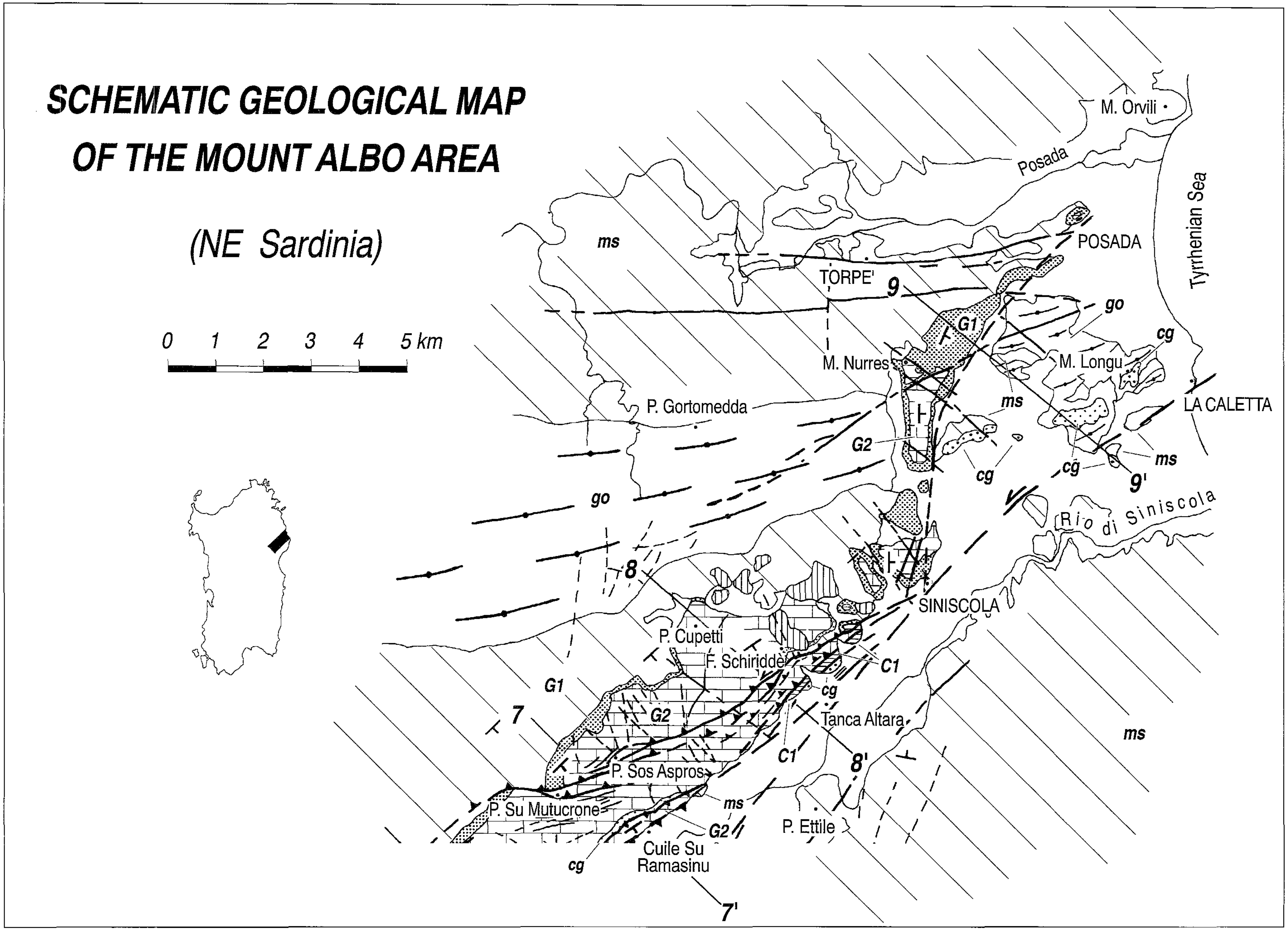
3. 3 *Southern Domain*

This domain includes Sarcidano, Southern Barbagia and Quirra regions. Here Mesozoic sequences unconformably cover the Paleozoic basement and are characterized by nearly horizontal carbonate outcrops ("Tacchi") affected only by N-S-trending normal faults.

4. Mount Albo Structure

4.1 *Nuoro Fault*

The Nuoro Fault extends for about 60 km from the Ottana - Orani zone in the southwest (Oligo - Miocene


SCHEMATIC GEOLOGICAL MAP
OF THE MOUNT ALBO AREA
(NE Sardinia)
0 1 2 3 4 5 km
M. Orvili
Posada
POSADA
Tyrrhenian Sea
TORPE'
M. Nurres
M. Longu
LA CALETTA
P. Gortomedda
Rio di Siniscola
SINISCOLA
P. Cupetti
F. Schiriddè
Tanca Altara
P. Sos Aspros
P. Su Mutucrone
Cuile Su
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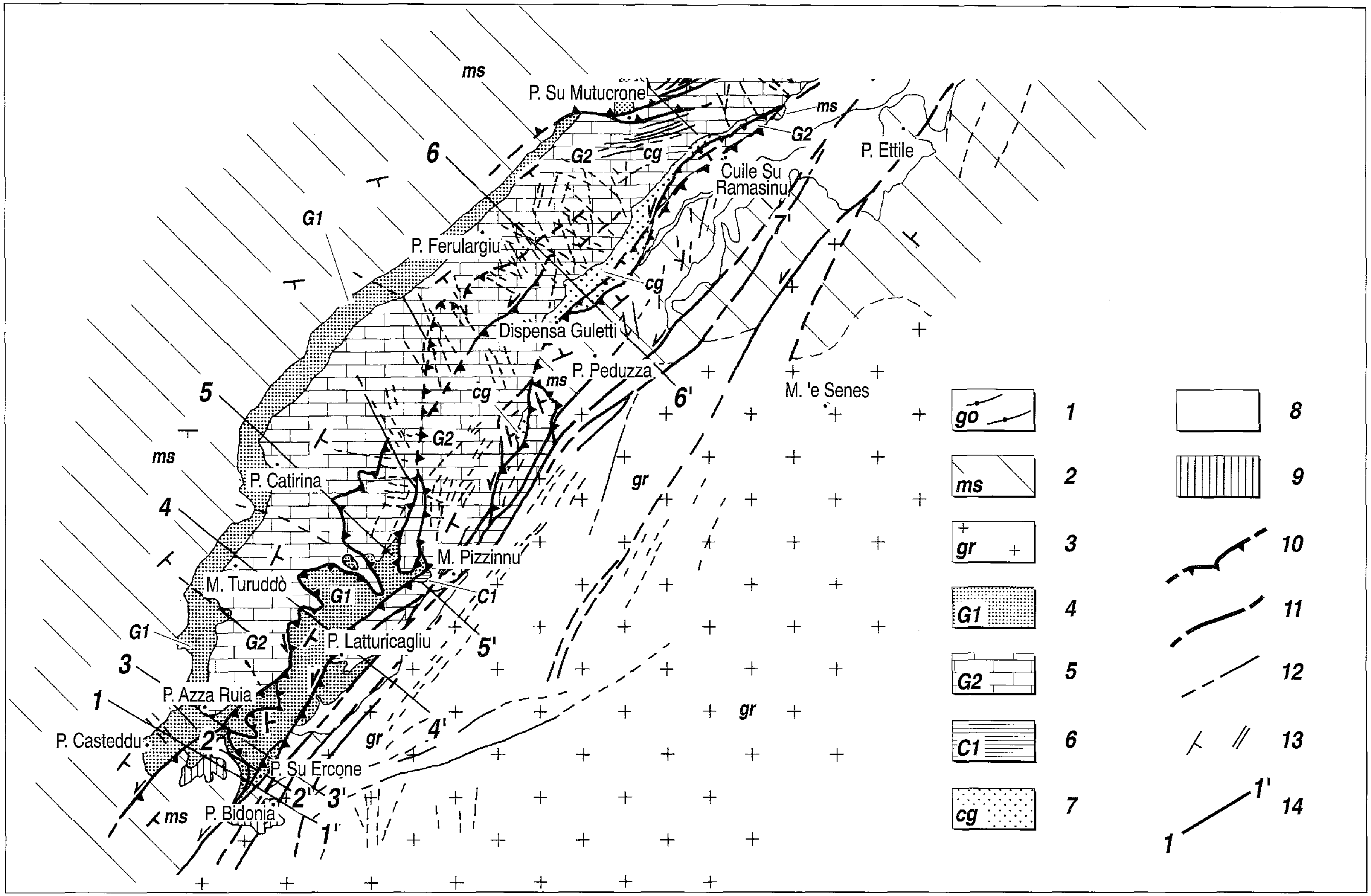


Fig. 2 - Schematic Geological map of the Mount Albo area; 1) granodioritic orthogneisses (go) (Paleozoic); 2) phyllites, micaschists, paragneisses and migmatites (ms) (Paleozoic); 3) M. 'e Senes Granite (gr) (Paleozoic); 4) Dorgali Dolostone (G1) (Bathonian-Late Kimmeridgian); 5) S'Adde Limestone (Callovian-Late Kimmeridgian) and M. Bardia Limestone (Portlandian-Berriasian) (G2); 6) Schiriddè Limestone and Marls (C1) (Early Valanginian-Earliest Hauterivian); 7) Cuccuru 'e Flores Conglomerate (cg) (Post Early Lutetian); 8) Recent detrital and alluvial deposits; 9) Landslides; 10) Strike-slip faults with reverse component; 11) Strike-slip and normal faults; 12) Minor faults and joints; 13) Bedding strike and dip (schistosity in the metamorphites); vertical beds; 14) Section traces.

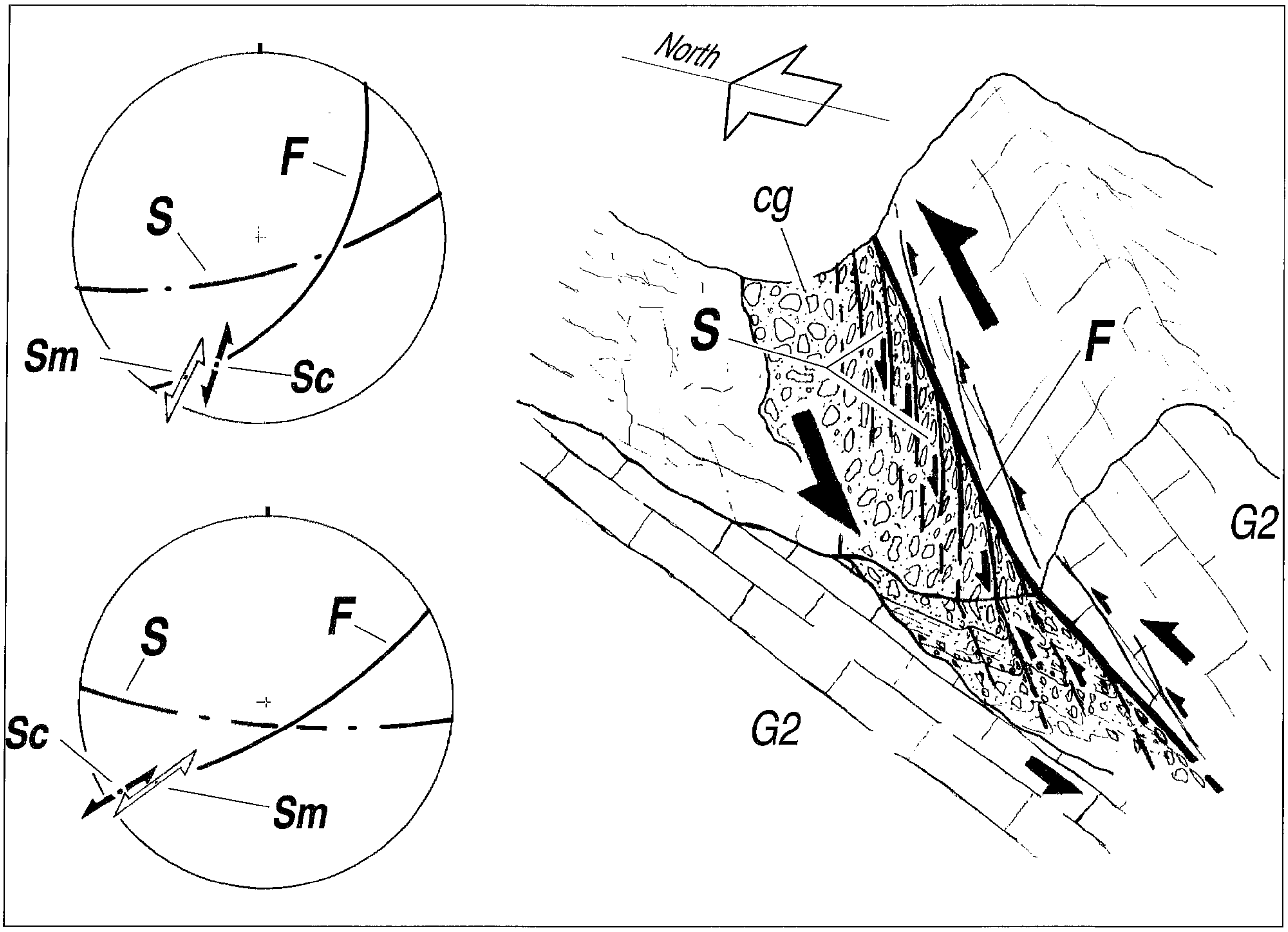


Fig. 3 - Synthetic faults near Cuile Su Ramasinu; F: fault plane; S: flattening plane; Sm: measured striae; Sc: calculated striae. The data indicate a left-lateral strike-slip movement with a minor reverse-slip component; (Wulff projection, lower hemisphere).

Sardinian basin) to the Siniscola alluvial plane in the northeast. Its direction varies from N60° in the southern sector to N40° in the northern one. It borders the entire southeastern margin of Mount Albo, where maximum deformation is recognized and constitutes an important regional lineament.

South of Mount Albo the Nuoro Fault affects only the Paleozoic basement. To the north, near M. Pizzinnu, this structure brings into contact the M. 'e Senes Granite and the M. Bardia Limestone. In this area Alvarez & Cocozza (1974) described the occurrence of limestones and shales in a tectonic slice lying subparallel to the Nuoro Fault. The limestones (partially recrystallized and with no definable age) and the shales (attributed to the Middle Jurassic on the basis of pollen and rare spores) are strongly sheared. The granites are cut by vertical or steeply SE-dipping cataclastic fault zones. These structures affect the granites for some tens of meters away from the tectonic contact and fault surfaces often show horizontal slickensides. On the basis of these features, Alvarez & Cocozza considered the Nuoro Fault a left-lateral strike-slip fault. We can add that the relationship between granites and limestones and the SE dip of the fault surfaces suggest a reverse separation (apparent relative movement) along the fault with the relative upward movement of the eastern area (granites) with respect to western one (Mount Albo).

Northward, from the P. Peduzza area, the Nuoro Fault probably splays out into branches which affect about a 2 Km-wide area.

Other important kinematic data concerning the Nuoro Fault have been collected near a quarry southeast of Siniscola: here the tectonic contact between the M. Bardia Limestone and the Schiriddè Limestone is characterized by a cataclasite with nearly vertical shear planes and structures that again reveal a left-lateral strike-slip movement with a smaller reverse-slip component.

4.2 *The Syntetic Uptrusts*

In the southern Mount Albo area, a set of tectonic slices rests on the Mesozoic sedimentary sequence, is composed of the same lithologies and is delimited by faults with generally reverse separation. These faults have an "en echelon" arrangement with respect to the Nuoro Fault (Fig. 2).

Important kinematic data about these synthetics are recognizable along the tectonic contact in the Dispensa Guletti-Cuile Su Ramasinu zone (Fig. 2). Here, in the western side, a polygenic Tertiary breccia (Cuccurru 'e Flores Conglomerate) stratigraphically overlies the M. Bardia Limestone, while in the east, it is tectonically covered by a tectonic slice of the same M. Bardia Limestone.

Along this tectonic contact, steeply dipping to the SE, the Tertiary breccia is deformed and sometimes crushed to a clayey cataclasite for a 1-1.5 m thickness. Also, limestones are cross cut by cataclastic surfaces that are nearly parallel to bedding. Kinematic indicators, such as the relationship between the main tectonic contact and the asymmetric shape of deformed clasts in the breccia, suggest a dominant left-lateral strike-slip movement with a minor reverse component (Fig. 3).

Fault planes with nearly horizontal slickensides confirm the predominance of the strike-slip component. On the basis of separation the fault appears to be a reverse fault, but kinematic indicators and actual relative movement (slip) show it is a left-lateral strike-slip fault with a component of reverse-slip.

Therefore, on the basis of the angular relationships of these faults with the main strike-slip fault (the Nuoro Fault) and their kinematic features, these faults are called "synthetics".

The geometry of the basement-cover contact in profile and the bedding orientation follow these fault trends (Fig. 5). They appear at depth near the Nuoro Fault as surfaces strongly dipping to the SE; at higher levels, generally not far from the basement-cover contact, their slope is shallower and they ultimately become subparallel with bedding in the carbonate formations: thus, these faults are upthrusts.

The essentially vertical throw affecting the metamorphic rocks changes to nearly horizontal slip in the higher, sedimentary levels. This geometry is accompanied by:

— strong shear lamination in the metamorphics and plutonics that probably represents "cataclastic flow" as described by Sylvester & Smith (1976) for basement crystalline rocks along the San Andreas Fault zone. This produces squeezing up of these rocks through subvertical tectonic surfaces.

— weak deformation in the stratified rocks; here, shear surfaces parallel or subparallel to bedding enable differential slip between various packages in the sedimentary cover (Fig. 5, sections 3-3', 5-5').

— development of minor tectonic slices which determine lithological repetition at the basament-cover contact; they can be observed near P. Su Ercone (Fig. 5, sections 1-1', 2-2').

Although basement and sedimentary cover are characterized by different deformation mechanisms, at the mesoscopic scale they behave as a unit and their contact is not a generalized surface of detachment. Locally, however, cover and basement elements are relatively detached (e. g. West of M. Pizzinnu). Here the basement-cover primary contact also represents a detachment surface (Fig. 5, section 5-5').

4.3 *The Schiriddè Fault*

The Schiriddè Fault is an important tectonic lineament at least 10 km long, stretching from the western border of Mount Albo and reaching the Siniscola area. west-northwest of the fault is an undeformed zone characterized by the sedimentary sequence dipping shallowly to the SE; est-southeast of the fault is a zone, cut by several faults linked to this main fault, where beds dip from 45° up to 90° and the Schiriddè Limestone is locally overtuned (Fig. 5, section 8-8'). Northwest of Tanca Altara the fault strongly dips to NW and tends to approach and parallel the Nuoro Fault. Even though in the Fontana Schiriddè Paleozoic micaschists lie on the M. Bardia Limestone, kinematic indicators (cataclasites, shear surfaces and slickensides) suggest that the Schiriddè Fault is here a left-lateral strike-slip fault with a minor component of reverse-slip. Westward, in the P. Su Mutucrone-P. Sos Aspros zone, its trend changes to N70-90°, the dip shallows and the fault becomes mainly a reverse fault.

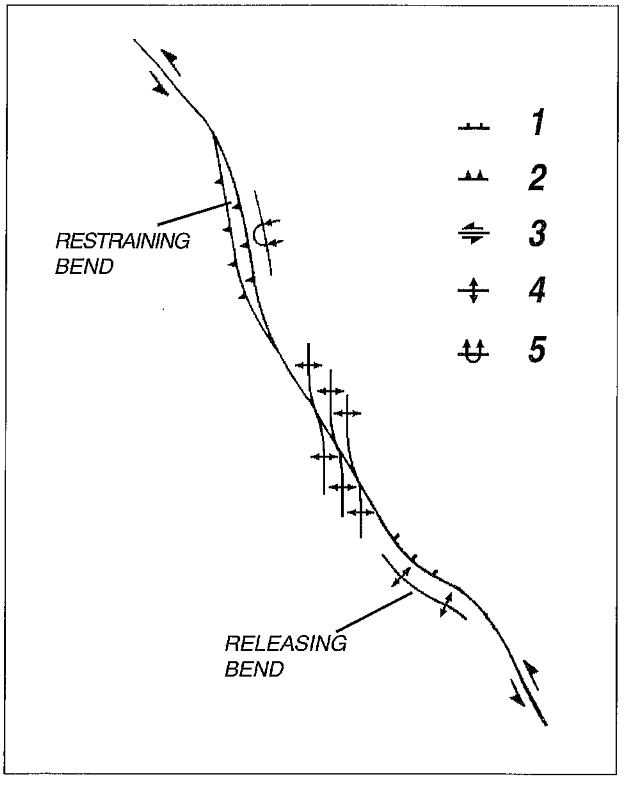
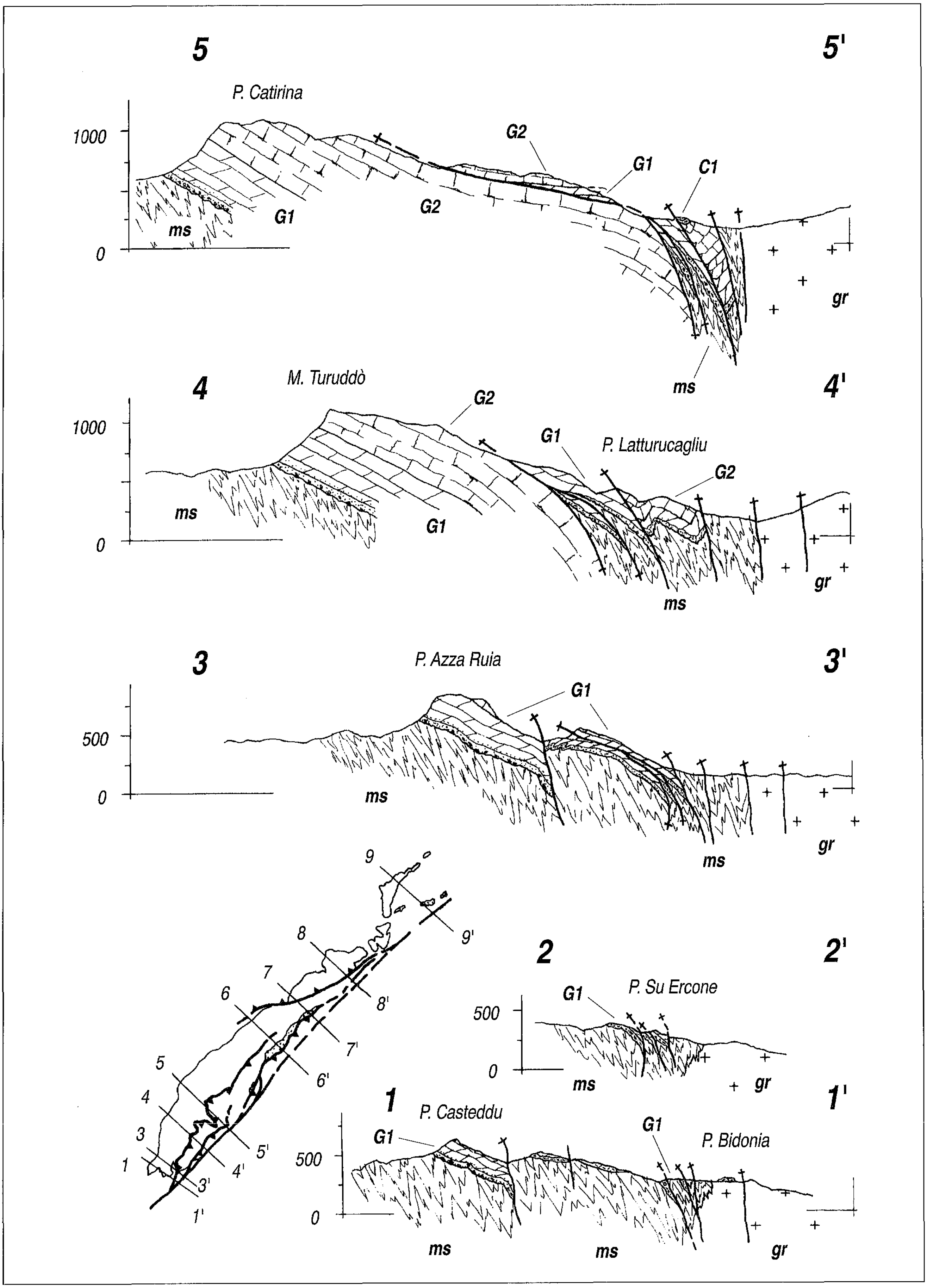


Fig. 4 - Structures developed along an ideal left-lateral strike-slip fault: 1) normal-separation fault; 2) reverse-separation fault; 3) fault with sense of strike-slip; 4) fold; 5) overtuned fold (after Sylvester, 1988). The bending of the Schiriddè Fault in the P. Su Mutucrone area (Fig. 2), can be ascribed to the so called "restraining bend" of this picture.


5
5'
P. Catirina
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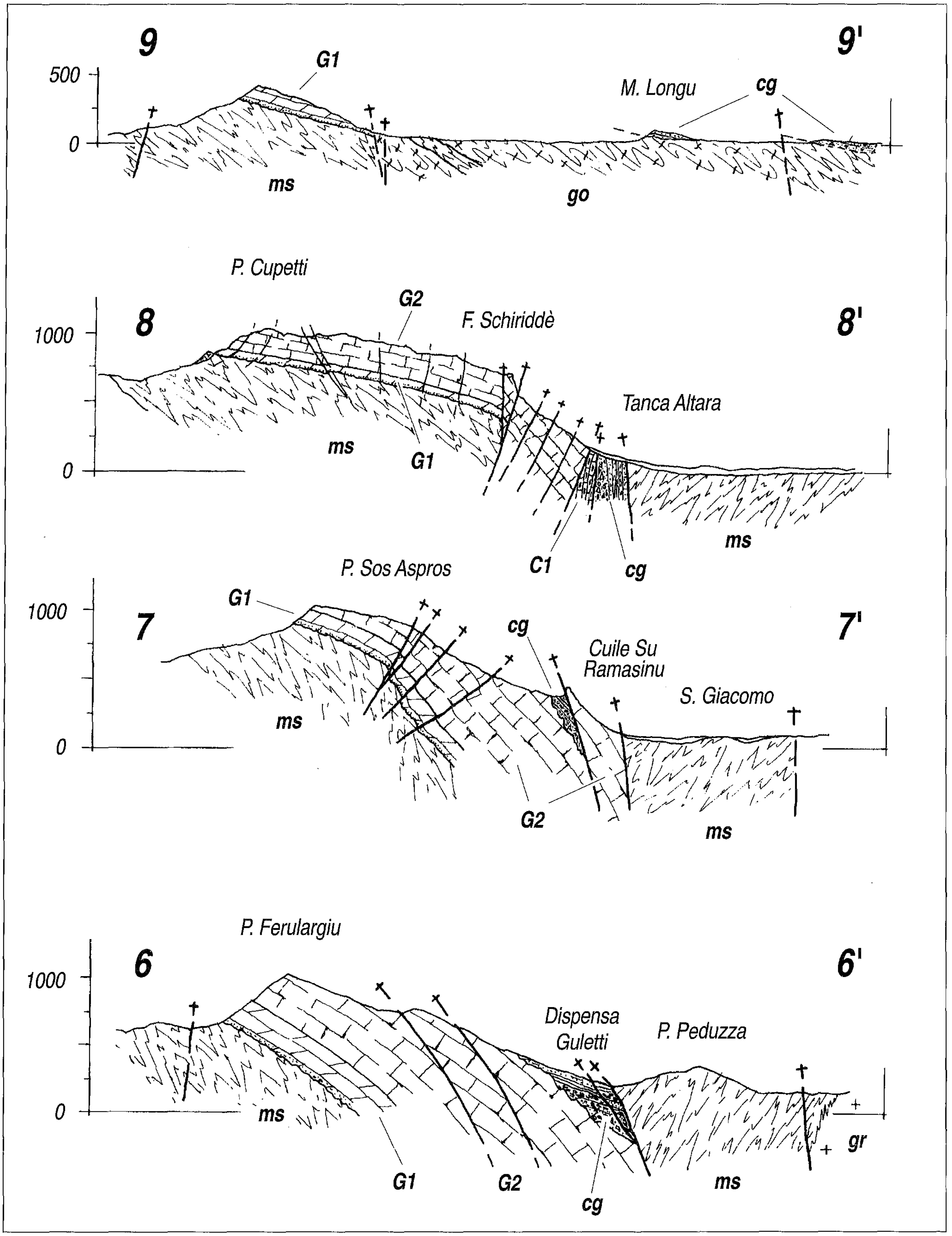

Fig. 5 - Geological sections; go: granodioritic orthogneisses (Paleozoic); ms: phyllites, micaschists, paragneisses and migmatites (Paleozoic); gr: M. 'e Senes Granite (Paleozoic); G1: Dorgali Dolostone (Bathonian-Late Kimmeridgian); G2: S'Adde Limestone (Callovian-Late Kimmeridgian) and M. Bardia Limestone (Portlandian-Berriasian); C1: Schiriddè Limestone and Marls (Early Valanginian-Earliest Hauterivian); cg: Cuccuru 'e Flores Conglomerate (Post Early Lutetian); section traces also in Fig. 2.

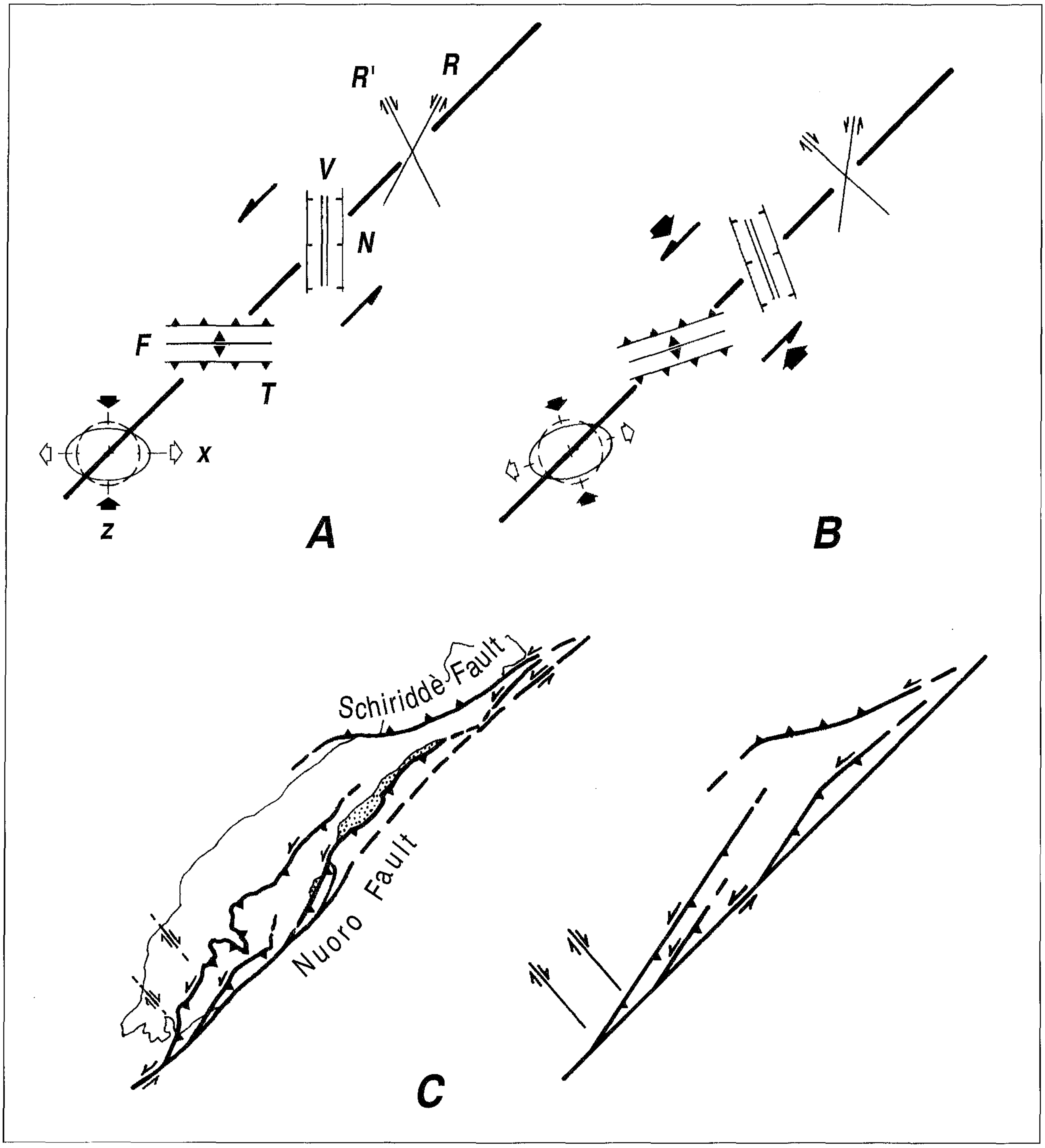

Fig. 6 - Tectonic synoptic sketch of the Mount Albo area; A) strain ellipse and main deformation features in a left-lateral strike-slip system; B) change in orientation of structures in a transpressional system; (after Sanderson & Marchini, 1984, modified); C) Mount Albo tectonic map and interpretative sketch-map; T: thrust and reverse faults; F: folds axes; N: normal faults; V: veins, dykes or extension fractures; R, R': synthetic and antithetic shears; x, z: maximum and minimum strain axis.

In this kinematic system the bending of the fault near P. Su Mutucrone produces a so called "restraining bend" (Fig. 4) where part of the strike-slip component is transformed to reverse-slip and the fault plane dip shallows.

4. 4. *Antithetics and Joints*

The Nuoro Fault and the synthetic upthrusts are approximately parallel to the long direction of Mount Albo and, in the massif, two secondary transverse tectonic

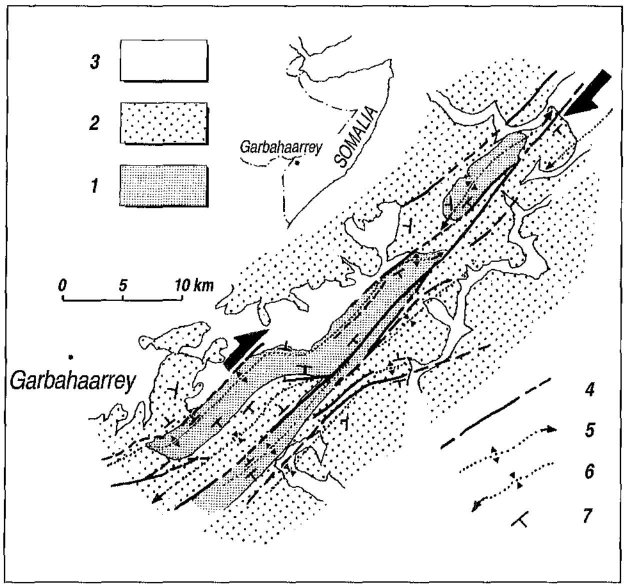

Fig. 7 - Tectonic sketch-map of the right-lateral strike-slip fault zone in the Upper Giuba Valley (Southwestern Somalia) (after Carmignani et al., 1984); 1) Upper Jurassic; 2) Upper Jurassic-Lower Cretaceous; 3) Tertiary-Quaternary; 4) Faults; 5) Anticlinal axial plane traces and axial dipping; 6) Synclinal axial plane traces and axial dipping; 7) Bedding strike and dip.

lineaments (Fig. 2) exist:

— a subvertical N110-140° trending fault system that is developed primarily in the central and northern parts of Mount Albo but doesn't appear to have important throw. South of M. Turuddò a fault belonging to this system is characterized by a right-lateral separation. The orientation and separation sense of these faults indicate that they could be antithetic faults with respect to the Nuoro Fault.

— a subvertical N150-170° trending fracture system that is less well developed and does not appear as a prominent lineament.

5. Kinematic Interpretation

In synthesis, the main features characterizing the Mount Albo structures are:

— Mount Albo is a Mesozoic carbonate massif along the Nuoro Fault, representing an important tectonic lineament;

— kinematic indicators along the Nuoro and Schiriddè Faults and along some synthetics, always indicate a left-lateral strike-slip movement with a minor component of reverse-slip;

— southern Mount Albo faults have an "en-echelon" arrangement and are subparallel to the Nuoro Fault; they are synthetic upthrust steepening toward the Nuoro Fault;

— two high angle (with respect to the Nuoro Fault) fault systems in the Mount Albo area do not show important throw;

— the intensity of deformation rapidly decreases away from the Nuoro Fault;

— even though basement and sedimentary cover were deformed by different mechanisms, they behave, on a mesoscopic scale, as a unit without relative detachment.

We suggest that all these data can be interpreted within the context of a typical left-lateral transpression regime (Harland, 1971), or convergent wrenching regime (Wilcox et al., 1973, Harding, 1974), in which the slip vector and fault trends are not parallel. Sanderson & Marchini (1984) ascribed transpression to a simple shear deformation plus a pure shear deformation with shortening oriented normal to the shear zone trend and with extension oriented vertically which corresponds to thickening of that same zone.

The geometry expected in a left-lateral simple shear deformation (Fig. 6A) is then modified in transpression as follows:

— folds and reverse faults tend to parallel the main fault while other structures (synthetic, antithetic and normal faults) develop at a larger angle to the shear zone (Sanderson & Marchini, 1984) (Fig. 6B);

— important vertical movement components and faults with reverse sense of apparent dip-slip separation (Harding, 1974) are developed;

— synthetic faults (trending vertically in simple shear) acquire a convex-upward shape (upthrusts) (Wilcox et al., 1973, Woodcock & Fischer, 1986, Naylor et al., 1986), defining the so-called "flower structures" (Harding & Lowell, 1979) described, for example, by Lowell (1972) (Spitsbergen) and Sylvester & Smith (1976) (Mecca Hills, S. Andreas Fault zone).

Finally, comparing a tectonic sketch-map of Mount Albo (Fig. 6C) with Figures 6A and B, we can assume the Nuoro Fault as the master fault of a left-lateral convergent strike-slip system. The southern Mount Albo synthetic upthrusts clearly represent "half flower structures" (they all dip to SE and are asymmetric with respect to the main fault) indicating transpression; the N110-140° and N150-170° trending fault systems can represent anthitetic faults and tension fractures, respectively. Even the character of the Schiriddè Fault agrees with this kinematic interpretation because it is dominantly a left-lateral strike-slip fault where it tends to parallel the Nuoro Fault, yet it is primarily a reverse fault normal to the shortening direction in the P. Su Mutucrone zone. Vertical movements are also important because the Paleozoic M. 'e Senes Granite reaches more than 800 m in heights (M. 'e Senes area) and the Mesozoic M. Bardia Limestone crops out along the Nuoro Fault at less than 200 m in height.

On the basis of these detailed structural data and in accordance with theoretical models and regional geological examples, we therefore regard the Mount Albo massif represents a "half-flower structure" developed along the transpressive left-lateral Nuoro Fault.

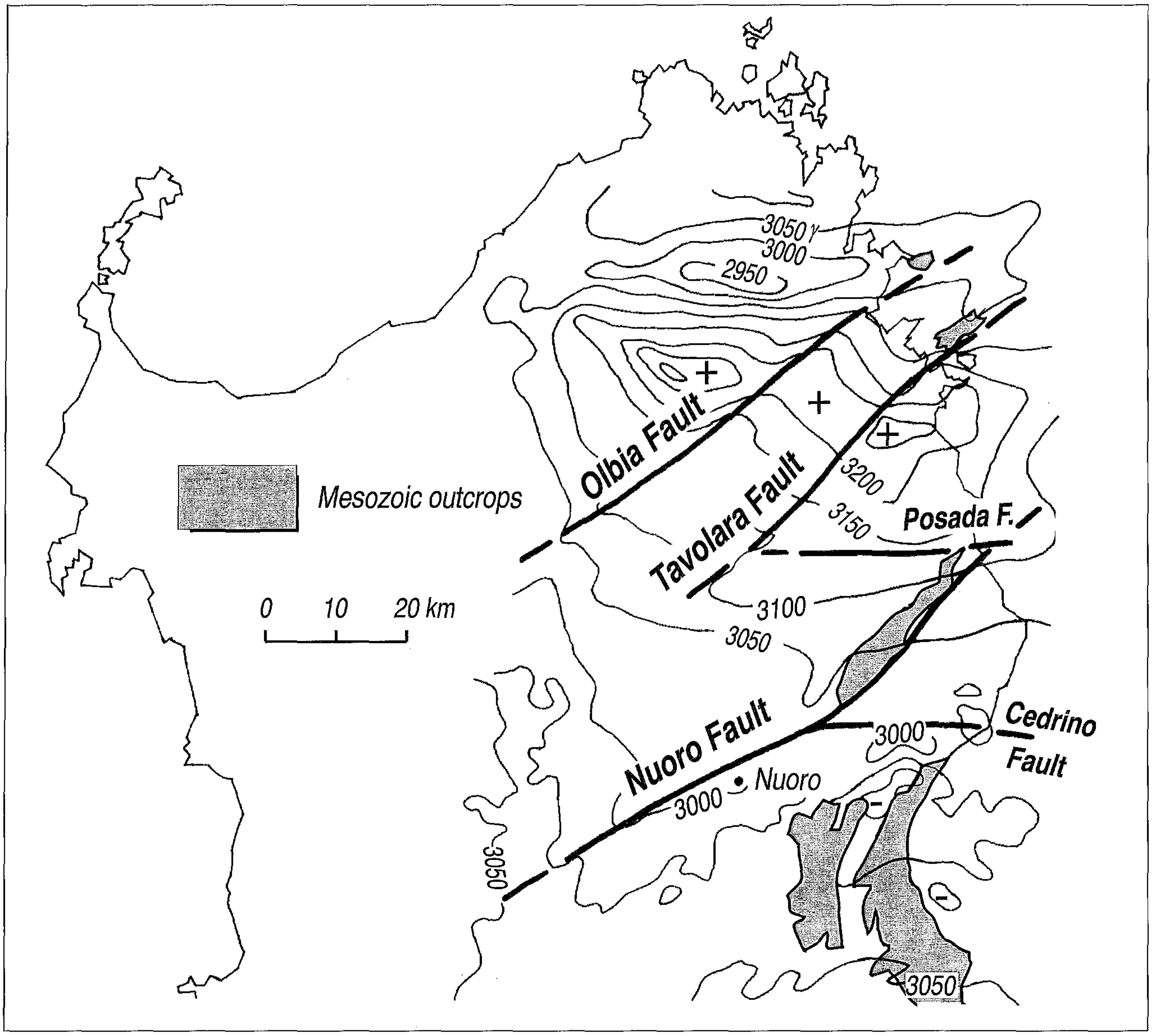

Fig. 8 - Main faults and aeromagnetic map of northeastern Sardinia.

6. The M. Nurres Structure

As previously described, the Mount Albo massif and its carbonate sequence trend NE south of Siniscola. From Siniscola to Posada (Fig. 2) a series of minor Jurassic highs joined to Mount Albo trend N-S from Siniscola to M. Nurres and turn back to the NE trend from M. Nurres to Posada (Fig. 2). In southern area (Siniscola) the change in orientation is rather gradual, while in the north (M. Nurres) it is sudden and occurs in connection with a NW striking fault.

We suggest this "S-shaped" trend of the northern termination of the Mount Albo massif can be explained as a drag fold in a left-lateral strike-slip system (the Nuoro Fault), in accordance with our kinematic interpretation (parag. 5.).

A similar interpretation has been suggested by Conti (1989) in work regarding a deformed, sedimentary Mesozoic belt in southwestern Somalia (Fig. 7). In this case, anticlinal-synclinal structures would have been dragged and deformed to a "Z-shape" because of the movement along a right-lateral strike-slip fault.

Analogies between these two structures are evident as shown in Figures 7 and 2.

7. The North Eastern Sardinia Fault System

The Olbia and Tavolara Faults have important analogies with the Nuoro Fault (Fig. 1):

— same direction (NE-SW);

— northwest of the Olbia and Tavolara Faults the

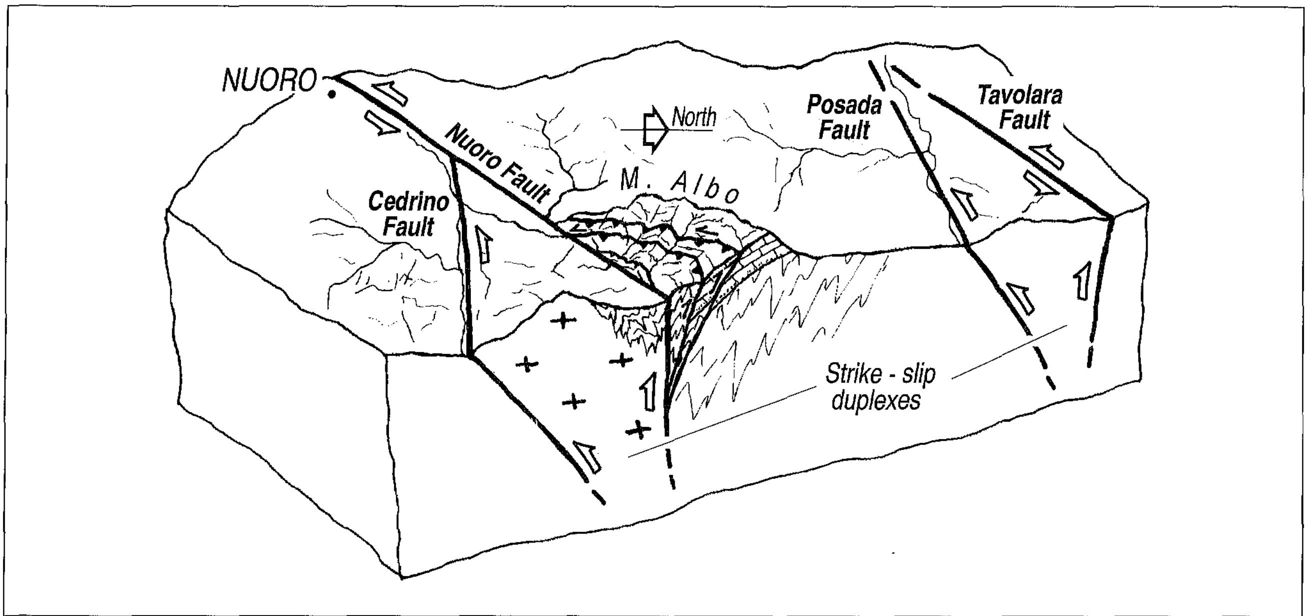

Fig. 9 - Schematic block-diagram of Mount Albo and surrounding areas; the two west-tapered wedges between the Nuoro and Cedrino Faults and the Tavolara and Posada Faults can be ascribed to strike-slip duplexes.

Mesozoic carbonate massifs of C. Figari and Tavolara Island crop out. Bedding orientation and fault relationships are the same as that of the whole Nuoro Fault-Mount Albo massif;

— in the aeromagnetic map of northeastern Sardinia (Cassano et al., 1979) (Fig. 8), some magnetic anomaly maxima are affected by a left-lateral offset that is probably due to the two faults; this is in accordance with the left-lateral movement along the Nuoro Fault.

We can therefore assume that all these faults represent a deformation system intersecting the E-W system (Posada and Cedrino Faults) thereby producing two west-tapered wedges (Fig. 1) that can be considered strike-slip duplexes (Woodcock & Fischer, 1986), as represented in Figure 9: one wedge, strongly uplifted, is bounded by the Nuoro and Cedrino Faults, the other one is bounded by the Tavolara and Posada Faults. Such a system results in a NNE-trending direction of maximum shortening.

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THE HERCYNIAN BACKTHRUSTS OF EASTERN IGLESIENTE (SW SARDINIA): AN EXAMPLE OF INVERSION TECTONICS

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Abstract

The Iglesiente area (SW Sardinia) was affected, during Hercynian orogenesis, by fold and thrust tectonics under very low grade metamorphic conditions. Upright N-S trending folds deformed a Cambrian-Devonian succession. The interference with the previous Caledonian E-W open folds (Sardinian phase), give rise to the development of Dome and Basin structures all over the area. Neverthless in the Eastern Iglesiente the main features are represented by eastward backthrusts overthrusting the Cambrian formations onto the Ordovician deposits. Their development is due to an inversion tectonics, consequent to the closure of the Iglesiente basin. Reactivation of steeply dipping synsedimentary extensional faults give rise to upright thrusts in the western Iglesiente and gently dipping backthrusts in the eastern Iglesiente, with the achievement of a "Pop Up" structure of the Iglesiente Dome. Furthermore all these features emphasize that the Iglesiente was an external portion of the Sardinian Hercynian belt in which the pre-existing extensional structures have strongly influenced the compressional history.

Key words: *Paleozoic, Hercynian orogenesis, inversion tectonics, backthrusts, Sardinia.*

1. Introduction

The thrust-fold belt cropping out in the Iglesiente area (Southwestern Sardinia) is regarded as the "External Zone" of the Hercynian chain in Sardinia (Carmignani et al., 1978, 1981). Very low grade metamorphism affected the paleozoic sequences during the Hercynian orogenesis. The main structures are typical "Dome and Basin" interference pattern, resulting from the overprinting of N-S Hercynian folds onto E-W Caledonian folds. In the Eastern Iglesiente N-S Hercynian backthrusts developed, with an opposite displacement respect to the overall W and SW vergenge of the Hercynian chain. The aim of this paper is to debate the origin of these backthrusts as the result of inversion tectonics due to the closure of the Iglesiente basin that developed since the Cambrian time.

2. Stratigraphic Outline

The oldest sequences of the Sardinian paleozoic basement crop out in the Iglesiente area (Southwestern Sardinia) (Fig. 1). Two sedimentary cycles have been recognized (Carmignani et al., 1982; 1986): an early Caledonian cycle, made up by a Cambrian - Early Ordovician sequence and a later Hercynian cycle (Upper Ordovician - Devonian) that unconformably covered the Caledonian one.

The Caledonian cycle started with the mixed terrigenous- carbonate deposits of Nebida Formation of Botomian age, consisting of two members. The lower one, Matoppa Member, is made up of green siltstones and sandstones alternating with an archaeocyaths bearing limestone of Lower Cambrian. The upper one, Punta Manna Member, starts with an oolitic horizon followed by a rhytmic alternance of sandstones, shales and limestones. In the upper part the carbonate intercalations are composed of early dolomite with algal mats and dessication structures. The following Gonnesa Formation is represented by a thick carbonate sequence, made up by dolostones and limestones. It begins with the Dolomia Rigata Member costituted by early dolomite with algal mats and dessication structures evolving to "vadose pisolite". The occurence of archaeocyaths in the upper half of this sequence allow to attribute a Botomian age to the Dolomia Rigata Member (Debrenne & Gandin, 1985). The overlying Calcare Ceroide Member (Lower Cambrian, Toyonian) is represented by pearl-gray or dark-gray limestone, generally massive and locally stratified. The transition from the Dolomia Rigata Member to the Calcare Ceroide is marked by an interval of Dolomia Blu (Galassi & Gandin, 1992). In the Iglesiente area it appears to shows a maximum thickness of nearly 200 m. Four typical lithofacies characterize this Member (Debrenne & Gandin, 1985); mudstone-wackstone with bioclasts; grainstone with oolites and/or oncolites, echinoderms and trilobite remnants; cryptalgal boundstone, locally skeletal-algal boundstone with archeocyaths and

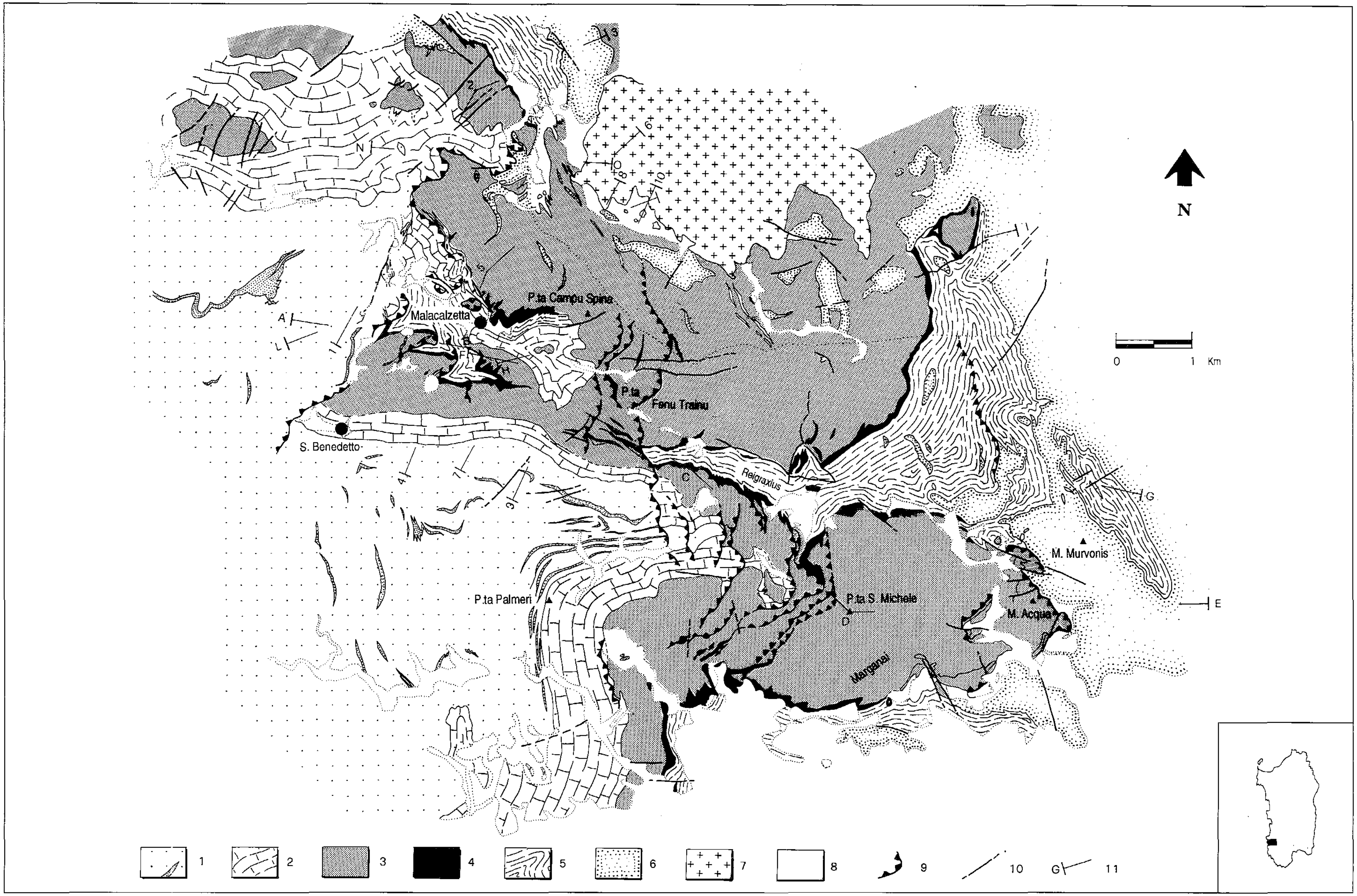


Fig. 1 - Geological sketch map of Eastern Iglesiente area; 1) Nebida Formation: sandstones with archaeocyaths bearing limestones lenses; 2) Dolomia Rigata Member; 3) Calcare Ceroide Member; 4) Nodular Limestone Member; 5) Cabitza Shale Member; 6) Ordovician Formation; 7) Late Carboniferous granite; 8) post Hercynian deposits; 9) Backthrusts; 10) Faults; 11) traces of geological cross sections.

"vadose pisolite"

The Cabitza Formation (Middle Cambrian - Early Ordovician) starts with gray, pink nodular limestone alternating with red and green shales. These facies represent the Nodular Limestone Member (Cocozza, 1980), and lie with parallel unconformity on the Calcare Ceroide Member. The upper member of the Cabitza Formation consists of alternating with shales siltstones with intercalations of fine-grained sandstones in the lower part. The bottom of the upper member has been attributed to Middle Cambrian (Rasetti, 1972) whereas the finding of Early Ordovician acritarchs and graptholites at the top of this formation indicate a Tremadocian age (Barca et al., 1987).

The Hercynian cycle begins with a polygenic unsorted conglomerate with a red-violet silty matrix ("Puddinga Ordoviciana Auct."). It lies with an angular unconformity on the underlying formations of the Caledonian cycle. The clasts represent some lithofacies of the Cambrian sequence (Calcare Ceroide and Cambrian breccias associated, Cabitza Shales), their grain size gradually decrease upwards into a microconglomerate facies followed by a sequence of alternating sandstone and siltstones.

The first well dated horizon, occurring toward the top of this sequence, consists of fossiliferous shales, siltstones, sandstones and carbonate shale of Caradocian - Ashgillian age. Thin and rare layers of basic volcanics, or their reworked subaerial products, intercalated in the upper part of the sequence (middle - lower Ashgill; Leone et al., 1991), represent the evidence of post-Caledonian magmatism (Beccaluva et al., 1981; Memmi et al., 1982). The Silurian deposits consists of siltstones, shales, black carbonaceous shales with graptolites and black shales with Orthoceras bearing limestones lenses, followed by Devonian shales and thin bedded limestone with lenses of "griotte" - type limestone with crinoids, cephalopods and conodonts (Gnoli et al., 1988).

2.1. *Paleogeographical Evolution*

The Paleozoic sediments of the Iglesiente area represent the depositional evolution of a continental shelf during the Cambrian/Lower Ordovician (Gandin, 1987). An extensional tectonics was active since al least the sedimentation of the Punta Manna Member, giving rise to the fragmentation of the Cambrian shelf that deepened along a ramp toward East and South. The deposition of the Calcare Ceroide was controlled by the extensional tectonics. A Bahamas-like carbonate platform, gently sloping toward East and South developed on the previuos formed ramp, (Gandin, 1987; Cocozza & Gandin, 1990). In the Middle Cambrian the platform drowned and the terrigenous sedimentation started again with the Cabitza Formation, representing a syn-rift deposit. Afterwards these sediments were sligthy deformed by the compressive tectonics of the Sardinian phase.

This compressive phase was followed by an extensional phase with the development of extensional and transtensional grabens and/or halfgrabens. Depositional features of the Puddinga Beds show an eastward transition from alluvial fan to marine shelf, evolving later towards a confined basin with euxinic sediments during the Silurian. The sedimentation occurred within narrow fault-bounded basins, one of them, according to Martini et al. (1991) had the NW-SE Gonnesa fault as a boundary fault. The Paleozoic Iglesiente basement belongs to the Gondwana continental margin that underwent to passive margin evolution since the Infracambrian (?) - Cambrian time (Carannante, et al., 1975; Carmignani et al., this volume)

3. Tectonic Evolution

The Iglesiente area represents the outermost part of the Hercynian chain in Sardinia (Carmignani et al., 1981). It was affected by very low grade metamorphism associated to fold and thrus tectonic during the Hercynian orogenesis. The Caledonian structures, recognizable in the Cambrian sequences, were deformed by the Hercynian folds. The following deformation history can be outlined (Arthaud, 1963; 1970; Carmignani et al., 1982; 1986; Dunnet 1969; Poll, 1966; Poll & Zwart, 1964):

— Sardinian phase: open large-scale folds with E-W axes (Fig. 2) affecting the Cambrian - Early Ordovician sequence, without axial plane foliations;

— First Hercynian phase: minor E-W trending folds, accentuating the Caledonian folds;

— Second Hercynian phase: main deformation, N-S trending folds and thrusts (Fig. 2). Accompanied by subvertical axial plane foliation;

— Third Hercynian phase: small deformations with minor folds.

The interference between the Caledonian and Hercynian fold systems (Fig. 2), give rise to typical Dome and Basin structures. The "arenarie dome" north of Iglesias town (Fig. 1) is a typical example of this interference pattern, with strongly dipping axial planes and nearly orthogonal axial directions.

3.1. *Caledonian Deformation*

The Sardinian phase, affecting the Cambrian sequence, was responsible of the occurence of large scale E-W trending folds such as the Iglesias syncline and the Marganai anticline (Fig. 3). Minor structures are represented by E-W concentric folds, without foliation and with vertical or strongly dipping axial plane. The Middle Ordovician age of this deformation is supported by the unconformity at the base of Puddinga Beds unconformably covering the E-W structures (Barca et al., 1987).

3.2. *Hercynian Deformation*

The Hercynian orogenesis began with a weak deformation phase giving rise to E-W folding structures overprinting the Caledonian one. Metric and decametric folds affecting the Puddinga Beds at north of Domusnovas and at south of Gonnesa, supported the occurence of this phase (Dunnet,1969; Arthaud, 1970). Furthermore these folds prevent the evaluation of importance and style of the Caledonian structures. Neverthless these latter are characterized by a major wavelenght and represent the main folding structures in the Iglesiente area. The major deformation took place during the second Hercynian phase. N-S trending folds and thrusts, affecting the whole sequence as far as the Silurian - Devonian deposits, gave rise to the greatest shortening.

3.2.1. *Folds and Thrusts*

N-S trending folds characterized by subvertical axial planes developed. Their wavelenghts range from hectometric to metric scale, but it result always shorter respect to the caledonian folds. The upright axial plane foliation evolves from a discontinuous spacing foliation in the metasandstones and limestones to well-developed foliation in the shale. A synkinematic low grade metamorphism affected the sequences during this phase. Fold axes have a variable plunge, due to the occurence of previus E-W structures, indeed on the limbs of the E-W folds, they achieve a subvertical plunge toward north or south. The main folding structures outcrop in the Western Iglesiente (Buggerru - Aquaresi area), represented by large scale uprigth open anticlines and narrow synclines. Minor folds (metric to decametric scale) occur at the contact between the Calcare Ceroide and Cabitza formation with a typical "cuspe and lobe" shape. The shales were "phinched" within limestones, giving rise to narrow synclines separated by open anticlines. Within these synclines the Calcare Ceroide limestones show a strong deformation with the development of a N-S cleavage. This structures already described by Zuffardi (1965) and Dunnet & Moore (1969), can be observed in correspondence of the limbs of the E-W synclines of Domusnovas, Reigraxius and Malacalzetta (Fig. 1). The narrow N-S synclines of Buggerru, Aquaresi and Masua, whose core is made of Cabitza Shale and Ordovicain Puddinga, represent similar structures at large scale.

The main structures of this phase are represented in the western Iglesiente (Masua - Gonnesa area) by N-S trending thrusts steeply dipping toward East. They are characterized by a westward displacement, giving rise to the tectonic superposition of Early Cambrian Gonnesa Formation onto the Cabitza Formation or Ordovician Formation. Otherwise the Eastern Iglesiente is characterized by numerous and well developed eastward backthrusts, moderate to gently dipping toward west. The main backthrusts in this area are represented by:

— M. Acqua backthrust (Fig. 3, sez. A-E): a slices of Calcare Ceroide is overthrusted onto the Ordovician Puddinga, with a tectonic displacement of at least 1 km. The bottom of Calcare Ceroide was intensely deformed with the development of a narrow foliated layer (0.5 - 2 m thick);

— S. Benedetto-Baueddu-Arenas (Fig. 3, sez. 1-3); in this area the E-W trending of Malacalzetta syncline is truncated by a major backthrust that moderate to gently dips toward West. This structure affected the whole sequence. The Early Cambrian sandstones (Nebida Formation), locally overturned and the dolostone and the limestone (Gonnesa Formation) were overthrusted onto the shales of Cabitza Formation. A detached slice of this backthrust is represented by the M. Cuccheddu klippe (near the Malacalzetta village; Fig. 3, sez. 4-6) where the Calcare Ceroide overlain the Cabitza Formation.

Other major backthrusts developed within the Marganai anticline between Punta Soa Martini and Punta San Michele (Fig. 1 and Fig. 3, sez. A-E). They cross-cut the E-W structures from the southern limb (Domusnovas syncline) to the northern limb (Reigraxius syncline). The sense of displacement of the backthrusts has been recognized by the occurence of cataclastic fault zones at the base of overthrusted block. Cataclastic and/or protomilonitic rocks developed within these fault zones with the occurence of S-C like structures (the S planes correspond to the foliation whereas the C planes represent the shear surfaces) indicating univocally a top the E and NE sense of shear. The slickensides striae on the shear planes give an overall NE-SW direction of displacement. These structures are clearly recognizable in correspondence of the above mentioned backthrusts. Sometimes the slikensides and striae have an horizontal component.

4. Inversion Tectonics

In the Iglesiente has been recognized the occurence of large backthrusts (i.e. hinterland vergent thrusts). They affected the Cambro-Ordovician sequence deposited in an extensional basin, westward bounded by a slope and eastward deepened along a ramp (Gandin, 1987). The geometry of this basin was achieved by nearly N-S trending listric faults, active since the Early Cambrian and partly reactivated during the Ordovician (Gonnesa Fault; Martini et al., 1991). The onset of Hercynian compressive deformation reactivated the Cambrian extensional faults as steeply dipping thrusts with a top to the west displacement, i.e. toward the foreland. Therefore the basin underwent to a progressive closure and, at the higher amounts of shortening, backthrusts developed all over the area resulting from the inversion of the basin. The influence of pre-existing extensional faults on compressional structural geometries is well documented in other complex orogenic belts such as the Alps and the Pyrenees (Hayward & Graham, 1989;

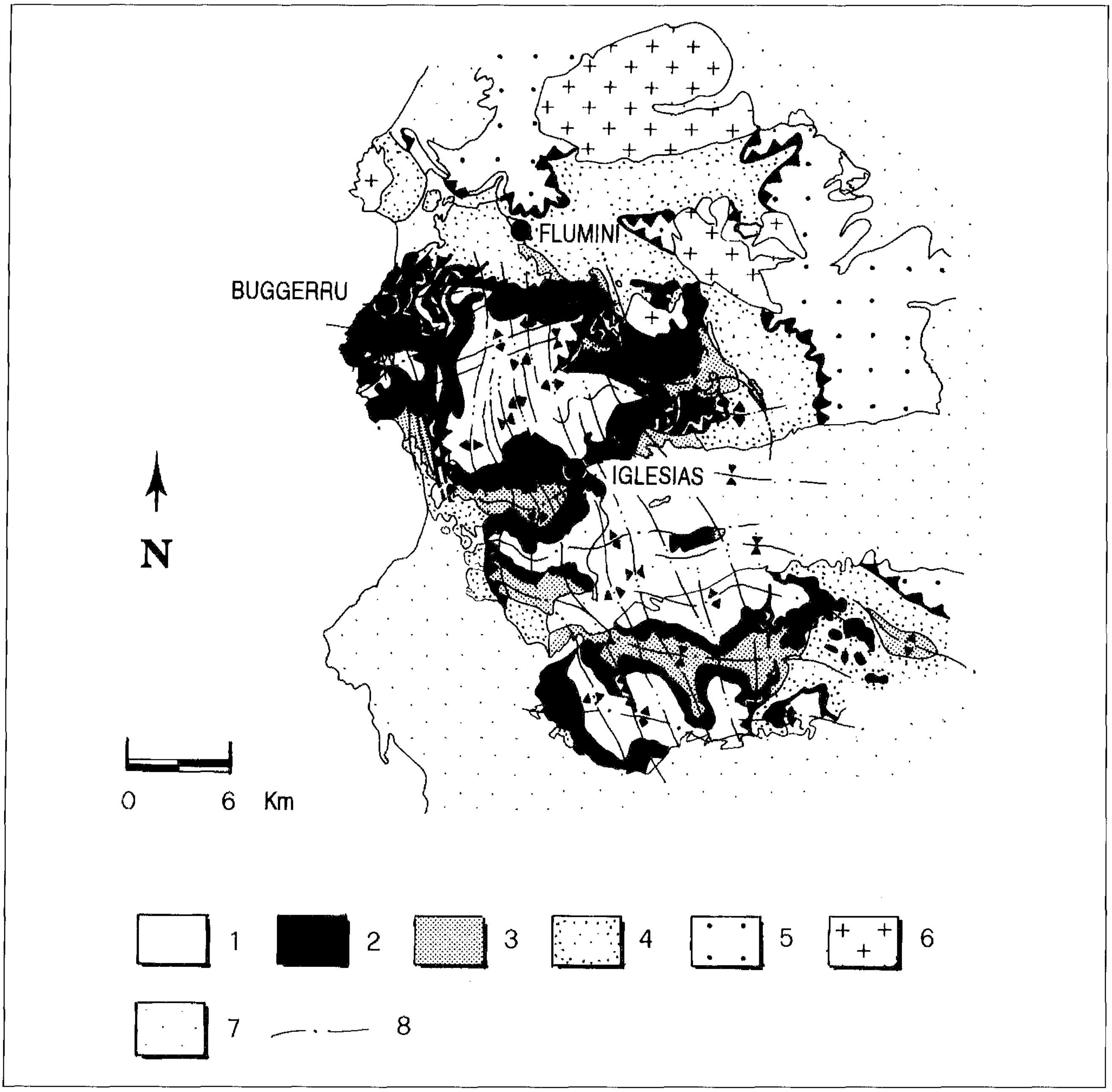


Fig. 2. Tectonic skecth map of SW Sardinia; 1) Nebida Formation; 2) Gonnesa Formation; 3) Cabitza Formation; 4) Ordovician Formation; 5) Arburese Unit; 6) Hercynian granites; 7) Post Hercynian deposits; 8) traces of axial planes of Caledonian (E-W) and Hercynian (N-S) structures.

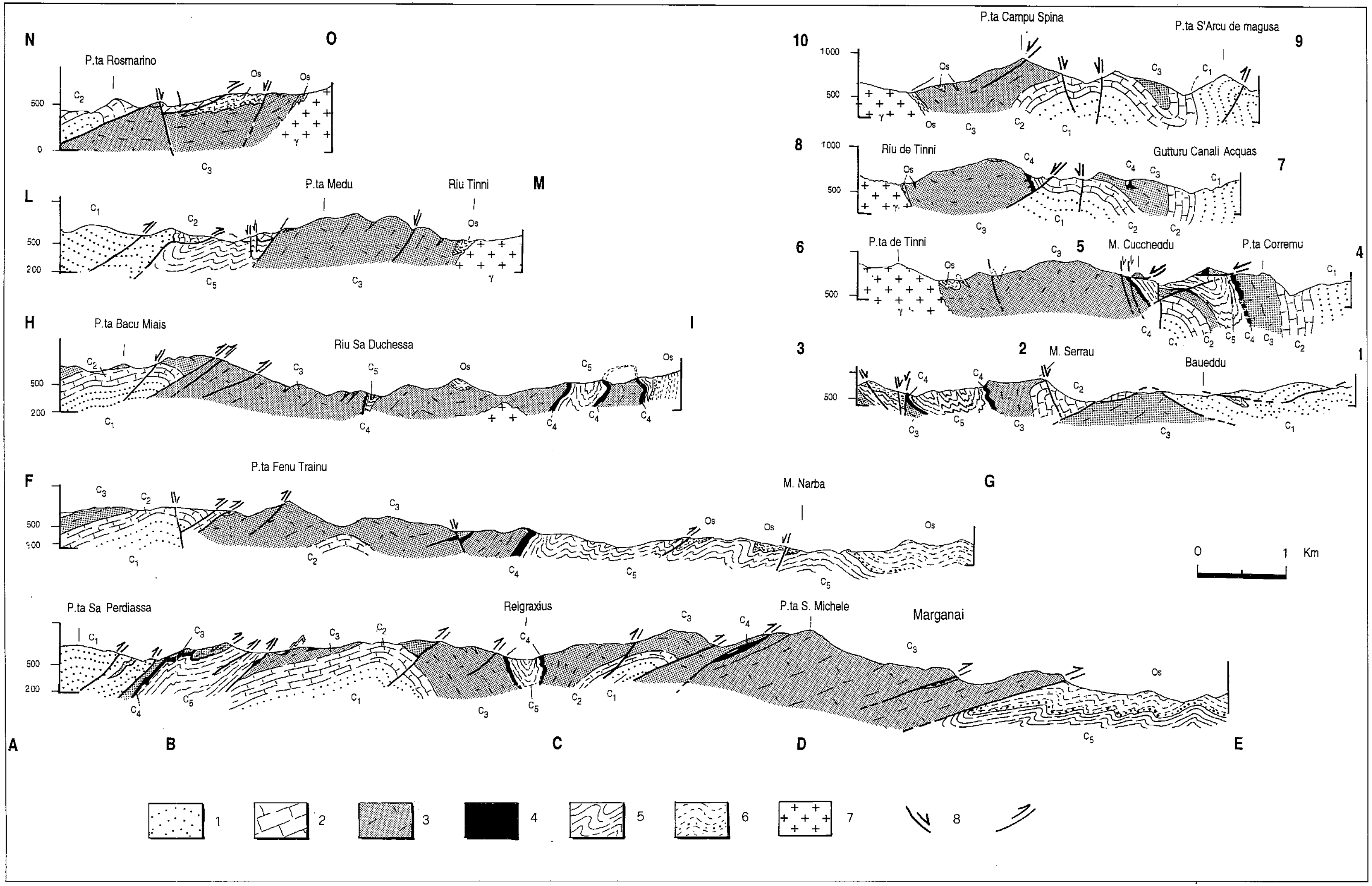
Velasque et al., 1989).

According to the experimental models of McClay & Buchanan (1992) and Buchanan & McClay (1991) the backthrusts may be caused by the buttressing effect of the concave upwards shape of the main listric or ramp flat extensional faults (Fig. 4) or by the reactivation of minor antithetic faults (Hayward & Graham, 1989). In the Iglesiente the backthrusts become progressively more numerous at later stages of inversion (previously formed N-S folds were truncated by thrusts on their reversed limbs). This result in a significant uplift that give rise to a "pop up" structure, made up by the "arenarie dome" of Iglesias (Fig. 5).

According to Hayward & Graham (1989) the extensive generation of backthrusts testify the high grade of inversion.

It is possible that the direction of compression did not coincided perfectly with the direction of extension. This "obliquity" can explain the presence of horizontal components of moviment on the surfaces of thrusts and bakthrusts

The westernmost margin of the Cambrian basin wasn't continue, but was probably affected by some "transfer-faults", nearly perpendicular to the direction of wideninig of the basin. This geometry influenced the development of the Hercynian structures and it is probably at the origin of



Fg. 3. Geological cross sections; 1) Nebida Formation (C_1); 2) Dolomia Rigata Member (C_2); 3) Calcare Ceroide Member (C_3); 4) Nodular Limestone Member (C_4); 5) Cabitza Shale Member (C_5); 6) Ordovician Formation (Os); 7) Late Carboniferous granite (g) 8) Backthrusts; 9) Faults.

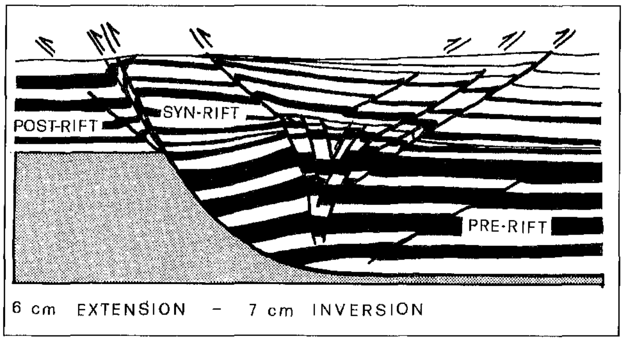


Fig. 4. Sandbox model of inversion tectonics of a simple listric fault (Mc Clay & Buchanan, 1992; simplified)

the alignement of foreland thrusts and backthrusts observed in the Western Iglesiente (Buggerru - Nebida area) as show in Fig. 2.

5. Conclusions

The SW area of the Hercynian chain in Sardinia may be regarded as the frontal part of an Early Carboniferous accretionary wedge. The wedge caused deformation and metamorphism in the continental crust, after the subduction of oceanic crust, happened in the northernmost parts of the chain in Devonian times (Carmignani et al., this volume; Matte, 1986; Pin & Vielzeuf, 1988).

In the frontal part the wedge exhibit Coulomb behaviour (Davis et al., 1983), well documented by the above described deformation structures such as thrusts and backthrusts.

The P-T conditions of the rock of Iglesiente (very low grade metamorphism) suggest that they have not suffered an important underplating during orogenic deformation. The active mechanism was that of frontal accretion of the orogenic wedge (Platt, 1986). There was accumulation of material at the tip of the wedge tending to lenghten the wedge itself. According to Platt's model, the response of the wedge was of shortening in order to maintain its stabilty. The frontal region was therefore in compression, causing the formation of folds, new thrusts, out of sequence thrusts and backthrusts. In the Iglesiente area the formation of backthrusts was facilitated by the occurrence of pre-existing extensional listric faults, whose concave upward shape caused the extensive formation of backthrusts.

According to Hayward & Graham (1989) the extensive generation of backthrusts is typical of case of severe inversion tectonics.

The response to shortenings caused also the activation of new westward thrusts (foreland vergent thrusts) joining up with the easteward backthrusts at the base of the "Arenarie Dome" (Fig. 5).

Moreover, it is possible that the seismic discontinuity observed at nearly 5 km of depth below the Iglesiente (Civita et al., 1983), might represent a relict of an extensional fault reactivated as a thrust during inversion in

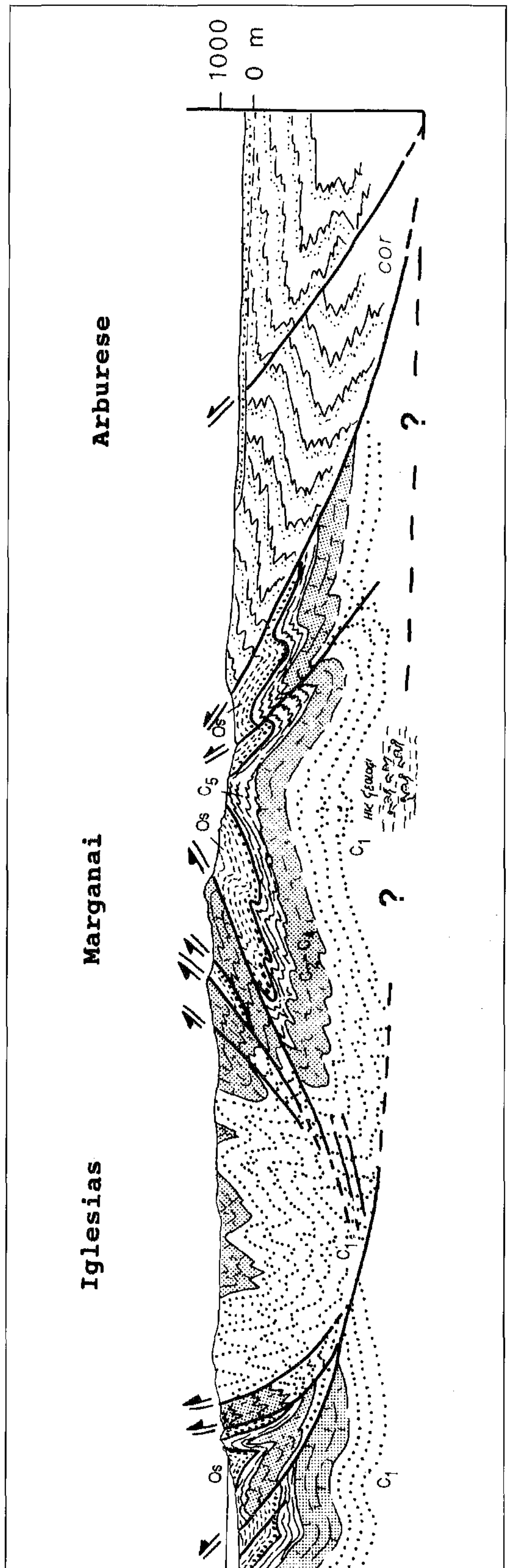


Fig. 5 - Interpretative schematic geological cross section of Iglesiente:1)Nebida Formation (C_1);2) Dolomia Rigata Member, Calcare Ceriode,Member and Calcescisti Member (C_2 - C_4); 5) Scisti di Cabitza Member (C_5); 6) Ordovician Formation (Os).

Hercynian orogenesis.

Therefore Iglesiente may represents a case of an external part of orogenic belt, with complex tectonic evolution, where the geometry of pre-existing extensional structures has exerted a strong control on compressional structures during Hercynian orogenesis.

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PRE-HERCYNIAN MAGMATIC ACTIVITY IN THE NAPPE ZONE (INTERNAL AND EXTERNAL) OF SARDINIA: EVIDENCE OF TWO WITHIN PLATE BASALTIC CYCLES

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Abstract

Recent field data on stratigraphic and structural relationships among the Paleozoic sequences in the "Nappe Zone" of the Variscan chain in Sardinia, from Sarrabus to Nurra, have made possible to set in a more detailed spatial and chronological framework the pre-Hercynian intermediate-basic magmatic activity.

Review of the outcropping patterns between metabasites and country rocks and analysis of their fabric, geochemistry, structural and paleogeographic setting, allow us to propose a new interpretation of the pre-Hercynian tectono-magmatic evolution.

The nappe belt has been divided into an internal and external zone (Carmignani et al., this volume) on the basis of the paleogeography of different, coeval stratigraphic sequences that existed prior to Variscan continental collision and shortening.

New sampling has been performed taking into account this division and paying particular attention to the effusive or intrusive character of the rocks, so that correct stratigraphic correlation can be made. We also have rehexamined several literature chemical data taking care to set them in their correct stratigraphic position. Results presented in this paper show that a first Within Plate (WP) basaltic cycle occurred between Caradoc and uppermost Ashgillian and a new cycle during uppermost Devonian and lower Carboniferous. Furthermore some chemical differences have been detected between the metabasalts moving from External Zones to Internal Zones; we discuss their meaning in the evolution of the Sardic Hercynian belt.

Key words : *Sardinia, Hercynian Chain, Nappe Zones, Magmatic Cycles, Within Plate Basalts.*

1. Introduction

The Paleozoic magmatism of Sardinia is widespread throughout the Variscan chain of the island. The most important magmatic cycle is characterized mainly by effusives and minor intrusives both showing a calc-alkaline character (Memmi et al., 1983). This volcanism has recently been tied to continental margin arc activity (Garbarino et al., 1981; Carmignani et al., this volume). Both radiometric (Ferrara et al., 1978; Di Simplicio et al., 1975) and stratigraphic data suggest it occurred in a time interval spanning between Arenig and Caradoc (Carosi et al., 1987).

Chronological data for the subsequent alkaline-to-transitional WP magmatism are less definitive but, for some authors, support initiation in the Upper Ordovician as result of continental rifting which continued through the Silurian (Ricci & Sabatini, 1973, 1978; Beccaluva et al., 1981; Memmi et al., 1983).

In this paper we present the results of new sampling and new correlation of previous literature data brought to light by new structural and stratigraphic evidence from different areas of the Nappe Zone (moving from the internal up to the external) as well as of the Inner Zone (Nurra) and of the External Zone (Iglesiente).

These results enables us to describe a succession of magmatic events that can be related to several different geodynamic models.

We refer to metamorphic rocks from all the nappes with a more internal provenance than the "Gerrei type nappe" as INZ, and those from External Zones and External Nappe Zones as ENZ (Fig. 1).

Until now data available on WP type metabasalts in Sardinia were considered to represent an upper Ordovician and Silurian magmatic cycle, related to intra-continental extension and crustal thinning possibly progressing to incipient oceanization (Ricci & Sabatini, 1978).

2. Presentation of Data

2.1. *Structural and Stratigraphic Relantionships*

Basic WP rocks of uppermost Ordovician age crop out as metabasites derived both from extrusions and from hypabissal intrusions.

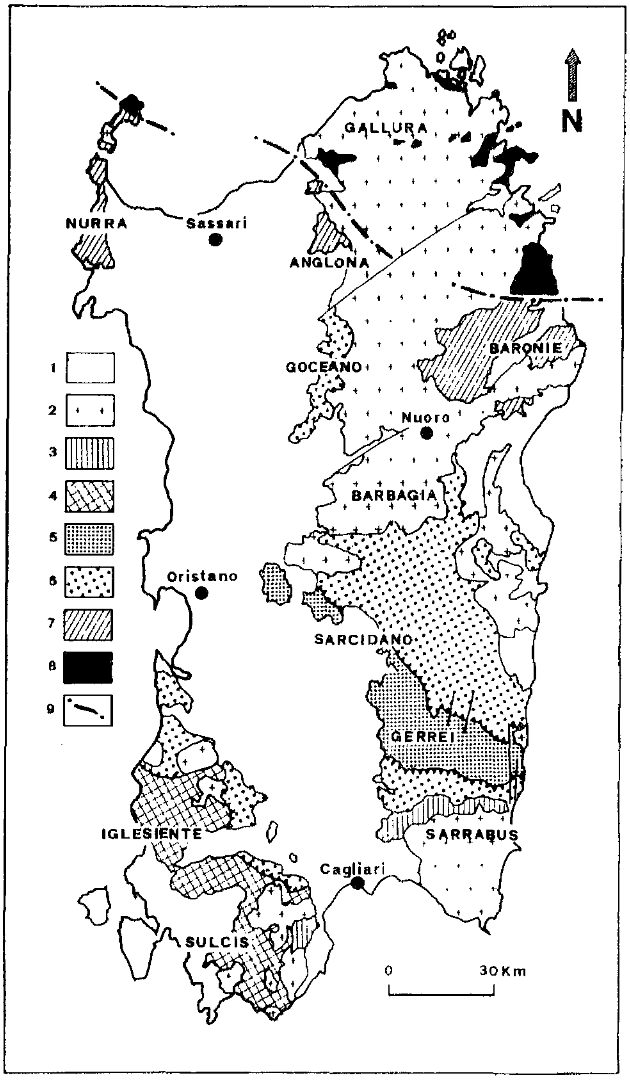

Fig. 1 - Geological sketch map of the Sardinian basement. 1: Post-Hercynian covers; 2: Hercynian batholith; 3: Foreland Carboniferous deposits; 4: External Zone (back-arc region, Iglesiente); 5: External arc Nappes; 6: Inner-arc Nappes; 7: Inner zone of the chain; 8: High Grade Metamorphic Complex; 9: Hercynian Suture Zone.

Metabasic effusives in the External Zone (Iglesiente) are embedded within upper Ashgillian terrigenous deposits (Beccaluva et al., 1981; Leone et al., 1991; type-1 in Fig. 2 and in Tab. 1); in the external nappes WP metabasalts are interlayered with transgressive clastic deposits of Caradocian age and in some cases they contain pillow structures (Gattiglio & Oggiano, 1990; type-1)

In the internal zone (Nurra; type-1) and in the internal nappes (Goceano; type-2) we have inferred a Caradocian age for the WP metabasalts on the basis of lithostratigraphic correlations (Di Pisa & Oggiano, 1984) to the well dated fossiliferous sequences of more external pertinence.

Moreover the upper Ordovician magmatic activity is also represented by dike swarms and hypabyssal intrusive bodies (type-2) hosted in rocks that range in age from Tremadoc-Arenig (Solanas formation) up to Llanvrin-Llandelio (Barbagia volcanoclastic complex).

The younger WP metabasalts crop out either within a Culm like deposit (Maxia, 1984) referable to the Dinantian in Sarrabus (Barca & Olivieri, 1991; type-3) or within a supposed Siluro-Devonian succession in the Nurra region (type-4). The former show mainly effusive or hypabyssal features; the latter are represented by metagabbroic stocks and sills: their intrusive relationships with a Siluro-Devonian succession point to a post-Silurian (or even post-Devonian) age and suggest that there was no WP basaltic activity during Silurian time.

The effusive character of type-1 and of part of type-3 rocks provides the strongest evidence for the existence of two WP magmatic cycles in uppermost Ordovician and lower Carboniferous, respectively.

2.2. *Petrographic and Geochemical Data*

In spite of the mineralogical and textural metamorphic transformations, intrusive and effusive fabrics can be easily distinguished: the textural features reflect the metabasite-contry rock relationships. Metamorphic facies range from very low grade up to amphibolite based on the typical mineral assemblages.

Effusives sometime underwent spilitization and in some cases show pillow-like structures suggesting a sub-marine emplacement. On the other hand, many effusives appear strongly massive with an high degree of cohesion. Grain sizes are always small if blasto-porphyritic textures are not present. Original magmatic minerals are occasionally preserved and consist essentially of albitized plagioclase, small, homogeneously distributed opaques and rarely pyroxenes as remnant grains normally enclosed in metamorphic amphibole.

Intrusives are always massive and cohesive in appearence and are medium to coarse grained. These contain the most well-preserved magmatic textures consisting of generally large laths of euhedral-subedral albitized plagioclase and opaques, often abundant fine to medium grained apatite prisms and rare amphibolitized pyroxenes. The metamorphic texture is generally blastophitic.

The analyzed rocks sometime show geochemical characters which appear tied to significative chemical modifications. Metamorphism, alteration and occasionally spilitisation concur to mobilize elements with low ionic potential. Any present consideration must be developed only taking into account High Field Strenght Elements (HFSE), which are generally considered immobile or slightly mobile during the above mentioned syn-to-post magmatic processes. In particular we have utilized ratios between HFSE to minimize even the possible modifications in the

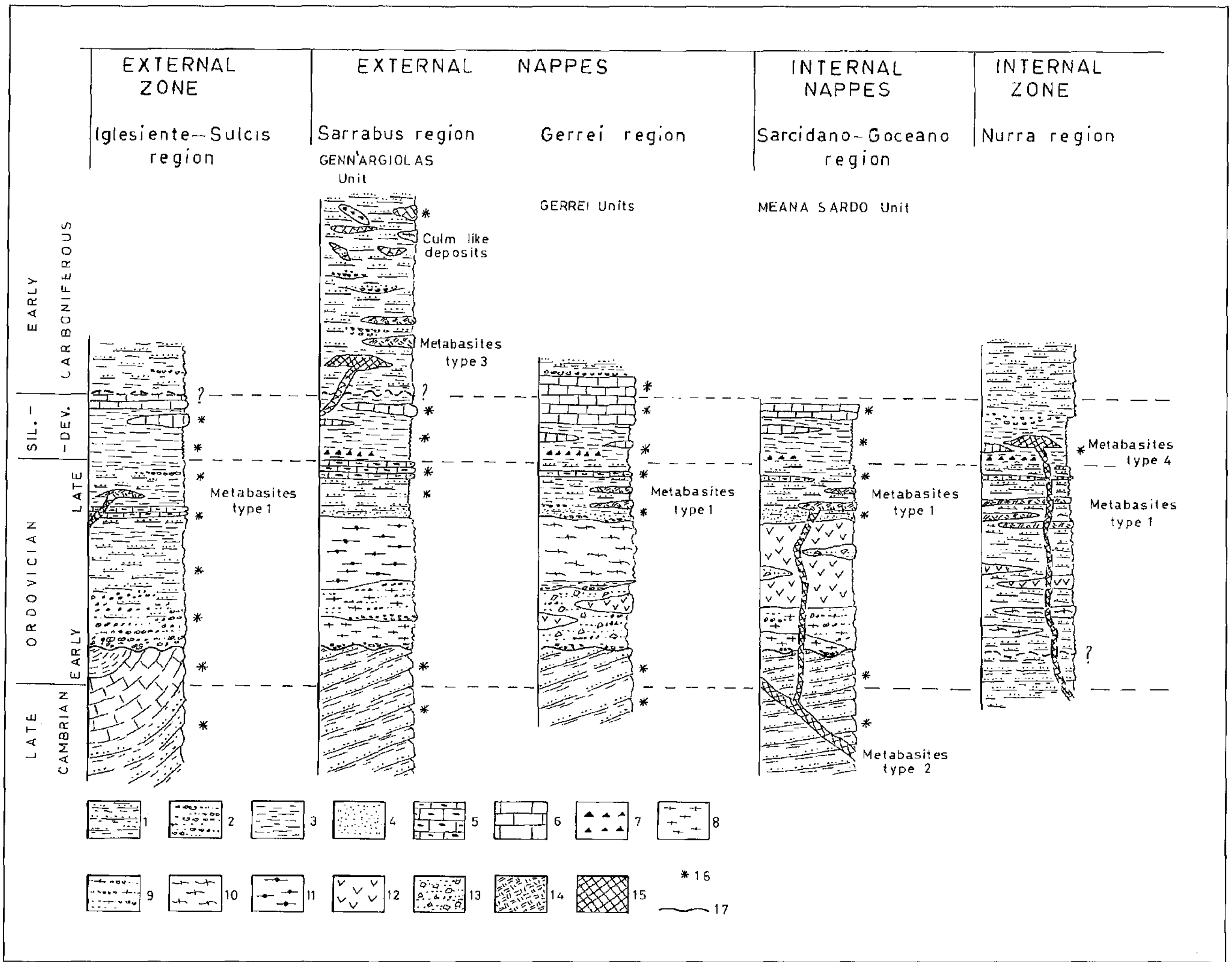

Fig. 2 - Schematic stratigraphic columns of External and Internal Zones and some Nappe Zones Tectonic Units showing the stratigraphic position of the metabasic rocks. 1: Metasandstones; 2: Metaconglomerates; 3: Metasiltites and Shales; 4: Metarkoses; 5: Metalimestones often silicified with Late Ordovician fauna; 6: Metalimestones and Metadolostones; 7: Black cherts; 8: Metaryolites; 9: Original pyroclastic products, volcanic metasandstones and reworked ryolites; 10: Metaryolites and metaryodacites with "augen texture"; 11: Metadacites and metaryodacites; 12: Metandesites and metadacites; 13: Metamorphosed reworked andesitic and dacitic lavas; 14: WP effusive metabasalts; 15: WP intrusive metabasalts; 16: Fossils; 17: Unconformity.

absolute content of them. The geochemical study of this type of rocks is always affected by these uncertainties but we believe that an accurate field screening followed by a laboratory data selection offer a reasonable releability for some interpretations of the data, expecially if some systematic features rise up. Furthermore, in this study we consider rocks with comparable degree of evolution. We take into account only basalts (rocks with $Zr/TiO2 \leq 0.010$; according to Winchester & Floyd, 1977 and Floyd & Winchester, 1978).

Original chemical data are reported in Tab. 1 together with several literature data which are here reproposed to facilitate the reader.

The analyzed rocks are separated into two groups according to the two different ages: Group A refers to the Upper Ordovician ones and Group B to the Upper Devonian-Lower Carboniferous ones.

2.3. *Ordovician Basalts vs. Carboniferous Basalts.*

The considered rocks, as a whole, plot within or near the WPB field of the Ti-Zr-Y diagram after Pearce & Cann (1973) (Fig. 3), in agreement with their environmental occurrence. The majority show a Nb/Y ratio > 1 and strongly differentiated REE (Rare Earth Elements) patterns.

2.3.1. *Group A (Upper Ordovician)*

These rocks show strongly differentiated REE patterns (La/Yb_n ratio 8.13 - 10.74) with quite similar $(La/Sm)_n$ and $(Ga/Yb)_n$ ratios (1.99 - 3.71 and 1.78 - 3.43, respectively) and with no or slightly positive Eu anomalies (Eu/Eu* 0.98 - 1.13) (Fig.4). Generally they show medium-high Ti/Y (295 - 1266) and Nb/Y (0.73 - 3.20) and low Zr/Nb (2.20 - 7.50).

Tab. 1a	INZ																									
	UPPER ORDOVICIAN																						LOWER CARBONIFEROUS			
	Nurra								Goceano										Barbagia	Sarcidano			Nurra			
Type	1	1	1	1	1	1	1	1	2	2	2	2	2	2	2	2	2	2	2	2	2	2	4	4	4	4
Sample	Vn 1*	Vn 3*	Vn 4*	Vn 5*	Mb 2*	334	336	338	S 1194	S 1199	S 1200	S 1201	S 1206	S 1208	S 1209	S 1268	S 1269	S 1356	Mb 36*	S 1655	S 1669	S 1670	Mb 5*	Mb 6*	M. Ru*	S 1703
SiO_2	45.31	46.62	45.47	44.17	43.92	46.09	45.50	46.36	44.28	48.83	42.97	47.95	45.46	48.58	44.15	47.69	48.52	47.30	47.34	36.49	45.15	42.28	46.37	50.50	46.18	45.32
TiO_2	3.11	2.35	3.00	2.29	2.78	2.89	2.30	2.57	2.91	2.35	3.38	1.87	2.27	3.20	3.14	3.86	2.16	2.33	2.27	2.39	2.34	3.94	3.28	3.11	2.78	2.96
Al_2O_3	14.13	19.27	16.34	18.39	15.15	-	-	-	15.48	17.59	14.92	15.50	16.84	13.58	15.67	12.89	16.92	14.88	18.95	-	-	13.70	15.50	13.46	16.97	15.95
Fe_2O_3	1.81	2.70	1.18	1.13	1.08	-	-	-	2.82	2.00	3.69	3.00	3.08	0.65	2.21	0.44	1.46	2.23	1.51	-	-	19.82	3.65	3.39	2.38	19.23
FeO	13.28	6.65	9.81	7.71	9.71	-	-	-	12.70	9.86	12.43	8.71	9.60	13.03	12.54	15.55	9.86	10.08	10.43	-	-	-	9.70	8.73	8.21	-
MnO	0.17	0.13	0.19	0.13	0.14	-	-	-	0.39	0.17	0.32	0.19	0.17	0.24	0.20	0.30	0.18	0.20	0.17	-	-	0.36	0.19	0.20	0.16	0.24
MgO	4.37	4.40	6.36	5.09	5.56	-	-	-	7.27	5.07	6.08	6.59	5.10	5.00	6.63	4.95	4.74	5.31	5.06	-	-	7.07	4.30	5.90	3.60	6.75
CaO	5.74	9.71	5.86	8.72	8.54	-	-	-	5.66	7.11	8.30	9.42	9.43	8.70	7.77	7.05	8.94	11.14	1.84	-	-	4.36	8.95	7.99	7.82	0.64
Na_2O	4.11	3.93	4.71	3.07	3.51	-	-	-	3.73	4.44	4.33	3.66	4.88	4.16	4.55	4.09	5.12	3.67	5.76	-	-	3.20	3.73	4.11	4.76	2.60
K_2O	0.61	0.00	0.00	0.82	0.10	0.58	0.43	1.18	0.20	0.21	0.43	0.29	0.22	0.13	0.36	0.19	0.14	0.53	0.53	0.20	1.33	0.82	1.07	0.03	1.56	0.02
P_2O_5	1.52	0.30	0.38	0.30	0.35	0.70	0.70	0.65	0.91	0.59	0.68	0.74	0.66	1.23	0.79	1.43	0.60	1.20	0.40	0.42	0.39	0.71	0.73	0.53	1.17	0.67
H_2O+	2.91	3.07	4.04	3.95	4.50	-	-	-	3.65	1.78	2.56	2.08	2.29	1.50	1.99	1.56	1.46	1.12	3.90	-	-	3.73	2.17	2.33	2.57	5.61
CO_2	2.63	0.62	2.50	3.23	4.39	-	-	-	-	-	-	-	-	-	-	-	-	-	1.57	-	-	-	0.03	0.06	1.88	-
Total	99.70	99.75	99.84	99.00	99.73	-	-	-	100.00	100.00	100.09	100.00	100.00	100.00	100.00	100.00	100.10	99.99	99.73	-	-	99.99	99.67	100.34	100.04	99.99
Ba ppm	659	180	66	1593	85	-	-	-	207	245	280	130	160	155	240	540	140	672	484	-	-	-	336	27	611	-
Cr	6	63	53	62	78	-	-	-	5	15	5	95	10	-	5	-	5	25	76	-	-	-	51	54	8	-
Nb	51	22	29	20	24	63	27	36	50	32	41	32	36	59	36	48	37	56	29	56	26	52	35	39	31	24
Ni	11	34	32	31	34	-	-	-	-	36	-	50	28	-	12	-	36	24	61	-	-	-	22	28	310	-
Rb	21	-	-	29	6	-	-	-	2	18	13	7	5	9	17	4	8	23	20	-	-	-	33	5	30	-
Sr	444	788	325	552	543	-	-	-	801	655	912	1470	815	667	908	428	878	2040	269	-	-	-	861	483	633	-
Th	3.94	1.67	2.31	1.33	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	6	-	-	-
V	179	196	263	222	269	-	-	-	332	342	495	200	197	205	440	302	287	227	236	-	-	-	233	317	139	-
Y	37	17	20	16	18	30	35	26	38	21	16	24	19	33	18	35	20	19	23	21	24	41	29	31	28	26
Zr	278	111	144	106	122	260	131	229	255	240	270	190	223	203	270	240	205	175	162	247	137	325	189	251	160	195
La	46.63	17.70	20.09	15.44	20.50	28	40	23	58	38	39	40	37	60	33	52	36	51	23.61	18	27	32	34.35	36.48	33.47	22
Ce	97.37	36.51	43.33	34.45	50.02	76	80	60	122	79	77	65	68	156	89	139	72	106	55.62	57	46	79	89.37	93.54	85.07	74
Nd	57.20	19.97	24.16	18.72	20.32	-	-	-	-	-	-	-	-	-	-	-	-	-	23.71	-	-	-	37.97	34.16	41.49	-
Sm	12.74	4.96	5.69	4.89	5.23	-	-	-	-	-	-	-	-	-	-	-	-	-	6.03	-	-	-	9.42	8.31	10.03	-
Eu	4.37	1.72	2.07	1.71	1.88	-	-	-	-	-	-	-	-	-	-	-	-	-	1.90	-	-	-	3.70	2.72	3.97	-
Gd	11.87	4.92	5.84	4.43	4.97	-	-	-	-	-	-	-	-	-	-	-	-	-	5.66	-	-	-	8.90	7.86	9.11	-
Dy	8.48	3.46	4.27	3.25	3.50	-	-	-	-	-	-	-	-	-	-	-	-	-	4.51	-	-	-	6.21	5.82	6.27	-
Er	3.46	1.58	1.78	1.50	1.74	-	-	-	-	-	-	-	-	-	-	-	-	-	2.36	-	-	-	2.79	2.95	2.71	-
Yb	2.93	1.16	1.47	1.11	1.34	-	-	-	-	-	-	-	-	-	-	-	-	-	1.96	-	-	-	2.14	2.56	1.99	-
Lu	0.42	0.19	0.23	0.19	0.37	-	-	-	-	-	-	-	-	-	-	-	-	-	0.41	-	-	-	0.41	0.60	0.39	-
Sum. REE	245.47	92.17	108.93	85.69	109.97	-	-	-	-	-	-	-	-	-	-	-	-	-	125.77	-	-	-	195.26	195.00	194.50	-
Zr/TiO_2	0.0089	0.0047	0.0048	0.0046	0.0043	0.0090	0.0057	0.0089	0.0088	0.0102	0.0080	0.0101	0.0098	0.0063	0.0086	0.0062	0.0095	0.0075	0.0071	0.0103	0.0058	0.0082	0.0058	0.0081	0.0058	0.0066
Ti/Y	504	829	899	858	926	578	394	593	459	671	1266	467	716	581	1046	661	648	735	592	682	585	576	678	601	595	683
Nb/Y	1.38	1.29	1.45	1.25	1.33	2.10	0.77	1.38	1.32	1.52	2.56	1.33	1.89	1.79	2.00	1.37	1.85	2.95	1.26	2.67	1.08	1.27	1.21	1.26	1.11	0.92
Zr/Nb	5.45	5.04	4.96	5.30	5.08	4.13	4.85	6.36	5.10	7.50	6.58	5.94	6.19	3.44	7.50	5.00	5.54	3.12	5.58	4.41	5.27	6.25	5.40	6.43	5.16	8.12
Ce/Y	2.63	2.15	2.17	2.15	2.78	2.53	2.29	2.31	3.21	3.76	4.81	2.70	3.58	4.73	4.94	4.97	3.60	5.58	2.42	2.71	1.92	1.93	3.08	3.02	3.04	2.85

Tab. 1 - Representative chemical analyses of the Hercynian cycle WP metabasaltic rocks in Sardinia. a) INZ: Internal Zone and Nappes of relative internal provenance; b) ENZ: External Zones and External Nappe Zones. (*) new data (present work; all the analyses were carried out at the CRPG, CNRS, Vandoeuvre les Nancy, France); other data come from: Di Simplicio et al. (1975), Ricci & Sabatini (1973, 1978), Beccaluva et al. (1981), Memmi et al. (1983).

Tab. 1b	ENZ															
	UPPER ORDOVICIAN								LOWER CARBONIFEROUS							
	Gerrei		Iglesiente						Sarrabus							
Type	1	1	1	1	1	1	1	1	3	3	3	3	3	3	3	3
Sample	Mb 53*	Mb 54*	SI 1	SI 4	SI 8	DS 19	DS 25	DS 17	Mb29*	Mb30*	Mb56*	Mb57*	Mb57*	Mb57*	Mb58*	M58b*
SiO2	48.01	43.18	47.38	41.85	45.32	38.97	38.11	45.23	38.15	46.79	46.02	50.51	46.16	51.28	44.83	45.66
TiO2	1.73	1.17	2.09	3.11	1.32	1.76	2.36	1.28	1.69	1.73	2.19	1.22	1.37	1.46	2.44	2.31
Al2O3	16.91	13.76	15.00	15.11	15.91	15.44	14.78	17.14	13.54	13.58	16.16	16.38	16.87	14.65	17.46	16.69
Fe2O3	1.48	0.97	7.16	10.69	10.44	2.93	7.68	2.44	0.95	1.11	1.65	1.32	1.81	1.26	1.24	1.63
FeO	6.83	5.40	-	-	-	6.80	9.77	6.37	7.97	6.63	10.11	5.62	8.90	5.85	10.27	8.94
MnO	0.04	0.10	-	-	-	0.20	0.29	0.14	0.29	0.15	0.24	0.17	0.22	0.20	0.28	0.28
MgO	9.39	9.81	3.84	5.85	8.77	9.20	5.01	10.64	5.06	9.70	7.38	8.93	9.90	8.19	10.28	8.32
CaO	2.97	9.89	19.37	18.08	14.26	9.60	16.00	8.25	13.97	6.75	8.88	10.52	6.82	11.57	5.25	11.16
Na2O	3.80	2.72	1.47	1.08	0.89	2.08	1.03	3.08	4.16	3.35	2.73	3.97	3.99	4.07	3.56	3.48
K2O	0.87	0.19	1.60	1.85	0.81	0.83	1.28	1.06	0.04	0.05	1.99	0.05	0.24	0.20	0.17	0.06
P2O5	0.35	0.28	0.32	0.40	0.24	0.56	0.49	0.38	0.43	0.43	0.32	0.16	0.17	0.17	0.43	0.40
H20+	5.48	5.54	1.75	1.97	2.02	11.63	3.18	3.98	2.97	5.70	1.56	0.72	3.19	0.99	3.45	0.91
CO2	1.89	6.76	-	-	-	-	-	-	10.50	4.03	0.06	0.07	0.08	0.09	0.07	0.08
Total	99.75	99.77	99.98	99.99	99.98	100.00	99.98	99.99	99.72	100.00	99.29	99.64	99.72	99.98	99.73	99.92
Ba ppm	343	172	337	717	149	630	354	670	73	124	579	105	167	161	40	55
Cr	283	418	549	536	974	300	-	214	353	554	312	1429	515	892	161	167
Nb	48	40	31	33	37	55	31	19	36	40	20	5	5	6	40	38
Ni	157	151	58	325	295	138	165	139	303	376	138	96	170	71	112	95
Rb	53	6	76	47	35	65	47	72	5	-	90	7	16	10	14	5
Sr	55	112	264	108	164	226	247	464	156	255	240	397	450	471	281	457
Th	-	-	-	-	-	-	-	-	7	6	-	5	-	-	-	-
V	222	208	142	221	236	208	295	200	157	145	250	245	232	269	271	238
Y	15	19	27	24	20	28	36	26	19	21	25	23	25	27	27	28
Zr	125	88	131	156	135	125	149	74	117	131	139	82	93	94	182	177
La	27.88	25.18	18	25	30	42	32	22	23.78	29.82	14.87	5.90	7.18	6.02	24.27	24.43
Ce	49.08	47.46	17	47	43	64	51	29	66.04	76.56	33.09	15.68	17.32	15.46	50.81	50.32
Nd	22.90	20.59	-	-	-	-	-	-	22.34	26.25	21.62	9.86	11.94	11.97	27.39	27.00
Sm	5.06	4.27	-	-	-	-	-	-	5.69	6.22	6.18	3.69	4.06	4.09	6.92	6.60
Eu	1.44	1.31	-	-	-	-	-	-	1.66	1.99	1.84	1.24	1.14	1.45	2.10	2.31
Gd	4.49	3.90	-	-	-	-	-	-	5.30	5.62	6.29	3.68	4.31	5.12	6.35	6.54
Dy	3.58	3.20	-	-	-	-	-	-	3.68	3.97	5.36	3.87	4.46	4.61	5.42	5.41
Er	1.87	1.71	-	-	-	-	-	-	1.86	2.02	2.91	2.00	2.27	2.49	2.53	2.70
Yb	1.93	1.77	-	-	-	-	-	-	1.51	1.72	2.62	2.05	2.44	2.51	2.47	2.53
Lu	0.34	0.30	-	-	-	-	-	-	0.35	0.35	0.46	0.28	0.34	0.45	0.33	0.44
Sum. REE	118.57	109.69	-	-	-	-	-	-	132.21	154.52	95.24	48.25	55.46	54.17	128.59	128.28
Zr/TiO2	0.0072	0.0075	0.0063	0.0050	0.0102	0.0071	0.0063	0.0058	0.0069	0.0076	0.0063	0.0067	0.0068	0.0064	0.0075	0.0077
Ti/Y	691	369	464	777	396	377	393	295	533	494	525	318	329	324	540	495
Nb/Y	3.20	2.10	1.15	1.38	1.85	1.96	0.86	0.73	1.89	1.90	0.80	0.22	0.20	0.22	1.48	1.36
Nb/Y	3.20	2.10	1.15	1.38	1.85	1.96	0.86	0.73	1.89	1.90	0.80	0.22	0.20	0.22	1.48	1.36
Zr/Nb	2.60	2.20	4.23	4.73	3.65	2.27	4.81	3.89	3.25	3.28	6.95	16.40	18.60	15.67	4.55	4.66
Ce/Y	3.27	2.50	0.63	1.96	2.15	2.29	1.42	1.12	3.48	3.65	1.32	0.68	0.69	0.57	1.88	1.80

Tab. 1 -(continued)

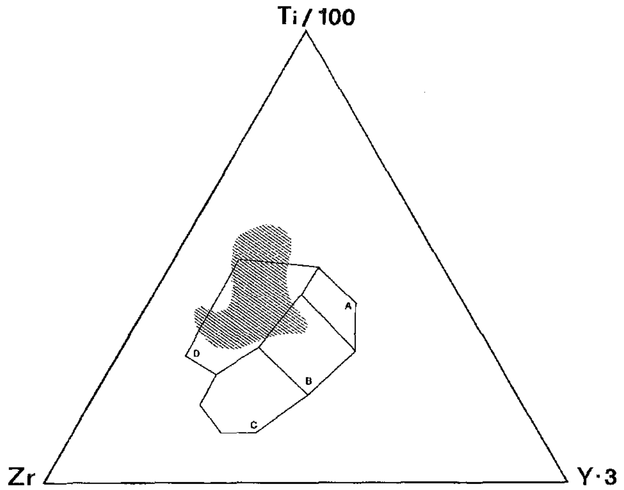

Fig. 3 - Ti-Zr-Y diagram (after Pearce & Cann, 1973) for the WP pre-Hercynian metabasic rocks in Sardinia (shaded area). Field: A+B= Low-K tholeiites; B+C= Calc-alkali basalts; B= Ocean floor basalts; D= Within-Plate basalts.

2.3.2. *Group B (Lower Carboniferous)*

The group B metabasalts show quite different compositional ranges for REE (La/Yb_n 1.62 - 11.70, La/Sm_n 0.93 - 3.02, Gd/Yb_n 1.43 - 3.70 ; Fig. 5 a,b)with occasionally more pronounced Eu anomalies (Eu/Eu* 0.83 - 1.27) and for Ti/Y (318 - 678), Nb/Y (0.20 - 1.90) and Zr/Nb (3.25 - 18.60) ratios, owing to the presence, among them, of some samples from Sarrabus showing a depleted signature (see also Fig. 6b).

2.4. *Inz Basalts vs. Enz Basalts.*

The most releavant features about the Paleozoic WP magmatism in Sardinia rise up when directly comparing the chemical characters of INZ metabasalts with those of ENZ metabasalts, disregarding their age.

Some compositional differences are worthy of note.

The INZ rocks frequently (70 %) show a Ce, P and, when data are available, also Sm selective enrichment,

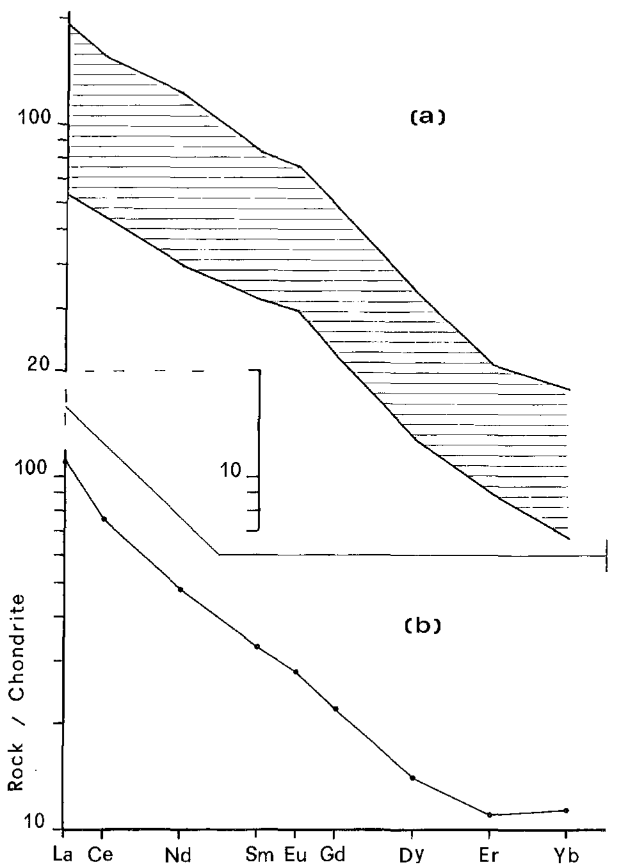

Fig. 4 - Chondrite REE normalized patterns (after Evensen et al;, 1978) of Upper Ordovician WP metabasalts in Sardinia. a): range of REE content found in INZ metabasalts; b): representative REE pattern for ENZ metabasalts (mean on two samples).

generally higher abundances in Zr and Ti and higher Zr/Nb ratios with respect to ENZ rocks.(Fig. 6a).

In the ratio-ratio plot of Fig 7, which describes the distribution of melts derived from different degree of partial fusion in the mantle (Fitton et al., 1988), samples from INZ, as a whole, occupy the higher right hand side of the diagram while samples from ENZ systematically plot in the lower left hand side. The area inside the dashed line represents the Ocean Island Basalt (OIB) field (after Fitton et al., 1988) and show a negative sloop with lower degrees and higher degrees of partial melting in areas with high Ce/Y-low Zr/Nb and low Ce/Y-high Zr/Nb respectively.

Carboniferous metabasalts from Sarrabus (ENZ, solid circles) well follow this pattern possibly indicating that mantle beneath this area underwent different degrees of partial melting. On the other hand the same samples well fit a calculated hyperbolic distribution (according to Langmuir et al., 1978) which points to a possible mixing process between two end members with a MORB-like depleted signature and an OIB-like enriched signature, respectively.

Upper Ordovician samples from ENZ (solid triangles) plot similarly to those from Sarrabus and, even if no clear trend appears, they plot always below the OIB field with a fairly good negative correlation.

INZ samples (open symbols) scatter in Fig. 7 without any appreciable correlation.

3. Discussion

The chemical characters of the Sardinian pre-Hercynian WP metabasalts suggest for the majority of them an origin from LREE and HFS incompatible elements enriched mantle sources.

In spite of their metamorphic character we want to stress the systematic chemical differences, between INZ and ENZ metabasalts.

These differences do not appear to be related to either magmatic differentiation or to crustal contamination, because of the general basaltic character of the considered rocks and the bulk crust composition (Taylor & Mc Lennan,

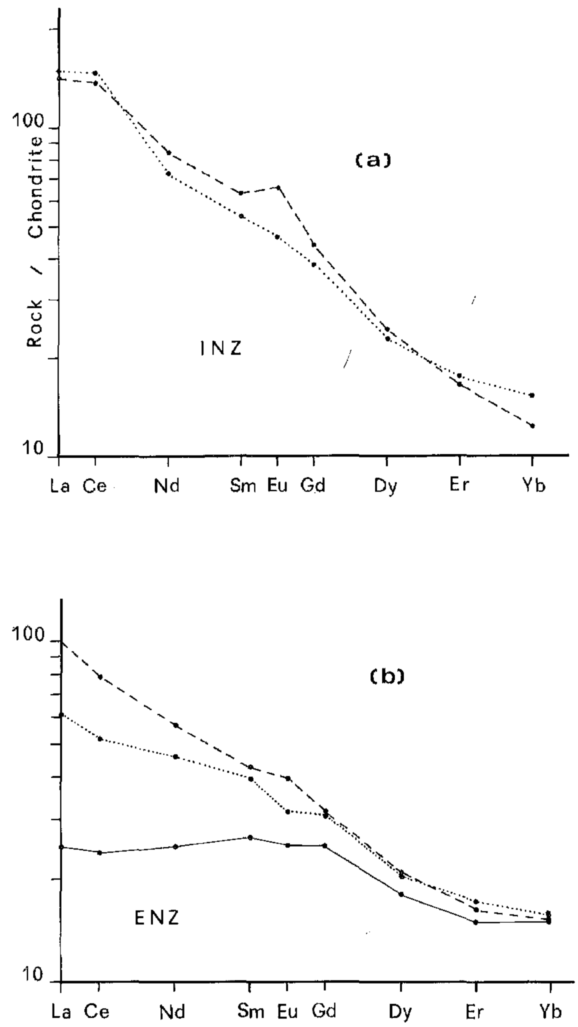

Fig. 5 - Chondrite REE normalized patterns of representative Lower Carboniferous WP metabasalts in Sardinia. a): Nurra (dotted line: sample Mb-6; broken line: sample Mb-5); b): Sarrabus (solid line: sample Mb-57b; dotted line: sample Mb-56; broken line: sample Mb-58b).

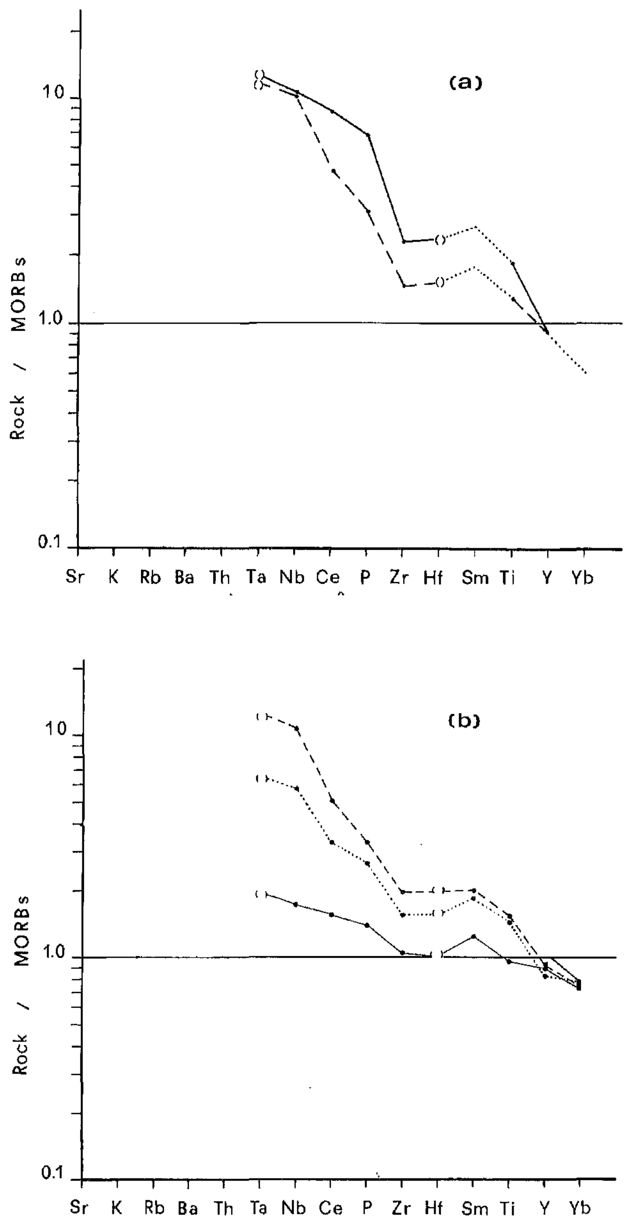

Fig. 6 - a): cumulative spider diagram for INZ and ENZ WP metabasalts of Sardinia; solid line: INZ metabasalts mean (only including 70 % of the samples, see text); broken line: ENZ metabasalts mean (samples Mb-57, Mb-57a and Mb-57b are excluded for clarity). b): spider diagram for some representative Lower Carboniferous samples from Sarrabus; line types as in fig. 5b. Values normalized to MORBs (after Pearce, 1983). Ta and Hf calculated according to Jochum et al. (1986).

1985) that appears unable to play an important role in their chemical composition, as can be seen in Fig. 7. On the other hand, to explain the differences between INZ and ENZ metabasalts, different degrees of partial melting in an homogeneous mantle seem inconsistent with the general positive trend in Fig. 7. Metabasalts with a depleted signature have never been found within INZ terrains neither among samples of group A or group B, possibly indicating for the sub-continental mantle beneath INZ a different nature with respect to mantle beneath ENZ.

The above differences may be tentatively explained in terms of chemical components contributing to the bulk chemistry of basalts. HFSE enrichments are expected from convecting asthenospheric mantle or from deep mantle reservoirs (OIB-type), while LREE enrichments may also develop in sub-continental mantle in response to "...earlier subduction-related magmatism and/or metasomatism" (Fitton et al., 1988; Ormerod et al., 1988; Leat et al., 1988).

Splitting basalts in their components (after Pearce, 1983) INZ basalts seem to show the influence of a subduction process in addition to a dominant OIB-like component. In fact the above selective enrichments resemble those occurring in subduction environments (Pearce, 1983; Pearce et al., 1984) and are not detected in basalts coming from within-plate environments where subduction is not involved (e.g. Afar, Barberi et al., 1975; Antarctica, Kyle, 1981; Azores, Pearce et al., 1984). Ormerod et al. (1988) and Fitton et al. (1988) described a similar situation in Western Great Basin (U.S.A.), where they demonstrated the contribution of a subduction related component in addition to an asthenospheric OIB-like component in the formation of a magmatic suite resulting from the tectonic transition from a supra-subduction zone to a within-plate environment.

The spatial and chronological distribution of the metabasic rocks as a whole (including calc-alkaline and tholeiitic MORB rocks) in Sardinia offer some constraints on interpretation of the chemical character of the WP metabasalts.

Remnants of N-MORB tholeiites in northern Sardinia (Cappelli et al., 1990) testify to the existence of an oceanic crust at least from about 950 Ma (Cappelli et al., 1991) the consumption of which is proposed to have partially occurred beneath the north Gondwanian continental margin in the lower Paleozoic (Carmignani et Al., this volume). The middle Ordovician calc-alkaline volcanics in Sardinia might represent the magmatic arc manifestation of this subduction

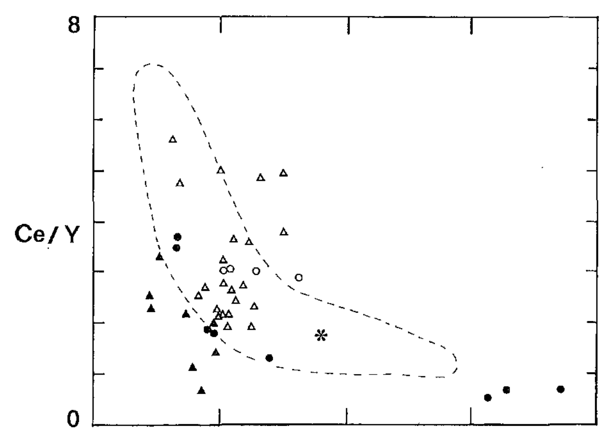

Fig. 7 - Ce/Y vs. Zr/Nb cumulative plot of the Paleozoic WP metabasalts of Sardinia (see text). Triangles: Upper Ordovician metabasalts; circles: Upper Devonian-Lower Carboniferous metabasalts; open symbols: INZ metabasalts; solid symbols: ENZ metabasalts; asterisk: bulk crust.

under the Gondwanian continental margin. Carmignani et al. (this volume) propose, on structural and stratigraphic bases, the reconstrution of the polarity of the arc that allows us to infer a NE to SW sense of subduction. The direction of deepening of the subducting oceanic slab coincides with the distribution of the paleogeographic zones, from internal to external, prior to the Variscan collision and shortening.

Calc-alkaline volcanism ceased in middle-late Ordovician times as a consequence of the cessation of subduction. A ridge-trench collision (Nelson & Forsythe, 1989) is proposed to explain this interruption (Carmignani et al., this volume). A detached slab of oceanic lithosphere continued to sink in the sub-continental asthenosphere while the compressive stress field linked to subduction (Scholz et al., 1971) was released. As consequences, a lithospheric window opened beneath the arc (causing the interruption of arc volcanism in a wide area: "no slab area" after Dickinson & Snyder, 1979) and extension dismembered the arc-back-arc system, respectively (Carmignani et al., this volume), allowing the ascent of sub-lithospheric magmas, coming from convecting upper mantle sources (Ormerod et al., 1988) , through the mantle wedge previously affected by subduction processes.

This scenario might represent a key to interpreting the geochemical differences between INZ and ENZ WP metabasalts.

During subduction, the sub-continental mantle wedge columns beneath INZ and ENZ must have had different thickness (small and large respectively, INZ being close and ENZ far from the inferred Middle Ordovician trench), and possibly accepted subduction components at different degrees of dilution. In this way the INZ basalts might have been derived by interaction of sub-lithospheric magmas with supra-subduction mantle volumes that were more significantly modified by subduction than volumes from which ENZ basalts were derived.

4. Conclusions

The new data presented here, together with those available from the literature, allow us to recognize, in the Variscan segment of Sardinia, two different magmatic suites that show WP affinity and occur in Upper Ordovician and Lower Carboniferous times, respectively.

The older one appears to be related to an extensional regime following B-type subduction possibly linked to the cessation of convergence due to an ocean ridge-continent collision (Nelson & Forsythe, 1989).

No evidence of volcanic activity from the Lower Silurian to the end of the Devonian has been found.

A new phase of crustal extension must have started in the Lower Carboniferous in the external zones and more likely in the Upper Devonian in the internal zone. As suggested by Vai (1982) and Vai & Cocozza (1986) such an extensional phase must be connected to wrench dynamics. The strike-slip events accompanying extension could have been due to the indentation mechanisms resulting from early collisional stages, like the ones proposed for the south-western european Variscides (Matte, 1986). Such a geodynamic model could explain the continental WP basic activity represented by metavolcanics embedded in the Lower Carboniferous deposits of Sarrabus, and by the matagabbros and metadolerites intruded into the Siluro-Devonian black shales of Nurra.

Each suite shows some systematic chemical characters according to their occurrence in the Internal Nappe Zone or in the External Nappe Zone.

The very similar geochemical nature of rocks with different ages that come from the same area (INZ or ENZ), might represent an essentially similar nature of the source region in the mantle. Chemical differences between INZ and ENZ basic rocks can be likely explained by a spatial variation of subduction related metasomatism in the mantle during Lower and Middle-Ordovician time.

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THE CALC-SILICATE MARBLES OF TAMARISPA (NE SARDINIA)

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Abstract

A structural and petrological study was performed of the calc-silicate marbles cropping out in the High-Grade Amphibolite-facies Complex in north-eastern Sardinia. Field and petrographical data have shown that marbles have the same structural features as the surrounding migmatites; the mineral assemblage consists of clinopyroxene + wollastonite + calcite + grossularite ± plagioclase ± quartz which formed in the range 650°<T<<850° C. Thus marbles formed during the same tectono-metamorphic event which formed migmatites.

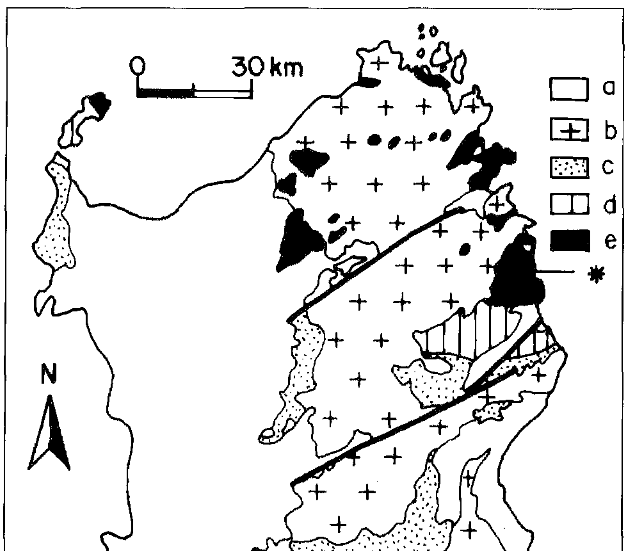

Fig.1 - Axial Zone of Hercynian Sardinia: a) Quaternary, Tertiary and Mesozoic covers; b) Granitoid Complex; c) Green Schist - facies Complex; d) Medium-grade Amphibolite-facies Complex; e)High-Grade Amphibolite-facies Complex; *) Tamarispa area where the calc-silicate marbles crop out.

Key words: *North-eastern Sardinia, High Grade Complex, calc-silicate marbles.*

1. Geological Setting

At Tamarispa area in north - eastern Sardinia, sparse calc-silicate marbles crop out in the High-Grade Amphibolite-facies Complex (Elter & Sarria,1989) (Fig.1), which consists of migmatites and orthogneisses pertaining to the sillimanite + K-feldspar zone (Franceschelli et al., 1982).

The outcrop consists of two elliptical lenses (Fig.2) that stretch northeast-southwest and whose thicknesses and lengths are about 3÷5 m, 10÷20 m respectively. The marble lenses are whitish and contain stretched garnets (size about 2÷20 cm),in a matrix of wollastonite.

The relationships between the lenses and the other lithologies (migmatites and a orthogneiss) can be defined as follows:

— the planar anisotropy of the calc-silicate marbles (Sx schistosity) and the mineralogical -extensional lineations in the Sx plane (Lmx1, Elter & Sarria 1989) are the same as that of the surrounding migmatites and orthogneiss (Fig.3);

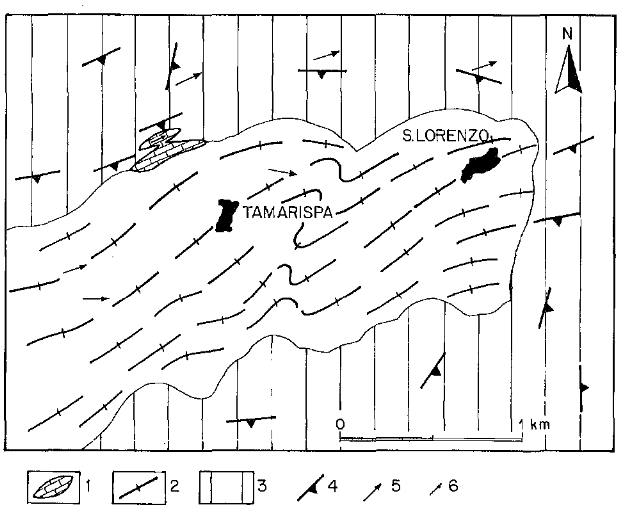

Fig.2 - Geological map of the area where the calc-silicate marbles crop out. 1: Calc-silicate marbles; 2: Granodioritic orthogneiss with Sx schistosity; 3: Gneiss s.l. and stromatic migmatites; 4: Sx schistosity in migmatites. 5: Mineralogical lineations (Lmx1) in the granodioritic orthogneiss and migmatites;6: Mineralogical lineations (Lmx1) in the calc-silicate marbles.

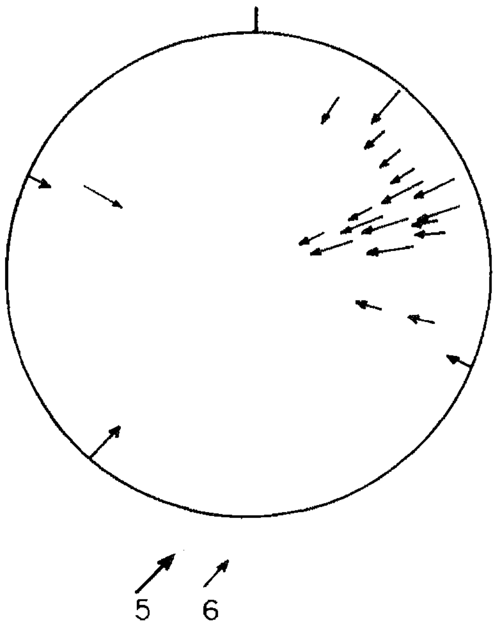

Fig.3 - Stereonet of the mineralogical lineations (Lmx1) in the granodioritic orthogneiss, migmatites (5) and calc-silicate marbles (6). (Schmidt net - lower hemisphere).

—the trend of the stretching lineations in all lithologies ranges from 30° to 85° and dips from 10° to 30° NE;

—the garnets are boudined in the YZ plane (40% maximum extensional rate) and in the XZ plane (20% maximum extensional rate);

—fractures filled with wollastonite can be present.

2. Petrographical Outlines

The matrix of the calc-silicate marbles show a grano-

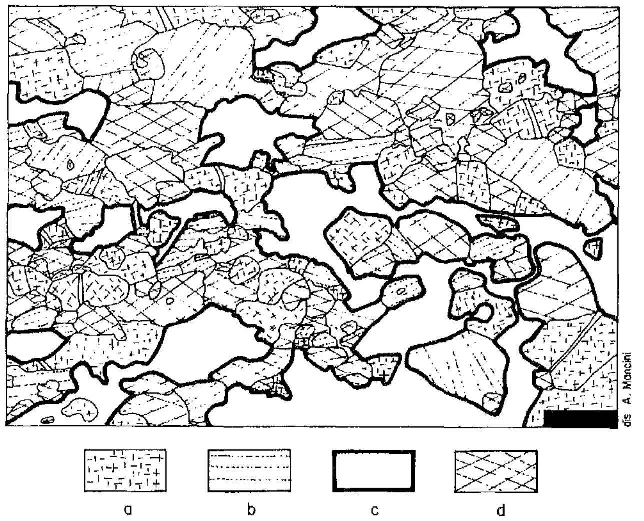

Fig.5 - Sketch of textural features of matrix in the calc-silicate marbles. a: clinopyroxene; b: wollastonite; c: garnet; d: calcite. Scale bar = 1 mm

nematoblastic texture with compositional layering and it consists of salitic clinopyroxene + wollastonite + calcite + grossularite ± plagioclase ± quartz, with sphene as accessory mineral (Fig.4,5). Clinopyroxene and especially wollastonite form unidirectional elongate crystals which define the main schistosity (Sx) (Fig.4). In some cases wollastonite contains small inclusions of pyroxene. Grossularite is both euhedral and anhedral.

Grossularite of different grain size (decimetric) is poikiloblastic, with small salitic clinopyroxene and wollastonite as inclusions, the latter are partly aligned in a direction forming a millipede structure (Bell & Rubenach, 1980) (Fig.6).

From the textural relationships the growth of decimetric garnet is referred as syn-late to the enclosed alignment but previous to the main schistosity Sx.

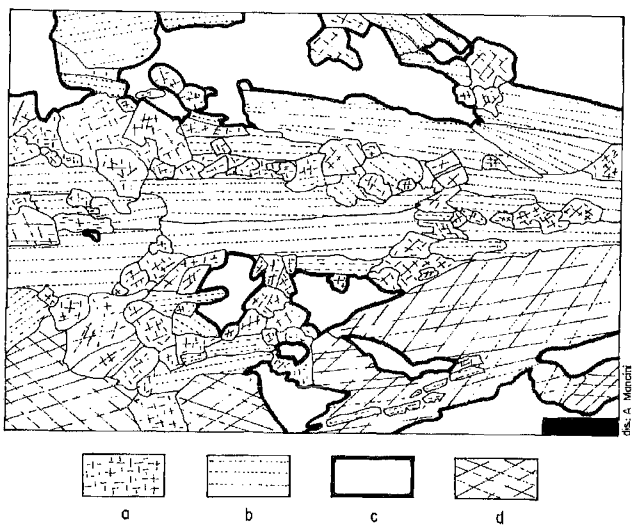

Fig.4 - Sketch of textural features of matrix in the calc-silicate marbles. a: clinopyroxene; b: wollastonite; c: garnet; d: calcite. Scale bar = 1 mm

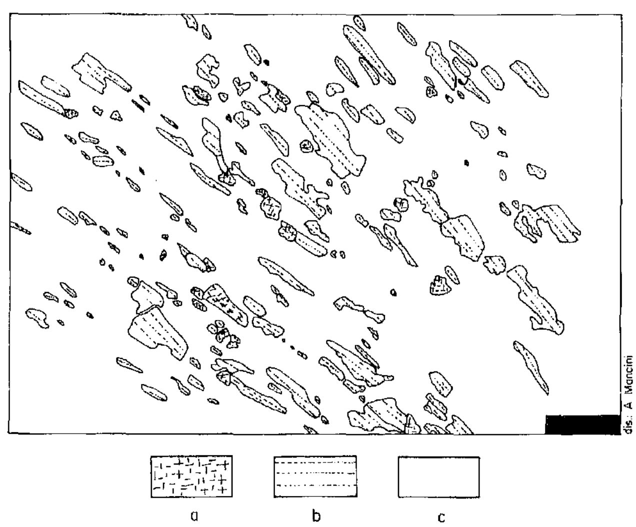

Fig.6 - Sketch of the millipede structure in the garnet. a: clinopyroxene; b: wollastonite; c: garnet. Scale bar = 1 mm

3. Mineral Chemistry

Clinopyroxene is a homogeneous salite (Tab.1) with X_{Mg}(=Mg/Mg+Fe) ranging from 0.69-0.67 in the matrix to slightly higher (0.71-0.69) in the inclusions within garnet.

Garnet is a homogeneous grossularite (Tab.1) with X_{Ca}(=Ca-1.5(Fe^{3+} +Ti) /Fe^{2+}+Mg+Ca+Mn) ranging from 0.896-0.883 in the matrix to slightly lower in the decimetric poikiloblasts (X_{Ca} = 0.835-0.851); in both the textural type of garnets spessartine is absent and pyrope content is very low, X_{Mg}(= Mg/Mg+Fe+Mn+Ca) is max 0.007. The almandine content $X_{Fe}{}^{2+}$(=Fe^{2+}/Mg+Fe^{2+}+Mn+Ca) ranges from 0.034-0.027 in the matrix to 0.038-0.023 in the poikiloblastic garnets. Finally, the andradite content is higher in the poikiloblasts $X_{Fe}{}^{3+}$= Fe^{3+}/ Fe^{3+} +Al+Ti = 0.089-0.095 than in the garnet of the matrix (0.050-0.069).

4. Phase Relations

In the calc-silicate-bearing assemblage, salite is the only mineral that contains significant quantities of FeO and MgO, and sphene is the only TiO_2-bearing phase. Thus, in first approximation the stability of the remaining phases can be modeled in the system CaO-Al_2O_3-SiO_2-CO_2-H_2O. Figure 7 is a qualitative diagram T-XCO_2 from Ellis (1978) calculated for pressures of 4 Kbar.

The diagram indicates that grossularite, wollastonite, anorthite, calcite and fluid coexist in the field constrained by the following reactions:

Gr+Q=An+Wo (4)
An+Cc=Me (1)
An+Cc+Wo=Gr+V (3)
An+Cc=Gr+Co+V. (5)

Grtm	Grtm rim	Grtp core	Grtp rim	Cpxe core	Cpxe rim	Cpxm core	
SiO_2	38.94	39.15	39.04	38.93	52.33	52.88	52.48
TiO_2	.52	.64	.99	1.07	--------	--------	--------
Al_2O_3	20.96	21.29	20.00	20.00	--------	--------	--------
Fe_2O_3	2.41	1.75	3.31	3.12	--------	--------	--------
FeO	1.27	1.59	1.09	1.79	10.06	9.54	10.45
MgO	.16	.07	.07	.19	12.33	12.94	12.26
CaO	35.67	35.64	35.93	35.11	24.25	24.37	24.16
MnO	--------	--------	--------	--------	--------	--------	--------
Na_2O	--------	--------	--------	--------	--------	--------	--------
Anhydrous total	99.93	100.13	100.43	100.21	98.97	99.73	99.35
Si	2.957	2.964	2.964	2.965	1.994	1.992	1.995
Al	1.875	1.899	1.790	1.795	-------	--------	--------
Ti	.030	.037	.056	.061	--------	--------	--------
Fe^{3+}	.138	.100	.189	.179	--------	-------	--------
Fe^{2+}	.080	.101	.069	.114	.320	.300	332
Mg	.018	.008	.008	.021	.700	.726	.694
Ca	2.902	2.891	2.923	2.864	.990	.983	.983
Mn	--------	--------	--------	--------	-------	-------	--------
Na	--------	--------	--------	--------	--------	--------	--------
X_{Mg}	0.184	0.073	0.104	0.155	0.686	0.708	0.676
Grs	0.883	0.896	0.851	0.835			
Alm	0.027	0.034	0.023	0.038			
Prp	0.006	0.003	0.003	0.007			
Sps	--------	--------	-------	--------			
Adr	0.069	0.050	0.095	0.089			

m=matrix; p=porphyroblast; e=enclosed in garnet; X_{Mg}= Mg+Fe^{2+}; Grs= grossularite; Alm= almandine; Prp= pyrope; Sps= spessartine; Adr=andradite

Tab.1 - Selected SEM/EDS analyses for garnet and clinopyroxene. Structural formulae are calculated on the basis of 12 oxygens for garnet and 6 for clinopyroxene.

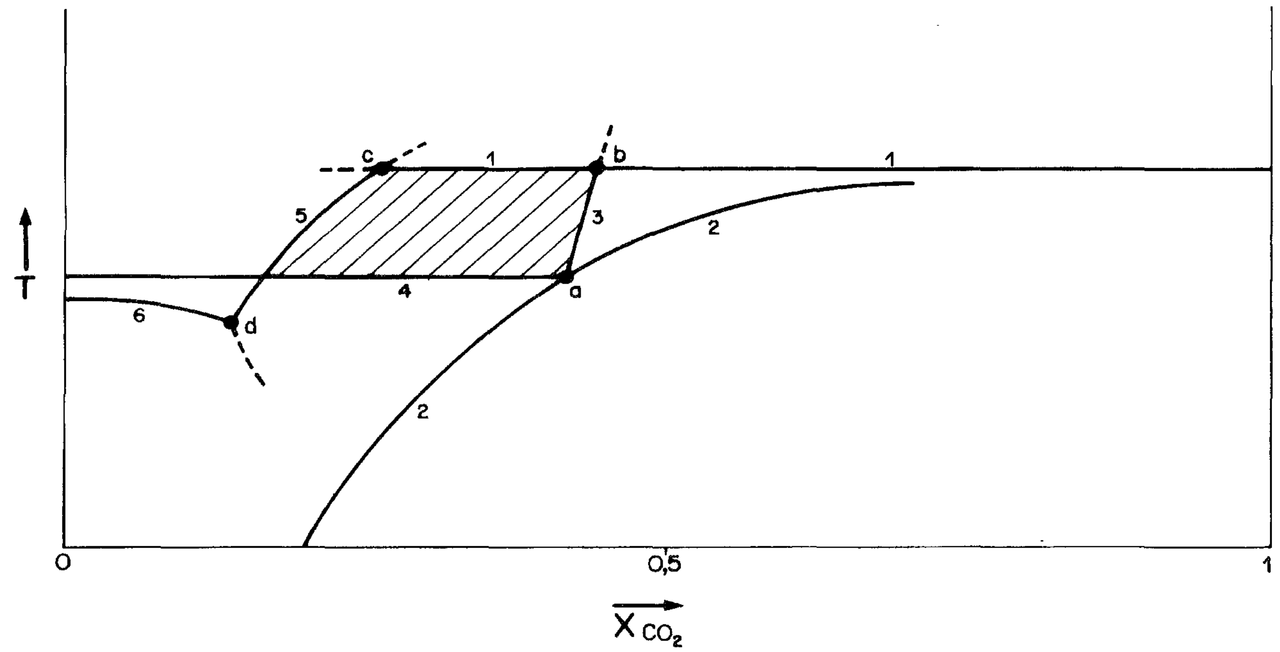

Fig.7 - T-XCO_2 diagram after Ellis (1978) showing phase relations in the system CaO-Al_2O_3-SiO_2-H_2O-CO_2 at 4Kb. (a), (b), (c), (d), invariant points. (1) An+Cc=Me; (2) Cc+Q=Wo+V; (3) An+Cc+Q=Gr+V; (5) Gr+Q=An+Wo; (5) Cc+An=Gr+Co+V; (6) Zo=Gr+An+Co+V. An=anorthite; Cc=calcite; Me=meionite; Q=quartz; Wo=wollastonite; Gr=grossularite; Co=corundum; Zo=zoisite; V=vapour.

The temperature of the invariant point (a) in Figure 7, at which the phases studied coexist, was calculated experimentally by Gordon & Greenwood (1971) and A.B. Thompson (1971) for P=2Kb; the former give a temperature of 590° C while the latter give a temperature of 630°C. The temperature of the invariant point increases with increasing pressure. Winkler (1979), reports an increase of approximately 150° C at 6 Kb (730° C). Moreover, according to Huckenholz et al.(1981) the slope of reaction (4) can be described by the equation of T(°)= 483+48P (Kb), so at P=2Kb T=579° C and at P=6Kb T=771° C.

The absence of scapolite constrains the maximum temperature; infact, according to Goldsmith & Newton (1977), the reaction An+Cc=Me is virtually pressure-independent and is realized in the 850°-875° C range.

The pressures estimated for the migmatites of the same area are around 4 Kb (Franceschelli et al. 1989), so using this pressure and the above considerations, the stippled area of figure 7 is constrained between 651°-699° C, up to 875° C. However, because plagioclase was not pure anorthite (An ≈ 70%), the maximum temperature was lower than 850°C. In conclusion the most probable range of temperatures in which the calc-silicate marbles formed is 650°< T << 850° C. Figure 7 can be used to evaluate XCO_2: Hoschek (1974), obtained XCO_2=0.13 for the reaction An+Wo+Cc=Gr+CO_2 (3) and XCO_2=0.06 for the reaction An+Cc=Gr+Co+CO_2 (5) under temperature and pressure conditions of T=750° C and P=4Kbar . Thus, XCO_2 probably ranges between 0.06 and 0.13.

5. Conclusions

The field, textural and petrological data indicate that the schistosity Sx is the same within marbles, migmatites and orthogneisses, and that the boudined deformation of poikiloblastic garnet is sin-kinematic to Sx.

The existence of millipede structure of poikiloblastic garnet and the coexistence of Wo+Grs+An+Cpx was also observed. The estimated temperatures are 650°<T<<850°C.

This data suggests that marbles formed together with the surrounding metamorphics (migmatites and the orthogneiss) in a continuous tectono-metamorphic event.

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PRESENCE OF CONTINENTAL DEPOSITS UNDERLYING THE SARDINIAN EOCENE MARINE SEQUENCE

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Abstract

Ferruginous unfossiliferous deposits outcropping at the base of marine Eocene successions are described. At the outcrop, they are marked by lack of bedding and by erosive boundaries. Detailed sedimentological and micromorphological analysis have pointed out typical quartz grains dissolution morphologies, the size reduction of quartz, the presence of kaolinite, mottles, nodules of Fe and coatings of Fe and clay around the clasts.

These sediments have been identified as paleosols evolved in hot wet climate.

Key Words: *Continental deposits, Paleogene, Southern Sardinia.*

1. Introduction

The first references in literature to the presence of "paleosols" at the base of Eocene successions are to be found in Vardabasso (1962), who recognizes evidence of "paleosols" on the southern border of M. Cardiga (Fig. 1). Pomesano Cherchi (1962) recognized evidence of more or less weathered "paleosols" at the base of the southern slope of the Eocene succession of M. Ixi. At M. Genis (Is Cantonis) Pomesano Cherchi (1964) describes unfossiliferous ferruginous sandstones probably of continental origin at the base of the Cuisian succession. Also Cocozza et al., (1974) refer to the occurrence of continental facies sediments in southern Sardinia within the Eocene sedimentary successions. Finally Dieni et al. (1979) have pointed out periods of subaerial exposure in the Orosei sector from a study of carbonatic clasts contained in the Post-Cuisian conglomerates of Cuccuru'e Flores. Particularly the first Paleogenic deposits, that are found only at Orosei, are made up of limestones containing Dasycladales (attributed to the Danian-Montian) and modified into pseudo-breccias by the settling of colonies of Paronipora (ex - Microcodium; Cherchi & Schroeder, 1988) during the continental phase between the Montian and the Thanetian. In the Orosei sector the Authors also show another two phases of subaerial diagenesis between the Thanetian and the Middle Ilerdian and between the Middle Ilerdian and the Cuisian. Also the pseudo-breccias of Orroli and Siurgus Donigala (south - eastern Sardinia) and in the Sulcis-Iglesiente (south - western Sardinia) can be referred to a phase of continentality previous to the deposit of overlying marine Ilerdian sediments (Murru, under study).

2. Description of the outcrops of the Monte Maraconis and Sant'Andrea Frius

During a new phase of research on the Lower Tertiary in Southern Sardinia, two outcroppings (Fig. 1) presenting unfossiliferous ferruginous deposits at the base of the Eocene successions have been recognized.

The following is a brief description of the lithostratigraphical successions from bottom to top:

a) Monte Maraconis, Lipparini 1938; Ferrara et al., 1991: 1 - Paleozoic bed rock; 2 - Matrix supported, unorganized granule-pebble conglomerates.They generally consist of quartz clasts with minor amounts of schist clasts embedded in a ferruginous clay matrix. Almost all clasts are subrounded (60 cm); 3 - Clast supported, moderately sorted, lense shaped, pebble conglomerates with clast imbrication. They generally consist of rounded schist clasts with minor amounts of subrounded quartz clasts and fragments of the underlying level. Matrix is a mixture of quartz silt-sand. The contact between the two conglomeratic levels is an undulating erosional surface (3,80 m); 4 - Siltstones, marly claystones and fine sandstones with gastropods, ostracods and, presumably, Ilerdian nummulitids (9,50 m); 5 - Conglomeratic sandstones (2,50 m).

b) Sant'Andrea Frius, Pecorini & Pomesano Cherchi 1969; Matteucci & Murru 1988; Ferrara et al., 1991; 1 - Paleozoic bed-rock; 2 - Red unfossiliferous claystones (12 m); 3 - Unfossiliferous massive ferruginous coarse sandstones (3 m); the contact with the underlying claystones is erosive; 4 - Claystones, marly claystones, marls, limestones with ostreids, charopyhtes, ostracods, nummulitids and alveolinids of Upper Thanetian-Ilerdian - Lower Cuisian? age (21 m).

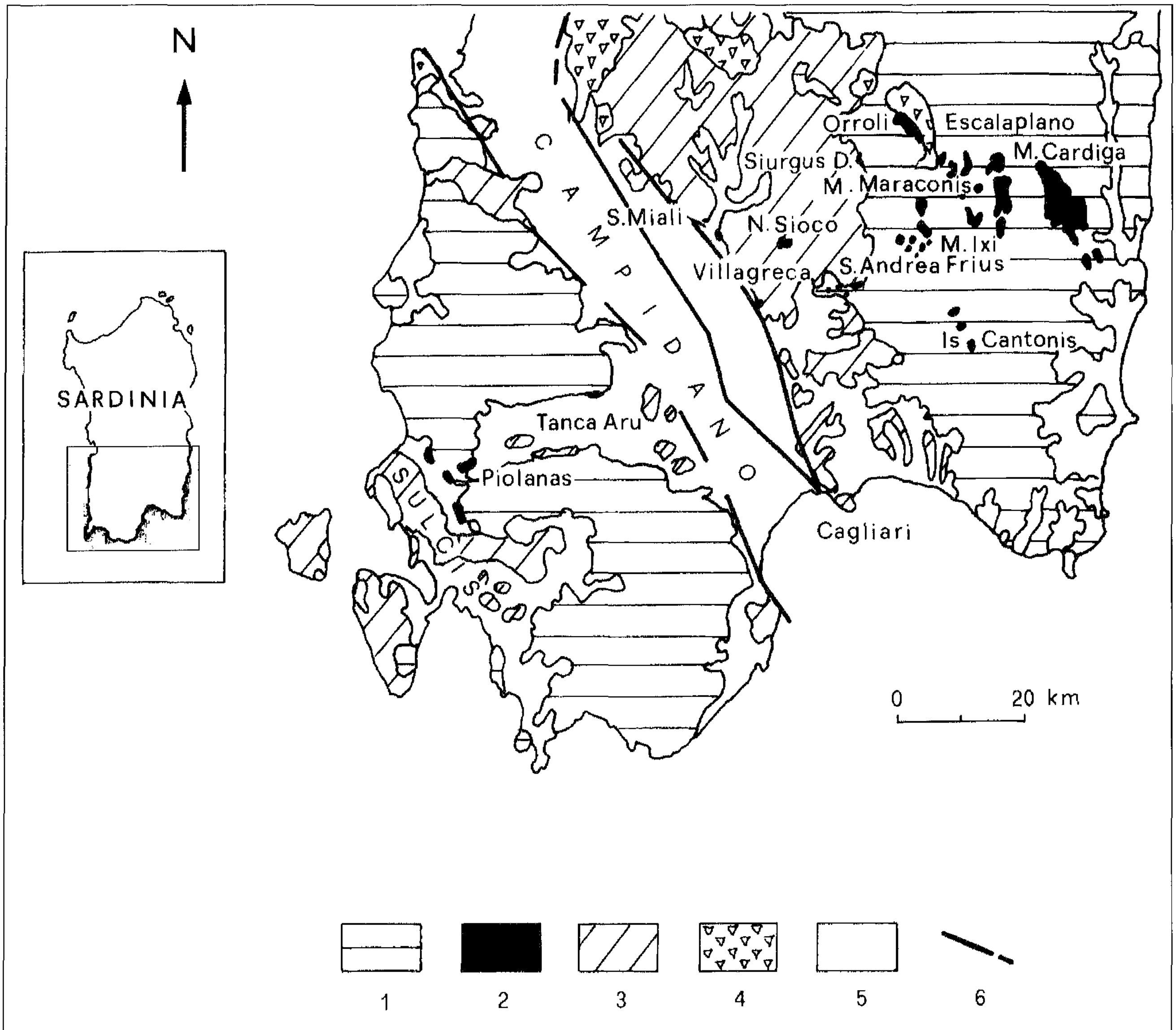

Fig. 1 - Paleozoic-Mesozoic bed-rock; 2 - Eocene outcroppings; 3 - Oligo-Miocene sediments and volcaniclastic deposits; 4 - Plio-Quaternary basalts; 5 - Quaternary sediments; 6 - Faults.

3. Results and Discussion

The outcropping Monte Maraconis (a-2) ferruginous conglomerates show some areas of weak red colour (lOR4/4 Munsell) and others of grey colours (5Y6/1 Munsell). The texture is clayey. The X-ray analysis has pointed out presence of haematite and the clay minerals consist mainly of kaolinite. In the coarse fractions, sedimentological and micropedological analysis (BULLOCK et al., 1985) have identified presence of subangular and embayed subrounded quartz grains with fractures filled with Fe sesquioxides, fine sandstone granules and lithorelicts with completely altered femic minerals. Rounded embayed quartz clasts display a variety of morphologies: "T" shaped reentrants or curvilinear reentrants and broad rounded embayments. Many of the grains show also rounded outlines with angular embayments (Fig. 2 - 3). The roundness of these clasts is partly inherited. These deposits came too from the reworking of the quartz conglomerates outcropping at the base of the marine Jurassic successions.

The pedofeatures are represented exclusively by macroscopic mottles which appear as a very dense network of Fe sesquioxidic hypo-coatings. On the basis of these characteristics these sediments can be identified as plinthites, corresponding to a horizon formed in hydromorphous environment by local segregation of iron in hot wet climate (Wood & Perkins, 1976).

The Sant'Andrea Frius (b-3) quartz-ferruginous sandstones are of reddish yellow colour (7.5YR6/8 Munsell) with red mottles (2.5YR4/Munsell), and have a sandy loam texture. X-Ray analyses have pointed out presence of haematite and the clay minerals consist mainly of kaolinite

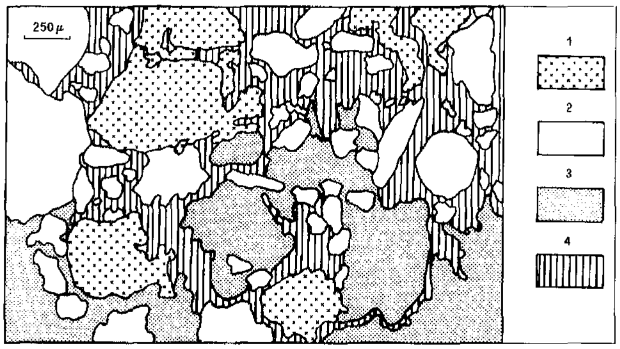

Fig. 2 - Photomicrograph of Monte Maraconis plinthite sample: 1 - Quartz grains with "T" shaped reentrant (typical of lower soil zones); 2 - Quartz grains; 3 - Matrix; 4 - Macroscopic mottles as a very dense network of sesquioxidic hypocoatings.

and subordinately illite (Zachos et al., 1989) By micropedological analysis it has been possible to ascertain that the coarse fraction is made up almost exclusively of fractured subangular and subrounded quartz. The pedofeatures are represented by compound coatings of clay and Fe around quartz grains and voids, nodules of Fe, mottles and hypo-coatings of Fe surrounding pores and grains.

These sediments have bimodal granulometric distribution curves with the primary coarse modes made up of subrounded quartz clasts and the secondary fine modes made up of angular quartz clasts (Fig. 4). The granulometric reduction takes place in situ and can be attributed to quartz solution within the soil environment (Cleary & Conolly 1972). These characteristics suggest that it is an argillic horizon (pedogenetic horizon).

4. Conclusions

In Sardinia, the Cretaceous marine sedimentation, which ended in the Lower Maastrichtian in Eastern Sardinia (Dieni & Massari, 1985), is followed by a long phase of continentality, which extended at least up to the Danian Montian (Dieni et al., 1979). The sediments of this age,

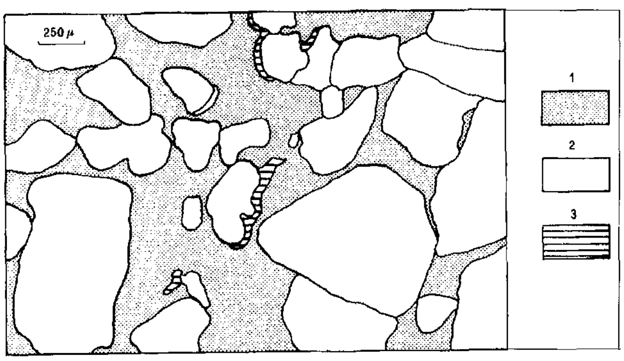

Fig. 3 - Photomicrograph of Sant'Andrea Frius sample: 1 - Matrix; 2 - Quartz grains; 3 - Compound coatings of clay and Fe.

which are found only within Post-Cuisian conglomerates of Cuccuru 'e Flores in the Orosei sector (central-eastern Sardinia), have been modified into pseudo-breccias by the settling of colonies of Paronipora during a period of subaerial exposure between the Montian and the Thanetian. The study of two outcroppings of unfossiliferous ferruginous deposits in southern Sardinia, has allowed to ascertain the presence of Pre-Eocenic paleosols. The ferruginous conglomerates of Monte Maraconis have been identified as plinthite (pedogenetic horizon very rich in iron) and are overlain by levels of nummulitids of probably Ilerdian age. The ferruginous sandstones of Sant'Andrea Frius are considered as paleosols and are covered by a level containing Upper Thanetian Charophyta.

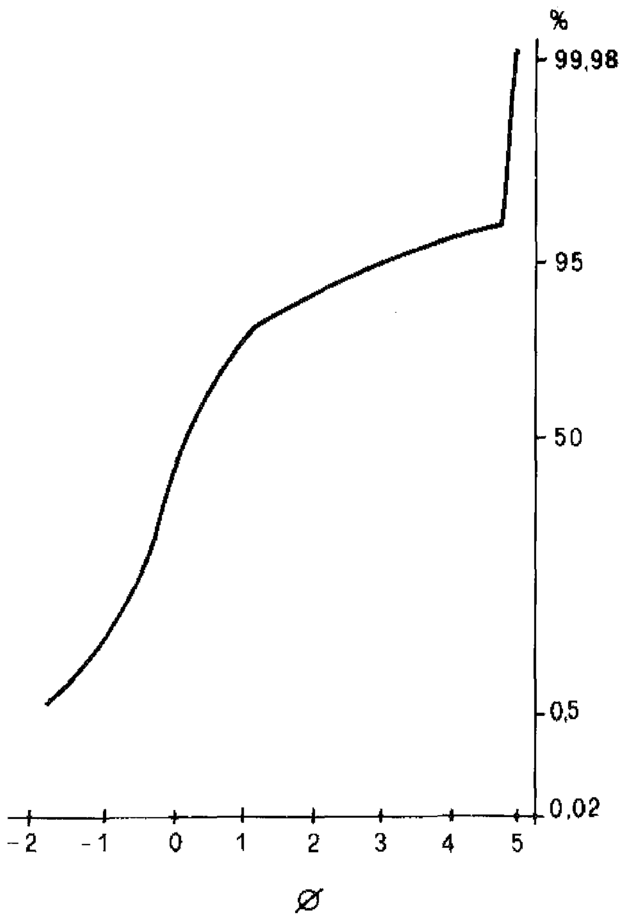

Fig. 4 - Granulometric distribution curve showing the bimodality of Sant'Andrea Frius quartz sandstones.

The dissolution morphologies, the roundness and the size reduction of quartz grains, observed in the samples, have been described by several authors (Crook, 1968; Cleary & Conolly, 1972; Freyssinet et al. 1990) in contemporary soils and paleosols and attributed to dissolution of SiO2 in situ.

During pedogenesis, in hot wet climate, the interaction between the organic and inorganic fractions of soils are sufficently widespread and intense to produce the features observed. The grain shapes are mostly formed along preexisting grain defects (partly healed fractures). The intimate relationships between embayments on quartz grains and matrix may reflect solution of quartz at its contact with clay-humus complexes.

In short the emplacement of the continental deposits that are to be found at the base of the Eocene successions of southern Sardinia can be referred to a continental phase established in the Upper Cretaceous and extended until the Thanetian-Ilerdian.

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ILLITE CRYSTALLINITY IN PELITIC ROCKS FROM THE EXTERNAL AND NAPPE ZONES OF THE HERCYNIAN CHAIN OF SARDINIA (ITALY)

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Abstract

Illite crystallinity was determined in more than 250 pelitic rocks from the Nappe and External zones of the Hercynian chain of Sardinia.

The regional distribution of illite crystallinity values indicates a clear Hercynian metamorphic zonation from anchizone to epizone moving northwards from the External to the Nappe zone.

In the Nappe zone illite crystallinity values also reveal a roughly metamorphic zonation from the lower to the upper tectonic units of the pile of nappes as well as from the front to the root zone in the Gerrei Unit and in the Metamorphic Complex of Barbagia Unit. In the External zone contrasting illite crystallinity values were detected at the Sardic phase unconformity between the Cabitza Formation (0.25 ± 0.01 °$\Delta 2\Theta$) and the overlying "Puddinga" (0.39 ± 0.03 °$\Delta 2\Theta$). The status of illite "crystallinity" in the Cabitza Formation seems to record pre-Hercynian metamorphic effects.

Key words: *Illite crystallinity, pelitic rocks, Paleozoic, Hercynian orogeny ,Sardic phase, Sardinia.*

1. Introduction

Sardinia is a complete section of a segment of the southern European Hercynian chain from the External (south-west Sardinia) ,through the Nappe (central Sardinia) to the Axial (north Sardinia) zones.

The degree of metamorphism slightly changes from the External to the Nappe zones over a horizontal distance of about 150 Km. In the Axial zone the degree of metamorphism rapidly increases from low to high grade towards north-east reaching the migmatite zone. (Franceschelli et al., 1982; 1989).

The low number of data on the incipient metamorphism in the External and Nappe zones lead us to draw up a plan for a systematic survey of illite crystallinity (IC) in pelitic rocks from these two structural sectors of the chain.

This paper is the first report on this survey. It contributes to a more detailed metamorphic Hercynian zonation and sheds some light on the effects of pre-Hercynian metamorphism on the Cambrian - early Ordovician rocks in SW Sardinia. Although the presence of pre-Hercynian deformation was recognized on these rocks a long time ago (Stille, 1939; Vardabasso, 1940; 1956; Arthaud, 1963; Poll and Zwart, 1964; 1965; Poll, 1966; Cocozza, 1980) the evaluation of possible pre-Hercynian metamorphic effects and their importance has been neglected so far.

2. Regional Geology

Detailed information on the stratigraphy and tectonic evolution of the rocks of the External and Nappe zones can be found in Carmignani et al. (1980, 1982, 1986a, b).

A structural sketch map of the Paleozoic basement of south and central Sardinia is shown in Fig.1. A concise scheme of Paleozoic sequences and their tectonic relationships are shown in Fig.2

The Nappe zone consists of six major tectonic units overthrusted from NE to SW onto the External zone (Fig.2). From bottom to top they are: Riu Gruppa, Gerrei, Genn'Argiolas, Meana Sardo, Arburese and the Metamorphic Complex of Barbagia ("Postgotlandiano Auct.", Vai and Cocozza, 1974).

The bottom of the Riu Gruppa Unit is made up of Cambrian-Ordovician metasandstones, intermediate to acidic metavolcanites and metagraywackes (Gattiglio & Oggiano, 1991). The top of the sequence consists of Siluro-Devonian marble and black phyllites (Carmignani et al., 1986b).

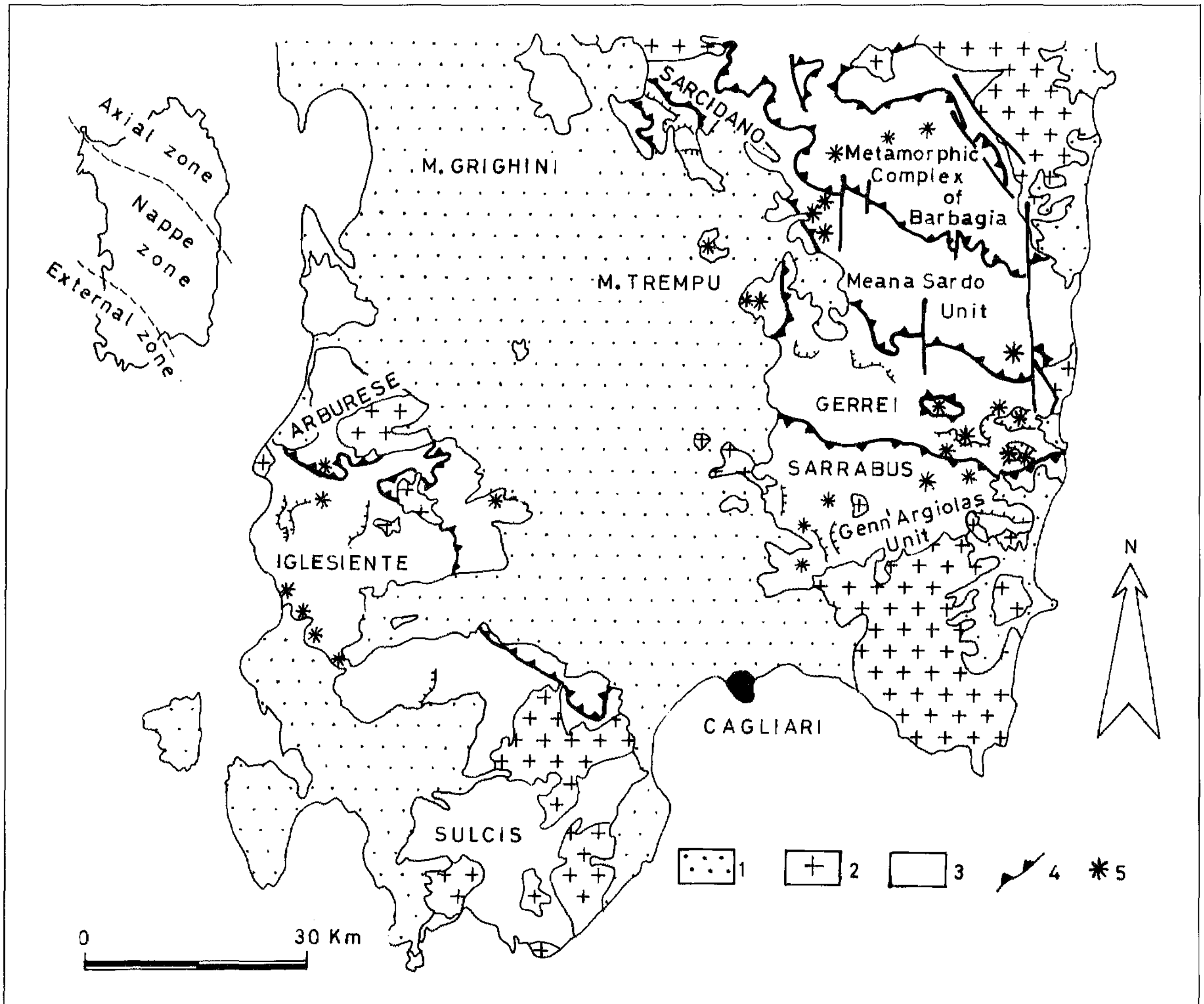

Fig.1 - Schematic geological map of the south Sardinian Paleozoic basement (modified from Carmignani et al., 1986a). Legend: 1: Post-Hercynian sediments and volcanites; 2: Hercynian granitoids; 3: Paleozoic metamorphites; 4: Major and minor overthrusts; 5: Sampled localities.

The Gerrei Unit is made up of Cambrian-Ordovician metasandstones, metagraywackes, metavolcanites, metaconglomerates and upper Ordovician -Devonian black quartzites,carbonaceous phyllites and metalimestones .

The Genn'Argiolas, Meana Sardo and Arburese units are mainly composed of Cambrian-Ordovician metasandstones, metavolcanites and Silurian-lower Carboniferous metalimestones, black slates and metasandstones. The stratigraphy and structural position in the pile of nappes of the three tectonic units show many similarities, supporting the hypothesis that they form part of a major tectonic unit translated southwards down to the External zone. (Carmignani et al., 1986a).

The Metamorphic Complex of Barbagia Unit is mainly composed of repeated alternations of metasandstones, metasiltites and phyllites. This unit has been considered Cambrian to Devonian in age (Carmignani et al., 1982).

The External zone consists of two major complexes separated by a sharp angular unconformity (Sardic unconformity). The first is composed of the Nebida, Gonnesa and Cabitza Formations of Cambrian to early Ordovician age (Barca et al.,1987). From bottom to top the Cabitza Formation is composed of nodular crystalline limestone slates and metasiltites, interbedded with minor massive fine grained metasandstones. The second complex is made up of middle Ordovician polygenetic conglomerates (Puddinga, Auct.) with a red-violet silty or shaly matrix. The upper part of the "Puddinga" is characterized by red silt and shale with minute conglomerate intercalations.

Two main tectonic phases have been recognized in the Nappe zone of the Hercynian chain by Carmignani et al. (1986a and bibliography therein) .The first phase (D_1) is related to a subduction and crustal thickening; the second

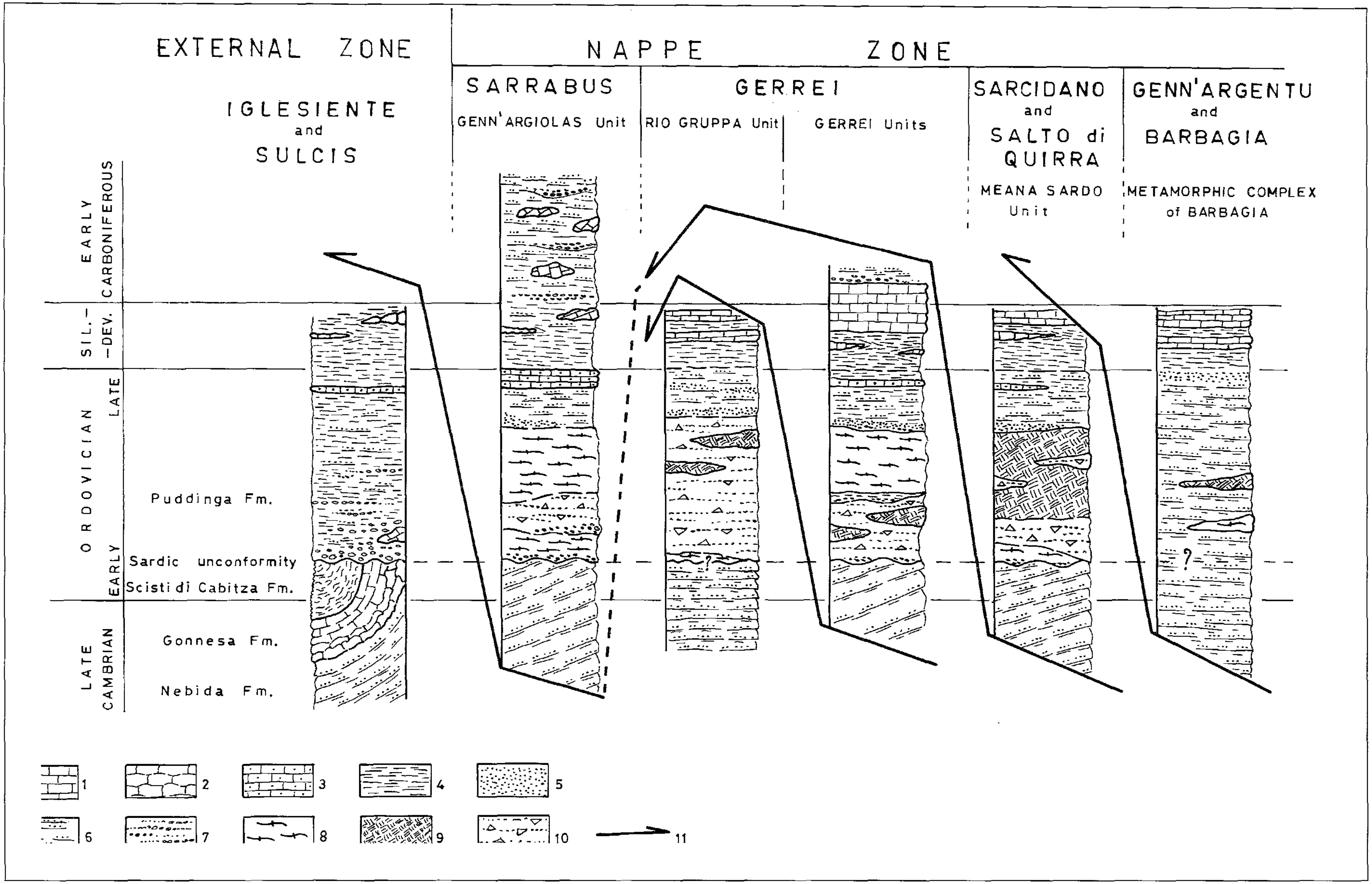

Fig.2 - Sketch showing the Paleozoic sequences and their tectonic relationships in the External and Nappe zones of the Hercynian chain of Sardinia. Legend : 1: Metalimestones; 2: Dolostones; 3: Carbonatic metasiltites,slates and metalimestones of Upper Ordovician age; 4:Slates and phyllites; 5: Metarkoses; 6: Metasandstones; 7: Metaconglomerates; 8: Metarhyodacites and metarhyolites; 9: Intermediate metavolcanites;10:Metavolcanoclastites and metamorphic products of reworked volcanites; 11: Tectonic contacts and vergence of overthrusts

(D_2) is mainly extensional in character and is related to the uplift of the chain (Carmignani et al., 1991). D_1 produced overturned isoclinal folds associated with axial plane S_1 schistosity. It was followed by important regional overthrusts. D_2 produced a NW-SE fold system associated with pervasive axial plane S_2 schistosity only in the deepest tectonic units of the Nappe zone.

The structural analyses in the External zone rocks was given by Arthaud (1963), Poll and Zwart (1964) and Cocozza (1980). The deformation history can be summarized as follows: i) minor E-W trending folds in the Arenig- Caradoc interval (Sardic phase) affecting only the Cambrian - early Ordovician rocks (Barca et al., 1987); ii) E-W trending folds without any schistosity development ,i.e. D_1 Hercynian phase; iii) N-S trending folds associated with strongly dipping penetrative (Sh) schistosity,i.e. D_2 Hercynian phase; iv) late Hercynian deformations .The tectonic characteristic of the Sardic phase is still largely unresolved. Two major hypotheses have been put forward: 1) a tectonic phase related to the Caledonian orogeny (Cocozza, 1980; Barca et al., 1984); and 2) a compressional event typical of continental back arc basins (Carmignani et al., 1991).

3. Sampling Procedure and Analytical Techniques

The sampling plan was made using the map of the "Structural Model of the Hercynian Basement of Sardinia" (Carmignani et al. 1987) and unpublished maps. In the Nappe zone the pelitic samples were collected from the Riu Gruppa , Gerrei , Arburese, Genn'Argiolas , Meana Sardo and the Metamorphic Complex of Barbagia units. In the External zone the pelitic samples were collected from the" Puddinga" and the Cabitza shale member of the Cabitza Formation. The sampled localities are shown in the tectonic sketch map in Fig. 1.

IC (half-height peak width expressed as °Δ2Θ) was determined using a fully automatic Rigaku Geigerflex Diffractometer. Measurements were performed on the < 2 μm fraction. The mounts on glass slides were prepared by sedimentation from aqueous suspensions. Air-dried preparations were used. The standard instrumental conditions were the following: CuKa, Ni-filtered radiation, Kv = 40 mA= 20 , divergence slit 0.5°, receiving slit 0.3 mm, scatter slit 1°, and goniometer speed 0.5°/min. IC was calibrated against standards kindly supplied by Prof. J. Hunziker (University of Lousanne). The estimated error in each single measurement was 0.02 °Δ2Θ.The diagenetic/ anchimetamorphic (0.42 °Δ2Θ) and anchimetamorphic/ epimetamorphic boundaries (0.25 °Δ2Θ) proposed by Kübler (1984) were used.

The mica polytypes were identified by the method suggested by Maxwell and Hower (1967) .

4. Illite Crystallinity

4.1 *The Nappe Zone*

The pelitic rocks of the Nappe zone are moderately to strongly schistose. The most common S_1 mineral assemblage is white K-mica ,quartz and chlorite. Other minerals possibly present include: calcite, chloritoid, paragonite, Fe-oxide, carbonaceous matter, tourmaline and zircon. No mixed layered minerals are observed. Detrital muscovite is sporadically present but in very low modal proportions and its effects are probably negligible in the < 2 μm fraction. The dominant muscovite polytype is 2M. The 2M/(2M+1M d) ratio is just below 1.00.

In the samples of the Riu Gruppa Unit, two schistosities were observed. The S_2 schistosity is moderately to tightly spaced. It is coupled by opacitic materials and rare phyllosilicates.

The IC data are presented in the histograms in Fig. 3. From the lowest to the highest tectonic unit of the Nappe zone the results are:

i) the Riu Gruppa Unit yielded homogeneous IC values in the range of 0.18- 0.26 °Δ2Θ ;

ii) the Gerrei Unit yielded IC values of 0.24- 0.32 °Δ2Θ in the front and 0.20- 0.26 °Δ2Θ in the root zones of the Unit;

iii) the geometrically intermediate tectonic units yielded the following IC values: Arburese Unit 0.20-0.26 °Δ2Θ, Genn'Argiolas Unit 0.20-0.28 °Δ2Θ, except one sample 0. 34 °Δ2Θ,. In the Meana Sardo Unit there is an appreciable dispersion of IC values.This is often observed in individual outcrops and may be ascribed to the influence of such factors as bulk chemistry and/or composition of illite on the IC (Frey, 1987;Franceschelli et al., 1991). Nevertheless the IC values show a faint tendency to decrease from the front to the root zones of the unit.

iv) the Metamorphic Complex of Barbagia yielded IC values of 0.24-0.34 °Δ2Θ in the front and 0.18-0.22 °Δ2Θ in the root zones of the unit.

4.2 *The External Zone*

The samples from the " Puddinga" are red and weakly to moderately schistose. White K-mica, chlorite, quartz, tourmaline and apatite are the most common components of the rocks. Sporadically mixed layer chlorite/ vermiculite was also detected. The minerals are mostly parallel to the main Hercynian schistosity. 2M is the dominant muscovite polytype. The mean b_0 value of the Puddinga samples given by Conti et al.(1978) is 9.000 A ± 0.005. The IC values range from 0.34 - 0.42 °Δ2Θ with an average value of 0.39 ± 0.03 °Δ2Θ (Fig. 4).

The pelitic rocks of the Cabitza Formation show two schistosities. They are composed of quartz,white k-mica

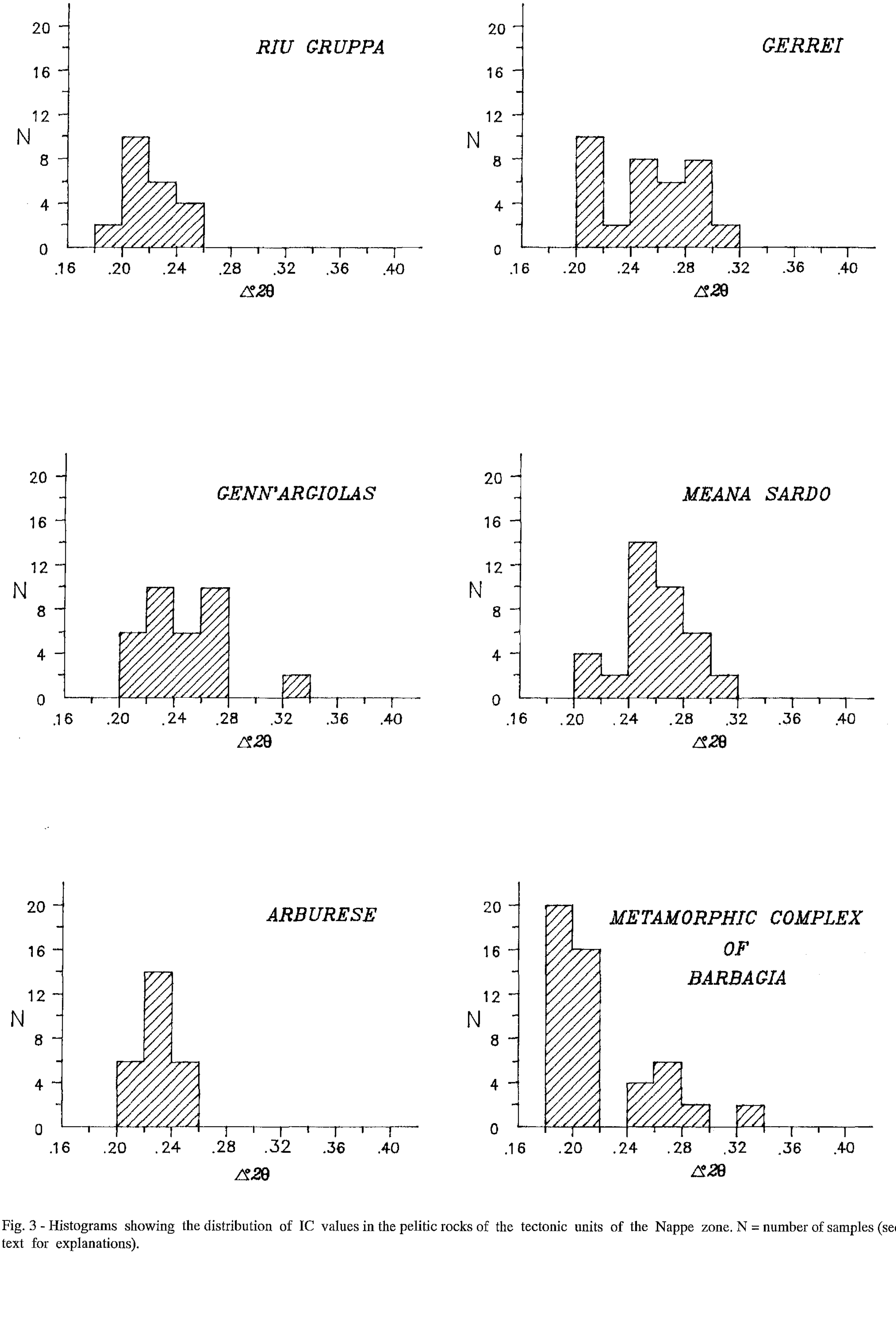

Fig. 3 - Histograms showing the distribution of IC values in the pelitic rocks of the tectonic units of the Nappe zone. N = number of samples (see text for explanations).

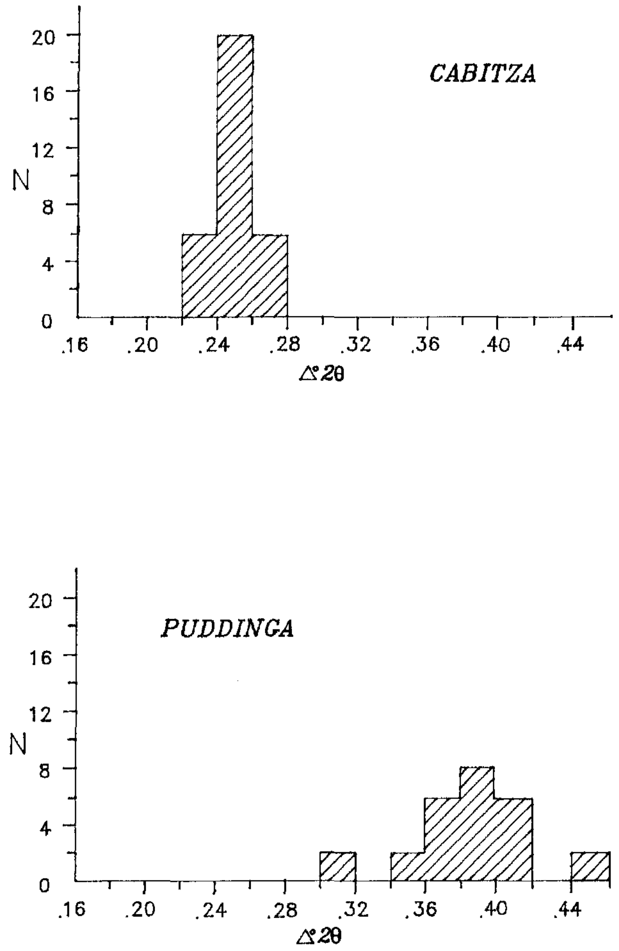

Fig. 4 - Histograms showing the distribution of IC values in the pelitic rocks of the " Puddinga" and Cabitza Formation of the External zone. N = number of samples (see text for explanations).

and chlorite mostly arranged parallely to the pre-Hercynian schistosity and minor tourmaline, apatite and carbonates. Other minerals sporadically observed are the mixed layer chlorite/vermiculite and minor illite/vermiculite. Detrital mica is scarce and in many samples absent. The 2M/ $(2M+1M_d)$ muscovite polytypes ratio is lower than 0.90. The mean b_0 value of the Cabitza samples given by Conti et al.(1978) is 9.025A $\pm$ 0.005. The IC values range from 0.22 to 0.28 $°\Delta 2\Theta$ with an average value of 0.25 $\pm$ 0.01 $°\Delta 2\Theta$(Fig. 4).

5. Discussion

5.1 *Hercynian Metamorphic Zonation*

The distribution of IC values in the Hercynian pelitic rocks from the External and Nappe zones shows the following trend:

i) relatively high $°\Delta 2\Theta$ values in the "Puddinga" (External zone) with respect to the Riu Gruppa,Arburese and other units of the Nappe zone;

ii) an increase in the pile of nappe of $°\Delta 2\Theta$ from the Riu Gruppa Unit (the deepest tectonic unit) to the front of the Gerrei Unit and the Metamorphic Complex of Barbagia Unit;

iii) a general decrease of $°\Delta 2\Theta$ from the front to the root zones in both the Gerrei and Metamorphic Complex of Barbagia units.

For an interpretation of the geological meaning of the regional distribution of the IC values, it is crucial to identify the original position of the various tectonic units before their piling up. Carmignani et al. (1982) suggest the following order from the External zone to the Nappe zone: "Puddinga", Riu Gruppa, Gerrei, Arburese, Genn'Argiolas, Meana Sardo and Metamorphic Complex of Barbagia (Fig.5a).

According to Carmignani et al.(1991) the late Devonian- early Carboniferous continent-continent collision was followed by crust-mantle detachment and crustal stacking with migration of the deformation and metamorphism from the axial (NE Sardinia) to the foreland zones (SW Sardinia).

During the emplacement of nappes, the sediments belonging to the outermost tectonic units of the Nappe zone (i.e. the Riu Gruppa and the Gerrei Units) were buried deeper and deeper within the crust than the internal ones (Fig.5b).

The thickness of the pile of nappes , in the central part of the Nappe zone, produced an increase in load pressure that could explain the slightly increase in metamorphic grade from the front of the upper tectonic unit to the lowermost (Riu Gruppa Unit). In each tectonic unit of the Nappe zone, the sediments of the root zones were buried deeper and deeper than those of the front; moreover the thickness of the pile of nappes decreases in the External zone.

This may explain the progressive northward increase in metamorphic grade recorded by the rocks of the Metamorphic Complex of Barbagia and Gerrei units and, more generally by the rocks of the Nappe zone with respect to those of the External zone.

The contrasting IC values observed between the Arburese Unit and the underlying "Puddinga" requires further comments. The IC values indicate epimetamorphic conditions in the Arburese Unit and anchimetamorphic conditions in the underlying "Puddinga".The discontinuity in metamorphic grade coincides with the tectonic contact between the two units. The discontinuous inverse metamorphic zonation may be explained by post-metamorphic thrusting of the Arburese Unit over the "Puddinga" (Fig.5c) during the late-Hercynian extensional tectonics.

5.2 *Evidence of Pre-Hercynian Metamorphic Effects*

Two main types of evidence suggest that the Cabitza

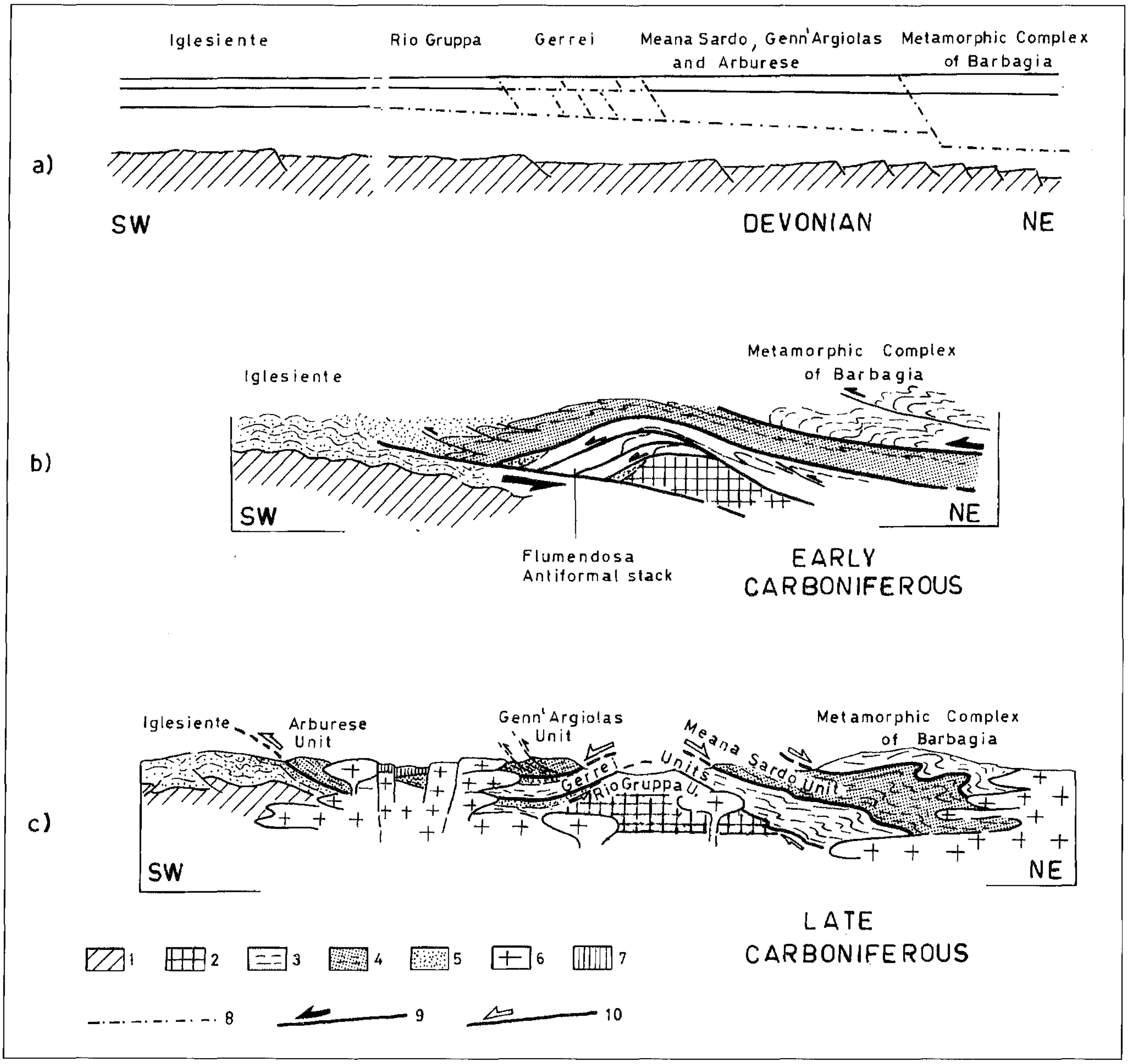

Fig.5 - Schematic cross sections illustrating the structural evolution of central and south Sardinia from the Devonian to the late Carboniferous (modified from Cappelli,1991) : a) Devonian : pre-collisional geometry showing the probable areas of sedimentation of Puddinga, Riu Gruppa, Gerrei, Arburese, Genn'Argiolas, Meana Sardo and Metamorphic Complex of Barbagia units . Restored state traces of principal thrust faults and ramp-flat geometry are also shown; b) Early Carboniferous: Hercynian continent-continent collision causing shortening and crustal thickening. The compressional tectonics developed an antiformal stack in the Flumendosa valley, central Sardinia; c) Middle - late Carboniferous: Post -collisional extensional tectonics, associated and followed by granitoid emplacement caused crustal thinning and uplift of the Paleozoic basement. Legend : 1: pre-Paleozoic substrata; 2: Rio Gruppa Unit; 3: Gerrei Unit; 4: Meana Sardo,Genna'Argiolas and Arburese units; 5: Syntectonic Flysh deposit; 6: Hercynian granitoids; 7: Post-Hercynian sedimentary sequences; 8: Traces of thrust faults , 9: Compressional thrust planes; 10: Early thrust planes reactivated in extensional regimes.

Formation underwent pre-Hercynian deformation and weak metamorphism. They are : i) structural data; ii) illite crystallinity .

Data collected by the authors of the present paper emphasize the existence of a pre-Hercynian schistosity parallel to sub-parallel to the bedding in the Cabitza Formation. This can be documented by a careful structural analysis of the Cabitza Formation at the angular unconformity near the Village of Nebida (Fig. 6).

The penetrative schistosity (Sh) related to the main Hercynian tectonic phase in the pelitic rocks of the Cabitza Formation is moderately spaced (Fig. 6) and cuts across an early schistosity (Sp), which we refer to the Sardic phase.

The IC values indicate anchimetamorphic conditions in the" Puddinga" and epimetamorphic-anchimetamorphic boundary conditions in the underlying Cabitza Formation. The discontinuity in the IC values, presumably caused by a discontinuity in the metamorphic grade, occurs at the

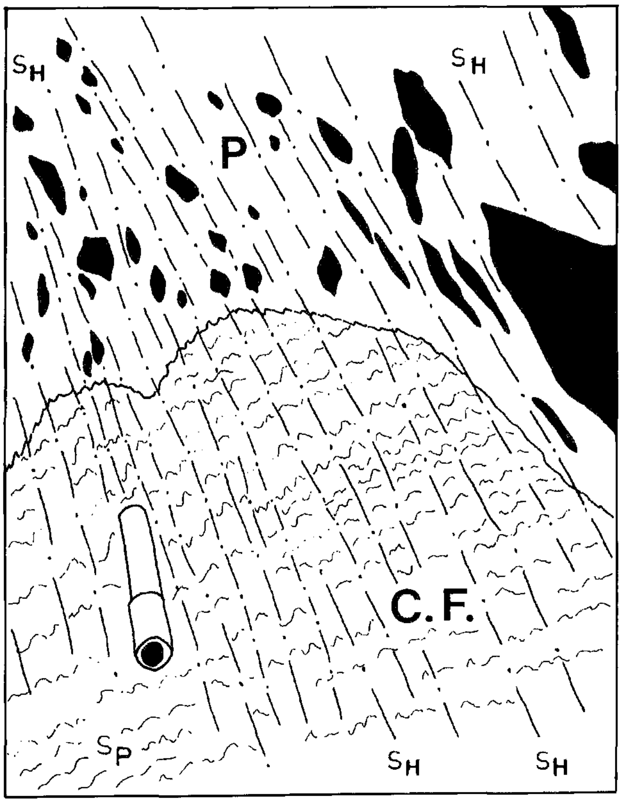

Fig. 6 - Sketch showing the structural relation at angular unconformity located about 1 Km south of the Village of Nebida between the Cabitza Formation (C.F.) and the overlying lower Ordovician conglomerates of "Puddinga" Auct. (P). The pencil shows the trails of the penetrative Hercynian schistosity (SH) in the shale of the Cabitza Formation. In the Cabitza Formation, note a pervasive pre-Hercynian schistosity (Sp) at a sharp angle with the S_H schistosity.

angular Sardic unconformity between the two Formations. This difference can be explained by the fact that the illite in pelitic rocks from the Cabitza Formation acquired their " good crystallinity" in the middle Ordovician before Hercynian metamorphism affected the "Puddinga". Further evidence confirming the pre-Hercynian metamorphism in the Cabitza Formation is the systematic difference in the white k-mica b_0 parameter at the Cabitza/ Puddinga contact detected by Conti et al.(1978). These Authors interpreted the b_0 data in the "Puddinga" and the Cabitza Formation as indicative of low and medium pressure- type metamorphism respectively.

According to Carmignani et al. (1991) at the Gondwanian margin during the middle Ordovician there was the following scenario: subduction of an ocean crust from NE to SW , a volcanic arc (Central Sardinia) and a continental back arc basin (SW Sardinia).

The conventional stratigraphical thickness overlying the Cabitza Formation in the back arc basin during the middle Ordovician may not be more than a few kilometers. This implies that the metamorphic effects are essentially due to the relatively high geothermal gradient caused by the lithospheric extension (Mason, 1990) and/or emplacement of the Ordovician magmatic bodies in the neighbouring areas (Sassi & Visona, 1989).

In the middle Ordovician the Cabitza Formation and the underlaying Cambrian sequences uplifted and were unconformably covered by the clastic sedimentation of the "Puddinga". The Sardic phase may be tentatively related to a compressional phase that produced the closure of the back arc basin .

During the early Carboniferous, the Cabitza Formation was newly deformed together with the" Puddinga", but the Hercynian metamorphism did not significantly modify the IC values.

6. Conclusions

Two main preliminarygeological conclusions can be drawn from the study of the IC in the External and Nappe zones of the Hercynian chain of Sardinia:

i) The IC values indicate anchimetamorphic conditions in the External zone and anchi - epimetamorphic conditions in the Nappe zone.In the Nappe zone the distribution of the IC values also reveals a roughly metamorphic zonation from the lower to the upper tectonic units of the pile of nappes as well as from the front to the root zones of the Metamorphic Complex of Barbagia and Gerrei units.

ii) In the External zone the pelitic rocks of the Cabitza Formation record metamorphic effects and deformation of middle Ordovician age.

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CALCIMICROBIAL-ARCHAEOCYATHAN BUILDUPS AND EVOLUTION OF THE NORTHWESTERN PLATFORM MARGIN IN THE LOWER GONNESA FORMATION (LOWER CAMBRIAN) OF SW-SARDINIA (ITALY)

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ABSTRACT

The northwestern margin of the Gonnesa platform is exposed in the Buggerru-Acquaresi region. Numerous reef-like buildups occur about 80 to 150 m above the base of the Gonnesa Formation, consisting of several types of calcimicrobial organisms (e.g. *Epiphyton*, *Renalcis*, *Girvanella*) and of regular, and irregular archaeocyathans. These buildups show early cementation. Early tectonics at the instable platform margin caused deformation and fragmentation of the buildups and the enclosing laminites. These structures and fossil content, indicating open marine conditions at the northwestern platform margin during Gonnesan time, are missing at the eastern, more restricted margin of the platform.

KEY WORDS: *Buildup, platform margin, Lower Cambrian (Gonnesa Formation), SW-Sardinia (Iglesiente)*

1. INTRODUCTION

In the Lower to Middle Cambrian sequence of Southwest Sardinia a carbonate platform, under- and overlain by terrigenous clastic sediments, is well developed (Fig. 1).

The Cambrian sequence shows various stages of platform evolution, from a homoclinal, siliciclastic-carbonate ramp in Nebidan time to an isolated carbonate platform in Gonnesan time, which was finally segmented and drowned in late Gonnesan and Campo Pisano time (Bechstädt et al., 1988; Bechstädt & Boni, 1989).

The Cambrian of Southwest Sardinia, according to a classification recently proposed by Pillola (1990), is subdivided in groups, formations and members. In this paper, however, the informal stratigraphical names are still used with the subdivision sensu Pillola (1990) in parenthesis.

The Lower to Middle Gonnesa Formation (Santa Barbara Formation) is subdivided in the platform sediments of "Laminated Dolomite" and "Grey Dolomite" (Arcu Biasterria Member) and the platform margin dolomites and limestones of the Planu Sartu Member (sensu Bechstädt et al., 1988).

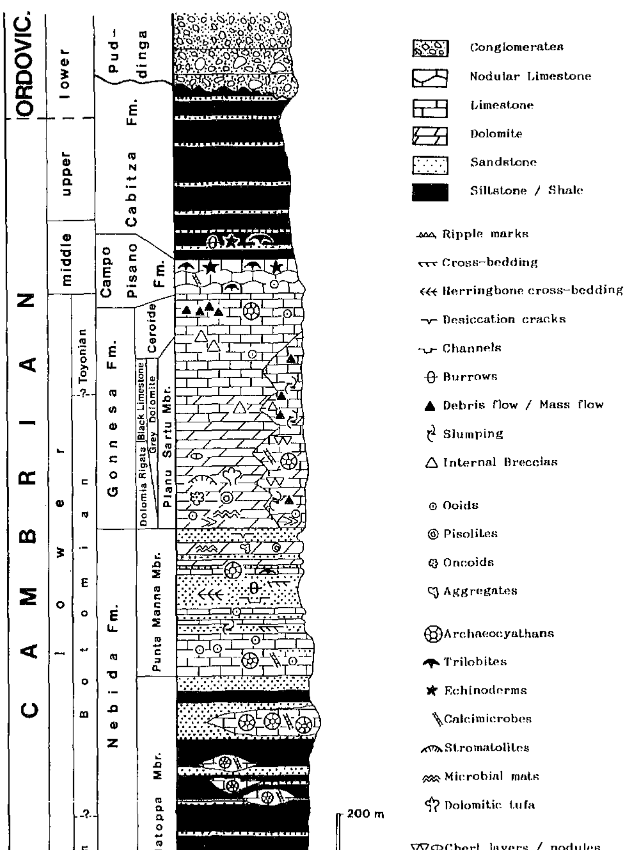


Fig. 1 - Generalized Cambro-Ordovician stratigraphic column, modified after Fröhler et al. (1991a). Biostratigraphic correlation after Debrenne & Gandin (1985) and Pillola (1990).

2. Lower to Middle Gonnesa Formation

With the onset of Gonnesa Formation (Gonnesa Group) terrigenous input was interrupted and the platform became isolated. Flanks, occasionally very instable, occur at southeastern and northwestern parts of the platform.

2.1 *Platform sediments*

Tidal to supratidal carbonates of the "Laminated Dolomite" (Arcu Biasterria Member) are arranged in shallowing upward cycles. Detailed descriptions are given by Gandin et al. (1974; 1987), Bechstädt et al. (1988), and Bechstädt & Boni (1989). Stromatolitic boundstones with intercalated mudstones prevail. They are associated with oncolites, grapestones, peloids and vadose pisolites. Subordinate layers of ooids are possibly originated in situ within microbial mats (comp. Dahanayake et al., 1985). Pseudomorphs after evaporites are scattered in the mudstones. Black coloured dolomitic tufa beds indicate temporary subaerial exposition of the Laminated Dolomite (Bechstädt & Boni, 1989).

2.2 *Platform Margin and Slope Sediments (Planu Sartu Member)*

In the Buggerru area (Fig. 2) the Lower to Middle Gonnesa Formation (Gonnesa Group) is developed in a platform margin facies and in the slope facies of the Planu Sartu Member. These facies associations are represented by low energy microbial mats and mud laminites (Debrenne & Gandin, 1985), with intercalated buildups (Fröhler et al., 1991; 1991a). Tectonic pulses caused synsedimentary deformations. Sedimentary fabrics within the Lower Gonnesa Formation of the western Iglesiente area exhibit slumping features, subaquatic mass flows and seismites (Bechstädt et al., 1988; Bechstädt & Boni, 1989; Fröhler et al., 1991). The dolomitic sediments alternate in the upper part with "Black Limestones" (Boni & Marinacci, 1980; Angiullì et al., 1985). Near Acquaresi the marginal facies passes into the eastward situated platform facies of the "Laminated Dolomite" and "Grey Dolomite" (Arcu Biasterria Member).

2.2.1 *Laminites*

Laminites of the lowermost Gonnesa Formation (Gonnesa Group) consist in the investigated area mainly of medium to dark grey stromatolitic bindstones with subparallel to wavy laminations, sometimes they show laterally linked hemispheroids. Alternating with them light grey, laminated to massive bedded mudstones occur. These sediments belong to a predominant shallow subtidal setting. Peloids, oolites and reworked laminites are intercalated only as thin layers. Spherulitic quartz occur scattered in the mudstones. Other typical features of the Laminated Dolomite (Santa Barbara Formation, Arcu Biasterria Member), e.g. pisolites, oncolites, and dolomitic tufa, are generally absent. They occur sparsely only in the eastern part of the study area (Fig. 2).

Toward the upper part of the Santa Barbara Formation the amount of light to dark grey, monotonous mm to dm varve-like laminated mudstones, with chert layers up to several cm in thickness increases in the westernmost areas. Near Cala Domestica and at M. Malfidano these dolomites alternate with recrystallized "Black Limestones" (Boni & Marinacci, 1980; Angiulli et al., 1985).

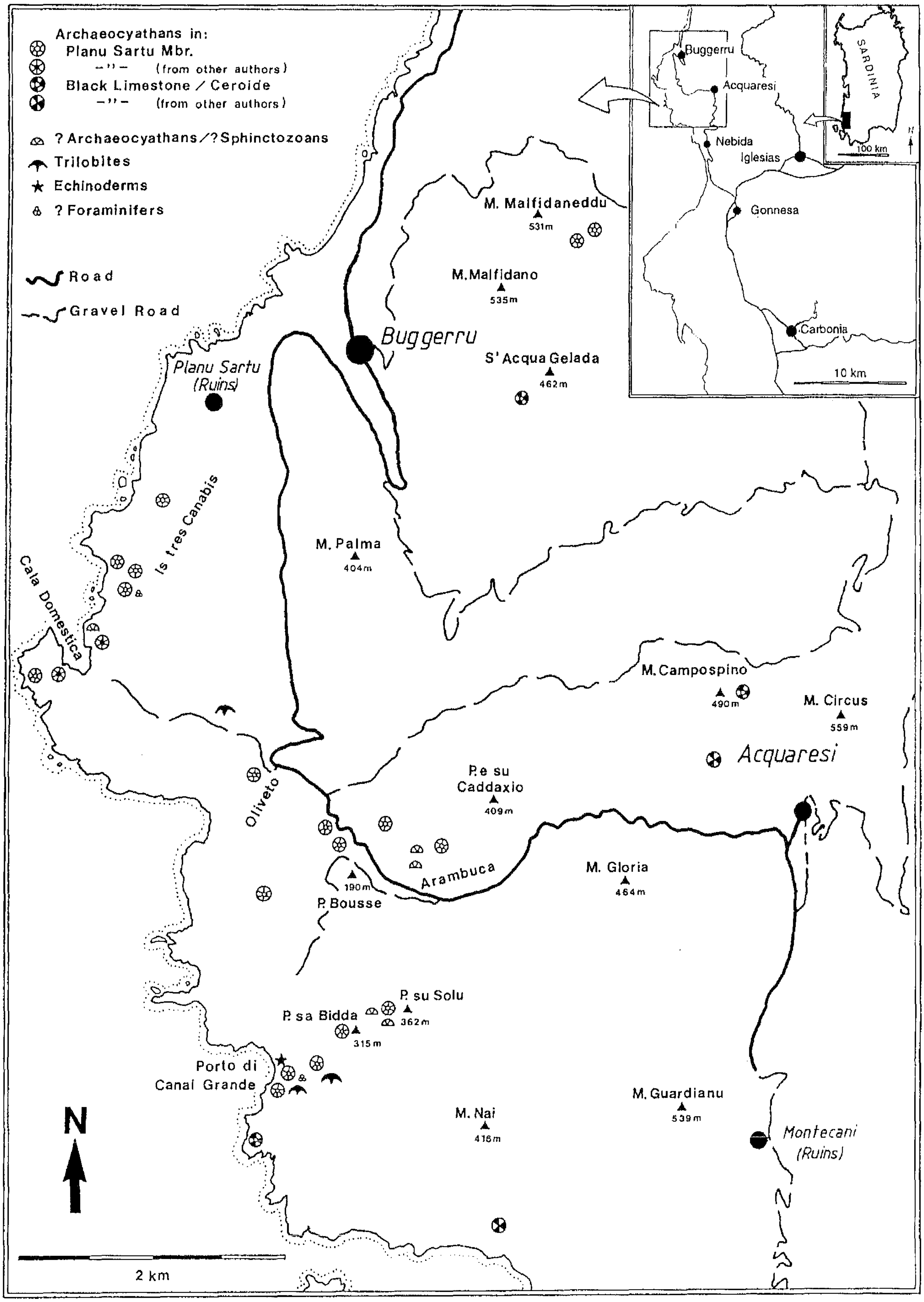
These varve-like laminites and Black Limestones are interpreted as slope sediments (Planu Sartu Member) by Bechstädt et al. (1988). Distinct structures in the Black Limestones indicate calcitization of former lower to middle Gonnesan dolomites (Zeeh et al., 1991).

2.2.2 *Calcimicrobial-Archaeocyatan Buildups*

Numerous reef-like buildups within the laminites, described for the first time, occur about 80 to 150 m above the base of the Gonnesa Formation (Gonnesa Group) in the Buggerru-Acquaresi region. These meter to some tens of meters thick buildups have been constructed by several groups of calcimicrobial organisms (*Epiphyton*, *Renalcis*, *Girvanella*). Due to the uncertain taxonomic affinity of some of these organisms, James & Gravestock (1990) use the general term "calcimicrobes" for *Renalcis, Epiphyton* and *Girvanella*, a shortened form of "calcified microbial microfossils". The crudely laminated, dendrolitic and clotted calcimicrobial structures are comparable to low energy buildups, described by James & Gravestock (1990) from open shelf and shelf edge environments of the Lower Cambrian of Australia. Transition to the enclosing laminites is frequently indistinct.

The buildups often show fenestral fabrics. Cavities with calcimicrobial encrustations at the roof are filled with internal sediments. *Renalcis* prevails in most of the buildups. Filamentous calcimicrobes, like *Girvanella*, dominate in the eastern parts of the study area. West of P. Bousse - P. su Solu and at M. Malfidaneddu, *Renalcis* is associated with different forms of *Epiphyton*. Archaeocyathans frequently show in situ encrustations by *Girvanella* and other calcimicrobes. At Cala Domestica archaeocyathans are

Fig. 2 - Fossil distribution within the Planu Sartu Member and additional archaeocyathans from "Black Limestone" and "Ceroide Limestone" in the study area, modified after Fröhler et al. (1991). The locations of the archaeocyathans "from other authors" are from Gross (1982), Laske (1983) and Debrenne & Gandin (1985).


Archaeocyathans in:
Planu Sartu Mbr.
-"- (from other authors)
Black Limestone / Ceroide
-"- (from other authors)
? Archaeocyathans /? Sphinctozoans
Trilobites
Echinoderms
? Foraminifers
Road
Gravel Road
Buggerru
Acquaresi
Nebida
Iglesias
Gonnesa
Carbonia
SARDINIA
100 km
10 km
M. Malfidaneddu
531m
M.Malfidano
535m
Buggerru
S'Acqua Gelada
462m
Planu Sartu (Ruins)
Is tres Canabis
Cala Domestica
M. Palma
404m
M. Campospino
490m
M. Circus
559m
Acquaresi
Oliveto
P.e su Caddaxio
409m
Arambuca
M. Gloria
464m
190m
P. Bousse
P. su Solu
362m
P. sa Bidda
315m
Porto di Canal Grande
M. Guardianu
539m
M. Nai
416m
Montecani (Ruins)
N
2 km

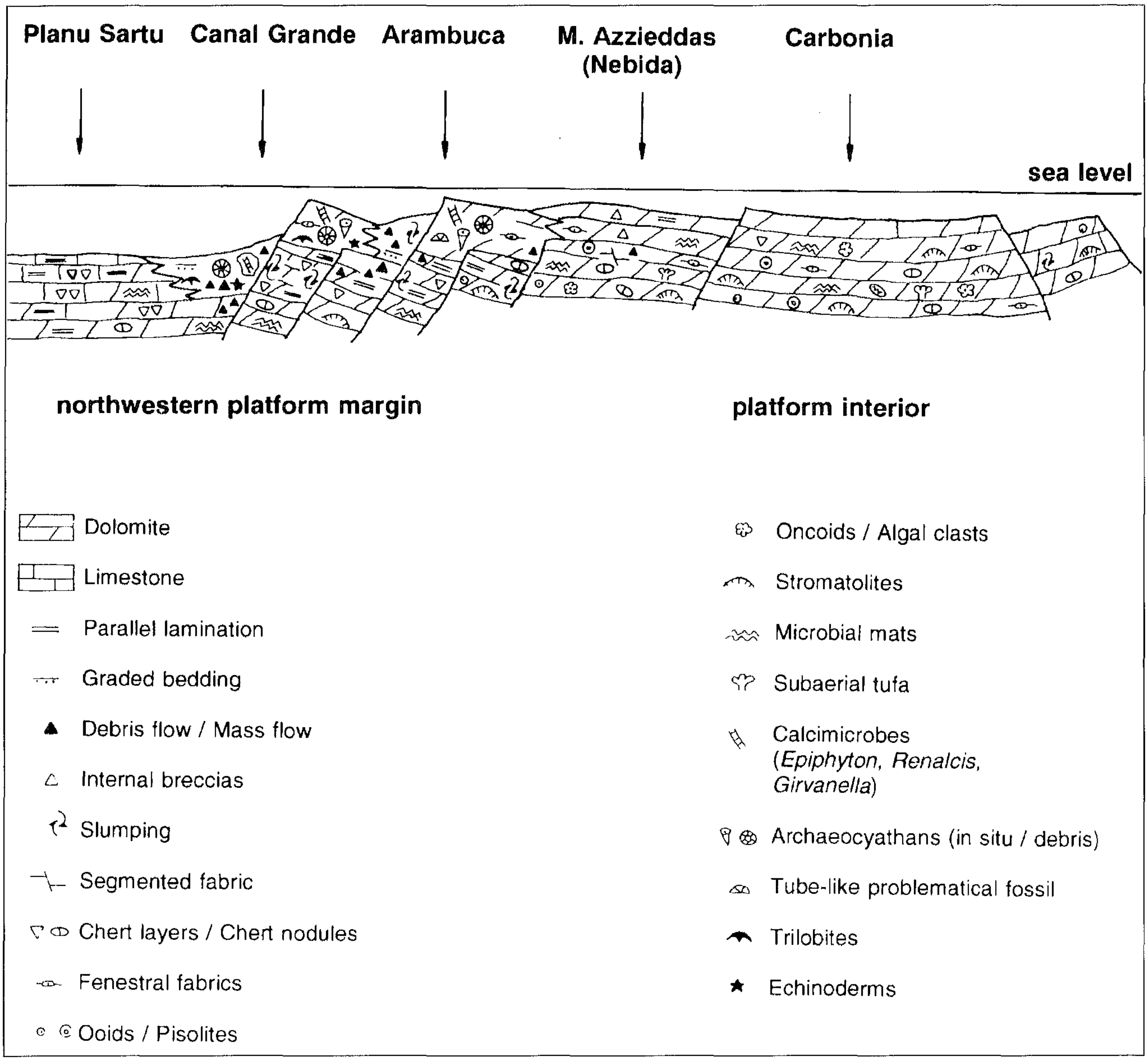


Fig. 3 - Cross section of northwestern and central parts of the Gonnesa carbonate platform, Middle Gonnesan time, after Fröhler et al. (1991).

associated with ball-shaped clusters of calcimicrobes, e.g. the delicate branching *Epiphyton crinitum*, and *Epiphyton fruticosum*. These clusters have been described as oncoids by Debrenne & Gandin (1985: Fig. 9). Except some micritic and/or microbial encrustations no real concentric layering can be observed within these "oncoids". On the contrary radial branching is frequent within these calcimicrobial balls. Structures corresponding to *Palaeomicrocodium devonicum* exist near Cala Domestica. These are reported for the first time from Sardinia and from the Cambrian in general (Schroeder, pers. comm., 1991).

From Cala Domestica archaeocyathans of lower Gonnesan age have already been reported by Bornemann (1891), whereas in eastern parts of the platform no archaeocyathans have been reported at all. The occurences at Cala Domestica have been rediscovered by Laske (1983) and Debrenne & Gandin (1985). The latter authors described four genera of regular archaeocyathans (*Coscinocyathus, Rasetticyathus, Densocyathus* and *Aldanocyathus*) from this locality. Lower Gonnesan archaeocyathans within debris flows near Canal Grande have been mentioned for the first time by Bechstädt & Boni (1989). Many additional occurences of archaeocyathans have been discovered during our recent investigations (Fröhler et al., 1991; 1991a). Occurrences of archaeocyathans in the Planu Sartu Mbr. range from Porto di Canal Grande in the south to M. Malfidaneddu in the north (Fig. 2). The recently discovered archaeocyathans from buildups and from debris flows in the study area belong to Dokidocyathina and to the genera *Capsulocyathus, Nochoroicyathus, Rotundocyathus,* *Sajanocyathus?, Mikhnocyathus?, Erismacoscinus, Antoniocoscinus, Coscinocyathus, Bicyathus, Chouberticyathus, Protopharetra* and *Agastrocyathus* (Perejón et al., in prep.). Within the buildups the archaeocyathans are generally encrusted by microbial layers and by *Renalcis*.

At Canal Grande and Cala Domestica, buildups and debris layers contain additional fragments of trilobites, echinoderms (comp. Fröhler et al., 1991: Pl. 1/6), sponges, *Chancelloria* (Mostler, pers. comm., 1991), and problematical microfossils with analogies to foraminifers.

Tubular fossils of uncertain affinity occur in small colonies at Arambuca, P. su Solu and Cala Domestica (Perejón et al., in prep.). The tubular fossils show analogies to some archaeocyathans and sphinctozoans but also to codiacean algae e.g. *Palaeoporella*, or dasyclads (comp. Fröhler et al., 1991: Pl. 1/7-8). These fossils are surrounded by crusts of early marine cements and/or calcimicrobes.

The fossil content within the buildups and in associated debris indicates temporary open marine conditions during Lower to Middle Gonnesan time in the Buggerru area (Debrenne & Gandin, 1985; Bechstädt & Boni, 1989; Gandin, 1990). None of these facies types occur at the eastern margin in eastern Sulcis. Contrary to western areas, the eastern basin probably was much more restricted (Bechstädt & Boni, 1989).

2.2.3 *Synsedimentary Tectonics*

Synsedimentary tectonics caused fragmentation of the buildups situated in the western part of the study area. The laminites frequently have been affected by soft sediment deformation, among them pull apart structures, and slumping (Bechstädt & Boni, 1989; Fröhler et al., 1991). Debris flows and mass flows consisting of reworked microbial laminites, mudstones and reef fragments can reach several m in thickness. Clast-sizes of graded beds range from mm to more than 30 cm. Laterally the debris sheets rapidly decrease in size (Bechstädt & Boni, 1989; Fröhler et al., 1991). Debris flows generated at the slope form the transition to probably basinal sediments.

3. Conclusions

Recent investigations permit a more detailed description of the evolution of the northwestern margin of the Gonnesa platform (Fig. 3). In earliest Gonnesan time only a few lithotypes of the typical shallow water facies (shallow subtidal to supratidal) are present in the Buggerru-Acquaresi area (Angiulli et al., 1985; Fröhler et al., 1991). These facies, on the other hand, are widespread in the inner parts of the Gonnesa platform (Gandin et al., 1974; Bechstädt et al., 1988). Tectonic pulses caused instability and marine ingressions at the northwestern platform flanks (Bechstädt & Boni, 1989; Fröhler et al., 1991).

On tectonically instable horsts within a low energy environment, structured calcimicrobial-archaeocyathan buildups started to settle. They experienced early marine cementation and fragmentation. "Reef debris" and "inter reef deposits" (laminated mudstones and bindstones) were often transported downslope in form of debris flows, mixing with probably anoxic slope sediments, represented by varve-like laminated dolomites and black limestones.

In the upper part of the Gonnesa Formation, during sedimentation of the "Ceroide Limestone", occasionally intercalated high energy deposits occured in the Buggerru-Acquaresi region (comp. Bechstädt et al., 1988). They contain few archaeocyathans (Debrenne & Gandin, 1985) and probably belong to still existing elevated areas of the northwestern margin. Again, archaeocyathans are missing at the eastern margin of the platform.

Throughout Gonnesan time open marine areas were situated somewhere further to the west (Angiulli et al., 1985; Bechstädt et al., 1988; Gandin, 1990). For the Eastern Sulcis Basin a more restricted situation is indicated by lithological and paleontological data (Bechstädt et al., 1988; Gandin, 1990).

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STRATIGRAPHICAL AND STRUCTURAL OUTLINE OF THE RIU GRUPPA TECTONIC UNIT (SOUTHEASTERN SARDINIA)

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Abstract

In the Hercynian nappe building of southeastern Sardinia the overlapping of different tectonic units is well exposed in corrispondence of the Flumendosa Valley antiform (Carmignani et al., 1982). Detailed mappings of the deeper unit (Riu Gruppa Unit) allowed the authors to identifie a lithostratigraphic succession which appeared correlative to those of other tectonic units of the nappe zone of the Sardic Hercynian chain and, as consequence to better define its structural frame.

Key words: *Hercynian Chain, nappe zone, Sardinia.*

1. Introduction

The main structure of Flumendosa Valley (Southern Sardinia) consists of a wide, N120E trending antiform (Flumendosa Valley antiform) which shows the superposition of the different tectonic units (Carmignani et al., 1982) of the nappe building of the Sardic Hercynian Chain (Fig. 1).

This antiform is essentially due to a first Hercynian deformative event which results from the collision-related shortening; it develops as a wide antiformal stack (Cappelli, 1991) with associated, isoclinal folds,thrusts and synmetamorphic slaty cleavage. During a second tectonic event which was extensional,the thrust planes were reactived as low angle detachments and asymmetric folds developed, both consistent with a sense of movement away from the culmination of the structure (Cappelli, 1989; Carmignani et al., 1991). Although at least five overlapped tectonic units have been envisaged, a correlation among them was still full of gaps; particularly one of the more superficial units (i.e. Bruncu Nieddu Unit) and the deepest one (Riu Gruppa Unit) retained sedimentary and volcanic successions different or incomplete, if compared to those of the Gerrei-type or Sarcidano-type units (Carmignani et al., 1982; Cappelli & Moretti in Carmignani et al., 1986). Recent detailed mapping carried out by the authors allow them to identify, both in the Riu Gruppa and Bruncu Nieddu Units, a sequence of sedimentary and volcanic events which are correlative, mainly on litostratigrafic basis, to those of the other units, pertaining to the nappe building.

In this short note we present a detailed lithostratigraphycal reconstruction of the Riu Gruppa Unit, already briefly described by the authors (Gattiglio & Oggiano, 1990), joined to a map with a cross section of the cropping area. The described volcanic and sedimentary successions are also comparable to that of other deeper tectonic units (i.e. Castello Medusa,) where also Carosi et al., 1991 roughly recognize the cambro-devonian type-sequence of the nappe zone.

2. Geology of Riu Gruppa Unit

The Rio Gruppa tectonic Unit crops out in a tectonic window along the easternmost culmination of the Flumendosa Valley antiform, below a stack of Gerrei type units.

Its metamorphic conditions are fearly higher than those of the overlapping tectonic units. The most common mineral assemblage along the oldest detectable cleavage in metapelitic lithologies is white K-mica, quartz, chlorite $\pm$ chloritoid $\pm$ paragonite (Fadda et al., this volume).

The crystallinity index average values of $\Delta 2\Theta = .20$ and $\Delta 2\Theta = .24$ (Fadda et al., Op. cit.) in metapelites of Riu Gruppa Unit and Gerrei Units respectively, reveal an increase of the metamorphic conditions downward as consequence of the crustal thickening owing to the stacking of different nappes.

Notwithstanding this slightly higher metamorphism, the preserved primary structures, particularly in metaarenaceous rocks, allowed the identification of a sequence of sedimentary and volcanic episodes correlative to that of other units which occupy a more superficial level in the nappe building of Eastern Sardinia.

The sequence, showing a thickness of about 800 mt, is exposed on the northern flank of the Flumendosa antiform (fig.1) and consists of:

— Turbiditic metasandstones showing decimetric to metric alternations of pelitic and arenaceous levels, graded

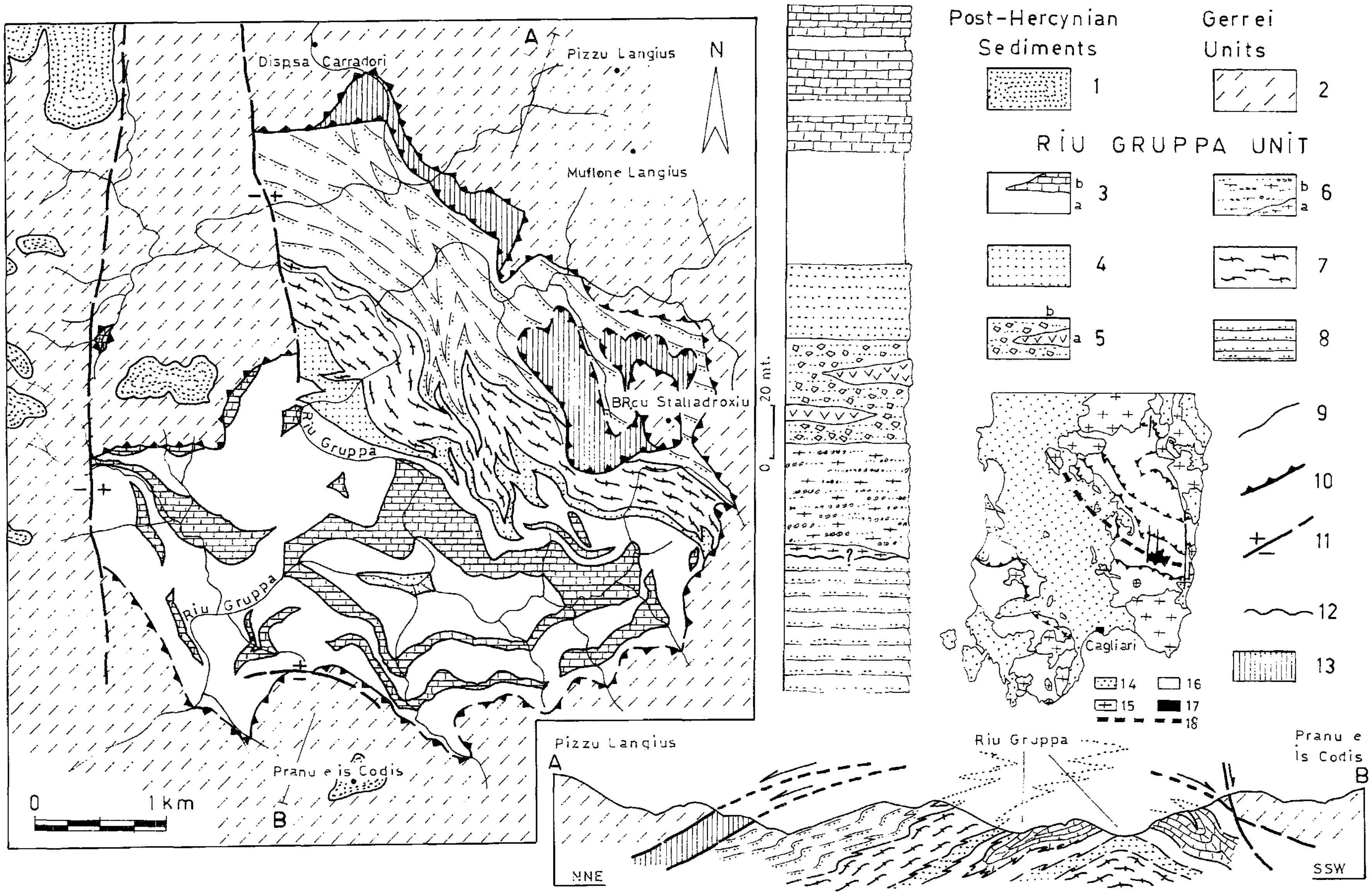

Fig. 1 - Schematic geological map, cross-section and sequence (Thicknesses on scale for the post-unconformity succession only) of Riu Gruppa Unit (South-East Sardinia).Post-Hercynian sediments: 1 - sandstones and conglomerates (Eocene).Gerrei Units: 2 - volcanic and sedimentary sequences of Late Cambrian to Early Carboniferous age.Riu Gruppa Unit: 3 - shales and black shales (a); marbles and calc-schists (b) (Silurian - Devonian). 4 - metarkoses, metasandstones and metasiltstones (Late Ordovician). 5 - intermediate-basic metavolcanics (a); metamorphic products of reworked original intermediate volcanics and carbonatic metagraywaches (b) (Middle Ordovician). 6 - aphiric metarhyolites (a); metavolcanoclastics and sub-aerial reworked conglomerates of acidic lavas (b) (Middle Ordovician). 7 - this symbol shows the (? Middle Ordovician) metavolcanic complex of 5 and 6 represented in the schematic geological map and in the cross-section. 8 - turbiditic metasandstones (? Late Cambrian - ? Early Ordovician). 9 - geological boundaries. 10 - tectonic contacts. 11 - faults. 12 - unconformity. 13 - cataclasites and mylonites.In the syntetic structural schetch of central-southern Sardinia: 14 - post-Hercynian sediments and volcanites. 15 - granitoids. 16 Hercynian metamorphites. 17 - metamorphites of Riu Gruppa Unit. 18 - trend of the axial surface of the Flumendosa Valley Antiform.

bedding, cross and parallel laminations. Intruded in this terrigenous sequence metadoleritic bodies occur showing WPB geochemical affinity (Di Pisa et al., 1991).

The similarity of sedimentary facies as well as the occurence of WPB metadolerites leads to correlate this deposit to metapelitic-metarenaceous S.Vito and Solanas formations referred to Middle Cambrian-Lower Arenigian on biostratigraphic basis (Barca et al., 1982, 1989 ; Albani et al., 1985; Naud & Pittau Demelia, 1985).

— A metavolcanic complex consisting of local basal aphiric metarhyolites capped by metavolcanoclastics and sub-aerial metaconglomerates due to reworking of acidic lavas; porphiritic intermediate-basic metavolcanics with phenocrysts of albitic plagioclase and carbonatic metagraywackes ended the volcanic cycle.

This metavolcanic complex rests on the previous metapelitic-metarenaceous succession along an horizon of metaconglomerates deriving from its sub-aerial reworking.The meaning of such a deposit is referable to the Sardic-Sarrabese unconformity and the age of the whole complex could be referred to the subalkaline magmatic activity (Memmi et al., 1983) occurred in a time interval spanning between Arenig and Caradoc.

— Metasandstones, mainly arkosic in composition, associated with metamicroconglomerates containing pebbles of acidic-intermediate volcanics. These rocks are interlayered with metasiltites and metapelites which are predominant toward the top of the succession . On a lithological basis, and considering their transgressive character, these metasandstones could be compared to the Caradocian-Ashgillian transgressive sediments of other units (e.g. the Pta. Serpeddì formation; Barca & Di Gregorio, 1980).

— Black metapelites and metasiltites with intercalations of black quartzites showing cross bedding. The typical euxinic facies and the position occupied by this deposit enable one to correlate it with the "black shales" which characterize the Silurian of the sardic Hercynian Chain.

— Marble and calc-schists, some times dolomitic, bearing crinoidal remnants and thin intercalations of carbonaceous metapelites, referable to the Devonian.

The pre-silurian described succession crops out as an overturned limb of an isoclinal anticline, referable to the first phase of Carmignani & Pertusati (1979), verging toward SW; wether the Siluro-Devonian terrains mainly crop in corrispondence of a contiguos sincline bearing devonian marble in the core (Fig 1). To This sincline-anticline couple minor parasitic folds are tied, pertaining to the same deformative event, which are preserved as " tête plongeante" folds in the southern flank of the Flumendosa antiform.

Asymmetric folds facing away from the top of the antiform as well as extensional composite structures (Rikkelid & Fossen, 1992), particularly developed in chlorite bearing marble, must be related to the second deformational event which developed in a non coaxial strain regime controlled by the post-collisional extension. This late tectonic event does not seem to significantly alterate the relationships among the different formations acquiered during the collision-related tectonic.

Although a study of the internal deformation of this unit is beyond the scope of this note, a crenulation cleavage, particularly pronunced near the thrusts , developed wich transposed the " S1 " slaty cleavage in the rocks of carbonatic and pelitic derivation. We suggest that this crenulation cleavage which is thought linked to the same collisional tectonic event which produced the first slaty cleavage (Carosi et al., 1991) - could have locally (close to the thrusts reactivated as low angle normal shears) undergone a new cycle of transposition during the tectonic inversion proposed by (Cappelli, 1989, 1991; Carmignani et al., 1991).

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AN EXOSCOPIC SURVEY OF UPPER ORDOVICIAN QUARTZ ARENITES IN SOUTHWESTERN SARDINIA

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ABSTRACT

Clasts belonging to quartz arenitic lithofacies of the Maciurru Member (Domusnovas Formation), of early Ashgill age (Leone et al., 1991) outcropping in southwestern Sardinia have been studied by means of a scanning electron microscope and EDS analyzer. By the EDS analysis it has been possible to determine a mineralogical association made up of quartz individuals and subordinate feldspars. The exoscopic survey, wich was carried out on clasts of sizes ranging between about 200 and 400 micron, shows irregular or subangular morphologies for the quartz clasts; their surfaces show a primitive extended alteration of pedogenetic type and several individuals show a successive partial clastation consisting of conchoidal fractures for wich a glacial origin is not to be excluded.

KEY WORDS: *Quartz exoscopy, Upper Ordovician, SW Sardinia, Glaciomarine sediments.*

Quartz clasts belonging to arenitic lithofacies of the Ordovician succession of the Iglesiente (southwestern Sardinia) were studied by a scanning electron microscope in order to check whether the hypothesis of their glacial (glaciomarine) origin, previously suggested by Cocozza et al. (1974) and Cocozza & Leone (1977), can be confirmed.

The analyses were carried out on samples taken from within the Maciurru Member of the Domusnovas Formation (Leone et al., 1991), wich corresponds to the informal unit "d" in Cocozza & Leone (Fig. 1).

The Maciurru Member is entirely made up of a succession of beds of prevalently quartzose sandstones, of a grain size that increases towards the upper part, and reaches fine conglomerates or conglomeratic coarse sandstones. In the Perda Muzza section (Domusnovas) the Member reaches a thickness of 37 metres (Fig. 2).

On the basis of the brachiopod fauna, the age of the member is entirely within the early Ashgill, and it does not go beyond the middle Ashgill either, which is documented as far as regards the overlying Punta S'Argiola Member by a trilobite and brachiopod fauna.

The samples were collected at different levels in

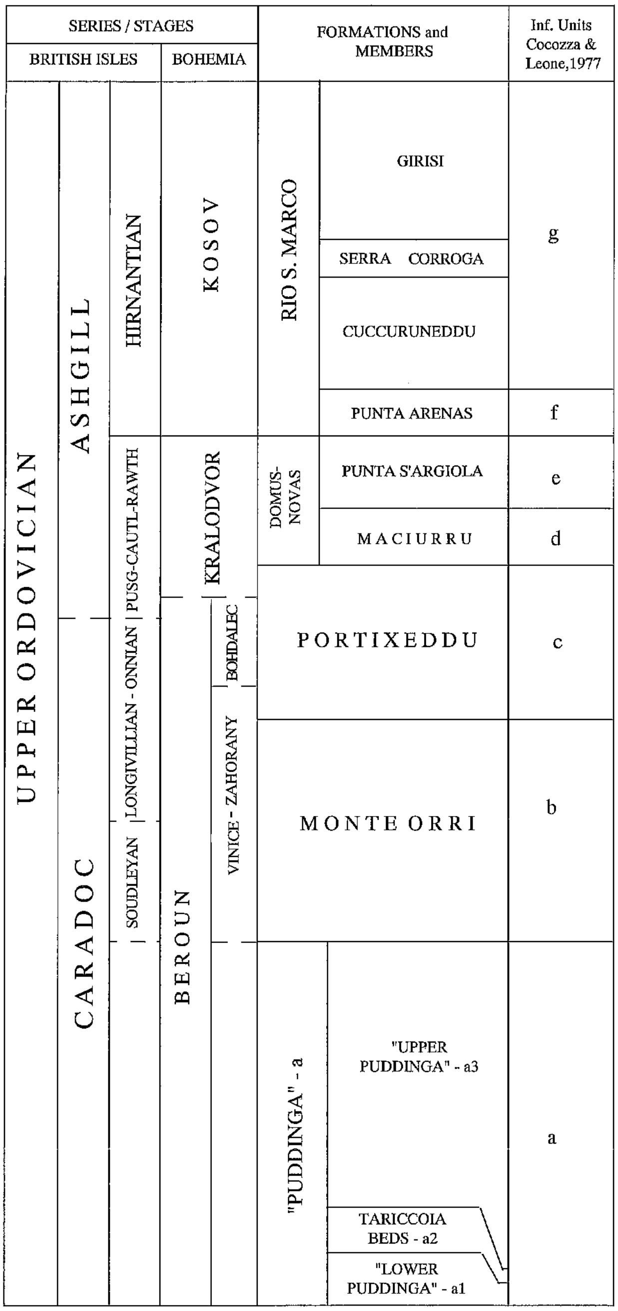

Fig. 1 - Formal lithostratigraphic units of the "post-sardic" Ordovician sequence in southwestern Sardinia (after Leone et al., 1991).

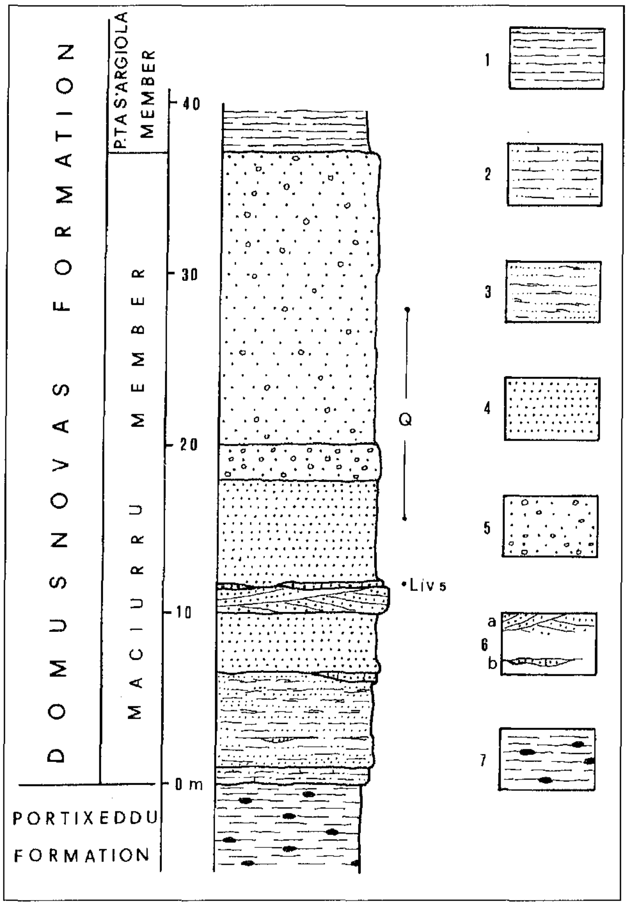

Fig. 2 - Perda Muzza section (Maciurru Member). 1: silty claystone; 2: sandy siltstone; 3: fine to medium sandstone with pelitic layers; 4: medium sandstone; 5: conglomeratic coarse sandstone; 6: a) cross bedding, b) lenticular bioclastic sandstone; 7: syltstone with nodules. Q and Liv 5 are analysed samples.

different times, on the outcrops east of Domusnovas. As shown in figure 2, the first series of samples "Q" refers to the middle-upper portion of the Member, while the second series "Liv5" refers to the lower part.

The first sampling suggested some interesting indications, which had to be interpreted with some caution, since the forms observed on the quartz clasts could have been produced by recent chemical processes or by the procedures used to separate the clasts. As a matter of fact, since these arenites do not have a carbonatic cement, except at rare levels, the first samples were taken where they appeared to be easily cleft by natural weathering of the outcrops .

The second sampling was carried out on fresh rock at a level with a significant amount of bioclasts and a weak carbonatic cement. In this case the clasts were more easily separated and the surfaces should not have been affected by any recent phenomena.

All the isolated clasts were first attacked with cold concentrated HCl. They were then submitted to the usual preparation, including ultrasound techniques. From these clasts more than 500 quartz individuals smaller than 400 micron were considered.

The exoscopic analyses were carried out by a scanning electron microscope connected to an EDS analyzer for a qualitative check of the microclasts.

The principal exoscopic characteristics observed on the surfaces of the clasts and the chronological evolution of the different ways of chemical and mechanical processes, that, moreover, proved common to the two series of samples, are summarized in the following:

a) irregular shape with surfaces presenting intensive chemical processes and partial recrystallization of silica, that can be attributed to pedogenesis, as defined and illustrated by Le Ribault (1977), probably in conditions of hot-wet climate;

b) about 10-15% of the clasts presents superimposed, partial, well-defined conchoidal fractures of different sizes and in broken relief. They originated from processes of clastation that certainly occurred after the above mentioned pedogenetic phenomena. They are generally unaltered surfaces that often present sub parallel, "marche d'escalier" type, "cisaillement" figures (Le Ribault, op. cit.). The less frequent figures produced by friction and the traces of sub circular grinding seem to point out a rotational component in the dynamic process that produced clastation. On the basis of literature reports (Krinsley & Doornkamp, 1973; Le Ribault, 1977), these figures are to be genetically related to glacial environment;

c) a subsequent, less intense mechanical action probably due to a non-prolonged transport in water environment, seems to have partially and locally rounded off the more pronounced outer edges of the fractures under b).

On the whole, the exoscopic characters and the phenomenological evolution of the different figures observed on the surfaces of the examined quartz microclasts are very similar to those described by Hamoumi et al. (1981) in the "Schistes du Cosquer", an Upper Ordovician glaciomarine formation of the Armorican Massif.

In the samples belonging to the "Q" series, besides quartz, microclasts made up of Al silicate (andalusite ?) were found, with the same exoscopic characteristics as those described for the quartz individuals.

In the samples belonging to the "Liv 5" series, besides quartz, rare zircon, rutile and tourmaline individuals, a small percentage of subspherical, elongated crystal individuals was observed. Qualitative analyses on a scanning electron microscope showed that these crystals are made up of intensely sericitized alkali feldspar.

The analyses carried out seem to corroborate the possibility that many of the clasts from these arenites are of glacial origin. There are various reasons for not entirely accepting the conclusion that the Maciurru arenites are to be considered a glaciomarine deposit.

The data obtained analyzing the European "glacial" deposits, suggest that the maximum of the glaciation at the end of the Ordovician is totally within the Hirnantian stage (Brenchley, 1988; Robardet & Doré 1988), but the Maciurru arenites are slightly older. The Hirnantian stage on the other hand is well documented by the overlying Rio San Marco Formation, where clear signs of its glaciomarine origin have been found (Leone et al., op. cit.).

Moreover, the possibility of relating the Maciurru arenites to a regressive (glacio-eustatic) phase does not find any confirmation in the paleogeographic reconstruction of either the Maciurru Member or the units including it. There is, in fact, no evidence whatsoever of a regressive phase in the areas considered most proximal (Western Iglesiente).

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BENTHIC PALEOCOMMUNITIES OF THE MIDDLE-UPPER MIOCENE LITHOSTRATIGRAPHIC UNITS FROM THE CAGLIARI HILLS (SOUTHERN SARDINIA, ITALY)

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Abstract

In the present paper, the first ecobiostratigraphic data regarding the Middle-Upper Miocene lithostratigraphic units from the Cagliari area are described. The basal bed of the investigated successions is lithologically made of clay marls and sandy marls ascribed to the Fangario Formation and well exposed in quarry fronts of the NW Cagliari periphery.

From bottom to top the following lithostratigraphic units have been described: “Arenarie di Pirri”, “Pietra Cantone”, “Tramezzario” and ”Pietra Forte”.

The first one is included in the "Areanarie di Pirri" Formation ; while the other three units are ascribed to the Cagliari Limestone Formation.

The “Arenarie di Pirri” (calcarenitic sandstone), Lower-Middle Serravallian in age, contain a macrofauna that is equivalent to that presently characterizing the infra-circalittoral zone.

The “Pietra Cantone” (sandy calcarenite), ascribed to the Tortonian-Lower Messinian Age, turned out to belong to a considerably deeper environment.

This unit is characterized by a lithological uniformity and a benthic population stability that seems to be similar to the populations presently living within the Circalittoral and the Bathyal.

For the uppermost portion, which is deeply bioturbated, an interruption in sedimentation can be hypothesized. The passage to the next unit occurs with an abrupt lithological change.

The “Tramezzario” (calcarenites and limestones often in lateral heteropy) related to the Lower (?)Messinian age, is characterized in the lower stratigraphic portion by the instability of the sedimentation basin. This is documented by slumping, discontinuity surfaces, synsedimentary faults and angular unconformities. The depositional environment is typical of the circalittoral zone.

The “Pietra Forte” (bioclastic, massive and *Lithothamnium* limestone), Messinian in age, lies over the previous unit with a striking unconformity. The depositional environment for the middle lower portion of the sequence can be referred to the circalittoral zone.

The “Pietra Forte” contains paleopopulations of a less deep environment.

Key words : *Stratigraphy, Paleoecology, Middle- Upper Miocene,Sardinia, Italy*

Riassunto

In questa nota vengono proposti i primi risultati delle ricerche sulla ecobiostratigrafia delle unità formazionali del Miocene medio-superiore affioranti nelle colline e lungo la fascia costiera della città di Cagliari.

Il letto delle successioni esaminate è costituito da marne argillose o sabbiose della Formazione di Fangario esposte in fronti di cava ubicate nella periferia NW di Cagliari. Le successioni comprendono le seguenti unità litostratigrafiche: “Arenarie di Pirri”, “Pietra Cantone”, “Tramezzario” e “Pietra forte”. La prima rientra nella Formazione delle Arenarie di Pirri; le altre tre nella Formazione del Calcare di Cagliari.

Le "Arenarie di Pirri" (arenarie calcarenitiche), di età serravalliana, hanno fornito una macrofauna equivalente a quella che attualmente caratterizza l’Infra-Circalitorale.

Di ambiente strettamente più profondo si è rivelata la Pietra Cantone (calcarenite arenacea), attribuita al Tortoniano- Messiniano inferiore. Essa è caratterizzata da uniformità litologica e da stabilità dei popolamenti bentonici che appaiono analoghi a quelli viventi attualmente nel Circalitorale e nel Batiale più alto.

Per la parte sommitale, intensamente bioturbata, si ipotizza un’interruzione della sedimentazione. Il passaggio all’unità successiva avviene con brusco cambiamento litologico.

Il Tramezzario (calcareniti e arenarie marnose spesso in eteropia laterale), riferito al Messiniano inferiore(?), è contraddistinto nella parte stratigraficamente più bassa da instabilità del bacino di sedimentazione documentata da slumping, superfici di discontinuità, faglie sinsedimentarie e discordanze sull’unità precedente. L’ambiente di

sedimentazione per la porzione medio- bassa della sequenza, è riferibile al Circalitorale . La parte sommitale della "Pietra Forte", di età messiniana, contiene paleopopolamenti di ambiente meno profondo.

1. Introduction

The Miocene marine formations of the Cagliari area are well- known in literature with the following informal terms: "Arenarie di Pirri", "Pietra Cantone", "Tramezzario" and "Pietra Forte". These lithostratigraphic units have not been properly studied from a paleobiological and paleoecological point of view, apart from the "Arenarie di Pirri" that have already been defined in details by Cherchi (1974). The present work contains the first results of a series of researches presently taking place at The University of Cagliari, Earth Science Department. They are part of the studies carried out by the "Gruppo Nazionale Paleobenthos" (National Paleobenthos Group) aiming at highlighting the ecobiostratigraphic features of such deposits. They refer to chronostratigraphic interval included between the Serravallian (Zone NN6 of Martini, 1971) and the Messinian (Cherchi, 1974; Cherchi & Tremolieres, 1984; Cherchi, 1985).

The transitions among the different units are all discontinuous and marked by sedimentation hiatuses whose chronostratigraphic intervals are impossible to define for the planktonic microfaunas scarcity, and/or by unconformity surfaces to which the erosion phases are likely to be related.

2. Litho-Biostratigraphy

2.1 *"Arenarie di Pirri"*

They lie on the Upper Langhian-Lower Serravallian "Argille del Fangario" (Pecorini & Pomesano Cherchi, 1969; Cherchi, 1974; Robba & Spano, 1978; Cherchi, 1985). They consist of some alternances of extremely fine and silty sandstones and of medium and/or coarse-grained sandstones(levels 1-6 and 1-3 of the sequences respectively outcropped at Cala Fighera and Cala Mosca (Fig.1). The rocks are mainly inconsolidated and well stratified, greyish-green sometimes verging on yellow. The stratification is generally uniform and it presents same unrhythmical alternances of unconsolidated and compact layers with variously evident contacts. The uppermost part of the unit generally verges on even coarser lithologies that are poorly cemented, scarcely stratified and beige-yellow in colour. It is characterized by the presence of abundant macrofossils that are often found in internal mold condition. The thanatocoenosis includes *Mollusca, Echinoidea, Crustacea* and Algae. *Turritella (Haustator) triplicata* (Brocchi) prevail among the *Gastropoda*, while the most common species among the *Bivalvia* are: *Flabellipecten fraterculus* (Sowerby), *Flabellipecten solarium* Lamarck, *Pecten aduncus* Eichwald, *Anomia (Anomia) ephippium* Linneo, *Anomia (Anomia) ephippium rugolosostriata* (Brocchi), *Cubitostrea frondosa* (De Serres).

The attribution to the Serravallian age is justified by the existence of planktonic *Foraminifera* (Cherchi, 1974) such as: *Orbulina suturalis* Bronn, *Orbulina universa* D'Orbigny, *Globorotalia mayeri* Cushman & Ellis, *Globorotalia praemenardii* Cushman & Stainforth, *Globoquadrina altispira* (Cushman & Jarv). (Cherchi, 1974).The preliminare data regarding a study about calcareous Nannofossils carried out by A. Negri (The University of Bologna), particularly allow to reffer the basal part (level 1) of the "Arenarie di Pirri" contained within the sequence outcropping Cala Mosca (Fig. 1), to the lower middle Serravalian age (zone NN6 of Martini's scale, 1971). It is not to be excluded, there fore, a Middle Upper Serravallian age for the stratigraphic higher part.

2.2 *"Pietra Cantone"*

This unit is placed between the "Arenarie di Pirri" and the "tramezzario".

It is beige-yellow, scarcely compact and generally without stratification. The top is characterized by extremely abundant flat-lying Ichnofossils, tracks and galleries, whose diametres are included between 1-3 cm. In the Cala Fighera section the "Pietra Cantone" is represented by level 8; by the basal levels within the sections 3-6 of Faro S. Elia; by the basal levels within the sections of Sa Spiaggiola and Torre Perdusemini (Fig. 1). A considerable paleontological component can be found in the internal mold condition. *Bivalvia* clearly prevail over *Gastropoda*, *Echinoida* and Bryozoa. Among the *Bivalvia*: *Amussium denudatum* Reuss, *Pecten (Flabellipecten) koheni* Fuchs, *Flabellipecten burdigalensis* (Lamarck), *Chlamys (Aequipecten) malvinae* Dubois, *Pecten (Gigantopecten) latissimus* (Brocchi), *Arca (Arca) tetragona* Poli, *Anadara (Anadara) pectinata* Brocchi, *Anomia (Anomia) ephippium* Linneo, *Palecyora (Palecyora) islandicoides* (Lamarck). Among the *Gastropoda: Cassidaria (Cassidaria) echinophora* Linneo, *Cassis mamillaris* Grateloup, *Conus dujardini* Deshayes, *Ficus (Ficus) conditus* Brongniart.

The basal portion of this unit is assigned by Cherchi (1974) and Cherchi & Tremolieres (1984) to the Tortonian age for the presence of: *Globigerina apertura* Cushman, *Globigerina microstoma* Cita, Premolisilva & Rossi, *Globigerina nepenthes* Todd, *Globorotalia acostaensis* Blow, *Globorotalia lenguaensis* Bolli, *Globorotalia menardii* (D'orbigny), *Globigerinoides obliquus* Bolli, the higher part instead is attributed, by the Authors themselvs, to the lower Messimian age for presence of species such as , *Globigerina nepenthes* Todd, *Globorotalia acostaensis* Blow, *Globorotalia menardii* (D'orbigny), *Globorotalia merotumida* Blow & Banner, *Globorotania mediterranea* Catalano & Sprovieri, *Globorotalia miozea saphoae* Bizon e *Globigerinoides obliquus extremus* Bolli & Bermudez.

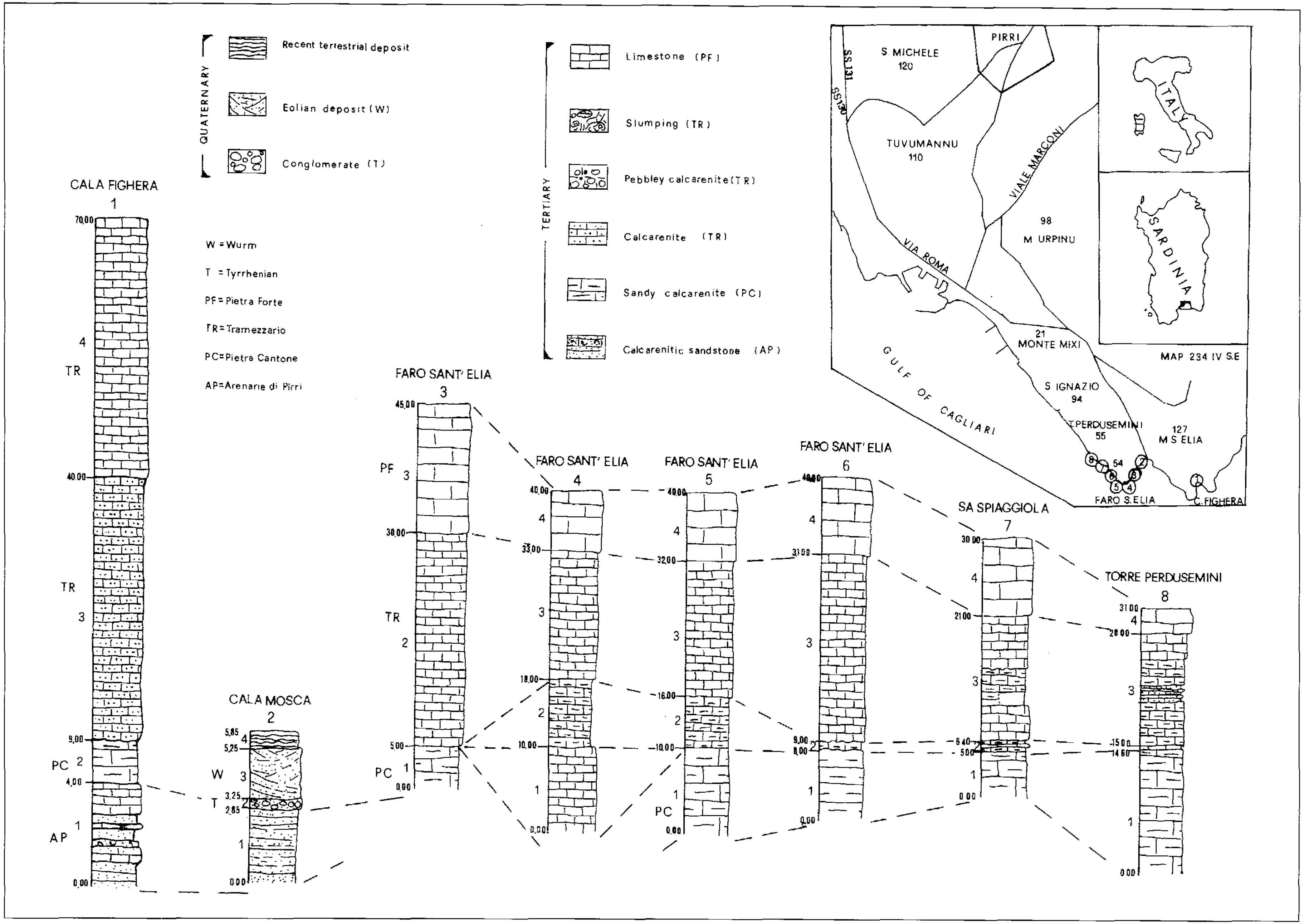

Fig. 1 - Location and biostratigraphic correlation of mapped section

2.3 *"Tramezzario"*

This unit outcrops in all the examined sections. It is placed between the "Pietra Cantone" (bottom) and locally, the "Pietra Forte" (top). It is found within levels 9 and 10 of Cala Fighera; level 2 of Faro S. Elia (3); levels 2 and 3 of Faro S. Elia (4) and (5); levels 2-4 of Faro S.Elia (6); levels 2-6 of Sa Spiaggiola and Torre Perdusemini (Fig. 1).

It is the product of a purely carbonate sedimentation.

The lithology is represented by a medium and coarse biocalcarenite, It is whitish and generally scarcely consolidated. Also, it is well stratified on the whole, sometimes in lateral heteropy with a whitish limestone that is compact and well stratified.

The paleontological content is given by a mostly Mollusk macrofauna. *Bivalvia* are extremely abundant as well as Echinoidea and red Algae, while *Gastropoda* are rarely found. Among the Bivalvia it has been collected: *Chlamys (Aequipecten) macrotis* Sowerby, *Chlamys (Aequipecten) malvinae* Dubois, *Flabellipecten solarium* (Lamarck), *Flabellipecten planosulcatus* (Matheron), *Flabellipecten fraterculus* (Sowerby), *Pecten (Gigantopecten) albinus* Von Teppner, *Pecten (Gigantopecten) latissimus* Brocchi, *Ostrea (Ostrea) edulis lamellosa* Brocchi, *Ostrea (Saccostrea) forscklii* Chemnitz, *Anomia (Anomia) ephippium* Linneo, *Anomia (Anomia) ephippium helvetica* Mayer, *Anomia (Anomia) ephippium costata* Brocchi.

The stratigraphic relationships suggest an age still messinian, probably lower, for the "Tramezzario" by considering what it has been said about the "Pietra Cantone" and also, because the latest is referred to the Messinian Age an the basis of what is affirmed in the following paragraph.

2.4 *"Pietra Forte"*

The "Pietra Cantone" is not always found within the studied area. It lays over the "Tramezzario" and characterizes level 3 of the Faro S. Elia (3) section, levels 4 of Faro S. Elia (4) and (5); level 5 of Faro S. Elia (6); level 7 of Sa Spiaggiola and Torre Perdusemini (Fig. 1).

The unit terminal part of the Miocene succession consists of a calcareous bioclastic lithology, whitish, extremely compact, massive rich mainly of *Lithothamnium sp.*. Within the "Pietra Forte" some unconformity surfaces and a breccia facies mixed with paleosoil, are recognizable. The macrofauna is extremely abundant although it is mainly represented by internal molds. It consists of *Bivalvia, Gastropoda*, Echinoidea, Coral and red Algae. Among the *Bivalvia: Lutraria (Lutraria) oblonga* Chemnitz, *Cardita (Cardita) jouanneti* Basterot, *Glycymeris (Glycymeris) insubrica* (Brocchi), *Glycymeris (Glycymeris) bimaculata* (Poli), *Venus (Ventricoloidea) multilamella* (Lamarck), *Perna soedanii* Deshayes, *Lithodomus appendiculatus* Philippi, *Lithodomus compressus* Menke. Among the *Gastropoda*: *Conus dujardini* Deshayes, *Conus ponderosus* Brocchi, *Murex (Muricantha) trunculus* Linneo. Among the red Algae: *Lithothamnium* sp.

The regional geological frame and recent studies by Cherchi et al.(1974), and Cherchi et al. (1978 b), suggest an evolutive parallelism occuring between the Cagliari Hills and the Sinis (middle-western Sardinia) series.

Cherchi et al. (1978) believes that these areas, as marginal parts of the meditteranean basin during the Miocene, were influenced, since the Messinian, by the regression of the deepest parts of the basin. The "Pietra Forte" should, therefore, be included within the Lower Messinian. According to the *Pectinida* scale proposed by Demarcq(1990), the *Pectinida* association found in this unit is unconsistent with an age more recent than the Messinian.

3. PALEOENVIRONMENTAL CONSIDERATIONS

3.1 *"Arenarie di Pirri"*

In the present unit, considered almost sterile by the previous Authors, a noteworthy macrofauna has been collected. It comes from its middle-upper part. The physical features of the sterile sediment are indicative of a littoral-infralittoral sandy environment. The Mollusk associations are essentially autochthonous although they present a certain degree of trasportation which is limited to their life environment. This transportation is results from the occurence of some species that are widely known for being related to the presence of fluvial materials and independent from the bathymetry. A low degree of transportation is also documented by the presence of thin and delicate shells such as *Anomia (Anomia) ephippium* and its subspecies (Tab. 1). They are still in a good state of preservation. The forms generally result consistent with the lithology they are incorporated in.

The thanatocoenosis found in this unit are widely represented by species and subspecies that are recognized, in literature, to belong to an Infra-Circalittoral enviroment widespread in the Infra-Circalittoral, such as: *Anomia (Anomia) ephippium* Linneo, *Anomia (Anomia) ephippium costata* Brocchi, *Anomia (Anomia) ephippium radiata* Pantanelli, *Anomia (Anomia) ephippium sulcata* Poli and by species typical of the Circalittoral such as: *Pecten (Flabellipecten) fraterculus* (Sowerby), *Pecten (Pecten) aduncus* Eichwald, *Pecten (Pecten) benedictus* Lamark, *Lutraria (Lutraria) oblonga* Chemnitz. This group set of species shows a paleobiocoenosis with a paleobiocoenotic meaning that is put in relation with the one concerning the present mediterranean associations of the Coastal Detritic zone (DC), typical of the Circalittoral.

The noteworthy abundance of *Turritellidae* (the following species have been identified: *Turritella (Haustator) desmarestinus* (Basterot), *Turritella*

TAXA	WATER COLUMN	BATHYMETRIC MEANING
GASTROPODA		
Diodora sp.		
Protoma (P.) cathedralis (Brongniart)		
Semicassis cf. subsulcosa Höern & Aning		
Turritella (H.) desmarestinus (Basterot)		
Turritella (H.) triplicata (Brocchi)		
Turritella (H.) tricarinata (Brocchi)		
BIVALVIA		
Anomia (A.) ephippium Linneo		
Anomia (A.) ephippium radiata Pantanelli		
Anomia (A.) ephippium rugolosostriata Brocchi	30 - 50 m	INFRALITTORAL (?) — CIRCALITTORAL
Anomia (A.) ephippium sulcata Poli		
Chlamys (A.) spinulosa (Münster)		
Cubitostrea frondosa (De Serres)		
Flabellipecten fraterculus (Sowerby)		
Flabellipecten cf. planosulcatus (Matheron)		
Flabellipecten solarium (Lamarck)		
Lutraria (L.) oblonga Chemnitz		
Modiolus sp.		
Pecten (P.) aduncus Pusch		
Pecten (P.) benedictus Lamarck		

Tab. 1 - Dominant thanatocoenosis and paleobathymetric meaning of the "Arenaria di Pirri".

(Haustator) triplicata Brocchi, *Turritella (Turritella) tricarinata* (Brocchi) indicates the settlement of an important hydrographical network around the sedimentation basin, presumably related to the sea level negative changes. The *Turritella* banks are presently linked to important stream flows in the Mediterranean Sea and they generates the Terrigenous Coastal Muds (VTC) associations.

The information just given suggest that the fossiliferous deposit took place at a maximum depth of 40-50 meters, while a bathymetry of about 30 meters is likely to have occured for the sterile deposits.

3.2 "*Pietra Cantone*"

The fossil communities essentially show conditions of autochthony and a good consistency with the sediment characteristics. A transportation can be hypothesized for the *Conidae* (Tab 2) ascribed to a bathymetry included between 0-30 m by Hall (in Davoli, 1972, p. 66). As far as the *Glycymeridae* are concerned, at the present time they are independent from the bathymetric zone for the most part. Also, they are widespread in the Infralittoral where they are transported from the environment hydrodynamism.

The thanatocoenosis belonging to the external shelf is documented by the prevailing abundance of species belonging to biocoenoses homologous to those presently living within the Bathyal the Mediterranean sea such as: *Ficus (Ficus) conditus* (Brongniart), *Amusium cristatum* (Bronn) or well-known fossils of analogous paleocommunities, such as: *Chlamys (Aequipecten) spinulosa* (Munster) and *Capulus (Capulus) hungaricus* (Linneo). A further rich group of forms belonging to the present Coastal Detritic, or the paleobiocoenoses that can be assimilated to it, confirm the Circalittoral nature of the whole thanatocoenosis. Among them there is: *Arca (Arca) tetragona* Poli; *Chlamys (Aequipecten) malvinae* Dubois; *Flabellipecten koheni* Fuchs; *Modiolus (Modiolus) adriaticus* (Lamarck).

For this unit, a bathymetry of about 60-80 m can be hypothesized on the basis of the highlighted malacological content and for the presence of numerous single corals typical of depth zone.

TAXA	WATER COLUMN	BATHYMETRIC MEANING
GASTROPODA		
Capulus (C.) hungaricus (Linneo)		
Cassidaria (C.) echimophora Linneo		
Cassis mammilaris Grateloup		
Conus antediluvianus Bruguiere		
Conus dujardini Deshayes		
Conus tarbellianus Grateloup		
Ficus (F.) conditus (Brongniart)		
BIVALVIA		
Amussium cristatum (Bronn)	60 - 80 m	CIRCALITTORAL
Amussium denudatum (Reuss)		
Anomia (A.) ephippium Linneo		
Arca (A.) tetragona Poli		
Chlamys (A.) malviane Dubois		
Chlamys (A.) spinulosa (Münster)		
Flabellipecten burdigalensis (Lamarck)		
Flabellipecten koeni (Fuchs)		
Glycymeris (G.) insubrica (Brocchi)		
Glycymeris (G.) fichteli (Deshayes)		
Modiolus (M.) adriaticus (Lamarck)		
Pecten (G.) latissimus Brocchi		
Pelecyora (P.) islandicoides Pantanelli		

Tab. 2 - Dominant thanatocoenosis and paleobathymetric meaning of the "Pietra Cantone".

3.3 *"Tramezzario"*

The paleontological content turned out to be totally autochthonous. *Pectinidae* and *Ostreidae* occur with both valves. There is no evidence of broken or selected faunas, or faunas inconsistent with the sediment characteristics. A terrigenous lithofacies, with a malacologic fauna, was found within the unit. This fauna is sufficiently differentiated and referable to a sedimentation depth that is higher than that of the most widespread lithology. The following species have been collected: *Amussium denudatum* (Reuss); *Cubitostrea frondosa* (De Serres); *Chlamys (Chlamys) varia* Linneo.. The thanatocoenoses with a biocoenotic meaning analogous to the biocoenoses of the present Coastal Detritic, belong to this unit. In abundance order there are: *Chlamys (Aequipecten) macrotis* Sowerby; *Paphia (Paphia)* cf. *vetula* (Basterot); *Pecten (Gigantopecten) albinus* Von Teppner; *Pecten (Gigantopecten)* cf. *holgeri* Geignitz; *Flabellipecten fraterculus* (Sowerby); *Lutraria (Lutraria)* cf. *oblonga* Chemnitz. The Tramezzario seems to have been deposited at a maximum sea depth of around 40 m.

3.4 - *"Pietra Forte"*

The stratigraphically lower part of this unit consists of a rich fossil association with a predominance of large *Pectinidae* with still articolated valves. It is often possible to notice also the early ontogenetic phases of the growth . The paleoassociations that can be recognized in the lower part have a biocoenotic meaning that can be referred to the present Coastal Detritic biocoenoses. The following species prevail (Tab 4): *Pecten (Gigantopecten) albinus* Von Teppner; *Pecten (Gigantopecten) latissimus* Brocchi; *Chlamys (Aequipecten) macrotis* (Sowerby); *Flabellipecten calaritanus* (Meneghini); *Flabellipecten fraterculus* (Sowerby); *Flabellipecten planosulcatus* (Matheron); *Flabellipecten solarium* (Lamarck); *Lutraria (Lutraria) oblonga* Chemnitz; *Modiolus (Modiolus) adriaticus* (Lamarck).

The upper part instead shows a thanatocoenosis characterized by the extraordinary abundance of red Algae (Lithothamnium sp.) and by the high rate of Mytiloidae and "Lithodomus". They can be related to the littoral and infralittoral zones because of the existence of living species having their life environment optimum in bathymetrical zones related to a relevant hydrodynamism and to a coarse-grained sandy and gravel substratum, such as: *Glycymeris (Glycymeris) bimaculata* (Poli), *Glycymeris (Glycymeris) insubrica* (Brocchi) and *Lithodomus compressus* Menke. To the same paleocommunity it is possible to ascribe the fossil species *Glycymeris fichteli* (Deshayes) and *Lithodomus appendiculatus* Philippi. The latter is signalled by the Authors in the littoral zone, in high energy environment.

The maximum paleobathymetry for the whole unit is likely to correspond to about 30 m.

TAXA	WATER COLUMN	BATHYMETRIC MEANING
BIVALVIA		
Amussium denutatum (Reuss)		
Anomia (A.) ephippium Linneo		
Anomia (A.) ephippium costata Brocchi		
Anomia (A.) ephippium elvetica Mayer		
Chlamys (A.) malvinae (Dubois)		
Chlamys (A.) macrotis (Sowerby)		
Chlamys (A.) seniensis (Lamarck)		
Chlamys (C.) varia (Linneo)		
Cubistostrea frondosa (De Serres)	20 - 40 m	INFRALITTORAL – CIRCALITTORAL
Flabellipecten fraterculus (Sowerby)		
Flabellipecten planosulcatus (Matheron)		
Flabellipecten solarium (Lamarck)		
Lutraria (L.) cf. oblonga Cheminitz		
Ostrea (O.) edulis lamellosa Brocchi		
Ostrea (S.) forsckalii Chemnitz		
Paphia (P.) cf. vetula (Basterot)		
Pecten (G.) albinus Von Teppner		
Pecten (G.) latissimus gibboplanus (Sacco)		
Pecten (G.) latissimus nodosiformis (Sacco)		
Pecten (G.) cf. holgeri Geinitz		

Tab. 3 - Dominant thanatocoenosis and paleobathymetric meaning of the "Tramezzario".

4. Stability and Instability

In the investigated sedimentation basin, it has been highlighted how sharp the passage among the different units is, excluded the boundary between the "Arenarie di Pirri" (Lower-Middle Serravallian age) and the "Pietra Cantone" (Tortonian-Lower Messinian) where sufficiently distinguished lithologies take place although a certain gradual transition occurs. The absence of gradual transition among the different units is also supported by sharp changes in the paleobiocoenotic and paleobathymetrical meaning of the fossil communities. Instability of the sedimentation basin variously marked, corresponding to these changes, can be hypotesized. We believe that these intabilities are believed to be related to the tectonic activity.

An extremely sharp discontinuity can be noticed at the contact between the "Pietra Cantone" and the "Tramezzario". It has been referred to a sedimentation hiatus regarding a great decrease in sedimentation that must have been associated to an erosion phase (Cherchi & Montadert , 1984; Cherchi, 1985). The "Pietra Cantone" upper part is characterized by an accentuated bioturbation, while erosion surfaces, angular unconformities and slumping can be clearly noticed in the "Tramezzario" basal part.

TAXA	WATER COLUMN	BATHYMETRIC MEANING
GASTROPODA		
Conus dujardini Deshayes		
Conus ponderosus Brocchi		
Murex (M.) trunculus Linneo		
BIVALVIA		
Cardita (C.) jouanneti Basterot		
Chlamys (A.) macrotis (Sowerby)		
Glycymeris (G.) bimaculata (Poli)		
Glycymeris (G.) fichteli (Deshayes)		
Glycymeris (G.) insubrica (Brocchi)	10-30 m	INFRALITTORAL — CIRCALITTORAL
Lithodomus appendiculatus Philippi		
Lithodomus compressus Menke		
Lutraria (L) oblonga Chemnitz		
Modiolus (M.) adriaticus (Lamarck)		
Flabellipecten calaritanus (Meneghini)		
Flabellipecten fraterculus (Sowerby)		
Flabellipecten planosulcatus (Matheron)		
Flabellipecten solarium (Lamarck)		
Pecten (G.) albinus Von Teppner		
Pecten (G.) latissimus Brocchi		
Perna soldanii Deshayes		
Venus (V.) multilamella (Lamarck)		

Tab.4 - Dominant thanatocoenosis and paleobathymetric meaning of the"Pietra Forte".

At the "Tramezzario (Lower(?) Messinian age)" - "Pietra Forte (Messinian age)" boundary, another clear discontinuity can be found.

The autoctonous thanatocoenoses regarding these two lithostratigraphic units present similar biocoenotic meanings and show a decreasing paleobathymetric trend. In this context, therefore, a short stand in sedimentation with a low sea level change can be assumed. This is also suggested by the different chemical and physical depositional conditions of the facies; in fact, they change from calcarenitic-calcareous facies ("Tramezzario") to decidedly massive facies ("Pietra Forte").

Given the autoecological and paleoecological data about the examined malacofaunas and the physical features of the sediment, it can be assumed that the formation of the Epibathyal in meaning and "Argille de Fangario", Langhian-Lower Serravallian in age, is followed by a regressive episode represented by the "Arenarie di Pirri". It was presumably deposited at a depth of 20-30 m and is Lower Middle Serravallian in age. The phases of maximum depth of the basin accurred during the deposition of the Tortonian-Lower Messinian ("Pietra Cantone") and of the Lower(?) Messinian ("Tramezzario"). A sedimentation depth respectively of 60-80 and 40 meters is attributed to them. A sea level decrease occurs when it corresponds to the sedimentation hiatus recognized at the boundary between the two units. The following "Pietra Cantone" deposition, Lower Messinian in age, which took place at a bathymetry of around 30 m, is also related to the sea level decrease.

5. Conclusion

The detailed stratigraphic survey of numerous Middle and Upper Miocene sequences exposed in the Cagliari urban area, permitted to define the lithological, preliminary paleoecological and depositional features of the different units constituting the Miocene units: "Arenarie di Pirri", "Pietra Cantone", "Tramezzario" and "Pietra Forte".

The country side survey pointed out the existence of sedimentation hiatuses probably associated with erosional phases corresponding to the contacts between the "Arenarie di Pirri" and "Pietra Cantone", the "Pietra Cantone" and "Tramezzario" and between the latter and the "Pietra Forte".

Synsedimentary faults have been recognized in the "Tramezzario", where slumping and unconformity surfaces also exist. The whole sequence was interested by a post-Lower Messianian tectonic phase.

The data coming from the study of the macrofossiliferous content and the sediment character of the different units permitted to reconstruct to the paleobiocoenotic and paleobathymetric meaning of the deposits.

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STORM DEPOSITS (PLACERS AND RHYTHMITES) IN THE CARADOCIAN TRANSGRESSIVE SEDIMENTS OF THE SARRABUS AREA (SE SARDINIA - ITALY)

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Abstract

The Caradocian to Ashgillian Punta Serpeddì Formation, outcropping in the Sarrabus area, is clearly transgressive and overlies Ordovician volcanic known as "Porfidi grigi". The coastal environment in the lower part, gradually evolved to an offshore environment in the upper part. The sedimentation appears to have been controlled by high energy mechanisms related to storm phenomena. Placers and rhythmites are typical of these deposits.

Key words: *Storm deposits, Placers, Ordovician, Sardinia.*

1. Introduction

This paper is devoted to the first results of a new stratigraphical and sedimentological study of the Caradocian to Ashgillian Punta Serpeddì Formation (Barca & Di Gregorio, 1980), outcroping in the Sarrabus area (SE Sardinia). Several sections have been sampled and detailed geochemical and petrographical studies are in progress in this area.

The Punta Serpeddì Formation is clearly transgressive ("trasgressione caradociana" Auct.) upon pre-Caradocian volcanic known as "Porfidi grigi", (Calvino, 1967). This formation contains a rich benthic fauna (Brachiopods, Bryozoans, Crinoids, Trilobites, Gastropods, etc.) Upper Ordovician in age (Barca & Di Gregorio, 1979; Giovannoni & Zanfrà, 1979).

From the stratigraphical and sedimentological points of view, the most significative sections are located near the city of Dolianova (Sa Murta section) and near Punta Serpeddì (Bruncu Spollittu section) (fig. 1).

2. Facies Description and Interpretation

The sections can be divided into three parts on the basis of granulometric and sedimentological characteristics.

2.1. *Lower Part*

The lower part of the formation (from about 20 to 60 m) comprises decimetric to metric beds of coarse-grained and microconglomeratic sandstones (lithic greywackes and subordinate feldspathic greywackes) (Loi et al.,1991b). A noteworthy feature is the presence of heavy mineral layers (placers) mainly constituted by zircon, rutile and tourmaline.

Various sedimentary structures are present such as graded bedding, low angle cross laminations and parallel laminations, relative to high energy environments (Harms et al., 1982). The coarse-grained lower part of the formation can be related to a nearshore deposit.

2.2. *Middle Part*

Upwards, the formation continues with rhythmical deposits (1 to 4 m thick), constituted by millimetric to centrimetric alternating layers of siltstones and well sorted conglomerates.

In the conglomerates the pebbles are very well rounded and their size varies from 3 mm to 4 cm . Pebbles are mainly constituted by quartz and by volcanic and quartzitic rocks.

Thin heavy mineral placers are frequent in this facies and are located between siltstone and conglomerate layers.

The good sorting of conglomerate is related to high energy traction processes on the sea bottom which are evidenced by the presence of erosional grooves at the upper part of siltite layers. These layers are the result of the decantation of fine-grained material suspended in agitated water.

Storm sequences (hummocky cross stratification) are present in this transitional environment, but they are incomplete because of the lack of the argillaceous fraction in the sediment.

In the whole Sarrabus area these rhythmical sediments are always located between the coarse-grained lower part of the formation and the upper part, mainly constituted by siltstone; they can be used as a lithostratigraphic reference and can be related to a storm dominated environment.

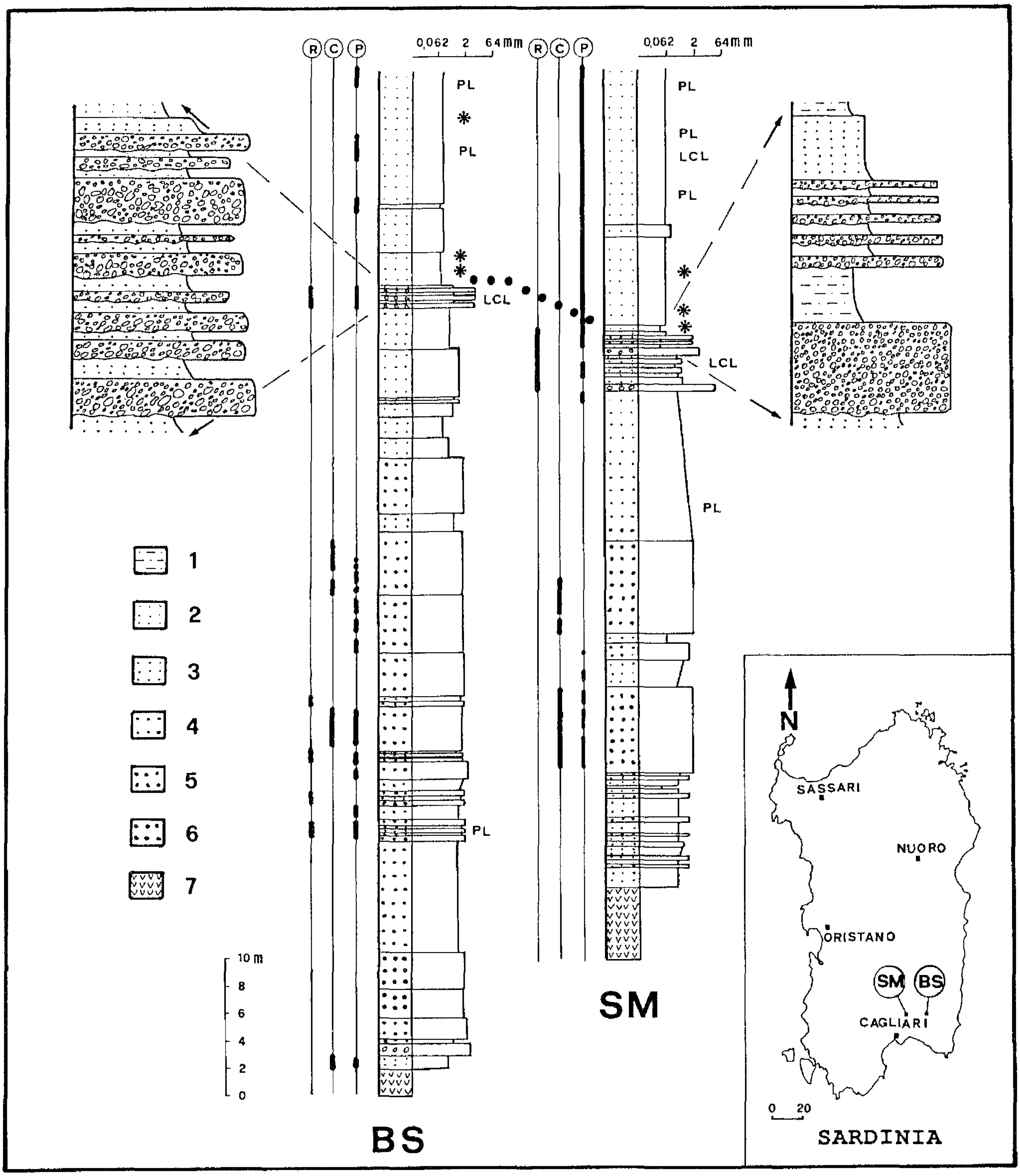

Fig.1: The Punta Serpeddì Formation in the Sa Murta section. SM: Sa Murta section; R: rhythmites; C: cross bedding; P: placer; PL: plane lamination; LCL: low angle cross lamination; *: fossils. 1-6; PuntaSerpeddì Formation. *1: micaceous siltite; 2: siltstone; 3: sandstone; 4: coarse grained sandstone; 5: microconglomeratic sandstone; 6: conglomerate.*
7: "Porfidi grigi".

2.3. *Upper Part*

The upper part of the Punta Serpeddì Formation is mainly constituted by fossiliferous (Caradoc-Ashgill) siltstones and fine-grained sandstones (lithic greywackes and subordinate feldspathic greywackes).

The thickest and most numerous heavy mineral placers are located in this upper part of the formation and locally (near Dolianova) these placers are the main sedimentary feature.

Usually, placers are constituted by millimetric parallel laminae, exclusively constituted by zircon, monazite, rutile, ilmenite and tourmaline; they are interlaminated with quartz-siltstone laminae with a high concentration of the same heavy minerals.

High energy and decantation processes result in bed surfaces underlined by bioclastic accumulations (Brachiopods, Crinoids) overlain by very fine-grained sediments.

Various sedimentary structures, evidenciated by placers, are present such as low angle cross laminations, planar laminations, hummocky cross bedding, etc..

This high frequency of placers in siltstones, in the upper part of the Punta Serpeddì Formation, can be interpreted as the result of storm activity in an offshore environment.

3. Conclusion

The main sedimentological characteristics of the Punta Serpeddì Formation indicate a high energy environment dominated by traction mechanisms. Decantation processes were also present and resulted in the emplacement of heavy mineral placers.

The sedimentation was controlled by high energy mechanisms such as storms (Loi et al., 1991a). Characteristics of deposits depend on the various environments in which the Punta Serpeddì Formation accumulated: coastal environment in the lower part gradually evolving to an offshore environment in the upper part.

Because of the argillaceous fraction scarcity, the Punta Serpeddì Formation storm structures (hummocky cross stratification?) are roughly preserved and incomplete. Besides, in the environment where hummocky cross stratifications have been produced, their good preservation was very difficult, because of sea-bottom reactivation by normal waves. On the contrary HCS were preserved in the transitional environment, under the normal waves action.

In the Punta Serpeddì Formation the frequent heavy mineral placers with clear structures of high energy traction processes (plane lamination and hummocky cross bedding) are the main result of storm actions. A detailed study is in progress and the preliminary results don't allow to go further in precision.

The comparison with storm deposits, such as the Grès Armoricain Formation (Arenig of western France), in which heavy mineral placers have been recorded, (Guigues & Devismes, 1969) confirms the idea of storm dominated environments for the deposition of Upper Ordovician Sarrabus deposits.

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THE PUNTA SERPEDDI FORMATION NEAR DOLIANOVA (SARRABUS - SE SARDINIA) RECENT PETROGRAPHICAL AND GEOCHEMICAL DATA

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Abstract

The Punta Serpeddì Formation (Upper Ordovician) outcropping in Sarrabus (SE Sardinia) overlies Ordovician lavas known as "Porfidi grigi".

The arenites of Punta Serpeddì Formation contain quartzarenite and quartzofeldspathic lithic fragments and heavy minerals concentrated in millimetric placers. The "Porfidi grigi" is the source of Ti-rich minerals and Y-poor zircon but Y-rich zircon, tourmaline and quartzofeldspathic fragments have been provided probably by a plutonic/ metamorphic Precambrian basement presently unknown in Sardinia.

Key words: *Palaeogeography, Placers, Caradoc-Ashgill, Sardinia.*

1. Introduction

The Punta Serpeddì Formation (Barca & Di Gregorio, 1980), located in the Sarrabus area (SE Sardinia), is a terrigenous clastic unit which has been assigned to the Caradoc-Ashgill (Barca & Di Gregorio, 1979; Giovannoni & Zanfrà, 1979). It transgressively overlies precaradocian volcanic rocks known as "Porfidi grigi" (Calvino, 1967), which include eruptive rock-types of intermediate to acid composition (Memmi et al., 1982). These volcanic rocks lie above the Cambro-Ordovician terrigenous unit known as "Arenarie di San Vito" (Calvino,1959; Barca & Di Gregorio, 1979; Barca et al., 1981; Barca et al.,1988).

In the Sa Murta section (fig. 1), 1Km to the East of Dolianova (Sarrabus - SE Sardinia), the Punta Serpeddi Formation is constituted by (from the bottom to the top):

— decimetric to metric beds of microconglomeratic and coarse grained sandstones (40m);

— centimetric to decimetric layers of rhythmically alternating siltstones and conglomerates (4m);

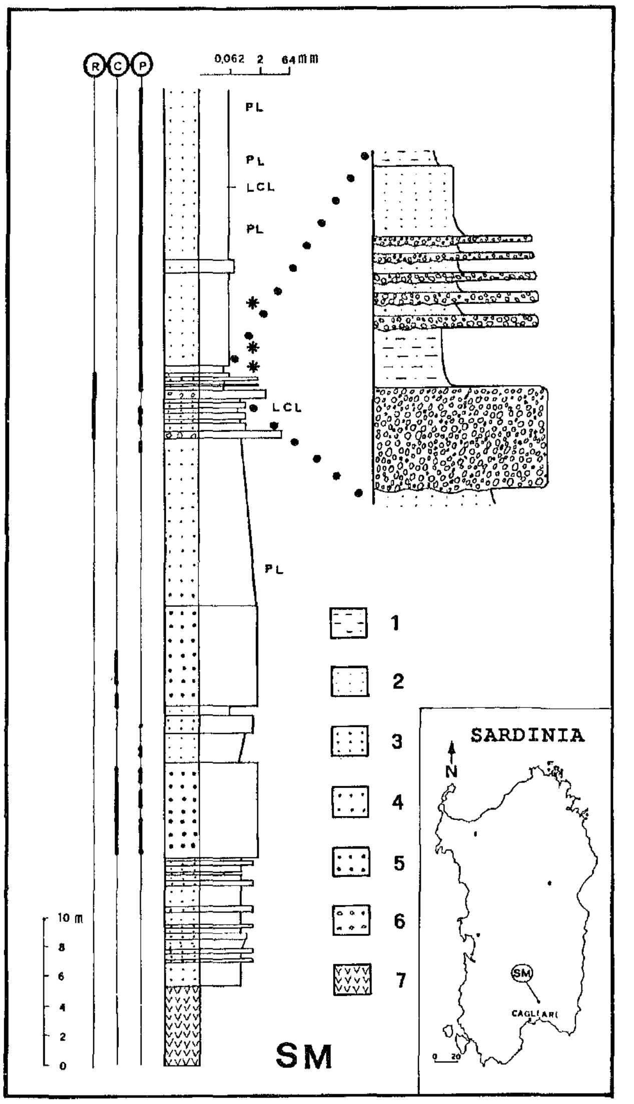

Fig.1: The Punta Serpeddì Formation in the Sa Murta section.
SM: Sa Murta section; R: rhythmites; C: cross bedding; P: placer; PL: plane lamination; LCL: low angle cross lamination; *: fossils.1-6; Punta Serpeddì Formation. *1: micaceous siltite; 2: siltstone; 3: sandstone; 4: coarse grained sandstone; 5: microconglomeratic sandstone; 6: conglomerate.* 7: "Porfidi grigi".

— fossiliferous siltstones (20m) (Caradoc-Ashgill).

Heavy minerals concentrations (placers) are the main feature of this formation which deposited in storm-dominated offshore and shoreface environments (Loi et al., 1991a).

2. Petrographical Analysis

The main constituents of the "Porfidi grigi" are globular quartz, sericitized feldspar, muscovite, biotite and accessory minerals (rutile, ilmenite, monazite and prismatic zircon). The fine grained matrix comprises mainly quartz and feldspar.

The mineralogical composition of siltstones and sandstones is characterized by the presence of various lithic fragments: "Porfidi grigi" matrix (1,8% to 30,6%), quartzarenite-siltstone (0 to 8%), quartzofeldspathic plutonic and/or metamorphic rocks (0 to 2,8%).

Monocrystalline quartz grains are rather abundant (33% to 73%) and some of them, which show etch-pits, are of volcanic origin.

The other constituents are muscovite, biotite, albite, K-feldspar and heavy minerals concentrated in placers.

The main minerals of placers are rutile, pseudorutile, leucoxene, ilmenite, zircon, monazite and tourmaline.

The heavy mineral frequency has been calculated in 50 thin sections, on the basis of point countings (1500 points/thin section).

Each point counting has been curried out perpendicular to the sedimentary lamination. Results show that heavy minerals are abundant to very abundant (2% to 12%).

The fine-grained matrix constitutes generally more than 15% of the sediments which can be related to the lithic wackes group (Loi et al., 1991b) (fig. 2).

3. Microprobe Analysis

Zircon grains have been analysed by microprobe beam (Microsonde Ouest - Brest - anal. M. Bohn), in some samples of "Porfidi Grigi" and of arenite of the Punta Serpeddì Formation .

The "Porfidi grigi" zircons are characterized by their low Y_2O_3 concentrations (than <0,5%). This can be explained by the fact that Y has a compatible behaviour with respect to monazite.

In samples collected in the upper part of the Punta Serpeddì Formation, zircon grains can be separated in two populations (Loi et al., 1991c) on the basis of morphology, colour and Y_2O_3 amount (Fig. 3).

In prismatic and colourless grains, the Y_2O_3 amount is low and close to the Y_2O_3 amount of "Porfidi grigi" zircon elements. On the contrary, well rounded zircon grains are coloured and their Y_2O_3 amount is about twice the amount in colourless grains.

Their rounded morphology, the colour and Yttrium concentration suggest a plutonic or metamorphic origin (Blatt et al.,1980).

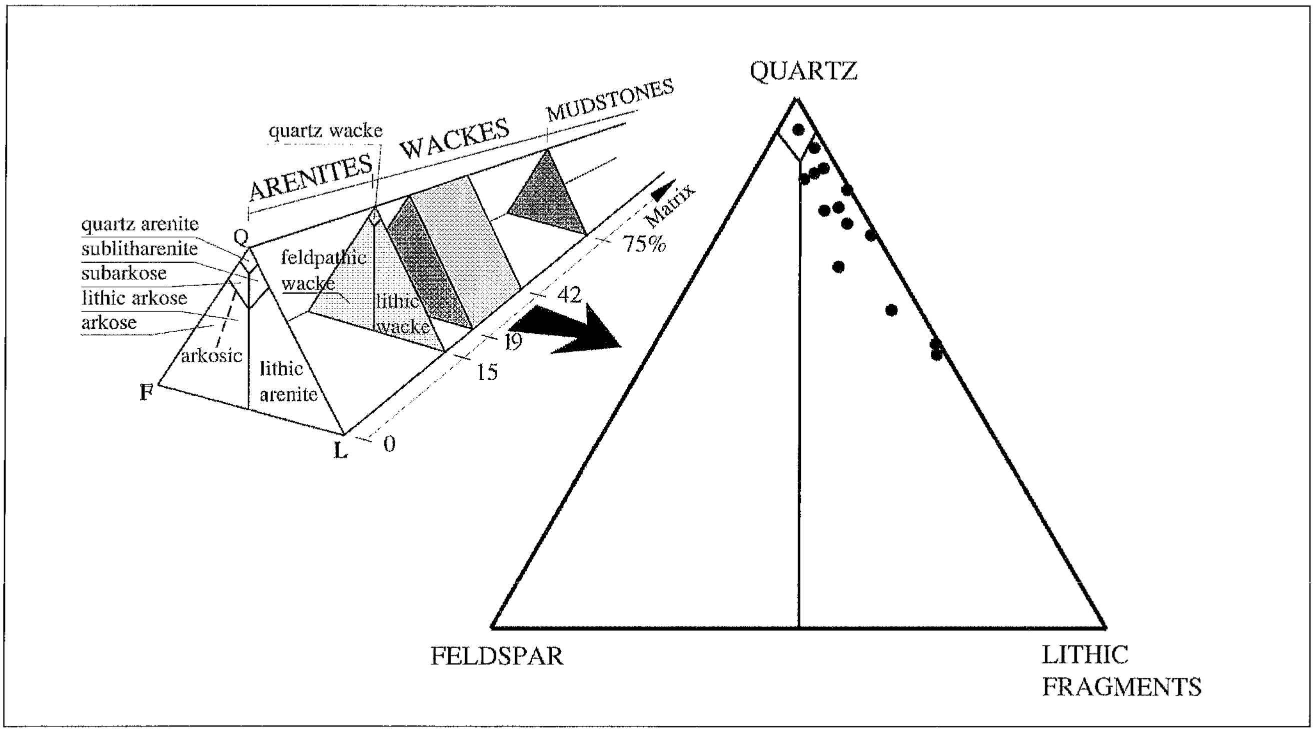

Fig. 2: Composition of sediments of the Punta Serpeddì Formation (quartz-feldspar-lithic fragments diagram after Pettijon et al., 1972)

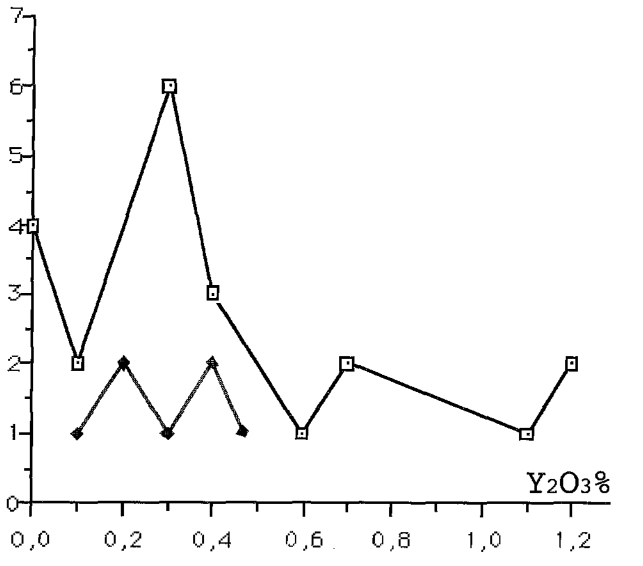

Fig. 3: Distribution of the Y_2O_3 contents of zircons grains. "Porfidi grigi" (◈) and Punta Serpeddì Formation (▣). Vertically: number of analysed zircons; horizontally: Y_2O_3%.

4. Sedimentary Evolution

The composition of sediments evolves from the bottom to the top of the Punta Serpeddì Formation (fig. 4).

In the coarse-grained lower part of the section, the concentration of polycrystalline quartz and quartzofeldspathic fragments progressively increases. It decreases in the upper fine-grained part of the formation.

The evolution in the siltitic upper part is linked to the granular structure of the polycrystalline grains which are easily disaggregated and so are not perceptible in the fine-grained upper part.

On the other hand, the amount of volcanic fragments decreases from the bottom to the top of the section.

This decrease is not linked to the granulometric evolution because these fragments are constituted by microcrystalline volcanic groundmass and are well preserved, even in fine-grained sediments.

So, the observed evolution for the volcanic fragments, can be related to a progressive removal of the source area linked to the transgressive phenomena.

In the coarse-grained lower part of the Sa Murta section heavy minerals are concentrated in centimetrical rich beds interlayering decimetrical to metrical poor beds.

The frequency of heavy mineral placers increases upwards and reaches its maximum in siltstones at the upper part of the section, where they are concentrated in millimetric layers very close one another.

So, the initially sandy sedimentation of the Punta Serpeddì Formation appears to have been mainly fed by the underlying "Porfidi grigi".

Then, sedimentary and plutonic and/or metamorphic source areas progressively participated to the clastic supply.

5. Conclusion

The present petrographic study of the Punta Serpeddì section shows that the clastic material was provided by three source areas.

The main source was the underlying "Porfidi grigi" which provided volcanic fragments, Ti-rich minerals, monazite and Y-poor prismatic zircon.

Presence of Y-rich rounded zircon, tourmaline and quartzofeldspathic fragments suggests the contribution of a plutonic and/or metamorphic source area.

The origin of quartzofeldspathic clasts remains uncertain. The oldest plutonic rocks known in Sardinia are the orthogneiss of Capo Spartivento outcropping in the South Sulcis and recorded by Delaperrière & Lancelot (1989) as Ordovician in age (478 +/- 16 MA - U/Pb on zircon). An older source cannot be excluded but, plutonic and metamorphic rocks of sure Precambrian age have never been recorded in Sardinia.

As for siltstone and quartzarenite clasts, their origin

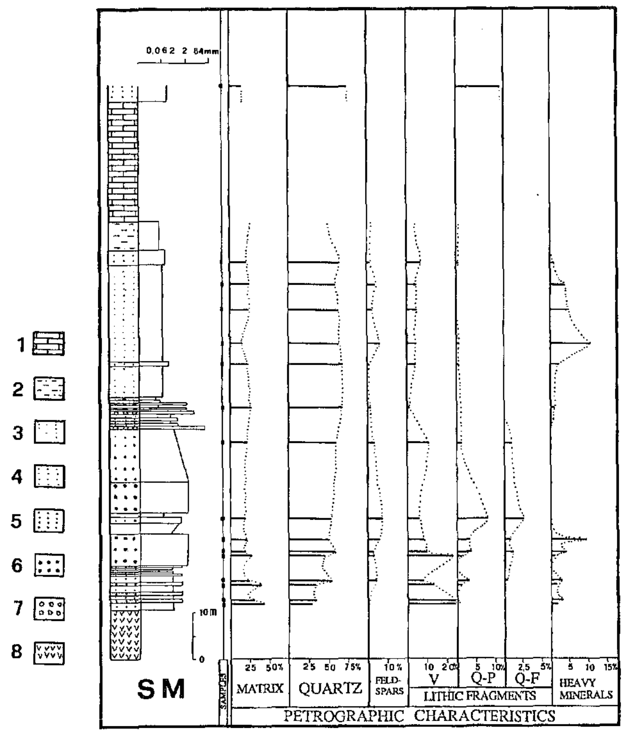

Fig. 4: Petrographic characteristics and sedimentary evolution of the Punta Serpeddì Formation in the Sa Murta section. SM: Sa Murta section; V: volcanic fragment; Q-P: polycrystalline quartz; Q-F: quartzofeldspathic fragments. 1: Tuviois Formation (siliceous limestone); 2-7; Punta Serpeddì Formation; *2: micaceous siltite; 3: siltstone; 4: sandstone; 5: coarse grained sandstone; 6: microconglomeratic sandstone; 7: conglomerate.* 8: "Porfidi grigi".

can be found in the largely outcropping Cambro-Ordovician "Arenarie di San Vito".

The Sa Murta section sediments are clearly constituted by a mixing of various clastic materials provided by several source areas. Fragments of volcanic "Porfidi grigi", of the underlying terrigenous "Arenarie di San Vito" and of a plutonic/metamorphic old shield are present in the early deposits of the Sarrabus Ordovician sedimentary basin. Besides, tourmaline and Y-rich rounded zircon are present from the bottom to the top of the formation so proving the permanent alimentation from an old shield source.

The distribution of sediments of the upper part of the Sa Murta section in the quartz-feldspar-lithic fragments diagram (fig. 2) is similar to the one observed by Potter et al.(1986) for the Brasil nearshore sands fed by the reworking of the South American shield.

On the basis of these data, the Sarrabus Ordovician Basin can be interpreted as a cratonic basin in which the clastic material is provided at once by an old Precambrian shield by the Cambro-Ordovician "Arenarie di San Vito" and by the Ordovician "Porfidi grigi".

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GEOLOGICAL SETTING OF THE NORTH-EASTERN IGLESIENTE AREA: AN EXAMPLE FOR DISCUSSING THE STRUCTURAL MODEL OF THE SARDINIAN HERCYNIAN CHAIN

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Abstract

The current interpretation of the structure and stratigraphy of the Hercynian Chain of Sardinia is strikingly contradictory in the northern part of the "Foreland" (NE Iglesiente area).

The Hercynian cleavage structures of the northern Iglesiente autochthonous area (beneath the allochthonous Arburese nappe) are overturned towards the North-East with a vergence opposite to the orogenic polarity of the chain as it was suggested by previous authors.

Moreover, the presence of metamorphic successions, belonging to more internal palaeogeographical domains, under the autochthonous, shows that the Iglesiente was emplaced on these more internal metamorphites before the overlapping of the Arburese Unit.

Lithostratigraphic considerations and fossil dating indicate that the tectonic evolution is more complex than that maintained up to now.

Keywords: *Structural Geology, Hercynian Orogeny, Palaeozoic, SW Sardinia.*

1. Introduction

The Iglesiente, like the whole south-west area of Sardinia, is characterized by outcrops of metamorphic successions (mainly believed to be of Early Cambrian to Early-Middle Devonian age), deformed during the Caledonian and Hercinian Orogenies.

These tectonic events, created very complex interfering structures, which can be observed on a detailed map level.

On the basis of Carmignani et alii's recent model (1980; 1982; 1986; etc.), the Iglesiente represents the most external area (Foreland Auct.) of the Hercinian Chain.

This hypotesis is mainly based on the following considerations:

— The low metamorphic grade of the outcropping successions, typical of the most external area of the "collision chain".

— The peculiar structural shaping , mainly characterized by subvertical axial plane folds, in contrast with the structure of the innermost tectonic areas, characterized by sub-horizontal axial plane sinmetamorphic folds.

— The geometrical relationships between the Iglesiente and the Most Internal Tectonic Units which overlap towards south-west above the authocthonous.

In the last years, however, on the basis of a number of field observations some geological elements, not easily explainable with the proposed model, have been found.

These elements, together with similiar situations observed in different parts of the north-east area of "Campidano", suggested for the Sardinian Hercynian chain the hypothesis of a much more complex deformation mechanism than the ones previously proposed (Maxia 1985-1987).

— North-east verging folds involving the Ordovician-Carboniferous succession, placed outside the Cambrian domain (in which folds verging in the same direction had partly been described by other authors) were found in the NE Iglesiente area together with new elements indicating the presence of upside down flans of kilometric isoclinal folds.

The study of this area, concerning above all the geometrical aspects of the most northern sequences and their relationships with the tectonically superimposed units, permied to point out as follows (fig. 1, 4,):

— A new structural situation for the whole NE Iglesiente area.

— Substantial differences concerning the geometrical and structural relationship with the units belonging to the more internal paleogeographic domains of the chain.

A more complex scheme of deformation which is different from the one previously observed.

2. Stratigraphy

The most ancient succession ("Formazione di Nebida") is characterized by terrigenous sediments mainly represented

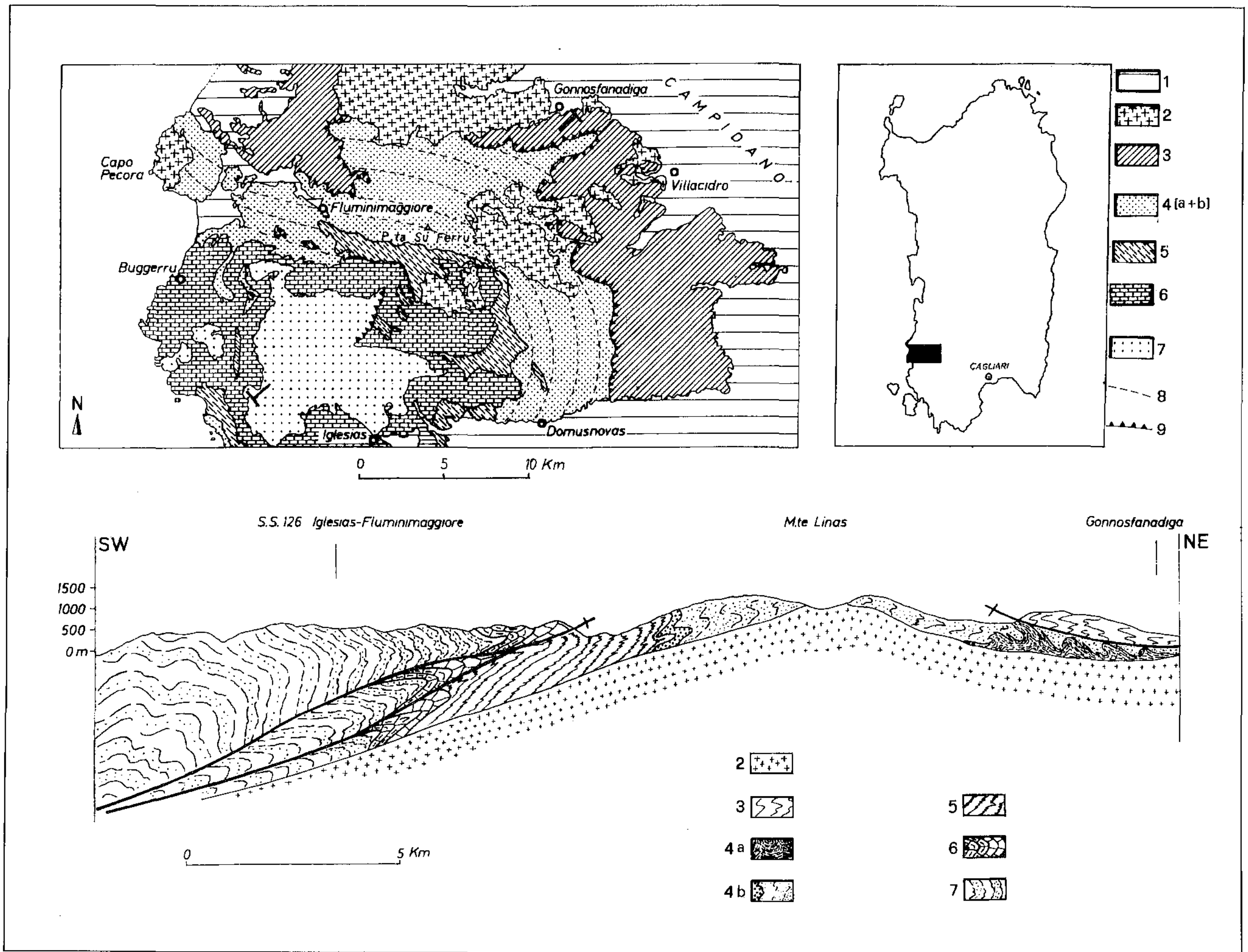


Fig. 1 - Schematic geological map of Iglesiente - Arburese area and interpretative cross - section. 1: Post Paleozoic sequences; 2: Late Hercynian granites. *Arburese Unit :* 3: Cambrian-Lower Ordovician sequence. *Autochthonous Succession :* 4a: Silurian-Devonian and Lower Carboniferous sequence; 4b : Middle-Upper Ordovician sequence; 5 : Cabitza Formation (Midle-Upper Cambrian-Tremadoc); 6: Gonnesa Formation (Lower-Midle Cambrian); 7 : Nebida Formation (Lower Cambrian); 8 : Axial plane traces of major deformation phase; 9 : Main thrust.

by meta-sandstones which, in the upper part, contain carbonatic lenses with archaeocyatinae and trilobites of early cambrian age (Rasetti F. 1972; Cocozza T. 1980) - (Fig. 2)

This sequence, whose bottom is unknow, is about 1000 m. thick; it was observed mainly in the nucleus of the dome structure which forms the central western part of Iglesiente.

It is followed by the calcareous-dolomitic carbonatic succession of the "Gonnesa Formation" (Metallifero Auct.) subdivided into the following members: "Dolomia rigata", "Dolomia grigia" and "Calcare Ceroide". These litotypes, 600-700 m. thick,form the principal mineral horizon in the whole "Anello Metallifero" of Iglesiente;

Stratigraphically above the carbonatic succession, some sequences constitued of prevailingly polychrome silty-clayed metasediments know as "Formazione di Cabitza" are found. They are divided into the two members of "Calcari Nodulari" or "Calcescisti" Auct. and "Scisti di Cabitza".

This last member, that was previously totally included between the higher part of Lower Cambrian and the Middle Cambrian, has recently been dated up to the Tremadocian (Barca S. et alii, 1986); its thickness reaches at least 500 m..

The above mentioned Cambrian-Lower Ordovician sequence is covered, unconformably, by a prevalently terrigenous succession (at least 800 m. thick encompassing the middle-upper Ordovician to the middle-lower Devonian and perhaps part of the lower Carboniferous).

The succession includes, from the bottom, polygenic metaconglomerates (Puddinga Auct.), varicolored terrigenous metasediments, azoic meta-sandstones passing upwards to siliciclastic to carbonate fossiliferous metasediments with a few volcanic interlayers (Cocozza T. & Leone F. 1977) of Caradocian to Ashgillian age (Leone F. et alii, 1991).

Black Graptolitic schists and Orthoceras limestone

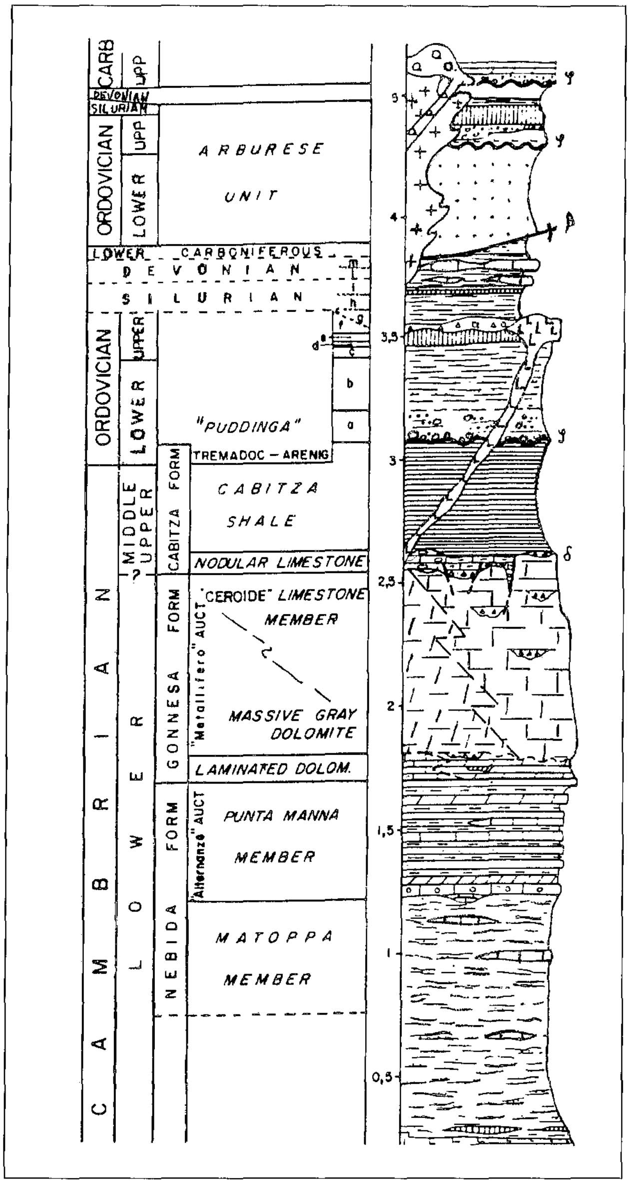


Fig. 2 - Palaeozoic sequence of Iglesiente area (Carmignani et alii, 1986, modified).

lenses (Silurian) which are followed by Devonian schists and limestones with Crinoids and Tentaculites. The sequence is closed by azoic terrigenous-conglomeratics metasedimentary outcrops, aged Lower Carboniferous.

3. Tectonics

The several hypothesis about the stratigraphic polarity and the geometric setting were underlined by the old writers (Novarese V. 1942; Geze. B. 1952; Vardabasso S. 1940-1956; Shwartzbach M. 1939; etc.).

Modern structural analyses studies (Arthaud F. 1963-1970 ; Dunnet D. 1969; Poll J.J.K. & Zwart H.J. 1964; Carmignani L. et alii 1982-1986) have lately shown a more precisely tectonic framework which led to the consideration that the Iglesiente area (inclusive of the SW part of the Island) is the more external area of the Sardinian Hercynian segment. The studies undertaken showed that this area has a typical high-crustal level hercynian structuration and a low grade deformation and metamorphism (Arthaud, 1970).

The structural style of the Iglesiente is the result of four main deformation phases:

A) Open Caledonian folding with E-W trending axes sealed by the middle-late Ordovician transgression (Sardinian Phase AUCT.)

B) Weak non-metamorphic deformations still with E-W axes, involving the previous structure and the superposed Ordovician-Carboniferous succession (first Hercynian phase)

C) Synmetamorphic deformations associated wich sub-vertical or very inclined schistosity or N-S axed folds interfering with the previous folds which caused the "Domes and Basins" structures (Principal Hercynian phase or 2nd Hercynian phase

D) Minor folding with different directions, perhaps associated to late faulting and generally expressed by conjugate kinks.

Among the problems caming out from the analysis of the above menditioned scheme, we can point out the followings:

— The meaning of the non-metamorphic deformations linked to the first Hercynian phase (Carmignani e Pertusati, lg96).

— The presence of sectors (north-eastern Iglesiente, and Southern Sulcis) in which the principal Hercynian phase (2nd Hercynian phase) was associated with a tangential style tectonics, characterized by anincreased degree of deformation and metamorphism (Arthaud, 1970; Carmignani et alii, 1982; 1986; ecc.)

— The correlation between the synmetamorphic principal phase of the Iglesiente area and the more internal areas of the chain.

With regard to the first problem the authors think that the first Iglesiente Hercynian unmetamorphic phase, which is characterized by E-W axed folding, must be considered much more important than previously thought. In fact, a deformation phase preceeding the main synmetamorphic one, (expressed according to the locality and the metamorphic zoneography with not schistogenous folds, overthrusts and schistogenous structures more or less linked to the deep horizontal shear- zone), would better explain, by the ipothesis of a extensive tectonic wich followed the main phase (Carmignani L. 1982, 1991, etc.), some geological problems pointed out in other localities of the Foreland, such as the two schistogenetic phases described in the deepest sequences of the Sulcis-Iglesiente area.

With regard to the other two problems, they are part of the next paragraph.

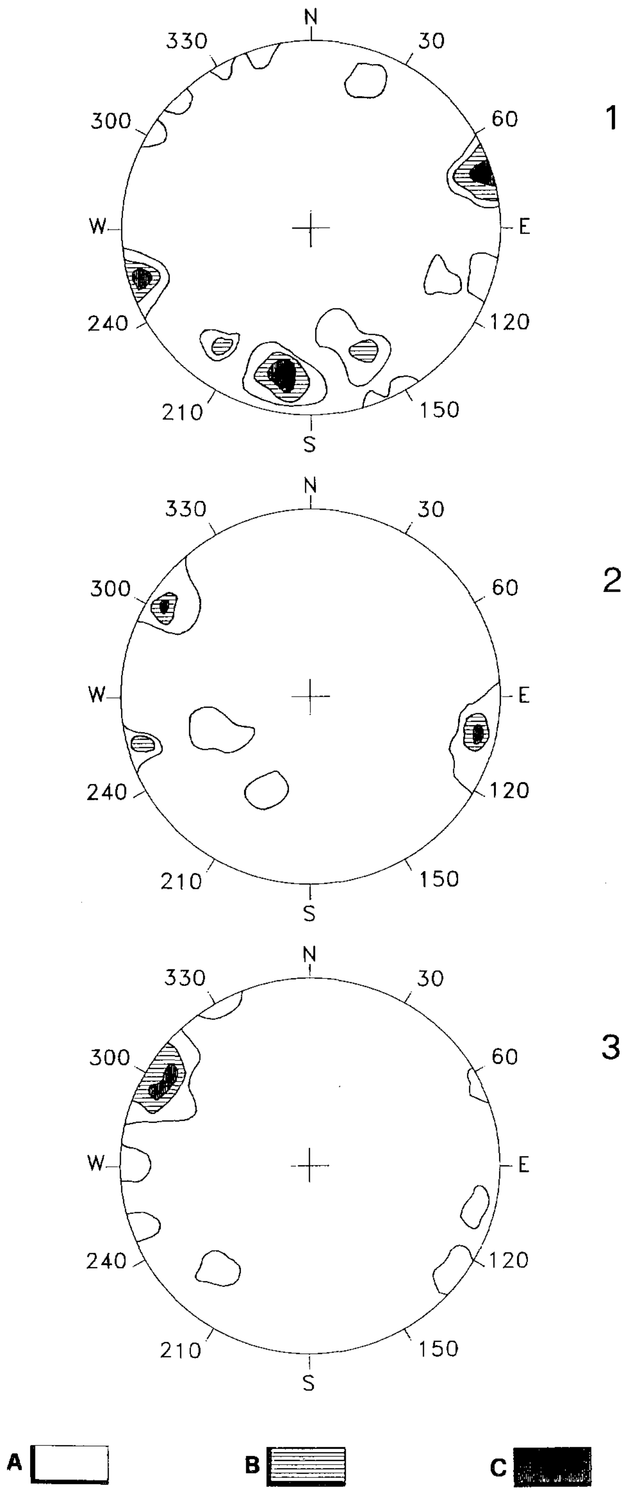


Fig. 3 - Stereograms (Schmidt projection, lower hemisfere) showing the axes of the main deformation phase. *Station: 1:* Domusnovas (110 m.); 2: P.ta Su Ferru (75 m.); 3: Fluminimaggiore (92 m.). *Class: A (1%* 3%); B (3% - 5%); C (> 5%).

4. New Structural Data

Recent surveys of the Ordovician to Carboniferous sequence in the north-eastern part of the Iglesiente, have pointed out as follows:

a) The synmetamorphic plicative structures of the main deformation phase, occurring in the Ordovician sequences, are constantly overturned towards the north-western quadrants. They describe a general kilometric turning carryng schistosity traces, axes and lineations of intersection, from approximately N-S (Domusnovas sector) to approximately E-Wdirections (Fluminimaggiore sector), (fig. 1).Statistical Schmidt stereograms (fig. 3) performed in three different sectors of the examined area show axial dispersion due to the interference between the main Hercynian phase and the preceeding structures of the unmetamorphic Hercynian phase (See also Arthaud, 1970).

b) Moving away from the oldest nucleus, the same schistosity surfaces become less and less inclined (M. Linas southern side etc). Then, they dip northward because of a late folding deforming the main structures and the schistosity itself (North-eastern Fluminese; M. Linas northern side; etc.).

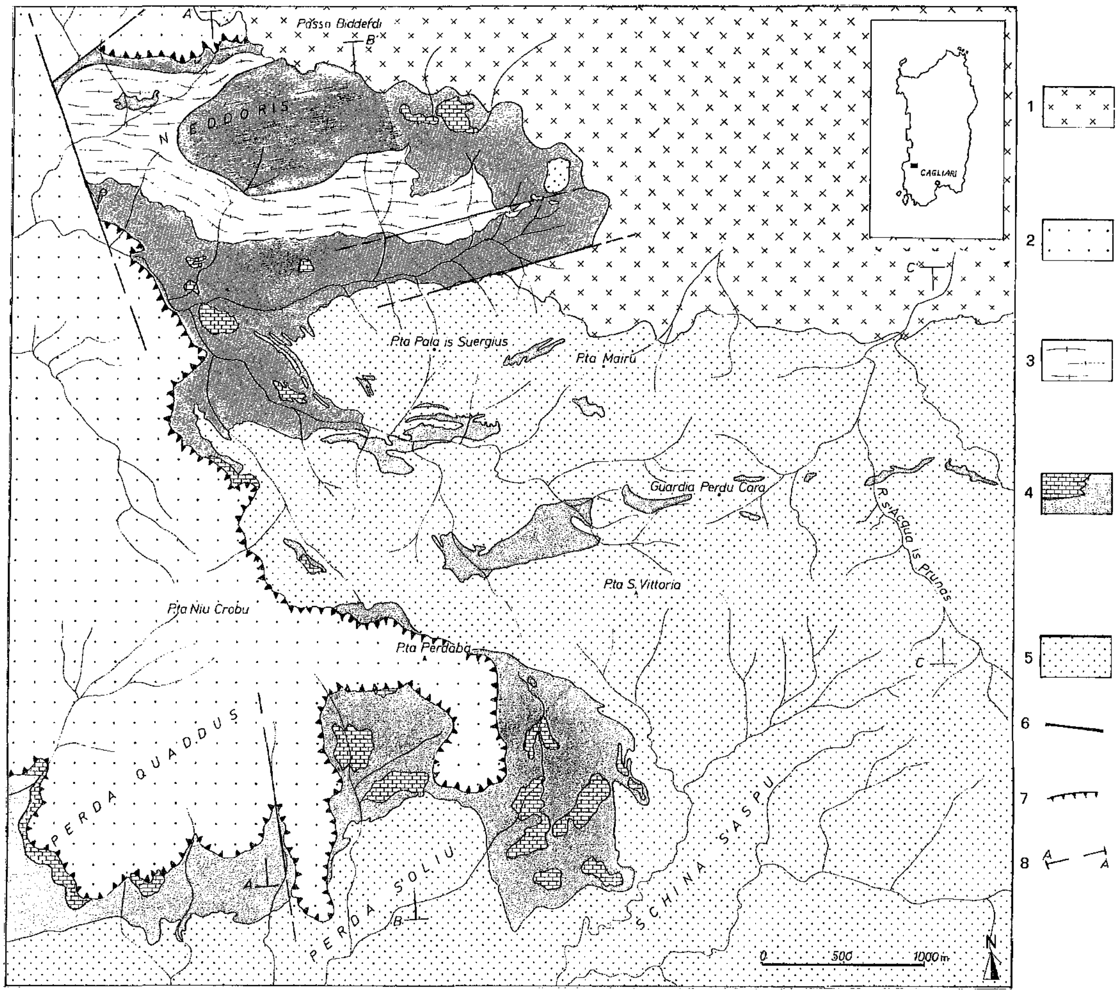
c) Following the same direction, an increase in the deformation and the metamorphic degree can be noticed. In agreement with Arthaud (1970), a fracture schistosity can be observed in correspondence of the structurally (and geometrically) higher Cambrian-Ordovician successions (the Marganai area, S. Benedetto, etc).It evolves into a flow schistosity in the geometrically underlying (and therefore structurally deeper) successions, which can be presently observed, because of erosion, around the Oridda and Linas intrusive domes.

Along the same direction the mechanisms of deformation change too, passing from flexure deformations and flexure and/or shear deformations which can be seen in the Cambrian - lower Ordovician successions of the central sectors (variable folding for intensity and style according with the lithology considered) to deformations marked by thikened hinges and thinned limbs, isoclinal folding and to a widespread flatting of the lithological types observed in the area.

In disagreement with Arthaud (1970), who attributes the increase of such deformation (and metamorphism) to the thermic effects due to the closeness of the above mentioned intrusive domes, we believe that the increase in deformation is due to the "lithodynamic" load generated by the growing structure , lately involved in contact metamorphism by the intrusive bodies.

d) In the northern zone of the examined area, apart

Fig. 4 - Schematic geological map of North-Eastern Fluminese area and interpretative cross sections; 1: Late Hercynian granites (Upper Carboniferous). *Arburese Unit* : 2: Metasandstones alternating with phyllites (Cambrian - Lower Ordovician). *Lithostratigraphic Unit belonging to internal domains:* 3: Metavolcanites with metaconglomeratic and metarenaceous levels (Midle-Upper Ordovician?7. Autochthonous Successions: 4:Metalimestones, carbonaceous phyllites, "Liditi" (Silurian-Devonian) and restricted terrigenous-conglomeratics metasedimentary outcrops (Lower Carboniferous); 5: Metaquarzites, metasandstones, metasiltites with fossil, silicizzed metalimestones, ecc. (Upper Ordovician); 6: Main thrusts; 7: Main faults; 8: Traces of geological cross sections.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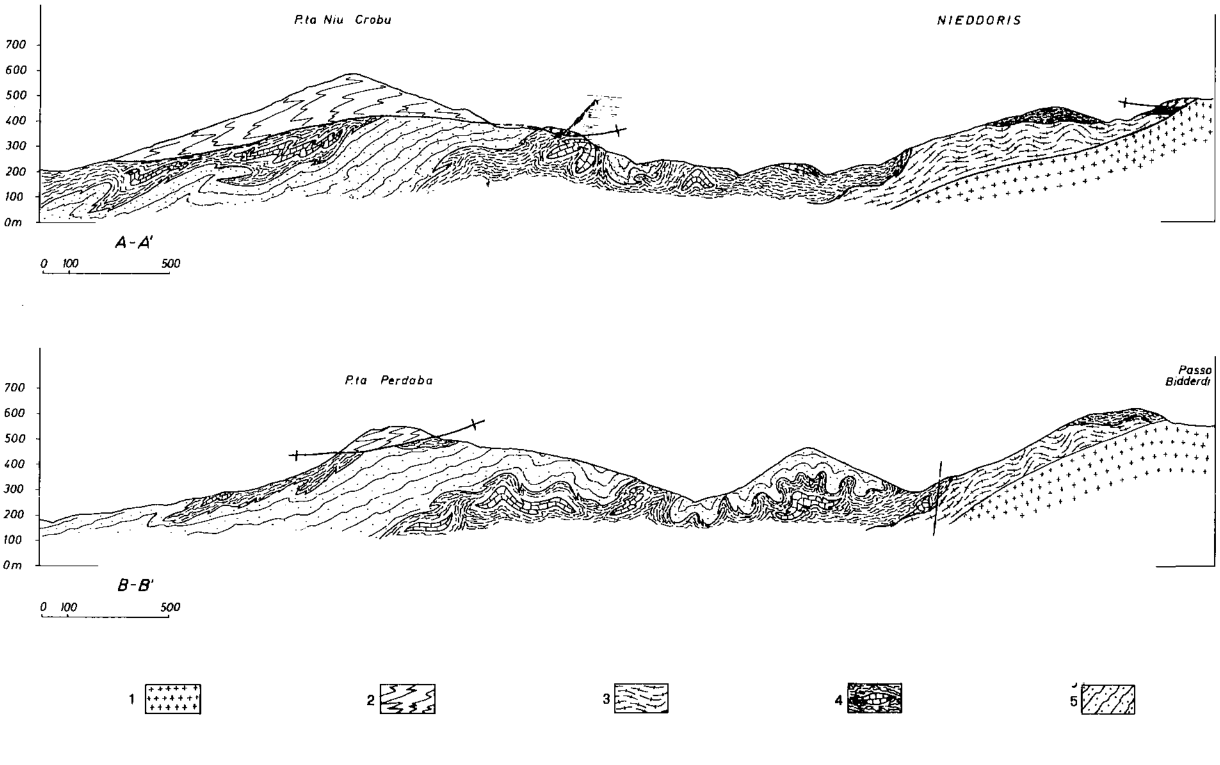
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from the metamorphic successions belonging to the Autochthonous sedimentary sequences, volcanic and sedimentary successions considered as being part of the most internal paleogeographical domains can be found (Fig - 4).

The Autochthonous succession (Iglesiente area) is characterized by the typical sequences, mainly arenaceous and sometimes fossiliferous , of the middle- upper Ordovician age. Upwards, they pass to pelitic-carbonatic sediments belonging to the Silurian- Devonian (Cocozza-Leone, 1977) and small terrigenous outcrops of lower Carboniferous age.

The succession belonging to the most internal areas, includes:

— The pelitic-arenaceous monothonous sequences of Cambrian-early Ordovician age belonging to the Arburese Unit (Barca S. et alii 1981), which overlay the first successions tectonally;

— Volcanic and subordinately terrigenous layers, similar to those found in the middle-upper Ordovician successions of central-eastern Sardinia (Calvino F.1961, 1967; Memmi et alii 1982 etc;).However, they are structurally refolded together with the autochthonous succession.

In the studied area a series of new geological elements were pointed out from a tectonic point of wiew.

Among the collected data it is possible to notice as follows:

l) The structural geometry of the Autochthonous: on the basis of lithostratigraphic reconstructions and paleontological dating, it can be interpreted as the overturned limbs of N to NE verging kilometric folds.

2) A late phase folding these first structures especially in the northern area, generating antiforms and synforms with approximate E-W axes.

3) The geometrical position of the above mentioned (mainly volcanic and volcanoclastic) Ordovician metamorphites maintained to belong to more internal domains.They lay beneath the Autochthonous lithostratigraphic succession of the Iglesiente area. They are also refolded with them in a later phase.

4) The Arburese unit overthrusts the Iglesiente structural complex allready deformed by the plicative structures of the two described phases.

5. Conclusions

The data since now examinated put forward some new obsevations about the Foreland structure and its evolutional history.

The north-eastern sector of the Iglesiente area, which has been described as “the relative Autochthonous most external area of the "Sardinian Hercynian Chain", on the basis of its metamorphic and structural features and its geometrical relations with the units of the internal areas, appears as a large scale NE verging overthrust. In the most northern part it is characterized by a kilometric fold overturned in the opposite direction of the chain vergence.

Thrusting occured during the main deformation phase (2nd Hercynian phase of the Iglesiente area) over the successions belonging to the paleogeographical domains of those areas believed to be the more internal ones. Subsequently, the Arburese Unit was transported (with a S-SW vergence) up on the described structural complex, simultaneously (or subsequently) with a late phase that refolded the structure (and schistosity) of the main phase.

The suggested structure cannot be easily explained by the current schemes, both those implying a compressive regime (Carmignani L. et alii 1978, 1982,1986; etc.) and those recently assuming an extension following the main deformation phase (Cappelli B. 1989; Carmignani L. et alii 1991).

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THE GNEISSES OF MONTE FILAU (CAPO SPARTIVENTO, SW SARDINIA): PETROGRAPHIC AND CHEMICAL FEATURES

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Abstract

The Monte Filau orthogneisses show peculiar metamorphic features. They are probably "Caledonian" granites of anatectic-crustal origin which, before being intruded by the late Variscan granites, underwent two different metamorphic events. The older event caused their alteration into gneisses, with crystallization of andalusite, sillimanite and garnet, and the appearance of pockets of melts, in sites of suitable composition (T > 650°C and P 3.5-4.8 Kb). The younger event had mainly cataclastic-mylonitic effects, with recrystallization in the Qtz-Ab-Ep-Chl subfacies. Local contact effects by the late Variscan granites concluded the metamorphic history of the M. Filau orthogneisses.

Key words: *"Caledonian" magmatism, Variscan metamorphism, SW Sardinia.*

1. Introduction

The oldest rocks in the southern Sardinian basement, i.e. the Bithia Formation, are infra-Cambrian and, in the sector of southern Sulcis, they tectonically overlie a mainly gneissic complex composed of the Monte Filau gneisses and Settiballas schists (Sassi et al., 1990). Field observations and petrographic data suggest that their protoliths were granitic in composition and plutonic in nature (Cocozza et al., 1977).

This work reports the chemical and petrographic characters of the M. Filau gneisses, in order both to characterize the protoliths, and to evaluate their metamorphic evolution in comparison with that recorded in the surrounding rocks, which are polymetamorphic and belong to the amphibolitic facies (Sassi et al., 1990).

2. Geological Setting

The M. Filau gneisses outcrop over an area of about 20 km^2 near Capo Spartivento (Fig. 1) . Together with the Settiballas Schists, they make up a structural element surrounded by the schists of the Bithia Formation, a low-grade metamorphic sedimentary sequence. Post-kinematic, Hercynian granites crosscut all metamorphic sequences.

On the basis of isotope data and field evidence, the age of these three rock formations has been established as follows:

i) The Bithia Formation is believed to be infra-Cambrian, on the basis of regional stratigraphic evidence (Cocozza et al., 1977; Junker & Schneider, 1979);

ii) The M. Filau gneisses represent an Upper Ordovician granitoid intrusion (Cocozza et al., 1977; Scharbert, 1978; Ferrara et al., 1978; Ludwig & Turi, 1989; Delapierre & Lancelot, 1989; Sassi et al., 1990);

iii) The post-kinematic granites belong to the late Variscan magmatism, on the basis of geological evidence and radiometric datings (Cocozza et al., 1977; Scharbert, 1978).

The Settiballas schists may belong to the M. Filau structural element. Their metamorphic grade is higher than that of the Bithia rocks: amphibolitic facies vs lower greenschist facies (Sassi et al., 1990).

3. Petrography of M. Filau Gneisses

The M. Filau gneissic body consists of several rock types: a) medium- to coarse-grained granitic gneisses, b) leucogranitic fine-grained gneisses and c) cataclastic gneisses. The former gneisses under point a) are the main rock type.

The cataclastic types occur in the external parts of the body, i.e. the topographically higher parts (e.g., see also Cocozza et al., 1977) and near the western and eastern contacts with the Settiballas Formation. In the latter case, decimetric xenoliths of biotite-rich schists occur along the boundary within the cataclastic gneisses.

The leucocratic fine-grained gneisses generally make up centimetric to metric dykes and apophyses within the coarser-grained gneisses.

All field observations indicate that the M. Filau gneissic body represents an original granitic intrusion, and the Settiballas schists their original country rocks (Sassi et al., 1990). In fact, gneissic apophyses and veins locally crosscut the Settiballas schists.

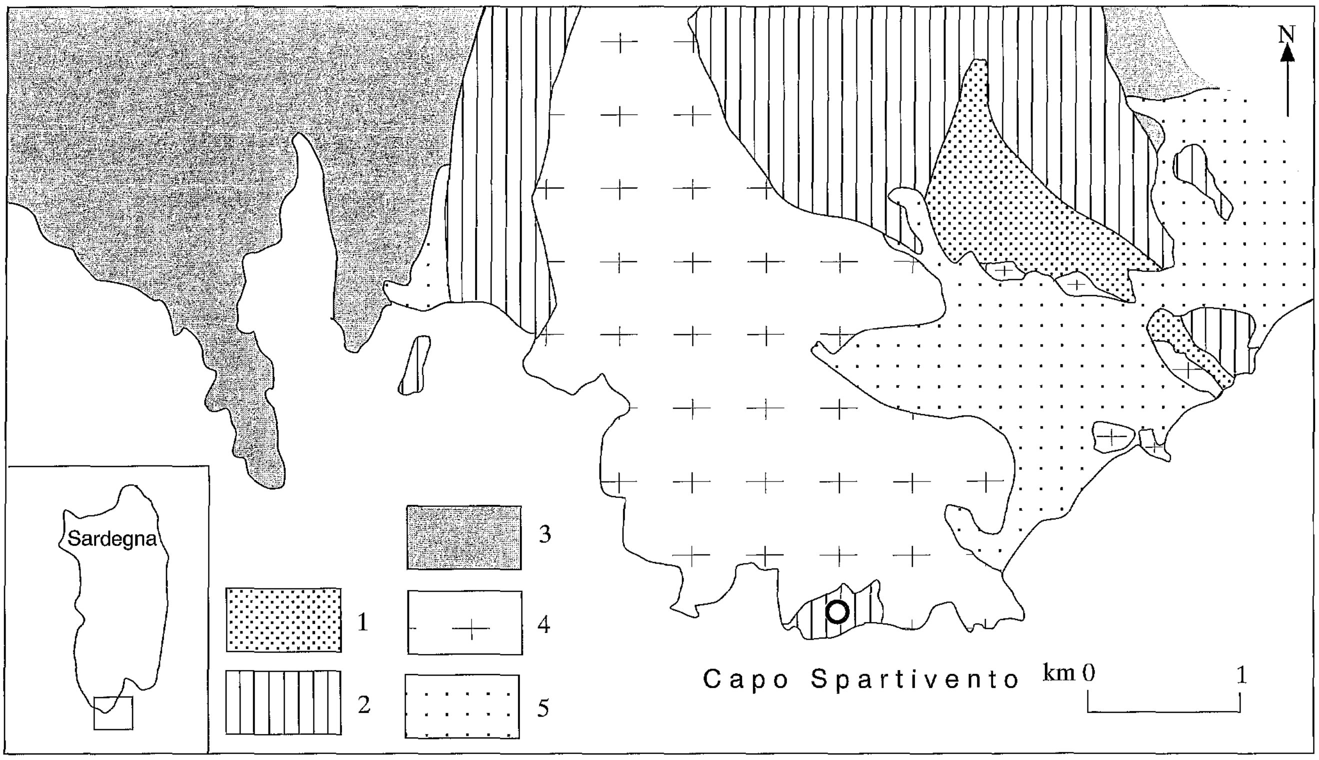

Fig. 1 - Geological sketch of Capo Spartivento area. 1) Settiballas Formation; 2) M. Filau orthogneisses; 3) Bithia Formation; 4) late Variscan granites (modified after Delapierre et al., 1989).

a) *The medium-to coarse grained gneisses.*

These gneisses often show an augen structure, and consist of feldspars + quartz + biotite ± muscovite, sometimes accompanied by small quantities of andalusite, sillimanite and garnet.

The augen are composed of pecilitic microcline-perthites. They are frequently associated with large lobed grains of quartz, aggregates of microcline + quartz + oligoclase, or large, irregular quartz crystals.

The groundmass consists of phyllosilicates, feldspar and quartz, and makes up discontinuous layers wrapped around the augen. Plagioclase commonly makes up small grains with a large saussuritized core and a thin clear rim of albitic composition. Albite also occurs locally as small individual crystals. Biotite is usually the most abundant (and sometimes the only) phyllosilicate. It defines the foliation, and frequently also occurs as decussate ovoids or small lenses. It shows strong post-crystalline deformations. Muscovite is either associated with biotite or replaces microcline phenoblasts and Al_2SiO_5 crystals as large flakes. It is abundant in the rock types containing andalusite or andalusite + sillimanite + garnet. Andalusite mainly occurs as crystals deeply altered into white mica, although some large, deep-pink coloured andalusite crystals also occur. Sillimanite makes up acicular crystals parallel to the schistosity, located within muscovite flakes. Garnet is very rare and makes up small, fractured but unaltered equant crystals.

These rocks also display contact metamorphic effects caused by the late Variscan granites: unoriented flakes of biotite and aggregates of white mica and biotite with decussate structure.

b) *The leucocratic fine-grained gneisses*

These rocks are characterized by white colour, a very poorly foliated structure locally grading to a granofels structure, and the occurrence of isoriented, dark, centimetric spindles consisting of a coarse-grained polymineralic aggregate (see below). Locally, these rocks display a layering due to the alternation of quartz- and feldspar-rich layers. The lineation defined by the elongation of the spindles lies within the foliation and layering surfaces. Quartz and feldspars make up the abundant granoblastic groundmass. The polymineralic spindles stand out from this granoblastic groundmass, due to their darkness and grain-size.

The granoblasts are quite regular with a low number of mainly rectilinear boundaries. Potassic feldspar (microcline) is an exception, since it tends to be pecilitic and larger in size, and is always parallel to the schistosity. Albitic plagioclase is weakly kaolinized. Garnet is frequent, and makes up fractured idioblasts sometimes containing small spots of quartz, often in aggregates parallel to the schistosity. The composition of garnet is peculiar (Tab. 1) due to the absence of MgO and scarcity of CaO; it is close to that of natural and synthetic garnets in granitic systems (Fig. 2; Green, 1977). Sillimanite occurs in fibrolitic aggregates or

	AMS1	AMS2		BP12	CPL2	IPL2		IGR4	BGR4	BGR5	HGR1
SiO_2	47.62	47.45		65.29	65.34	64.46		37.13	36.60	37.88	36.00
TiO_2	0.36	0.36		-	-	-		-	-	-	-
Al_2O_3	34.49	34.70		21.96	20.89	22.19		21.18	21.28	20.78	20.73
Fe_2O_3	0.24	0.08		-	-	-		-	-	-	-
FeO	2.64	2.71		0.32	0.45	0.34		36.53	31.52	31.10	20.73
MnO	-	-		-	-	-		4.62	9.91	8.98	4.67
MgO	-	-		-	-	-		-	-	-	-
CaO	-	-		1.78	1.52	2.60		0.53	0.68	0.29	0.52
Na_2O	1.21	0.66		10.27	10.90	9.97		-	-	-	-
K_2O	9.55	9.73		0.03	0.35	0.33		-	-	-	-
H_2O	4.51	4.51		-	-	-		-	-	-	-
TOT	100.61	100.20		99.65	99.45	99.89		99.99	99.99	99.96	100.01
#Si ^{IV}	6.30	6.29	Si	11.51	11.59	11.39	Si	6.07	6.00	6.18	5.96
#Al^{IV}	1.70	1.71	Al	4.56	4.37	4.62	Al	4.08	4.11	3.99	4.04
T site	8.00	8.00	Fe	0.05	0.07	0.05	Fe	5.00	4.32	4.24	5.27
#Al^{VI}	3.67	3.71	Ca	0.34	0.29	0.49	Mn	0.64	1.38	1.37	0.65
#Ti^{VI}	0.04	0.04	Na	3.51	3.75	3.42	Ca	0.09	0.12	0.05	0.09
#Fe^{+3}	0.02	0.01	K	0.01	0.08	0.07	Σ	15.89	15.94	15.83	16.02
#Fe^{+2}	0.29	0.30	Σ	19.97	20.14	20.00					
							adr	1.70	2.10	0.90	2.70
O site	4.02	4.06	*an%*	8.80	7.00	12.30	*sps*	11.70	24.60	23.30	18.90
#Ca	-	-					*grs*	0.00	0.00	0.00	11.00
#Na	0.31	0.17					*alm*	86.60	73.30	75.80	78.40
#K	1.61	1.65									
A site	1.92	1.82									
#O	20.02	20.01									
#OH	3.98	3.99									

Tab. 1 - Microprobe analysis of some minerals from spindles in fine grained M. Filau gneisses: white micas (col. 1-2), plagioclase (col. 3-5) and garnets (col. 6-9).

bands of thin crystals, always included in white mica flakes. Sometimes it occurs along the rim of andalusite crystals, thus indicating its later crystallization. It is very abundant in some samples, along the schistosity. Biotite only occurs in some samples, in the form of small isoriented flakes, partially decolored, containing fine-grained ores, with more or less well-developed rims of white mica. This white mica is occasionally more abundant and may also contain a fine felt of sillimanite.

The blackish, medium-grained spindles are mainly composed of tourmaline aggregates; the white ones are composed of andalusite, quartz and white mica, with minor quantities of potassic feldspar and (albitic) plagioclase. In these aggregates the single andalusite crystals form a few large domains in which they are isoriented, as if they were old porphyroblasts containing large quartz crystals. Andalusite crystals are commonly replaced by white mica, which also makes up large intergranular flakes. The composition of the large intergranular flakes is shown in table 1; the high TiO_2 contents suggest a magmatic crystallization (Fig. 3) (Monier et al., 1984) even though MgO content is extremely low, depending on protolith bulk composition. Although only in small quantities, interstitial potassic feldspar and plagioclase (euhedral crystals of albitic composition, never included in muscovite; see Tab. 1) always occur. The above described spindles sometimes also contain garnet, biotite and sillimanite, the latter always included in the white mica replacing andalusite.

c) *The cataclastic gneisses*

These rocks have a characteristic structure, locally of augen type, defined by alternation of medium- to coarse-grained bands and fine-grained layers. These rocks are better defined as mylonites (Bell & Etheridge, 1973; Spry, 1974).

The medium-grained bands are normally polymineralic, and the crystals are highly deformed in them. Pecilitic microcline perthites sometimes occur with cemented fractures, as well as irregularly shaped quartz and rare saussuritized plagioclase. Garnet is occasionally found, always in deformed but never altered idioblasts. Monomineralic augen are frequent and may either consist

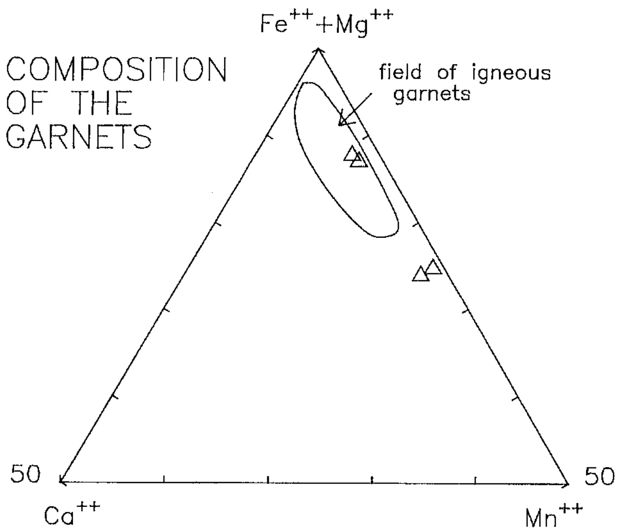

Fig. 2 - Composition of garnets in fine-grained gneisses. Igneous field of natural and synthetic garnets of granitic systems (Green, 1977).

of microcline perthitic single crystals, or large flakes of muscovite (sometimes intensely deformed and with sericitized rims), or quartz aggregates, or rare oligoclase single crystals.

The fine-grained layers are made up of quartz, microcline, albite and sericite, the latter sometimes associated with chloritized, decolored biotite containing abundant small granules of ores.

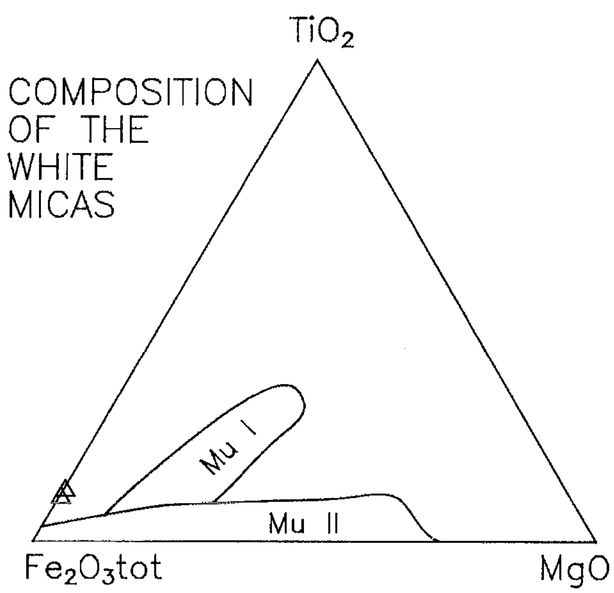

Fig. 3 - Composition of muscovites from andalusite-bearing lenses in fine-grained gneisses. Mu I and Mu II fields are those of magmatic and secondary muscovites respectively according to Monier et al., 1984.

4. Chemistry of M. Filau Gneisses

Twenty-eight samples representing the three main rock types were analysed.

The analytical data are shown in table 2 and in the Harker's diagrams of figure 4. They are similar to those of low Ca granitoids having high normative corundum. The data points concerning the medium- to coarse-grained and the cataclastic gneisses fall in the same fields, while the leucocratic fine-grained gneisses fall within a smaller, Si-richer and Ti, Fe, Mg poorer area of the same compositional fields.

Assuming that during metamorphism Q:Or:Ab ratios did not change significantly, the mesonorm of Mielke & Winkler (1979) was calculated and the Q, Or, Ab values plotted in the Q-Or-Ab (Fig. 5) projection of the Q-Or-Ab-An tetrahedron (Winkler et al., 1975), at $P_{tot}=P_{H_2O}=5$ Kb. Most of the data points plot along or near the cotectic line.

Some of the data points fall in the quartz field. This fact may be due to a loss of alkalis during metamorphism, as also suggested by the high values of normative corundum. *c*-values in the M. Filau gneisses are in the range 2-7, close to those of "granites rich in restites" (c = 4-6: Clemens & Wall, 1981); however the M. Filau gneisses are very poor in xenoliths and restitic minerals, showing that the high c values may be an effect of synmetamorphic alkali depletion. The Sr contents are unusually low and Rb is high, in comparison with the values commonly found in anatectic granitoids. This feature may be related either to geochemical magmatic fractionation or to metamorphic alteration (e.g. crystallization of white mica after feldspars).

5. Discussion

The above petrographic and geochemical observations allow us to reconstruct the metamorphic evolution of the M. Filau gneisses and to estimate the P and T values during the different metamorphic stages.

The oldest recorded event produced the schistosity of these rocks, and the characteristic appearance of andalusite, sillimanite and garnet. The And → Sil sequence is to be considered as a typical prograde metamorphic sequence of low pressure, rather than a relic of the magmatic mineral assemblage.

More detailed information on this old event may be obtained by considering the polymineralic spindles occurring in the leucocratic gneisses. The crystallization of large muscovite flakes after andalusite in the spindles requires water and high temperature. Water was made available from the andalusite-producing dehydration of the magmatic muscovite. Temperature may have been as high as that required in the KNASH system by melt producing reactions like b and d shown in figure 6. Therefore, beginning anatexis may have been possible. Where the magmatic muscovite was in contact with Qtz, Ab and Kfs, melt pockets and andalusite were produced at temperatures of about 650°C and pressures higher then 3.5 Kb (Fig. 6). In agreement with the experimental data of Brown & Fife (1970), the first melt should have incorporated the available water making further melting impossible and, after further increase in temperature, allowing to crystallization of sillimanite (fibrolite). Then, the cooling history began, the melts crystallized and in the latest stage muscovite formed at the expenses of andalusite, in large flakes due to high H_2O pressure and still high temperature.

The second event involved only a small part of the gneissic complex and produced cataclasites and mylonites. The minerals related to this stage, specifically sericite + albite + quartz, are fine-grained and indicate temperatures around 400°C.

The last event recorded in the M. Filau gneisses and their mylonites is confined to the immediate neighbourhood of the Hercynian intrusions. The related paragenesis (albite + muscovite + biotite) indicates relatively low temperatures in this case too.

6. Conclunding Remarks

New field observations combined with petrographic and chemical data indicate the plutonic nature of the protoliths of the M. Filau gneisses. Therefore, these gneisses were originally granitic rocks.

Their composition is near that of the granitic minimum for medium to low pressure values. Their original peraluminous character, which is consistent with the fact that the xenoliths are only pelitic in composition, was

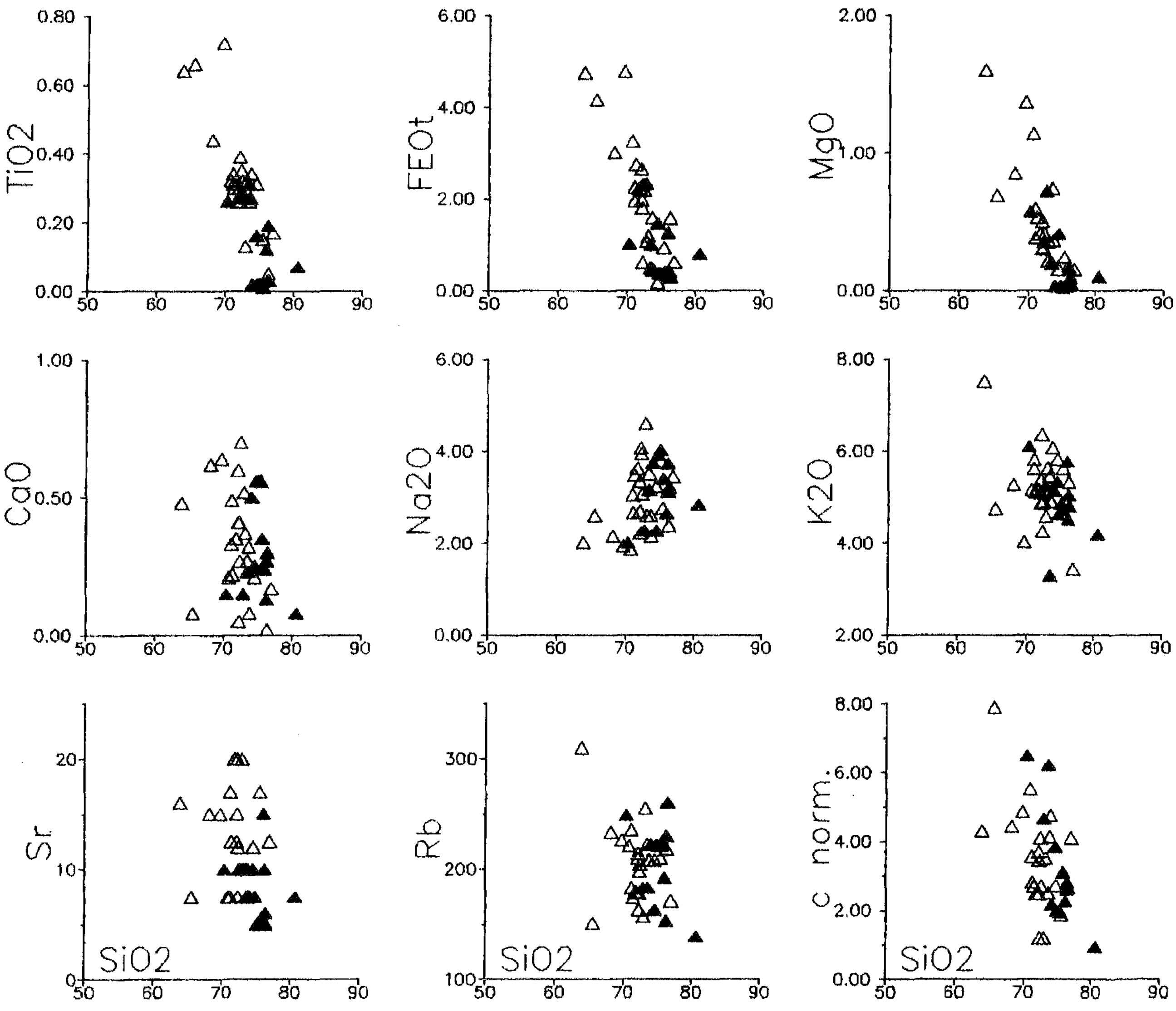

Fig. 4 - Harker diagrams for major elements Rb and Sr and normative corundum. Symbols as in Fig. 5.

Sample	151-A	154-A	161-A	249-A	250-A	97-B	162-B	185-B	186-B	294-B	301-B	302-B	308-B	324-B	326-B	155-C	156-C	157-C
SiO_2	76.20	72.44	72.16	72.10	71.16	65.53	73.66	72.20	73.82	76.30	74.52	72.40	73.15	71.18	73.51	75.60	75.14	74.79
TiO_2	0.19	0.32	0.30	0.39	0.34	0.66	0.27	0.26	0.34	0.05	0.31	0.35	0.32	0.31	0.26	0.01	0.02	0.02
Al_2O_3	13.67	13.99	14.53	14.16	14.41	17.03	13.93	13.98	14.84	12.41	14.58	15.46	14.26	14.47	14.53	14.23	14.41	14.35
Fe_2O_3	0.28	1.10	1.83	2.53	1.95	4.07	0.66	1.76	0.26	1.55	0.07	0.38	0.83	0.75	0.35	0.20	0.15	0.19
FeO	0.16	1.20	0.32	0.36	0.20	0.50	1.04	0.21	0.22	0.19	0.12	0.27	0.44	1.58	0.18	0.23	0.22	0.20
MnO	0.01	0.02	0.01	0.03	0.02	0.02	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.02	0.01	0.02	0.02	0.01
MgO	0.15	0.41	0.30	0.50	0.38	0.69	0.74	0.30	0.36	0.09	0.15	0.35	0.35	0.59	0.20	0.03	0.02	0.03
CaO	0.13	0.70	0.60	0.41	0.33	0.08	0.32	0.05	0.08	0.02	0.21	0.27	0.37	0.49	0.27	0.35	0.56	0.56
Na_2O	3.73	3.07	3.34	2.69	2.65	2.60	2.59	2.22	2.15	2.38	3.24	3.95	2.60	3.05	3.51	3.40	4.02	3.92
K_2O	4.51	4.91	5.37	5.08	5.82	4.75	4.89	6.35	6.07	5.33	5.81	4.25	5.63	5.62	5.45	4.64	4.58	4.62
P_2O_5	0.07	0.13	0.14	0.06	0.13	0.06	0.11	0.07	0.03	0.02	0.05	0.06	0.10	0.12	0.04	0.04	0.05	0.04
Total	99.10	98.29	98.90	98.31	97.39	95.99	98.23	97.41	98.18	98.36	99.07	97.75	98.06	98.18	98.31	98.75	99.19	98.73
LOI	1.04	1.35	1.37	2.09	1.97	3.38	1.65	1.96	1.81	1.21	1.20	1.69	1.41	1.34	1.22	0.72	0.69	0.81
Rb	153	204	214	210	183	151	208	178	208	218	175	198	255	235	222	221	221	222
Sr	10	10	13	15	18	8	10	8	8	5	13	13	10	8	17	5	5	8
Q	39	40	35	42	37	40	45	39	42	45	35	36	40	36	34	40	35	36
OR	27	30	34	32	37	32	30	40	38	33	35	26	35	34	34	29	28	29
AB	34	30	31	26	26	28	25	21	20	22	30	38	25	30	32	31	37	35

Sample	158-C	163-C	150-C	152-C	214-C	215-C	218-C	159	160	164	165	166	153	168	171	172	174	180
SiO_2	74.05	70.38	74.52	72.78	76.40	76.27	76.00	73.51	76.95	71.76	71.40	72.28	80.59	63.88	68.16	69.70	75.40	72.93
TiO_2	0.02	0.26	0.16	0.27	0.03	0.03	0.12	0.31	0.17	0.26	0.30	0.28	0.07	0.64	0.44	0.72	0.15	0.13
Al_2O_3	14.71	16.20	13.46	13.99	13.31	13.57	13.10	15.07	13.64	14.17	13.99	13.64	10.30	16.32	14.43	13.34	13.34	14.26
Fe_2O_3	0.22	0.76	1.85	2.23	0.15	0.13	1.05	0.81	0.44	2.14	2.70	2.32	0.82	3.89	2.75	4.25	0.73	0.74
FeO	0.23	0.34	0.25	0.31	0.15	0.17	0.18	0.26	0.22	0.30	0.31	0.21	0.09	1.25	3.18	0.96	0.27	0.40
MnO	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.01	0.01	0.02	0.01	0.03	0.01	0.03	0.03	0.03	0.01	0.01
MgO	0.03	0.57	0.41	0.72	0.05	0.05	0.17	0.19	0.15	0.41	0.53	0.31	0.10	1.60	0.85	1.37	0.24	0.21
CaO	0.50	0.15	0.25	0.15	0.30	0.27	0.24	0.23	0.17	0.35	0.22	0.41	0.08	0.48	0.62	0.64	0.56	0.52
Na_2O	3.74	2.01	2.27	2.27	3.20	3.12	2.64	3.15	3.44	3.62	3.47	4.07	2.84	2.01	2.15	1.94	2.76	4.60
K_2O	5.18	6.10	5.32	5.24	4.79	5.04	5.77	3.29	3.43	5.18	5.12	4.87	4.19	7.52	5.28	4.03	5.62	4.58
P_2O_5	0.04	0.11	0.10	0.11	0.12	0.12	0.07	0.05	0.04	0.20	0.13	0.11	0.10	0.08	0.08	0.04	0.06	0.17
Total	98.73	96.89	98.60	98.08	98.51	98.78	99.37	96.88	98.66	98.41	98.18	98.53	99.19	97.70	97.97	97.02	99.14	98.55
LOI	0.76	2.14	1.50	2.08	0.66	0.68	0.95	2.55	1.51	1.56	1.62	1.08	0.61	2.25	1.72	2.53	0.81	1.09
Rb	222	249	163	183	261	231	192	183	171	179	174	163	139	310	233	226	-	157
Sr	8	10	10	10	3	5	15	8	13	20	13	20	8	16	15	15	-	20
Q	34	41	45	45	42	41	41			34	35	32			43		40	30
OR	32	39	33	33	30	31	36			32	32	38			35		35	28
AB	34	20	22	22	28	28	23			36	33	30			22		25	42

Tab. 2 - Major element (wt%), Rb and Sr data of M. Filau orthogneisses. A and B: medium to coarse grained gneisses; C: fine grained gneisses; others: cataclastic gneisses.

enhanced by metamorphic processes. The anatectic-crustal hypothesis which can be formulated basing on our new data does not fit derivation from an alkaline magma, as proposed by Delapierre & Lancelot (1989), on the basis of zircon typology and normative mineral composition. It should also be noted that, in interpreting the genesis of these rocks, the composition of the feldspars cannot be used because it was controlled by metamorphism.

The observed mineralogical and structural alterations suggest the following metamorphic and deformational history:

a) medium- to high-grade metamorphism (P=3.5-4.8 Kb and T>650°C) which caused the following effects: (i) gneissification of the original granitic body; (ii) crystallization of andalusite, sillimanite and garnet; (iii) local anatexis in the muscovite sites within the leucocratic fine-grained gneisses;

b) local cataclasis and mylonitization, with related local recrystallization: the new paragenesis (Qtz+Ab+Ms±Chl) indicates low temperatures;

c) contact metamorphism around the late-Hercynian granites: at the maximum grade observed, this event produced aggregates of white micas with decussate structure and large biotite flakes.

The above outline reveals the contrasting character of the first two events. The first one was high-grade, with all features which may be found in an intermediate crust affected by high thermal gradient (sequence And → Sil).

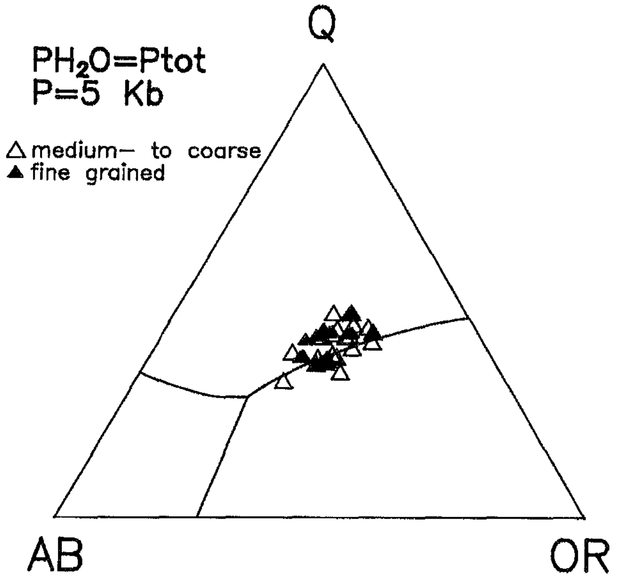

Fig. 5 - Location of the data points in the Q-Or-Ab triangle, supporting the medium to low pressure character of the melts.

The second event only display deformational-cataclastic and low-temperature effects and may find an appropriate location in a shallower crustal level.

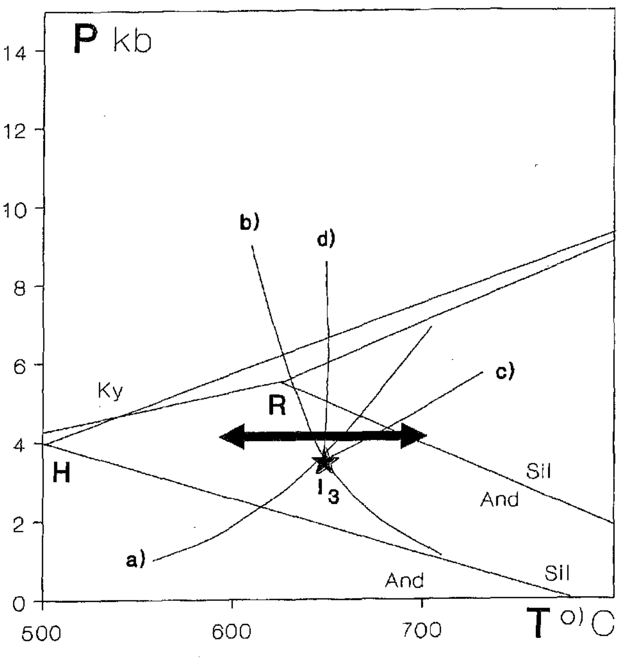

Fig. 6 - Sketch of the possible melting and cooling paths (bold arrow). Reaction grid is from Thompson & Algor (1977). a) Ms + Ab + Qtz = Kfs + Al_2SiO_5 + V; b) Ms + Kfs + Ab + Qtz + V = L; c) Ms + Qtz + L = Kfs + Al_2SiO_5 + V; d) Ms + Ab + Qtz + V = Al_2SiO_5 + L. H and R represent the location of triple point according to Holdaway (1971) and Richardson (1969), respectively.

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PETROGRAPHY AND GEOCHEMISTRY OF SOME MIGMATITES FROM NORTHEASTERN SARDINIA (ITALY)

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ABSTRACT

The detailed petrographical and geochemical studies carried out on seven samples of migmatites from northeastern Sardinia, outline a complicated petrological picture that suggests that migmatites probably formed in response to different processes. In particular, both congruent and incongruent partial melting reactions can be envisaged in the different type of migmatites cropping out at Punta Ottiolo /Punta de li Tulchi, while the peculiar mineralogical and geochemical composition of the leucosome of the stromatic migmatite cropping out at Brunella induce to think that, there, leucosomes probably formed through a process of subsolidus quartzo-feldspatic differentiation.

1. INTRODUCTION

Preliminary geochemical data on migmatites cropping out at Punta Ottiolo/Punta de li Tulchi and Brunella (south of Olbia), in the northeastern Sardinia, are presented and discussed.

The two areas belong to the Sillimanite+K-feldspar zone (Fig.1) of the Hercynian axial zone metamorphic basement (Franceschelli et al. 1982) and are composed of stromatic migmatites, nebulites, orthogneisses, Biotite-Sillimanite gneisses, calc-silicate rocks and retrogressed eclogite .

Field studies enabled Franceschelli et al.(1982,1989) and Elter & Sarria (1989) to recognize in the areas two main deformational events, the first of which is more important than the later one. The main foliation, observed in the field, is linked to the first deformational event. Mineralogical and structural relics, such as armoured kyanite grains within plagioclase or intrafolial hinges transposed along the main foliation, have been found in some gneisses and mesosomes and suggest a previous tectono-metamorphic history.

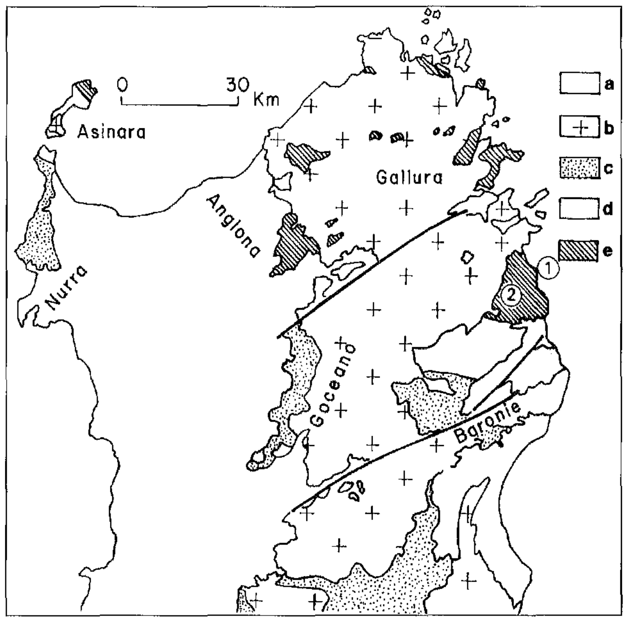

Fig.1 - Schematic map of northern Sardinia (after Carmignani et alii, 1982 partly modified). a: Post-Hercynian volcanics and sediments; b:Granitoids; c: Greenschist facies rocks; d: Amphibolite facies rocks; e: Migmatites with calc-silicate pods and bodies with relics of granulite and/or eclogite facies parageneses. 1) Punta Ottiolo/Punta de li Tulchi. 2) Brunella.

2. PETROGRAPHY

Seven samples from the outcrops of Punta de li Tulchi, Punta Ottiolo and Brunella were selected for detailed petrologic and geochemical studies. Locality, lithology and leucosome composition are shown in Tab.1 (nomenclature from Henkes & Johannes 1981).

The migmatitic gneiss and the mesosomes of stromatic migmatites are mainly composed of Bt+Pl+Qtz±Kfs (mineral symbols according to Kretz, 1983); sillimanite, garnet and muscovite are often present. Zircon, monazite and apatite are the most common accessory minerals. The leucosome RS2B (Tab. 1) consists of Qtz+Kfs+Pl with rare biotite and idioblastic garnet while leucosome RS9B is constituted by Pl+Qtz with rare K-feldspar, garnet, sillimanite and relics of kyanite.

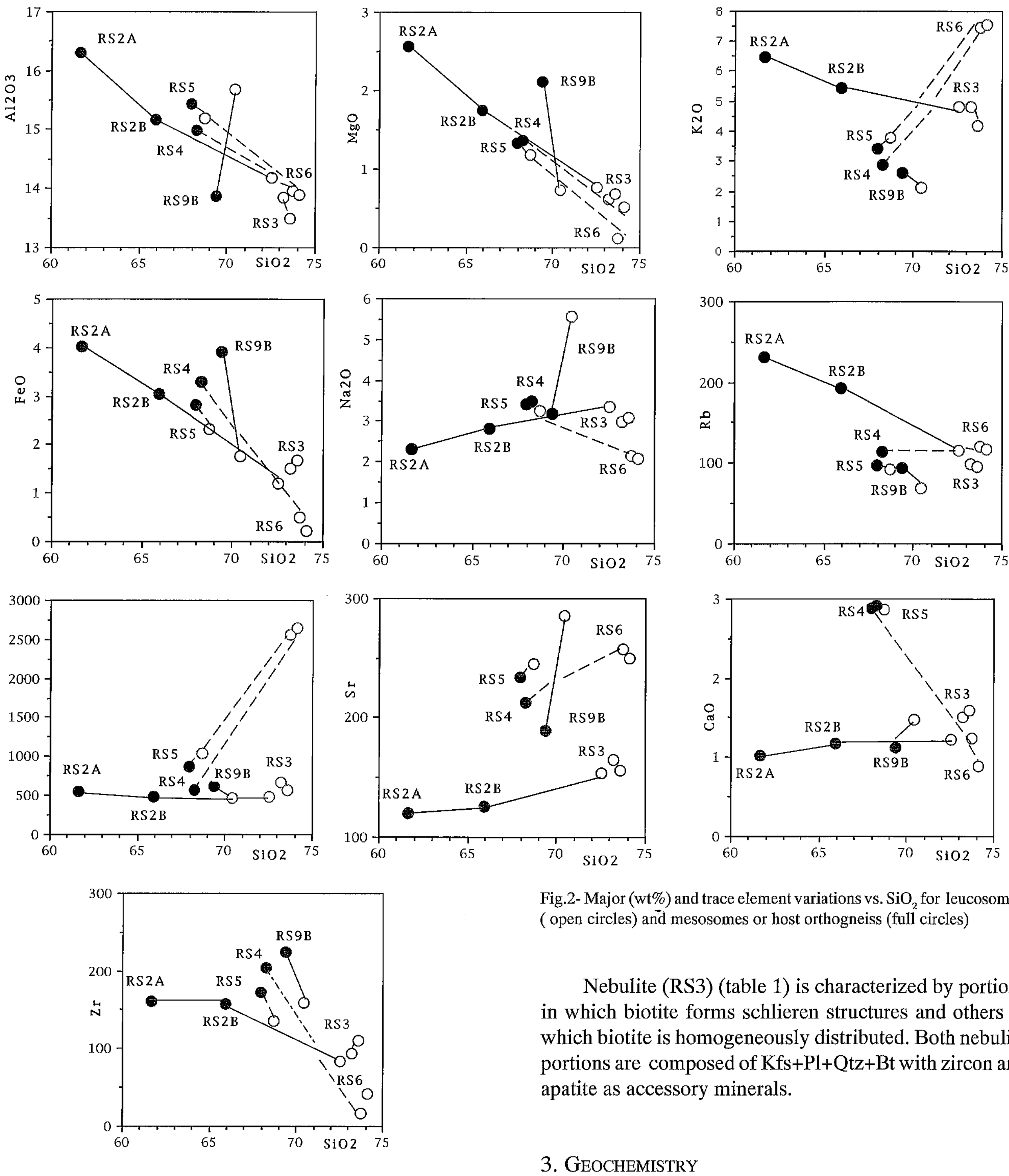

Fig.2- Major (wt%) and trace element variations vs. SiO_2 for leucosomes (open circles) and mesosomes or host orthogneiss (full circles)

Samples RS4 and RS5 consist of Qtz+Pl+Kfs with biotite, muscovite and very rare garnet. This lithology hosts three types of leucosomes: i) the first, granodioritic in composition, is associated with sinistral shear zones and outlines dictyonitic textures; ii) the second consists of a leucosome which constitutes disharmonic folds subconcordant with respect to the main foliation; iii) the third is definitively discordant in respect to all structures of the orthogneiss. The last two leucosomes are granitic with abundant K-feldspar, very rare biotite and relics of garnet.

Nebulite (RS3) (table 1) is characterized by portions in which biotite forms schlieren structures and others in which biotite is homogeneously distributed. Both nebulite portions are composed of Kfs+Pl+Qtz+Bt with zircon and apatite as accessory minerals.

3. Geochemistry

The samples were treated as follows: they were serially sawn, in each slice, leucosome was separated both from mesosome in stromatic migmatites and from the host rock in orthogneiss. The nebulite was subdivided into two parts, the first containing biotite-schlieren and the second with homogeneously distributed biotite. All leucosomes are Corundum normative with a $Al_2O_3/CaO+Na_2O+K_2O$ molecular ratio (ASI) ranging from 1.04 to 1.10 without apatite correction.

Figures 2 and 3 report the variation diagrams of the

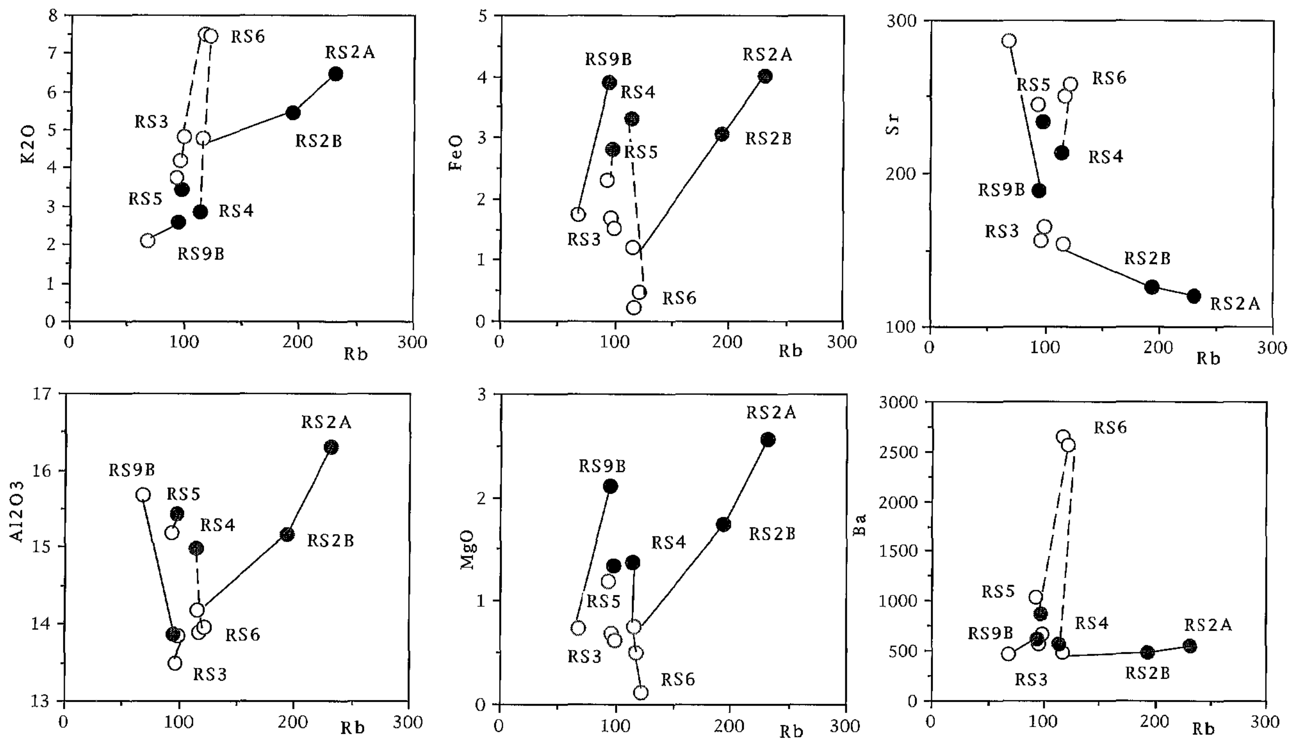

Fig. 3 - Major (wt%) and trace elements variations vs. Rb (symbols as Fig.2) .

major and trace elements in respect of SiO_2 and Rb respectively, between leucosomes and corresponding mesosomes or hosts orthogneisses. Figure 4 reports the chondrite-normalized REE patterns of samples RS2B, RS9B, RS5 and RS6. In the stromatic migmatites, leucosomes have lower and mesosomes correspondingly higher content of Si, Al, K, Mg, Ti, Fe, P, Rb, Zr, Nb, and total REE; while Na, Ca, Sr, Ba, Al(RS9B) is higher in the leucosomes than in the mesosomes (Fig.2 and 3). This reflects the partitioning of ferromagnesian and accessory phases in the mesosomes and of quartz and feldspars in the leucosomes.

Migmatitic gneiss (RS2A) shows higher K,Al, Mg, Ti, Fe, P, Rb, Zr, Nb, and lower Na, Ca, Sr, Ba, with respect to the mesosome RS2B. For most of the considered elements the migmatitic gneiss RS2A, the mesosome and leucosome of sample RS2B form a straight line (Fig.2 and 3).

The two parts of nebulite (RS3) show little chemical differences and both plot near the leucosome RS2B.

Dictyonitic leucosome (RS5) shows virtually the same chemical features as the host orthogneiss. With respect to the latter, it has only a minor increase in the K, P, Sr, Ba content reflecting the partitioning of K-feldspar and apatite in the dictyonitic leucosome (Fig 2 and 3).

The leucosomes forming subconcordant disharmonic folds and the discordant ones show lower Na, Ca, Mg, Ti, Fe, Zr, and higher K, Sr, Rb, Ba, with respect to the host, reflecting the partitioning of plagioclase, biotite and zircon in the restitic portion and of K-feldspar in the leucosome (Fig.2 and 3). Host orthogneisses and the different associated leucosomes usually form straight lines for most of the considered elements (Fig.2 and 3).

Total REE content of the leucosomes is always lower with respect to mesosomes and host orthogneiss. With the exception of sample RS9B, all the specimens show a moderately fractioned pattern (Fig.4). All the samples show a $(La/Sm)_N$ ratio ranging from 2.5 to 4.4; the $(Gd/Yb)_N$ ratio ranges from 0.6 to 0.9 for leucosomes RS6, RS3, RS2B, and from 1.3 to 2.09 for mesosomes and host orthogneiss RS5, RS9B, RS2B. Europium anomalies are always negative with (Eu/Eu*)=0.5-0.9.

SAMPLE	LOCALITY	LITHOLOGY	LEUCOSOME COMPOSITION
RS2A	Punta de li Tulchi	Migmatitic Gneiss	---------
RS2B	" " " "	Stromatic Migmatite	Granitic
RS3	" " " "	Nebulite	Granitic
RS4	Punta Ottiolo	Orthogneiss with discordant leucosome	Granitic
RS5	" "	Orthogneiss with dictionitic leucosome	Granodioritic
RS6	" "	Discordant leucosome in RS5	Granitic
RS9B	Brunella	Stromatic Migmatite	Trondjemitic

Tab. 1

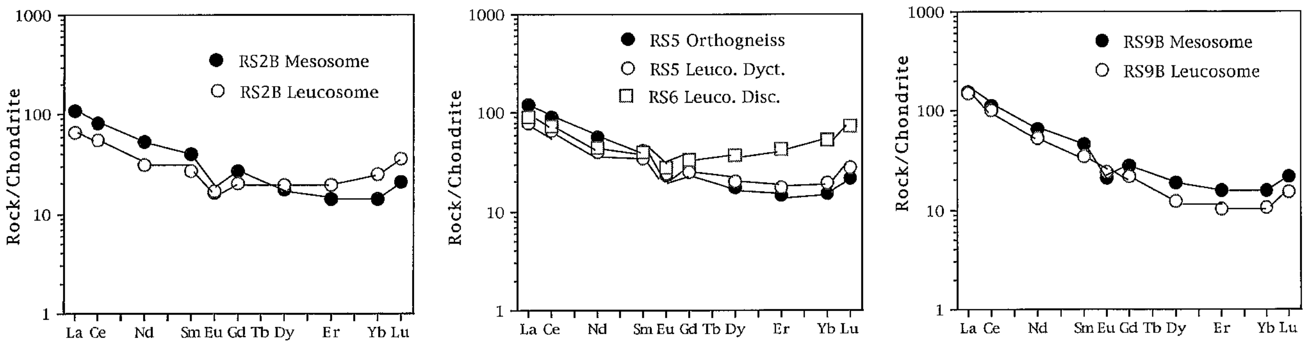

Fig.4 - Chondrite-normalized REE patterns (chondritic values from Evensen et alii,1978).

Leucosome RS9B shows a REE pattern subparallel to the one of the corresponding mesosome with a significant decrease of the Europium anomaly (Eu/Eu* = 0.58 in respect to 0.91).

The other leucosomes show LREE pattern subparallel to that of the corresponding mesosome and host orthogneiss. The HREE pattern is different as far as the leucosomes which have higher HREE contents with respect to mesosomes or host orthogneiss. Infact, the $(Gd/Yb)_N$ ratio is lower in leucosomes than in mesosomes. This fractionation is particularly evident in the samples RS5 and RS6, where host orthogneiss has $(Gd/Yb)_N$=1.74, while the discordant leucosome has $(Gd/Yb)_N$ =0.64.

4. Conclusions

The data presented for migmatitic rocks of northeastern Sardinia outline a complicated petrographic and geochemical picture that suggest that they probably formed in response to different processes.

However, using textural features, petrographic and geochemical data, some genetic hypotheses on the formation of leucosomes can be put forward. Leucosome of the stromatic migmatite RS2B can be regarded as a product of in-situ partial melting in which the parent rock is represented by the migmatitic gneiss RS2A. A dehydration melting of the biotite-muscovite-quartz assemblage through a model reaction such as Bt+Ms+Qtz=Grt+Kfs+L (Thompson, 1982) can be envisaged.

Nebulite RS3 and leucosomes RS4 and RS6 might be formed through the same mechanism, while the former possibly had a parent rock similar to the migmatitic gneiss RS2A and the latter two a parent rock like the orthogneiss.

Dictyonitic leucosome RS5 which formed along the shear zone shows chemical and mineralogical similarities to the host orthogneiss. In this case, a congruent melting reaction involving the quartzo-feldspatic component of the host orthogneiss, possibly favoured by fluids channeled along the shear zone, can be assumed.

Finally, due to the particular mineralogical and geochemical composition, the leucosome RS9B probably formed through a process of subsolidus quartzo-feldspatic differentiation.

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FROM CRUSTAL THICKENING TO EXHUMATION: PETROLOGICAL, STRUCTURAL AND GEOCHRONOLOGICAL RECORDS IN THE CRYSTALLINE BASEMENT OF NORTHERN SARDINIA

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Abstract

Northern Sardinia represents the axial zone of the Hercynian chain of Corsica and Sardinia. The metamorphic complex is well exposed along the northeastern and northwestern coasts of Sardinia where the continuity from low to high-grade metamorphics can be followed. In the other areas it is fragmented and consists of scattered outcrops of medium- to high-grade rocks "floating" in the Hercynian batholith. Widespread preserved records indicate that this part of the chain underwent crustal thickening consequent to continental collision. An age of $\geq$ 344 Ma, measured for a stromatic migmatite, is tentatively attributed to this stage.

After thickening, the metamorphic rocks were uplifted towards the surface. The process completed in a relatively short time as documented by Permian volcanites resting on medium- to high-grade rocks and radiometric dating.

Available petrological, structural and geochronological records are used to reconstruct the exhumation history of this crystalline basement. It arises that the various portions of the metamorphic basement followed different P-T-t paths which are interpreted as evidence of differential uplift and different unroofing processes.

The different processes, tentatively assumed to have operated during the exhumation, are:
— erosion and isostasy (northeastern Sardinia)
— tectonic unroofing (western Gallura: Giuncana-Tungoni area)
— extensional tectonics (Palau-S.Teresa di Gallura area).

1. Introduction

This note reports a few ideas presented at the meeting "Geologia del Basamento Italiano" held in Siena, 21-22 March 1991, to commemorate our friend Tommaso Cocozza who passed away on July 26, 1989.

In choosing a topic like the exhumation history of the crystalline basement of northern Sardinia, I was strongly influenced by a discussion Tommaso and I started in 1977 during a field trip organized in Sardinia by the working group "Paleozoico" of the Italian National Research Council. The main point was how to reconcile his stratigraphic and my petrologic approach to the Hercynian orogenic belt of Sardinia. In those years, we were collecting early radiometric data and performing the first thermobarometric estimations of the metamorphism in northeastern Sardinia. What was intriguing for Tommaso was to accept the idea that the whole process of regional metamorphism, plutonism and exhumation in northern Sardinia developed in a relatively short time (40-50 Ma) (Ferrara et al., 1978), and, in particular, that the removal of several-kilometer-thick sequences above the medium- to high-grade rocks - we assumed without demonstrations "through processes of both erosion and tectonic denudation" - did not leave any evidence in the upper Carboniferous-Permian sedimentary sequences (lack of Hercynian molasse). Moreover, at that time, a large part of the geological community did not only refuse the idea that the Hercynian orogeny implied deep-crustal processes but also the nappe structure of the chain.

In this paper I review the available - indeed scarce - petrological, structural and geochronological data of northern Sardinia and their possible meaning on the uplift history of this basement. Aims were, on one hand, to stress the potentiality of P-T-t trajectories to solve geotectonic problems and, on the other hand, to show that petrological, structural and geochronological data are still too scarce to conclusively support any tectonic model able to reconstruct the evolution of the Hercynian chain in Sardinia.

2. Regional geology

Corsica and Sardinia form a complete section of a segment of the Hercynian chain from the foreland, in northern Corsica, to the axial zone, in southern Corsica and northern Sardinia, and to the external zone in southwestern Sardinia (Carmignani et al., 1984, 1986). In Sardinia, the metamorphic grade slightly changes from subgreenschist facies to greenschist facies, moving from the external to the nappe zone. In the axial zone, over a distance of about 50 km, the grade rapidly increases - in a northeastern direction - from greenschist to amphibolite facies with migmatites (Fig. 1).

The metamorphic complex is well exposed along the northeastern and northwestern coasts of Sardinia where continuous transition from low to high-grade metamorphics can be followed. In the other areas, it is fragmented and consists of scattered outcrops of medium-or high-grade metamorphics "floating" on the Hercynian batholith (Carmignani et al., 1984).

In northern Sardinia, metasedimentary rocks (pelitic, quartz-feldspathic and minor carbonatic) largely prevail over metaigneous rocks (metarhyolite, orthogneiss, amphibolite). The only palaeontologically documented ages are those of the Ordovician-Silurian terrains in northwestern Sardinia. Emplacement ages of the protoliths of granodioritic and granitic orthogneisses are 458 $\pm$ 31 and 441 $\pm$ 33 Ma, respectively (Ferrara et al., 1978). The involvement of Precambrian rocks has repeatedly been advocated, but never conclusively demonstrated (Beccaluva et al., 1985).

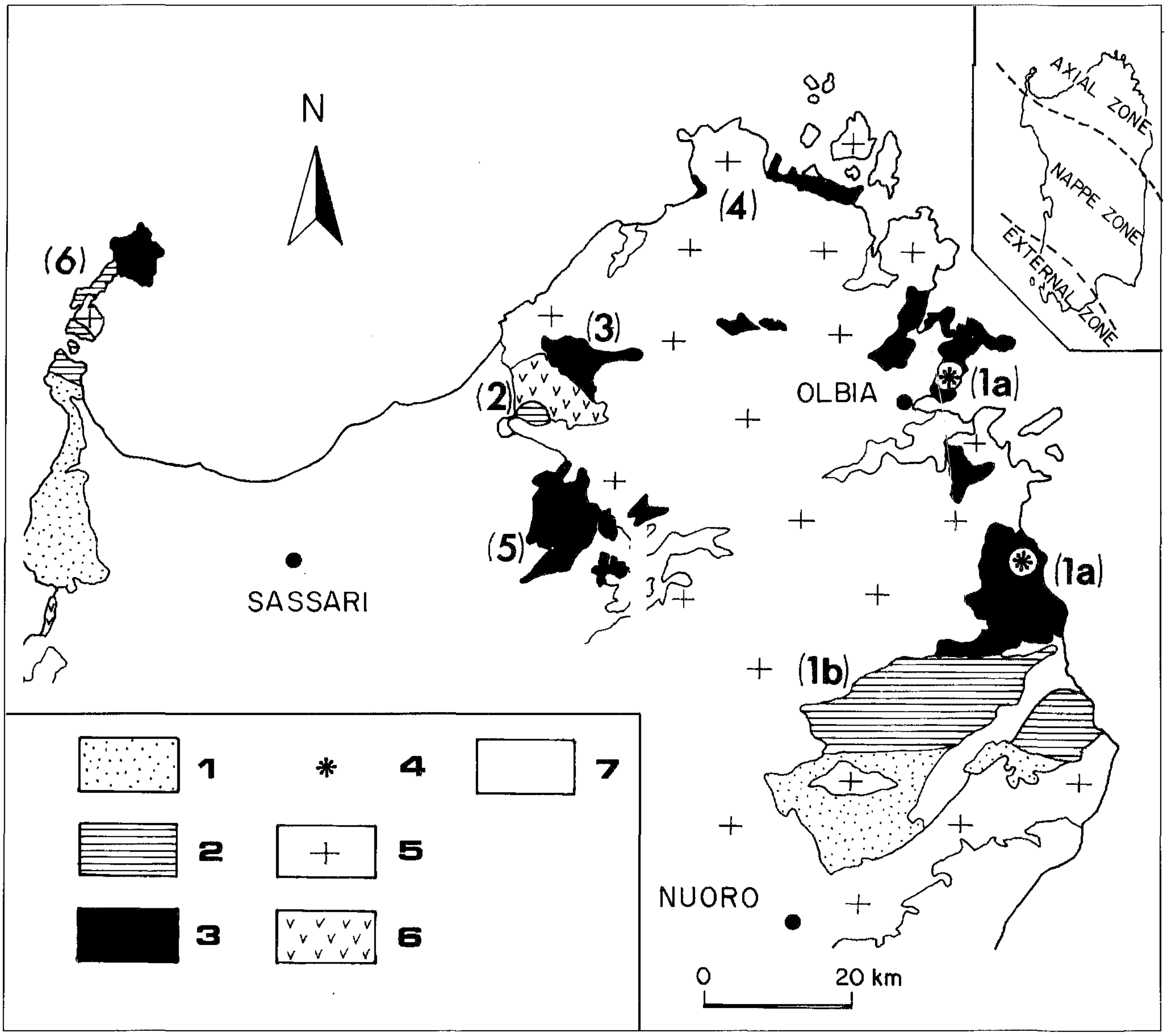

Fig. 1 - Sketch of the zones of regional metamorphism of northern Sardinia (simplified after Franceschelli et al. 1989): 1) low-grade zones (Chlorite, Biotite, Garnet + Albite zones); 2) medium-grade zones (Garnet + Oligoclase, Kyanite + Biotite, Staurolite + Biotite zones); 3) high-grade zones (Sillimanite + Muscovite, Sillimanite + K-feldspar zones); 4) pods and lenses of eclogites and/or mafic granulites; 5) Hercynian granitoids; 6) Permian sediments and volcanites; 7) Post-Permian sediments and volcanites. Numbers within parenthesis refer to the various portions of the basement considered in this paper: 1a) northeastern Sardinia (high-grade zones); 1b) northeastern Sardinia (medium-grade zones); 2) Western Gallura: Giuncana-Tungoni area; 3) Western Gallura: Tarra Padedda area; 4) Palau-S.Teresa di Gallura area; 5) Anglona; 6) Asinara Inlet.

3. Metamorphic evolution of the Sardinian basement

3.1. *Metamorphic zones: The classical overview*

Seven metamorphic zones have been mapped on the basis of mineral assemblages in pelitic and quartz-feldspathic rocks (Franceschelli et al., 1982). Moving from south to north (Fig. 1) they are Chlorite, Biotite, Garnet, Staurolite + Biotite, Kyanite + Biotite,.Sillimanite + Muscovite and Sillimanite + K-feldspar zones.

The transition from medium- (Staurolite and Kyanite zones) to high-grade zones (Sillimanite + Muscovite and Sillimanite + K-feldspar zones) is continuous from the metamorphic point of view, but abrupt from the lithological point of view, as it is characterized by a rapidly decreasing amount of pelitic schists and appearance of more or less migmatitic quartz-feldspathic gneiss. Furthermore, the high-grade complex contains pods and lenses of retrogressed mafic granulite and eclogite, unknown in the medium-grade complex.

3.2. *Metamorphic zones: A critical discussion*

The basement of northern Sardinia experienced polyphase Hercynian deformation. Three superimposed systems of folds (D_1, D_2, D_3), characterized by different axial orientation and associated with three axial planar foliations (S_1 S_2, S_3), have been identified (Carmignani et al., 1986).

Textural and petrological relationships underline diachronous mineral growth in the various zones during the main D_1 and D_2 deformational episodes: in the low - medium-grade zones the index minerals (biotite, garnet, staurolite) grew preferentially, often as porphyroblasts, after D_1 and during early D_2, with matrix minerals aligned in the S_2 foliation; in the high-grade zones the main parageneses are parallel to the S_2 foliation and the porphyroblasts grew during late D_2 (Franceschelli et al., 1982, Palmeri, 1991). In the high-grade complex, evidence of pre-D_2 (S_2) history is represented by rounded and armored relics of staurolite, garnet and kyanite - generally included in plagioclase - and by relics of granulitic and eclogitic parageneses - sometimes present in metabasites (Ghezzo et al., 1979; Miller et al., 1976; Palmeri, 1991).

Therefore, metamorphic reactions are clearly diachronic within the different zones. The large-scale metamorphic zonation represents thus only a complex finite pattern, i.e. the accumulation of various phase trasformations through time. Moreover, when exposed, mylonitic belts have been recognized between different zones.They developed during the uplift of the basement (Elter et al., 1986), but probably reactivated previous thrust belts. I consider therefore that the metamorphic zonation reflects contrasted thermal reequilibration during a progressive orogenic evolution suffered by a stacked pile of crustal slices. As previously underlined by Franceschelli et al. (1989), the metamorphic evolution of the Sardinian basement gives rise to clockwise P-T-t paths consequent to crustal thickening and exhumation during continental collision-type orogeny. In the following paragraphs, I briefly review the metamorphic records of the thickening episode and focus the attention upon the contrasted P-T evolutions related to the uplift history.

4. The thickening stage

The existence of thickening stage is mainly supported by the evidence that most outcrops in northern Sardinia suffered regional high- to medium-pressure metamorphism.

This is particularly true for the medium-grade complex of northeastern Sardinia. Here, physical conditions range from 400°C-9 kbar (Garnet zone) to 550°C-10 kbar (Staurolite zone) (Franceschelli et al., 1989). In the Sillimanite zones, this assumption is suggested by the widespread presence of relict kyanite in gneiss and by the presence of metabasites containing relics of granulite and/ or eclogitic parageneses. Assuming that kyanite coexisted with K-feldspar in migmatite during this stage, the minimum pressure of 10-12 kbar at 700°C was estimated (Ghezzo et al., 1982).

Metamorphic conditions for mafic granulite and eclogite have been estimated to be close to 750°C - 10 kbar (Ghezzo et al., 1979) and 850°C - 17 kbar (Memmi & Palmeri, personal communication), respectively. The evidence of this medium-high-pressure event is also relatively well preserved in western Gallura (Giuncana-Tungoni area). Some relics of kyanite and/or staurolite have been found in Anglona and Asinara Island. This kind of relics have not been found yet in the outcrops of high-grade metamorphics of western Gallura, Palau and S.Teresa areas.

The age of this metamorphism is poorly constrained. The 344 $\pm$ 7 Ma measured by Rb/Sr method for six layers of a stromatic migmatite of NE Sardinia (Ferrara et al., 1978) could provide minimum age as representing closure of isotopic exchange among the different bands. The 350 and 336 Ma obtained by Rb/Sr method on two muscovites of staurolite + kyanite-bearing micaschists of western Gallura (Del Moro et al., 1991) could also indicate a minimum age, as due to cooling below the specific "blocking temperature". I can just propose that this thickening episode developed some time before 344 Ma ago.

5. Contrasted retrogressive evolutions

5.1. *Northeastern Sardinia*

Most of the data on the metamorphic basements of Sardinia actually come from this area. The Hercynian barrovian-type metamorphism and its zonation in northern Sardinia has been recognized here for the first time (Ricci, 1972, 1978) and later extrapolated to the whole of northern Sardinia (Franceschelli et al., 1982).

Franceschelli et al. (1982, 1989) described the petrological effects and estimated the P-T conditions of the retrogressive evolution. According to the metamorphic zone, they consist of a retrogression of variable extent under amphibolite down to greenschist facies conditions.

No regional developments of low-pressure metamorphic minerals has been observed.

Four mineral isochrons have been constructed on two samples of granodioritic orthogneiss, one sample of augen gneiss and one layer of banded migmatite.

The ages obtained are very similar: 304 $\pm$ 1 and 308 $\pm$ 24 Ma for the granodioritic orthogneiss, 306 $\pm$ 10 Ma for augen gneiss, and 298 $\pm$ 2 for the two-point isochron of a banded migmatite (Di Simplicio et al., 1974; Ferrara et al., 1978). Rb/Sr and K/Ar ages have been determined by Di Simplicio et al. (1974) and Ferrara et al. (1978) on separated metamorphic minerals like hornblende, muscovite, biotite. The ages range from 319 to 284 Ma, with an average of 297 Ma.

The results obtained on different minerals and different methods do not show systematic differences, but rather suggest a quite simultaneous and rapid closure of the various isotopic systems.

5.2. *Western Gallura: Giuncana and Tungoni area*

According to Del Moro et al. (1991) and personal communications by G. Oggiano, a metamorphic evolution which is quite similar to that of northeastern Sardinia can be assumed for this area. The intermediate pressure mineralogy is in general well preserved and no low-pressure overprint has been detected during this stage.

The few radiometric age determinations give values of 350 $\pm$ 16 and 336 $\pm$ 8 for two muscovites and 316 $\pm$ 5 and 310 $\pm$ 5 for two biotites (Rb/Sr method: Del Moro et al., 1991). It must be stressed that they are significantly older than those obtained in northeastern Sardinia.

5.3. *Western Gallura: Tarra Padedda*

This outcrop is mainly constituted by high grade rocks (migmatites). Among Al_2SiO_5 polymorphs, only sillimanite has been reported. The retrogressive evolution seems to have been similar to that of high-grade rocks of northern Sardinia. The main difference is the widespread occurence of peraluminous syntectonic (D_2) granitoids (Oggiano & Di Pisa, 1988; Macera et al., 1989). The emplacement age has been assumed to be 300-305 Ma. The cooling age of muscovite (6 samples) ranges from 308 to 298 and the age of biotite (3 samples) from 298 to 293 Ma (Macera et al., 1989).

5.4. *Northern Gallura: Palau - S. Teresa*

Here, several outcrops of high-grade rocks (migmatites) - widespreadly intruded by syntectonic peraluminous granites - occur. As evidence of an early metamorphic episode, the metamorphites contain scarce sillimanite + K-feldspar and garnet. Cordierite is instead widespread in the peraluminous granitoids (Barrabisa, Monte Biancu) and, together with andalusite (+ biotite), overprints the previous metamorphic minerals in migmatites (Innocenti, 1990; Folco, 1991; Ghiribelli, in preparation).

The emplacement age of the granitoids is assumed to be > 300 Ma; the cooling age for muscovite (3 samples) is 299 Ma and the age of biotite 288-290 Ma (Innocenti et al. 1991).

Non-coaxial deformation (extensional shear zones?) which affected peraluminous granitoids and D_2 structures of migmatites, but not the Palau monzogranite (290 Ma; Innocenti et al., 1991), developed under cordierite + andalusite + muscovite stability conditions.

5.5. *Anglona*

The metamorphism is dominated here by low-pressure minerals (andalusite, sillimanite, cordierite) that developed late $D_2 \approx$ syn D_3 and overprint an intermediate-pressure metamorphism evident as relics of staurolite and garnet (Di Pisa & Oggiano, 1987).

According to Del Moro et al. (1991), the appearance of andalusite + K-feldspar is due to prograde dehydration at the expense of muscovite and quartz.

Radiometric Rb/Sr age determinations - limited to one muscovite and two biotites separated from micaschists - yield ages of 303 $\pm$ 6 Ma and 277 $\pm$ 4 - 281 $\pm$ 4 respectively (Del Moro et al., 1991).

5.6. *Asinara Inlet*

Here, a metamorphic history, very similar to Anglona, has been reconstructed by Di Pisa & Oggiano (unpublished data): low-pressure metamorphism with growth of cordierite and andalusite porphyroblasts overprints a medium-pressure metamorphism evidenced by staurolite, garnet and perhaps kyanite relics.

No radiometric data are available.

5.7. *Hercynian Plutonism*

In the latest stages of its evolution, the metamorphic complex was invaded by the huge amount of granitoids constituting the Hercynian batholith. The batholith is a composite one and consists of scanty mafic precursors and syntectonic tonalites and monzo-granodiorites (sometimes with peraluminous character), about 75% of late tectonic intrusions (~10% tonalites and tonalitic granodiorites; ~65% monzogranitic granodiorites to leucomonzogranites) and ~25% of post-tectonic leucogranites (Ghezzo & Orsini, 1982; Carmignani et al., 1984)

All these plutonites emplaced at relatively shallow depth, although they intruded high-grade rocks. The ages obtained by Del Moro et al. (1975) for the three main suites of the Sardinia batholith - tonalitic-granodioritic and monzogranitic-granodioritic suites - (307, 302 Ma, respectively) and the post-tectonic leucogranitic suite (289 Ma) give an idea of the ages of late stages of the regional uplift.

6. Discussion: Tectonic significance of the P - T - t evolutions of the Sardinian basement

The thermomechanical modelling of orogenic processes (England & Thompson 1984; Davy & Gillet 1986; Thompson & Ridley 1987; Ridley 1989) and the calculation of metamorphic P-T-t paths (i.e. Spear & Selverstone, 1983; Droop, 1985) have been extensively used (Collins & Vernon, 1991; Harley, 1989; Gardien et al., 1990; Lardeaux & Spalla, 1991; De Yoreo et al., 1991; Mercier et al., 1991a, Mercier et al., 1991b; Spear & Peacock, 1989; Spear et al., 1984) to derive - from the shape of the P-T trajectories versus time - the contrasting mechanisms assisting the exhumation of the thickened portions of the orogens.

In order to compare and interpret the previously presented metamorphic evolutions, I have tried to extract from the available data (summarized in table 1) significant information for reconstructing the P-T-t paths followed by the rocks of the different outcrops.

6.1. *Constructions of the P-T-t paths*

The temperature and pressure values have been obtained (Franceschelli et al., 1989; Memmi & Palmeri, personal communication) by traditional thermobarometric methods (Ganguly & Saxena, 1984; Hodges & Crowley, 1985; Massonne & Schreyer, 1987; Newton & Haselton, 1981; Richardson, 1968; Ellis & Green, 1979; Erling, 1988; Holland, 1980; Lee & Ganguly, 1987; Newton & Perkins, 1982; Perkins & Chipera, 1985; Sen & Bhattacharya, 1984) and/or estimated by using petrogenetic grids based on well calibrated reactions. The most important of these reactions are reported in fig. 2 (simplified from Yardley, 1989).

In order to construct the P-T-t paths for the different outcrops I made the following assumptions:

— The crustal thickening developed at the same time in the whole of northern Sardinia some time before 344 Ma ago, that is the isotopic closure of the stromatic migmatite.

— The highest pressure value estimated for the various rocks from the different metamorphic zones correspond to those experienced at this stage.

— All the other temperature and pressure estimates represent transient P-T conditions experienced by the rocks during uplift.

Other assumptions regard the interpretation and geological meaning of the radiometric data which, at least for metamorphic rocks, are at present controversial.

Discussion is limited also by further assumptions:

— Rb/Sr whole-rock isochrons on plutonic rocks from different areas could provide approximate estimations of the emplacement age of the plutonic suites, whereas Rb/Sr whole-rock isochrons on single pluton or orthogneiss are interpreted as emplacement age of the single igneous body.

— Two-point Rb/Sr isochrons (total rock + muscovite or biotite) and Rb/Sr age of muscovite and biotite are assumed to represent cooling age below the "specific" mineral closure temperature of 500°C and 300°C, respectively (Purdy & Jäger, 1976; Cliff et al., 1985; Zeitler, 1989).

— K/Ar mineral ages are assumed to represent cooling age, too: Hbl ~ 500°C, Ms ~ 400°C and Bt ~ 300°C.

The intrepretation of the P-T-t evolutions implies to compare the constructed P-T loops with the current thermal models in which the evolving geotherms of a thickened crust are calculated for a range of well accepted values of thermal parameters (see England & Thompson, 1984). Therefore, in fig. 2, I have reported, together with the P-T shapes, four reference geotherms (Thompson, 1981; Thompson & England, 1984):

V_s: perturbed geotherm due to underthrusting of lithosphere into the mantle;

V_c: perturbed geotherm of thickened crust through continental collision.

V_0: stable geotherm of unperturbed crust;

V_∞: maximally relaxed geotherm for a reasonable post-thickening heat-supply.

6.2. *The initial conditions of burial*

The thermal signature of subduction-like processes is the perturbation of the steady-state geotherm due to the underthrusting of cold lithosphere into the mantle. Considering the Sardinian example, it is important to note that the recorded temperatures at the maximum pressures are higher than the temperature reflecting a perturbated geotherm during subduction (Tab. 1; Fig. 2 and 3). Depending on the degree of confidence we give to the available metamorphic records, two interpretations must be envisaged:

— Due to the thermal relaxation (i.e. increase of the temperature during uplift), the first registrated P-T conditions do not correspond to the maximum conditions. In such a case, the T_{max} does not correspond to the P_{max} of the P-T-t path and the thermal relaxation had started before the P-T conditions recorded by the minerals. Following this idea, we cannot exclude, as recently proposed by Carmignani et

AREA	ROCK UNIT	THICKENING STAGE		RETROGRESSIVE STAGE (S)						
		Highest P conditions	Highest T conditions	P and T conditions		Cooling ages				
						MS (Rb/Sr) 500	Hbl (K/Ar) 500	Miner. Isochron (Rb/Sr) ? 450	Ms (K/Ar) 400	Bt (Rb/Sr&K/Ar) 300
	Eclogite	P 15-17 T 770-825 t ?		~10-12 700-750	5 650 ≤ 344					
North-	Granulite	P 10-12 T 750 t ?			5 650 ≤ 344		296(9)			
eastern	Migmatite	P 10 ? T 700 t ?	7 750 344		4-5 600-650 < 344			298±2		
	Ky ± St Micaschist	P 10 T 580 t ?	8 620							
Sardinia	Grt Micaschist	P 10 T 450 t ?	7.5 500 ?		5 450					
	Augen-gneiss	P 10 T 450 t ?	7.5 500 ?		5 450			306±10	311(2)	294(2)
	Orthogneiss	P 10 T 450 t ?	7.5 500 ?		5 450			308±24		296(4)
Anglona	Micaschist	P 8-10 T 550-650 t ?		2-3 ? 400 ?	2-3 550-600	303(1)				279(2)
Asinara	Micaschist	P 8-10 T 550-650 t ?		? ?	2-3 550-600					
Giuncana Tungoni	Micaschist	P 8-10 T 550-650 t ?			~3 ~400	343(2)				313(2)
Tarra Paddedda	Migmatite	P 7 ? T 700 ? t ?		1-2 ?	650 ?					
	Al-granite emplacement	P T t			1-2 300	303(6)				295(3)
Palau-S.Teresa	Migmatite	P 7 ? T 700 ? t ?			1-2 ? 650 ?					
	Al-granite emplacement	P T t			1-2 >300	299(3)				289(3)

Tab. 1 - Thermobarometric and geochronologial data for northern Sardinia (P = Kbar, T = °C, t = Ma; cooling ages for minerals are average values, between parenthesis number of mesures; mineral abbreviations after Kretz, 1983).

al. (1991) and Cappelli et al. (1991), that continental collision was preceded by lithospheric subduction.

— On the other hand, if we accept that the maximally reached P-T conditions are well recorded by the metamorphic assemblages of eclogites and/or mafic granulites, the crustal thickening episode must be related only to continental collision. According to the present petrological data and to the lack in the Corsica-Sardinia segment of true ophiolitic complexes, an ensialic environment can also be envisaged (see also Ricci & Sabatini, 1978, Memmi et al., 1982; Carmignani et al., 1986).

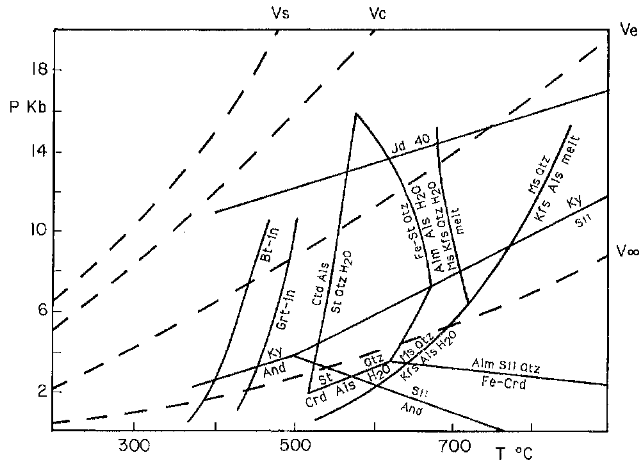

Fig. 2 - Petrogenetic grid for the metamorphism of pelitic rocks (simplified after Yardley, 1989). The curve Ab=Jd+Qtz for XJd=0.40 is also reported. Four reference geotherms V_s, V_o, V_e and V_∞ are also shown as dashed lines (see text for discussion).

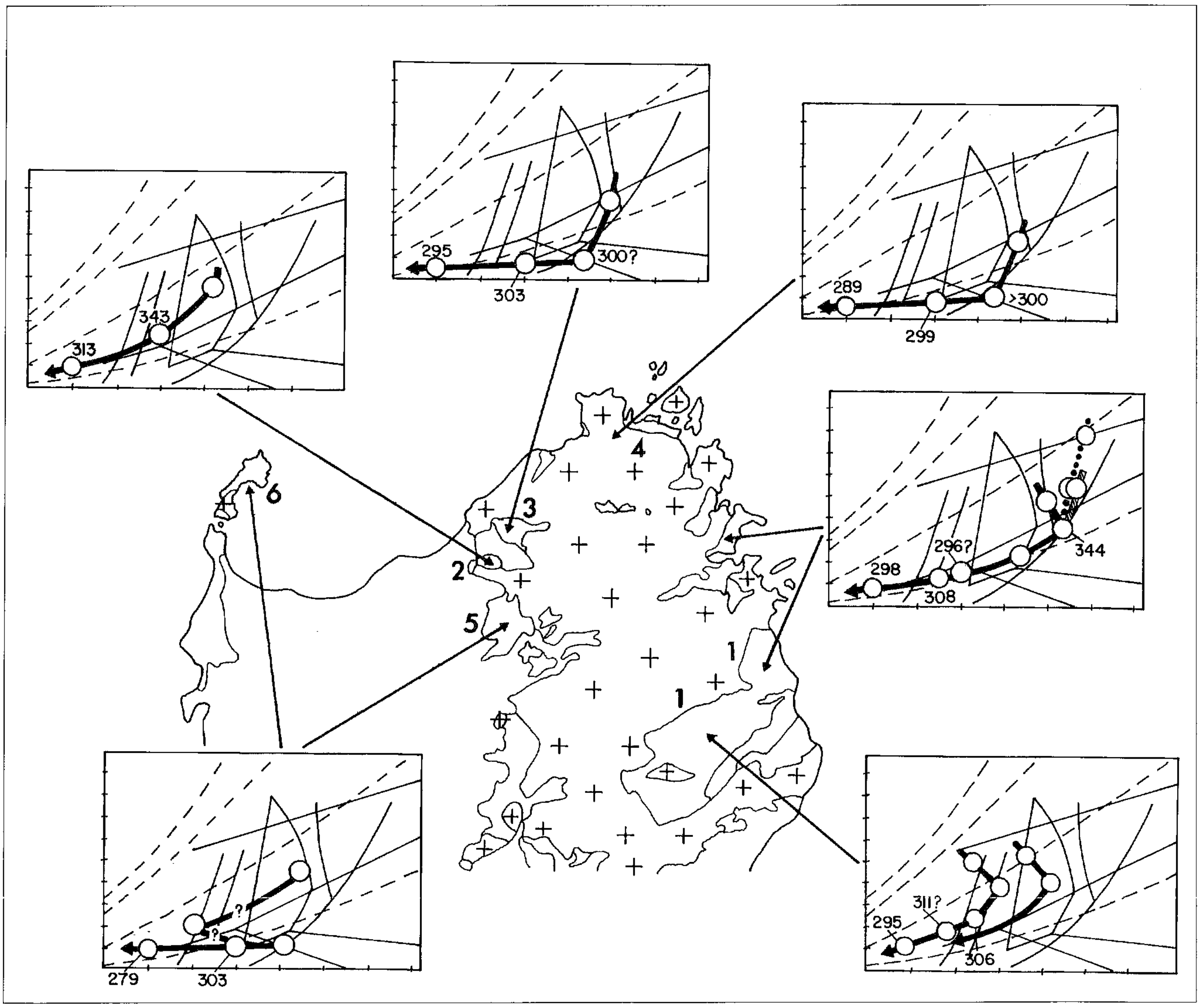

Fig. 3 - Pressure-Temperature-time paths (solid curves with arrows) followed by various portions of the metamorphic basement of northern Sardinia. Voids are petrological and/or radiometric records (radiometric age determinations are indicated with numbers). The dotted line refers to eclogites, the striped line to mafic granulites.

6.3. *The contrasted P- T - t paths*

Whatever is the model we consider for the burial stage, it is also apparent that the different units show different P-T shapes and time evolutions (see P-T-t paths constructed for the different outcrops in fig. 3).

Because the assumptions, listed above, oversimplify the problem, on one hand, they allow to begin a discussion even with such a few data, but, on the other hand, deliberately exclude the possibility to investigate the consequence of the possible non-simultaneous thickening, as well as the effects induced by the lower and upper units (Davy & Gillet, 1986), over a given tectonic unit in a stacked nappe-pile.

However, the contrasted P-T loops (Fig. 3) suggest that different mechanisms controlled the unroofing of the different blocks.

The available data are not sufficient to allow an evaluation of the uplift rate either for the different outcrops or for the whole uplift history within the single outcrop.

I can only obtain a rough idea of the uplift rate of the high-grade areas of northeastern Sardinia, where the bands of a stromatic migmatite closed at 344 Ma, whereas biotite closed at 298 Ma (Ferrara et al., 1978). Assuming a ΔT of 350-400°C between the two closures, and considering that petrological data indicate that the exhumation path did not transgress the geotherm of maximum relaxation V_∞, this portion of the cooling history followed a $\Delta T/\Delta P$ gradient of about 80°C/kbar. Namely ~13 km were removed in that time span, that is an uplift rate of about 0.29 mm yr^{-1}. This value is consistent with a model which assumes that this part of the uplift history was controlled by erosion and isostatic processes (England & Thompson, 1984; Thompson & Ridley, 1987).

The previous uplift history, that is the immediate post-thickening stage, has no radiometric constraints.

Furthermore, no data allow to envisage the mechanism of juxtaposition of rocks recording so different highest pressure conditions such as migmatite (~700°C; ~10 kbar), mafic granulite (~750°C; ~10-12 kbar) and eclogite (~850°C; ~17 kbar). As a working hypothesis, I can assume that these rock units were tectonically juxtaposed through an intracrustal delamination process which followed the crustal thickening stage.

In the medium-grade zones of northeastern Sardinia, there is no possibility of evaluating the uplift rate.

The fairly coincident age values ($\Delta t < 20$ Ma) obtained on different minerals (hornblende, muscovite, biotite) notwithstanding their different "blocking" temperatures led Ferrara et al. (1978) to recognize a sudden temperature fall, due to rapid uplift probably favoured by the emplacement of the main granitoids of the batholith. For the immediate post-thickening stage, the petrological data indicate that an important decompression ($\Delta P \simeq -3$ kbar) was accompanied by a temperature increase of about 30°C-70°C, later followed by a concomitant decrease of both P and T (Franceschelli et al., 1989). This initial part of the P-T path is suggestive of slow uplift.

For the Giuncana and Tungoni areas in western Gallura, no thermobarometric data, useful to reconstruct the initial P-T path, are available. The very precocious radiometric closure of muscovite (and biotite) suggests faster initial cooling rate than in northeastern Sardinia. This might suggest that tectonics, instead of erosion, was the main process assisting uplift.

Therefore, in the two so far considered cases, exhumation totally developed above the V_∞ geotherm. In the other outcrops, I am now considering that the late low-pressure type metamorphic imprint implies the transgression of this geotherm. However, it is worth noting that the actual P-T path is difficult to be reconstructed. Furthermore, the petrological data seem to indicate different histories among different areas.

In Anglona, for instance, the low P-high T metamorphism is described as temperature prograde (Del Moro et al., 1991). This could mean that it developed concomitantly with and consequently to the emplacement, at high crustal level of some early plutonic intrusions. Under this circumstance, the Anglona metamorphites should already have been uplifted to upper crustal condition and cooled below amphibolite facies condition before being reheated at low-pressure condition (Fig. 3).

A different picture arises from the Tarra Padedda (western Gallura) and Palau-S.Teresa di Gallura outcrops. Here, no evidence indicates that these units have been almost completely cooled before high-temperature/low-pressure metamorphism. Instead, the data point to a prolonged residence at relatively high temperature (late closure of muscovite and biotite of granite rocks ~300 and ~290 Ma ago, respectively), even at shallow depth where the rocks underwent ductile deformation and syntectonic emplacement (~300-310 Ma) of peraluminous granitoids (Innocenti et al., 1991; Macera et al., 1989).

The petrological data in this case indicate that the maximally relaxed geotherm (V_∞) was transgressed in the late stage of uplift. Considering also the non-coaxial style of the late deformation, we can reasonably conclude that this stage was controlled by extensional tectonic process. (A process of post-thickening crustal extension has also been proposed by Cappelli, 1991, for the Nappe zone of central Sardinia.)

7. Summary and conclusions

As a matter of fact, the sad main conclusion is that the available data are still too scarce for a satisfactory reconstruction of the evolution of the Hercynian chain in Sardinia.

However, those data allow to propose the following conclusions and working hypotheses.

1) Northern Sardinia exhibits evidence for crustal

thickening which possibly developed through continental collision processes.

2) The close association, in northern Sardinia, of migmatites, mafic granulites eclogites recording different early P-T conditions suggest that crustal thickening was, at least locally, followed by intracrustal delamination giving rise to juxtaposition of lower-crust to mid-crust rocks.

3) The data indicate that after crustal thickening the various portions of the basement of northern Sardinia followed contrasted P-T-t paths. In some cases, the P-T paths developed entirely above the maximally relaxed geotherm for a reasonable post-thickening heat supply. In other cases, the rocks show evidence that this critical geotherm was trangressed. Thus, northern Sardinia probably underwent differential uplift. It is apparent that post-thickening exhumation of the various portions of the medium- to high-grade zones was controlled by different processes such as erosion and isostasy, tectonic unroofing and tectonic extension.

The modelling and understanding of the post-thickening evolution of northern Sardinia could benefit from checking the following working hypotheses by future research :

a) In northeastern Sardinia, P-T-t paths are consistent with an erosion-controlled exhumation.

b) In the Giuncana-Tungoni area, P-T-t is indicative of rapid cooling (and rapid uplift); a tectonic unroofing process may be expected to have assisted the uplift.

c) In the Anglona, Asinara and Palau-S. Teresa di Gallura areas, P-T data suggest that V_{∞} was transgressed. This could be induced either by:

— reheating of a previously cooled and uplifted block by late intrusions (Anglona);

— extensional tectonic processes accomplished by the emplacement of syntectonic, peraluminous granitoids (Palau-S.Teresa di Gallura) at a relatively shallow depth.

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MACROFAUNA AND PALEOENVIRONMENT OF THE LANGHIAN-SERRAVALLIAN DEPOSITS FROM THE CAGLIARI AREA (SOUTH SARDINIA, ITALY)

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ABSTRACT

A series of samples, collected at Fangario and Simbirizzi (NW periphery of Cagliari) and at Is Foradas (Nw of Sestu), with a semiquantitative method, was analyzed from the chemical, physical and biological points of view. This analysis made possible to establish the correlation among the sequences and to attempt the reconstruction of the evolution of the Middle-Miocenic sedimentary basin concerning the Campidano area of Cagliari as well as the reconstruction of those benthic paleocommunities contained in it.

The examined sequences belong to the Fangario Formation.

The sedimentation regarding the stratigraphically lower portion of the Is Foradas section (whose whole section is related to the *Orbulina suturalis* Subzone) and the Cava Giuntelli section (except for level 7 deposits) is continuous and typical of the bathyal zone top portion. Differently, level 7, belonging to the "Arenarie di Pirri" Formation, consists of sandy lithofacies of the *Orbulina universa* Subzone, with a *Pectinidae* and *Ostreidae* macrofauna signalling an infralittoral environment.

The uppermost section of the Is Foradas sequence, instead, consists of repeated alternances of decimetric sandstone banks, fragments of circalittoral malacofauna and marls with macrofauna of bathyal environment.

The turbidity of this part of the sequence is well documented by sedimentological and sedimentographical features of the sandy levels. Their bathymetry, taken from the analysis of the malacofauna, is considerably lower than that regarding the pelitic levels situated above and below. The turbidity is likely to be related to the tectonic instability of the sedimentary basin caused by tectonic stresses that are likely to be related to the volcanic activity occurred in southern Campidano during the Upper Langhian *(Orbulina suturalis* Subzone).

The Epibathyal is attested at Sestu, within the highest levels of Fangario as well as at Simbirizzi, by their paleobiocoenoses consisting of certain species at present marking the beginning of the bathyal zone, such as *Aporrhais (Aporrhais) uttingeriana* (RISSO), *Eudolium fasciatum* (BORSON), *Xenophora infundibulum* (BROCCHI), *Abra (Syndosmya) stricta* (BROCCHI), *Korobkovia denudata* REUSS; *Flabellipecten burdigalensis* (LAMARCK). The Mesobathyal and deeper Bathyal zones haves been recognized in the lower stratigraphic part of the Fangario sequence, where the presence of *Abra (Syndosmya) longicallus* and *Malletia (Malletia) caterinii* dominates.

KEY WORDS: *Paleontology, Paleoecology, Middle Miocene, Sardinia, Italy.*

RIASSUNTO

Una serie di campioni, prelevati con metodo semiquantitativo, a Fangario e Simbirizzi (periferia NW di Cagliari) e a Is Foradas (NW di Sestu), é stata analizzata dal punto di vista chimico, fisico e biologico. Tali analisi hanno consentito di effettuare la correlazione di dette sequenze e di tentare la ricostruzione dell'evoluzione del bacino di sedimentazione meso-miocenico del Campidano di Cagliari e delle paleocomunita' bentoniche in esso contenute.

Le varie sequenze considerate rientrano nella Formazione di Fangario.

La sedimentazione relativa alla porzione stratigraficamente piu' bassa della successione di Is Foradas (la cui intera sezione è riferita alla Subzona a *Orbulina suturalis*) e della sezione di Cava Giuntelli (esclusi i depositi relativi al livello 7) risulta continua e propria della parte alta del piano batiale. Il livello 7 appartenente alla Formazione delle "Arenarie di Pirri", e' costituito invece da litofacies sabbiose della Subzona a *Orbulina universa* con macrofauna a Pectinidi e Ostreidi indicatrice di un ambiente infra-circalitorale.

La parte sommitale della successione di Is Foradas e' invece caratterizzata da ripetute alternanze di banchi decimetrici di arenarie con frammenti di malacofaune circalitorali e marne con macrofaune batiali.

Il carattere torbiditico di questa parte della sequenza e'

documentato dalle caratteristiche sedimentologiche e sedimentografiche dei letti arenacei. La loro batimetria, desunta dall'analisi delle malacofaune, risulta apprezzabilmente minore rispetto a quello dei livelli pelitici soprastanti e sottostanti.

Gli eventi torbiditici vengono posti in relazione con un'instabilita' del bacino di sedimentazione dovuta a stress tettonoci da collegare, presumibilmente, all'attivita' vulcanica esplicatasi nel Campidano meridionale durante il Langhiano superiore (Subzona a *Orbulina suturalis*).

L'Epibatiale e' suggerito a Sestu, nei livelli piu' alti di Fangario ed a Simbirizzi, dalla dominante presenza di una paleobiocenosi costituita da specie che caratterizzano l'inizio del piano batiale quali *Aporrhais (Aporrhais) uttingeriana* (RISSO); *Eudolium fasciatum* (BORSON), *Xenophora infundibulum* (BROCCHI), *Abra (Syndosmya) stricta* (BROCCHI), *Korobkovia denudata* REUSS, *Flabellipecten burdigalensis* (LAMARCK).

Il Mesobatiale ed il Batiale, sono stati riconosciuti nella parte stratigraficamente più bassa della sequenza di Fangario dove è predominante la presenza di *Abra (Syndosmya) longicallus* e *Malletia (Malletia) caterinii.*

PAROLE CHIAVE: *Paleontologia, Paleoecologia, Miocene medio, Sardegna, Italia.*

1. INTRODUCTION

The present work is to be considered as a part of the paleontological and paleoecological research on the Neogene in Sardinia.

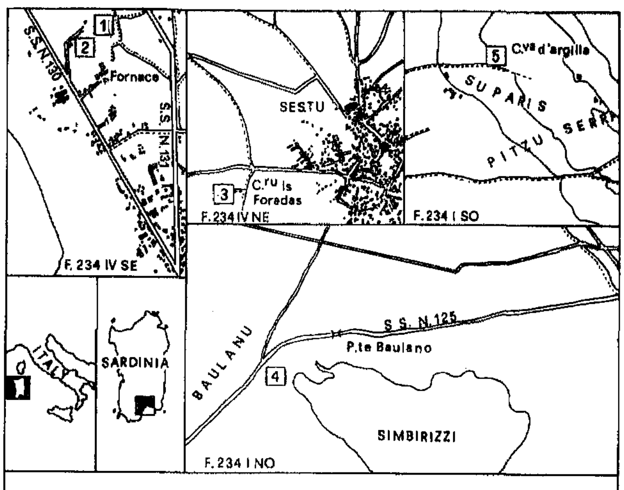
It follows the papers of ROBBA & SPANO (1978) and SPANO (1989; 1990) on the macrofaunas contents in the stratigraphic macrofauna successions of the areas of Fangario (Cava Giuntelli and Cava Cementaria) and of the Cuccuru Paris (Quartu S. Elena).

Previous research on the macrofossils belonging to the Miocene of Fangario had been performed by GENNARI (1867), PARONA (1887) and COMASCHI CARIA (1958). Recently, the microfaunas of Fangario were studied by CHERCHI (1974) (*Foraminifera*), ROBBA & SPANO (1978)(*Foraminifera and Pteropoda*), BARBIERI & D'ONOFRIO (1985) (*Foraminifera*) and BARBIERI et al. (1985) (*Foraminifera* and *Calcareous nannofossils*). CORRADINI et al. (1985) (*Foraminifera, Dinoflagellata* and *Calcareous nannofossils*), instead, have studied the microfossils of the Is Foradas sequence (Sestu).

On the basis of the micropaleontological data provided by these Authors the examined sections are referred on the whole refer to the Middle Langhian - Lower Serravallian time. Also, they would be included between the Zones N 8 and N 11 (probably up to of the Zones N 11/ N 12) of the BLOW scale (1969).

The sequences just examined are those regarding Cava Giuntelli and Cava Cementeria, outcropping in the Fangario area; Cava Is Foradas, nearby Sestu, and Cuccuru Paris and Simbirizzi, in the periphery of Quartu S. Elena (Fig. 1). The paleoecological analysis regards instead the localities of the Giuntelli, Is Foradas and Simbirizzi.

As for the description of sections , see ROBBA & SPANO (op. cit.) and what has been reported in the tables 4-6.



SECTIONS:
1 Section Cava Giuntelli (Fangario - Cagliari). 2 Section Cava Cementeria (Fangario - Cagliari). 3 Section Cava Is Foradas (Sestu). 4 Section Simbirizzi (Quartu S. Elena). 5 Section Cuccuru Paris (Quartu S. Elena).

Fig. 1 - Location of the studied Middle Miocene sections

2. METHODOLOGY

The sampled lithotypes were defined by physical (granulometries) and chemical (value in carbonates) laboratory analysis.

Moreover, the numerical data concerning some of the sedimentological features of the sediment were calculated as follows: the median (MZ); the sorting coefficient (<F128M> σ <F255D>); the asymmetry coefficient (SK1).

The Plankton/Benthos ratio (P/B) was calculated for each sample.

As for Cava Giuntelli, the thanatocoenoses were collected at the same levels as the ones sampled by ROBBA & SPANO (1978) for the Pteropodes and the ones sampled by SPANO (1989; 1990) for the Bivalvia and the Gastropoda respectively.

They bear the same. The chosen levels had been symbols examined before, since they had been the subjects being the object of an accurate visual examination aiming at revealing possible "anomalies" in the relationships among the different fossil constituents and between them

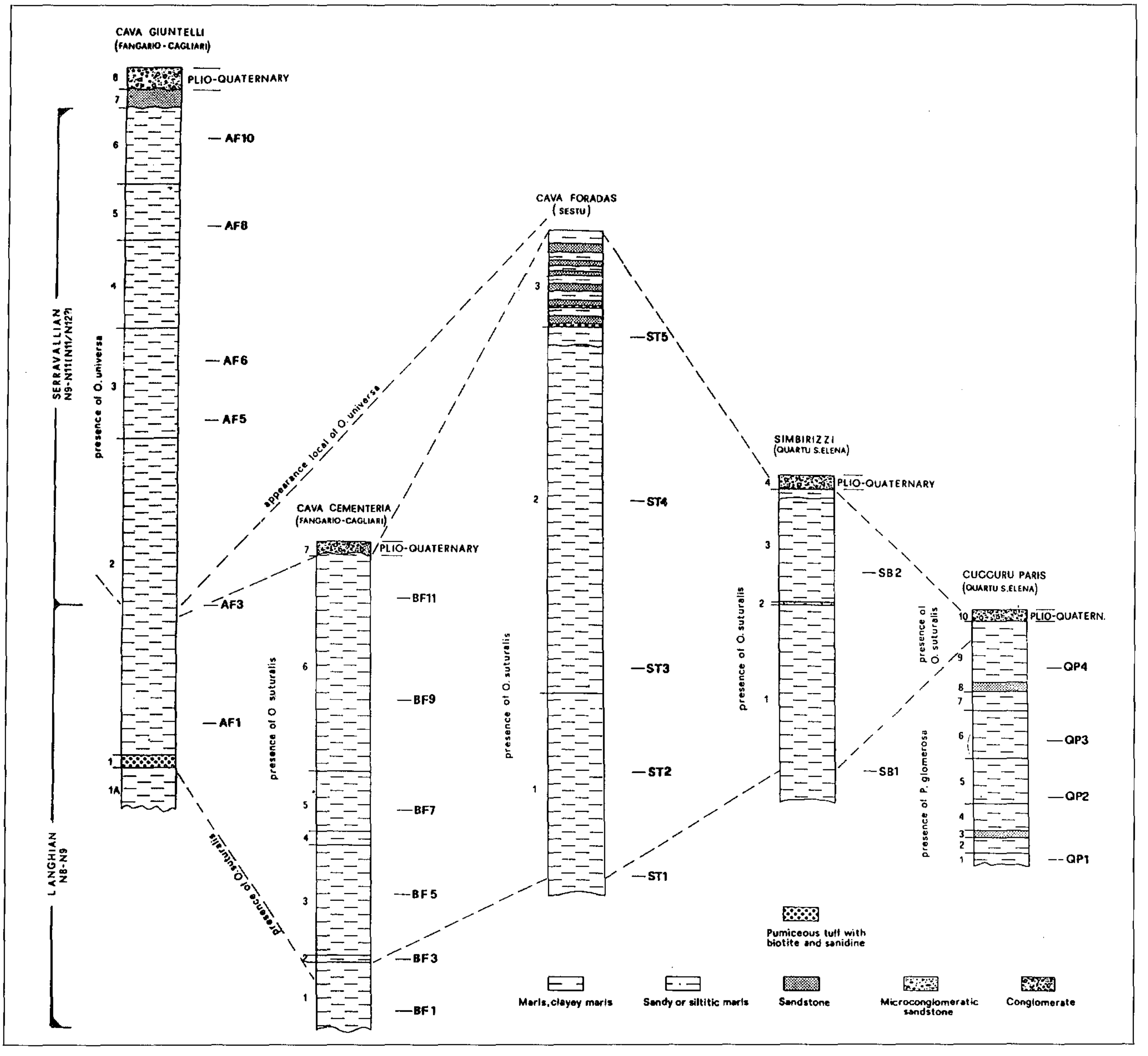


Fig. 2 - Biostratigraphic correlation of studied Middle Miocene sequences

and the rocks incorporated.

As a attempt to prepare a volumetric sample of 30 dmc with H_20_2 did not give any good result because of the extreme fragmentation of the findings, some appriopriate lenses were used in order to highlight the smallest individuals when collecting the findings on the layer heads limited to a surface of 0.90 sq/m corresponding to a rectangle with the longest side of 6 m parallel to the layer surface and the shortest side of 0.15 m.

The following paleoecological analysis, and particularly, the Abundance values (A.), Average Abundance (AM), Dominance (D) and Average Dominance (DM) are all related to the fossiliferous contents gathered in this way.

The method followed suggested was the one by DI GERONIMO & ROBBA (1976) for the definition of the fossiliferous content and for the Abundance values was followed.

The consistency between sediment and fossil communities, the balance between right and left valves in the *Bivalvia*, the presence of intact shells and of still entire delicate valves and the absence of dimensional sorting among the different samples, all confirm the different thanatocoenoses autochthony.

3. The Sediment

The numerical data on the granulometries (Tab. 1), also reported in graphics (Fig. 3-6), ascribe the examined

PARAMETERS	F A N G A R I O						S E S T U					S I MB.	
	AF1	AF3	AF5	AF6	AF8	AF10	St1	St2	St3	St4	St5	SB1	SB2
SAND	4,00	2,00	2,16	2,93	3,26	2,61	12,19	5,13	5,06	9,20	19,81	9,10	10,93
SILT	66,50	68,14	78,22	65,72	77,06	63,54	64,33	70,91	63,09	71,66	74,39	90,64	88,98
CLAY	29,50	29,86	19,62	31,35	19,69	33,85	23,48	23,96	31,85	19,14	5,80	0,26	0,09
MODE	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT	SILT
MEDIAN	6,20	6,60	6,30	6,40	6,90	6,70	6,00	5,50	6,50	5,60	4,10	6,20	5,40
Mz	6,10	6,10	6,15	6,10	6,15	6,35	5,75	5,65	6,05	5,85	4,70	5,20	5,10
$\sigma 1$	2,10	2,00	1,95	2,10	1,85	1,85	2,35	2,25	2,15	2,25	1,90	1,40	1,10
SK1	-0,02	-0,12	-0,06	-0,07	-0,20	-0,09	-0,05	-0,08	-0,10	+0,05	+0,16	-0,36	-0,14

Tab. 1 - Physical and chemical analysis of the sediment

sediments to the shaley silt field and, limited to the samples SB1 and SB2, to the sandy silt field.

Mainly arenaceous levels exist in the Cava Is Foradas and Cuccuru Paris sections. With the silt and pumice levels, they constitute, in Sestu, about 4 meters of alternances. Their deposition, as it will be widely reported in a proper paragraph, is to be related to volcanic-tectonic events.

The carbonates values estimated, on the rock as a whole as well as on the section passing through 62,5 <F128M> μ <F255D>meshes, locate the corresponding lithotypes among the silty-fine or sandy-silty marls (Tab. 4-6).

Similar conditions can be found in all the examined sequences.

The highest percentages of silt refer to levels 3 and 5 of Cava Giuntelli with 78.22 and 77.06; 3-2 of Cava Is Foradas with 74.39 and 71.66; 1 of Simbirizzi with 90.64. The fine fraction is more abundant in levels 6 and 3 of Cava Giuntelli with percentages of 33.85 and 31.35; 2 and 1 of Is Foradas with 31.85 and 23.96. In Simbirizzi this component is meaningless with the highest peak of 0.26 in the level 1. On the contrary, the section related to the latter locality shows the highest values in sand with 90.64% again in the same level 1.

These data show lithologies that are similar to those characterizing the Pliocenic bathyal muds in Western Liguria (ROBBA, 1981) and those presently existing in the Mediterranean Sea (CARPINE, 1970).

4. The Middle Miocene Malacofaunas of the Fangario,Sestu and Simbirizzi Areas

The stocks of the species existing in the different successions examined, coming from "homogeneous surfaces" (Tab. 2,3), do not constitute the totality of the species actually existing in the sampled sequences.

It is necessary to add those species found by SPANO (1989; 1990) with the traditional sampling method, and those preserved within the "Museo Sardo di Geologia e Paleontologia" "D. LOVISATO", which are not found in the "homogeneous samples".

In Cava Giuntelli 39 species are represented to the sample AF1 the following must be added: *Amusium cristatum* (BRONN); *Gastrana fragilis* LINNEO; *Eudolium subfasciatum* (SACCO); *Eudolium subfasciatum unituberculifera* (SACCO); *Amyclina* cf. *oblita* (BELLARDI). To the sample AF3: *Eudolium subfasciatum unituberculifera* (SACCO); *Columbella (Columbella)* cf. *curta* (DUJARDIN). To the sample AF6: *Amyclina* cf. *semistriata dertonensis* (BELLARDI); *Amyclina* cf. *connectens* BELLARDI; *Amyclina* cf. *clathurella* (BELLARDI); To the sample AF8: *Chlamys (Aequipecten) haueri* (MICHELOTTI); *Loripes* sp.; *Trachycardium (Trachycardium)* cf. *multicostatum* (BRONN); *Azorinus (Azorinus) chamasolen* (DA COSTA); *Teredo* sp.; *Cassidaria echinophora* (Linneo); *Eudolium subfasciatum unituberculifera* (SACCO); *Amyclina* cf. *semistriata dertonensis* (BELLARDI); *Amyclina* cf. *connectens*

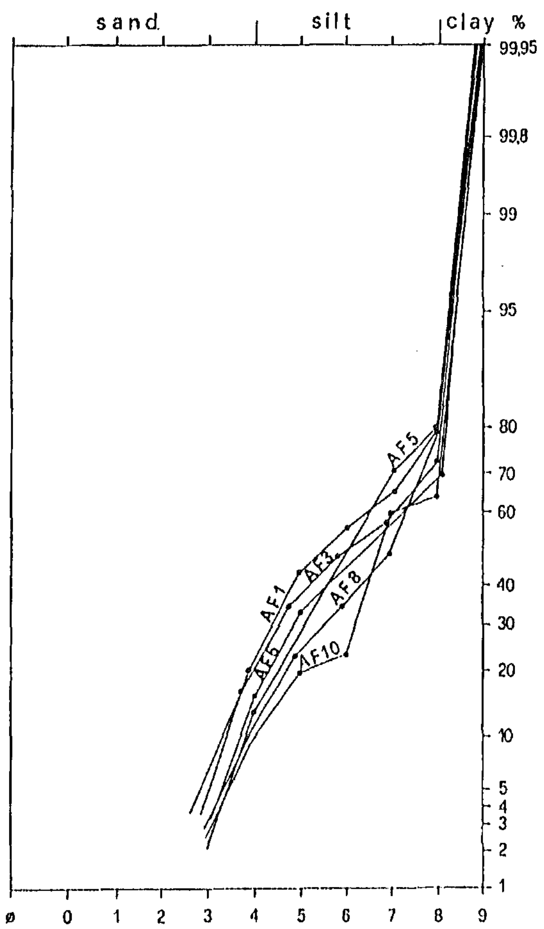


Fig. 3 - Cumulative frequency curves of the Fangario section sediment

BELLARDI. To the sample AF10: *Barbatia (Cucullaearca)* cf. *candida idae* (FUCINI); *Barbatia (Barbata)* cf. *barbata* (LINNEO); *Anadara (Anadara) diluvii* (LAMARCK); *Limopsis (Limopsis)* sp.; *Glycymeris (Glycymeris) insubrica* (BROCCHI); *Atrina pectinata* (LINNEO); *Amusium cristatum* (BRONN); *Chlamys (Aequipecten) haueri* (MICHELOTTI); *Trachycardium (Trachycardium)* cf; *multicostatum* (BRONN); *Lutraria (Lutraria) lutraria* LINNEO; *Lutraria (Lutraria) oblonga* (CHEMNITZ); *Tellina (Arcopagia) cf. corbis* (BRONN); *Tellina (Peronidia)* cf. *albicans* (GMELIN); *Azorinus (Azorinus) chamasolen* (DA COSTA); *Callista (Callista)* cf. *italica* (DEFRANCE); *Callista (Callista) erycinoides* (LAMARCK); *Clausinella cf. basteroti* DESHAYES; *Teredo sp.; Poromya (Poromya) tauromagna perumbonata* (SACCO); *Cuspidaria (Cuspidaria)* cf. *maxima proboscidea* (SISMONDA); *Pleurotomaria* cf. *pedemontana* (SACCO); *Astraea (Ormastralium) carinata* (BORSON); *Scala(Fuscoscala) turtonis alternicostata* (BRONN); *Cirsotrema (Cirsotrema) crassicostata pedemontana* (SACCO); *Cirsotrema (Stenorhytis)* cf. *proglobosa* (SACCO); *Polinices (Polinices)* cf. *redemptus* (MICHELOTTI); *Cassidaria* sp.; *Semicassis* *(Echinophoria)* cf. *rondeletti taurinensis* (SACCO); *Charonia (Charonia) nodifera* (LAMARCK); *Eudolium subfasciatum unituberculifera* (SACCO); *Columbella (Macrurella)* cf. *elongata* (BELLARDI); *Amyclina* cf. *semistriata dertonensis* (BELLARDI); *Amyclina* cf. *oblita* (BELLARDI).

13 species were identified at Cava Is Foradas; the following must be added to the sample St1: *Amusium cristatum* (BRONN); *Chlamys (Aequipecten) haueri* (MICHELOTTI); *Lucina (Lucina) michelottii* (MAYER); *Lucina (Lucina) tumida* (MICHELOTTI); *Trachycardium (Trachycardium)* cf. *multicostatum* (BRONN); *Azorinus (Azorinus) chamasolen* (DA COSTA); *Cassidaria echinophora* (LINNEO); *Eudolium subfasciatum unituberculifera* (SACCO); *Amyclina* cf. *semistriata* dertonensis (BELLARDI). To the sample St1: *Azorinus (Azorinus) chamasolen* (DE COSTA); *Eudolium*

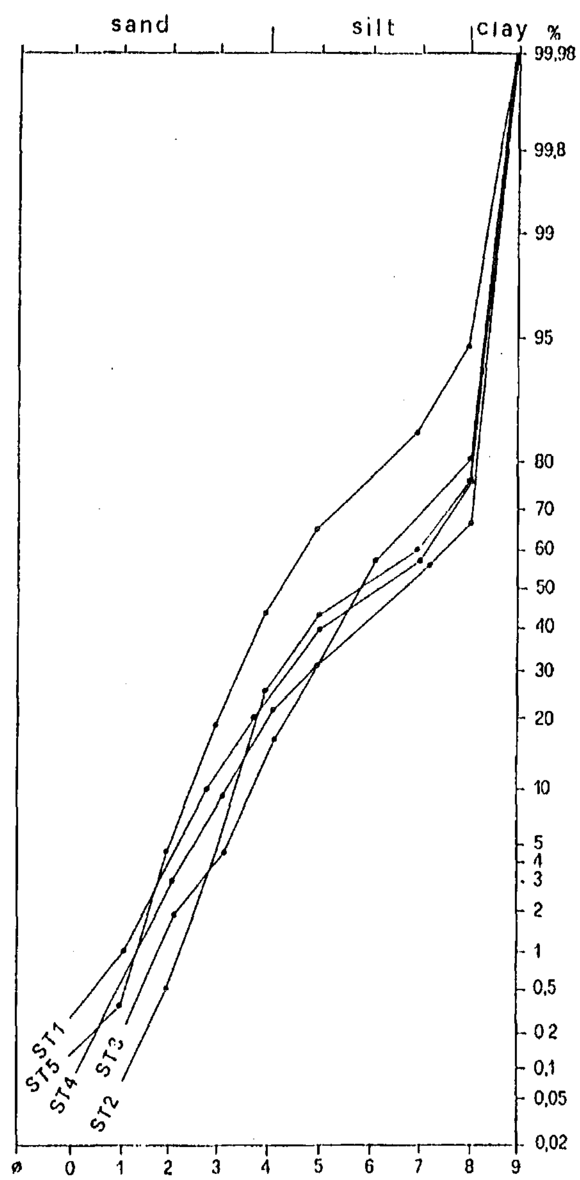


Fig. 4 - Cumulative frequency curves of Sestu section sediment

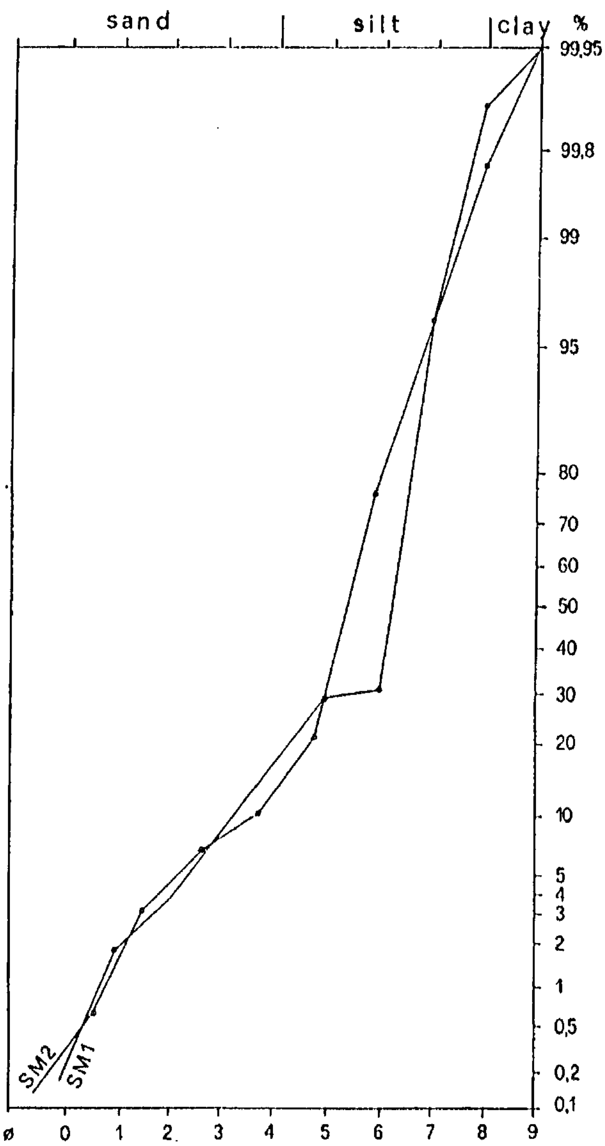


Fig. 5 - Cumulative frequency curves of the Simbirizzi section sediment

subsfasciatum unituberculifera (SACCO). To the sample St3: *Azorinus (Azorinus) chamasolen* (DA COSTA); *Bittium (Bittium)* cf. *exiguum* (MONTEROSATO); *Amyclina* cf *semistriata dertonensis* (BELLARDI). To the sample St4: *Chlamys (Aequipecten) haueri* (MICHELOTTI); *Lucina (Lucina) michelottii* (MAYER); *Trachycardium (Trachicardium)* cf. *multicostatum* (BRONN); *Azorinus (Azorinus) chamasolen* (DA COSTA); *Bittium (Bittium)* cf. *exiguum* (MONTEROSATO); *Amyclina* cf. *oblita* (BELLARDI). To the sample St5: *Chlamys (Aequipecten) haueri* (MICHELOTTI); *Lucina (Lucina) michelottii* (MAYER); *Trachycardium (Trachycardium)* cf. *multicostatum* (BRONN); *Phaxas (Phaxas)* cf. *pellucida* (PENNANT).

The "homogeneous surfaces" sampling of layer hesds show instead a stock of 45 new species for the section. The following 17 species were found in the Fangario area: *Nucula (Nucula) jeffereysi* BELLARDI; *Lucina (Lucina)* cf. *miocenica* MICHELOTTI; *Thyasira (Thyasira) flexuosa* (MONTAGU); *Laevicardium (Laevicardium) norvegicum* (SPENGLER); *Gastrana lacunosa* (CHEMNITZ); *Donax* sp.; *Glossus (Meiocardia) moltikianoides* (BELLARDI); *Saxicava aff. rugosa* (LINNEO); *Semicassis laevigata* (DEFRANCE); *Mitra (Tiara) bronni* MICHELOTTI; *Vexillum (Uromitra) recticosta* BELLARDI; *Lyria sp.; Genota (Genota)* cf. *ramosa* (BASTEROT); *Eburna* sp.; *Oniscidia* cf. *cythara (costa)*; *Drillia matheroni* BELLARDI; *Clavatula semimarginata* LAMARCK. For the Sestu area: *Nucula (Nucula) nitida* SOWERBY; *Arca sp.; Plagiocardium hirsutum* (BRONN); *Abra (Syndosmya) stricta* (BROCCHI); *Solecurtus sp.; Semicassis (Semicassis) laevigata* (DEFRANCE); *Conus* sp..

All of the 21 species found in Simbirizzi have been mentioned here for the first time.

The semiquantitative analysis distinguished 58 species in Cava Giuntelli as well as in Cava Is Foradas. The silty Middle - Miocene marns of the Fangario area show the existence of 97 species while 71 are those referring to the Sestu area.

5. Autoecological Data About the Extant Species

From the semiquantitative analysis (bulk samples) and from those performed by SPANO (1989; 1990) concerning Cava Giuntelli and Cava Is Foradas, a large group of species (36) still living in the Mediterranean Sea has been identified. Autoecological data inferred from the literatature are reported hereafter. The Average Abundance (Am) and Average Dominance (Dm) relative values

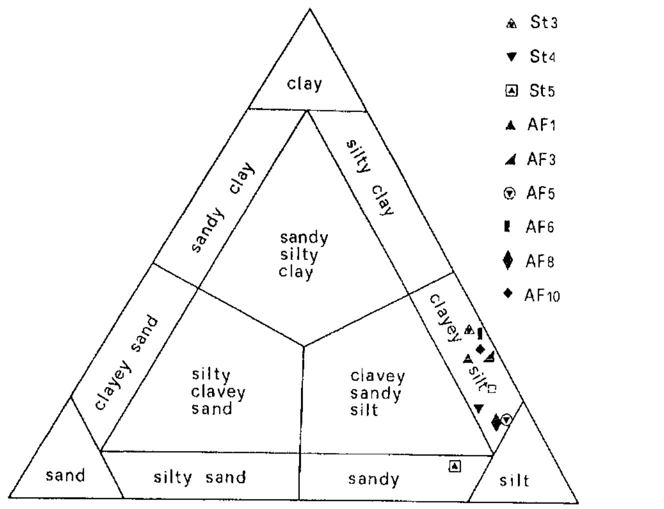


Fig. 6 - GORSLINE diagram: definition of the granulometrical intervals belonging to the analyzed samples

TAXA	FANGARIO						A_m	D_m
	AF1	AF3	AF5	AF6	AF8	AF10		
	cf A D	cf A D	cf A D	cd A D	cf A D	cd A D		
Nucula (Nucula) nucleus (LINNEO)			lib 1 2.78				0.17	0.46
Nucula (Nucula) nitida SOWERBY	2 vd 2 4.26						0.33	0.71
Nucula (Lamellicula) jeffreysi BELLARDI						1vs 1 2.04	0.17	0.34
Malletia (Malletia) caterinii (APPELIUS)			1vs 1 2.78		2vs 2 3.57	2sv 3 6.12 2vd	1.00	1.06
Tindaria (Tindaria) arata BELLARDI	1 fr 1 2.13						0.17	0.35
Nuculana (Nuculana) höernesi (BELLARDI)						1vd 1 2.04	0.17	0.34
Nuculana (Jupiteria) brocchii BELLARDI					1 gc 1 1.79		0.17	0.30
Nuculana (Lembulus) pella (LINNEO)						1vd 1 2.04	0.17	0.34
Yoldia (Yoldia) longa BELLARDI					1vd 2 3.57 1vs		0.33	0.60
Yoldia (Yoldia) nitida (BROCCHI)						1vd 1 2.04	0.17	0.34
Solemya (Solemya) doderleini (MAYER)	2vs 3 6.38 1gi		3gi 4 11.11 1gc .		1gc 1 1.79	2gi 2 4.08	1.66	3.21
Bathyarca pectuncoloides (SCACCHI)					1vs 1 1.79		0.17	0.30
Loripes sp.				1vd 1 3.33			0.17	0.55
Pinna cf. ventilabrum ROVERETO		1fr 1 6.25					0.17	1.04
Korobkovia denudata REUSS		5ip 6 37.50 1vd	2m 2 5.55	1vd 2 6.66 1gi	2ip 2 3.57	1ip 2 4.08 1vd	2.33	9.56
Flabellipecten burdigalensis (LAMARCK)	5m 6 12.77 2vd	1vs 2 12.50 1ip	1vd 2 5.55 1vs	2vd 4 13.33 3ip		2fr 2 4.08	2.66	8.04
Chlamys spinulosa (MUNSTER)	1vd 1 2.13			1fr 1 3.33		2fr 1 2.04	0.50	1.25
Ostreinella negleta (MICHELOTTI)						1vd 1 2.04	0.17	0.34
Lucina (Lucina) cf. miocenica MICHELOTTI			1vd 1 2.78				0.17	0.46
Lucinoma borealis (LINNEO)			1vd 2 12.50 2fr			1vs 2 3.57 1gi	0.67	16.07
Linga (Linga) columbella (LAMARCK)					2vs 2 3.57	1vd 1 2.04	0.50	3.61
Megaxinus (Megaxinus) cf bellardianus (MAYER)						1vs 2 4.08 1fr	0.33	1.28

Tab. 2 - Qualitative table of the thanatocoenosis at Fangario locality

concerning the semi-quantitative samples are also quoted (Tab. 4- 6). The frequency data already determined are next to the species reported only by SPANO (op. cit.).

Nucula (Nucula) nucleus (LINNEO) (Fangario: Am= 0.17; Dm= 0.46). It is a mistophilous species, preferential with respect to the biocoenoses DC-DE-DL. It is meaningful both in the "*Abra alba* and *Corbula gibba*" and in the "*Syndosmya community*" biocoenoses (in DI GERONIMO & COSTA, 1980).

Nucula (Nucula) nitida SOWERBY (Fangario Am= 0.33; Dm= 0.71. Sestu: Am= 0.60; Dm= 1.66). Epibion moving a few centimeters under the sea floor surface, saprovore. It lives in the Circalittoral and in the Bathyal.

Nuculana (Lembulus) pella (LINNEO) (Fangario: Am= 0.17; Dm= 0.34). This species indicates the mobile bottoms instability (PAVIA et al., 1989). It is circalittoral-bathyal and mainly lives within the deep sublittoral zones.

Bathyarca pectuncoloides (SCACCHI) (Fangario: Am= 0.17; Dm= 0.30). BUCCHERI et al., (1987) define it as being circalittoral and bathyal. DI GERONIMO & LI GIOI (1980) relate it to the set of eurybathic species shortly distributed.

Anadara (Anadara) diluvii (LAMARCK) (Fangario - AF10:R). Seminfaunal, sospensivore, mistophilous and preferential of the VTC. It is often found in the DC and, above all, in the DE.

Glycymeris (Glycymeris) insubrica (BROCCHI) (Fangario- AF10: R). Psammophilous, sospensivore and epibion. It is preferential of the SGCF; it is often found in the SFBC.

Atrina pectinata (LINNEO) (Fangario - AF10: R). It is a seminfaunal species sticking to sandy and muddy beds or in hard substratum cracks. It is widespread in the Infralittoral and in the Circalittoral. According to NORDSIECK (1969) it can go up to 600 m in depth.

Lucina (Lucina) orbicularis (DESHAYES) (Simbirizzi: Am= 1.50; Dm= 3.33). It is tolerant psammophilous and mobile epibion.

Linga (Linga) columbella (LAMARCK)(Fangario: Am= 0.50; Dm= 3.61. Sestu: Am= 1.20; Dm= 2.68): In Sardinia it is found in pelitic facies of the Circalittoral and the Bathyal.

Linga (Linga) columbella (LAMARCK) (Fangario: Am= 0.50; Dm= 3.61. Sestu: Am= 1.20; Dm= 2.68). It is

Megaximus (Megaximus) transversus (BRONN)				1vd 1 3.33			0.17	0.55
Gonimyrtea meneghinii (DE STEFANI & PANTANELLI)	2vd 2 4.26			1vs 2 6.66 1vd			0.66	1.82
Gonimyrtea meneghinii crassolamellata SACCO	1fr 1 2.13				1fr 1 1.79	1vd 1 2.04 1fr	0.50	0.65
Thyasira (Thyasira) flexuosa (MONTAGU)					1vs 4 7.14 3vd	3vd 4 8.16 2fr	1.33	2.55
Diplodonta (Diplodonta) rotundata (MONTAGU)						1vs 1 2.04	0.17	0.34
Laevicardium (Laevicardium) norvegicum (SPENGLER)					1vs 1 1.79	1vs 1 2.04	0.33	0.64
Donacilla cornea (POLI)						1vd 1 2.04	0.17	0.34
Ervilia castanea (MONTAGU)						1vd 1 2.04	0.17	0.34
Telline (Arcopagia) crassa PENNANT	1vd 2 4.26 1gi		3gi 4 11.11 1vs	1vd 1 3.33			1.17	3.12
Tellina (Moerella) donacina LINNEO	2vd 3 6.38 2vs				1ip 1 1.79		0.66	1.36
Gastrana lacunosa (CHEMNITZ)	1vd 1 2.13						0.17	0.35
Donax (Cuneus) sp.	1vd 2 4.26 1gi				1vd 2 3.57 1vs		0.66	1.30
Abra (Syndosmya) longicallus (SCACCHI)	5vs 7 14.89 6vd		4vs 6 16.67 4vd	6vd 6 20.00	8vd 11 19.64 6vd	7vd 9 18.37 4vd	6.50	14.93
Abra (Syndosmya) stricta (BROCCHI)	2vd 2 4.26	1vd 2 12.50 1vs			1gc	2vd 3 6.12 1vs	1.17	3.81
Abra prismatica (MONTAGU)				1vd 1 3.33	1vs 1 1.79		0.33	0.85
Glossus (Meiocardia) moltkianoides (BELLARDI)					2vd 2 3.57		0.33	0.59
Dosinia (Pectunculus) exoleta (LINNEO)	1vd 2 4.26 1fr			1gi 1 3.33			0.50	1.26
Corbula (Caryocorbula) carinata DUJARDIN					vd 1 1.79		0.17	0.30
Corbula (Varicorbula) gibba (OLIVI)	1vd 1 2.13		2vd 3 8.33 1vs		1vs 1 1.79		0.66	2.04
Saxicava aff. rugosa (LINNEO)					1vd 2 3.57 1ip	1vd	0.33	0.60
S BIVALVIA	36 76.63	13 81.25	24 66.66	20 66.63	40 71.45	39 79.57		
Dentalium badense PARICH			1gi 1 3.33				0.17	0.55

Tab. 2 - (continued)

found in the outer shelf muds.

Megaxinus (Megaxinus) transversus (BRONN) (Fangario: Am= 0.17; Dm= 0.55. Sestu: Am= 0.20; Dm= 0.77). It is mistophilous, seminfaunal, sospensivore, essentially circalittoral.

Lucinoma borealis (LINNEO) (Fangario: Am= 0.67; Dm= 16.07. Sestu: Am= 0.40; Dm= 0.57. Simbirizzi: Am= 1.50; Dm=3.33). It is unanimously recognized as preferential of the PE and indicative of the mobile beds instability.

Thyasira (Thyasira) flexuosa (MONTAGU) (Sestu: Am:= 0.40; Dm=.57) (Simbirizzi: Am= 0.50; Dm= 1.11). According to GIACOBBE & LEONARDI (1988) It is a narrow vasicola species and preferential of the VTC.

Diplodonta (Diplodonta) rotundata (MONTAGU) (Fangario: Am= 0.17; Dm= 0.34. Sestu: Am= 020; Dm= 0.45). Narrow pelophilous and infaunal. it is found in the Infralittoral and the Circalittoral.

Donacilla cornea (POLI) (Fangario: Am= 0.17; Dm= 0.34). PAVIA et al. (1989) undoubtedly relate it to the SFBC preferentials.

Ervilia castanea (MONTAGU) (Fangario: Am= 0.17; Dm= 0.34). It exists within the tortonian deposits of Capo S. Marco ascribed by SPANO (1988) to the DC.

Lutraria (Lutraria) lutraria (LINNEO) (Fangario-AF10: R). This sospensivore species is indicated by the Authors as being preferential of the DC.

Lutraria (Lutraria) oblonga (CHEMNITZ) (Fangario-AF10: R). It is sospensivora and preferential of the DC.

Phaxas (Phaxas) pellucida (PENNANT) (Sestu-St5: F). This species is considered to be typical of the SFBC and of the VTC. It is peculiar of the "Abra alba and Corbula gibba associations" and of the "Syndosmya community" (in DI GERONIMO & COSTA, 1980).

Tellina (Arcopagia) crassa (PENNANT) (Fangario: Am= 1.17; Dm= 3. 12. Sestu: Am= 1.40; Dm=2.64). It is frequently found in the SGCF biocoenosis.

Tellina (Moerella) donacina (LINNEO) (Fangario: Am= 0,66; Dm= 1.36. Sestu: Am= 5.20; Dm= 12.13). This species is detritivore, mobile endobion and mistophilous. It is believed to belong to the stock of species that are exclusive or preferential of the DC.

Tellina (Peronidia) albicans (GMELIN) (Fangario-

AF10: FF and Sestu-St1: R; St5: R). Its autoecological features are not well known.

Gastrana fragilis (LINNEO) (Fangario- AF10: R). It is prevailingly an internal sublittoral species.

Gastrana lacunosa (CHEMNITZ) (Fangario: Am= 0.17; Dm= 0.35. Sestu: Am= 1.20; Dm= 3.08). This species has been extinct in the Mediterranean since the Pleistocene (MARASTI & RAFFI, 1980). It presently lives within the Atlantic ocean.

Abra (Syndosmya) alba (WOOD) (Sestu: Am=0.40; Dm= 1.22). It is believed to be a LRE species (PAVIA et al., 1989). In fact, it is for the SFBC and, above all, for the DC and the DE.

Abra (Syndosmya) longicallus (SCACCHI) (Fangario: Am= 6.50; Dm=14.93. Sestu: Am=4.80; Dm=12.34. Simbirizzi: Am=1.50; Dm= 3.33). It is detritivore, mobile endobion and lelofila. It has a deep circalittoral and bathyal distribution. This species is exclusive of the VP.

Abra (Syndosmya) prismatica (MONTAGU) (Fangario: Am=0.33; Dm=0.85. Sestu: Am=0.20; Dm=0.29). It is saprovore and mobile endobion. Most of the Authors believe it to be preferential of the DC; it exists in the DE.

Azorinus (Azorinus) chamasolen (DA COSTA) (Fangario-AF8:R; Af10:FF and Sestu-St1:R; St2:R; St3:R; St:R). This species is mistophilous and preferential of the DC and DE.

Venus (Ventricoloidea) multilamella (SACCO) (Sestu: Am=0.20; Dm=0.29). PAVIA (1975) believes it to be eurybathic and typical of the Sublittoral muddy facies. CALDARA et al. (1989) believe it to be preferential of the DC-DE. DI GERONIMO (1984) relates it to the DE and strictly to the VTC. BUCCHERI et al. (1987) include it in a stock of species that are preferential of the biocoenoses VTC-VP.

Dosinia (Asa) lupinus (LINNEO) (Fangario: Am=0.50; Dm=1.26). The present species is strictly sabulicole (GIACOBBE & LEONARDI. 1987) and sublittoral. It is quoted in biocoenoses of the inner (SFBC) and outer (DE) shelf. It is also considered to be preferential of the PE.

Corbula (Varicorbula) gibba (OLIVI) (Fangario: Am=0.66; Dm=2.04. Sestu: Am=3.20; Dm=7.51). It characterizes the PE biocoenoses. It is infaunal, sessile, sospensivore and eurybathic. In the Atlantic ocean it is part of the Abra alba and *"Syndosmya community"*. In the Adriatic Sea it is part of similar associations (VATOVA, 1949-in DI GERONIMO & COSTA, 1980).

Hiatella (Hiatella) artica (LINNEO) (Sestu: Am=0.20; 0.77). It is a species of large ecological diffusion species.

Cassidaria echinophora (LINNEO) (Fangario-AF6:R; AF8:R and Sestu-St1:F). It belongs to the ecocline VTC-

Pleurotomaria sp.					1gi 1 1.79		0.17	0.30
Xenophora testifera (BRONN)		3fr 3 18.75				3fr 3 6.12	1.00	4.14
Semicassis sp.				1gi 1 3.33	1fr 1 1.79	2gi 2 4.08	0.67	1.53
Eudolium fasciatum (BORSON)			4gc 6 16.67 4gi	4gi 4 13.33	1gi 4 7.14 3fr	3gi 3 6.12	2.83	7.21
Ficus (Ficus) conditus BRONGNIARI	5fr 5 10.64		1gc 1 2.78 2gi	1gc 3 10.00	1gc 2 3.57 1ip		1.83	6.50
Ficus (Ficus) geometra (BORSON)	5gi 5 0.64	1gi			1gi 1 1.79		1.00	2.07
Ficus sp.					3gi 3 5.36		0.50	0.90
Murex cf. spinicosta BRONN						1ip 1 2.04	0.17	0.34
Genea sp.						1gi 1 2.04	0.17	0.34
Mitra (Tiara) bronni MICHELOTTI	1gi 1 2.13						0.17	0.35
Vexillium (Uromitra) recticosta BELLARDI				1gi 1 3.33			0.17	0.55
Lyria sp.					1gi 1 1.79		0.17	0.30
Eburna sp.				3gi 3 8.33			0.50	1.40
Drilla matheroni BELLARDI					2gc 3 5.36 1ip		0.50	0.90
Clavatula semimarginata LAMARCK				1gc 1 2.78			0.17	0.46
S GASTROPODA	11 23.41	3 18.75	12 30.56	10 33.32	16 28.59	10 20.40		
TOTALS	47 100.00	16 100	36	30	56 100	49		
AVERAGE ABUNDANCE FOR SAMPLES*	2.6	2.7	2.6	2.0	2.1	2.5		
NUMBER OF SPECIES	18	6	14	15	27	25		

Tab. 2 - (continued)

*Calculated on the Mollusca. cf=content Fossiliferous; A=Abundance;D=Dominance;Am=Average abundance; Dm=Average Dominance;vd=right valve;Vs=left valve;gc=complete shell; ip=externalmold;m = internal mold;fr=Fragment;

TAXA	SESTU					SIMBIRIZZI		SESTU		SIMBIRIZZI	
	St1	St2	St3	St4	St5	SB1	SB2	Am	Dm	Am	Dm
	cf A D	Cf A D	cf A D	cf A D	cf A D	cf A D	cf A D				
Nucula (Nucula) nitida SOWERBY		2vd 2 4.44			1vs 1 3.85			0.60	1.66		
Tindaria (Tindaria) arata BELLARDI	2vd 3 1.29 1vs							0.60	0.86		
Nuculana (Nuculana) höernesii (BELLARDI)				1vd 1 2.27				0.20	0.45		
Nuculana (Jupitarial) brocchi BELLARDI	1vs 1 1.43			2vs 3 6.82 1vd	1m 1 3.85			1.00	2.42		
Solemya (Solemya) doderleini (Mayer)		2gi 2 4.44				1gc 2 4.44 1fr		0.10	0.33	1.00	2.22
Arca sp.		1vd 1 2.22						0.20	0.44		
Loripes lacteus LINNEO		1vd 1 2.22						0.20	0.44		
Pinna sp.						1fr 1 2.22	2fr 2 66.67			1.50	1.11
Korotkovia denudata REUSS	6ip 3 11.62 3vd							1.60	2.28		
Flabellipecten burdigalensis (LAMARCK)	2ip 2 2.36	5fr 5 11.11		1ip 4 9.09 3gi	3fr 2 7.69			2.60	6.15		
Chiamys spinulosa (MUNSTER)	1vd 2 2.86			2fr 1 3.35				0.80	1.80		
Ostrenella negleta (MICHELOTTI)			1gi 1 3.85		2vd 3 11.54 1g1	3vs 4 8.39 2vd		0.30	3.08	2.00	4.44
Lucina (Lucina) orbicularis DESHAYES						4vd 6 13.33 2gc				3.00	6.66
Lucina (Lucina) rollei (MICHELOTTI)						2gc 3 6.66 1vs				1.50	3.33
Lucina sp.	1ip 1 1.63					6fr 3 6.66 6fr		0.20	0.29	1.50	3.33
Lucinoma borealis (LINNEO)	1vs 2 2.86 1vd					2vs 3 6.66		0.40	0.57	1.50	3.33
Linga (Linga) columbella (LAMARK)	1vs 2 2.86 1vd	1vs 2 4.44 1vd		1vs 1 2.27	1vd 1 3.35	1vd		1.20	2.68		
Megaxinus (Megaxinus transversus (BRONN)					1vd 1 3.85			0.20	0.77		
Gonymyrtea meneghinii (DE STEFANI & PANTANELLI)	1vd 1 1.43							0.20	0.29		
Gonymirtea meneghinii crassolamellata SACCO		1vd 1 2.22 1fr	2fr 2 7.69 1vs	1vd 2 6.56 2fr	1vs 1 3.85	2fr 2 4.44		1.20	3.66	1.00	2.22
Thyasira (Thyasira) flexuosa (MONTAGU)	2vd 2 2.86					1vd 1 2.22		0.40	0.57	0.50	1.11
Diplodonta (Diplodonta) rotundata (MONTAGU)				1vs 1 2.27				3.30	0.45		
Parvicardium hirsuta (BRONN)					1vs 1 3.85			0.20	0.77		
Laevicardium (Laevicardium) norvegicum (SPENGLER)	1vd 1 1.43			4vs 6 13.64 3vd	1vd 1 3.85			1.60	3.78		
Cardium sp.					1m 1 3.85			0.20	0.77		
Lasaenia inaequilateralis (COSSMAN)			2vc 3 11.54 1vs					0.60	2.31		
Lianocardium cf. bollenensis (MAYER)				1vs 1 2.27				0.20	0.45		
Spisula (Memimactra?) astensis (SACCO)						1vd 2 6.66 1vs				1.00	2.22
Tellina (Arcopagia) crassa PENNANT	1ip 3 4.29 2vd	2vs 3 6.56 1gi		1vd 1 2.27				1.40	2.64		
Tellina (Moerella) donacina LINNEO	3vd 7 10.00 5vs	7vs 8 17.78 2vd	2vd 3 11.54 1sv	5vs 6 13.66 1vd	1vd 2 7.69 1ip			5.20	12.13		
Macoma (Macoma) elliptica (BROCCHI)		1vd 1 2.22						0.20	0.44		
Macoma (Macoma) sp.	1vc 1 1.43							0.20	0.29		
Gastrana lacunosa (CHEMNITZ)			3vs 4 15.38 1vc	1vs 2 6.56 1fr				1.20	3.08		
Tellina sp.	1fr 1 1.43							0.20	0.91		
Abra (Syndosaya) alba (WOOD)				1vs 1 2.27	1vs 1 3.85			0.40	1.22		
Abra (Syndosaya) longicallus (SACCHI)	3vs 6 8.56 4vd	4vd 7 15.55 5vs	3vs 4 15.38	2m 3 6.82 1vs	3vs 4 95.38 1vd	2vd 3 6.66 1fr		4.80	12.34	1.50	3.33
Abra (Syndosaya) stricta (BROCCHI)						2gc 3 6.66 1gi				1.50	3.33
Abra prismatica (MONTAGU)	1vs 1 1.43							0.20	0.29		
Solecurtus sp.	1vd 1 1.43	1vd 1 2.22	2vd 2 7.69	1vd 2 2.27				0.80	2.72		
Venus (Ventricoloidea) multilamellata (LAMARCK)	1vd 1 1.43							0.20	0.29		
Corbula (Caryocorbula) carinata DUJARDIN	1gc 1 1.43							0.20	0.29		

Tab. 3 - Qualitative table of the thanatocoenosis at Sestu and Simbirizzi localites

VP (PAVIA et al. 1989).

Murex spinicosta BRONN (Fangario: Am=0,17; Dm=0.34). PAVIA et al. (1989) include it in a group of preferential species of the VTC.

Charonia (Charonia) nodifera (LAMARCK) (Fangario-AF10:FF). It is preferential of the VTC.

Corbula (Varicorbula) gibba (OLIVI)	2vs 4 5.70 3vd	2vs 3 6.66 2vd	2vs 3 11.51 2vd	1vs 6 13.64				3.20	7.51		
Miactella (Hiactella) arctica (LINNEO)					1gi 1 3.85	1		0.20	0.77		
Saxicava aff. rugosa (LINNEO)			1vd 1 3.85					0.20	0.77		
Cuspidaria (Cuspidaria) miocenica PARONA				1vd 1 2.27				0.20	0.45		
Cuspidaria sp.	1vd 1 1.43							0.20	0.29		
Clavagella (Clavagella) brocchii LAMARCK					1ip 1 3.85	1		0.20	0.77		
S BIVALVIA	52 74.29	37 84.40	23 88.66	41 93.16	23 88.50	33 73.28	2 66.67				
Xenophora infungibulum (BROCCHI)	1gi 1 1.43							0.20	0.29		
Aporrhais (Aporrhais) uttingeriana (RISSO)						1m 1 2.22		0.50	1.11		
Cypraea sp.		1gi 1 2.22						0.20	0.44		
Semicassis (Semicassis) laevigata (DEFRANCE)	2gi 3 4.29 1ip						2fr 1 3.33	0.60	0.36	0.50	16.66
Eudolius fasciatum (BORSON)			1gi 1 3.35			2gi 3 6.66 2fr		0.20	0.77	1.50	3.33
Ficus (Ficus) conditus BRONGNIART	9gi 11 15.70 2ip	4gi 4 8.89		1gc 2 6.56 1fr	1gc 2 7.69 1fr	3fr 3 6.66		3.80	7.36	1.50	3.33
Ficus (Ficus) geometra (BORSON)	9gi 1 1.63	1gi 1 2.22	1gi 2 7.69			2gi 2 6.66		0.80	2.27	1.00	2.22
Genota (Genota) cf. ramosa (BASTEROT)	1gi 1 1.43							0.20	0.29		
Oniscidia cf. cythara (COSTA)	1ip 1 1.63							0.20	0.29		
Drillia matheroni BELLARDI					1ip 1 3.85			0.20	0.77		
Cunus sp.						1fr 1 2.22				0.50	1.11
Pharus legumen LINNEO		2gi 2 4.44		11i 1 2.27		1gi 2 4.44 1ip		0.60	1.34	1.00	2.22
S GASTROPODA	18 25.71	8 17.77	3 11.54	3 6.81	3 11.54	12 26.64	1 33.33				
TOTALS	70 100.00	45	26 100.00	44	26 100.00	45	3 100.00				
AVERAGE ABUNDANCE FOR SAMPLE*	2.5	2.6	2.4	2.4	1.4	2.5	1.5				
NUMBER OF SPECIES	28	17	11	18	18	18	2				

Tab. 3 - (continued). *Calculated on the total Mollusca. cf= Fossiliferous content ; A= Abundance; D=Dominance;Am=Average abundance;vd=right valve;vs=left valve;gc=complete valve;gi=incomplete shell;ip=external mold;m=internal mold;fr=Fragment.

6. Paleoecology of some Extinct Species

Nucula (Nucula) jeffereysi BELLARDI (Fangario: Am=1.17; Dm= 0.34). It has frequently been quoted for the Circalittoral.

Malletia (Malletia) caterinii (APPELIUS) (Fangario:Am=1.00; Dm=1.06). It is considered to be preferential of the VP (PAVIA et al., 1989). According to MARASTI & RAFFI (1976) it is a well- known species also within the circalittoral muds. It was saprovore, purely pelophilous, mobile endobion immediately under the sea bottom surface of the sea (ROBBA, 1981).

Tindaria (Tindaria) arata BELLARDI (Fangario: Am=0.17; Dm=0.35. Sestu: Am=0.60; D=0.86). The species belonging to the same genus are the most preferential of the deep muds. The T. solida is exclusive of the VP.

Nuculana (Nuculana) höernesi (BELLARDI) (Fangario: Am=0.17; Dm=0.30. Sestu: Am=0.20; Dm=0.45). This species is considered to be preferential of the VTC. (CARPINE (1970) includes it within a stock of circalittoral-epibathyal species, or eurybathic poorly distributed. PAVIA et al. (1989) include it among the VTC preferential species.

Yoldia (Yoldia) longa BELLARDI (Fangario: Am=0.33; Dm=0.60). According to PAVIA et al. (1989) it is often found in Pliocene and Pleistocene pelitic circalittoral ZONES. The same Authors believe it to be probably typical of the VTC.

Yoldia (Yoldia) nitida (BROCCHI) (Fangario: Am=0.17; Dm=1.04). It probably was seminfaunal and sticking to dandy and muddy bottoms or in hard substrata cracks. It is found in infra and circalittoral deposits.

Amusium cristatum (BRONN) (Fangario-AF5:F; AF10:R and Sestu- St1:R). The Authors believe that this species belongs to the VTC- VP. In Sardinia it is always found in pelitics of the external Circalittoral or Bathyal.

Korobkovia denudata REUSS (Fangario:Am=2.33; Dm=9.56. Sestu: Am=1.60; Dm=2.28). SPANO (research in progress) found it in outer shelf facies and in the Sardinia epibathyal facies.

Flabellipecten burdigalensis (LAMARCK) (Fangario:Am=2.66; Dm=8.04. Sestu: Am=2.60; Dm=6.15). The same as the K. denudata RESS.

Chlamys spinulosa (MUNSTER) (Fangario: Am=0.50; Dm= 1.25. Sestu: Am=0.80; Dm=1.80). The same as K. denudata REUSS.

Ostreinella negleta (MICHELOTTI) (Fangario: Am=0.17; Dm=0.34). Sestu: Am=0.80; Dm=3.08. Simbirizzi: Am=2.00; Dm=4.44). It is often found within pelitic circalittoral and bathyal deposits.

Megaximus (Megaximus) bellardianus (MAYER) (Fangario: Am=0.33; Dm=1.28). PAVIA et al. (1989) propose it in a stock of LRE and SSPR species. It is diffused in platform muddy facies.

Glossus (Meicardia) moltikianoides (BELLARDI) (Fangario: Am=0.33; Dm=0.59). It does not have any particular paleontological location. It is frequently found in pelitic circalittoral and bathyal facies.

Gonimyrtea meneghini (DE STEFANI & PANTANELLI) (Sestu: Am=0.20; Dm=0.29). PAVIA et al., (1989) find that this a species preferential of the Circalittoral. Its ecological needs are not sufficiently known.

Gonimyrtea meneghini crassolamellata SACCO (Sestu: Am=1.20; Dm=3.66. Simbirizzi: Am=1.00; Dm=2.22). The same as the above species.

Aporrhais (Aporrhais) uttingeriana (RISSO) (Simbirizzi: Am=0.50; Dm=1.11). It is a LRE species, especially widespread in the Circalittoral.

Xenophora infundibulum (BROCCHI) (Fangario: Am=1.00; Dm=4.14. Simbirizzi: Am= 0.50; Dm=1.11). In Sardinia it is always found in the Circalittoral-Bathyal (SPANO, study in progress). The analogous species X. crispa (KOENIG) is considered to be preferential of the DC and DE biocoenoses by DI GERONIMO & COSTA (1980).

Xenophora testigera (BRONN) (Fangario-AF10:FF). The same as Y. (Yoldia) longa BELLARDI.

Eudolium fasciatum (BORSON) (Fangario: A=2.83; Dm=7.21. Sestu: Am=0.20; Dm=0.77. Simbirizzi: Am=1.50; Dm=3.33). The some as X. infundibulum (BROCCHI).

Eudolium subfasciatum (SACCO) and unituberculifera subspecie (SACCO) (Fangario-AF1:R; AF8:R; AF10:FF and Sestu- St1:R; St2:R). The same as E. fasciatum (BORSON).

Ficus (Ficus) conditus BRONGNIART (Fangario: Am=1.83; Dm=4.50. Sestu: Am=3.80; Dm=7.36. Simbirizzi: Am=1.50; Dm=3.33). The same as Xenophora infundibulum (BROCCHI).

Ficus (Ficus) geometra (BORSON) (Fangario:Am=1.00; Dm=2.07. Sestu:Am=0.80; Dm=2.27. Simbirizzi:Am=1.00; Dm=2.22). The same as Xenophora infundibulum (BROCCHI).

Amyclina semistriata dertonensis (BELLARDI) (Fangario-AF6:R; AF8:F; Af10:F and Sestu-St1:F; St3:F). It is typically preferential of the VTC (BUCCHERI et al., 1987).

Fusinus (Fusinus) longiroster (BROCCHI) (Fangario-AF6:R). It is quoted in literature chiefly for the Bathyal.

Athleta (Athleta) ficulina (LAMARCK) (Fangario-AF10:FF). In Sardinia it exists in outer shelf deposits and the beginning of the bathyal.

Gemmula (Gemmula) rotata (BROCCHI) (Fangario-AF10:R) It can be found in pelitic circalittoral facies of the Pliocene and Lower Pleistocene. According to PAVIA et al. (1989) it could be considered as preferential of the VTC.

The abbreviations used in this work iare from PERES & PICARD (1964): SFBC = Biocoenosis of Well Calibrated Fine Sands; DC = Biocoenosis of Coastal Detritic; DE = Biocoenosis of Muddy Detritic; DL = Biocoenosis of Large Detrital; CB = Biocoenosis of White Corals; VP = Biocoenosis of Bathyal Muds; PE = Biocoenosis of heterogeneous peoplings (of instable soft bottoms); VTC = Biocoenosis of Terrigenous Coastal Muds; SGCF = Biocoenosis of Coarse Sands and Fine Gravels submitted to Bottom currents; SSPR = Biocoenosis without precise meaning; LRE = Biocoenosis with a Large Ecological Distribution; R=rare; F= frequent; FF= very frequent.

7. Thanatocoenoses Autochthony Degree

The thanatocoenoses of each sample show a considerable degree of autochthony. A first analysis of the fossiliferous content, carried out immediately after it was collected, permitted to exclude some of the findings showing allochthony signs. Therefore, only the stock of autoctonous individuals was affected by the statistic elaboration.

The latter phase allows to make some reservations about the autochthony of certain species such as: *Glycymeris insubrica, Tellina crassa, Donacilla cornea* among the extant species; *Pinna ventilabrum* among the extinct ones. Given the autoecological and/or paleoautoecological data collected by the Author, their biocoenotic meaning is believed to be inconsistent with the one regarding most of the other species. Particularly, *Glycymeris insubrica* and *Donacilla cornea* presently belong to the SFBC biocoenoses; *Tellina crassa* belongs to the SGCF biocoenosis. Their presence within the investigated samples is likely to be related to gravitational slidings, especially documented, in fact, in the Cava Is Foradas area. The mostly represented biocoenotic categories, according to their number of species and individuals, in order are Bathyal CB-VP; deep Circalittoral DL-DE-DC and VTC. The latter biocoenosis, as known, is independent from the bathymetric plane. The belonging of many extant species to the stock coming from the Fangario Mollusk collection preserved in the "D. LOVISATO" Museum, suggested that an analysis also based upon the trophism of the same species would be, inappropriate. On the contrary, a deeper investigation of the relationships between substratum and fossil communities, took place. As shown in the biostratigraphic correlations reported in Fig. 1, the stratigraphically lower samples than the whole Langhian- Lower Serravallian sequence of the Cagliari surrounding areas, are those belonging to the sequence exposed in Cuccuru Paris (QP1; QP2; QP3; QP4). As already shown somewhere else in the same work, they have not been the subject of a paleoecological analysis the same as the samples collected in the Cava Cementeria area.

The samples AF1 and Af3 of Cava Giuntelli; St1, St2, St3, St4, St5 of Cava Is Foradas; SB1 and SB2 of Simbirizzi can be essentially related to same chronostratigraphic interval.

AF5, AF6, AF8 and AF10 of Cava Giuntelli show the upper part of the same series.

7.1 *The Cava Giuntelli Sequence (Fangario Area)*

It begins with a layer (1a) of compact silty-clayey marls, with a limited thickness, followed by a layer (1) of pomiceous tuff with biotite and sanidine, and another one (2) also consisting of compact silty-clayey marls, to which the samples AF1 and AF3 belong (Fig. 2).

The thanatocoenoses included consist, in Abundance and Dominance order (Tab. 4), of *Abra longicallus* (clearly prevailing), *Flabellipecten burdigalensis, Ficus geometra, Solemya doderleini* and *Tellina donacina.*

This stock of species is integrated with other species (Tab.2) that are part of the deep or "pure bathyal" ones (sensu CARPINE, 1970). The value of the ratio between the number of planktonic and benthonic *Foraminifera is* 6.5.

The maximum depth of the sedimentary basin is ascribed to these values (500-1000 m, corresponding to the Middle-Bathyal). It takes place in the levels that provided the richest Pteropodes assemblages (ROBBA & SPANO, 1978). This information is also consistent with the one estimated by BARBIERI et al., (1985) on a sampled coming from the same section and, presumably, they share the same stratigraphic height.

Then the sequence evolves towards silty-clayey marls and fossil communities with a less deep bathymetric meaning.

Abra longicallus is the species characterizing the samples AF5 and AF6 of layer 3, while *Endolium fasciatum* acquires more importance. White Corals exist in considerable quantity. The ratio P/B decreases to 4.0.

The malacofauna most epibathyal connotation, together with the data already quoted, makes their belonging to the Bathyal upper horizon possible (about 300-600 m).

The sample AF8 shows similar features. The same role as the one assumed in the previously examined samples by *Flabellipecten burdigalensis, Solemya doderleini* and Tellinidae, is now assumed by *Thyasira flexuosa* and *Drillia matheroni.* The P/B ratio (3.5), although it showing a continuous decreasing trend, is not too far from the one between AF5 and AF6.

The reduction in sedimentation depth continues with the sample AF10 where *Malletia caterinii, Xenophora infundibulum* and *Abra stricta* are found within the stock of dominant species. Their different biocoenotic (DL-DE-DC) and bathymetric meanings show paleocommunities at present living within the Bathyal- Circalittoral transition zones, at a depth of around 300 meters.

7.2 *The Cava is Foradas Sequence (Sestu Area)*

The samples St1 and St2 (level 1) present a fossil association characterized by *Tellina donacina, Abra longicallus, Korobkovia denudata, Corbula gibba, Gastrana lacunosa* and *Flabellipecten burdigalensis* (Tab. 3). The lithology is given by silty-clayey marls; The P/B ratio is 3.7-3.9. It is important here the noteworthy presence (A= 4 and 3; D = 5.70 and 6.60) of *Corbula gibba* peculiar of the

LEVELS	SAMPLES	LITHOLOGY	* % CARBO-NATES	** % CARBO-NATES	THANATOCOENOSIS	P/B+	BATHYMETRIC ZONES	ESTIMATED PALEOBATHYMETRY (m)
7		Sandstone						
	AF10	Silty-clayey marls	30.00	22.75	Abra (S.) longicallus, Molletia M. caterinii, Thyasira (T.) flexuosa, Eudolium fasciatum, Xenophora infundibulum, Abra (S.) stricta.	3.4	EPIBATHYAL	300-400
6								
5	AF8	Silty-clayey marls	40.25	31.75	Abra (S.), longicallus, Eudolium fasciatum, Thyasira flexuosa, Drillia matheroni.	3.5	EPIBATHYAL	300-600
4		Silty-clayey marls						
3	AF6	Silty-clayey marls	38.25	32.75	Abra (S.) longicallus, Eudolium fasciatum, Flabellipecten burdigalensis, Tellina (A.) crassa, Solemya (S.) doderleini, Coralli bianchi.	4.0	EPIBATHYAL	300-600
	AF5		33.25	27.75				
2	AF3	Compact silty-clayey marls	32.25	30.25	Abra (S.) longicallus, Flabelli-pecten burdigalensis, Ficus (F.) geometra, Solemya (S.) doderleni, Tellina (M.) donacina	6.5	LOWER-MIDDLE BATHYAL	600-1000
	AF1							
1		Feldspar ash-tuff 0.30 cm thick	38.25	36.00				

Tab. 4 - Physical, chemical and biological charateristics of the samples from theFangario locality
*Carbonate content of total rock;**Carbonate content of sediment corser than 62?5u;+Planktonic and Benthonic Foraminiferal the ratio

PE and linked to the sedimentary basin instability (Fig. 2).

The lower bathymetry assigned to these two samples (500-600 m), compared with the one assigned to the Cava Giuntelli samples AF1 and AF3 to wich they can be correlated , is likely to be explained with the higher closeness of the Is Foradas sequence to the sedimentation basin edge.

St3 and St4 can be also correlated to AF1 and A3 of Cava Giuntelli. They show a thanatocoenosis which is close to the St1 and St2 one from which they differ for the presence of *Laevicardium norvegicum* and for the numerical reduction of *Abra longicallus*.

The P/B ratio drops to the lowest values of the sequence (1.3 and 1.5). For the samples St3 and St4, therefore, a paleobathymetry between 300 and 500 m is estimated.

As to sample St5, the presence of Circalittoral-Bathyal transition species becomes greater. The best represented biocoenotic categories, in fact, are the DL,DE and DC. Among the dominant taxa, there is *Ostreinella negleta*. A water column thickness of about 300-400 m is assumed.

7.3 *The Simbirizzi Sequence (Quartu S. E. Area)*

The sediments referring to levels 1 and 5, from which the samples SB1 and SB2 are derived, are more sandier than the Cava Giuntelli levels where the samples AF1 and AF3 were collected (Fig. 2).

In SB1, the stock of dominant species consists of White Corals, *Lucina orbicularis, Ostreinella negleta, Ficus conditus, Eudolium fasciatum, Abra longicallus* and *Lucina rollei*. These species are preferential of the Circalittoral and Bathyal. This thanatocoenosis can be believed to be analogous to the present CB associations whose bathymetric zone is at about 300 m; the P/B ratio is about 3.3.

Despite the low number of species and individuals, also the sample SB2, suggests the same paleoenvironmental considerations as for SB1.

8. Sedimentary Basin Stability and Instability

The Cava Is Foradas (level 3) and Cava Giuntelli (level 1) sequences show pomiceous tuff with biotite and sanidine. Given their paragenesis and biochronological location, they are analogous to those recognized by PECORINI (1975) in the Simbirizzi area (level 2 in the present work) and in many other parts of southern Sardinia.

Cava Is Foradas, particularly, shows deposits of a pyroclastic material alternated with sandstones, slightly graded, clearly reworked fossils that are "infra- circalittoral" in environment, siltites and clayey marls of bathyal deposition.

The recurrent presence of feldspar beds, together with their mild thickness, allows to establish a relationship with the coeval "minor interflows" as shown by PECORINI (op. cit.).

The biostratigraphic information about the Fangario and Sestu sequences, taken from CHERCHI (1974), BARBIERI & D'ONOFRIO (1985), BARBIERI et al.

LEVELS	SAMPLES	LITHOLOGY	% CARBONATES *	% CARBONATES **	THANATOCOENOSIS	P/B+	BATHYMETRIC ZONES	ESTIMATED PALEOBATHYMETRY (m)
3	St5	Compact silty-sandy marls with several well cementated intercalations of 20 cm. in thickness	23.25	20.75	Korobkovia denudata, Abra (S.) longicallus, Ostreinella negleta, Tellina (M.) donacina	1.9	EPIBATHYAL	300-400
2	St4	Silty-clayey marls	35.75	30.25	Laevicardium norvegicum, Tellina (M.) donacina, Gastrana lacunosa, Koro-bkovia denudata, Corbula (V.) gibba, Abra (S.) longicallus	1.5	EPIBATHYAL	300-500
	St3		39.25	33.75		1.3		
1	St2	Silty-clayey marls	35.75	33.75	Tellina M.) donacina, Abra (S.) longicallus, Korobkovia denudata, Corbula (V.) gibba, Gastrana lacunosa, Flabellipecten burdigalensis		EPIBATHYAL	500-600
	St1		38.25	35.25		3.7		

Tab. 5 - Physical, chemical and biological charateristics of the samples from the Sestu locality.

LEVELS	SAMPLES	LITHOLOGY	* % CARBO-NATES	** % CARBO-NATES	THANATOCOENOSIS	P/b+	BATHYMETRIC ZONES	ESTIMATED PALEOBATHY METRY (m)
5	SB2	Finely laminated silty-sandy marls	28.25	25.75	Pinna sp Semicassis (S.) laevigata, White corals	3.3	EPIBATHYAL	300-500
4		Well cementated silty-sandy marls						
3		Silty-sandy marls with interbedded arenaceous conglomeratic thin level						
2		Silty-sandy marls						
1	SB1	Silty-sandy marls	28.25	25.75	White Corals, Lucina (L.) orbicularis, Ostrenella negleta, Ficus (F.) conditus, Eudolium fasciatum, Abra (S.) longicallus, Lucina (L.) rollei.	3.3	EPIBATHYAL	300-500

Tab. 6 - Physical, chemical and biological charateristics of samples from SIMBIRIZZI locality

(1985) and CORRADINI et al. (1985), together with the stratigraphic relation between these sequences and affected by the products recognized by PECORINI (op. cit.) and ROBBA & SPANO (1978), confirm that the alternances under discussion can be related to the Langhian Age (*Orbulina suturalis* Subzone).

The sandy and pyroclastic units are here interpreted as the product of the instability of the sedimentary basin substratum that is related to the latest volcanic-tectonic phases of the calc-alkaline volcanism. The pelitic and terrigenous sediments should be related to periods of relative tectonic stability.

Furthermore, also the thanatocoenoses seem to confirm the instability moments thanks to the recurring presence of species typical of heterogeneous associations (PE) such as *Corbula gibba, Lucinoma borealis, Nuculana pella* and *Dosinia lupinus*.

The sedimentological, sedimentographic and biostratigraphic observation already made, would lead to the exclusion of a genesis linked to possible eustatic shifts because of litofacies of a "sublittoral" in environment. Moreover, such an origin would not be consistent with the phase of the advancement of the sea level suggested by Haq et al. (1987) for if the chronostratigraphic interval to which the examined sequences are referred is taken into consideration.

9. Conclusion

The paleoecological study of the fossil communities provided the basic data for the definition of the sedimentary basin evolutive trend, through the recognition of their paleobiocoenotic and paleobathymetric meaning.

The stratigraphically lowest layers, in Dominance order, consist of *Abra longicallus, Flabellipecten burdigalensis, Ficus geometra, Solemya doderleini* associations.

The average layers consists of *Abra longicallus, Eudolium fasciatum, Flabellipecten burdigalensis, Tellina crassa* and *White Coral paleobiocoenoses*.

The highest layers consists of *Abra longicallus, Malletia caterini, Thyasira flexuosa, Eudolium fasciatum, Xenophora infundibulum* and *Abra stricta.*

The analysis of the sedimentological features of the incorporating lithotypes (Tab. 1 and Fig. 3-6) and the paleobiocoenotic and paleobathymetric meaning of the shown tanatocoenoses, suggest the existence of some sedimentation environments evolving from the lower-middle bathyal (AF1 and AF3 of Cava Giuntelli) to the epibathyal (AF8 and AF10 of Cava Giuntelli; St1-St5 of Cava Is Foradas; Sb1 and Sb2 of Simbirizzi).

A regressive trend is therefore documented within the sedimentary basin. In Sardinia, it ends at the Serravallian turn (LEONE et al., 1991) and it is concomitant with the final "Arenarie di Pirri" deposition.

This regressive evolution is recurrently influenced by volcanic-tectonic events that are related to the latest phases of the calc-alkaline, oligo-miocenic volcanism. The tectonic instability could be responsible for those recurrent, although limited, torbiditic episodes (included in the *Orbulina suturalis* Subzone) that are now documented for the first time in the present work on Cava Is Foradas.

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REE DISTRIBUTION IN THE LATE HERCYNIAN DYKES FROM SARDINIA

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ABSTRACT

The late Hercynian dyke rock magmatism of Sardinia develops in two main stages, with different compositional and tectono-magmatic features, one preceding and the other following Permian volcanism.

The first stage -which is present all throughout the island- is represented by orogenic-type products ranging in composition from basaltic andesites to rhyolites and peraluminous rhyolites. These were erupted during two phases: one Carboniferous, and the other Permian in age. In the successive, late Permian, stage, basaltic products are dominant.

These are mostly transitional, from mildly sodic to mildly potassic, and are associated with subalkaline tholeiitic basalts in the central-southern areas and with basic alkaline rocks, in the vicinity of shear zones in the northern part of the island. Scarce peraluminous rhyolites are also found in the late Permian stage.Such a varied geographical distribution of basaltic types seems to be linked to the different crustal thinning of the northern and central-southern parts of the island.

The REE patterns and the values of the initial Sr isotopic ratio make it possible to differentiate the contamination phenomena present in the magma source zones from those accompanying the ascent, and hence the evolution, of the melts.

In the calcalkaline orogenic suite two groups of rocks with different values of ΣREE and initial Sr isotopic ratio have been identified; these can be traced back to the difference in composition of the source zone.

Furthermore geochemical evidence has enabled us to define more precisely the primitive nature of a few basaltic andesites, and to rule out any lack of correlation between calcalkaline types and subalkaline tholeiitic basalts.

The peraluminous rhyolitic dykes show an evident anatectic character connected with crustal melting of source rocks referable partly to pelitic-arenaceous types and partly to much less radiogenic protoliths, a feature which reveals a certain affinity with the rocks of the calcalkaline sequence.

The late-Permian basaltic dyke magmatism, successive to the Permian ignimbritic volcanism of Sardinia, can be entirely attributed to mantle sources.

Low values and small variations of initial Sr isotopic ratio in the alkaline types, correlated with REE distribution, may be referred to primary differences in an undepleted, astenospheric source. Subalkaline tholeiitic basalts reveal mantle metasomatism, probably caused by fluids of crustal origin. Also for the group of transitional basalts, contamination phenomena seem to be connected with a source metasomatism.

KEY WORDS: *Sardinia, dykes, magmatism, Hercynic, REE.*

1.INTRODUCTION

The late Hercynian dyke magmatism of Sardinia is represented by many petrographic types with varying petrogenetic and tectono-magmatic meaning (Traversa, 1968; 1969; Atzori & Traversa,1986; Pasquali, 1990; Pensi, 1990; Vaccaro, 1990; Vaccaro et al.,1991; Traversa et al.,1991).

Taking into account mineralogical and chemical characters, Atzori and Traversa (1986) have distinguished the following main groups of dikes:

1-Calcalkaline types (mainly basaltic andesites and andesites with rare dacites and rhyolites);

2-Peraluminous rhyolites;

3-Basic alkaline types (from alkali basalts to hawaiitic and mugearitic rocks);

4-Transitional basalts;

5-Tholeiitic basalts.

Petrographic features of these rocks are in Atzori and Traversa (1986). With regard to basic rocks, we should specify that alkaline dykes may partly be referred to alkaline lamprophyres (AL) of Rock (1987) and the camptonitic types of Yoder (1979), while the transitional types are closer to the calcalkaline lamprophyres (CAL) of Rock (1987) and the spessartitic types of Yoder (1979).

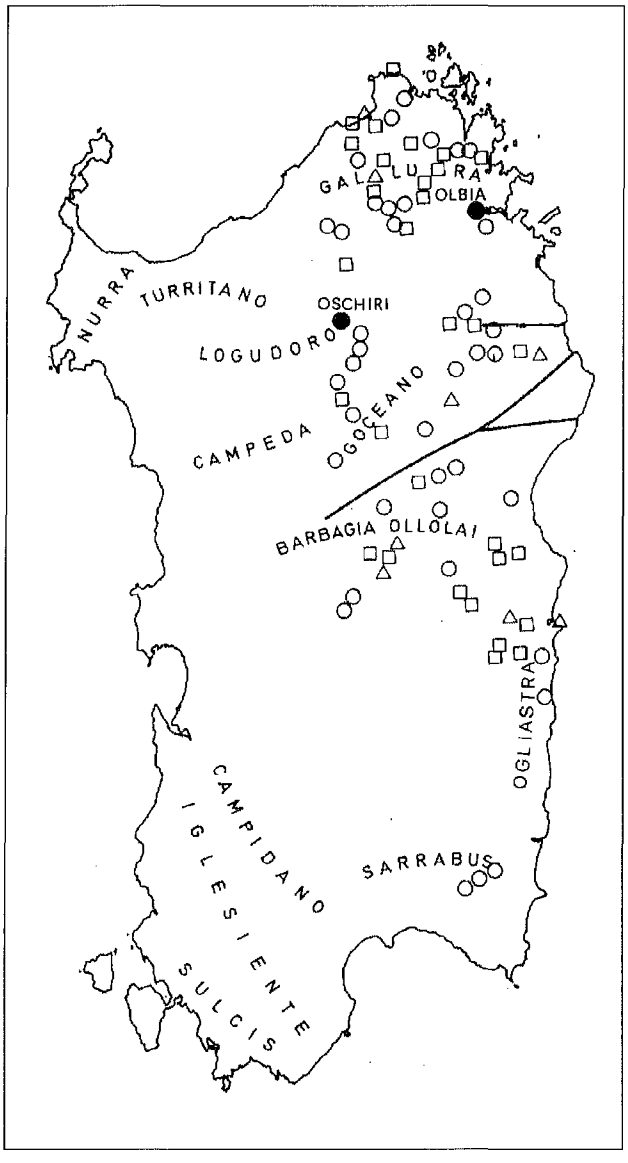

Fig. 1 - Distribution of Sardinian dyke outcrops studied in this paper (Atzori & Traversa, 1986). Circles=single outcrops; triangles=2 to 3 outcrops; square=4 to 6 outcrops; solid lines=important tectonic lines.

The Sardinian dykes (Fig.1) are unevenly distributed and intrude almost all the basement rocks following preferred directions.

While calcalkaline and peraluminous types are distributed throughout the island, the subalkaline tholeiitic basalts are found almost exclusively in the south.

Transitional types dominate in the north where they are associated with the alkaline facies, in the immediate vicinity of shear zones. (Pasquali,1990; Pensi,1990;Vaccaro,1990; Traversa et al.,1991). Basic alkaline dykes, are found almost exclusively in the northern part of Sardinia.

Integrated geochronological and isotopic Sr studies (Vaccaro,1990; Vaccaro et al.,1991) have enabled us to obtain a detailed picture of the closing stages of the Hercynian chain's tectono-magmatic evolution.

The oldest dyke group (298 to 289 m.y.) is made up of peraluminous and calcalkaline types with high initial Sr isotopic ratio.These are contemporaneous to slightly younger than late-tectonic batholith intrusions (Ghezzo & Orsini,1982).

The majority of these dykes are weakly deformed.

Two tectonic domains, separated by a chronological and chemical discontinuity, respectively occur in the north,and in the south of Sardinia. Characteristic of the southern domain is a limited uplift. In the south-eastern zone, magmatic dyke activity younger than 290 m.y. is very limited, and may -as far as basic types are concerned- be found exclusively in the Sarrabus subalkaline dykes. Roughly contemporaneous with these are the peraluminous rhyolitic dykes in the Mount Ferru zone (Ogliastra, southeastern Sardinia), which even cut across the local ignimbritic units (Atzori & Traversa,1986).

The remarkable uplift of the northern sector can be seen in the progressive change of tectonic style and in the chemical and isotopic composition of dykes.

The crustal thinning -perhaps already evident from the low initial Sr isotopic ratio of 270 m.y. old calcalkaline dykes,(Vaccaro et al.,1991)- is confirmed by the transitional basaltic dykes, which are of clear mantle origin.

The transitional stage is, at least partly, to be attributed probably to the same period as the alkaline one, since the two dyke groups are associated in non intersecting structures parallel to E-W shear planes (Traversa et al.,1991). The alkaline phase, dating to around 230 m.y. (Baldelli et al.,1985a;1985b; Vaccaro et al.,1991) and represented by magmas of asthenospheric origin (Vaccaro, 1990), is post-orogenic in character, and closes the Sardinian Hercynian cycle.

In this paper we examine REE distribution in the various groups of dyke rocks, also with the aim of constraining petrogenetic hypotheses put forward at an earlier date (Atzori & Traversa,1986).

2. Analytical Results

Tables 1,2,3,4 and 5 report REE and Y measured (ppm) by plasma emission on a group of samples selected among those studied by Atzori and Traversa (1986). Previously published major and trace (ppm) elements (Atzori & Traversa, 1986) and initial Sr isotopic ratios (Sri) (Vaccaro,1990; Vaccaro et al.,1991) are also given. Analyses are ordered by increasing ΣREE values.

	254	277	65	310	42	172	109	107	192	63	50	40	210
SiO2	52.42	53.83	52.91	54.25	52.43	56.80	56.22	58.85	56.86	60.27	55.67	63.38	70.50
TiO2	0.78	0.93	0.85	0.88	1.17	0.58	0.92	0.76	1.01	1.01	1.55	1.04	0.30
A1203	15.12	19.04	16.51	16.63	15.91	15.49	16.30	14.82	15.34	16.17	15.45	14.39	14.12
Fe203	2.87	5.97	3.09	4.08	4.05	2.20	3.12	2.14	2.01	2.28	4.33	2.10	0.74
FeO	5.69	2.71	5.60	4.14	4.47	4.60	4.45	4.37	5.50	4.51	4.89	4.33	1.30
MnO	0.17	0.26	0.21	0.22	0.15	0.14	0.15	0.14	0.12	0.17	0.20	0.10	0.03
MgO	8.39	3.54	6.12	4.73	6.34	6.49	4.30	4.70	4.55	2.94	3.93	1.50	0.65
CaO	6.64	7.40	6.47	6.83	6.21	6.62	5.30	6.03	5.84	5.29	5.53	4.06	2.44
Na20	2.65	3.40	2.63	3.25	2.91	2.75	2.87	2.71	2.71	2.46	2.87	3.20	2.70
K20	1.52	2.10	1.97	1.52	2.08	1.63	2.95	2.94	2.18	2.89	2.13	3.37	5.53
P205	0.20	0.24	0.17	0.18	0.30	0.13	0.18	0.14	0.30	0.22	0.56	0.32	0.27
L.O.I.	3.56	0.44	3.46	3.29	3.99	2.58	3.23	2.40	3.58	1.79	2.87	2.21	1.41
Cr	624	n.d.	250	n.d.	341	365	157	194	162	n.d.	1	11	n.d.
Ni	159	n.d.	79	n.d.	111	96	30	40	47	n.d.	4	7	n.d.
V	196	n.d.	190	n.d.	156	177	179	146	145	n.d.	297	121	n.d.
Ba	257	n.d.	485	n.d.	530	217	463	598	464	n.d.	595	719	n.d.
Sr	251	n.d.	292	n.d.	256	190	318	254	292	n.d.	326	224	n.d.
Rb	61	n.d.	99	n.d.	88	82	122	166	97	n.d.	84	112	n.d.
Zr	105	n.d.	142	n.d.	208	66	131	126	216	n.d.	213	327	n.d.
Nb	7	n.d.	8	n.d.	12	6	9	8	12	n.d.	14	18	n.d.
Y	19.93	20.29	28.50	21.83	33.53	18.95	27.45	24.37	30.01	30.97	38.76	48.52	19.66
La	14.84	14.78	23.09	25.66	31.77	9.26	24.67	25.36	33.44	35.40	36.52	53.90	32.40
Ce	31.44	32.23	50.42	51.25	64.93	22.37	45.01	48.78	70.41	71.54	75.39	108.17	63.98
Nd	14.80	16.86	23.12	23.20	31.06	9.40	23.60	22.81	33.18	31.11	35.04	48.44	26.68
Sm	3.81	3.99	5.55	5.07	7.20	2.72	5.57	5.29	7.49	6.73	8.10	10.58	5.43
Eu	0.99	1.17	1.29	1.26	1.57	0.69	1.29	1.11	1.45	1.39	2.10	1.89	0.98
Gd	3.24	3.38	4.81	4.04	5.65	2.45	4.59	4.21	5.70	5.47	6.88	8.22	4.06
Dy	2.93	2.94	4.03	3.26	4.87	2.59	3.95	3.50	4.52	4.47	5.66	7.09	3.07
Er	1.67	1.70	2.31	1.84	2.54	1.63	2.24	1.98	2.36	2.54	2.96	3.72	1.54
Yb	1.61	1.55	2.19	1.67	2.33	1.56	2.16	1.86	2.12	2.39	2.68	3.34	1.43
Lu	0.30	0.23	0.36	0.34	0.37	0.35	0.34	0.30	0.33	0.40	0.41	0.51	0.19
Q	3.24	6.55	5.27	8.28	5.15	9.86	9.47	11.94	11.91	17.55	13.35	20.35	27.63
Or	8.98	12.41	11.64	8.98	12.29	9.63	17.43	17.37	12.88	17.07	12.58	19.91	32.67
Ab	22.41	28.75	22.24	27.49	24.61	23.26	2427	22.92	22.92	20.80	24.27	27.06	22.83
An	24.88	30.50	27.43	26.31	24.21	25.11	22.89	19.60	23.26	24.55	22.99	14.96	10.08
Di	5.51	3.73	2.96	5.20	3.83	5.58	1.89	7.66	3.22	0.31	0.92	2.75	0.36
Hy	25.35	7.08	20.53	12.49	17.18	19.41	14.15	13.26	16.76	12.21	12.56	7.05	2.78
Mt	4.16	6.89	4.48	5.92	5.87	3.19	4.52	3.10	2.91	3.31	6.28	3.04	1.07
Hm	0.00	1.22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Il	1.48	1.77	1.61	1.67	2.22	1.10	1.75	1.44	1.92	1.92	2.94	1.98	0.57
Ap	0.43	0.51	0.36	0.38	0.64	0.28	0.38	0.30	0.64	0.47	1.20	0.68	0.58
ΣREE	75.63	78.83	117.17	117.59	152.29	53.02	113.42	115.20	161.00	161.44	175.74	245.86	139.76
D.I.	34.6	47.7	39.1	44.8	42.0	42.7	51.2	52.2	47.7	55.4	50.2	67.3	83.1
Sri	.70627	.70667	.70708	.70592	.71036	.70682	.70671	.70711	.70738	.71048	.70919	.70889	.70973

254 to 42=BasalticAndesites; 172 to 50=Andesites; 40=Dacite; 210=Rhyolite

Tab. 1 - Major and traces elements, REE and Y and CIPW Norms of calcalkaline dykes.

For the sake of convenience, we shall comment on the data regarding REE according to the main petrochemical groups, reported above.

2.1.*Pre-Permian Volcanism Dykes*

2.1.1.*Dykes of the Calcalkaline Suite*

Within the composite group of calcalkaline rocks (Tab.1),Vaccaro (1990) distinguished two groups based on different initial Sr isotopic values (Sri): one had lower Sri (average value for 8 example was 0.70674$\pm$ 0.00047), the other had higher Sri (average value for 5 samples was 0.70973 $\pm$ 0.00070). In the former group, only poorly evolved rocks such as basaltic andesites and andesites are found; whereas the latter group comprises all petrographic types, from basaltic andesites to rhyolites. In the lower Sri group -which includes types from both the normal calcalkaline and high-K series (Atzori and Traversa, 1986; Vaccaro, 1990)- a poorly defined positive correlation can be observed between fractionation, expressed by the Differentiation Index (D.I.) value of Thornton and Tuttle (1960), and the ΣREE (Fig.2). Various evolutive trends can be discerned, however, on the basis of major and trace element variations, as already pointed out by Atzori and Traversa (1986).

The ΣREE of the low Sri rock group is quite modest,

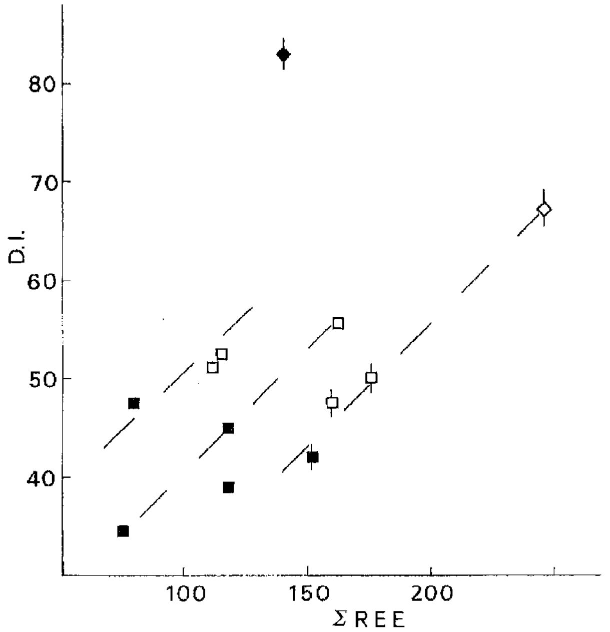

Fig. 2 - D.I. vs. ΣREE diagram for calcalkaline dyke. Full square=basaltic andesites; empty square=andesites; open diamond=dacite; full diamond=rhyolite. Barred symbols=high Sri dykes

considering the evolutive degree of the rocks; it varies from 53 to 161, with an average of 97 $\pm$ 23 for basaltic andesites and 111 $\pm$ 44 in the case of andesites. In the REE pattern diagram (Fig.3) the low Sri group presents enrichments of LREE ranging from 20 to 100 times chondritic and LREE/HREE ratios increasing with fractionation. Negative Eu anomaly is, on the whole, not much marked.

The high Sri dykes, all belonging to the high-K calcalkaline series, are richer in REE; ΣREE varies from 152 for a basaltic andesite to 246 in the case of a dacite. Only one rhyolite (sample 210) gives anomalous value (140) with respect to its evolution degree, considering that separation of REE-rich phases has not been found. These,

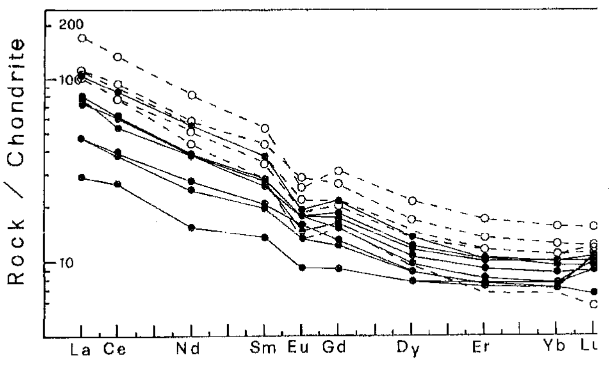

Fig. 3 - Chondrite normalized REE patterns for calcalkaline dykes. Full circles and solid line=low Sri rocks; empty circles and dashed lines=high Sri rocks

then, are primary differences, further supported by the absence of reports of fractionation among high Sri types.

Chondrite-normalized REE patterns (Fig.3) show that this high Sri group presents values which range from 100 to 170 times chondritic for Light REE, a high value of Light/Heavy ratio, marked negative Eu anomaly, and pronounced fractionation of Heavy REE. The LREE/HREE ratio -a feature peculiar to this group- tends to remain constant, or to diminish, if rhyolite (sample 210) is also taken into consideration.

To conclude, the lower Sri group, in comparison with the higher Sri calcalkaline rocks, is characterized, for analogous levels of evolution, by lower ΣREE, lower enrichments of LREE with respect to chondrites, lower LREE/HREE ratios and less pronounced Eu anomaly.

The differences encountered between the two calcalkaline dyke groups could be attributed to the different composition of the source zone. In fact, crustal contamination during uprising of melts would appear to be excluded by the lack of correlation between Srppm and Sri.As concerns the high-Sri group, the overall homogeneity of Sri values (0.70889 to 0.71048) and the lack of any positive correlation between D.I. and Sri (Fig.4) in variously evolved rocks (D.I. from 42 for basaltic andesites to 83 for rhyolites) are against crustal contamination processes during ascent of the magmas. The high Sri values can be attributed to the

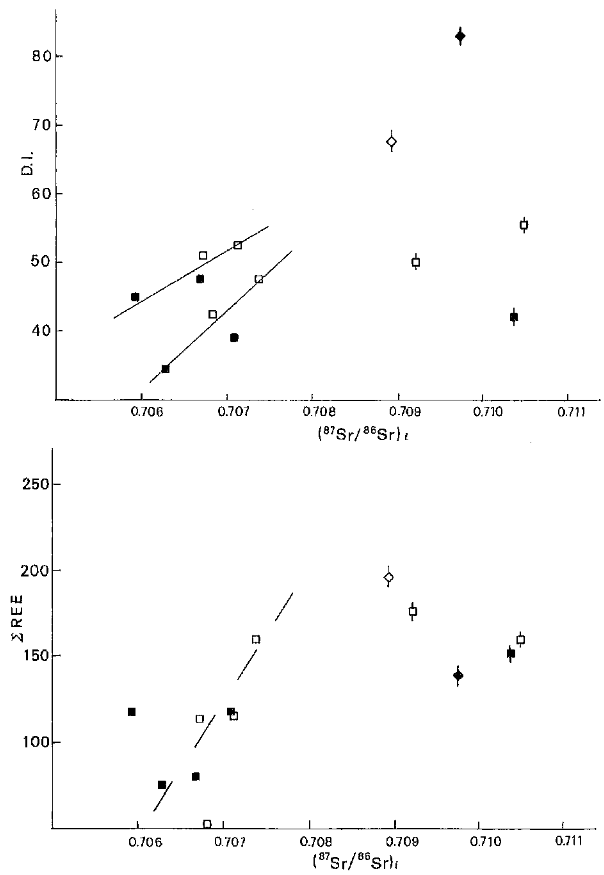

Fig. 4 - D.I. vs. Sri and ΣREE vs.Sri diagrams for calcalkaline dykes. Symbols as in fig. 2.

nature of the source. The same can be said regarding the relationships ΣREE-Sri (Fig.4), which even exhibit a negative correlation. For the low Sri types, a crustal assimilation could be supported on the basis of the poor positive correlation between fractionation (D.I. and ΣREE) and Sri (Fig.4)

The values of the ΣREE of lower Sri rocks(97 $\pm$ 23 for basaltic andesites and 111 $\pm$ 44 for andesites) are not only lower than those of the transitional basalts, both olivine (122 $\pm$ 31) and quartz-normative (133 $\pm$ 31): they are also lower than those of the subalkaline tholeiitic basalts (115 $\pm$ 23). This is further proof for the primitive character of many basaltic andesites, and for the absence of a link between calcalkaline types and tholeiitic basalts, as already pointed out by Atzori and Traversa (1986).

Lastly, it must be stressed that the high Sri calcalkaline dykes are entirely similar to Hercynian granitoid plutons from Sardinia (Poli et al.,1989) regarding both age (298 to 289 m.y.; Vaccaro et al.,1991) and compositional features, including Sri values and REE content. The lower-Sri types, on the other hand, can be referred to a more recent intrusive cycle (ca. 270 m.y.), characterized by less evolved products (from basaltic andesites to andesites: D.I from 35 to 52): this cycle shows the lowest Sri isotopic ratios (0.70592 to 0.70738) reported so far for calcalkaline products of Hercynian magmatism in Sardinia.

2.1.2. *Peraluminous Rhyolitic Dykes*

Several intrusive events were identified among peraluminous dykes (Atzori & Traversa, 1986; Pasquali, 1990; Pensi, 1990; Vaccaro, 1990; Vaccaro et al.,1991; Traversa et al.,1991) occurring contemporaneously with both the calcalkaline suite and at least part of the late-Permian basaltic phase.

REE distribution (Tab.2) reveals a division into two groups, unrelated to age, one with extremely low ΣREE values (22 to 46), the other presenting higher values (139-174).

The low-value ΣREE group is also to be distinguished on account of the very low LREE enrichment (about 20 times chondritic) (Fig.5), minimal LREE/HREE fractionation, and the typical, slightly negative Nd and Eu anomalies. These peraluminous dykes present the highest Sri values (0.71650-0.71490) (Vaccaro, 1990).

The high-value ΣREE group shows ΣREE similar to those of the non-peraluminous acid types in the calcalkaline series. Moreover, the pronounced LREE enrichment with respect to chondrites (from 70 to 100 times chondritic) (Fig. 5), the marked negative Eu anomaly, and the LREE/HREE fractionation all confirm the analogy between the high ΣREE peraluminous types and the calcalkaline rhyolitic dykes.

The only Sri datum available for the high ΣREE dyke group gives a significantly lower value (0.71121 $\pm$ 0.00003 Vaccaro, 1990), which could once again suggest a connection with the calcalkaline suite.

	108	211	269	128	44
SiO2	74.21	73.29	73.39	72.88	72.61
TiO2	0.05	0.08	0.11	0.18	0.21
A1203	14.47	14.71	14.63	13.94	13.81
Fe203	0.37	0.57	0.48	1.01	1.16
FeO	0.49	0.29	0.31	1.13	1.17
MnO	0.03	0.03	0.04	0.05	0.05
MgO	0.15	0.19	0.21	0.31	0.49
CaO	0.53	0.50	0.48	0.78	0.87
Na20	3.74	3.65	3.74	3.95	4.03
K20	4.57	5.18	5.09	4.01	4.09
P205	0.25	0.24	0.15	0.11	0.08
L.O.I.	1.14	1.26	1.37	1.65	1.44
Y	11.44	15.64	12.60	35.35	33.25
La	4.22	7.18	8.60	29.92	39.71
Ce	7.91	15.89	20.41	58.88	78.01
Nd	3.44	6.05	8.23	26.98	32.50
Sm	1.13	1.97	2.42	6.65	7.07
Eu	0.28	0.52	0.60	1.00	0.94
Gd	0.13	1.95	2.15	5.36	5.45
Dy	1.58	2.32	1.90	4.87	4.66
Er	0.16	0.97	1.03	2.70	2.54
Yb	0.91	0.77	0.84	2.59	2.51
Lu	0.16	0.09	0.13	0.34	0.32
Q	34.01	31.51	31.21	32.26	30.72
C	2.95	2.71	2.45	1.95	1.35
Or	27.00	30.60	30.08	23.70	24.16
Ab	31.63	30.87	31.65	33.42	34.08
An	1.16	1.07	1.40	3.15	3.84
Hy	1.35	0.47	0.59	1.81	1.93
Mt	0.54	0.80	0.70	1.46	1.68
Hm	0.00	0.02	0.00	0.00	0.00
Il	0.09	0.15	0.21	0.34	0.40
Ap	0.53	0.51	0.36	0.26	0.17
ΣREE	21.62	37.71	46.31	139.29	173.71
D.I.	92.6	93.0	92.9	89.4	89.0
Sri	.71650	n.d.	.71490	.71121	n.d.

Tab. 2 - Major elements, REE and Y and CIPW Norms of peraluminous rhyolitic dykes

To conclude, REE distribution and Sr data regarding peraluminous rocks clearly indicate that very different sources played their part in crustal melting processes. These can partly be attributed to former pelitic-arenaceous rocks with the highest Rb/Sr ratios, and partly to much less radiogenic protolites (Vaccaro et al., 1991). A possible explanation for the remarkably low values of ΣREE (22 to 46) displayed by the first group can be found in an extremely modest degree of melting of the pelitic-arenaceous source, which has brought about its non-participation in the melting processes of the carrier phases of REE such as zircon, apatite, etc. remained among restitic phases.

2.2 *Late-Permian Dykes*

2.2.1 *Basic Alkaline Dykes*

These are represented by variously-evolved types (D.I. 29.2 to 55.1). Some are cumulitic in nature and so the rocks pass from picritic types right through to mugearitic ones, across basaltic and hawaiitic lithotypes.

All samples examined (Tab.3) were characterized by high ΣREE values (194 to 257), the lower values being

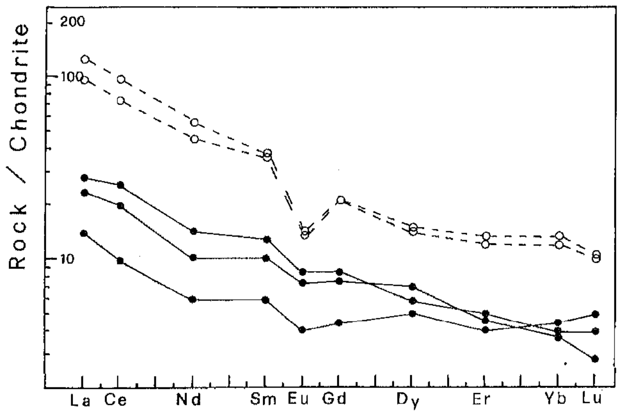

Fig. 5 - Chondrite-normalized REE patterns for perluminous rhyolitic dykes. Full circles and solid lines : Sri= 0.71650-0.71490; empty circles and dashed lines: Sri=0.71121;

typical of the rocks with intermediate value of D.I. (Fig.6). The high ΣREE value of cumulitic sample 221 can be related to apatite crystals included in cumulus biotite and amphibole.

Chondrite-normalized REE patterns for the alkaline rocks are reported in Fig.7. All the analyzed samples are strongly enriched with respect to chondrite, especially in LREE and are characterized by a strong LREE/HREE fractionation. Sample 221 shows a lower LREE/HREE ratio in relation to the presence of cumulus amphibole which slightly partitions HREE.

Sri values (average of 5 samples 0.70389 ± 0.00041) (Vaccaro,1990; Vaccaro et al., 1991) indicate that the alkaline melts definitely derived from an undepleted astenospheric source. Sri variations (0.70332 to 0.70443) can be referred to an isotopically inhomogeneous source. In fact, the lowest Sri value of the most evolved rock is against crustal contamination processes during the ascent of the melts.

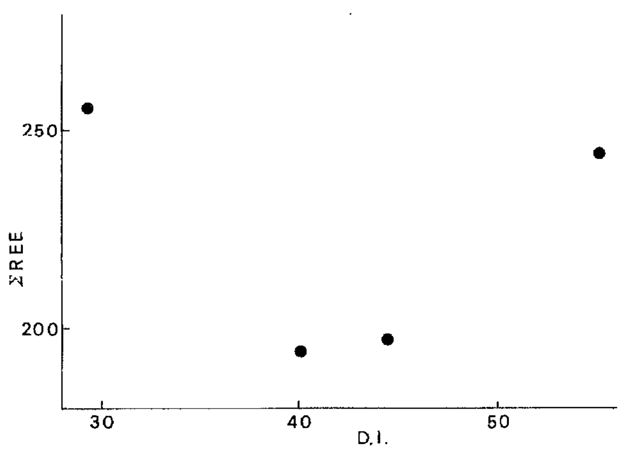

Fig. 6 - ΣREE vs. D.I. diagram for alkaline basic dykes

2.2.2. *Transitional Basaltic Dykes*

These form the most numerous group of basic dykes (Tab.4), with olivine- and quartz-normative types, and variable degree of evolution (D.I. 24.0 to 40.5) (Atzori & Traversa, 1986).

ΣREE has a good positive correlation with the degree of evolution (Fig.8). The less evolved types show a rather flat trend and a 30 times chondritic LREE enrichment. A variation in REE trend can be observed, characterized by more evolved types: about 60-70 times chondritic enrichment, with pronounced LREE/HREE fractionation; and the presence of negative Eu anomaly (Fig.9), typical of plagioclase fractionation-linked evolutions.

Sri values (Vaccaro, 1990; Vaccaro et al.,1991) are highly variable, ranging from 0.70385 to 0.70770 (average value for 11 samples = 0.70533 ± 0.00116) in the case of olivine-normative basalts and from 0.70450 to 0.70736 (average value for 6 samples = 0.70617 ± 0.00098) for the quartz-normative basalts.

The D.I. vs. Sri diagram (Fig.10), in particular, reveals the presence of two main series with positive correlation, one with lower Sri (< 0.70550) and the other with higher Sri (> 0.70550): the first of these is characterized by a decrease, the second by an increase in Na2O/K2O ratio (Fig.11). The majority of quartz-normative types fall in the group with Sri >0.70550, and can be distinguished also on account of their generally potassic affinity (Fig.12).

On the basis of the positive D.I.-Sri correlation (Fig.10), the Sri variations could be explained by crustal contamination during the ascent of melts (AFC). However, this is to be excluded, on account of the lack of positive correlation both between ΣREE and Sri (Fig.10) and between K2O and SiO2 (Table 4). Sri variations can be primary without any relation to the evolution degree. The strong heterogeneity of the source is evidenced by the Na2O/K2O ratios (1 to 5.5), silica being equal (Fig.13), and by the lack of Ni-K2O negative correlation which typifies the fractional crystallization processes.

In conclusion, variations of Sri and some major and trace elements testify that transitional melts were generated

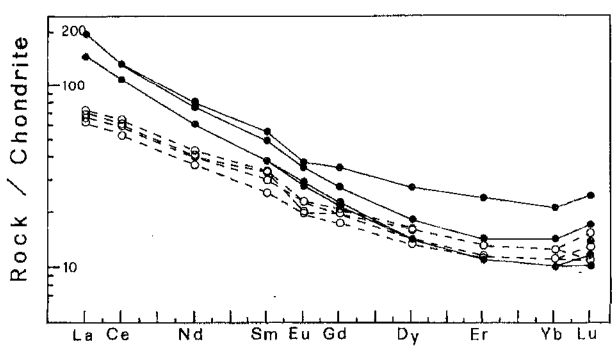

Fig. 7 - Chondrite-normalized REE patterns for alkaline basic dykes (full circles and solid lines) and tholeitic basaltic dikes (empty circle and dashed lines).

	216	218	219	221
SiO2	46.46	47.77	49.80	42.12
TiO2	2.27	2.17	2.29	2.31
A1203	15.31	15.60	16.76	13.49
Fe203	4.33	5.45	4.61	5.64
FeO	7.40	7.02	6.32	8.69
MnO	0.28	0.31	0.22	0.34
MgO	6.21	4.31	3.51	7.87
CaO	8.33	7.32	5.42	9.03
Na20	4.07	4.11	3.92	2.26
K20	1.44	1.63	3.77	2.02
P205	0.72	0.75	0.66	0.74
L.O.I.	3.19	3.55	2.71	5.49
Cr	178	127	31	357
Ni	88	75	38	238
V	197	176	138	200
Ba	349	260	939	724
Sr	759	986	922	939
Rb	17	38	89	49
Zr	308	323	365	247
Nb	75	80	86	71
Y	28.85	28.73	36.92	70.88
La	46.11	47.72	60.65	59.28
Ce	87.19	89.03	106.71	108.65
Nd	36.02	36.39	45.46	47.22
Sm	7.36	7.28	9.38	10.61
Eu	2.09	2.01	2.54	2.67
Gd	5.84	5.55	7.10	8.98
Dy	4.58	4.57	5.90	8.87
Er	2.33	2.40	3.00	5.15
Yb	2.10	2.11	2.92	4.42
Lu	0.37	0.33	0.55	0.81
Or	8.51	9.63	22.27	11.93
Ab	27.95	34.76	32.46	15.21
An	19.26	19.31	17.01	20.70
Ne	3.51	0.00	0.37	2.11
Di	14.38	10.20	4.85	15.89
Hy	0.00	4.69	0.00	0.00
01	11.06	6.01	7.84	14.48
Mt	6.28	7.90	6.68	8.18
Il	4.31	4.12	4.35	4.39
Ap	1.54	1.60	1.41	1.58
ΣREE	193.99	197.39	244.21	256.66
D.I.	40.0	44.4	55.1	29.2
Sri	.70433	.70423	.70332	.70374

Tab. 3 - Major and trace elements, REE and Y and CIPW Norms of basic alkaline Dykes

from a scarcely to greatly metasomatized mantle, intermediate between the source of low-Sri calcalkaline magmas and that of alkaline magmas. Such variations in mantle composition reflect the geodynamic evolution from an orogenic to a continental setting.

2.2.3 *Tholeitic Basaltic Dykes*

The types analyzed (Tab.5) represent scarcely-evolved (D.I. 23.2 to 28.9), REE-relatively rich basalts. The ΣREE,

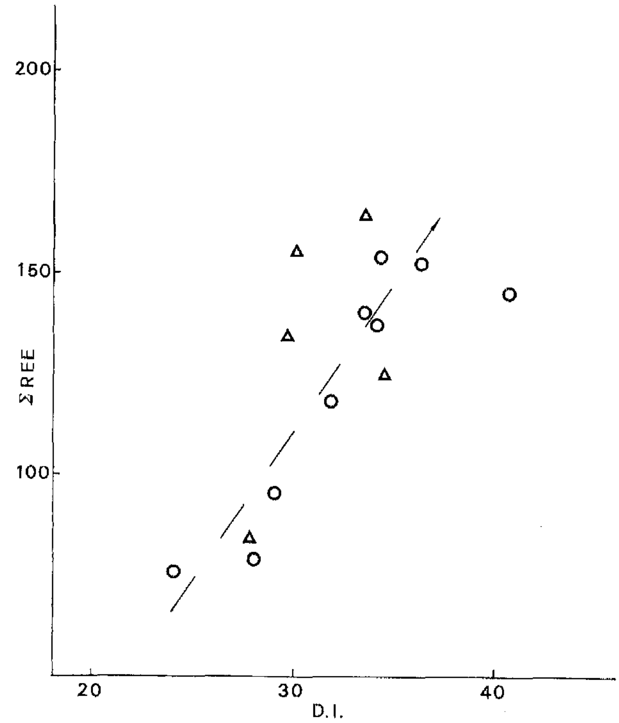

Fig. 8 - ΣREE vs. D.I. diagram for transitional, Ol-normative (empty circles) and Q-normative (empty triangle) basaltic dikes.

in fact, which ranges from 105 to 123, is higher than in transitional basalts, evolutive degree being equal. LREE enrichment ranges from 50-70 times chondritic (Fig.7), which is marked in the case of scarcely-evolved subalkaline types, moreover there is a consistent LREE fractionation with respect to HREE, comparable with that of the more evolved types among transitional basalts. Particularly as regards HREE, significant analogies with alkaline types are evident (Fig. 7).

Average Sri value (0.70568 ± 0.00046, for 4 samples) (Vaccaro,1990) and the lack of a link between Sri and fractionation (D.I. and ΣREE) and **Srppm** and Sri (Table 5) suggest probable contamination at source. Considering the primitive character of these subalkaline basalts,the high

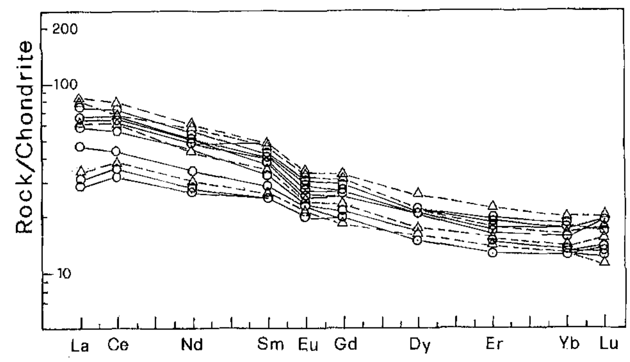

Fig. 9 - Chondrite-normalized REE patterns for transitional, Ol-normative (empty circles and solid lines) and Q-normative (empty triangle and dashed lines) basaltic dykes.

	204	233	147	226	196	242	198	220	95	240	227	255	162	252
SiO2	45.29	45.60	47.07	47.68	47.97	46.27	48.45	46.80	47.12	46.58	48.97	46.31	46.69	46.43
TiO2	1.63	1.55	1.65	1.68	1.86	1.63	1.48	1.86	2.44	1.86	1.62	2.31	2.14	1.89
A1203	17.06	17.62	16.66	16.29	16.29	16.69	16.60	16.41	16.34	16.43	16.41	16.08	16.55	16.21
Fe203	3.92	3.87	4.57	4.21	2.74	5.34	5.07	5.82	5.45	5.81	4.45	4.92	6.15	5.49
FeO	6.17	6.67	6.94	6.75	7.86	6.48	5.84	6.32	7.15	6.12	6.57	8.17	6.91	7.36
MnO	0.19	0.18	0.19	0.19	0.19	0.21	0.20	0.24	0.20	0.20	0.18	0.24	0.22	0.21
MgO	9.14	7.25	6.65	7.28	5.67	6.47	5.17	6.11	5.10	7.04	6.16	5.22	6.28	5.18
CaO	9.71	9.38	8.87	7.60	8.54	7.60	7.54	6.94	8.29	8.16	7.27	8.42	7.40	7.69
Na20	2.49	2.74	2.96	2.77	3.49	3.16	3.83	3.73	3.44	2.39	2.78	2.70	2.33	2.60
K20	0.50	0.80	0.69	1.42	0.76	1.14	1.37	0.79	0.88	1.02	1.60	0.92	1.21	1.77
P205	0.25	0.29	0.34	0.39	0.47	0.54	0.47	0.61	0.54	0.28	0.44	0.57	0.51	0.84
L.O.I.	3.65	4.06	3.38	3.73	4.16	4.48	3.98	4.36	3.05	4.11	3.55	4.14	3.60	4.33
Cr	153	107	184	247	87	177	166	94	138	87	209	143	133	139
Ni	95	106	56	131	54	93	97	51	60	76	120	80	57	76
V	219	214	263	228	217	180	170	226	277	228	222	264	269	202
Ba	150	139	173	205	201	314	267	211	330	174	221	214	467	363
Sr	291	346	337	269	299	273	332	403	337	291	237	269	315	248
Rb	38	34	35	57	39	60	49	36	51	70	41	43	62	62
Zr	164	153	174	206	289	293	386	268	291	186	224	259	276	293
Nb	5	7	10	11	12	12	18	14	15	7	13	12	15	19
Y	35.33	32.99	34.35	38.22	47.35	47.36	47.79	46.12	49.13	35.60	37.87	48.19	67.28	46.95
La	9.24	9.71	15.04	19.06	20.78	21.67	24.21	24.37	24.42	11.10	20.28	19.45	25.55	27.01
Ce	27.60	30.00	36.50	47.38	54.34	56.18	57.87	61.30	61.08	31.50	51.71	52.70	55.42	67.27
Nd	16.62	17.15	20.66	25.52	29.94	30.55	30.68	34.53	33.98	18.53	26.88	30.36	35.00	35.44
Sm	4.90	4.91	5.62	6.52	7.83	7.73	7.48	8.40	8.60	5.07	6.72	7.99	9.54	8.90
Eu	1.52	1.51	1.69	1.76	1.98	2.02	1.81	2.12	2.24	1.60	1.71	2.11	2.45	2.31
Gd	4.81	4.83	5.11	5.86	6.81	7.07	6.84	7.35	7.95	4.87	5.97	7.31	8.72	8.03
Dy	5.03	4.90	5.01	5.67	6.88	6.92	6.88	6.83	7.25	5.22	5.66	7.04	8.60	7.10
Er	2.84	2.77	2.85	3.14	3.88	3.90	3.93	3.67	4.04	2.99	3.13	3.81	4.79	3.75
Yb	2.64	2.53	2.60	2.89	3.66	3.72	3.83	3.35	3.67	2.80	2.90	3.49	4.14	3.41
Lu	0.46	0.40	0.43	0.52	0.64	0.59	0.54	0.61	0.61	0.37	0.52	0.63	0.66	0.56
Q	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.50	1.58	1.31	3.12	0.91
Or	2.95	4.73	4.08	8.39	4.49	6.73	8.09	4.67	5.20	6.03	9.45	5.44	7.15	10.46
Ab	21.06	23.17	25.03	23.43	29.52	26.72	32.39	31.55	29.09	20.21	23.51	22.83	19.71	21.99
An	33.90	33.42	30.14	27.83	26.54	27.99	24.06	25.71	26.55	31.09	27.58	29.04	31.13	27.34
Di	10.19	9.16	9.54	6.08	10.66	5.22	8.48	4.09	9.17	6.14	4.75	7.61	2.05	4.75
Hy	8.97	8.09	11.33	13.77	9.34	10.65	7.33	11.45	9.49	18.36	19.10	16.85	19.13	16.85
01	11.89	10.80	3.27	3.36	6.08	4.99	3.75	2.81	0.07	0.00	0.00	0.00	0.00	0.00
Mt	5.68	5.61	6.63	6.10	3.97	7.74	7.35	8.44	7.90	8.42	6.45	7.13	8.92	7.96
Il	3.10	2.94	3.13	3.19	3.53	3.10	2.81	3.53	4.63	3.53	3.08	4.39	4.06	3.59
Ap	0.53	0.62	0.73	0.83	1.01	1.15	1.01	1.30	1.15	0.60	0.94	1.22	1.09	1.80
ΣREE	75.66	78.71	95.51	118.32	136.74	140.35	144.07	152.53	153.84	84.05	125.48	134.89	154.87	163.78
D.I.	24.0	27.9	29.1	31.8	34.0	33.4	40.5	36.2	34.3	27.7	39.5	29.6	30.0	33.4
Sri	.70385	.70487	.70677	.70511	.70450	.70514	.70418	.70770	.70552	.70450	.70664	.70658	.70621	.70572

Tab. 4 - Major and trace elements; REE and Y and CIPW Norms of transitional basaltic dykes.

REE grade could be related to the same mantle metasomatism, probably connected to fluids of crustal origin, responsible for the enrichment of low ionic potential elements as well as incompatible ones (Atzori and Traversa, 1986; Vaccaro, 1990).

3. Conclusions

REE distribution in the late Hercynian dykes from Sardinia, together with field evidence and previously published major- and trace-element data (Atzori & Traversa, 1986; Pasquali, 1990; Pensi, 1990; Traversa et al., 1991) and geochronological and Sr isotopic results (Vaccaro, 1990; Vaccaro et al., 1991), allow us to draw the following conclusions.

The late Hercynian magmatism took place during two phases, each with distinct compositional and tectono-magmatic features, respectively older and younger than Permian volcanism (Traversa, 1978). The first phase is represented by products with evident orogenic to late-orogenic character which consist of calcalkaline, normal to high potassium, and peraluminous anatectic lithotypes.It developed during two distinct stages, one carboniferous (298 to 289 m.y.), and the other Permian (ca.270 m.y.) (Vaccaro, 1990; Vaccaro et al., 1991). The two stages of the first phase both belong to the same thermal crisis that affected the Sardinian basement between 300 and 270 m.y. bp. and was responsible for the batholith emplacement. The second phase of the late Hercynian dyke magmatism of Sardinia, is represented almost exclusively by basaltic products, and by rare peraluminous rhyolitic dykes. The latter are linked to crustal melting phenomena, probably due to the uprising of these basic magmas. The basaltic dykes, mainly transitional, cut across the Permian ignimbrites (Traversa, 1968;1969) which, in Gallura, are dated at about 260 m.y. (Edel et al., 1981).

Subalkaline basalts are found together with the

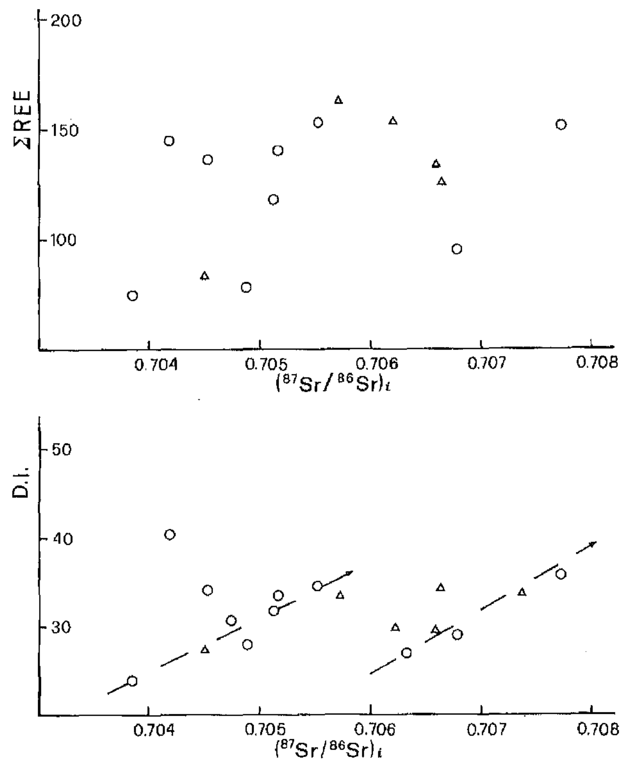

Fig. 10 - ΣREE vs. Sri and D.I. vs. Sri diagrams for transitional basaltic dykes. Simbols as in fig. 8

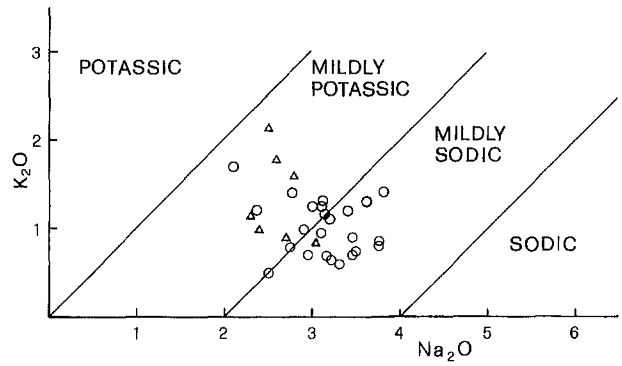

Fig. 12 - K2O vs Na2O diagram for transitional basaltic dykes. Symbols as in fig.8

transitional types in the central-southern sector, while alkaline types, always associated with the transitional ones near the shear zones, are restricted to the northern area (Atzori & Traversa, 1986; Pasquali, 1990; Pensi, 1990; Traversa et alii, 1991).

Age relationships among the different basaltic types are not yet altogether clear, partly because it had not been possible till now to date the transitional types. However, the following evidences lead us to believe, at present, that the overall basaltic phase is most probably younger than Permian volcanism:

1 - the lack of correlation between subalkaline basalts and calcalkaline rocks;

2 - the 245 m.y. isochrone defined by the Sarrabus comagmatic subalkaline tholeiitic dykes (Vaccaro, 1990);

3 - the age of approximately 230 m.y. of the alkaline dykes in the north (Baldelli et al.1985a;1985b; Vaccaro, 1990; Vaccaro et al., 1991);

4 - field evidence for the different types of dykes, particularly the reciprocal and repeated intersections of leucocratic and melanocratic hypoabyssalites (Atzori & Traversa,1986) ;

5 - the contemporaneity between alkaline and transitional types in the Concas area (Pasquali,1990; Pensi, 1990; Traversa et al., 1991); and concluding,

6 - the later emplacement of the transitional basalts (formerly called "diabases";Traversa,1968) with respect to the Permian volcanic rocks, as observed for the first time by Traversa (1968;1969) in Gallura, and successively also noted by Lombardi et al. (1974) and Atzori and Traversa (1986) in other parts of Sardinia.

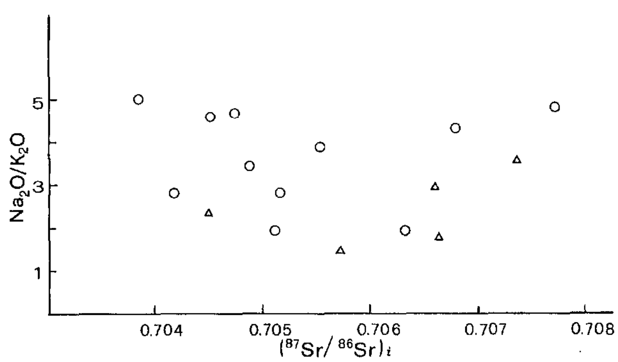

Fig. 11 - Na2O/K2O vs. Sri diagram for transitional basaltic dykes. Symbols as in fig.8

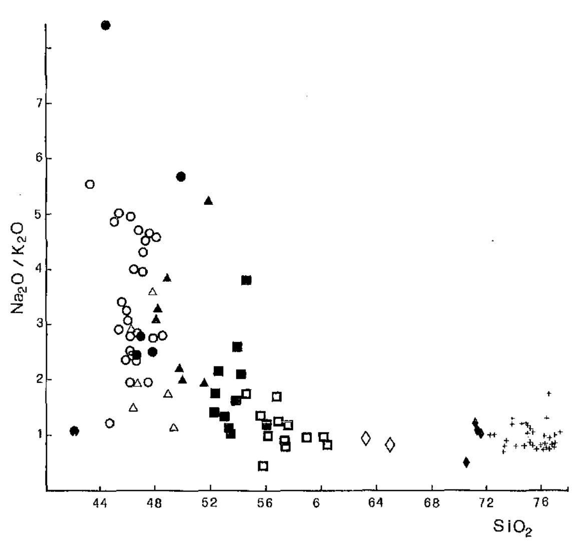

Fig. 13 - Na2O/K2O vs. SiO2 diagram for transitional basaltic dykes. Symbols as in fig. 8

	10	53	8	9
SiO2	48.16	48.91	46.61	48.23
TiO2	1.51	1.37	1.40	1.56
A1203	15.56	16.90	15.58	15.68
Fe203	3.16	3.94	4.17	3.15
FeO	6.19	5.97	5.65	6.76
MnO	0.18	0.17	0.23	0.21
MgO	7.33	6.25	6.23	5.90
CaO	9.78	9.16	10.98	10.41
Na20	2.30	2.63	1.86	2.32
K20	0.74	0.68	0.75	0.70
P205	0.26	0.28	0.29	0.32
L.O.I.	4.83	3.73	6.24	4.75
Cr	193	152	220	181
Ni	91	85	111	74
V	195	198	186	203
Ba	230	332	207	227
Sr	324	361	326	319
Rb	44	29	104	75
Zr	191	183	183	206
Nb	8	8	9	9
Y	31.01	31.24	36.00	36.38
La	19.17	20.78	22.81	22.03
Ce	43.34	46.84	48.59	51.12
Nd	21.47	24.08	24.59	25.13
Sm	5.29	5.69	6.39	6.11
Eu	1.40	1.61	1.47	1.66
Gd	4.55	5.06	5.32	5.34
Dy	4.45	4.54	5.22	5.13
Er	2.49	2.43	2.80	2.88
Yb	2.27	2.22	2.61	2.67
Lu	0.41	0.34	0.36	0.49
Q	1.43	2.66	3.06	2.52
Or	4.37	4.02	4.43	4.14
Ab	19.45	22.24	15.73	19.62
An	29.95	32.30	31.95	30.31
Di	13.67	9.22	16.67	15.76
Hy	18.28	16.89	12.58	14.67
Mt	4.58	5.71	6.05	4.57
Il	2.87	2.60	2.66	2.96
Ap	0.56	0.60	0.62	0.68
ΣREE	104.84	113.59	120.16	122.56
D.I.	25.2	28.9	23.2	26.3
Sri	.70541	.70637	.70546	.70547

Tab. 5 - Major and trace elements, REE and Y and CIPW Norms of tholeiitic basalts

The calcalkaline suite, dominated by basaltic-andesites and andesites is widespread throughout the island; however there is a compositional variation with increase in potassium percentage towords the south, i.e. the outer orogenic zone. In this area, in fact, the basaltic andesites belonging to the high-K calcalkaline series are normally found;these are accompanied by evolved types, andesitic and dacitic, always with high-K character.

In the calcalkaline suite, two groups can be recognized with different Sri and REE distribution: their characteristics are basically to be attributed to the composition of the source zone. The high-Sri types all belong to the high-K calcalkaline series, whereas those with low Sri prevalently belong to the normal calcalkaline series.

The peraluminous group,consisting of micrograntes and rhyolites (Auct. granite porphyrs),is widespread throughout Sardinia, and -like the calcalkaline group- mainly developed in both Carboniferous and Permian times. As mentioned above, occurrences of late-Permian peraluminous dykes are rare. The high Sri values confirm the anatectic origin of these rocks, as suggested by Atzori & Traversa (1986).Sri values and REE distribution bring out two crustal source types, unrelated in age, one with original pelitic-arenaceous composition, the other attributable to less radiogenic protolites, wich reveal a link with the calcalkaline suite (Vaccaro et al.,1991).

The transitional types -as observed before- represent the most widespread basalts in the late Hercynian dyke magmatism of Sardinia. Comparative analysis of REE distribution and Sri reveals the presence of different parental magmas from mildly potassic to mildly sodic, olivine- or quartz-normative- which developed from a metasomatized mantle.

The marked Nb anomaly in the spidergrams of the transitional basalts (Atzori & Traversa, 1986; Vaccaro, 1990) explains how even the apparently more primitive samples have recorded the metasomatic event.Even the Ba/Y vs. Zr/Y diagram (Fig.14) immediately emphasizes the enriched character of the source of the transitional basalts which fall within the field of the "Enriched-Morb" (Piccirillo et al.,1989).In other words, in the source of the transitional basalts a mantle metasomatism analogous to that linked with a subduction of a lithospheric slab during an orogenic or late orogenic phase can be recognized.

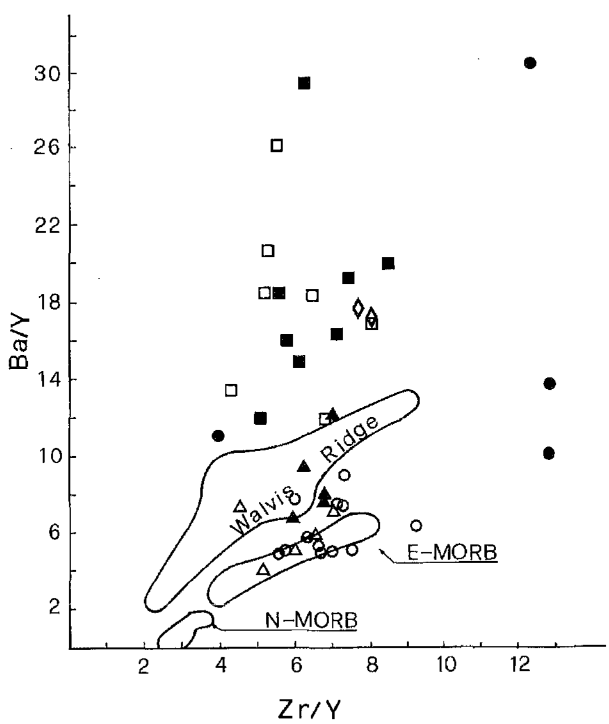

Fig. 14 - Ba/Y vs. Zr/Y diagram for late Hercynian dikes from Sardinia. Full triangles=subalkaline basaltic dikes; full circles=alkaline basic dykes; Other symbols as in figg.2 and 8.

The varying geographical distribution of the subalkaline and alkaline basalts -the former in the south, the latter in the north- is directly linked to the different structural meaning of the two Sardinian subcontinents (Edel et al., 1981; Vaccaro, 1990; Vaccaro et al., 1991; Traversa et al.,1991), and hence to the different crustal thickness of the two areas made evident by the lack of reexhumation in the south, and by the presence of migmatites with granulitic and eclogitic relicts in the north (Ghezzo et al., 1979; 1982).

The subalkaline types have characteristics which indicate that they originated from a lithospheric mantle source, probably metasomatized in an orogenic-late orogenic phase, although they are in a structural situation which we consider post-orogenic -as also suggested by the lack of crustal contamination,inhibited by the rapid ascent of the melts.

The alkaline types, on the other hand, display characteristics which are typical of magmas originating from an asthenospheric, uncontaminated source, in a structural "shear" situation which favours rapid uprising of magmas of deep origin in a sector characterized by marked crustal thinning.

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MAJOR ELEMENTS, 3d TRANSITION ELEMENTS, Cu AND Sr GEOCHEMISTRY OF PERIDOTITIC ROCKS WITHIN THE AUSTRIDIC CRYSTALLINE BASEMENT, NONSBERG AREA, NORTHERN ITALY

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Abstract

The peridotites from the Nonsberg area, eastern Italian Alps are subdivided on a petrographic and textural basis into four groups: coarse-type, porphyroclastic, serpentinized and fine-type. The fine-type peridotites have been previously considered the metamorphic derivatives at lower crustal conditions of coarse-type, spinel lherzolite protoliths and some geochemical characteristics of the former have been ascribed to crustal metasomatism. Eighty samples of peridotites (13 coarse-type, 7 porphyroclastic, 5 serpentinized and 55 fine-type) were analysed for major elements, Cu, Sr and 3d transition elements in the aim to confirm whether or not such metasomatism took place. Few chemical differences exist in terms of major elements between the coarse-type and fine-type peridotites except for K and Na which are significantly higher in the second group. The observed chemical differences, plots of percent SiO_2, MgO, FeOtot and Na_2O+K_2O vs. the MgO/FeOtot ratio and CaO vs. Al_2O_3 for all the peridotites represent the trends expected for residua after variable degrees of melt extraction. The same features are shown by the Ni-Co-Sc fractionation. The 3d transition element contents are quite similar in all four groups of peridotites. The patterns of 3d transition elements normalized to primitive mantle abundances show perturbation at Ti and they are generally flat from V to Ni in all the peridotites. The amount of Cu is comparable in all the peridotites, on the contrary the Sr content and distribution are quite different. The fine-type peridotites have about three times the Sr content of the coarse-type peridotites. It emerges that the 3d transition elements and Cu were not mobilized during the recrystallization of mantle material (coarse-type) in the lower crust while K, Sr and to a lesser extent, Na enrichments indicate crustal metasomatism.

Key words: *peridotites, Nonsberg area, Eastern Italian Alps, geochemistry, crustal metasomatism.*

1. Introduction

Garnet-bearing, alpine type peridotites occur within medium-high grade metamorphic rocks at several localities (Carswell,1990). Among them, some small garnet-bearing peridotites outcrop within garnet- and kyanite-bearing gneisses and migmatites of the palaeozoic Austroalpine terrane from the Non and Ultimo valleys, Nonsberg area, Eastern Italian Alps (Morten & Obata,1983; Obata & Morten,1987). Since their mineral assemblage is unstable under crustal physical conditions (O'Hara et al.,1971), several and conflicting hypotheses have been proposed to explain their origin. In summary, they are thought to represent i) mantle material intruded as solid slices within the crust, ii) in situ metamorphic products of a subducted slab of continental crust, iii) the result of subsolidus and isobaric cooling reactions which transform spinel to garnet lherzolite.

In the Nonsberg case the garnet-bearing peridotites are considered to be derived from spinel lherzolite protoliths through syntectonic recrystallization under lower-crustal conditions (Obata & Morten,1987). A geochemical study of the REE, minor elements and Sr-Nd isotopes (Morten & Obata,1990; Petrini & Morten,1990) on few (ten) but selected samples has suggested that some geochemical characteristics of the garnet-bearing and garnet-free recrystallized peridotites can be explained as due to a crustal metasomatic event suffered during their syntectonic recrystallization. The aim of this paper is to evaluate the above hypothesis, i.e. crustal metasomatism, on the basis of a larger amount of data, i.e. 80 samples.

2. Outlines of regional geology and petrography

The peridotites studied outcrop in an area located about 40 km west of Bolzano and 70 km north of Trento (Fig. 1). In the area, the Palaeozoic crystalline rocks belonging to the "Scisti del Tonale" series of the Austrides

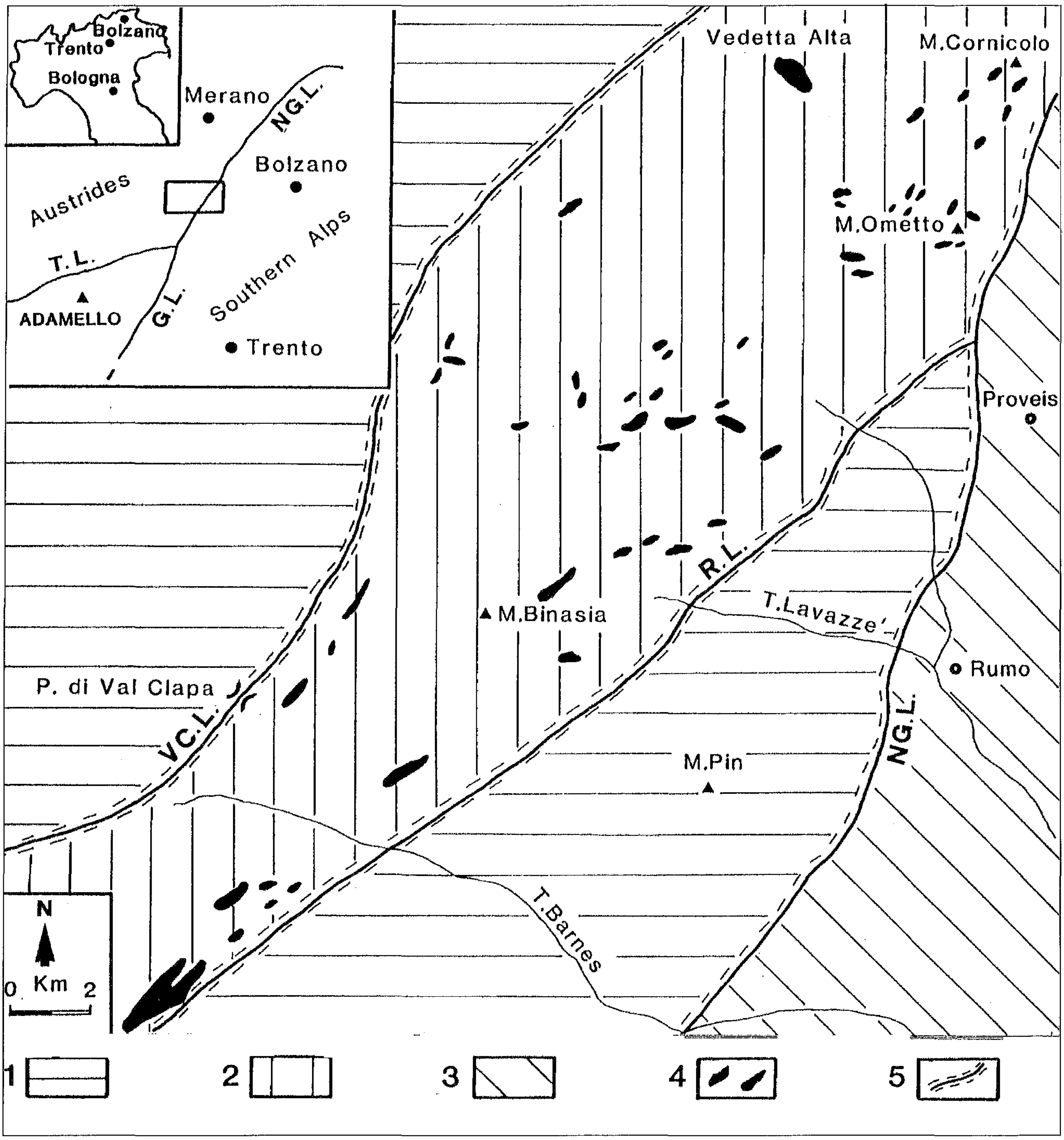

Fig. 1 - Geologic sketch map of the Nonsberg area. 1- biotite-muscovite micaschists and paragneisses; 2- garnet-kyanite paragneisses, migmatites and orthogneisses; 3- Southern Alps Mesozoic sedimentary rocks; 4- ultramafic rocks (some outcrops are not in scale and are exaggerated); 5- faults with associated cataclastic rocks; T.L.- Tonale Line; G.L.- Giudicarie Line; NG.L- North Giudicarie Line; R.L.- Rumo Line; VC.L.- Val Clapa Line.

are separated from the Mesozoic Southern Alps by the North Giudicarie Line (Linea delle Giudicarie Nord, Trevisan, 1939) which is part of the Insubric Lineament. The crystalline rocks have been subdivided into tectonic units and, among them, the unit bounded by the Rumo Line to the southeast (Morten et al., 1976) and by the Val Clapa Line to the northwest (Casolini, 1986) includes almost all the ultramafic bodies (Fig. 1). It is formed of garnet- and kyanite-bearing gneisses, migmatites and orthogneisses. The small and lens-shaped peridotitic bodies (300m x 100m average size) are generally concordant to the northeast-southwest directions of the main planar anisotropy of the surrounding gneisses. The contacts between the peridotites and country-rocks, where observable, are always sharp and at the contacts the peridotites are foliated and sometimes serpentinized.

The Nonsberg ultramafic rocks are predominantly peridotites of harzburgitic to lherzolitic compositions (Rost & Brenneis, 1978; Obata & Morten, 1987), while subordinate websterites, clinopyroxenites and dunites are present as small lenses or bands within the peridotitic bodies (Morten & Obata, 1983). On a textural and petrographic basis, they have been subdivided into two groups: coarse- and fine-type (Obata & Morten, 1987). The peridotites in the first group are coarse-grained with a protogranular texture (Mercier & Nicolas, 1975) and are formed of spinel-lherzolite. The second group consists of garnet-bearing and garnet-free peridotites which are more abundant and dominate the Nonsberg peridotites. The rocks are fine-grained and their texture is tabular or mosaic equigranular. They have been considered to be derived from the coarse-type peridotites by syntectonic recrystallization with the transformation from spinel to garnet peridotites occurring in the lower crust (Obata & Morten, 1987). The absence of clinopyroxene and garnet in many fine-type peridotites has been ascribed to the hydration reactions, in the CMASH system (Obata & Thompson, 1981):

cpx + opx + grt + H_2O $\rightarrow$ Ca-amph + ol

and at higher partial pressure of H_2O

grt + ol + H_2O $\rightarrow$ Ca-amph + opx + sp

Some samples have a porphyroclastic texture and they are considered transitional from the coarse- to the fine-type peridotites as documented by large garnet porphyroblasts surrounding spinel grains and by the large olivine and orthopyroxene porphyroclasts which occur in a fine-grained matrix.

Other samples are heavily serpentinized.

3. Geochemistry

Eighty samples subdivided on a petrographic and textural basis into four groups, i.e. coarse-type (13 samples), porphyroclastic (7 samples), serpentinized (5 samples) and fine-type (55 samples), were analysed for major elements, Cu, Sr and 3d transition elements by X-ray fluorescence analysis following the procedures of Franzini et al. (1972,1975) and Leoni & Saitta (1976). The loss on ignition at 1000°C was determined by gravimetric analysis.

3.1 *Major elements*

Table 1 shows the average, standard deviation and range for major elements and L.O.I. of the four groups of the Nonsberg peridotites. It emerges that few chemical differences exist for almost all the major elements between the two main groups of peridotites, i.e., coarse- and fine-type. Only the K_2O content is quite different, i.e., the average K_2O content of the fine-type peridotites is about three times that of the coarse-type peridotites. Furthermore the frequency per cent distribution of K is quite different

	coarse-type	porphyroclastic	serpentinized	fine-type
SiO_2	45.44 ± 1.41	45.31 ± 1.70	46.78 ± 0.78	45.34 ± 2.05
	46.95 - 41.59	46.91 - 42.57	47.97 - 45.85	54.11 - 39.31
TiO_2	0.04 ± 0.02	0.03 ± 0.01	0.03 ± 0.01	0.06 ± 0.09
	0.09 - 0.02	0.06 - 0.02	0.05 - 0.02	0.58 - 0.01
Al_2O_3	2.45 ± 0.90	2.42 ± 0.52	1.92 ± 0.76	2.80 ± 1.11
	4.03 - 1.06	3.16 - 1.68	2.79 - 0.81	7.60 - 1.53
FeO tot.	8.35 ± 0.51	8.59 ± 0.37	7.72 ± 0.93	8.17 ± 0.77
	9.23 - 7.41	9.08 - 7.98	8.87 - 6.39	10.97 - 5.68
MnO	0.13 ± 0.02	0.12 ± 0.01	0.10 ± 0.02	0.13 ± 0.02
	0.16 - 0.10	0.14 - 0.11	0.12 - 0.06	0.17 - 0.09
MgO	40.48 ± 1.83	40.94 ± 2.48	41.48 ± 2.01	40.38 ± 2.94
	43.82 - 37.33	44.31 - 36.75	44.44 - 38.97	44.87 - 28.69
CaO	2.31 ± 0.74	1.74 ± 0.59	1.28 ± 0.71	2.21 ± 0.82
	4.07 - 1.31	2.71 - 1.15	1.83 - 0.07	5.37 - 0.44
Na_2O	0.18 ± 0.03	0.13 ± 0.07	0.08 ± 0.03	0.20 ± 0.09
	0.24 - 0.12	0.26 - 0.04	0.13 - 0.05	0.53 - 0.01
K_2O	0.03 ± 0.01	0.05 ± 0.02	0.03 ± 0.02	0.10 ± 0.33
	0.05 - 0.01	0.08 - 0.03	0.04 - 0.01	2.52 - 0.02
L.O.I.	2.12 ± 1.44	4.62 ± 2.68	7.72 ± 3.44	3.92 ± 2.40
	4.85 - 0.62	8.44 - 2.09	13.62 - 5.48	11.76 - 0.59
MgO/FeO tot.	4.87 ± 0.37	4.78 ± 0.40	5.45 ± 0.88	4.99 ± 0.58
	5.47 - 4.33	5.37 - 4.05	6.96 - 4.67	6.43 - 2.61
The data are obtained from the analyses recalculated 100% on an anhydrous basis				

Tab. 1 - The data are obtained from the analyses recalculated 100% on an anhydrous basis

in the four groups of peridotites (Fig. 2). More in detail, the K distribution of the porphyroclastic peridotites is intermediate between those of the coarse- and fine-type peridotites, respectively. A phlogopite-bearing fine-type sample has a K_2O content as high as 2.52 wt%. The serpentinized rocks are depleted of alkalies, FeOtot, CaO and Al_2O_3 and slightly enriched in MgO in comparison to the other peridotites. Figures 3a-d show the variations of some elements, e.g. SiO_2, MgO, FeOtot and Na_2O+K_2O, against the MgO/FeOtot ratio taken as a reliable index of the degree of fractionation or of partial melting. The majority of the Nonsberg peridotites cluster around 4.5 to 5.5 for the MgO/FeOtot ratio. It is worth noting that some estimates of undepleted upper mantle compositions (Carter,1970;

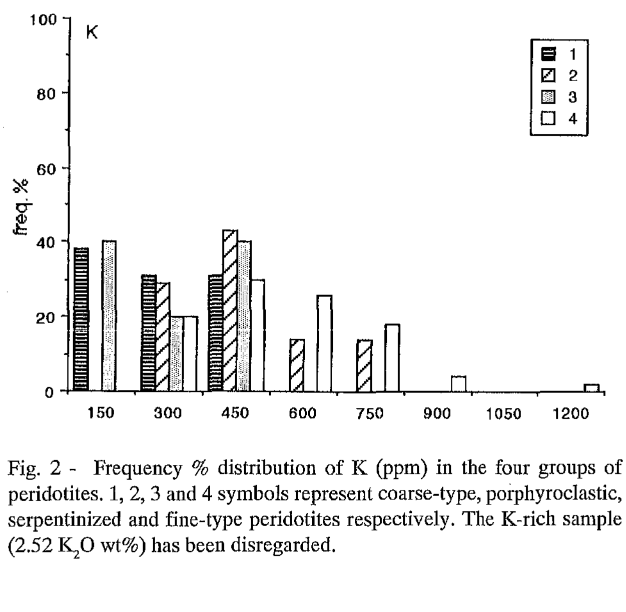

Fig. 2 - Frequency % distribution of K (ppm) in the four groups of peridotites. 1, 2, 3 and 4 symbols represent coarse-type, porphyroclastic, serpentinized and fine-type peridotites respectively. The K-rich sample (2.52 K_2O wt%) has been disregarded.

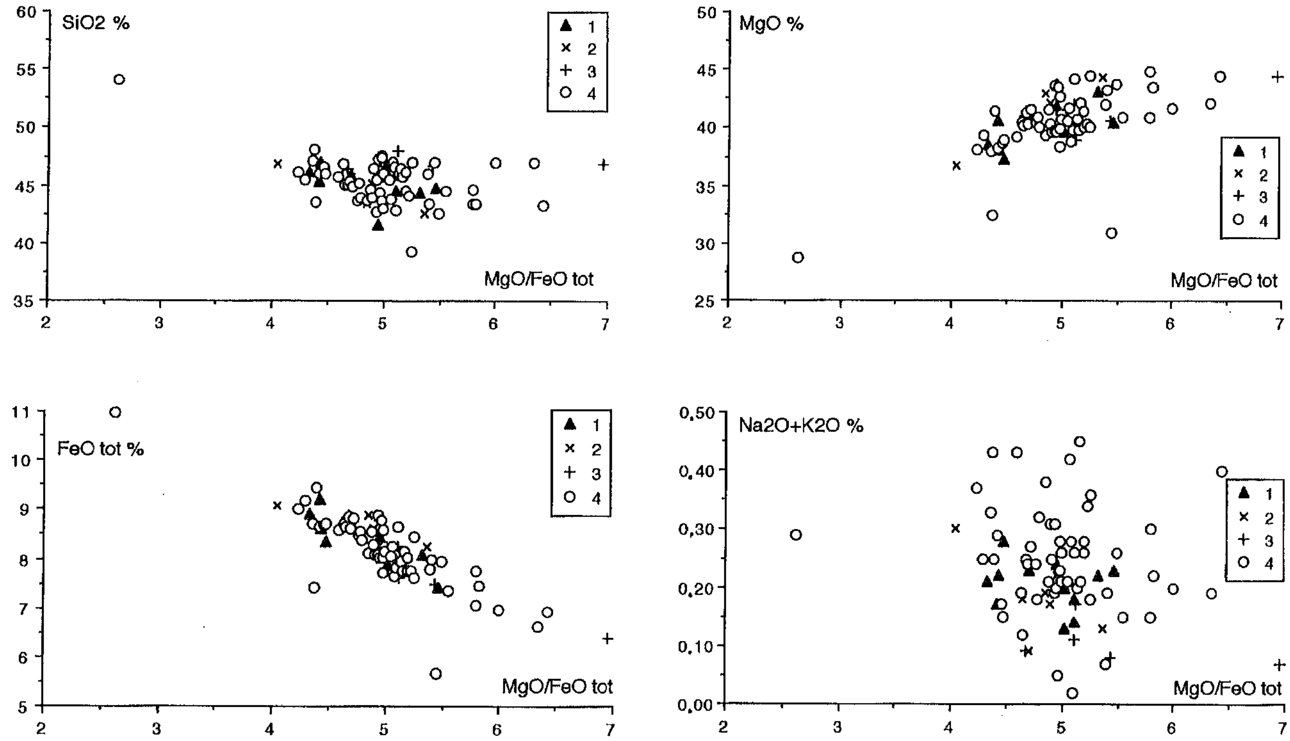

Fig. 3a-d - Weight per cent of oxide components vs. MgO/FeOtot ratio in the Nonsberg peridotites. 1, 2, 3 and 4 symbols represent coarse-type, porphyroclastic,serpentinized and fine-type peridotites, respectively.

Maaløe & Aoki,1977; Ringwood,1975; Jagoutz et al.,1979) are in the same range as regards the MgO/FeOtot ratio. The trends observed (Figs. 3a, b and c) for both the coarse- and fine-type peridotites represent the trends expected from residua after variable degrees of melt extraction. A further confirmation of this is given by the positive correlation between CaO and Al_2O_3 (Fig. 4). Conversely, the alkalies distribution does not show any significant correlation with the MgO/FeOtot ratio (Fig. 3d) thus implying alkalies perturbations.

	coarse-type	porphyroclastic	serpentinized	fine-type
Sc	11 ± 4	9 ± 3	9 ± 4	11 ± 3
	15 - 5	14 - 6	15 - 6	16 - 6
V	69 ± 22	54 ± 11	58 ± 8	64 ± 17
	107 - 38	66 - 39	70 - 50	117 - 20
Cr	3047 ± 490	2806 ± 709	3164 ± 363	2719 ± 574
	3967 - 2169	3721 - 1757	3475 - 2579	3568 - 723
Co	110 ± 7	103 ± 6	103 ± 10	106 ± 10
	123 - 98	109 - 96	112 - 87	121 - 65
Ni	2228 ± 152	2157 ± 94	2218 ± 67	2156 ± 210
	2499 - 2013	2271 - 1971	2273 - 2102	2695 - 1356
Cu	12 ± 3	11 ± 2	10 ± 2	13 ± 4
	15 - 8	15 - 9	13 - 9	30 - 8
Sr	8 ± 5	11 ± 10	6 ± 3	22 ± 15
	20 - 2	33 - 3	9 - 1	95 - 4

Tab. 2 - Average, standard deviation (±1s) and range for the minor elements of the Nonsberg peridotites.

3.2 *Minor elements*

Table 2 shows the average, standard deviation and range for some minor elements not corrected for L.O.I., of the four groups of the Nonsberg peridotites. No significant differences exist between the four groups of peridotites as regards the Sc, V, Cr, Co and Ni contents. Nevertheless the Cr content is slightly higher in coarse-type than in fine-type peridotites. The abundances of the 3d transition elements in the Nonsberg peridotites were normalized to primitive mantle estimates (Jagoutz et al.,1979) and plotted against element atomic number in figures 5a-d. Except for a fine-type sample, all the peridotites have more or less flat

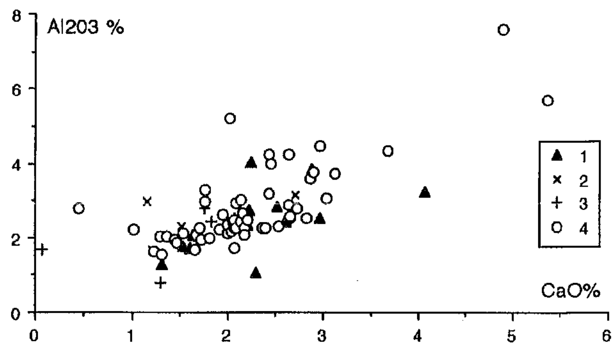

Fig.4 - CaO vs. Al_2O_3 diagram of the Nonsberg peridotites. Symbols as in figure 3.

patterns from V to Ni (Fig.5a-d) and show a distinct depletion in Ti with the exception of two fine-type samples whose patterns show a positive Ti anomaly (Fig. 5d). The general Ti depletion in comparison to the mantle source suggests a residual character for the Nonsberg peridotites due to the solid/liquid partition coefficient of Ti during partial melting of peridotitic material. The residual character is also supported by the trend of Sc depletion in the Ni-Co-Sc diagram (Fig. 6) since, as discussed by Ottonello et al. (1984), fractional and equilibrium melting of a peridotitic source drives the residua towards the Sc-poor region along essentially the same trend at a roughly constant Ni/Co ratio. The slight negative Mn anomaly in the serpentinized samples (Fig. 5c) is possibly due to Mn mobilization during hydrous alteration under moderate to low Eh and pH.

The Cu contents in the four groups of peridotites (including the serpentinized ones) are similar in spite of the well-known mobility of Cu during hydrous alteration at certain Eh and pH values. On the contrary, the Sr contents of the two main groups of peridotites, i.e. coarse- and fine-type, are significantly different and the latter group has on the average about three times the Sr content of the former. Figure 7 indicates that the Sr distribution of the coarse-type peridotites clearly differs from that of the fine-type peridotites. Furthermore the Sr distribution in the porphyroclastic peridotites is similar to that of the coarse-type but shifted towards higher Sr contents.

4. Concluding Remarks

The geochemical study of a large number of peridotites subdivided on a petrographic and textural basis into four groups, shows that they are indistinguishable as far as almost all their major elements, Cu and 3d transition elements. The geochemical variations and trends observed

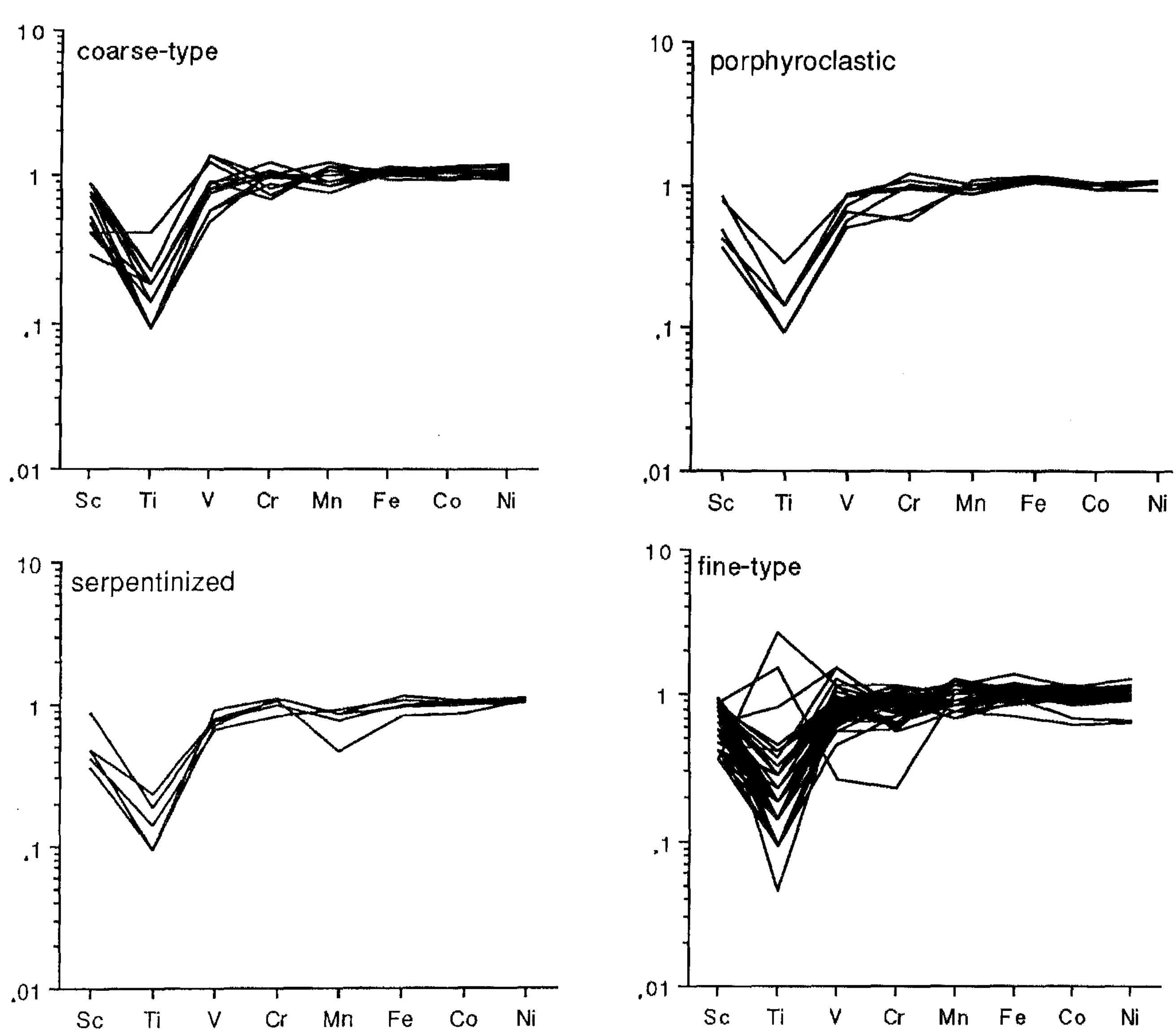

Fig.5a-d - Levels of 3d transition elements in the Nonsberg peridotites normalized to primitive mantle abundances (Jagoutz et al., 1979).

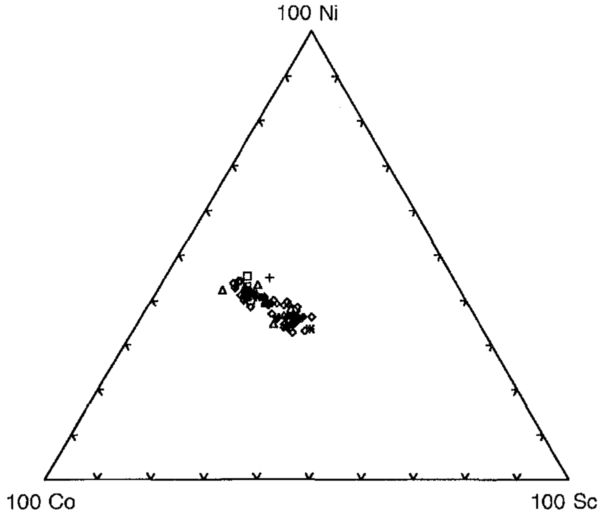

Fig. 6 - Ni-Co-Sc diagram for the Nonsberg coarse-type (open triangles), porphyroclastic (open square), serpentinized (crosses) and fine-type (diamonds) peridotites. Asterisk: primitive mantle source. The sample are plotted after normalization to primitive mantle abundances (Jagoutz et al., 1979).

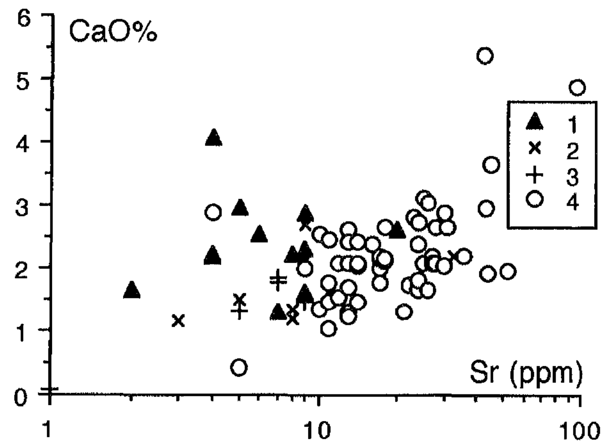

Fig.8 - Sr vs. CaO diagram of the Nonsberg peridotites. Symbols as in figure 3.

are explained as due to different degrees of fertility or better residuality of the protoliths (Figs. 4 and 6). This agrees with the isotopic depleted signature ($^{143}Nd/^{144}Nd$ from 0.51315 to 0.51280) of some coarse- and fine-type peridotites (Petrini & Morten,1990). On the other hand the behaviour of alkalies, but mainly of K_2O, does not support the above view. The scattering of alkalies (Fig. 3d) and the frequency per cent distribution of K (Fig. 2) in the four groups of peridotites suggest a metasomatic event. It should be noted that the K contents increase systematically from the coarse-type to fine-type peridotites via the porphyroclastic ones. This indicates a relationship between K content and degree of recrystallization.

The abundances and frequency per cent distribution of Sr (Table 2, Fig.7) indicate an enrichment from coarse-type to fine-type via porphyroclastic peridotites similar to the K behaviour. Conversely the average Sr content of the coarse-type peridotites (8 ppm, Table 2) is comparable to that (12 ppm) of six "primitive lherzolite xenoliths" by Jagoutz et al. (1979). On the other hand the higher Sr content of the fine-type peridotites cannot be correlated to a different character of the protoliths. In fact figure 8 shows that Sr is not positively correlated, as expected, with CaO but, at a more or less constant amount of CaO, the Sr content varies considerably, e.g. at about 2 wt% CaO, Sr ranges from 9 to about 50 ppm. Furthermore, Petrini & Morten (1990) have noted that the Sr isotopic composition of the Nonsberg peridotites are roughly correlated with their alkalies content and they hypothesized that such a relationship indicates a different degree of the alteration processes and not a different character of the mantle source. The same rough correlation exists between Sr and alkalies contents (Fig. 9). It is possible to suggest that the enrichment of Sr and alkalies, mainly K_2O, is related to the syntectonic recrystallization of the peridotites and possibly caused by the inflow from

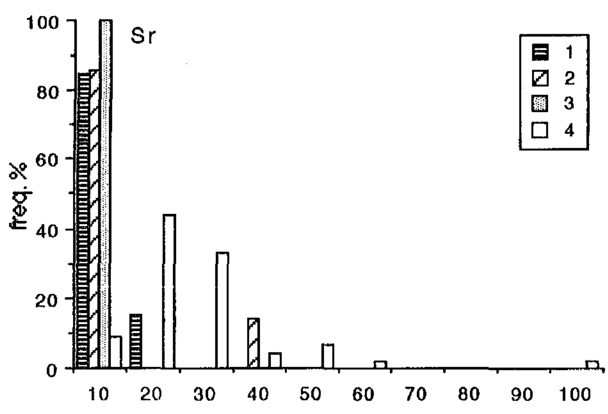

Fig.7. Frequency % distribution of Sr (ppm) in the four groups of peridotites. 1, 2, 3 and 4 symbols represent coarse-type, porphyroclastic, serpentinized and fine-type peridotites respectively.

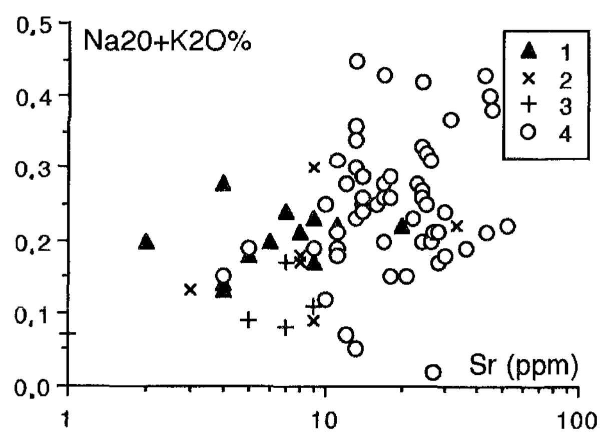

Fig. 9. Sr vs. (Na_2O+K_2O) diagram of the Nonsberg peridotites. Symbols as in figure 3.

the country-rocks of H_2O necessary to give the amphibole-forming hydrous reactions. Likely, the H_2O released during the recrystallization of the surrounding gneisses acted as a carrier for Sr and alkalies mobilized during the same metamorphic crustal event. In conclusion, the Sr and alkalies enrichments in the Nonsberg peridotites happened during the syntectonic recrystallization and transformation, at lower crustal conditions, from spinel-lherzolite (coarse-type) to garnet-bearing and garnet-free peridotites (fine-type). This event did not change and influence the other major elements, Cu and 3d transition elements.

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OCCURRENCE OF METAVOLCANICS IN THE SOUTHERNMOST DORA-MAIRA MASSIF (ITALIAN WESTERN ALPS)

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Abstract

The Dora-Maira Massif (Penninic Domain) is well known as the host of coesite-pyrope assemblages formed under very-high- pressure conditions. It is a portion of continental crust composed of (1) a pre-Variscan, polymetamorphic, mainly metapelitic, basement, (2) probably Permian-Carboniferous, monometamorphic, detrital covers, (3) acidic to intermediate metaintrusives traditionally related to the late-Variscan magmatic cycle, and (4) Mesozoic carbonates.

In the southernmost part of the Massif (Maira Valley), the occurrence of layers (a few decimeters to several meters in thickness) of leucocratic gneisses, with phenocrysts and fragments of K-feldspar embedded in a foliated groundmass, in a sequence of monometamorphic metasediments (chloritoid ± garnet ± alkali amphibole micaschists, fine-grained gneiss, tourmaline- bearing, quartz-micaschists and micaceous quartzite), strongly suggests an origin from tuff and/or clastic sediments derived from erosion of acidic lava. Support for this diagnosis is provided by zircon typology.

Furthermore, a basic volcanism, probably connected with early tensional tectonics following the end of the Variscan cycle, has been responsible for limited occurrence of fine- grained, porphyritic metabasics (already known from the same sequence).

This finding supports the interpretation of the Dora-Maira Massif as a composite unite in keeping the peculiar PT path proved by the very-high-pressure assemblages occurring only in the southern sector.

Key words: *Western Alps, Dora-Maira Massif, monometamorphic covers, metavolcanics, Alpine metamorphism.*

1. Introduction

Over the last ten years, the Dora-Maira Massif (Penninic Domain) has been in the limelight since the report of coesite within pyrope in early-Alpine, high-pressure assemblages, indicative of very-high-pressure (> 28 kbar) metamorphic conditions (Chopin, 1984; Chopin et al., 1991 with references), and thus a metamorphic evolution unlike that of the other "internal" massifs, i.e. Gran Paradiso and Monte Rosa (Dal Piaz & Lombardo, 1986). The coesite bearing rocks occur nevertheless only in a small area straddling the Po and Varaita valleys near the southern end of the Massif.

The Massif is composed of (1) a pre-Variscan, polymetamorphic, mainly metapelitic, basement, (2) monometamorphic (mainly detrital) covers referred to Permian- Carboniferous age (Vialon, 1966; Michard, 1967; Sandrone et al., in press), (3) acidic to intermediate metaintrusives traditionally related to the late-Variscan magmatic cycle (Sandrone et al., 1988; Sandrone et al., in press), and (4) a Mesozoic, mainly carbonate cover.

In the whole Massif the pre-Alpine continental crust was heavily modified by a strong tectonic and metamorphic Alpine overprint with initial eclogite and subsequent green schist metamorphic condition (see e.g. Sandrone et al., in press with references).

The nature, protoliths, and age of part of the monometamorphic cover are controversial (Sandrone et al., in press). According to Vialon (1966), the sequence of leucocratic augen- and micro-augen-gneiss, aplitic gneiss and "micascisti argentei" (phengitic schists) known as "Pietra di Luserna" (Sandrone et al., 1982), was produced by metamorphic alteration of a Permian, acidic, volcanic-detrital complex. However, fully convincing evidence for the presence of metavolcanics has never been put forward, and this has led other authors (Bortolami & dal Piaz, 1970; Barisone et al., 1979; Sandrone et al., 1982; Cadoppi, 1990) to regard these rocks as metagranite.

Rocks of volcanic, or more probably volcanic-detrital origin are here reported from the Maira Valley (southernmost part of the Massif).

2. The Metavolcanics

2.1 *The acidic rocks*

Gneisses derived from a recognisable volcanic protolith occur near the Santa Maria and San-Costanzo-al-Monte sanctuaries (Dronero district, Maira Valley) as intercalations a few dm to a few m in thickness within a metasedimentary

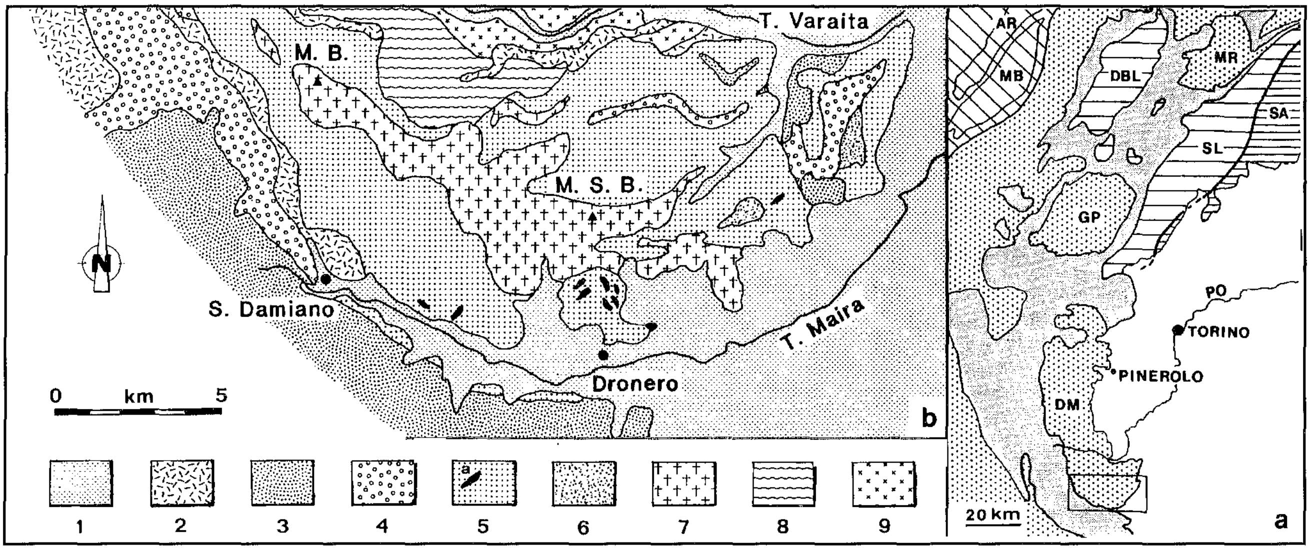

Fig. 1 - a) Simplified tectonic map of Western Alps. Oblique ruling: Helvetic domain (AR = Aiguilles Rouges; MB = Monte Bianco). Dotted area: Penninic basement (MR = Monte Rosa, GP = Gran Paradiso, DM = Dora-Maira massifs). Grey: Piemonte Ophiolite Nappe. Spaced horizontal ruling: Australpine system (DBL = Dent Blanche Nappe; SL = Sesia Lanzo Zone). Dense horizontal ruling: Southern Alps (SA). The boundary of the geological sketch map (b) is indicated.

b) Geological sketch map of the southernmost part of the Dora- Maira Massif (south of the "very-HP zone" of Chopin et al., 1991) (from Sandrone et al., in press, modified). 1: Quaternary deposits; 2: Mesozoic Schistes Lustrés and ophiolites (Piemonte domain); 3: Mesozoic cover (marble and metadolomite); 4: phengite-quartzite (± conglomerate) grading into quartzitic micaschist and phengite schist ("Ensemble de Sampeyre" p.p., Vialon, 1966) (Permo-Trias?); 5: Monometamorphic chloritoid ± garnet micaschist, fine-grained gneiss, micaceous quartzite, thin layers of tourmaline-rich micaschist, with intercalation of detrital acidic metavolcanics and (a) layers of porphyritic metabasics (Permian ?); 6: graphite-bearing micaschist (Complesso Grafitico del Pinerolese, *auct.*) (Carboniferous?); 7: Porphyritic metagranite, coarse augengneiss, flaggy leucocratic, fine-grained gneiss with thin intercalations of "silvery micaschist" ("porphyroides arcosiques" after Vialon, 1966); 8: polymetamorphic garnet-chloritoid ± glaucophane micaschist with relics (garnet, staurolite and muscovite; see Chopin et al., 1991) of a medium grade, pre-Alpine assemblage; 9: polymetamorphic gneiss with large, polycrystalline nodules (mainly albite and relics of K-feldspar) ("gneiss amygdalaires", Vialon, 1966). M.B. = Monte Birrone; M.S.B. = Monte San Bernardo.

sequence (unit n° 2 here above) (fig. 1). The latter includes chloritoid- garnet- micaschist locally rich in opaque minerals, quartz-micaschist, micaceous quartzite and albitized micaschist ("gneiss minuti" auct.).

The whole sequence displays a marked foliation which has transposed the primary fabric; it also includes rocks substantially composed of tourmaline with subordinate quartz, white mica and accessories (ore mineral, apatite) forming nodules and thin horizons (up to a few dm thick), generally near the contact of tectonically overlying metagranitoids (fig. 1).

The best evidence of a volcanic or volcanic-detrital origin is preserved by some leucocratic gneiss with K-feldspar (now, microcline) porphyroclasts 0.1 mm to a few mm in size displaying a twinning typical of high-T feldspar (sanidine?). The matrix, completely recrystallized by Alpine metamorphism, is made of quartz, phengite (Si = 3.45 p.f.u.), albite and accessory minerals, such as epidote, apatite, opaque minerals and zircon.

A study of zircons based on the method of Pupin (1976) was performed to clarify the nature of the protoliths. The typology distribution of a population of 140 zircons (fig. 2) displays a conspicuous presence of high-T (S_{24}, $S_{19\text{-}20}$) types compatible with a volcanic rather than granitic origin. The crystals, clear and colourless, lacking overgrowths, occasionally display "canalicular" inclusions typical of crystallization in a volcanic environment. It should be stressed that in leucogranite and highly differentiated granite, high-T types are generally scarce to absent; see e.g. the metagranite and phengite gneiss of Susa Valley in the northernmost part of the same massif, described

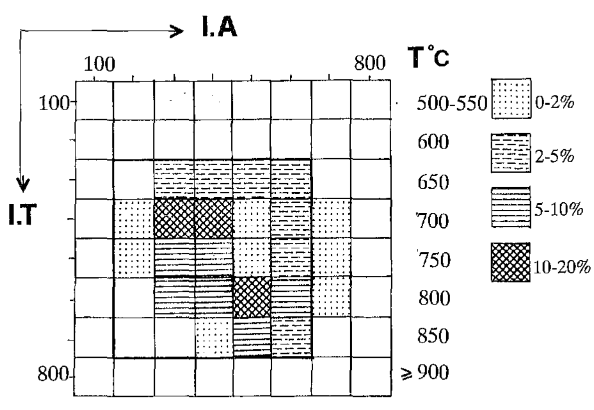

Fig. 2 - Typologic frequency distribution of zircon population from an acidic volcano-detrital rock in the I.A vs. I.T diagram of Pupin (1976); the indices I.A and I.T represent relative developments of pyramids ({101} and {211}) and prismes ({100} and {110}) respectively, related by Pupin (1980) to chemical composition of the magma (I.A) and to temperature (I.T) prevailing during zircon crystallization.

by Cadoppi (1990). In the sequence described here, however, some rocks also carry low-T (S_{2-5}) zircon types. The implication is that we are dealing with metasediments that drew detritus from a basin containing both volcanics and granite. The characters of the high-T zircons are suggestive of derivation from calc- alkaline to sub-alkaline magma, though a trend cannot be recognized owing to the the detrital nature of most of the rocks.

2.2 *The basic rocks*

In the studied area , mafic magmatic rocks forming thin intercalations (up to a few m thick) also occur within the metapelites. Their original magmatic fabric is still recognisable in spite of penetrative foliation and a new, metamorphic mineral assemblage. Some relic porphyritic textures with a fine-grained groundmass are indicative of dike- or subvolcanic bodies. Heterogeneous rocks with epidote nodules are reminiscent of a tufaceous protolith. The phenocrysts of pyroxene and\or amphibole and of plagioclase are completely replaced by aggregates of epidote + glaucophane ± white mica and by epidote ± white mica respectively. The groundmass is mainly epidote + albite ± glaucophane (and\or a greenish blue amphibole) + sphene + ore mineral. White mica and chlorite are present in the most foliated rocks. Garnet, when present, forms minute idioblasts in the groundmass. It is essentially almandine with ±20% grossular, ±10% spessartine and <10% andradite + pyrope. The presence of glaucophane in the metamorphic assemblage and the absence of a sodic pyroxene suggest that eclogitic conditions were not attained.

The metabasites here dealt with are thus different from those described further to the north (see e.g. Pognante & Sandrone, 1989; Cadoppi, 1990; Borghi et al., 1991). In fact, they display relic volcanic textures (whereas the others do not); they lack relic eclogite assemblages (whereas the others do not); lastly, they occur within a "cover" sequence (and not within polymetamorphic "basement"). Most of the thin metabasite intercalations within the "basement" probably derive from amphibolites representing the products of a pre-Variscan magmatic event (Sandrone et al., in press).

3. Summary and Conclusion

A few points should be stressed.

1) There are previous, presumptive reports of volcanics from the Dora-Maira Massif, but the occurrence here described is the first well-documented observation. No such evidence has been reported from the northern Dora-Maira, where the existence of volcanic protoliths was suggested by Vialon (1966) mainly on the ground of corroded quartz grains (besides some general stratigraphic considerations). As to the presence of abundant tourmaline forming thin intercalations in a metasedimentary sequence, it might be primary and due to circulation of boron- rich fluids derived from volcanic and\or hydrothermal processes in the sediment pile (Slack et al., 1984).

2) The volcanic protoliths strengthen the idea of a stratigraphic difference of the southernmost Dora-Maira massif, which is one outcome of our surveys now in progress. This has a bearing on the fact that this sector hosts the very-high-pressure assemblages. It also supports the hypothesis that the "massif" is tectonically composite (Chopin et al., 1991).

3) The volcanism that yielded the protoliths is fully in keeping with the tensional phases that closed the Variscan cycle and were the forerunners of the opening of the Piedmontese ocean. In the "internal" massifs, reports of volcanic protoliths are scanty, the most documented being that of Bertrand (1968) (from the Bonneval "gneiss minuti", western Gran Paradiso). Little textural and mineralogical evidence has presumably survived the Alpine metamorphism. It can thus be observed only in exceptionally favourable circumstances.

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LATE VARISCAN INTERMEDIATE VOLCANISM IN THE LIGURIAN ALPS.

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Abstract

The geological setting and the petrographic and chemical characters of the following groups of the Late Variscan intermediate volcanics have been studied :

— metavolcanics interbedded within Permian-Carboniferous metasediments from inner to outer Ligurian Briançonnais (Eze volcanics, E.V.) and from outermost Briançonnais (Aimoni volcanics, A.V.);

— dykes and volcanics crosscutting or covering the pre-Namurian basement in the Calizzano (C.V.) and Savona (S.V.) innermost Briançonnais "massifs";

— volcanic clasts from Briançonnais Upper Carboniferous conglomerates (O.V.), from Lower-Middle Jurassic Prepiedmont M.Galero Breccias (G.V.), from Upper Jurassic outer Piedmont Montaldo Breccias (M.V.).

A number of evidences and arguments indicate that the intermediates volcanic activity began soon after the main Variscan orogenic events (S.V. and O.V.), probably during Namurian-Early Westphalian times, and that it lasted at least until Stephanian-Autunian times (A.V. and possibly, G.V. and M.V.).

Intermediate minor lavas and dominant, usually fine-grained, pyroclastites were poured out, ranging in composition from rare basaltic andesites (E.V., S.V.), through commoner andesites (E.V., C.V.), up to trachy-andesites and dacites (G.V.). They have a subalkaline character with calcalkaline affinity, except for G.V., showing a shoshonitic affinity.

A parent magma originating from partial melting of a possibly modified upper mantle and suffering later contamination with crustal materials, fits with petrographic and chemical data. These indicate that melts were generated under a thick crust.

The geological setting evidences that the outpouring was favoured by a subsequent extensional regime, a possible forerunner of the crustal attenuation preceding and accompanying the Mesozoic pre-Tethyan rifting.

Finally, disregarding orogenic and/or subductive models, which have no geological support inside the region, the shoshonitic affinity of G.V., compared with the low-K to high-K andesitic character of E.V., has its counterpart in the geologic frame, which may indicate that the Prepiedmont inner sector behaved at first as a more stable crustal block, where volcanic and tensional tectonic activity developed later.

Key words :*Ligurian Alps, Permo-Carboniferous, andesites, WR analyses, clinopyroxenes.*

1. Introduction

In an earlier note (Cortesogno et al., 1983) intermediate volcanics (Eze Formation) interbedded within fine detrital Permian - Carboniferous sediments, belonging to the Ligurian Briançonnais domain were described. Data known at that time seemed to indicate that Eze Formation was present only in the inner -and possibly intermediate- sector of the Briançonnais Zone. Later findings (Cabella et al., 1988) proved that it crops out also in the outer sector.

As its increased areal distribution appeared to stress the regional importance of this volcanic event, new field and laboratory work has been undertaken, also including various detrital Briançonnais and Prepiedmont formations containing -among others- clasts of intermediate volcanics.

Meanwhile, the general stratigraphic and structural field work on Briançonnais tectonic units has led to a less uncertain reconstruction of the paleogeographic situation of the different sampled outcrops of Eze Formation.

Thus the availability of an increased number of stratigraphic, petrographic and chemical data, coupled with a hopefully more reliable paleogeographic frame, have given us the tools for reconsidering this important Late Variscan event.

The purposes of this note are to provide direct and indirect documentation for the considerable time span and space width occupied by the intermediate volcanism, to underline its link with coeval tectonic activity, to give more details on its chemical characters and, finally, to discuss how and to what extent this data may fit into petrogenetic and geodynamic correlated models.

To avoid useless repetitions, readers interested in the general geology of the Ligurian Alps are referred to

Vanossi et al. (1986) and Vanossi (1991); a more specific presentation of the Permian-Carboniferous events is to be found in Cortesogno et al. (1988 a).

2. The Geological and Petrographic Frame

The Permian-Carboniferous volcanites are grouped on the grounds of their different provenances:

I) Eze metavolcanics (E.V.), interbedded within Permian - Carboniferous metasediments, represent the Eze Formation from inner to outer Briançonnais domain.

II) The Aimoni metavolcanics (A.V.) represent the Eze Formation in the outermost Briançonnais (Viozene area).

III) In the Calizzano "massif" the intermediate activity (C.V.) is represented by dykes crosscutting the pre-Namurian basement and by lavas and pyroclastites resting on it; also in the Savona "massif" dykes (S.V.) occur crosscutting a pre-Namurian basement.

IV) The M. Galero volcanics (G.V.) are found as clasts in the Lower-Middle Jurassic M.Galero Breccias (Prepiedmont Arnasco-Castelbianco nappe).

V) Volcanic clasts (O.V.) in Briançonnais Upper Carboniferous conglomerates (Ollano Formation).

VI) Volcanic clasts (M.V.) in Upper Jurassic syn-radiolarite breccias (outer Piedmont Montaldo tectonic Unit).

Chemical data concerning O.V. and M.V. are not available due to small size and alteration of the clasts.

The volcanics pertaining to the Eze Formation (E.V. and A.V.) are widespread and well represented in the region (fig. 1); their Permian-Carboniferous age and attribution to the Briançonnais domain are well documented (Cortesogno et al., 1983; Cabella et al., 1988). Only the proper paleogeographic relative location of some samples within the domain is questionable.

The volcanic clasts in the Ollano Formation are scarce; they testify a Late Variscan acid to intermediate volcanism likely developed in the Briançonnais domain.

C.V., S.V., G.V. and M.V. all together lack any direct relationship with Permian-Carboniferous sedimentary sequences and cannot be dated but indirectly. Moreover, the paleogeographic setting is inferred but not proved as they are known only from rootless tectonic Units.

2.1 *The Eze Formation (E.V., A.V.)*

The Eze Formation is represented by lenses of metavolcanics from parental lavas and pyroclastites interbedded within detrital sequences. In some cases deformation and metamorphic re-crystallization have completely obliterated primary textures; pyroclastic products are generally more sensitive to the textural "reconstitution".

The present size of the metavolcanic bodies (both from lavas and pyroclastites) ranges from some m^2 to many km^2 and thickness varies from some decimeter to some hundreds of m.

Pyroclastic materials commonly grade into mixed volcanogenic and sedimentary products.

Pyroclastites are generally fine or very fine-grained. Only in the Viozene area (A.V.) coarse-grained volcanic agglomerates, probably pyroclastic flows or lahars, are found (Cabella et al., 1988).

The associated metasediments (Murialdo Formation, Viola schists and Gorra schists) are all detrital with fine-grained primary textures; they were deposited in a continental environment. Heteropic relationships are probable. From stratigraphic position and inter-regional comparison, a Stephanian (-Autunian) age is inferred.

According to Vanossi et al. (1986) and Vanossi (1991), two main groups -lower and upper- of Briançonnais tectonic units can be distinguished (fig. 1).

The upper group is only formed by rootless klippen, whose paleogeographic pertinence is uncertain. The lower one consists of more or less widely superposed nappes, which, according to the present geometric relationships and to similar Alpine tectono-metamorphic history, do not seem to have lost their original proximity.

The Eze Formation only outcrops within the lower group of nappes. As the northern, inner ones are piled on the southern, outer thrust sheets, the paleogeographic location (outer, intermediate, inner) of each sample (tab. 1) has been roughly inferred from the relative geometric position of the nappes.

The schematic geological cross-sections (fig. 2) show, besides regional structure, the position of some of the sampled metavolcanic outcrops.

In the Eze Formation (E.V.) at least some primary textural characters are preserved from inner and outer Briançonnais zone, but for some outcrops in the internal Briançonnais (tab. 1). Volcanites are mainly andesites and minor basaltic andesites, commonly at the boundary with trachyandesites; also dacitic compositions are found (fig. 4 a, b).

The porphyritic textures, often seriate and glomeroporphyric, characterize all the lava flows. The P.I. varies from 5 to 20; plagioclase phenocrysts, always altered to albite $\pm$ epidote $\pm$ pumpellyite $\pm$ lawsonite, are ubiquitous and generally overwhelming.

The mafic phenocrysts are generally represented by

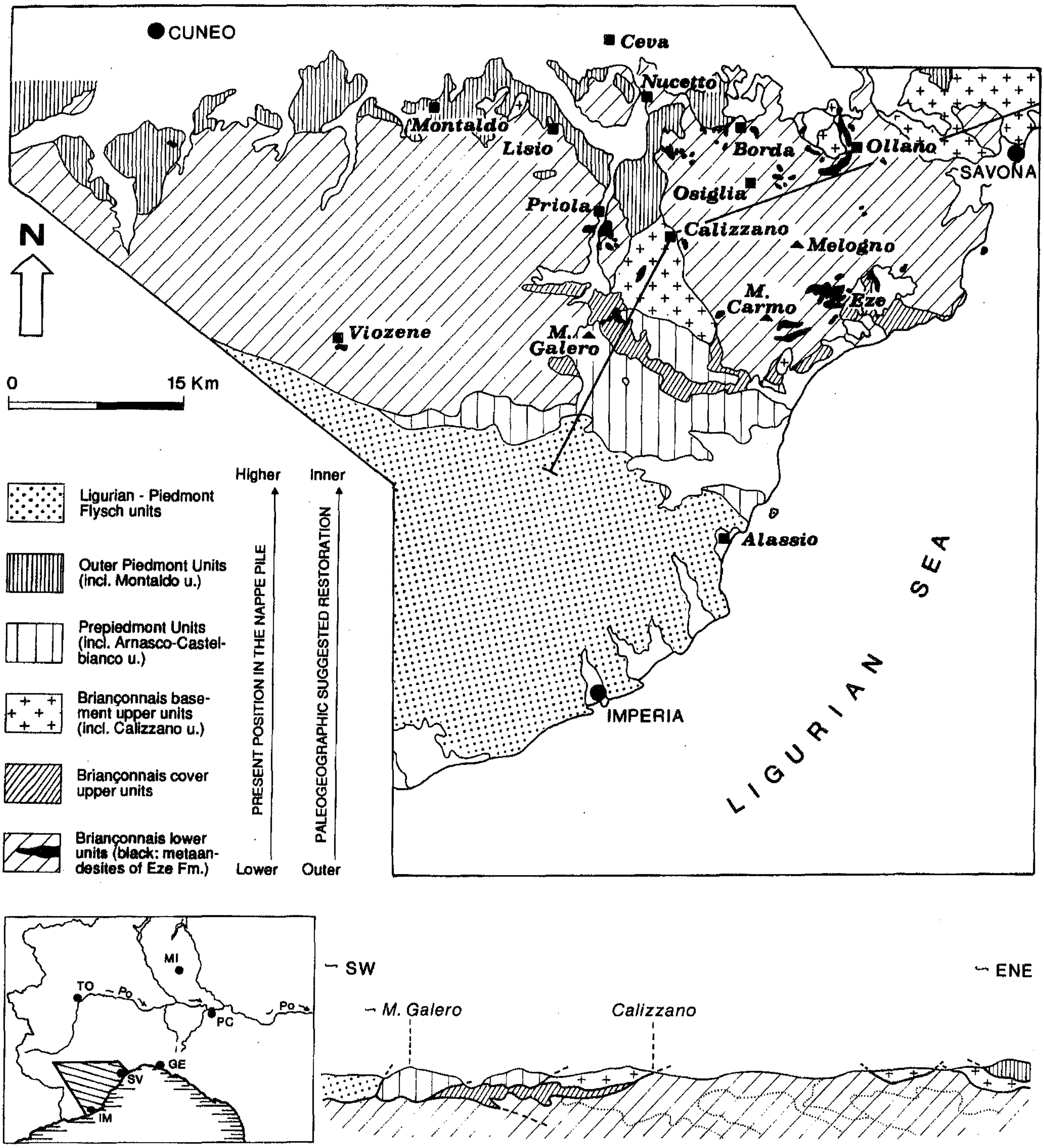


Fig. 1 - Schematic tectonic map and cross-section of the Ligurian Briançonnais and Piedmont Zones. White areas: Oligocene-Quaternary unconformable late and post-Alpine deposits.

brown hornblende; mainly in the eastern outcrops, large idiomorphic hornblendes are common, rarely including small plagioclase grains. The composition of unaltered grains ranges from edenitic hornblende to Mg hornblende.

Clinopyroxene is subordinate in andesites, but it is the only mafic phase in some basaltic andesites. The compositions generally plot in the augite field (fig. 3), but in basaltic andesites the compositions at the boundary salite-endiopside are also common. Compositions near the boundary of subcalcic augites are found at the rim of phenocrysts, possibly resulting from exsolved augite and pigeonite (Mellini et al., 1988). Pigeonitic compositions, however, are not found.

The orthopyroxene is rare, always rimmed by

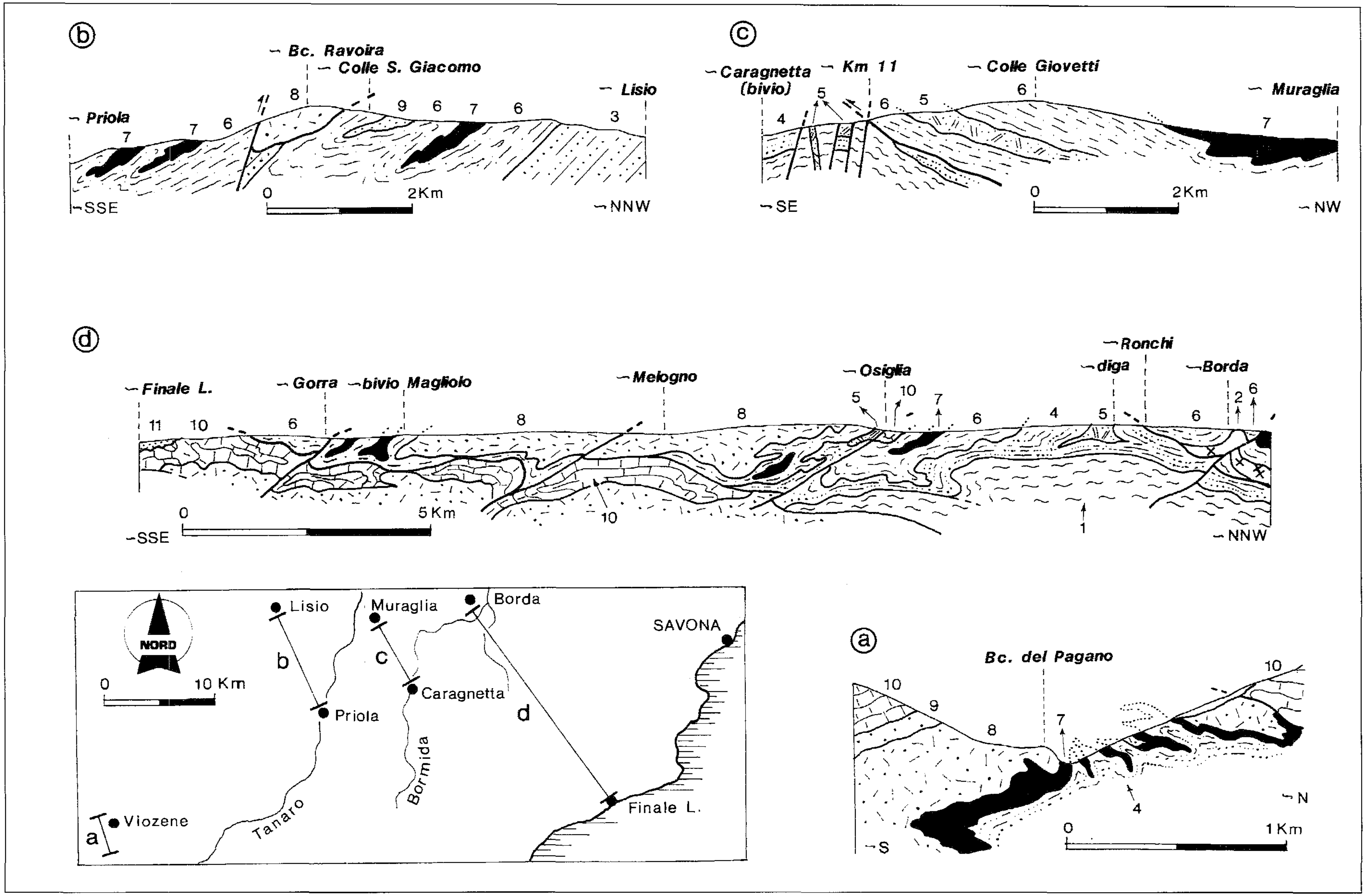


Fig. 2 - Schematic cross-sections across the lower Ligurian Briançonnais tectonic units, showing some of the late Variscan metaandesitic outcrops. 1: Pre-Namurian basement. 2: Late Variscan granodiorites. 3: Lower Late Carboniferous coarse continental metasediments (Lisio Formation). 4: Upper Carboniferous coarse to fine continental metasediments (Ollano Formation). 5: Upper Carboniferous metarhyolitic ignimbrites, agglomerates and dykes. 6: Stephanian-Autunian fine (Viola and Gorra Schists) to very fine, graphitic (Murialdo Formation) continental metasediments. 7: Stephanian-Autunian metaandesites (Eze Formation). 8: Lower Permian, mainly rhyolitic metaignimbrites (Melogno Porphyroids). 9: Upper Permian coarse continental metasediments (Verrucano). 10: Mesozoic covers. 11: Oligo-Miocene post-orogenic sediments.

	WEST		EAST
Piedmont domain			
Prepiedmont domain	*A* Galero breccias (G.V.)		
	B Calizzano (C.V.) Savona (S.V.)		
Briançonnais			
internal	*C*	*D*	*E*
		Eze volcanites (E.V.)	
intermediate	*F*	*G*	*H*
external	Aimoni volcanites (A.V.) *I*		

Analyzed samples

A	Galero breccias		2829/a, 2827/4, 2826/7, 2826/3, 2827/2, 2827/3 2838, 2824/2, 2826/1, 2826/5, 2835/3, 2826/8 49G, 49F, 3815b, 3815c, 3815d
B	Calizzano massif	dykes	*2815, 2817, 2819*
		lava flows	*2804b, 2806a, 2808a, 2809a, 2807a* 2813, 25717, 2695, 24715
	Savona massif	dykes	TBN8, TBN9, TBN17
C	Tetti Rosbella		2908
D	Muraglia,		*2924, 2512, 2514*
	Pamparato		*2512, 2514*
	Viola		*2913, 2915* 3759
E	Murialdo		3747,3746,GHI5,MU2,P211,P195,P193A,P192,GHI1,P180,P178
	Codevilla		*2951*
	Fornelli		*2953a, 2953b*
F	Colla Casotto		VC32, VC32'
G	Pianchiosso		*2920*
H	Feglino		*2957, 2958, 2960*
	Gorra		3814
	Eze		*2954*
	Tovo		GJ87,GJ142,GJ106,GJ24D,GJ128,GJ104
	Verzi		C9, C9b, C10, C11, C18, C19, C23, C24
I	Aimoni		3637,G4-6,3636,F4-31,3631,E3-7',E3-8,3632,3635,3627

Tab. 1 - Paleogeographic position of Permo-Carboniferous metavolcanics, distribution and frequency of analysed samples. Samples in italic from Cortesogno et al. (1983).

hornblende and altered to chlorite during Alpine metamorphism. Skeletal ilmenite and titanomagnetite are common in the groundmass and allanite sometimes occurs. Corroded quartz xenocrysts seldom occur in andesitic lavas. Pyroclastic levels are commonly enriched in tourmaline.

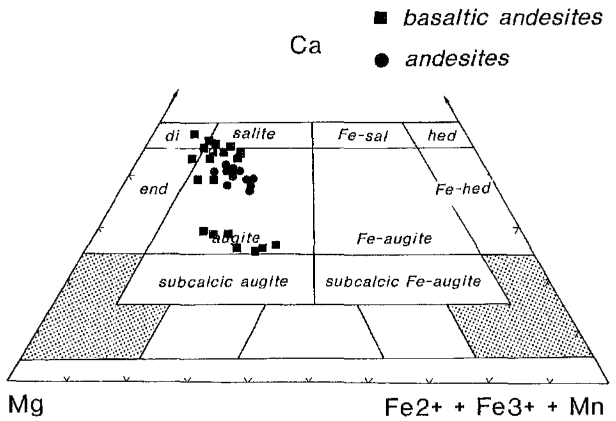


Fig. 3 - Composition of clinopyroxenes from basaltic andesites (full squares) and andesites (full circles).

The Eze Formation from Aimoni zone (A.V.) is mainly represented by volcanic breccias and pyroclastites, associated with minor lava flows of andesitic composition (fig. 4 a, b) and showing porphyritic seriate textures (P.I. 20-30).

Plagioclase and biotite are the prevailing phenocrysts, whereas hornblende is very scarce and pyroxene is lacking. Ilmenite, apatite, titanomagnetite and sometimes zircon are common accessory phases. Corroded quartz xenocrysts are widespread. Both plagioclase and biotite were completely altered during Alpine metamorphism.

2.2. *The Metavolcanics from the Pre-Namurian Basement*

2.2.1. *The Dykes and Volcanics from the Calizzano "Massif" (C.V.)*

The metavolcanics, andesitic to trachyandesitic in composition (fig. 4 a, b), occur as dykes crosscutting the pre-Variscan basement and as extrusives (lava flows and pyroclastites) directly overlying it.

Both dykes and extrusives show microporphyritic textures and low P.I. (5-20). Phenocrysts are plagioclase, augitic clinopyroxene and scarcer hornblende, biotite and orthopyroxene.

Abundant ilmenite and apatite are accessory phases. Corroded quartz xenocrysts are commonly found.

Analytical data on the primary mineral phases are not available due to the metamorphic alteration.

The C.V. have not been affected by Variscan metamorphism and dykes crosscut pre-Alpine metamorphic textures; as well as gneisses and amphibolites of the "massif",

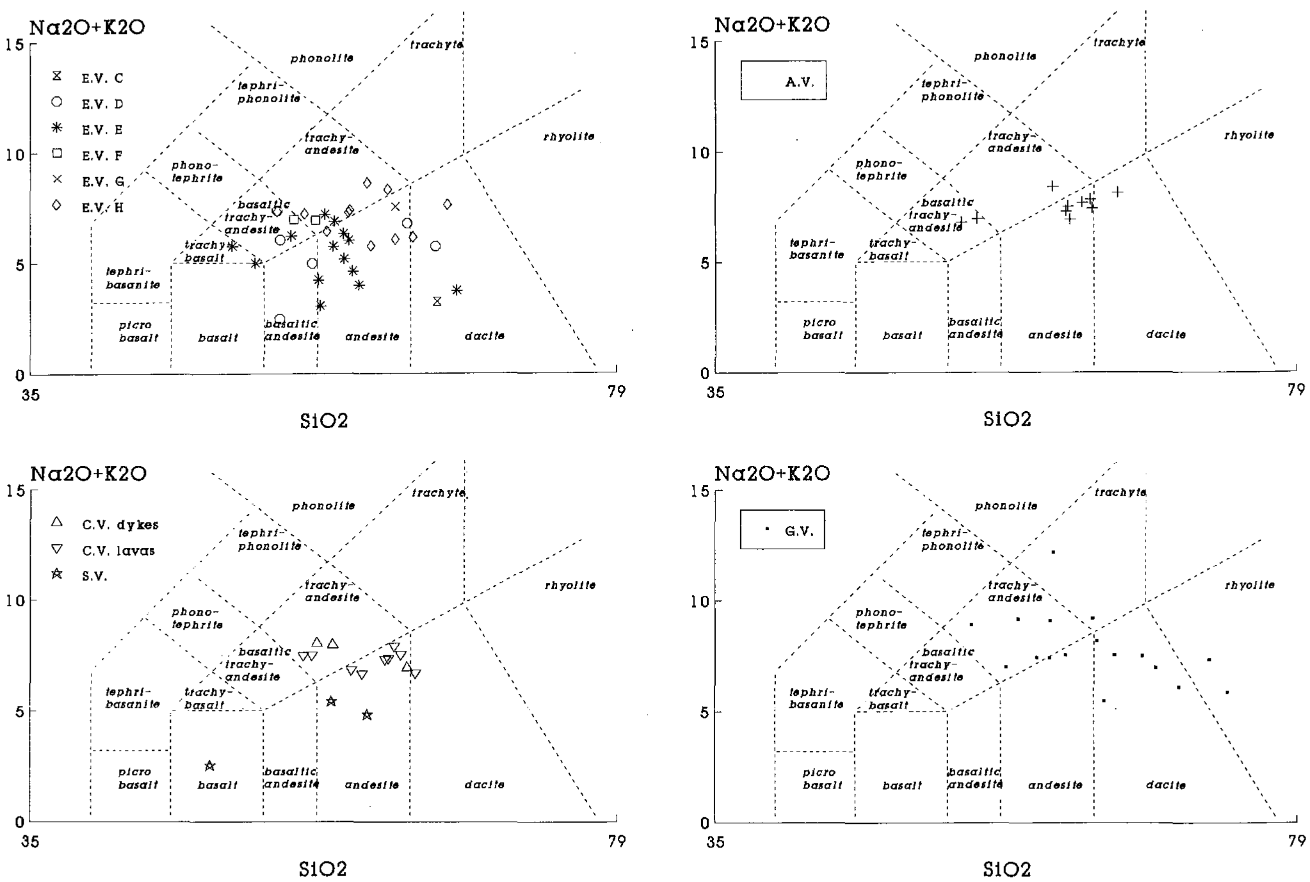


Fig. 4a - Total alkali/silica diagram (TAS; Le Bas et al., 1986) they underwent high pressure Alpine metamorphism.

Their age ranges therefore between Permian-Carboniferous and Paleocene times. Various arguments (Cortesogno et al., 1983) though not conclusive, led to choosing -among the different magmatic events that occurred in the Alpine region during that time span- a Permian-Carboniferous age: thus they should be more or less coeval with the Eze Formation.

In the nappe pile the Calizzano "massif" geometrically lies between lower Briançonnais and Prepiedmont tectonic units (fig. 1). According to Dallagiovanna (1988), the most probable paleogeographic restoration would be just at the boundary between these two domains.

2.2.2. *The Dykes from Savona "Massif" (S.V.)*

The Savona "massif" forms today an isolated outcrop of pre-Namurian basement (fig. 1). Its geometric position in the Alpine nappe pile and its polymetamorphic evolution are the same as those of the Calizzano "massif"; both "massifs" are therefore ascribed to the Calizzano-Savona nappe, which also include some other pre-Namurian slices and represents a Briançonnais basement upper unit.

Basic to intermediate dykes (fig. 4 a, b), where pseudomorphs are still recognizable after phenocrysts of plagioclase, pyroxene, hornblende and rarely olivine, occur crosscutting anatectic migmatites (M.te Cucco, NE of Savona). Textural evidences indicate that the emplacement of the dykes follows the main schistogenous Variscan event affecting migmatites; as a consequence the dykes are probably penecontemporaneous with the post-kinematic development of andalusite in migmatites and gneisses and become re-equilibrated during this late high temperature - low pressure metamorphic phase.

S.V. should likely be somewhat older than dykes in the Calizzano "massif" and possibly contemporaneous with the volcanic activity which produced O.V. (see par. 2.4.). Anyway, S.V. and O.V. evidence that early production of melts immediately follows the main Variscan orogenic events.

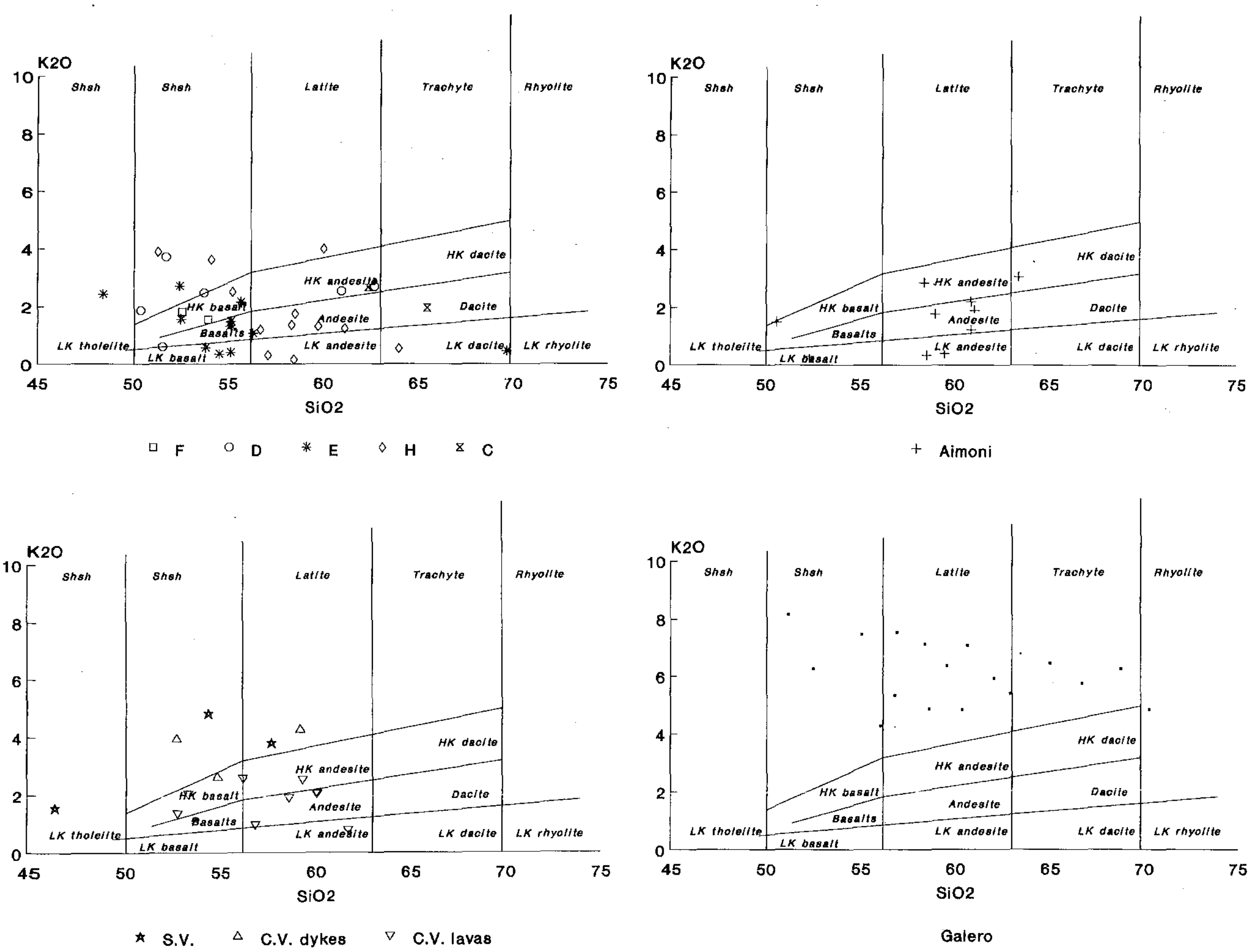


Fig. 4b - K_2O vs. SiO_2 (Peccerillo & Taylor, 1976) for C.V.+S.V., G.V., E.V. and A.V.

2.3. *The Clasts from the M.Galero Breccias (G.V.) in the Prepiedmont Arnasco-Castelbianco Nappe.*

The Arnasco-Castelbianco Unit is a rootless sheet lacking pre-Upper Triassic substratum; it is superposed on upper Briançonnais nappes in the southern part of the Ligurian chain (fig. 1).

Structural analysis (Dallagiovanna & Vanossi, 1983) shows that the unit was tectonically transported from inner sectors. Its stratigraphic sequence, from Norian dolostones to Upper Jurassic radiolarites, records the steps of the progressive sinking of a continental passive margin. For these reasons, a Prepiedmont paleogeographic pertinence is considered highly probable.

Huge volumes of breccias and mega-breccias (M. Galero Breccias) have been deposited during Early and Middle Jurassic times along sub-marine listric faults.

The lithic components come from older and older formations as sedimentation goes on: in the upper part of the Breccias Formation fragments and blocks from Permian-Carboniferous and from pre-Variscan rock bodies become prominent (Dallagiovanna & Lualdi, 1986).

The chaotic texture of the breccias, the angularity of the components, the absence of sorting and the abundance of matrix indicate that the source area was nearby: a Prepiedmont location, probably inner than the Calizzano "massif", may be inferred.

Among the clasts in the upper levels of the M. Galero Breccias, intermediate volcanites are common. Textures are porphyritic (P.I. 5-10) with phenocrysts of plagioclase and biotite and fringes of K-feldspar seldom rimming the plagioclase.

Corroded quartz phenocrysts are common and sometimes predominant in part of the samples. Alkaline plagioclase and K-feldspars are the main, still recognizable components in the microcrystalline mesostasis. Bulk-rock compositions range from trachyandesites to dacites (fig. 4 a, b), the highest SiO_2 contents mainly depending from quartz phenocrysts concentrations.

G.V. cannot be older than Permian-Carboniferous age as they did not suffer any Variscan metamorphism; on the other hand, they obviously cannot be younger than Early Jurassic age.

Moreover, volcanic fragments are missing in the lower levels of the Breccias, which rework Mesozoic, mostly carbonatic formations, while they become a frequent component of the upper Breccias reworking pre-Mesozoic formations; as a consequence, their Mesozoic age appears unlikely. Jurassic rhyodacites, coeval with the Breccias and often reworked immediately after their eruption, also occur (Cortesogno et al., 1981); they strongly differ from G.V. clasts in composition and in occurrence as clusters of soft pebbles or as matrix of the breccia. Thus, G.V. may be considered more or less contemporary with, but paleogeographically inner than, the Eze Formation.

2.4. *The Volcanic Clasts from the Ollano Formation (O.V.)*

The Ollano Formation (Late Westphalian-Stephanian) is made up of fluvial-lacustrine rhythmic detrital sequences.

They contain some intercalations of pyroclastic agglomerates ("Bric Crose Tuffs"), whose lithic components may be divided into: a) rhyolitic clasts as components of the tuffs (Cortesogno et al., 1983); b) polygenic fragments, torn away from the basement during the eruptions (Cortesogno et al., 1988 a).

Some of the polygenic fragments from middle Briançonnais sector (Cortesogno et al., 1988 b) are represented by intermediate, probably andesitic and dacitic volcanites showing porphyritic textures; phenocrysts are mainly represented by plagioclase and biotite.

As O.V. lack any pre-Alpine metamorphism, the parental andesitic lavas must have nearly the same age of the Ollano Formation, which is believed to be somewhat older than the Eze Formation.

These fragments, together with the dykes from the Savona "massif" (see par. 2.2.2.), therefore represent the oldest intermediate volcanic activity in the Ligurian Alps.

2.5. *The Volcanic Clasts from the outer Piedmont Montaldo Unit (M.V.)*

As well as the Arnasco-Castelbianco Unit, also the Montaldo Unit lacks any pre-Triassic substratum. It lies on Briançonnais nappes in the northern part of the Ligurian Alps (fig. 1).

Triassic shelf carbonates, followed by basinal limestones, then by Upper Jurassic radiolarites represent the stratigraphic sequence (Dallagiovanna & Vanossi 1988; Vanossi, 1991).

Some levels of polygenic breccias were deposited since the Lower Jurassic; the clasts are mainly of continental provenance, but serpentinitic and basaltic blocks from the nearby Piedmont-Ligurian oceanic domain are known in the middle and upper part of the sequence.

Sedimentary evolution and tectonic position match with a paleogeographic situation on the European passive continental margin, maybe somewhat nearer to the ocean than the Arnasco-Castelbianco Unit.

The polygenic levels of breccias alternated with radiolarites are very coarse and should then have been fed by an intra-basinal source.

Among the fragments and the blocks, pyroclastic metavolcanics of intermediate composition are not rare.

Petrographic analogies suggest a possible correlation with the Eze Formation; some monomineralic chloritic fragments, comparable with the Late Variscan chloritic dykes known in the French Western Alps, point to the same conclusion.

On the other hand, the global character of all the clasts found in the breccias (Cerro et al., 1975) do not allow a reliable and complete reconstruction of the stratigraphy of the source area, as it has been done with the M. Galero Breccias.

Thus, it can only be safely said that the M.V. might refer to Permian-Carboniferous volcanic activity located in a region corresponding to the innermost Briançonnais or - more probably- to the Piedmont domain.

3. Inferences from Geological Data

Two main conclusions can be drawn:

a) but for O.V. and S.V. (possibly as old as Namurian or Westphalian), the intermediate volcanism should have mainly a Stephanian (-Autunian) age.

b) the intermediate volcanism developed not only all over the Ligurian Briançonnais domain; it affected at least also the Prepiedmont and outer Piedmont sectors.

The outermost Briançonnais offers the best opportunities for the restoration: the Alpine tectono-metamorphic events were not severe and moreover, the Mesozoic cover is still adherent to the underlying Permian-Carboniferous tegument whose stratigraphy can thus be better established.

According to Cabella et al. (1988), thicknesses of late Variscan formations display both abrupt and gradual changes; these are related with graben and half-graben structures that were subsiding along conjugate sets of faults presently trending N30° and N110°. This model -similar to those suggested for various Late Variscan european regions- has been tentatively extended by present authors to the intermediate and inner Briançonnais sectors (fig. 5), though its reliability is lowered owing to the polyphasic Alpine tectonometamorphic events and lack of adherent Mesozoic covers.

According to the previous statements, the Calizzano "massif" has been located in the innermost sector. Obviously, it has not been possible to restore the Late Variscan stratigraphic section for the Piedmont domain. In the lower part of fig. 5, the different restored segments have been assembled, to show the subsidence rates in the Ligurian Briançonnais during Permian-Late Carboniferous times.

As the model suggests that volcanites and sediments mainly filled up depressions and did not create reliefs, the highest thicknesses coincide with the highest subsidence rates and, as a consequence, with the tectonically most active areas.

It follows that since Late Carboniferous times the innermost Briançonnais sector (between A and C in fig. 5) was mainly behaving as a source-area for the clastic deposits which were being sedimented in the outer, actively subsiding sectors (between D and G). The separation between outer and inner sector may have been sharp.

The transitional zone between the Briançonnais and Prepiedmont domains, where the Calizzano "massif" (A in fig. 5) should be located, is to be found in sectors more internal than the source area.

In a still inner situation the Prepiedmont zone, with its Jurassic M. Galero Breccias, should be found. For areas more internal than B, data -as already said- are not enough to allow any late Variscan paleogeographic restoration. Yet, it is possible to state that these zones were not (or were poorly) subsiding.

The existence of andesitic lavas directly covering the pre-Namurian basement demonstrates in fact that no Carboniferous sediments were deposited on the Calizzano "massif".

Likewise, the M. Galero Breccias contain no clasts either from Carboniferous sediments (Ollano Formation) or from Permian ignimbrites (Melogno Porphyroids).

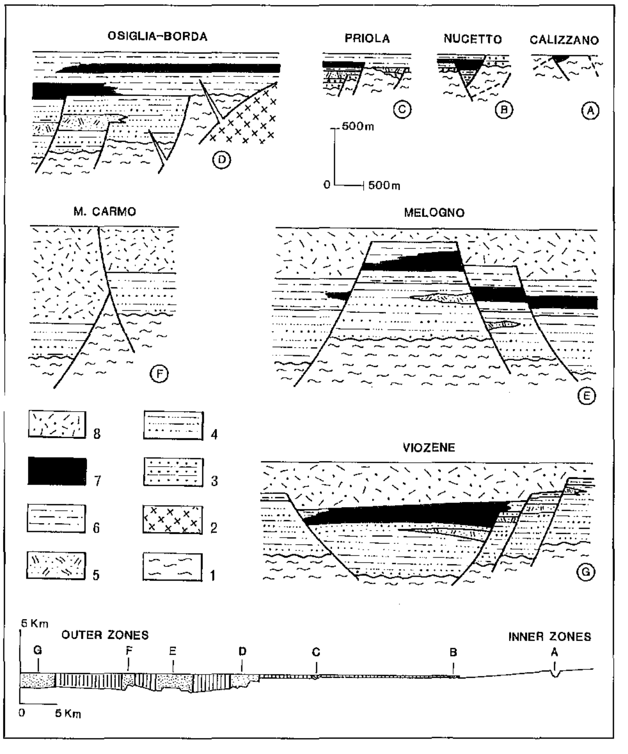


Fig. 5 - Highly speculative paleogeographic setting in the Ligurian Briançonnais domain at the end of early Permian times. 1: Pre-Namurian basement. 2: Late Variscan granodiorites. 3: Lower Late Carboniferous coarse continental metasediments (Lisio Formation). 4: Upper Carboniferous coarse to fine continental metasediments (Ollano Formation). 5: Upper Carboniferous metarhyolitic ignimbrites, agglomerates and dykes. 6: Stephanian-Autunian fine (Viola and Gorra Schists) to very fine, graphitic (Murialdo Formation) continental metasediments. 7:Stephanian-Autunian metaandesites (Eze Formation). 8: Lower Permian, mainly rhyolitic metaignimbrites (Melogno Porphyroids).
Lower cross-section. A to G schemes have been projected onto the same vertical plane; intervening distances might have been even much wider. Small dots and vertical hatching: presumed thickness of Permian-Carboniferous tegument, respectively in A to G sectors and interposed area. In regions more internal than B only erosion is presumed.

4. Chemical Features

Sixty-three selected samples have been analysed for major, minor, and trace elements; twenty-four analyses for E.V. and C.V. are available from Cortesogno et al. (1983).

Attention must be paid to the secondary mobilizations mainly affecting Ca, Na and K, probably during pre-metamorphic diagenetic events (Cortesogno et al., 1983). According to the same authors Al, Ti and Fe seem generally the least mobile elements also in the more altered samples.

On the whole, the detailed petrographic interpretation of primary mineralogy, pseudomorphic overgrowth and alterations processes allow to appreciate the amount of mobilization effects.

In G.V. SiO_2 values higher than about 62 wt.% depend from concentration of quartz phenocrysts (up to about 7 % in volume), rather than from the liquid evolution as no or very little quartz appears in the mesostasis. As a consequence, lower values for most major and trace elements and a negative correlation alkali-silica result (fig. 6 a).

The relative dispersion with respect to the magmatic trend of Ca, alkali and rarely MgO for part of the samples probably depends on secondary mobilization.

However, the average trends for each group are significant and well accord to the inferred primary mineral compositions.

On the whole, the Permo-Carboniferous volcanites (fig. 6 a-d) are characterized by high Al_2O_3 contents (15-22 wt.%) and low total Fe_2O_3 (5-10 exceptionally 12 wt.%), but very low in G.V. (1.2-5.1 wt.%); Mn ranges from traces to 0.25 wt.%, inversely correlated to SiO_2. TiO_2 is relatively low ranging from 0.6 to 1.0 wt.%, but for C.V. (1.0-1.8 wt.%) and S.V. (0.8-2.3 wt.%; fig. 6b). Also Zr is relatively low (131-395 ppm) with values as low as 62 ppm in basaltic andesites. Y generally ranges from 18 to 47 ppm with rare lower values (about 12 ppm) in basaltic andesites and an anomalous 190 ppm value (fig. 6e).

Compared to mean values, G.V. show relatively higher Nb, Zr and Y contents and C.V. high Zr, Y, Ti and P. Also part of the samples from E.V. E shows incompatible element contents higher than mean values (fig. 6e).

REE patterns (fig. 7) normalized according to Nakamura (1974) are relatively homogeneous, showing positively fractionated patterns for LREE and less fractionated trends for HREE. Negative Eu anomaly is common to A.V., E.V. and more evident in G.V., but lacking in C.V.. As well as for Ca and Sr some Eu leaching is hypothizable; however, the homogeneous values in each group and the low correlation with rock alteration, suggest that Eu anomaly is mainly a primary feature.

The different characterization of volcanites from different zones as well as the relative internal homogeneity are confirmed by comparing REE abundances (tab. 2): the E.V. show high relative fractionation REE (La/Yb: 10.5 in E.V. C, 7.8 in E.V. D, 19.6 in E.V. E, 12.1 in E.V. G, 14.0 in E.V. H), HREE fractionation (Tb_N/Yb_N: 2.3 in E.V. C, 1.60 in E.V. D, 3.0 in E.V. E, 2.0 in E.V. G, 1.80 in E.V. H) and negative Eu anomaly (Eu/Eu* about .3 in E.V. C, E.V. D, E.V. E, E.V. G, but for .78 in E.V. H). A rhyodacitic sample from E.V. H shows lower $\sum$REE and lack of Eu anomaly.

A.V. show homogeneous trends (fig. 7) and generally high relative REE fractionation (La/Yb: 7.47), high HREE fractionation (Tb_N/Yb_N: 1.30) and slight negative Eu anomaly (Eu/Eu*: 0.73).

C.V. differ for slight to lacking Eu anomaly (fig. 7), high relative REE fractionation (La/Yb: 8.8 dykes, 7.7 flows) and low HREE fractionation (Tb_N/Yb_N: 1.42 dykes, 1.34 flows).

In G.V. the variable $\sum$REE strongly depend on quartz xenocrysts concentrations in part of the samples. On the whole, the relative REE fractionation is high (La/Yb: 9.1), HREE fractionation low (Tb_N/Yb_N: 1.2) and Eu anomaly evident (Eu/Eu*: 0.31).

Samples affected by major mobilization effects (also evidenced on petrographic ground) mainly show increase in Ce, Lu, sometimes Sm and La, and Eu decrease.

	G.V. (10)	G.V. (10)	E.V.C. (1)	E.V..D. (3)	E.V.E. (3)	E.V.G. (1)	E.V.H. (8)	A.V. (4)	Oceanic Island arc			Continental island arc		Andean	
									LEand	And	Shosh. and.	And.	Shosh. and.	And.	Shosh and
La ppm	28.53	27.57	17.0	23	34.0	40	24.92	20.61	3.0	11.7	-	17	-	28.5	-
Ce ppm	74.54	65.73	18	58.33	55.67	84	58.29	42.59	6.9	23.5	-	37	-	60.7	-
SREE	132.62	139.72	48.69	113.49	123.86	165.48	129.32	107.50	35.2	74.2	-	94.4	-	146	-
La/Yb	9.14	7.68	10.49	7.81	19.57	12.08	14.04	7.47	1.2	6.4	-	8.9	-	16.5	-
La/Y	1.08	1.03	.49	.87	.90	2.0	1.42	.80	.11	.58	.8-1.8	.93	1.4	1.46	4
Th ppm	60.2	5.63	2.20	6.70	8.43	12.80	6.70	5	-	-	<3	-	5-11	-	-
Zr/Y	11.21	11.35	5.14	10.40	9.03	18.45	8.92	6.44	2.2	4.7	4	5.4	=10	14.6	12-16
Ni ppm	21	22.32	75.	71.67	63.67	72	37.63	28.25	7.0	18	<15	21	10-15	27	20.5
Ni/Co	2.2	.67	-	.6	5.1	-	1.9	1.83	.29	.52	-	.95	-	1.4	-
Sc/Cr	1.4	4.83	-	1.7	1.0	-	.3	.29	3.8	.8	-	.61	-	.36	-
Sc/Ni	7.5	2.27	-	2.8	1.4	-	.44	1.07	3.4	2.0	-	1.1	-	.55	-
Eu/Eu	.31	.54	.29	.31	.28	.27	.78	.73	1.0	.98	-	.87	-	.83	-
K/La	1908.76	737.53	908.23	813.03	206.93	531.26	259.19	627.76	1950	1150	1000-3000	814	800-900	715	400-800
P/La	18.62	52.22	7.66	29.44	17.91	13.20	50.90	42.28	180	86	70-230	49	25-60	47	18
Tbm/Yb	1.16	1.34	2.28	1.60	2.95	1.95	1.78	1.30	.95	1.76	-	1.31	-	2.03	-
Th/U	-	4.52	5.50	7.03	6.79	3.12	7.31	-	-	-	2-2.5	-	2.5-4	-	-
Rb ppm	146	80	-	88	61.4	-	30.6	67.50							
Ba ppm	447	352.50	-	-	305.8	-	164.6	346.75							
SLREE	126.53	128.26	45.20	107.70	119.23	159	115.69	90.15							
Sr ppm	13.2	78	-	-	157.9	-	267.8	182.75							

Table 2: Diagnostic parameters and element ratios for andesites (Bailey, 1981).

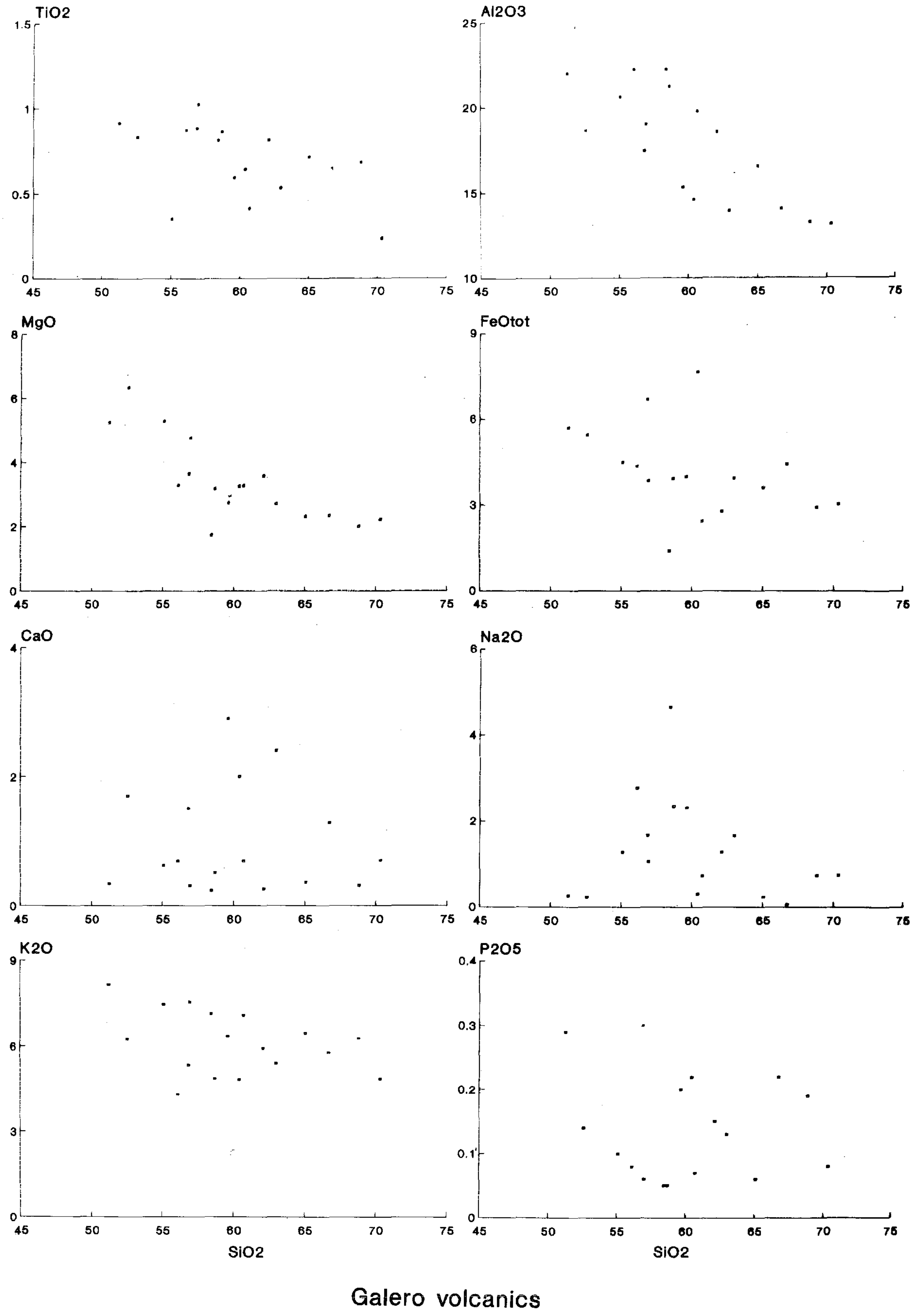
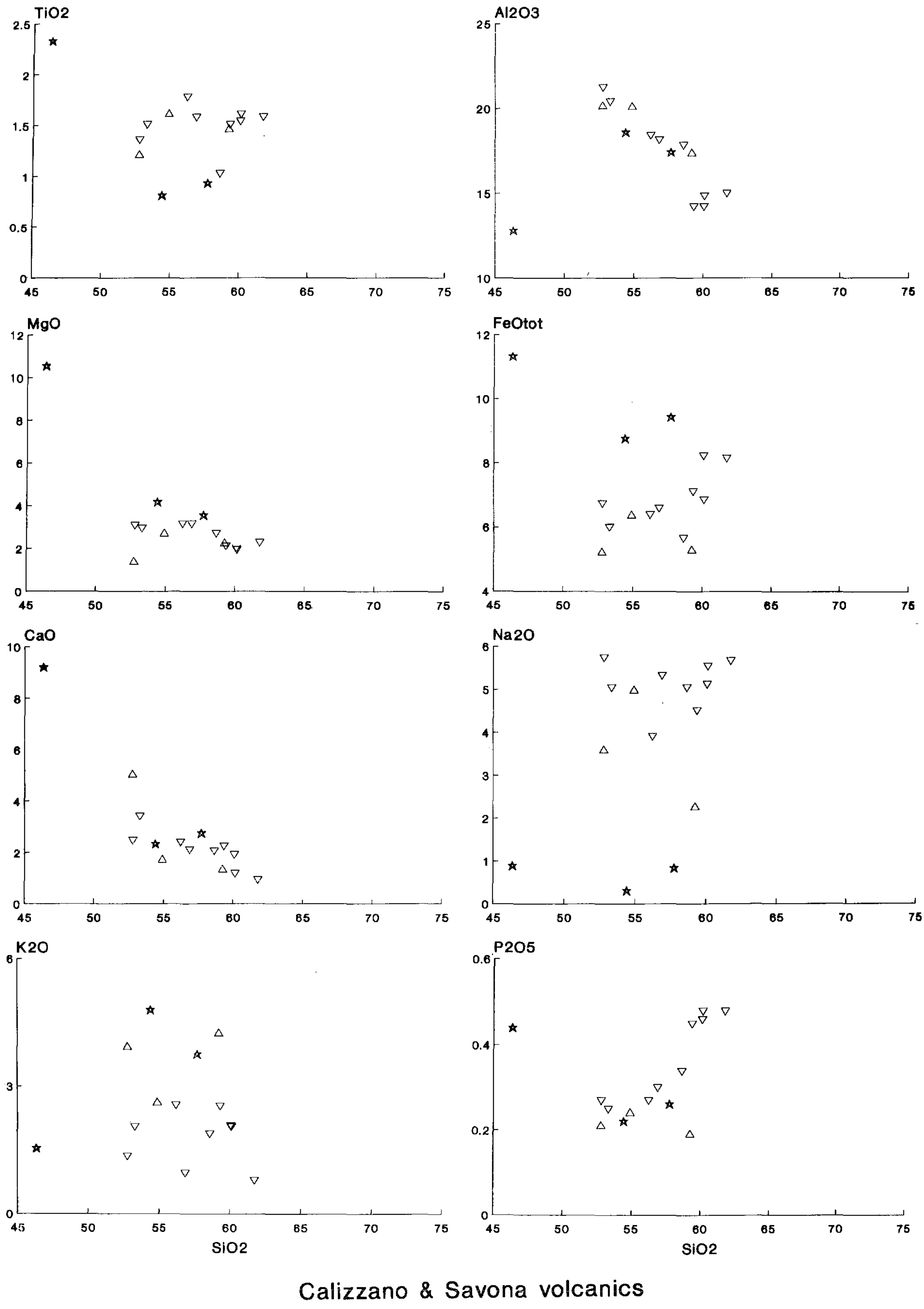


Fig. 6a,b,c,d,e - Major and minor element covariances vs. SiO_2 for C.V.+S.V., G.V., E.V. and A.V.



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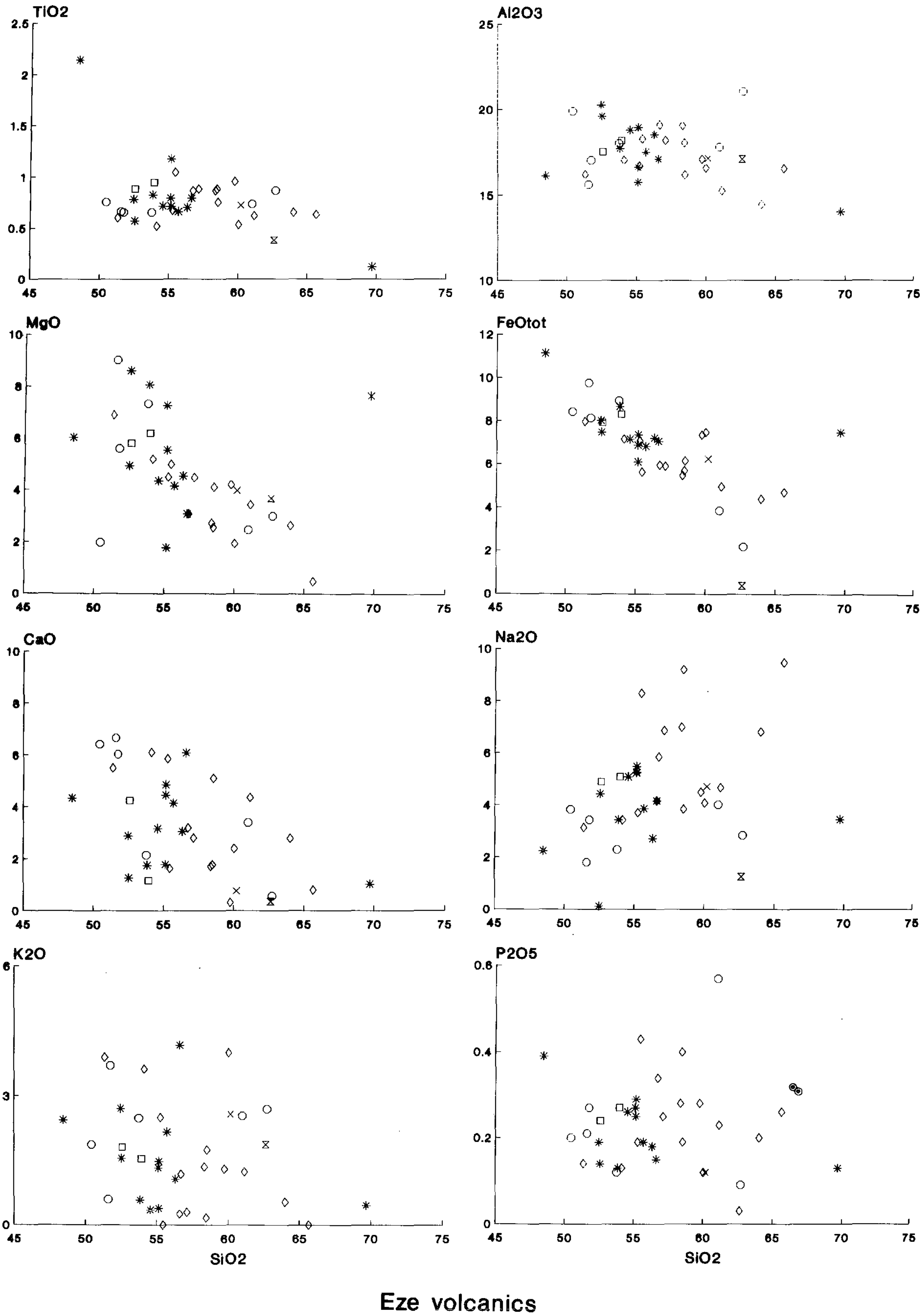
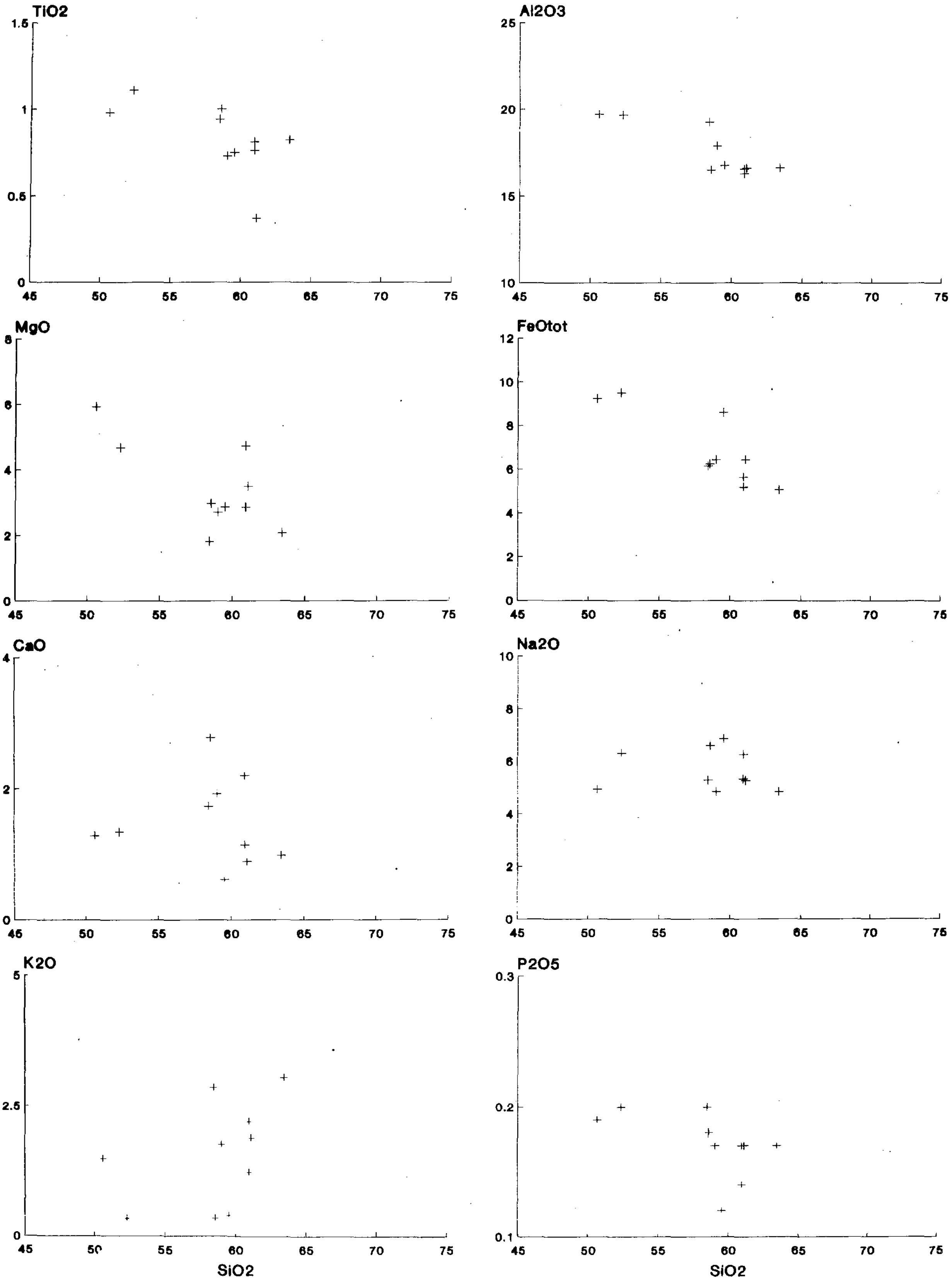


Fig. 6a,b,c,d,e - Major and minor element covariances vs. SiO_2 for C.V.+S.V., G.V., E.V. and A.V.



Aimoni volcanics

Fig. 6a,b,c,d,e - Major and minor element covariances vs. SiO_2 for C.V.+S.V., G.V., E.V. and A.V.

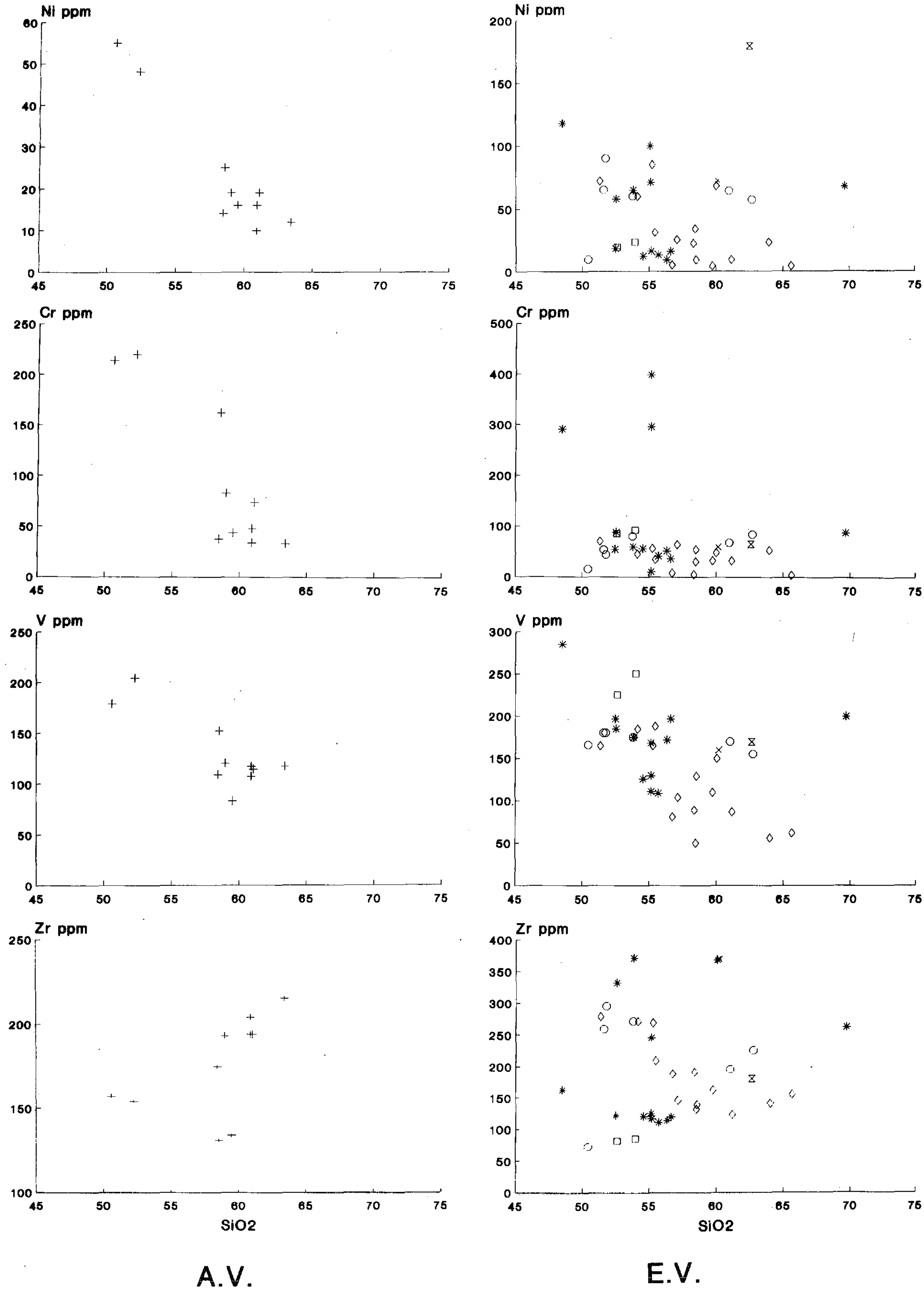


Fig. 6a,b,c,d,e - Major and minor element covariances vs. SiO_2 for C.V.+S.V., G.V., E.V. and A.V

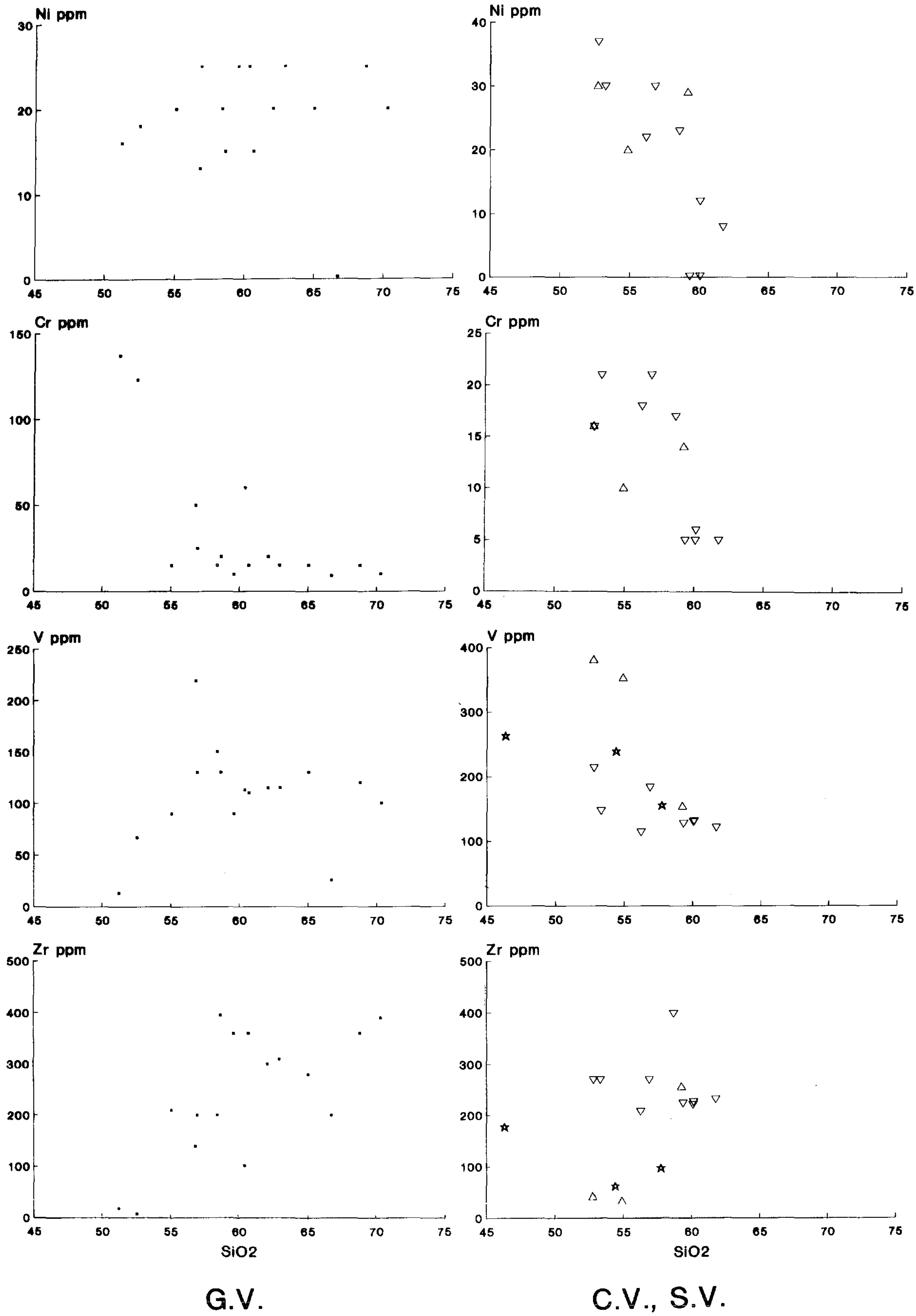


Fig. 6a,b,c,d,e - Major and minor element covariances vs. SiO_2 for C.V.+S.V., G.V., E.V. and A.V

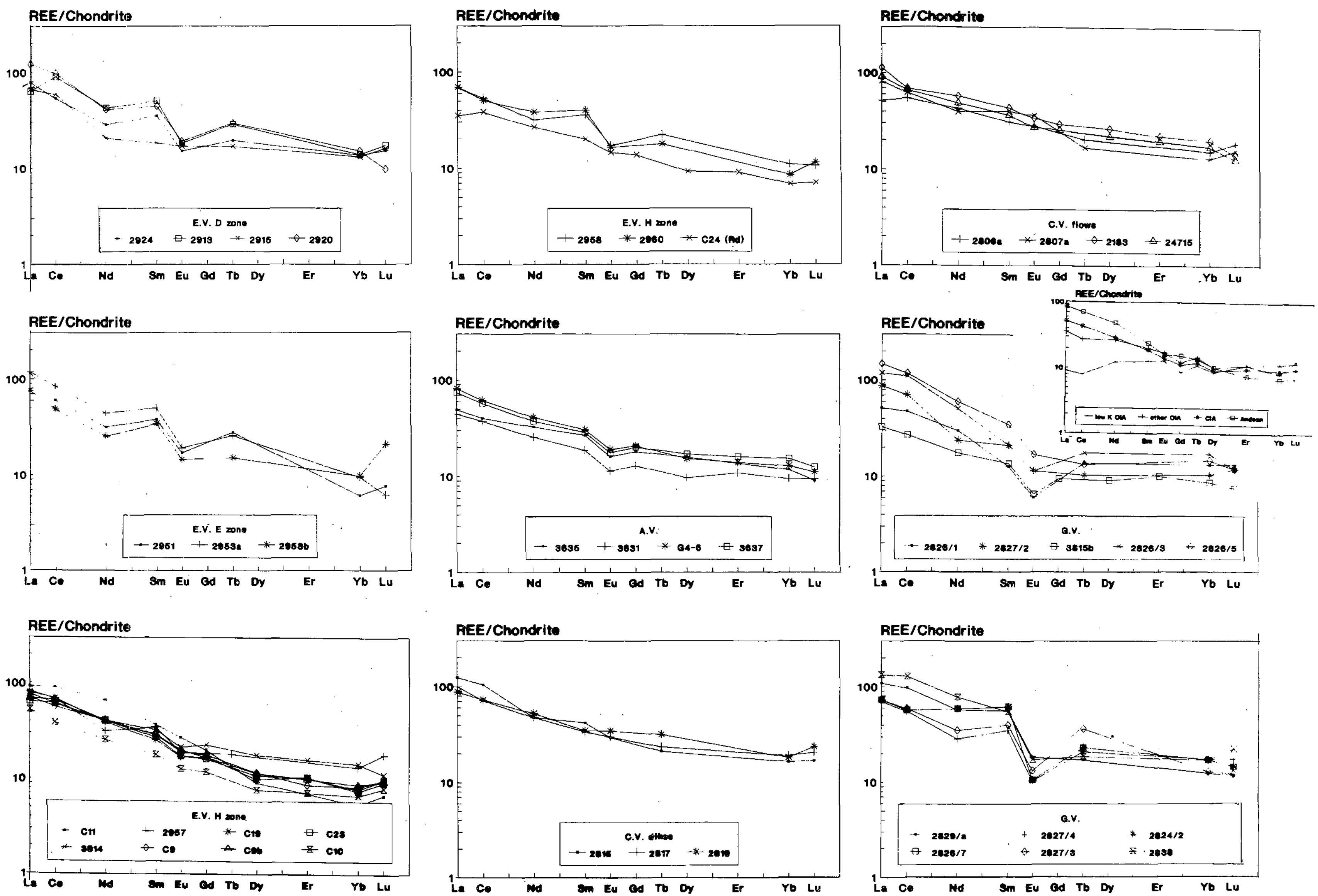


Fig. 7 - REE chondritic patterns, normalized according to Nakamura (1974). Window on the right: average REE patterns for low K Ocean Island, Ocean Island, Continental Island and Andean Andesites (Bailey, 1981).

5. Petrogenetic Constraints

The Briançonnais Permo-Carboniferous volcanism shows, on the ground of major, minor and REE, a subalkaline character and calcalkaline affinity (fig. 8) but for G.V. which have a shoshonitic characterization (fig. 3, 4, 9), also supported by relatively high incompatible element contents. On the whole, chemical data well accord with orogenic andesites except for the occurrence of negative Eu anomaly (Taylor, 1969; Bailey, 1981); E.V. are generally calcalkaline, showing strong similarities with Andean andesites, in the outer/western zones (F, I in tabs. 1, 2) whereas high K-andesitic affinities prevail in the inner/eastern zones (D, E, H in tab. 1).

The transition zone between the two types runs nearly parallel to paleogeographic directions inferred from the present tectonic setting. Trends are therefore indirectly and indipendently confirmed.

Some tendencies toward transitional series also result for C.V., characterized by higher TiO_2 and Zr contents (tab. 2 and figs. 9, 10). Intermediate trends toward shoshonitic series result from trace elements in fig. 9 for E.V. H, E.V. D, C.V. and part of E.V. E..

Hornblende phenocrysts commonly predominate in samples showing K-andesitic affinity, moreover biotite and K-feldspar rimming plagioclase in G.V. accord with the shoshonitic characterization; on the contrary it is noteworthy that quartz phenocrysts lack in typical shoshonitic series (Morrison, 1980).

Significant parameters for tectonic discrimination have been compared (Tab. 2) with data from Morrison (1980) and Bailey (1981).

The major and trace element abundances as well as REE fractionation are, on the whole, comparable with andesites generated at thick continental margins (see also Thorpe et al., 1976, 1979, 1982; Dostal et al., 1977).

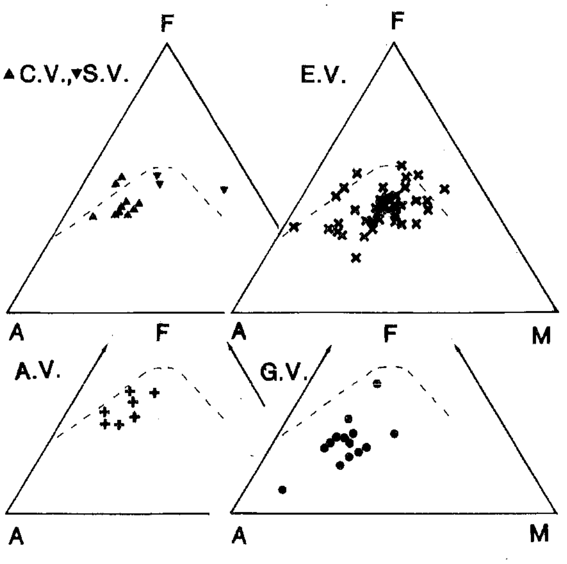


Fig. 8 - Alkali-FeO-MgO diagrams for C.V.+S.V., G.V., E.V. and A.V.

At the regional scale, the subalkaline character and calcalkaline affinity of the E.V., A.V. and C.V. well correspond to the Lower Permian volcanites from Alto Adige (Bargossi & Calanchi, 1984; Bargossi et al., 1981, 1983; Di Battistini et al., 1988) and Lugano (Stille & Buletti, 1987) areas; moreover K-andesitic affinity also results for part of volcanites from Trentino-Alto Adige region (Di Battistini et al., 1988). Quartz xenocrysts in E.V. and A.V. suggest contamination processes by crustal material possibly accounting for average high Rb, Sr and Ba contents. For the coeval Lugano area andesites, important mantle-crust interaction including wall-rock assimilation and fractional crystallization are suggested on the ground of isotopic data (Stille & Buletti, 1987).

MORB-normalized trace element patterns have been compared (fig. 9) with average values for orogenic rocks (Bailey, 1981). A.V. well match with Andean andesites except for two dacitic samples; a good affinity with Andean andesites is also shown for E.V. F and part of E.V. E samples .

An origin of melts below thick continental crust also matches with HREE fractionation which can be interpreted as due to equilibrium with garnet in the residuum (Arth, 1976).

Partial melting in equilibrium with garnet also consists with the observed trends in fig. 11, which cannot derive from fractional crystallization involving olivine, plagioclase, clinopyroxene and hornblende.

The negative Nb and Ti anomaly (fig. 9) shown by Permo-Carboniferous andesites but for S.V. and part of C.V. and common to many orogenic andesites, has been ascribed to the stability of Fe-Ti oxides during partial melting at high PH_2O, or alternatively to early fractionation of Fe-Ti oxides (Venturelli et al., 1984).

The last hypothesis contrasts with petrographic evidences of late crystallization of ilmenite in the studied volcanites.

Different hypotheses are proposed for Eu anomaly (Cortesogno et al., 1983):

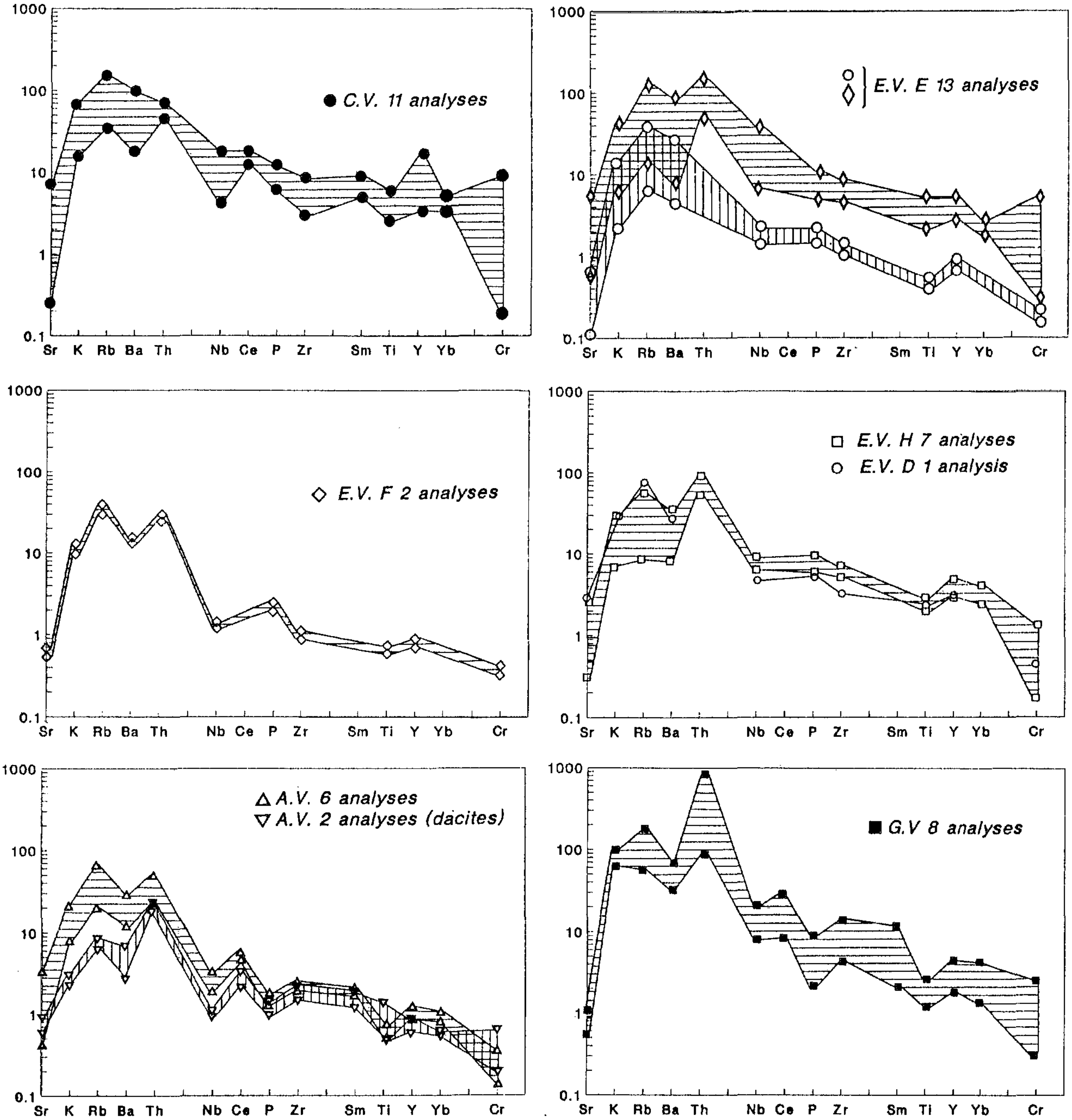
1) partial melting of pl-bearing rocks and plagioclase concentration in the residuum; 2) heritage from the source; 3) important plagioclase fractionation during shallow level crystallization.

The first hypothesis seems to contrast with suggestions of garnet stability in the source, whereas petrographic evidence of early crystallization of plagioclase can support the third hypothesis.

The occurrence of hydrate phases (hornblende, biotite) in most lavas except for some more primitive basalto-andesitic compositions is common also to volcanics outpoured on thick continental crust (Coulon & Thorpe, 1981), suggesting an early crystallization step at high PH_2O and temperatures (Green, 1982).

Also the rare occurrence of basaltic compositions - restricted to dykes intruded at relatively deep levels (S.V.)- is considered to match with conspicuous crustal thicknesses.

In the Ligurian Briançonnais the Late Variscan



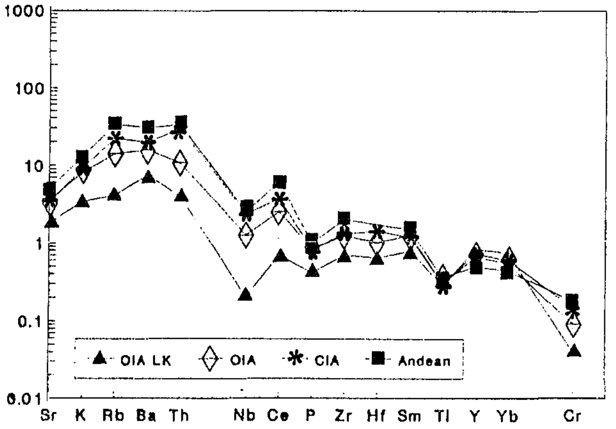


Fig. 9 - MORB-normalized (values from Pearce et al., 1981) spiderdiagrams for C.V., E.V., A.V. and G.V.; for comparison, average values for low K Ocean Island, Ocean Island, Continental Island and Andean Andesites (Bailey, 1981).

volcanism is characterized by a sequence of early rhyolitic-rhyodacitic ignimbrites, followed by andesites, then by Lower Permian rhyolitic ignimbrites associated with dacites and finally by alkaline rhyolites.

This suggests a tight association of intermediate and acidic magmas developing along two evolutive trends: mixing processes are hypothized for the origin of dacites from Aimoni zone (Cabella et al., 1988) as suggested also for the Trentino- Alto Adige region (D'Amico, 1986).

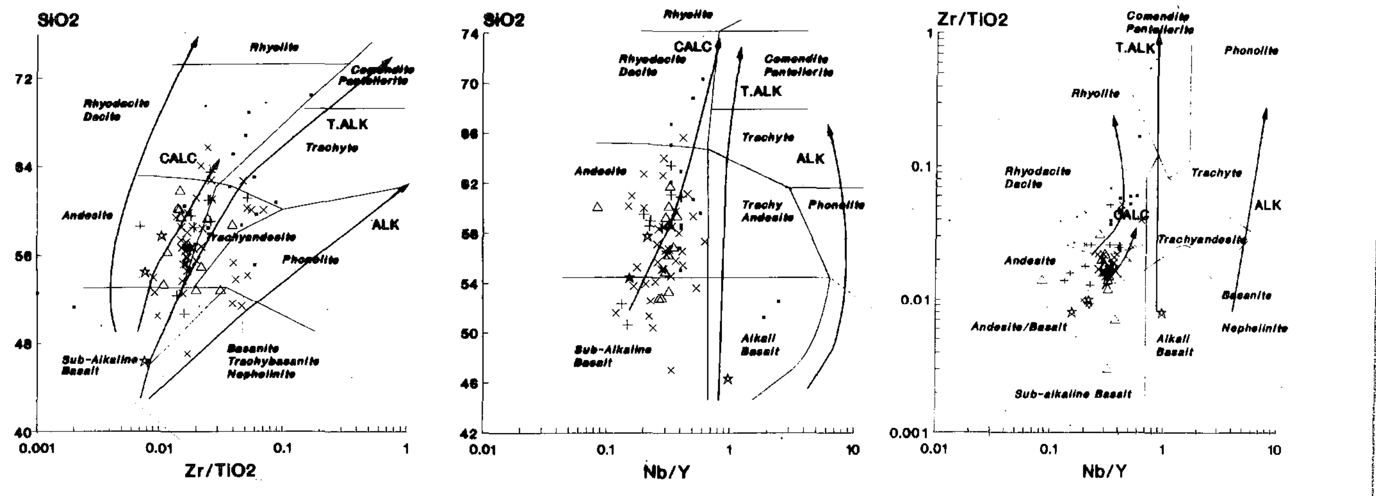


Fig. 10 - SiO_2-Zr/TiO_2, SiO_2-Nb/Y and Zr/TiO_2-Nb/Y diagrams (Winchester & Floyd, 1977). Typical trends of calcalkaline (CALC), alkaline transitional (T.ALK), and alkaline (ALK) are reported. Symbols for zones as in fig. 8.

Silicic ignimbrites occur only on continental crust, generally exceeding 25 km and are commonly associated with extremely thick crust (Central Andes, Tibetan plateau) or with crust affected by extensional or transcurrent tectonics (New Zealand, Sumatra, Sardinia).

The space-time relationships between the thickness of continental crust and the occurrence of silicic ignimbrites associated with andesitic lavas are focused by Coulon & Thorpe (1981), Brown (1982), Pitcher (1987): in the Andes ignimbrites associated with andesites are abundant mainly in the Central Zone where crustal thicknesses of 70 km are recognized .

The affinity with a shoshonitic trend of the Prepiedmont G.V. evidences a geodynamic environment different from that of Briançonnais E.V., A. V., C.V. and might be related with a longer stay of magmas in the crust. Anyhow it may be underlined that this volcanism occurred at or near to the presumed site of the future pre-Tethyan rifting.

6. Conclusions

When trying to correlate the paleogeographic picture (chap. 3) with a geodynamic frame, two main sectors should be distinguished: an outer sector including most of the Briançonnais domain, and an inner sector, covered by the innermost Briançonnais area and by the adjoining Prepiedmont domain.

The outer sector represents an actively subsiding, unstable region, submitted to tensile stress (a transcurrent component can be neither excluded nor demonstrated) producing graben and half-graben structures.

The inner sector can be regarded as relatively stable, owing to the lack of conspicuous subsidence; this, on the other hand, does not mean that stress was missing. On the contrary, a tensile stress field should be admitted, as extensional fractures were produced filled by andesitic dykes or offering an easy way to rising andesitic lavas (Calizzano "massif" and, possibly, also the M. Galero zone).

The geological difference between outer and inner sector might then be correlated with a different crustal rigidity, which in turn might have conditioned their lesser or greater tectonic stability.

Petrographic and chemical data indicate that, on the whole, petrogenetic processes occurred under a thick, or very thick, crust.

This does not contrast with late orogenic frame deduced from field evidences: in fact the former Variscan events may have produced crustal thickening as a consequence of nappes superposition.

It may be assumed that the outpouring of melts generated under thick crust was favoured by subsequent extensional regime, which in turn might be regarded as the first step of the later crustal attenuation, preceding and accompanying the Mesozoic pre-Tethyan rifting.

On the contrary the orogenic and/or subductive picture that, on the ground of actualistic models, is often correlated with andesitic volcanism, has no geological support either in Ligurian Alps or in adjoining areas during the Permian-Carboniferous times.

Disregarding models and taking into account only the link between geochemical parameters and crustal thickness (and/or stability), the shoshonitic affinity of G.V. compared with the low-K to the high-K andesitic compositions of E.V. and A.V. might correlate either with the greater tectonic stability of the inner sector (where only fractures were produced), compared with the outer sector, were conjugate sets of active normal faults were generated, and/or with a later development, in the inner sector, of the volcanic and tensional tectonic activity.

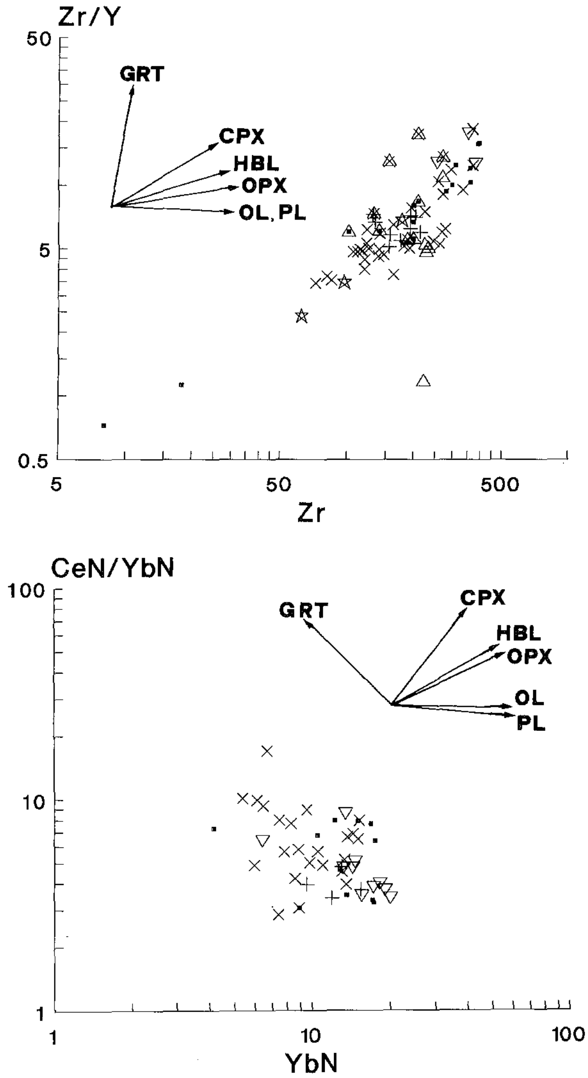


Fig. 11 - Zr/Y-Zr and Ce_N/Yb_N-Yb_N covariances for Briançonnais andesites. Trends for fractional crystallization of garnet (Grt), clinopyroxene (Cpx), hornblende (Hbl), orthopyroxene (Opx), olivine (Ol) and plagioclase (Pl) from Venturelli et al. (1984).

In the same context, the average more evolute compositions of A.V. may indicate a longer stay in a subsurface environment and therefore a somewhat younger age if compared with E.V.: this conclusion matches with their relatively high position in the stratigraphic column of the Permian-Carboniferous sequences of the region (Cabella et al., 1988).

Analytical Tecniques

The composition of clinopyroxenes was determined using a SEM-EDS microprobe, installed at the DISTER - sezione Mineralogia-Petrografia of Genoa.

Analyses were recalculed on 6 oxygens to 4 cations and Fe^{3+} = 12-total cation charge.

Whole rock analyses (major, minor, trace and RE elements) were carried out using an ICP spectrometer JY70 at the CRPG (Vandeuvre-les-Nancy).

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NEW CHEMICAL DATA ON THE UPPER ORDOVICIAN ACIDIC PLUTONISM IN THE AUSTRIDES OF THE EASTERN ALPS

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Abstract

The Upper Ordovician acidic magmatism is the most important magmatic activity in the Eastern Alps, both as concerns amount of melts and size of the area. It took place both under plutonic and volcanic conditions.

New chemical data on this plutonism are presented here. They concern rock samples from Oetztal, Pitztal and Casies Valley (Gsiestal). These new chemical data cover the SiO_2 range 67-77 wt%. The data points cluster along a more or less well-defined pattern in several variation diagrams concerning major and trace elements, including REE.

These new data are compared with previously published data concerning: (i) similar Austridic orthogneisses; (ii) Austridic sheet-like gneisses, augengneisses and "porphyroids"; (iii) Southalpine "porphyroids". The variation diagrams based on all these data support the following considerations consistently with the conclusion reached by some previous authors:

— Upper Ordovician plutonism and volcanism display identical chemical features, and can be related to a unique cycle of magma generation;

— numerous patches of melts formed in different places in a relatively short time range under identically constrained conditions at expenses of similar parent rocks, so that all these patches of melts could have taken similar geochemical features;

— these melts display a calc-alkaline affinity, covering almost continuously the 62-77 SiO_2 range without any significant gap;

— the lack of cogenetic basic rocks and the insignificant amount of intermediate rocks represent a feature which is meaningful for the genetic interpretation;

— crustal anatexis seems to be the most appropriate process for explaining all available data.

Key words: *Eastern Alps, acidic magmatism, Ordovician, granitoids.*

1. Introduction

The "Caledonian" (Upper Ordovician) acidic magmatism is the most important magmatic activity recorded in the Eastern Alps, both in terms of amount of melts and size of the areas in which it developed (Fig. 1) (Sassi & Zirpoli, 1989, and refs. quoted therein).

It is represented both by volcanic and plutonic products: a large acidic volcanic plateau, now metarhyolites (also reported as "porphyroids" in the literature) in low-grade areas, medium- to high-grade sheet-like leucocratic gneisses and augengneisses elsewhere, and numerous granitic to granodioritic and tonalitic bodies now occurring as sharply-boundaried bodies of orthogneisses.

The "Caledonian" age of the acidic plutonism is based on radiometric data obtained from numerous granitoid masses by several authors (Borsi et al., 1973, 1980; Satir, 1975). It turns out to be in the range 450-420 m.y. (Sassi & Zirpoli, 1979). Similar radiometric ages were obtained from the sheet-like acidic gneisses (Satir, 1975; Brack, 1977; Borsi et al., 1980; Hammerschmidt, 1981; Söllner & Schmidt, 1981). These radiometric age values are consistent with the palaeontologically obtained age (Flajs & Schönlaub, 1976) of the metarhyolites from the "Northern Grauwacken Zone", with which the acidic gneisses have strong geochemical analogies (Sassi & Zirpoli, 1979; Bellieni & Sassi, 1981).

The Upper Ordovician magmatism has been interpreted as a typical granite-rhyolite association (Bellieni & Sassi, 1981) of calc-alkaline affinity (Peccerillo et al., 1979; Sassi & Zirpoli, 1979), with chemical features indicating a crustal origin (Peccerillo et al., 1979).

In order to better understand the chemistry of these rocks for constraining their genetic model, new chemical data concerning the orthogneisses from Oetztal, Pitztal and Casies Valley (Gsiestal) are presented below, and they are compared with previously published data regarding: i) other similar Austridic orthogneisses (sharply-boundaried bodies); ii) Austridic sheet-like leucocratic gneisses, augengneisses and metarhyolites ("porphyroids"); iii) Southalpine metarhyolites ("porphyroids"). The whole set of data (199 data points) is then discussed, in order to better frame the genetic interpretation.

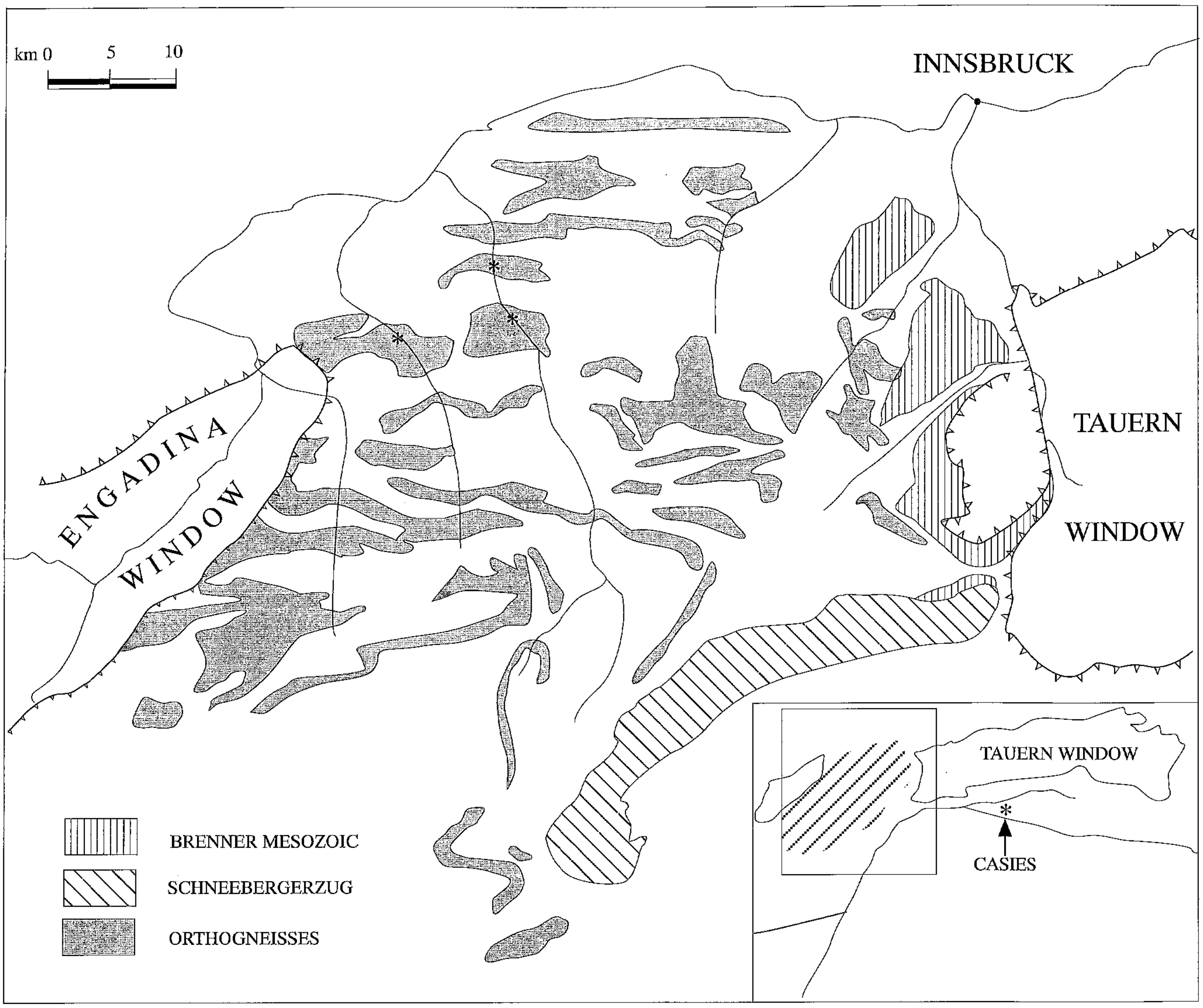

Fig. 1 - Geologic sketch map of the investigated area and sample location (asterisk)

2. Geopetrographic Outline

An old crystalline basement (OCB for short), assumed to be pre-Caradocian, and an overlying monometamorphic Variscan sequence have been recognized within the Austrides of the Eastern Alps (Sassi & Zirpoli, 1989, and refs. quoted therein).

The OCB consists of a medium- to high-grade metasedimentary complex (mainly paragneisses and micaschists) in which acidic leucocratic gneisses, amphibolites and quartzites are conformably interlayered, and sharply boundaried granitoid orthogneisses occur.

A pre-Alpine polymetamorphic history is recorded in this complex. Two main metamorphic events have been recognized (Sassi & Zanferrari, 1972; Borsi et al., 1973, 1978; Purtscheller & Sassi, 1975; Sassi & Schmidt, 1982; Becker et al., 1987; Sassi et al., 1987 and refs. quoted therein):
i) an Ordovician metamorphism of intermediate thermal gradient, developed under amphibolite facies conditions;
ii) a Variscan metamorphism of higher thermal gradient, occurring in greenschist facies in some places, amphibolite facies in others.

Migmatites formed by anatexis in situ locally occur in OCB (Grauert, 1969; Söllner & Schmidt, 1981); the upper Ordovician acidic plutonic bodies are also intruded into OCB. These granitoids (now orthogneisses) never occur within the Variscan monometamorphic sequence, which has been interpreted as the Paleozoic (Caradocian to Devonian) cover of OCB (Sassi et al., 1974; Becker et al., 1987; Ebner et al., 1987; Frisch et al., 1987; Sassi & Zirpoli, 1989; and refs. quoted therein).

The boundaries between the orthogneisses and the surrounding rocks commonly are sharp, as expected in injected bodies (Gregnanin & Sassi, 1969; Borsi et al., 1973; Purtscheller & Sassi, 1975). During their emplacement, a widespread acidic volcanism took place. It is recorded as metarhyolites both in the Southern Alps (Bellieni & Sassi, 1981) and in the Austrides (Heinisch, 1981), and as augengneisses (Heritsch & Teich, 1975) and sheet-like white gneisses (Tumulo gneisses: Gregnanin et al., 1968) in the Austrides.

All products of the Upper Ordovician acidic magmatism were affected by the Variscan metamorphism.

The composition of the granitoid bodies ranges from acidic to intermediate, but acidic types definitely prevail. Oetztal and Pitztal masses substantially consist of (biotite or two mica) granitic orthogneisses, while Casies orthogneisses are mostly tonalitic, bearing biotite ± horneblende.

Xenoliths are present both along the margins and in the inner parts of the granitoid bodies. They are not abundant, with the exception of some few masses (Gaislehn, in the central Oetztal; Parcines, west of Merano). The size of the xenoliths ranges from a few centimetres to some metres. They commonly have a flat shape, and show a foliation parallel to that occurring in the surrounding orthogneisses. A folded foliation has been described in rare cases, contrasting with the unfolded Variscan foliation of surrounding orthogneisses. The occurrence of these xenoliths is very important, because it indicates that the original rocks from which the granitoids derived were already metamorphosed and folded when some of their fragments were incorporated within the melt.

The xenoliths mainly consist of quartzites, amphibolites, quartz- or plagioclase-rich gneisses, gneisses, garnet-bearing gneisses. In many cases their lithology is consistent with the expected composition of the refractory resisters in a process of partial melting of rock-types similar to those making up the present country-rocks.

3. Chemical Data

A set of 65 new chemical analyses is presented and discussed in this paper. These analyses refer to rock samples collected from the Casies (12 analyses), Pitztal (19 analyses) and Oetztal (34 analyses) orthogneiss masses. All analytical data are processed jointly with 62 chemical analyses taken from the literature (8 from the Parcines massif: Gregnanin & Sassi, 1969; 12 from the Anterselva massif, 20 from the Casies massif, 4 from the Dellach massif, 5 from the Trotten massif: Peccerillo et al., 1979; 6 from the Oetztal orthogneisses, 7 from the Stubai orthogneisses: Borsi et al., 1980). Therefore, a total data base of 127 chemical analyses of Ordovician granitoids is considered here. A comparison with further 72 chemical analyses of Ordovician volcanics (and related rocks) is also presented, based on a total of 199 data points.

The samples chosen for the new geochemical investigation were selected so as to represent the main rock types as well as the compositional variability of each massif. Their location is indicated in Fig. 1.

Major and minor element concentrations were determined by X-ray fluorescence method while trace element and REE analyses of 30 selected samples were carried out by a plasma excitation source (ICP) spectrometer (JY70). The chemical data are here only shown as plots in the diagrams. The complete table of all analytical data is available on request.

a) Major, minor and trace elements.

The new data from Oetztal and Pitztal. Fig. 2 shows the variation patterns of some major and minor elements vs. SiO_2 in the Oetztal and Pitztal sample data points. These diagrams allow us to point out some observations:

— all data display regular patterns, with a reasonably good correlation for almost all elements, except P_2O_5 and alkalies;

— the variation range is relatively small; however, there is a defined, single trend in each plot (excluding P_2O_5);

— no significant compositional differences have been detected among the considered masses; however, the samples from Oetztal display a slightly two-modal character.

Similar pieces of information may be deduced from Fig. 3, which refers to the same Oetztal and Pitztal gneisses. Fig. 3 also shows a calc-alkaline AFM affinity, although the compositional range is too small to allow us to assume this affinity as a well established feature.

The Oetztal and Pitztal data points lie within the field of the subalkaline rocks in the Alk-SiO_2 diagram. This character must be considered with caution, because alkali mobilization might have occurred during metamorphism: however, the scatter of the Na_2O/K_2O ratio seems to be not important, as shown in the Na_2O - K_2O - CaO diagram (Fig. 3).

Trace elements vs. SiO_2 diagrams (Fig. 4) give consistent pieces of information, including a defined single trend and a two-modal character for the Oetztal orthogneisses. The correlation is negative for all the considered elements, except for Rb.

The new data from the Casies body. Figs. 5, 6 and 7 are based on 12 new data points. The following observations can be pointed out concerning these rock samples:

— they are compositionally homogeneous, with a variation range which is so narrow that the data points do not define a clear variation pattern;

— a variation trend may be only detected if the SiO_2-richest, single data point is considered; such a trend is consistent with that deduced from the whole data base, including the data from the literature.

Diagrams in Fig. 6 are consistent with the observations deduced from Fig. 3: Casies data points lie along the trend inferred from the Oetztal and Pitztal gneisses.

Comparison with data from the literature concerning granitoid gneisses. The 65 new data are now compared with 62 chemical analyses concerning granitoid gneisses from the same or other similar bodies from the Eastern Alps. All 127 data are shown in Figs. 8 and 9, in which all the above outlined features are confirmed, and acquire a higher level of certainty. Specifically, the variation range becomes larger, and consequently the variation patterns are better defined, although the scatter of the P_2O_5 data is also confirmed. The calc-alkaline AFM affinity is now evident, and the distribution of the data points in the Na_2O-K_2O-

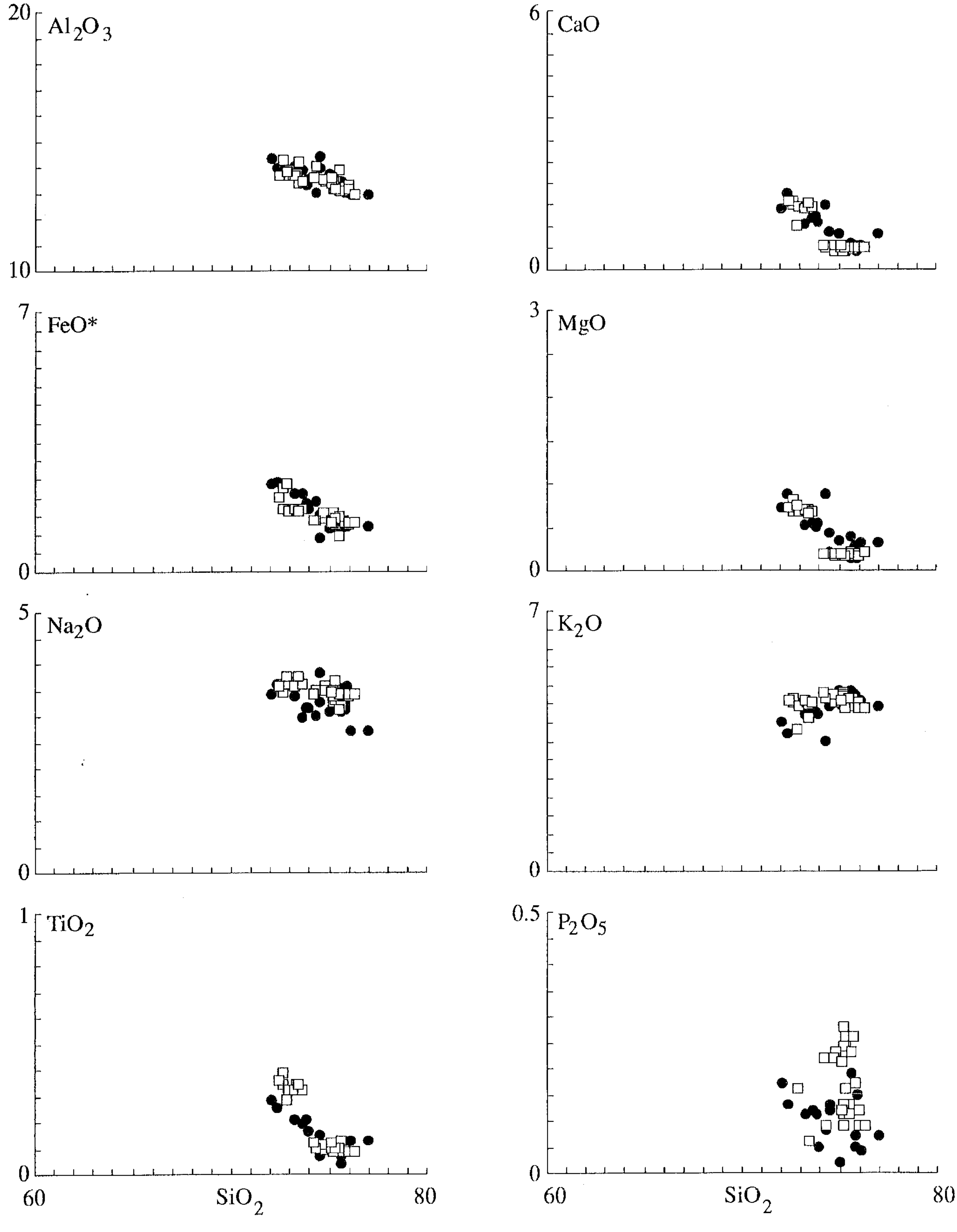

Fig.2 - Variation diagrams of major and minor oxides (wt%) vs. SiO_2 for the Oetztal (open square) and Pitztal (filled circle) orthogneisses.

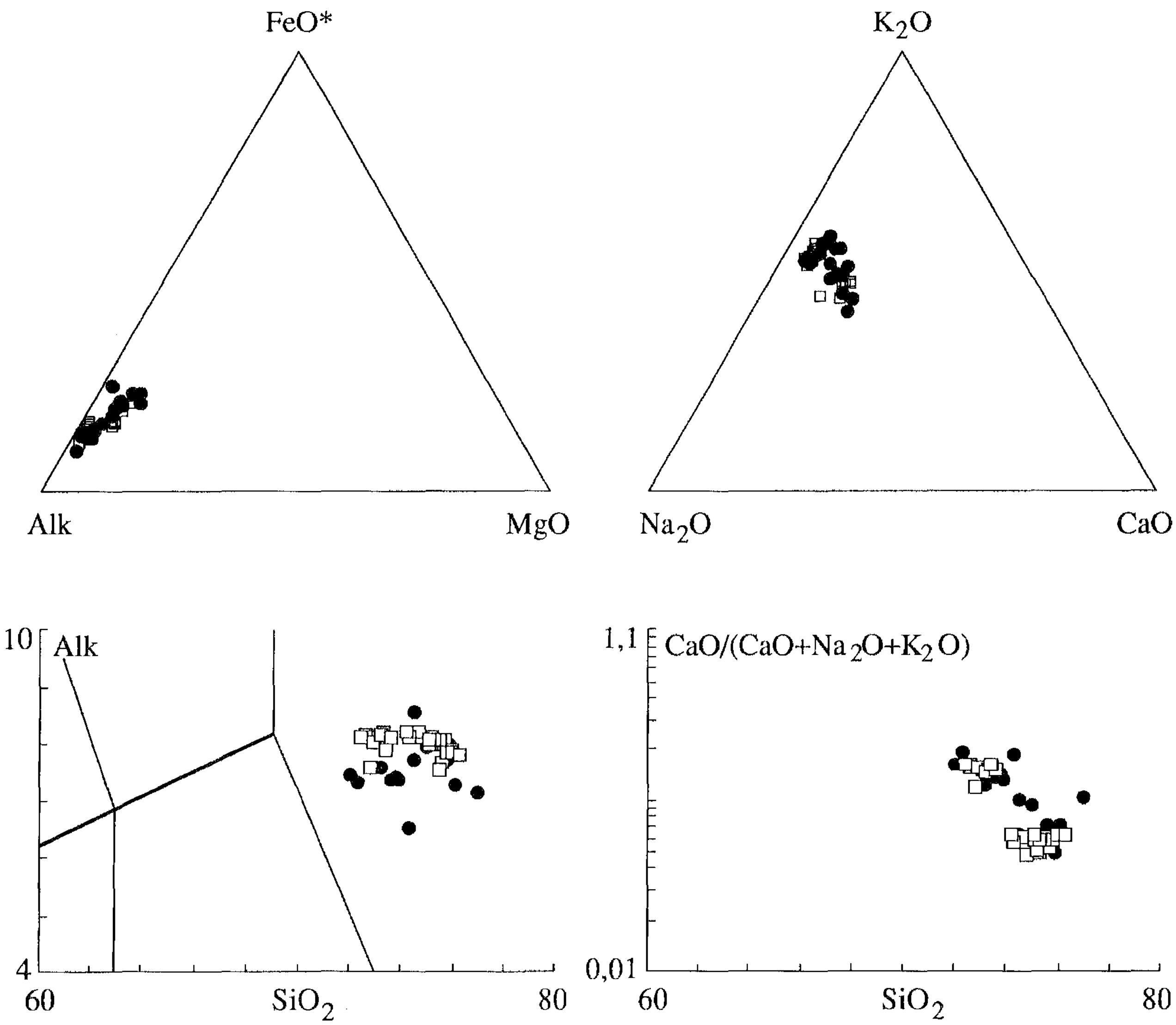

Fig.3 - Other chemical features of the analyzed Oetztal (open square) and Pitztal (filled circle) orthogneisses.

CaO diagram indicates that the Na_2O/K_2O ratio did not change significantly during metamorphism.

The data points from the Parcines body are scattered in some diagrams, but do not deviate from the main trends.

The Anterselva and the Casies data points cluster into two separate sub-areas in all plots, displaying a significant compositional difference between the two bodies.

No significant gaps exist within the whole compositional range, and the most acidic rock types quantitatively prevail. Even considering the scattering of the data points, the correlation of the chemical contents vs. SiO_2 is clear and consistent with a liquid line of descent. All these features indicate that an origin of most of these rocks as products of differentiation, at different degrees, of intermediate magmas is not probable; they suggest a crustal anatectic process.

Peccerillo et al. (1979) suggested indeed that the concerned granitoid orthogneisses represent magmas formed in a crustal anatectic context, the more intermediate and less abundant rocks being co-products formed in extreme environmental conditions. The patterns of the analytical data presented in this paper as well as the occurrence and nature of xenoliths all over these granitoid masses, do agree with this genetic hypothesis. However, doubts still persist for the most basic rock types. In fact, Peccerillo et al. (1979) could not definitely exclude for their origin a fractionation of melts formed by mixing of basic sub-crustal magmas and acidic crustal ones. Further studies are in progress for having further constraints about this hypothesis.

Comparison with data from the literature concerning the coeval volcanism. As mentioned in the introduction, the intrusion of the Upper Ordovician granitoids (now orthogneisses) was associated with a synchronous volcanism, widespreadly represented both in the Austridic and in the South Alpine basements. Mostler (1970), Sassi et al. (1974) and Peccerillo et al. (1979) suggested that this

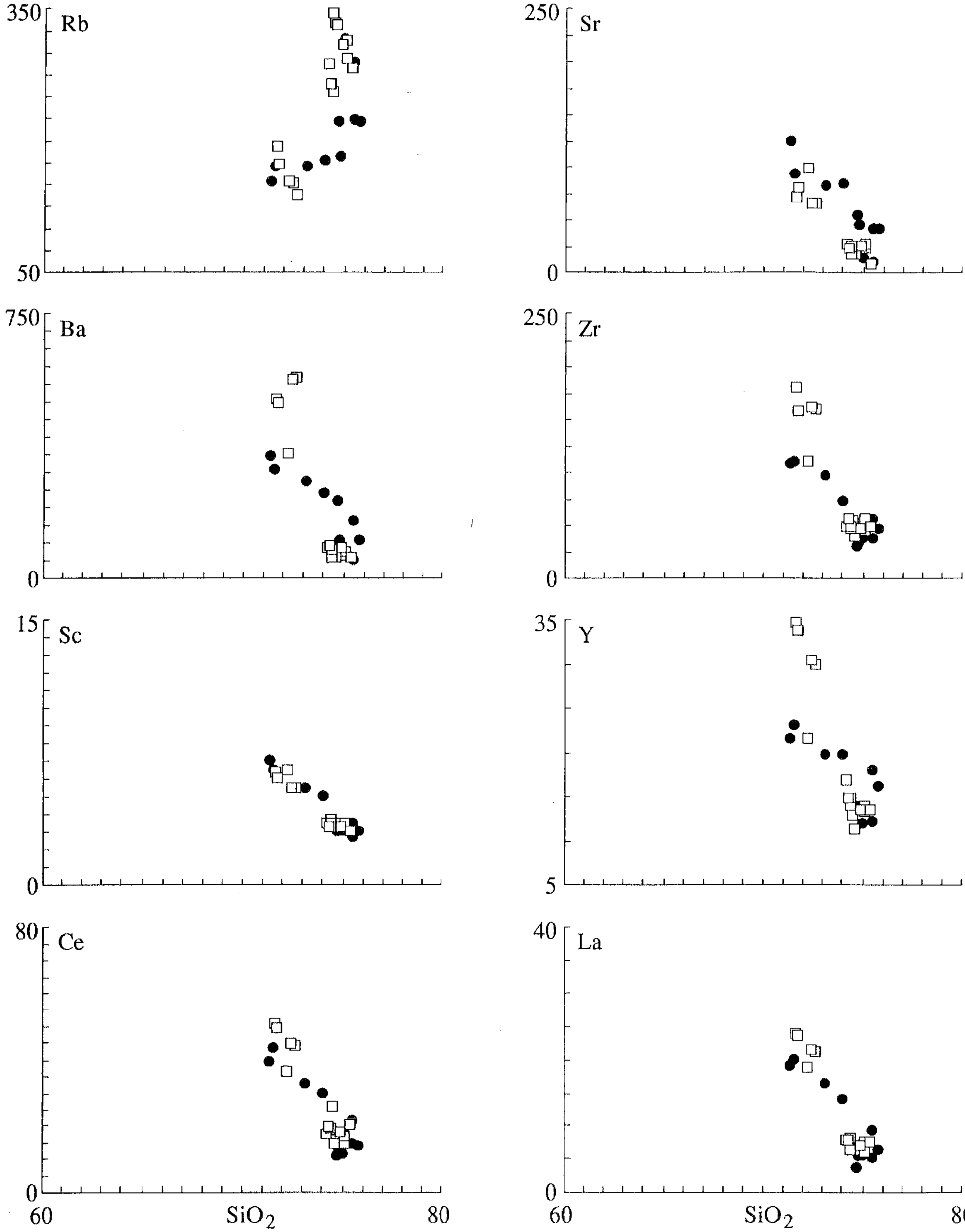

Fig.4 - Variation diagrams of trace elements (ppm) vs. SiO_2 (wt%) for the Oetztal (open square) and Pitztal (filled circle) metagranitoids.

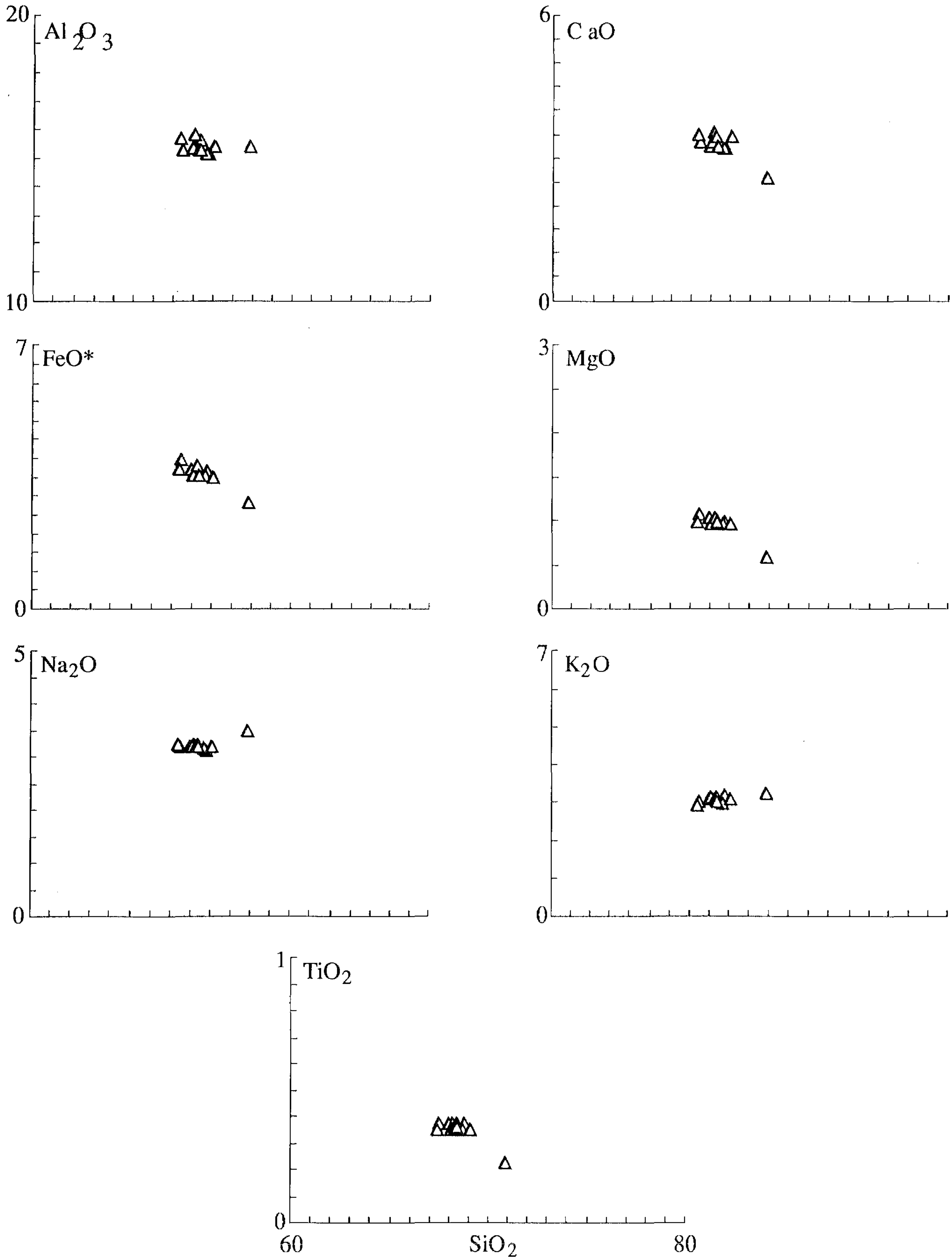

Fig.5 - Variation diagrams of major and minor oxides (wt%) vs. SiO_2 for the Casies metagranitoids.

volcanism was fed by the same melts that elsewhere crystallized in plutonic conditions. Bellieni & Sassi (1981) supported this hypothesis and proposed the name of "Upper Ordovician Granite-Rhyolite Association" of the Eastern Alps for this magmatic activity as a whole.

A comparison of all data concerning the Ordovician granitoids with the chemistry of the coeval, above mentioned volcanics is shown in Figs. 8 and 9: the 127 data points in these diagrams refer to the granitoids, while the dashed line shows the compositional field of the Austridic metavolcanics (metarhyolites, sheet-like white gneisses, augengneisses) as defined by Bellieni & Sassi (1981) utilizing 72 chemical analyses taken from Sassi et al. (1979), Gregnanin et al. (1968), Bellieni & Sassi (1981), Heinisch (1981) and Heritsch & Teich (1975).

The chemical features of the Upper Ordovician plutonic and volcanic rocks are substantially identical in Figs. 8 and 9, if the obvious scatter of the Na and K contents in the metavolcanics is disregarded. The abundance of the acidic versus the intermediate rock types is a peculiarity also in the volcanic products. Therefore the idea that the Upper Ordovician, acidic plutonics and volcanics are cogenetic is confirmed.

b) Rare Earth Elements.

30 samples were analyzed for REE, 5 of which from the Casies massif, 10 from the Pitztal and 15 from the Oetztal orthogneisses.

The samples to be analyzed were selected in order to represent, for each of the three groups, the whole chemical variability as deduced on the basis of the major elements.

Studying ancient magmatic rocks affected by metamorphism, we face the problem of a possible REE mobilization during metamorphism. Contrasting opinions exist among the authors as regards this problem (Green et al., 1972; Cullers et al., 1974; Hellman et al., 1977). Taking into account these opinions, we critically analyzed each

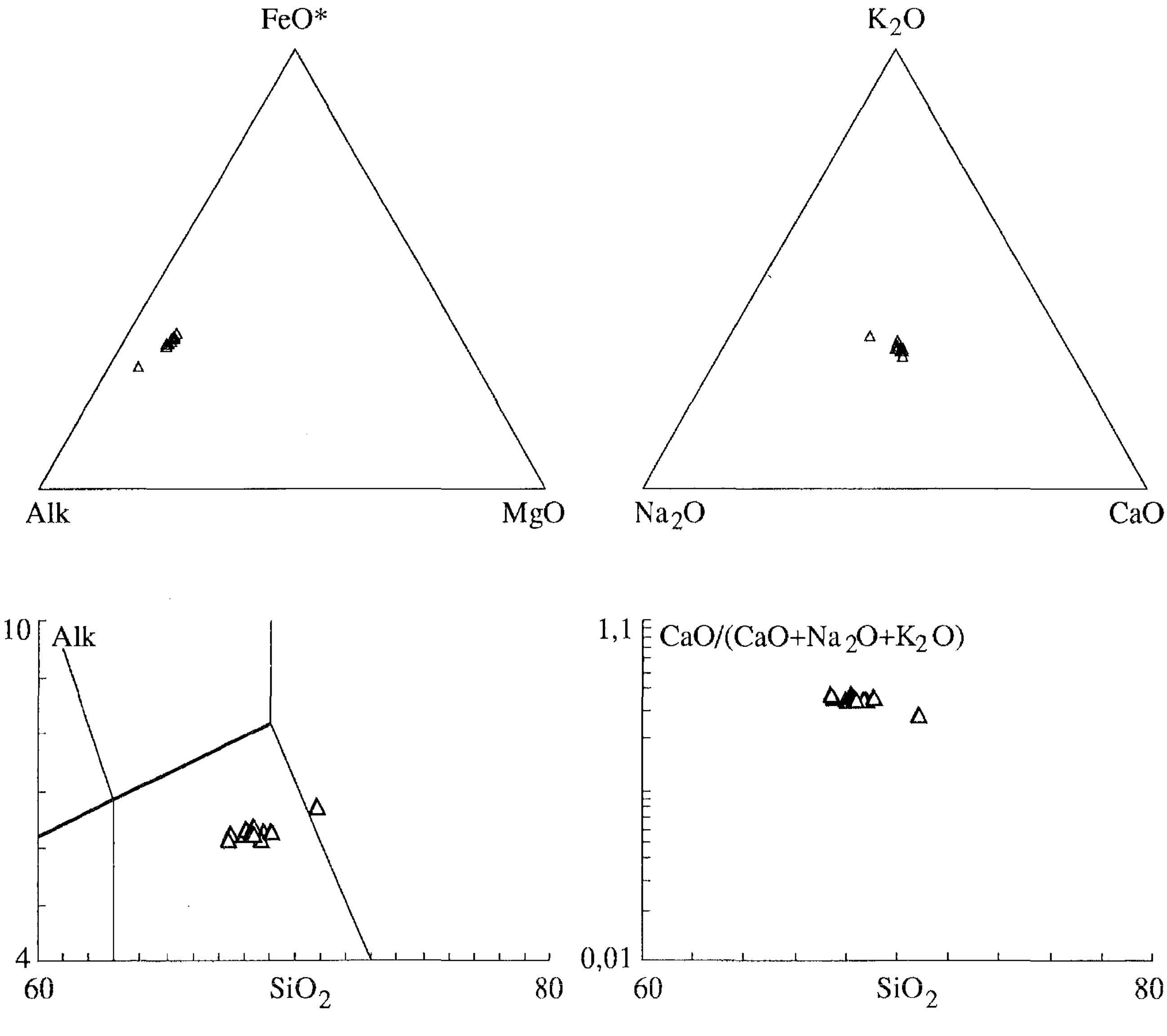

Fig.6 - Other chemical features of the analyzed Casies orthogneisses.

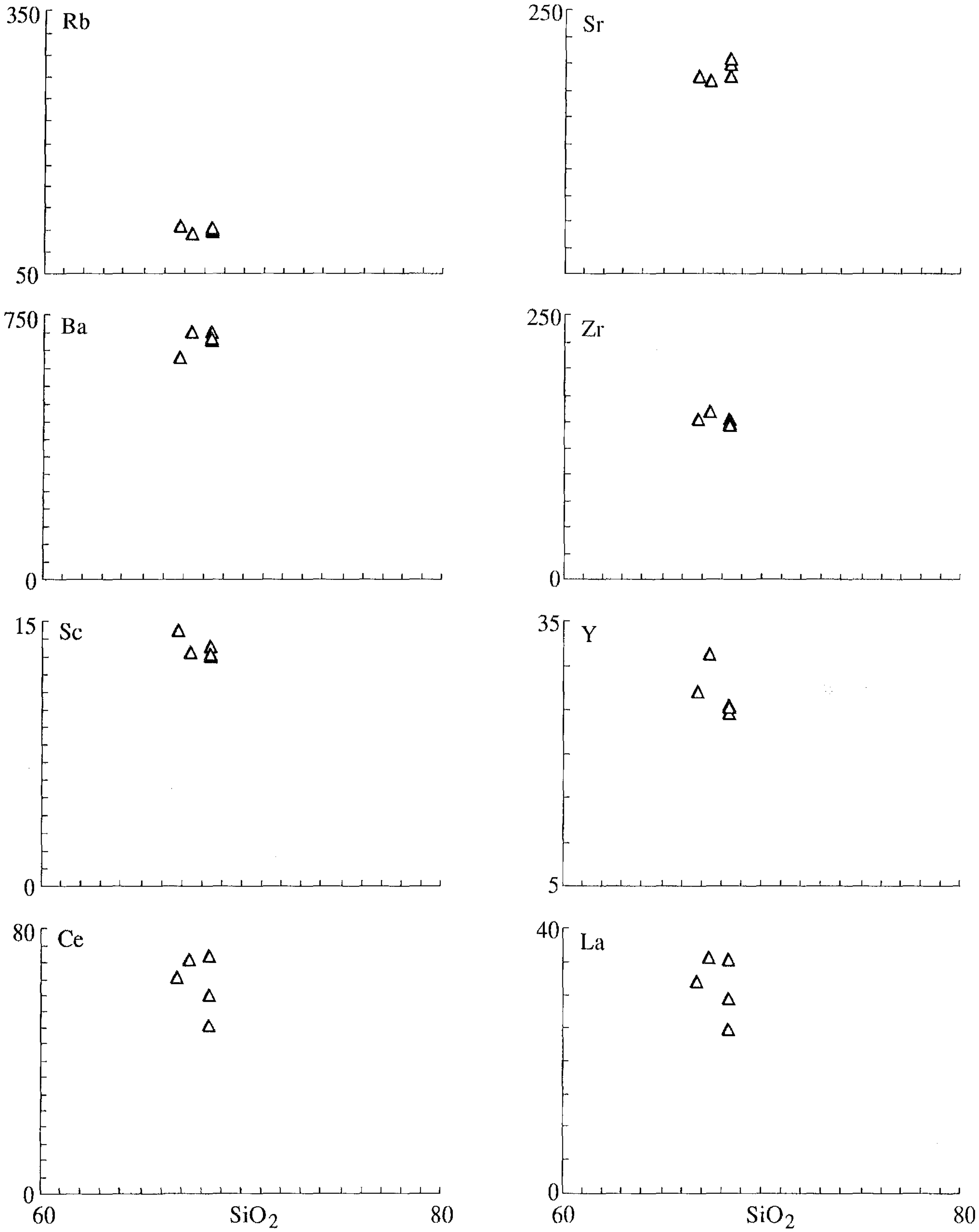

Fig.7 - Variation diagrams of trace elements (ppm) vs. SiO_2 (wt%) for the Casies orthogneisses.

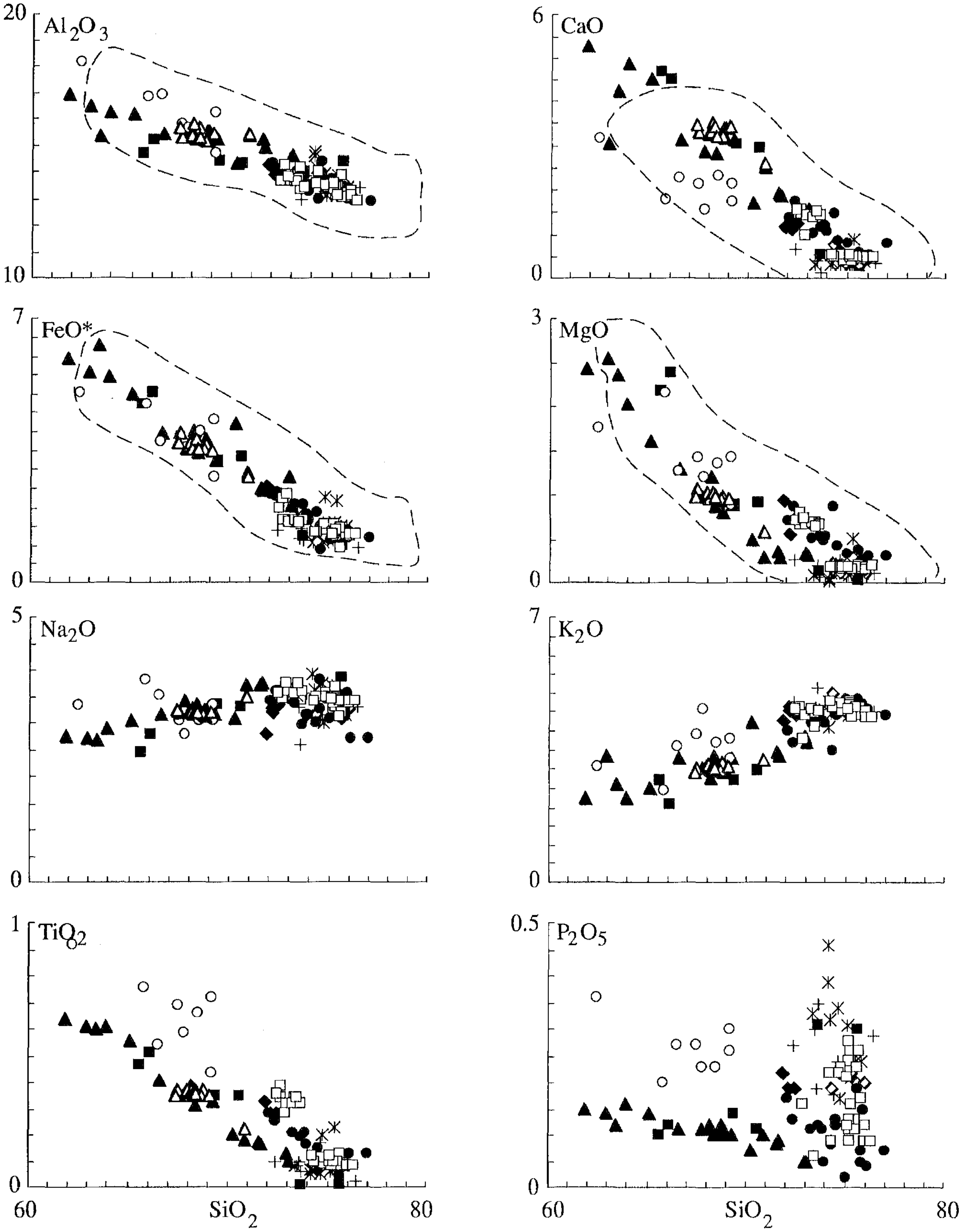

Fig.8 - Variation diagrams of major and minor oxides vs. SiO_2 for the whole set of data concerning the Upper Ordovician granitoids (65 new data and 62 data from the literature). Sources of the data from the literature: Gregnanin & Sassi (1969), Peccerillo et al. (1979), Borsi et al. (1980), and refs. quoted therein. Symbols for data from the literature: Casies (filled triangle); Oetztal (filled square); Stubai (cross); Anterselva (asterisk); Trotten (open diamond); Dellach (filled diamond); Parcines (open circle). Symbols for new data: as indicated in Figs. 2 and 5. The dashed line represents the compositional fields of the Ordovician acidic volcanics as defined by Bellieni & Sassi (1981) on the basis of 72 chemical analyses taken from Sassi et al. (1979), Gregnanin et al. (1968), Bellieni & Sassi (1981), Heinisch (1981), Heritsch & Teich (1975) and refs. quoted therein.

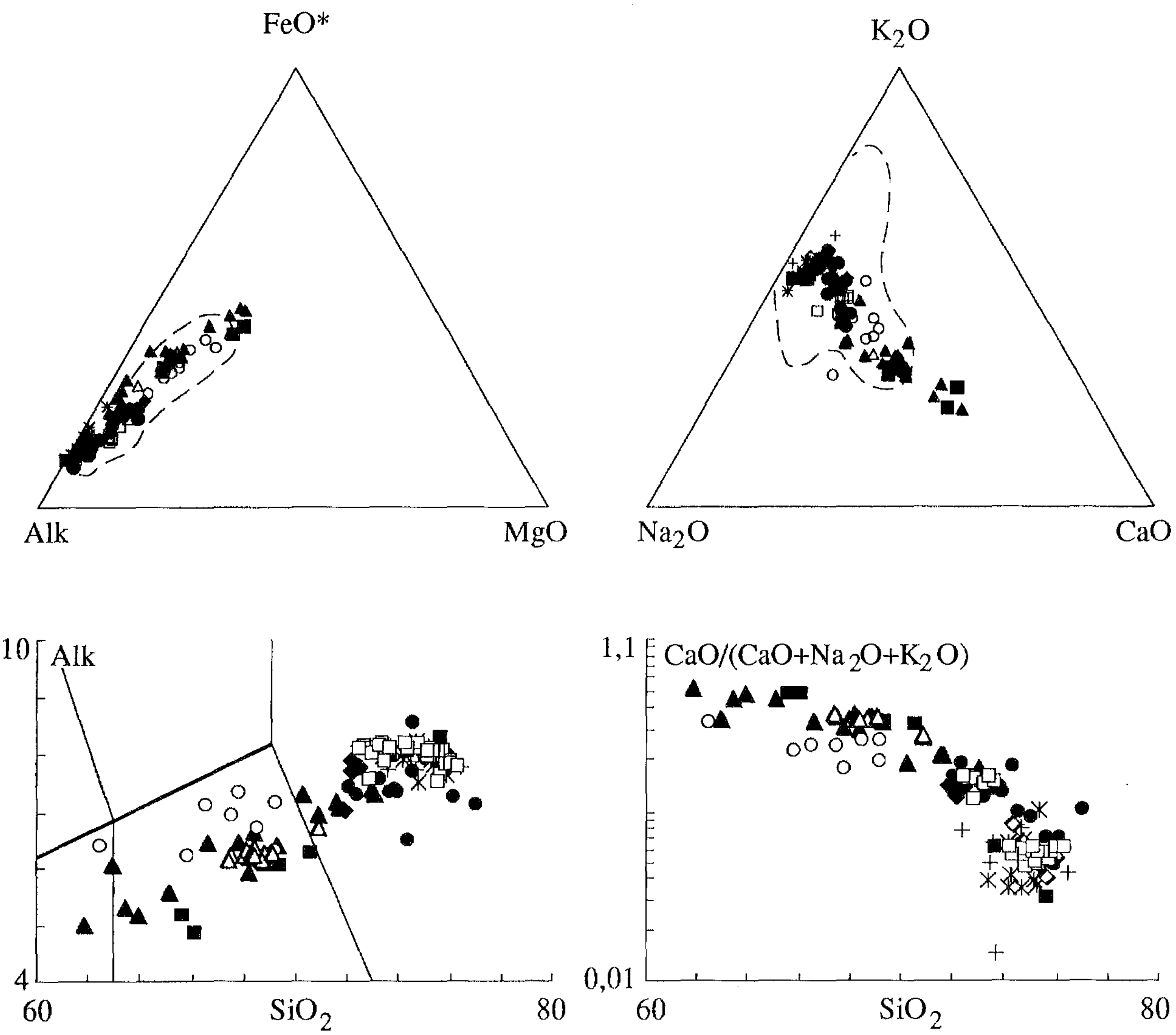

Fig.9 - Other chemical features of the whole set of analysed samples (new data and data taken from the literature). Symbols as in Fig. 8.

normalized REE pattern (normalization factors from Evensen et al., 1978) in order to detect possible records of metamorphic mobilization. The REE patterns are regular and very similar within each sample group (or sub-group), the slight spread being within the analytical error. We interpreted this similarity and regularity as an evidence that an important REE mobilization did not occur during metamorphism in our case: therefore we may assume that the REE contents and distribution patterns in the analyzed rocks conform to those of the protoliths.

The normalized REE patterns for each group of samples are shown separately in Fig. 10a, c, d. For comparison the patterns obtained by Peccerillo et al. (1979) are also displayed (Fig. 10b).

The Casies data. It is immediately evident that all 4 samples from the Casies body gave the same REE pattern (Fig. 10a) and that these new data are similar to those of group A published by Peccerillo et al. (1979) for the same body (open squares in Fig. 10b).

All the Casies data have fractionated LREE and poorly fractionated HREE. Furthermore, a small but significant negative Eu anomaly occurs, and may be explained by assuming:

— separation of plagioclase during crystallization of the magma;

— origin of the magma through melting of a plagioclase-bearing rock, with a certain amount of plagioclase in the unmolten residuum.

Alternatively or in addition, the Eu anomaly could represent a character inherited from the parent rock.

Some of the more acidic samples studied by Peccerillo et al. (1979) from the same Casies massif show a peculiar pattern (black squares in Fig. 10b), with a significantly higher HREE fractionation. These authors suggested a model to explain this peculiar behaviour, based on the fact that garnet is one of the few very common rock-forming minerals that can fractionate HREE: partial melting of a garnet-bearing parent rock, leaving a garnet-bearing

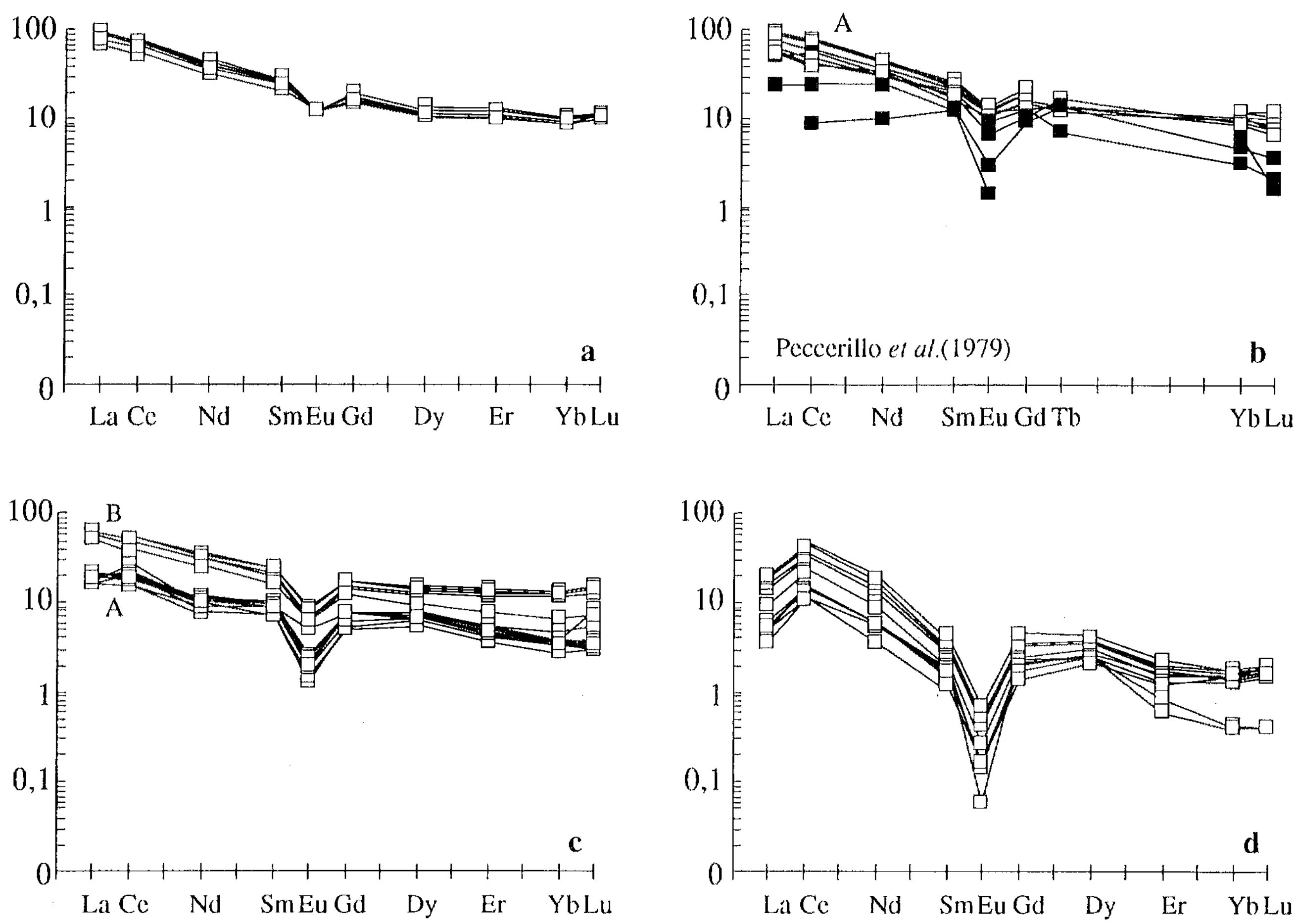

Fig.10 - Chondrite-normalized REE patterns in the orthogneisses. (a) Casies; (b) Peccerillo et al. (1979); (c) Oetztal; (d) Pitztal.

residuum, gives a REE pattern similar to the observed one. Alternatively, but less probably, the magma could have fractionated garnet.

The Oetztal data. The REE data from the Oetztal gneisses clearly show the two-modality detected by the major elements (Fig.10c): two sub-groups, one more acidic with lower REE content, may be recognized within the Oetztal samples. However, both of them show a poor LREE fractionation and a poor or no HREE fractionation. The negative Eu anomaly is slighty stronger in the more acidic rock types. As also shown in Fig. 11, there is a negative correlation between total REE content and differentiation index.

These data agree with a model in which these two different rock subgroups are considered as anatectic products at two different degrees of melting of the same apatite-bearing parent rock.

As well known, solid/liquid partitioning coefficients of common rock-forming minerals are normally < 1: therefore, partial melting of a rock made up of these minerals gives a liquid richer in LREE than the parent rock; and the amount of REE decreases as melting increases.

In the case of partial melting of an apatite-bearing rock, an initial melt relatively poor in REE may be formed (sub-group A) when apatite remains in the solid residuum; when apatite begins to melt, a sudden increase of REE occurs in the liquid phase (sub-group B). The amount of SiO_2 in the samples of the two sub-groups fits with this model.

Moreover, both these sample subgroups are more acidic than the Casies samples and are similar to the Pitztal samples. If we assume for their genesis, alternatively, a partial melting of garnet-bearing parent rocks, we may expect in a primitive magma (like the one that should be related to the Oetztal granitoids) a HREE fractionation even stronger than that actually observed in the Casies samples. Therefore, a garnet-free parent rock is more appropriate for the considered Oetztal granitoid samples.

The Pitztal data. The REE patterns from the Pitztal samples (Fig. 10d) are very similar to those obtained from the Oetztal samples, although they also display, in addition, a peculiar pattern.

This peculiar behaviour is similar to the above described peculiar pattern detected by Peccerillo et al. (1979) in the Casies gneisses (Fig. 10b: black squares): a strong HREE fractionation and a significantly high negative Eu anomaly.

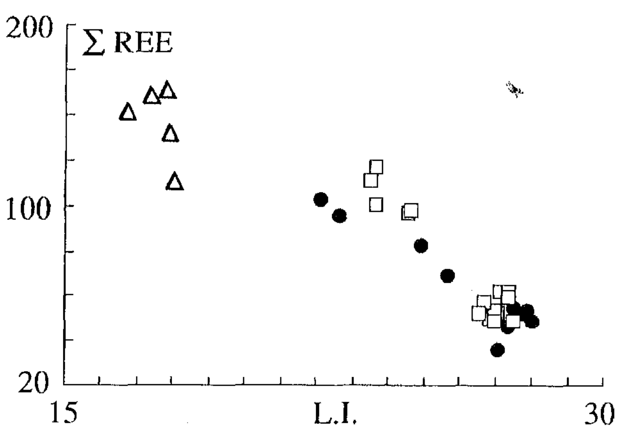

Fig.11 - ∑ REE contents vs. Larsen Index (L.I.) in the orthogneisses. Symbols as in Figs. 2 and 5.

Also in this case a process of melting involving a garnet-bearing parent rock and a garnet-bearing solid residuum can explain the observed REE pattern.

The other REE pattern shown by the Pitztal orthogneisses can be explained considering a garnet-free parent rock, as discussed for the Oetztal samples.

4. Concluding Remarks

Combining the new geochemical data presented in this paper with data taken from the literature, a more complete picture of the chemical character of the "Caledonian" magmatism has been obtained and the genetic models previously proposed by Peccerillo et al. (1979), Sassi & Zirpoli (1979) and Bellieni & Sassi (1981) are better focused, on the basis of the following remarks.

1. The Upper Ordovician plutonism and volcanism display identical features, so that they can be considered syngenetic, i.e. related to a unique cycle of magma generation.

2. This magmatic cycle only produced a huge amount of acidic melts, which display calc-alkaline affinity and cover almost continuously the 62-77 wt% SiO_2 range, without any significant gap: the lack of cogenetic basic rocks, the very small amount of intermediate rocks, and the constraint given by the REE contents and patterns, represent key-elements for the genetic interpretation.

3. Crustal anatexis seems to be the most appropriate process for explaining all petrographic, geochemical and field data. The subcrustal contribution suggested by the initial $^{87}Sr/^{86}Sr$ data (Borsi et al., 1973; Satir, 1975) does not necessarily imply direct mantle contributions, but only incorporation of mantle-derived crustal materials during anatexis.

4. In referring all protoliths of the considered rocks to a unique magmatic cycle, we do not imply a unique mass of magma, parts of which should have fed the volcanic activity, and parts should have been intruded to give numerous plutonic bodies; we mean that several, physically distinct patches of melts formed in different places in a relatively short time under identical constrained conditions at expenses of similar parent rocks, so that all these patches of melts could have taken similar geochemical features.

5. Sassi & Zirpoli (1979), Borsi et al. (1980) and Sassi et al. (1985) showed that the emplacement of the granitoid bodies developed in a relative short time; similar conclusions was given by Flajs & Schönlaub (1976) and Schönlaub (1979) for the emplacement of the volcanic products. Therefore this magmatic activity is to be considered as a short-lasting event.

6. The production in a short time of a huge amount of melts within a relatively large crustal volume requires a peculiar thermal situation and an appropriate geodynamic scenario. From the thermal point of view, we must admit that the involved Ordovician crust was affected by a quantitatively important increase of heat flow. From the geodynamic point of view, this thermal situation may be explained in a context of tensional stresses and related crustal thinning; a scenario which also explains the rise through the crust, within a short time range, of the numerous masses of high viscous melts, and their outpouring to feed the widespread Upper Ordovician volcanism.

Acknowledgements

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Howchinia bradyana (HOWCHIN) EMEND. DAVIS 1951 AND *Mediocris mediocris* (VISSARIONOVA) WITHIN THE BOMBASO FORMATION, CARBONIFEROUS OF THE CARNIC ALPS

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Abstract

Howkinia bradyana and *Mediocris mediocris*, both Late Visean - Earlier Namurian marker fossils, have been found within the Bombaso Fm. (Westfalian, Carnic Alps) in some reworked clasts, relicts of a Variscan carbonate platform till now unfound in the region . The microfacies show that *H. bradyana* seems to belong to a more protected habitat than *M. mediocris.*

Key words: *Foraminifers, Carboniferous, Southern Alps.*

The Bombaso Fm. has been formally defined by Venturini (1990). It can correspond, partially or entirely, to the Waidegg Fm. of the Austrian Authors (Fritz, Boersma & Krainer, 1990) and it is constituted by the rudites of the earlier "tardo-Hercynian sequence" of the Auernig Group, transgressive on the uplifted Paleocarnic Chain. The age of the Formation is Westfalian B - C (Moscovian).

The section outcropping near "Casera Auernig", immediately to the South of the Auernig Mt. (Fig. 1b), one of the classical exposures of the upper part of the Formation, shows two conglomerate beds, separated by up to 100 m of pelitic sediments. The higher bed includes a few limestone clasts with sizes varying from 10 to 25 cm. Only some of these reworked rocks present fossiliferous microfacies with Foraminifer associations.

Two sorts of Carboniferous assemblages have been found. One is characterized by the presence of *Howchinia bradyana*, a marker-fossil of the Late Visean (and Early Namurian?), and the other one by some primitive Fusulinids belonging to the genera *Endostaffella*, *Eostaffella*, *Mediocris*, *Dainella* and *Pseudoendothyra*.

There appears to be a definite association of some Foraminifers with the type of lithology. Infact the *H.bradyana* assemblage occurs only in mudstones and mudstone-wackestones, while the Fusulinids in grainstones, packstones and wackestones.

Mudstone and mudstone-wackestone microfacies present *H.bradyana* accompanied by *Diplosphaerina inaequalis* Derville, D.*sphaerica* Derville, *Endothyranopsis crassa* (Brady), *Endothyra* sp., *Haplophragmina loeblichi* Conil & Lys and rare *Nodasperodiscus* sp.

H.bradyana is contained also in a coral-builtstone pebble with rare *Pseudolituotubella* sp. and very rare *Endothyra* sp.

Grainstone-, packstone- and wackestone-microfacies contain *Tetrataxis* sp. ex gr. *T. conica* Ehrenberg, *Valvulinella* sp., *Pseudolituotubella tenuissima* (Vdovenko), *Lituotubella glomospiroides* Rauser-Chernoussova, *Quasiendothira nibelis* Durkina, *Endothyra omphalota* Rauser & Reitlinger, *E. apposita* Ganelina, *Endothyranopsis crassa* (Brady) some *Nodasperodiscus* with stellate central region and primitive Fusulinids as *Endostaffella fucoides* Rozowskaja, *Eostaffella* cf. *parastruvei* Rauser-Chernoussova, *Mediocris mediocris* (Vissarionova), *M.* sp., very numerous *Pseudoendothyra* sp. and very rare *Dainella* cf. *elegantula* Brazhnikova.

In spite of their differences both associations seem to be of Middle-Late Visean (V2-V3) - Early Namurian age. H. *bradyana* is useful marker of the British Asbian and Brigantian stages (Jenkins & Murray, 1981) and of the equivalent (?) Oka and Sepukov of the Russian Early Carboniferous (Reitlinger, 1956).

Among the Fusulinids, *Danella* belongs to the earlier Visean stages, but the "*Mediocris* zone" marks the Late V2 and V3 in the Uralian Primoria Carboniferous (Rozowskaja, 1975) (Fig. 1d). In addition both association contain stellate *Nodasperodiscus*, typical from V3 to Earlier Namurian.

If all these pebbles have the same age, the differences of the microfauna could probably be due to the habitat condition and to the type of sediment.

Therefore H. *bradyana* assemblage seems to indicate a more sheltered sea-habitat than the one with the Fusulinids.

Limestone clasts with H. *bradyana* have been found by Flügel & Schönlaub (1990) within the Badstub Breccia of the Nötsch serie some kilometres to the North of the Pramollo-Nassfeld Basin, North of the Gail Line (Austria).

These clasts have been related to a Visean-Early Namurian carbonate platform developed on the Variscan margin of the Southern European plate (Flügel & Schönlaub, 1990), wich till now has not been found probably because of tardo-Hercynian tectonic events (Flügel & Schönlaub, 1990; Vai, 1991).

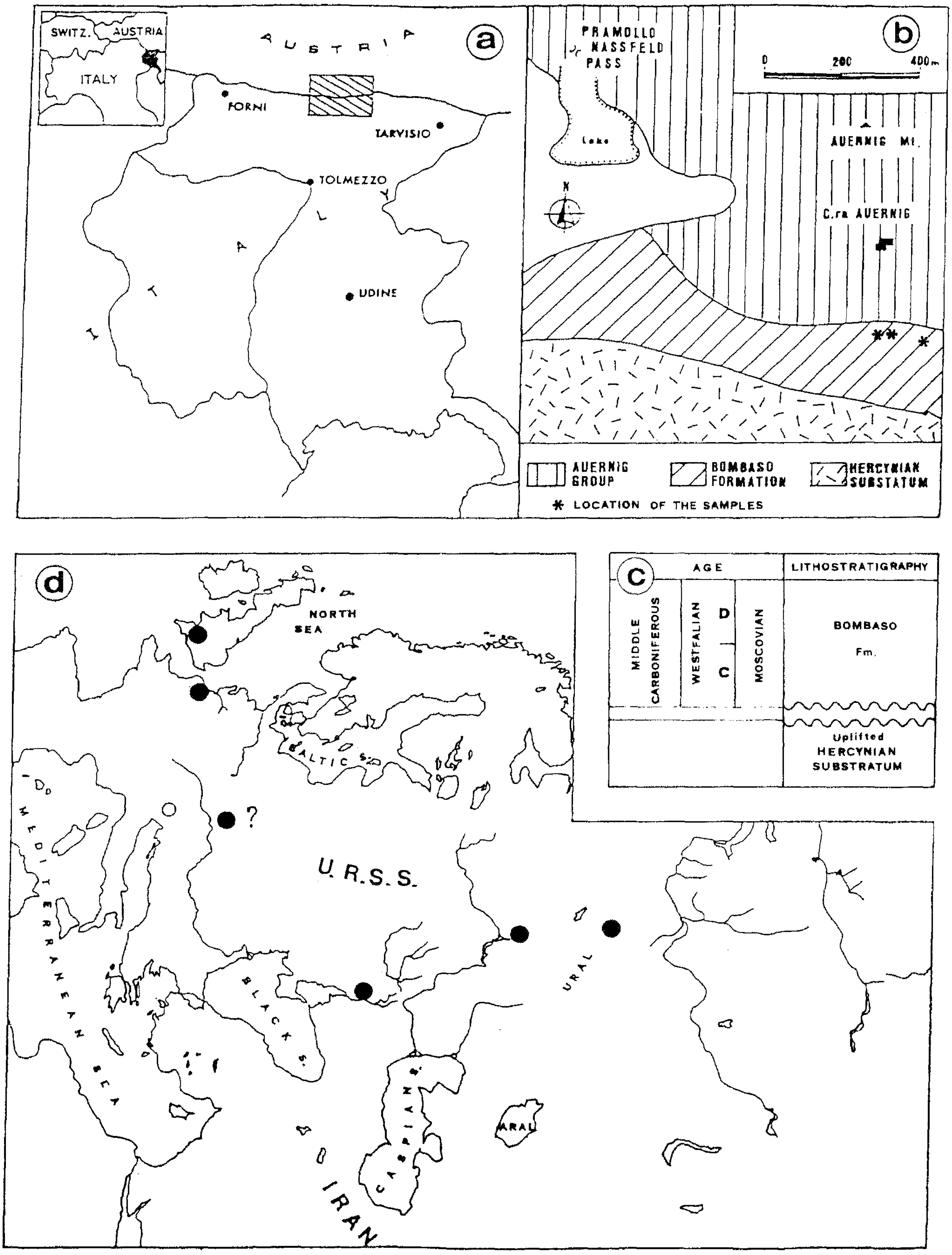

Fig. 1 - a) Location of the Pramollo-Nassfeld area; b) Simplified geologic map of the outcrops of the Bombaso Fm.; c) Simplified stratigraphic position of the Bombaso Fm.; d) Location of the till now well know findings of the *Howkinia bradyana* and/or *Mediocris mediocris* assemblages (filled circles) and location of the new findings (open circle).

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THE AGORDO BASEMENT (NE ITALY): A 500 MA - LONG GEOLOGICAL RECORD IN THE SOUTHALPINE CRUST

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Abstract

This paper describes and debates the ten main stages of the geological evolution of the Agordo metamorphic core (Hercynian basement of the Dolomites, eastern Southern Alps), when possible set in a geodynamic framework on a regional scale. The events are arranged as follows.

1. Sedimentation of Acritarch-bearing pelitic-psammitic deposits (Col di Foglia Fm.) and other epicontinental terrigenous units of the Lower Group: Cambrian.
2. A regional-scale tectonic event, possibly coeval to the Sardic phase: Early - Middle Ordovician.
3. Emplacement of acidic and basic volcanics and deposition of terrigenous units of Upper Group: Late Ordovician - Silurian.
4. First Hercynian phase (D_1), developing metamorphism (from Chl- to Alm-subfacies), axial-plane foliations, folding from micro- to regional scale, two ductile shear zones and three thrust sheets: Early Viséan.
5. Uplifting to a shallower crustal level: between D_1 and D_2.
6. Second Hercynian phase (D_2), developing metamorphism (from Chl- to Bt-subfacies), axial-plane foliations, folding from micro- to regional scale, one ductile shear zone and two east-vergent thrust sheets: Middle Namurian.
7. Uplift and first exhumation, without important effects of external rotation and/or tilting: Westphalian - Stephanian.
8. Slow burial, not exceeding 1.5 kbar and 150°C: from Earliest Permian until Early Miocene.
9. Thrusting towards SE along Valsugana thrust (D_3), developing faults and kinks, northwestward tilting and a gentle antiform, and second exhumation: Serravallian - Late Pleistocene.
10. Sinistral wrenching along subvertical upper ramp of Valsugana thrust (D_4), reactivating faults and developing kinks, en èchelon and drag folds from meso- to regional scale and a second gentle antiform: Middle (?)-Late Pleistocene - Holocene.

Key words: *pre-Hercynian, Hercynian and Alpine history, eastern Southalpine basement, Neogene-Quaternary Southalpine Chain, Eastern Alps, NE Italy.*

1. Introduction

The metamorphic core of Agordo is part of the Hercynian basement of the Dolomites and outcrops along the Valsugana thrust, about 20 km NW of Belluno (Fig. 1), in the middle of the Southalpine basement of the Eastern

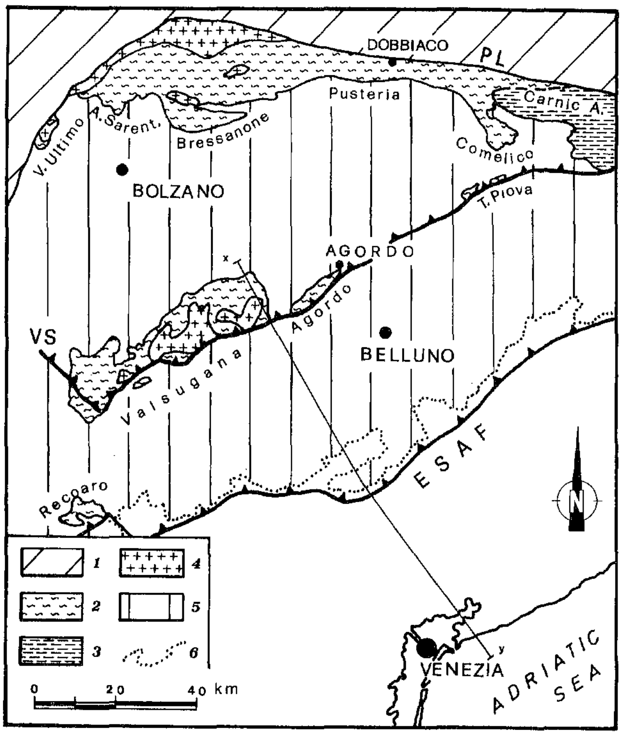

Fig. 1 - Agordo and other outcrops of Southalpine basement of Eastern Alps. Legend: Austroalpine (1); Southern Alps: metamorphic basement (2), non-metamorphic or very-low metamorphic Paleozoic of Paleocarnic Chain (3), late-Hercynian intrusive bodies (4) and non-metamorphic, late- and post-Hercynian cover (5); boundary of Veneto-Friuli plain (6); Periadriatic Lineament (PL); Valsugana thrust (VS); front of Neogene-Quaternary eastern Southalpine thrust-belt (ESAF); trace of geological cross-section of Fig. 10 (x - y).

Alps (ESB). This core consists of five metapelitic, metapsammitic and sometimes metapsephitic units and of one acidic and one basic metavolcanic unit (Fig. 2). The age of the protoliths ranges from Cambrian to Silurian. Their metamorphic imprint is linked to the two Hercynian tectono-metamorphic phases, both of which developed in the greenschist facies.

The Hercynian metamorphic framework is non-conformably covered by the non-metamorphic late-Hercynian molasse (Lower Permian) and by the Alpine sequences of the eastern Southern Alps (Upper Permian-Holocene) (Fig. 1).

During the post-collisional evolution of the eastern Southern Alps, the Agordo basement was involved in the development of the Neogene-Holocene Southalpine Chain. Therefore, the rocks of Agordo also show the effects of two late-Alpine deformations.

The aim of this paper is to define the evolution of the Agordo basement over this 500-Ma period and to debate still open problems.

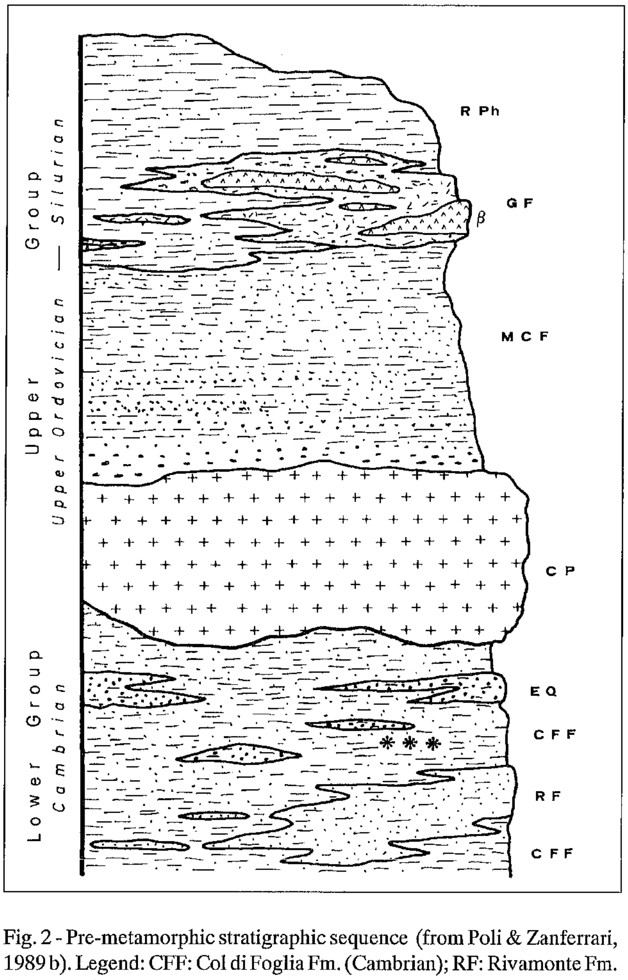

Fig. 2 - Pre-metamorphic stratigraphic sequence (from Poli & Zanferrari, 1989 b). Legend: CFF: Col di Foglia Fm. (Cambrian); RF: Rivamonte Fm. (Cambrian); EQ: Eores quartzite (Cambrian); CP: Comelico Porphyroids (Caradocian); MCF: M.Cavallino Fm. (Ashgillian); GF: Gudon Fm. (Lower Silurian); RPh: Recoaro Phyllite (Silurian); ***: Upper Cambrian Acritarchs; ß: basalts. Not to scale.

2. Sedimentation of Lower Group Protoliths

The Lower Group is composed of three metapelitic-psammitic formations (Poli & Zanferrari, 1989). Their protoliths were (Fig. 2): 1) pelites and minor psammites (*Col di Foglia Fm.*), 2) psammites and sometimes microrudites (*Eores Quartzite*) and 3) mainly psammites (*Rivamonte Fm.*). The first two formations are rich in carbonaceous matter and Fe-Cu sulfides, the third is very poor.

Acritarch assemblages of Late Cambrian age have been found in some pelitic levels of the Col di Foglia Fm. (Kalvacheva et al., 1986; Sassi et al., 1984).

From field evidence, the Eores Quartzite is a local, more psammitic, lateral change of the Col di Foglia Fm., which is the most widespread unit. Moreover, between the Rivamonte and Col di Foglia Fms., vertical and lateral transitions exist, while primary contacts between the Rivamonte Fm. and Eores Quartzite have never been observed.

The occurrence of Fe-Cu sulfides (mainly pyrite and chalcopyrite) is interpreted as pre-metamorphic strata-bound mineralization of Kieslager-type (Brigo, 1977; Brigo & Omenetto, 1979).

Owing to their lithological, stratigraphic, micropaleontological and metallogenic features, we believe that the terrigenous protoliths of the Lower Group were deposited on an epicontinental shelf.

The overall age of the Lower Group also probably includes the rest of the Cambrian, while possible extension to the Early Tremadoc is not excluded (Kalvacheva et al., 1986).

Lastly, the occurrence of Kieslager-type mineralizations is a peculiar feature (Brigo, 1971) of much of the "Bressanone (Brixen) phyllite", widespread in the western ESB. This fact and the close stratigraphic relationships of the Eores Quartzite with both the Col di Foglia Fm. and the "Bressanone phyllite" allow us to correlate the two latter units (Zanferrari, in prep.).

3. Sardic Phase ?

As a first basic problem, the stage between the Latest Cambrian (or Early Tremadocian ?) and the Caradocian must be discussed. This stage ranges between sedimentation of the Acritarch-bearing sequences and deposition of the parent rocks of the Comelico Porphyroids (Fig. 2).

The question is divided into two parts:

1. is there a gap which includes the Lower and Middle Ordovician ?

2. could a dynamic event (possibly also metamorphic) have developed in this time ?

As regards the Agordo basement, there is the following evidence:

a) absence of Lower and Middle Ordovician sequences;

b) lack of pre-Hercynian metamorphic signatures.

Therefore, we may assert that in Agordo a pre-Hercynian event might have occurred, but only tectonic in type.

Of course, the question cannot be solved in the small Agordo core, but must be examined on the scale of the whole ESB at least. The following elements exist.

c) There is no proof for or against the existence of Lower and Middle Ordovician sequences.

d) In the whole ESB, neither micro- nor mesoscopic evidence for a pre-Hercynian metamorphic imprint has yet been found.

e)However, in some areas of the ESB, under the former subaerial surface of deposition of the Comelico Porphyroids, some formations outcrop which in other cores are deeper, because of the interposition of one or more different units. This fact suggests possible pre-Hercynian tectonic reworking of the Lower Group sequences caused by folding on a regional scale.

f) Lastly, in the Pusteria basement near Brunico (Bruneck, Pustertal) a single radiometric age of 463 Ma has been recorded (Del Moro et al., 1984). These Authors believe this value is proof of "Caledonian" metamorphism in the Pusteria phyllites. Instead, it is more likely that this age only represents the thermal effect of some dacitic subvolcanic bodies on the surrounding pelitic rocks. Our interpretation is based:

— on point d);

— on the fact that the dacitic bodies belong to the Upper Ordovician magmatic cycle (Del Moro et al., 1984) as well as the Comelico Porphyroids (Bellieni & Sassi, 1981);

— on the fact that this age (463 Ma: Latest Llandeilian) is very close to the Caradocian age of emplacement of the Comelico Porphyroids (Sassi et al., 1979; Heinisch, 1981).

Summing up, everything strengthens the hypothesis that there is a gap which more or less completely covers the Early and Middle Ordovician between the Lower and Upper Groups. Furthermore, there are some indications that this gap may be linked with a non-metamorphic folding event on a regional scale.

We suggest that this Middle-Early Ordovician phase corresponds to the coeval Sardic Phase, well recognized in the basements of Sardinia and Tuscany (Carosi & Gattiglio, 1990; Gattiglio et al., 1990 and references therein).

4. Deposition of Upper Group Protoliths

Four metavolcanic, volcanoclastic and/or terrigenous units have been recognized in the Agordo basement (Poli & Zanferrari,1989), as well as in the whole ESB (Zanferrari, in prep.), forming the Upper Group.

The pre-metamorphic sequence (Fig. 2) starts with rhyolitic-rhyodacitic ignimbrites and related volcanoclastics (*Comelico Porphyroids*) of Caradocian age (Sassi et al., 1979; Heinisch, 1981). They are followed by a fining-upward, acidic volcanoclastic-terrigenous sequence of often graded micropsephytes and psammites, and pelites (*M. Cavallino Fm.*), Ashgillian in age (Sassi et al., 1979).

Two heteropic formations of probably Silurian age lie on the M. Cavallino Fm.: the *Gudon Fm.* and the *Recoaro Phyllite.*

The *Gudon Fm.* was originally formed of alkaline basalts and related volcanoclastics, closely interfingered with prevailing pelites, locally with an important carbonate content, and psammites, probably Early Silurian in age (Visonà & Zanferrari, 1987; Poli & Zanferrari, 1989; Poli et al., in prep.).

Pre-metamorphic, stratiform polymetallic mineralizations (Fe, Cu, As, Sb, Sn, Zn, Pb, Ag, Co, Ni, Au) are typical of the Gudon Fm. in Agordo as well as throughout the ESB (Brigo, 1971, 1977; Brigo & Omenetto, 1979). In the Agordo basement, they formed the famous Valle Imperina orebody (mainly pyrite and chalcopyrite, with minor galena, black sphalerite, arsenopyrite, cassiterite), intensely mined from at least 1420 until 1962 (Omenetto, 1968).

The *Recoaro Phyllite* is partially heteropic with the Gudon Fm., as shown in Fig. 2. Its protoliths consist of pelites and minor psammites with a medium-low carbonaceous content. Minor, pre-metamorphic, stratiform Fe-sulfide occurrences are widespread.

The higher pelite/psammite ratio and the lower contents in carbonaceous matter and Fe-sulfides are basic guidelines distinguishing the present Recoaro phyllites from those of the Col di Foglia Fm.

A sedimentation age probably comprising most of the Silurian may be inferred from stratigraphic relationships with the M. Cavallino and Gudon Fms. and from regional correlations with similar sequences in the Eastern Alps and Sardinia.

The depositional environment of the Upper Group sequences is inferred from the following points.

1) Revision of the Comelico Porphyroids type-section (Poli & Zanferrari, in prep.) supports the absence of intercalations of phyllites (assumed to be of marine origin) within the porphyroids; thus, the whole sequence is formed of primary ignimbrites and pyroclastics deposited in a subaerial environment (Heinisch, 1981).

2) From lithological features, the parent rocks of the M. Cavallino Fm. are closely related to subaerial erosion of acidic volcanics (i.e., the present porphyroids), suggesting sedimentation in land environments (mainly alluvial plain, alluvial fan and probably also lacustrine). Analogies between the protoliths of the Upper Ordovician Comelico Porphyroids + M. Cavallino Fm. and the Permian Bolzano Volcanic Complex + Val Gardena Sandstone (Bozener Quarzporphyr + Gröden S.) are impressive from the lithological and sedimentological points of view.

3) The lack of radiolarian chert ("lydites") within the phyllites of the M. Cavallino, Gudon and Recoaro Formations in both the Agordo basement and the entire

ESB suggests sedimentation of the terrigenous protoliths in a shallow-water environment.

4) The lateral and vertical relations between terrigenous deposits and mafic volcanoclastics of the Gudon Fm. (Poli & Zanferrari, 1989) and the large dispersal of the volcanoclastics require a subaqueous (but not necessarily deep-water) environment.

5) The intraplate alkaline basalts of the Gudon Fm. indicate an extensional regime during the Earliest Silurian in the crust of the present ESB (Poli et al., in prep.), a common situation in many Mediterranean and South-European Paleozoic areas.

However, this first Hercynian rifting was scanty, as suggested by the small amount of basaltic lavas. Deep basins did not form, but only shallow depressions, still in the neritic environment.

6) In Pusteria, from Brunico eastwards and in the Western Carnic Alps, the Recoaro Phyllite is primarily overlain by Silurian p.p. - Devonian sequences of carbonate-platform facies.

Accordingly, we believe that, in Agordo as in the whole ESB, the protoliths of the Upper Group were deposited on a peneplain which evolved to a slowly subsiding, terrigenous, epicontinental shelf. Carbonate platform(s) locally developed from the Silurian p.p. - Devonian onwards. That is, deposition began on land with the protoliths of the Comelico Porphyroids and probably most of the M. Cavallino Fm.. Neritic conditions certainly started from the Ordovician-Silurian boundary, with moderate bathymetric differentiations especially in the Early Silurian.

These shallow-water conditions persisted throughout the Silurian, allowing the establishment of Silurian p.p. - Devonian carbonate platform(s) at least in the NE zone of the ESB.

5. First Hercynian Tectono-Metamorphic Phase

The metamorphic effects of phase Ph_1 are radiometrically dated at around 350 Ma (Cavazzini et al., 1991; Del Moro et al., 1980, 1984), corresponding to the Early Viséan. Ph_1 is recorded by greenschist-facies mineral assemblages (D'Amico, 1962; Poli & Zanferrari, 1989). The metamorphic grade increases from NE (Qtz-Ab-Ms-Chl subfacies) to SW (Qtz-Ab-Ep-Alm subfacies) owing to the present attitude of the Ph_1 structures (Fig. 3a).

The mineralogical compatibilities of the Qtz-Ab-Ep-Alm subfacies are shown in Fig. 4.

The main synkinematic minerals are Qtz + Ab + Chl + Ms ± Bt ± Cld ± Alm in the metapelites, metapsammites and metarhyolites, and Ab + Ep + Act/Tr ± Hbl ± Bt in the metabasites. Only Ab has been definitely identified as a Ph_1 postkinematic mineral.

Widespread mylonites characterize all lithotypes, especially in two hectometric shear zones (Fig. 3a).

Penetrative axial-plane foliations, ranging from well-developed slaty cleavage to schistosity, are coupled with

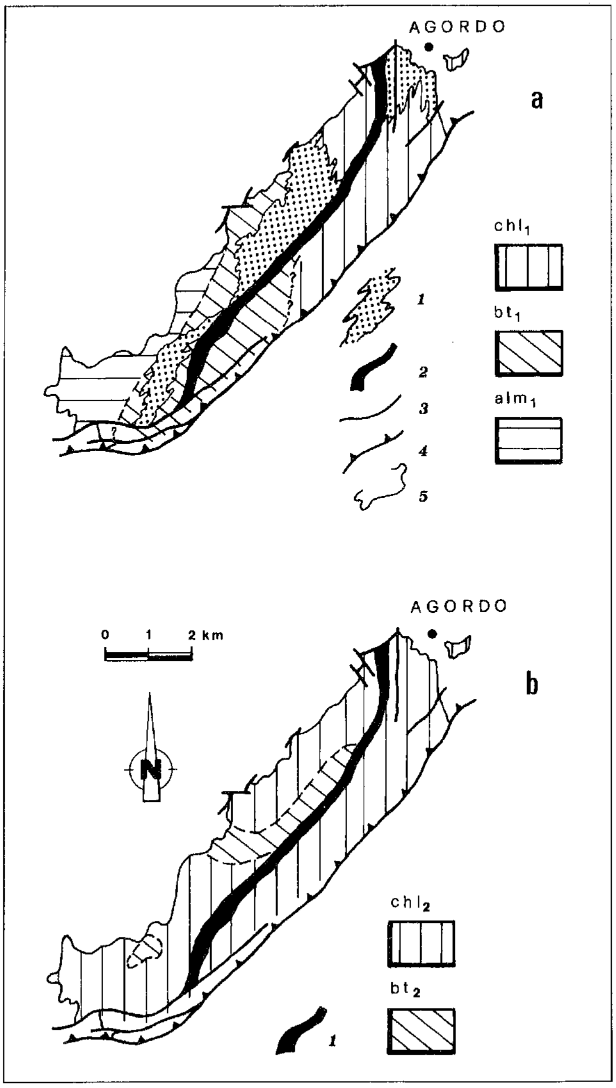

Fig. 3 - Sketch of Agordo basement (from Poli & Zanferrari (1991 b), modified). a: metamorphic zoneography (Chl_1, Bt_1, Alm_1) and two shear zones (1) of Ph_1, stretched and overturned eastwards along D_2 shear zone (2); main Alpine faults (3); Valsugana thrust (4); cover-basement boundary (5). b: zoneography (Chl_2, Bt_2) and shear zone (1) of Ph_2.

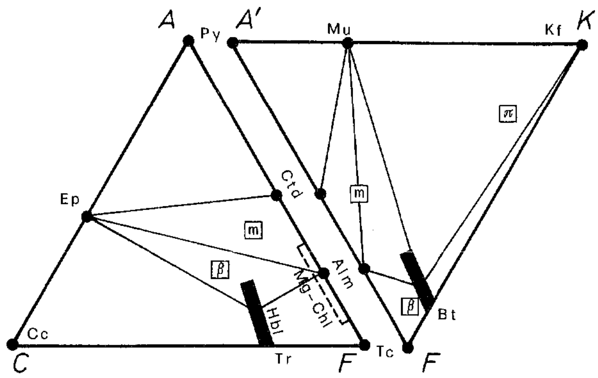

Fig. 4 - Mineralogical compatibilities of Qtz-Ab-Ep-Alm subfacies of Ph_1 phase in metapelites (m), metabasites (ß) and metarhyolites (π).

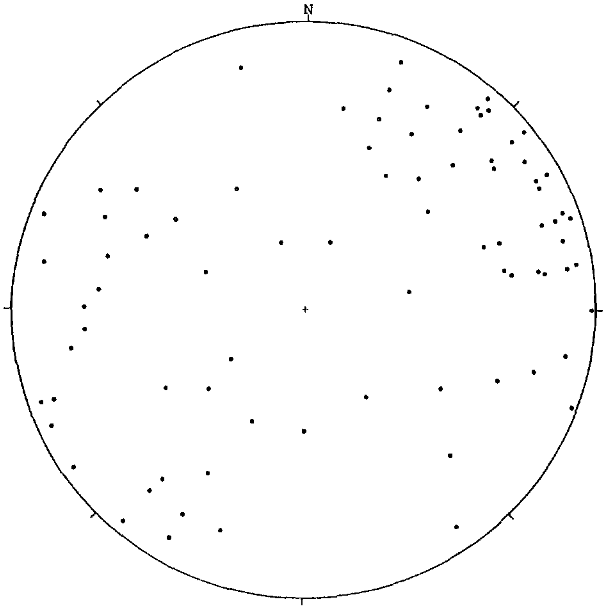

Fig. 5 - Attitudes of 80 axes of F_1 mesoscopic folds (Schmidt net, lower hemisphere).

the F_1 fold system (Poli & Zanferrari, 1991b).

The F_1 folds are always isoclinal and were originally probably recumbent. They have been recognized to the hectometric scale, and on all scales they are often rootless. B_1 axes are scattered along the NE-SW direction (Fig. 5).

The two ductile shear zones separate three thrust sheets, at present stretched and overturned towards the east along the Ph_2 shear zone (Fig. 3a). The relationships between mylonites and folds and foliations of Ph_1 show that the mylonitization developed towards the end of the Ph_1 metamorphism.

The above-mentioned features suggest a regime of non-coaxial deformation through probably subhorizontal simple shear.

6. Uplift to Shallower Crustal Level

There are no records of the evolution of the Agordo basement between the two Hercynian tectono-metamorphic phases.

We only know that phase Ph_2 occurred at a shallower crustal level than Ph_1, as inferred from the Ph_2 assemblages (see section 7). This uplift may be referred to a hypothetical geodynamic process between Ph_1 and Ph_2, which brought about crustal uplift and erosion, or to Ph_2, as an initial effect of the process of piling up and crustal shortening.

7. Second Hercynian Tectono-Metamorphic Phase

The metamorphic effects of Ph_2 in the ESB are radiometrically dated at around 320 Ma (Del Moro et al., 1980, 1984), corresponding approximately to the Middle Namurian.

In this case too, the metamorphic imprint belongs to the greenschist facies (from Chl- to Bt- subfacies). Bt-subfacies rocks outcrop in the central area (Fig. 3b).

The main synkinematic minerals are Qtz + Ab + Ms ± Chl ± Bt in metapelites, metapsammites and metarhyolites, and Ab + Ep ± Act ± Bt in the metabasites. The postkinematic minerals are mostly Ab ± Ms ± ore minerals.

Retrometamorphic effects on Ph_1 parageneses brought about the almost complete alteration of Alm_1: only garnet relics in decussate chlorite pseudomorphs have been observed.

Mylonitic processes also developed during Ph_2, but were discontinuous and less intense than in Ph_1. Mylonitic effects are important only within a shear zone about 100 m thick (Poli & Zanferrari, 1991 a).

The mean b_0 value of the potassic white micas (9.000 Å) of the Agordo basement (unpubl. data) coincides with those measured in the rest of ESB (Sassi et al., 1974). All these b_0 values of ESB, which can be referred only to Ph_2, suggest conditions of relatively high thermal gradient.

7.1 *Present Lithology of the Agordo Basement*

A - Lower Group

A.1 - Col di Foglia Fm.

This characterizes the NE and central areas. The unit is formed of very fine-grained metapelites and gray to black phyllites, with Ms ± Chl, rich in graphite and Fe-Cu sulfides. Quartzites and sometimes coarse-grained Ms ± Chl ± Gr metapsammites, greenish to black in colour, are frequently intercalated.

A.2 - Eores Quartzite

This widely outcrops in the southern-central area, and is formed of fine- to coarse-grained Ms + Chl ± Cld quartzites and minor fine-grained phyllites. The rocks are always black and very rich in graphite and Fe-Cu sulfides.

A.3 - Rivamonte Fm.

This is found in the central-eastern area and is formed of silvery-grey, Ms + Chl metapsammites and minor phyllites. Graphite is almost absent and sulfides are scanty.

B - Upper Group

B.1 - Comelico Porphyroids

This unit outcrops over the whole area. It is formed of massive Ms + Chl ± Bt metarhyolites characterized by numerous phenocrysts of Qtz, perthitic Kfs and sometimes Ab. Colours range from grey to greenish or pink.

B.2 - M. Cavallino Fm.

In the NE and central areas fining-upward rhythmic sequences occur. They are formed of sometimes graded Chl + Ms ±Bt metapsammites (with common, small phenocrysts from acidic volcanics), followed by Chl + Ms ± Bt fine-grained metapsammites and phyllites. Colour variability is peculiar to this unit, colours ranging from silvery-grey to pinkish-white, greenish and light grey. The significant content in carbonate (mostly Ank and Sd) is an important tool to distinguish the rocks of this unit.

In the SW zone these sequences consist of white-greenish Ms + Chl ± Bt paragneiss and silvery-grey, fine-grained Ms + Chl ± Bt micaschists.

B.3 - Gudon Fm.

This unit occurs over the whole area and is characterized by several, variously organized lithotypes: epidote-amphibolites (corresponding to the "metadiabasi albitico-epidotico-cloritico-anfibolici" of D'Amico, 1962), chlorite-actinolite schists, often alternating with very thin brownish carbonate levels (Cal + Ank ± Sd), yellowish-gray calc-schists, and dark grey Chl + Ms ± Bt phyllites.

The epidote-amphibolite and the chlorite- actinolite schists are strongly mineralized everywhere (see section 4). The phyllitic levels laterally pass to the Recoaro phyllites and have a middle-low content in graphite and Fe-sulfides or oxides.

B.4 - Recoaro Phyllite

This occurs over the whole basement and therefore shows significant lithological variations passing from the central area to the western one. In the former area, the Recoaro Phyllite appears as gray-greenish or silvery Ms + Chl ± Bt phyllites (sometimes passing to black graphite-rich phyllites), and minor metapsammites. Fe-sulfides and oxides contents are low.

Fine-grained, grey-greenish Ms + Chl ± Bt ± Alm micaschists typically outcrop in the western area. Fe-Cu sulfides, Fe-oxides and graphite are scanty.

7.2 *Tectonic Framework of* Ph_2

Penetrative axial-plane foliation S_2 ranges from slaty cleavage to crenulation cleavage, mostly between the 2nd stage (within massive rocks) and the 4th stage (in lithotypes with more ductile behaviour and/or for higher temperatures) of the scale of Bell & Rubenach (1983).

The present attitude of S_2 foliations is shown in Fig. 6, which also shows the most frequent attitudes of the layering of the Lower Permian sequences along the edge of the

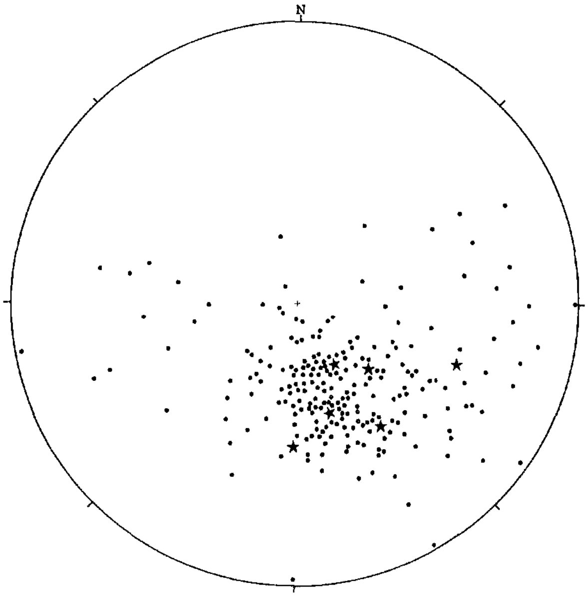

Fig. 6 - Attitudes of 221 poles of S_2 foliation planes (Schmidt net, lower hemisphere). Note folding effects of D_3 (NW tilting of 30°-40° and slight arching with axis approx. N50°) and those of D_4 (gentle antiform with axis approx. 310° and en échelon mesofolds with axes 290°-320°). Stars: most frequent attitudes of layering of Lower Permian sequences along edge of basement in hanging wall of Valsugana thrust.

Agordo core (see also Fig. 10). Since Alpine deformations D_3 and D_4 are weak (sections 10 and 11) and without effects of distortion and change in volume, the attitude of the S_2 system may be restored, although not completely (Poli & Zanferrari, 1991 b). We thus gain clear information regarding

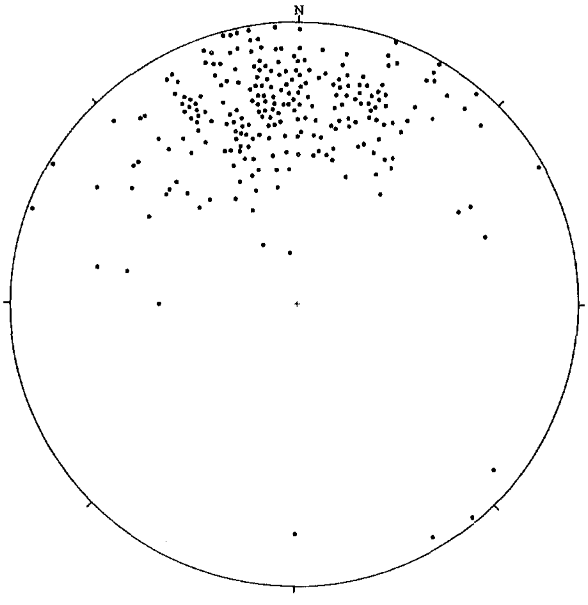

Fig. 7 - Attitudes of 240 axes of F_2 mesoscopic folds (Schmidt net, lower hemisphere). Scattering of poles is linked to D_3 and D_4 late-Alpine deformations.

subhorizontal trends for the planar structures of Ph_2.

The F_2 mesofolds range from isoclinal to open types, mostly depending on lithology, temperature, fluid phase and shear stress level. They are shown on the geological map on a kilometric scale and are rootless on S_2 only within the Ph_2 shear zone.

Owing to the D_3 and D_4 deformations, the B_2 axes of the mesofolds mostly plunge at 20°- 40° towards the north (Fig. 7). Figs. 6 and 7 also reveal that the F_2 folds were recumbent, B_2 axes striking originally N-S.

The interference patterns between F_1 and F_2 mesofolds are mostly of 3 I and 2 H Type (Ramsay, 1967). The angular value between B_1 and B_2 ranges between 10° and 45°, while the AP_1 and AP_2 axial planes form medium-high angles with each other (Poli & Zanferrari, 1991 b).

On the scale of the whole ESB (Fig. 8), the trend of B_1 and B_2 in Agordo is subparallel to that of the Recoaro, Bressanone (Brixen) and Alpi Sarentine (Sarntaler Alpen) areas. In Pusteria and Comelico they are bent clockwise, with B_1/B_2 angles approximately constant. This again confirms the coherent beaviour of the ESB during its Hercynian history.

A single, formerly subhorizontal ductile shear zone has been detected here, separating two thrust sheets of Ph_2. The kinematic indicators (s-c surfaces, σ-type and δ-type porphyroclasts, bookshelf-sliding structures), as well as the stretching and reversal of three thrust sheets and three metamorphic zones of Ph_1 (Fig. 3a), indicate eastward thrusting of the upper unit (Poli & Zanferrari, 1991 a). The minimum value of this thrusting has been calculated at least 5 km long.

Therefore, for the Agordo area too, the existence of thrust sheets of Ph_2, previously reported only from the Bressanone - Alpi Sarentine area (Zanferrari, 1987), is proved.

The fabric of the S_2 foliation and the shear zone again indicates a regime of non-coaxial deformation through subhorizontal simple shear.

With these data available, the geodynamic evolution of the ESB during the second Hercynian phase becomes less puzzling.

1. The eastward and southeastward thrusting of the ESB over the non- or very low-metamorphic Carnic foredeep and the more southerly non-metamorphic foreland (the present Veneto plain and Adriatic offshore: 445 Ma old, granite of the Assunta 1 well near Venice), as hypothesized by Vai & Cocozza (1986), is proved.

2. The shifting of 10-5 Ma between the Ph_2 of the ESB (Middle Namurian, with the approximation of the radiometric datings) and the "Carnic" phase (Westphalian C), which built up the Paleocarnic Chain, may be correlated to the tectonic shifting between the easternmost part of the Hercynian "Internides" (present ESB) and the "Externides" (Paleocarnic Chain), which were finally involved in the orogenic front.

With the end of the second Hercynian tectono-metamorphic phase, the Agordo basement gained its fundamental lithologic and structural characteristics. These features were to remain substantially unchanged throughout the following history, also including the late Alpine phases D_3 e D_4.

8. First Exhumation

In Agordo too the lithologic and sedimentologic features of the basal sequences of the non-metamorphic late-Hercynian molasse (Ponte Gardena (Waidbruck) Conglomerate of Earliest Permian age) demonstrate deposition in an intermontane molasse basin genetically linked to taphrogenesis of Stephanian and Early Permian age (Kreiner, 1989).

This fact leads us to set the Agordo area, as well as the better-known northern part of the eastern Southalpine realm (Venturini, 1983; cf. also Wopfner, 1984 and Massari, 1986), within the Upper Paleozoic network of dextral megashear between the African and American-European (= Laurasian) plates (Arthaud & Matte, 1977).

However, within the small Agordo core no evidence has yet been found of the movements which brought the present basement out of the metamorphic environment and then up to the surface. Particularly, at least until now, it has not been possible to discriminate between the effects of brittle or semi-brittle deformation of this phase and of the following one. This depends on repeated reactivations of faults in the basement, including those which may have been formed during the Late Carboniferous.

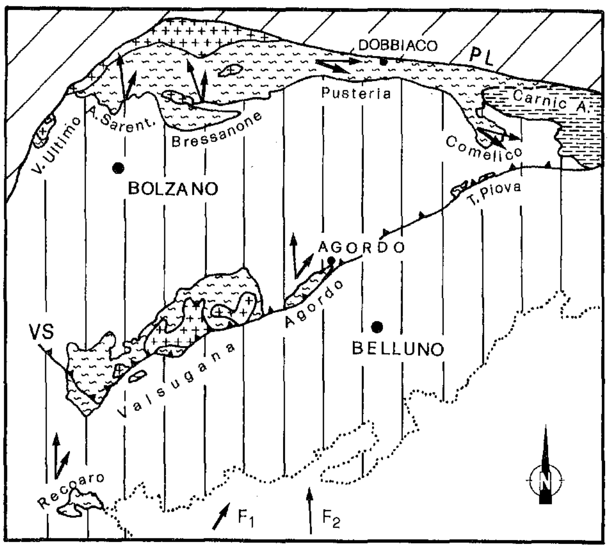

Fig. 8 - Mean trends of axial directions of F_1 and F_2 mesofolds in eastern Southalpine basement (Recoaro: from Blackburn et al. (1968); Agordo: this paper; other areas: unpubl. data). Symbols as in Fig. 1. Note uniform trends of F_1 and F_2 in central and western ESB and their clockwise rotation in eastern ESB and western Carnic Alps.

The problem regards not only this interval and the following one, but also the late Alpine phases D_3 and D_4. Both these intervals and D_3 and D_4 developed at various crustal depths and in different kinematic settings, as shown by:

— cataclasites of different grain-size and/or cementation degree;

— re-crushed cataclasites or tectonic breccias;

— the very frequent superposition of kinematic indicators of different conditions of deformations and structural meaning.

Therefore, a long polyphasic deformation history is revealed in most of the faults of the basement. To sum up, as regards the first exhumation of the Agordo basement, after Ph_2 the Hercynian structural framework did not change, but was locally disturbed by mostly minor faults.

The identical trends of the axes of mesoscopic folds F_1 and F_2 in the central and western ESB (Fig. 8) and the parallelism between the Ph_2 planar structures and layering in the cover (Fig. 6), suggest that the post-Ph_2 uplift of the Agordo basement took place without any important effects of external rotation and/or tilting. At least the end of its upwards path probably occurred in a transpressive regime linked to the above-mentioned late-Hercynian network of dextral megashear.

9. Slow Burial

The slow burial of the basement started with the sedimentation of the late-Hercynian molasse of the Earliest Permian. It lasted up to the Early Miocene, but undoubtedly alternated with positive movements, mostly in the Paleogene.

The maximun thickness of the sequences forming the Alpine cover can be estimated at not more than 4-5 km, corresponding to lithostatic pressures not exceeding 1.2-1.5 kbar and temperatures of 120°-150°C, which may have affected the present basement during maximum burial.

As already mentioned in the previous section, it is not possible to distinguish related structures, which in any case can only be of brittle or semi-brittle type.

Lastly, it should be stressed that the small metamorphic core is affected by none of the large normal faults which dismembered the Southalpine realm in basins and shelfs during the Mesozoic E-W extension.

10. Thrusting Toward SE and Second Exhumation

Starting from the Serravallian (Massari et al., 1986), the Agordo basement was also involved in strong uplift movements which accompanied the large horizontal displacement towards the SE on the listric surface of the N50°-striking Valsugana thrust (Fig. 9).

This caused the Agordo basement to be exposed again during the Late Pleistocene, probabily during the Würm.

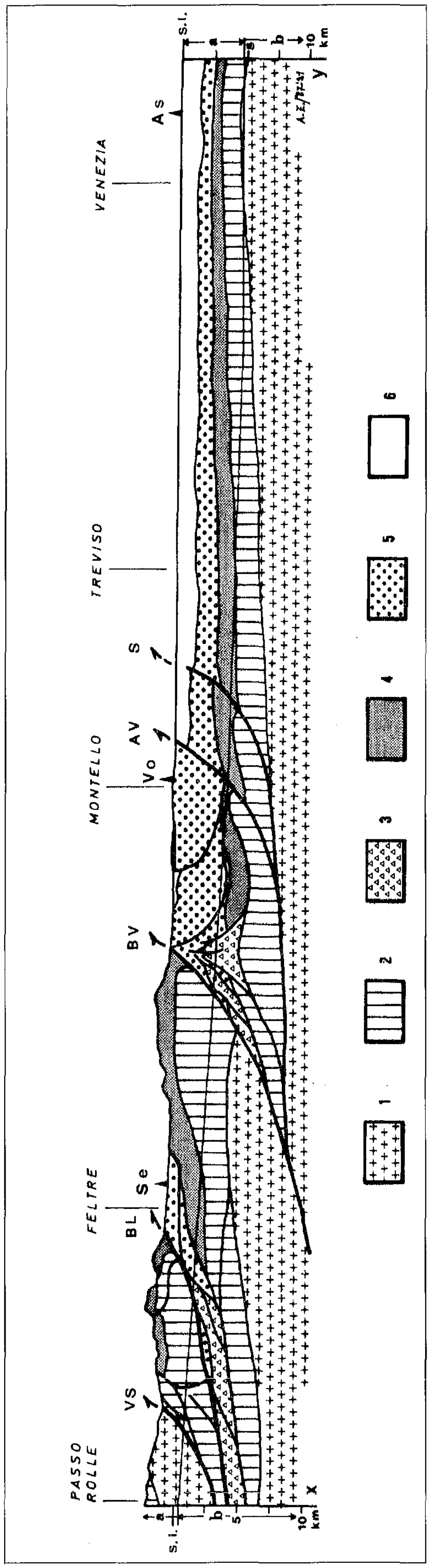

Fig. 9 - Geological cross-section through Neogene-Quaternary eastern Southalpine thrust-belt, from Passo Rolle (Dolomites) to Assunta 1 well in Adriatic offshore (—a—: mostly documented by geological and geophysical data; —b—: interpreted). Legend: pre-Permian basement (1), Permian-Triassic sequences (2), undifferentiated Mesozoic sequences (3), Jurassic-Cretaceous sequences (4), Paleocene-Miocene sequences (5), Pliocene-Quaternary sequences (6); deep-well (Se: Sedico 1 (projected), Vo: Volpago 1, As: Assunta 1); Neogene to Quaternary thrusts (VS: Valsugana, BL: Belluno, BV: Bassano-Valdobbiadene, AV: Aviano, S: Sacile). Trace in Fig. 1.

The age proposed here is based on the following evidence and considerations.

1. Post-second exhumation erosion, measured above the central belt, reaches a maximum of 150 m.

2. Within the very thick Messinian conglomerates (*Conglomerato del Montello*) outcropping in the eastern Veneto foothills, pebbles of Agordo metamorphic rocks have never been reported.

3. Pebbles of metamorphic rocks coming from an unidentified source have been found in the very rare intramontane fluvial conglomerates of Middle Pleistocene age, south of Agordo in the Piave basin.

4. In this drainage basin there are abundant Agordo lithotypes only in Late Würmian fluvioglacial and Holocene alluvial deposits.

5. Pre-Würmian glacial deposits are lacking in Agordo and the surrounding areas (B. Castiglioni, 1939; Fellerer, 1971).

6. Incoherent glacial drifts of generic "Würmian" age are widespread on the core and overlap locally outcropping fluvial (fluvioglacial ?) conglomerates. In the past, the latter were usually defined as "Interglacial alluvium" (=pre-Würm or even "pre-Glacial") or as "Riss-Würm Interglacial alluvial deposits" (B. Castiglioni, 1939). On the basis of the same arguments developed by G.B. Castiglioni & Trevisan (1973), we propose that these conglomerates may belong to the Würmian interstade preceding the last and major Würmian pleniglacial phase, i.e., Paudorf Interstade = Val Caltea-I of the Veneto and Friuli Alps.

Therefore, a Würmian age seems to be suitable for the second exhumation, if the high erodibility of most of the Agordo metamorphic rocks is also considered.

As regards the structures which developed during phase D_3, we must first state that the response of the basement was of brittle and semi-brittle type, mostly through reactivation or forming of the usual fault systems (e.g. Fellerer, 1971) related to a large thrust-fault zone.

First of all, various sets of subvertical faults occur, mostly NW- to NNE-striking. They show a more or less important lateral slip and are characterized mainly by metric to decametric displacement.

Less frequently, and always with minor displacements,

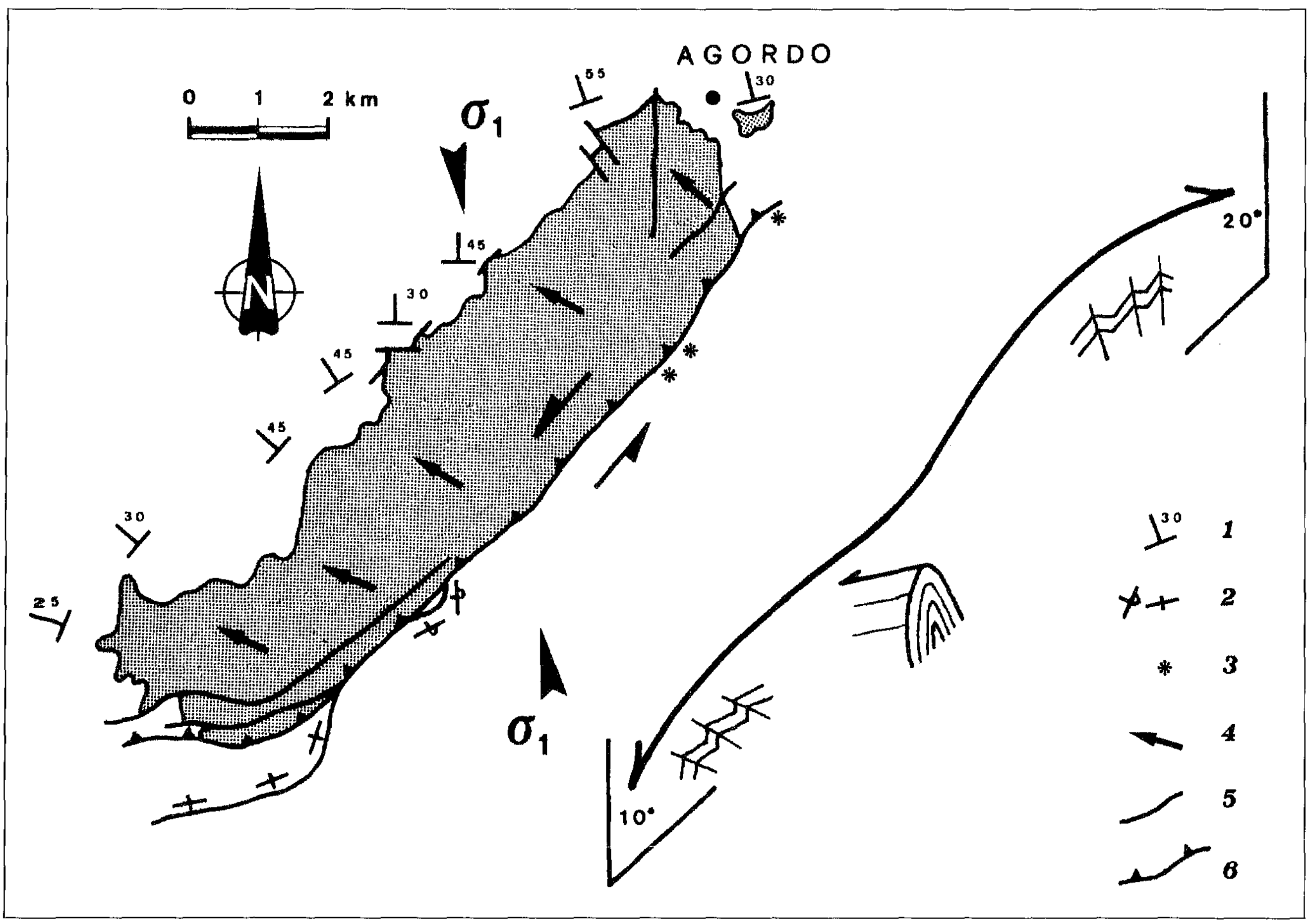

Fig. 10 - Sketch of Agordo basement showing main D_4 structures (from Poli & Zanferrari, 1991 b). Symbols: most frequent attitudes of layering of Lower Permian sequences (1); subvertical attitudes of layering in Permian-Triassic cover, which describe asymmetric, vertical drag folds (2); vertical or steeply plunging mesofolds in cover along subvertical ramp of Valsugana thrust (3); sets of en échelon F_4 mesofolds (4); main Alpine faults (5); Valsugana thrust (6). Patterns of F_4 kinks and en échelon F_4 mesofolds, and axis undulations of F_4 antiform are outlined in vertical section.

there are high-angle reverse faults, synthetic or antithetic with respect to the mainly N50°-striking Valsugana thrust.

The displacement of only a few low-angle faults (NW-dipping at 10°-25°), trending subparallel to S_2 and outcropping in the deepest part of the core (SW zone in Figs. 3a and 3b), is greater.

Metric-decametric F_3 kinks, mostly asymmetric and often conjugate, are common. They are often joined to some of the above-mentioned faults.

Steeply-inclined, angular F_3 mesofolds with mainly NE-striking axes are located along the front of the thrust.

The most important effect of D_3 is represented by a northwestward tilting of 30°-40°, as shown in Figs. 6 and 7. Because this inclination value decreases by about 10° towards the SE, a gentle macro-antiform formed, the axis being subparallel to the strike of the Valsugana thrust. Owing to the superposed D_4 deformation (section 11), the axis of the D_3 antiform, on average N50°-striking, ranges between N25° and N90°. Identical changes of axial direction can also be found in the F_3 kinks and mesofolds.

According to the most frequently accepted geodynamic model, we also set the SE-thrusting of the Agordo basement along the large Valsugana thrust in the framework of the Neogene-Quaternary evolution of the eastern Southern Alps (e.g., Massari et al., 1986; Castellarin & Vai, 1986; Ambrosetti et al., 1987; Boccaletti et al., 1990 a, 1990 b).

The formation and development of the "south-vergent Southalpine Chain" (SSC) of Ambrosetti et al. (Fig. 9) are linked to the crustal shortening which is developing in a compressive regime with σ_1 = NNW-SSE (Ahorner, 1975, 1978). In this kinematic setting, the Southalpine-Dinaric foreland (Fig. 11) acts as a rigid indenter, which underlies the Southalpine thrust-front.

The convergence between foreland and Southalpine thrust-front is progressively consuming the northern margin of the foreland itself, through tectonic accretion of further thrusts in front of the SSC (Fig. 9).

The evolution of the Serravallian to Holocene SSC and of its SSE-migrating foredeep-thrust belt system is well documented by stratigraphic and sedimentological data (Massari et al., 1986; Stefani, 1987) and by structural and geophysical evidence (Zanferrari et al., 1982; Ambrosetti et al., 1987; Slejko et al., 1989; and references therein).

11. Left-Lateral Crustal Escape

At present, the Valsugana thrust is characterized by a very large upper ramp, which is subvertical near Agordo (600 m a.s.l.), becomes NW-dipping at approx. 45°at an altitude of 1300-1400 m a.s.l., and then rapidly flattens to subhorizontal toward the SSE. On this frontal flat the Permian-Mesozoic sequences have been partly duplicated with a shortening at least of 11 km (Zoldo-Longarone area) (Antonelli et al., 1988).

Along the base of the ramp, the basement exhibits highly tectonized contacts with the Norian Dolomia Principale and with numerous thrust slices of Permian and Triassic formations, always with subvertical attitude. The geologic sections of the Vallalta mine (Baccos, 1968) and that of the Valle Imperina (Ogniben, 1968) accurately reflect the geometry of the thrust-fault zone.

Various kinematic indicators within the fault zone and subvertical drag folds along the upper ramp in the cover at the footwall, observed from meso- to macroscales (Fig. 10), display a left-lateral movement along the upper ramp of the Valsugana thrust (D_4 phase).

In the basement of the hanging wall, the following structures linked to sinistral wrenching were found (Poli & Zanferrari, 1991 b).

1. High-angle faults, striking subparallel to the Valsugana ramp and showing horizontal or oblique sinistral slip (Fellerer, 1971; Doglioni & Castellarin, 1985).

2. Concentric upright-plunging en échelon folds, observed from micro- to metric scales, and moderately asymmetric kinks with prevailing opposite vergence in the SW zone and in the NE area of the core (Fig. 10). Both have axes plunging 10°-25° toward N290° - N320°. The en échelon folds sometimes show a rough non-penetrative fracture cleavage in the metapelites.

The F_4 folds deform all the previous planar and linear structures (cf. section 10 and Figs. 6 and 7) and are thus the youngest. Their fabric and morphology are typical of very superficial conditions of deformation.

As sketched in Fig. 10, the F_4 mesofolds in the hanging wall participate in a larger structural pattern. It is a further gentle antiform of the whole basement, with axis striking approx. 310° and plunging more steeply at the eastern edge of the core (20° eastward). Because of this plunge the basement rapidly disappears a few km east of Agordo.

We observed a basin-and-dome interference pattern of regional extent along the whole Valsugana-Fella-Sava thrust (VFS). Therefore, we believe that the alternate appearance and disappearance of the metamorphic basement (or of the Carnic Paleozoic) in the hanging wall of VFS are also related to this pattern and not only to inherited structures, as asserted by Bosellini & Doglioni (1986).

To define the problems concerning the chronological relationships between D_3 and D_4 and the kinematic setting of the D_4 phase in the framework of the Quaternary geodynamic evolution of the eastern Southern Alps, we stress that:

1) left-lateral movements started and became predominant only when the large upper ramp of the Valsugana thrust had become subvertical and so stopped SE-thrusting, at least within the upper structural level;

2) sinistral wrenching along a N50°-striking subvertical fault is consistent with the present stress field with σ_1 = NNW-SSE (Ahorner, 1975, 1978);

3) the second exhumation of the basement probably took place during the Würm;

4) the features of the F_4 folds show very superficial

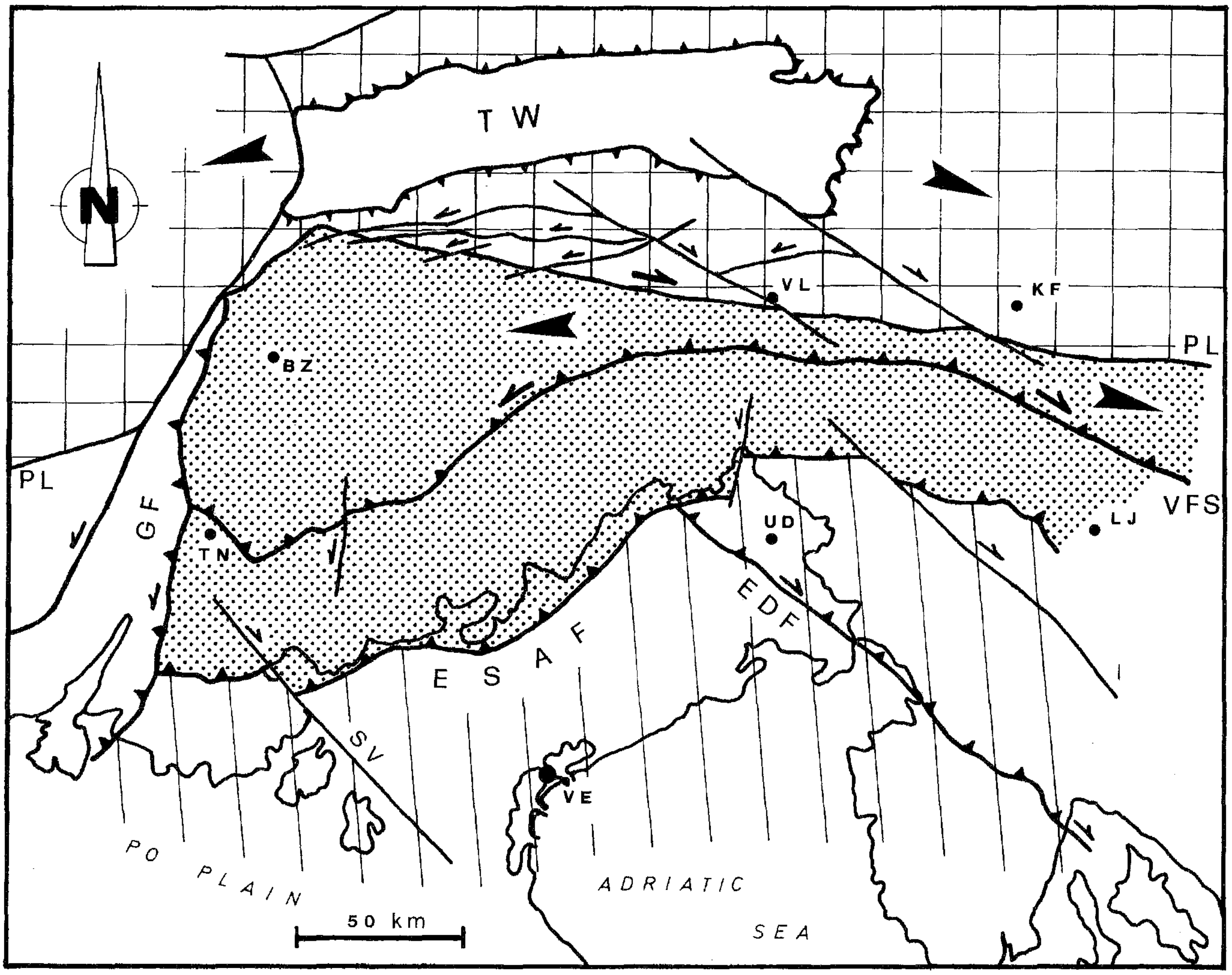

Fig. 11 - Crustal escape (arrows) in Austroalpine and Penninic units (from Ratschbacher et al., 1989) and northern part of eastern Southern Alps (this paper). Legend: squared area = eastern Austroalpine cover and basement nappes; TW = Penninic nappes of Tauern Window; dotted area = eastern Southern Alps; oblique lines = indenter of Southalpine-Dinaric foreland; PL = Periadriatic Lineament; VFS = Valsugana-Fella-Sava thrust; GF = front of Giudicarie thrust-belt; ESAF = front of eastern Southern Alps thrust-belt; EDF = front of External Dinarides thrust-belt; SV = Schio-Vicenza fault; I = Idrija fault; VL = Villach; KF = Klagenfurt; BZ = Bolzano/Bozen; TN = Trento; LJ = Ljubljana; UD = Udine; VE = Venice. VFS and ESAF define Neogene- Quaternary south-vergent Southalpine Chain. Main tectonic activity in Giudicarie thrust-belt is Messinian in age (Castellarin et al., 1987) and Priabonian (Cousin, 1981) in External Dinarides thrust-belt; both are mostly acting in a transpressive regime.

conditions of formation and they are the youngest structures.

From these points, we therefore deduce that D_4 started in relatively recent times (Middle ? - Late Pleistocene) and that it is still active.

Therefore, phases D_3 and D_4 represent different but congruent effects of the present stress field. Their partial chronological overlap is related to the more recent geometrical evolution of the Valsugana thrust which changed, at least in the upper structural level, from a low-angle, SE-vergent thrust into a vertical, left-lateral wrench fault.

As pointed out at the end of section 10, the present evolution of the eastern Southern Alps and the External Dinarides is ruled by the NNW drift of the indenter of the Southalpine-Dinaric foreland (Fig. 11). In this structural framework, the broad shear zone along the eastern part of the Periadriatic Lineament and the VFS fault act as a conjugate system with opposite strike-slip movements.

The observed movements are as follows: a) mostly right-lateral in the broad belt along the Periadriatic Lineament (mean direction of belt: N100°); b) right-lateral along the eastern part of the Fella-Sava fault, east of the meridian of Trieste (direction: N120°); c) left-lateral along the Valsugana fault s.s. (mean direction: N50°).

To sum up (Fig. 11), the large, curved VFS fault represents the main "escape way" of the northern part of the eastern Southern Alps towards both west and east, while the broad, mostly right-lateral shear zone which is developing along the Periadriatic Lineament act as a kinematic disengagement.

12. Final Remarks

The long and complex evolution of the Southalpine basement of Agordo is mostly a pre-Hercynian and Hercynian history. The best-defined stages are the deposition of the Cambrian-Silurian sequences and the second tectono-metamorphic phase.

Two main open questions concern the occurrence of a Sardic phase in the eastern Southern Alps and the geodynamic setting of the two Hercynian tectono-metamorphic phases. The solution of these problems needs further researches in the whole ESB and precise correlations with the sourrounding areas.

The late-Alpine history is also quite well-defined. Particularly, our model of crustal escape within the eastern Southern Alps fits well the seismotectonic frame of the eastern Alps, and the model of crustal escape in the whole eastern Alps suggested by Ratschbacher et al. (1989) on the basis of very different data.

One of the major problems concerning the late Hercynian and Alpine stages is to decipher the brittle and semi-brittle history revealed by the numerous faults of the basement. Nevertheless, although various gaps remain, the history of the Agordo basement is a tool for an understanding of the pre-Hercynian, Hercynian and late Alpine evolution of the whole eastern Southern Alps.

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FURTHER DATA ON THE PRE-ALPINE METAMORPHIC PRESSURE CONDITIONS OF THE AUSTRIDIC PHYLLITIC COMPLEXES IN THE EASTERN ALPS

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Abstract

The crystalline basement of the Austrides in the Eastern Alps mainly consists of pre-Alpine, amphibolite facies, polymetamorphic rocks partly affected by Alpine metamorphic overprints.

Phyllitic sequences lie over these amphibolite facies rocks in several areas, sometimes also affected by Alpine metamorphic overprints. However, their main metamorphic imprint is Variscan.

The sedimentation age of these sequences mainly is Upper Ordovician to Devonian. As regards the relationships with the underlying basement, these sequences are considered here to represent an original stratigraphic cover of the "Caledonian" (Panafrican?) basement, partly overthrust on their substratum due to Alpine tectogenesis.

Basing on data from the literature and new petrographic analyses of selected samples, all these phyllitic sequences are described, focusing the attention on the metamorphic features, and particularly on the baric character. The Variscan metamorphism recorded in these metapelites covers the whole temperature range of the greenschist facies, and display low pressure features. The muscovite composition (estimated through X-ray measurements of the *b* cell parameter) indicates a Variscan metamorphic thermal gradient of about 40°C/Km.

Key words: *Austrides, low-grade complexes, geothermobarometry, Variscan metamorphism.*

1. Introduction

The crystalline basement of the Austrides in the Eastern Alps mainly consists of pre-Alpine, amphibolite facies, polymetamorphic rocks partly affected by Alpine metamorphic overprints (Sassi et al., 1978; Becker et al., 1987; Sassi & Zirpoli, 1989; and references quoted therein).

Pre-Alpine greenschist facies complexes lie over this amphibolite facies substratum in several areas (Fig. 1). Mainly pelitic metasediments, acidic and basic metavolcanics and metavolcanoclastics, and subordinate carbonate metasediments make up these complexes, in which quartzphyllites however prevail.

The sedimentation ages recorded in these low-grade complexes mostly range from Upper Ordovician to Devonian (Sassi et al., 1978; Ebner, 1982; Ebner et al., 1989; and references quoted therein). The data in the literature indicate that the main metamorphic imprint in this sequence is Variscan; however, Alpine overprints also occur in some of them.

These phyllitic complexes are considered here to represent an original stratigraphic cover of the underlying "Caledonian" (Panafrican?) basement, partly overthrust on their substratum as an effect of the Alpine tectogenesis. However, the relationships between these complexes and their present substratum have been questioned. Some authors focused their attention on the tectonic surface occurring at the bottom of some of these greenschist facies sequences (Heinisch & Schmidt, 1976); others stressed the original stratigraphic relationships between these sequences and their present substratum (Sassi & Zanferrari, 1972).

A general remark on this problem is that, if the existence of a tectonic disconnection at the bottom of the greenschist-facies sequences cannot be questioned, data supporting the exotic nature of these sequences are completely lacking. Therefore, it can be reasonably assumed that the protoliths of these sequences were directly deposited on their present substratum, approximately in the same place where they presently are. The above mentioned tectonic, locally overthrusting, situations may be well explained as due to the differential deformation of the plastic phyllites and the more rigid, largely dehydrated basement during the Alpine compression, rather than to a large-scale nappe overthrusting. Such an interpretation is also supported by two facts:

(i) the original stratigraphic relationships between low grade complexes and their substratum are locally preserved (Neubauer, 1985);

(ii) metamorphic overprints occur in the substratum underneath the low grade complexes: these overprints are petrologically consistent with the Variscan metamorphic

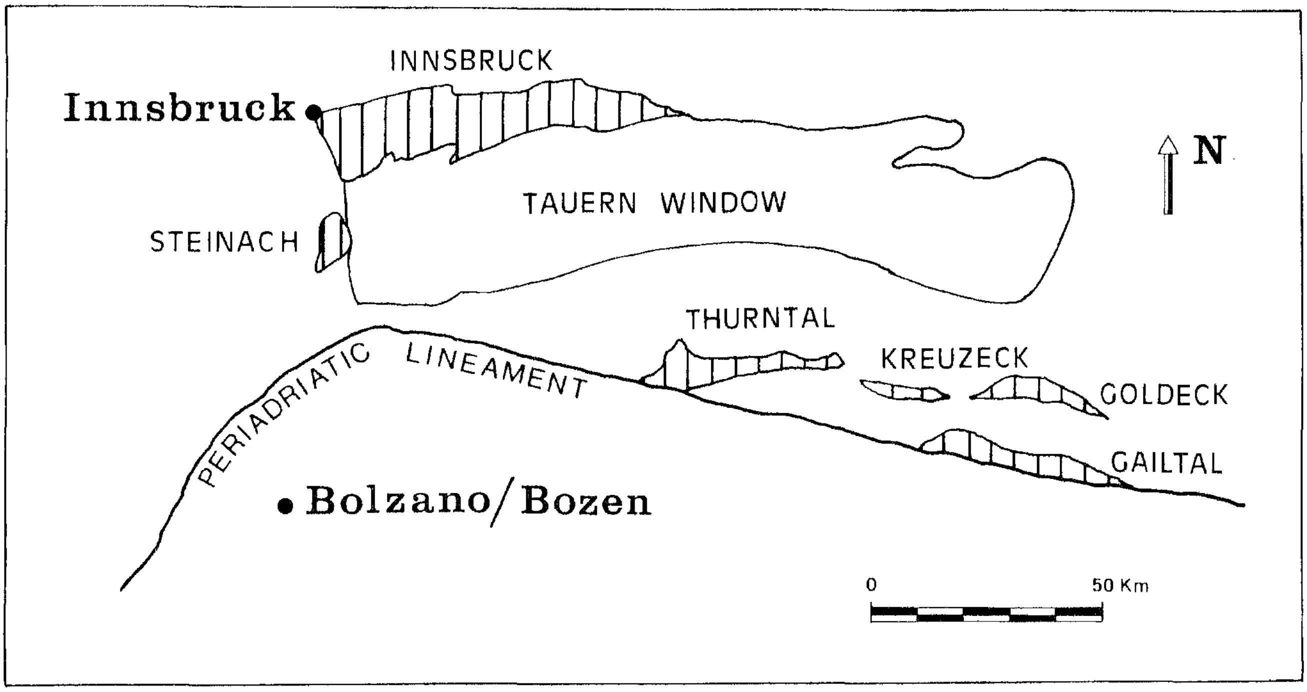

Fig.1 - Simplified geological sketch of the Eastern Alps showing the location of the 6 low grade phyllitic complexes taken into consideration.

imprint of the overlying greenschist facies complexes (e.g. Purtscheller & Sassi, 1975), displaying that basement and low-grade sequences jointly underwent the Variscan metamorphism, i.e. their contact is pre-Variscan, whatever its nature is.

Numerous papers exist in the literature dealing with lithology, lithostratigraphy, biochronology, etc., of these greenschist-facies sequences. However, systematic data concerning the physical conditions of their metamorphism are not yet available. Such type of data, and particularly the pressure character of metamorphism, is very useful for better understanding the nature and the history of these rock complexes, and for interregional correlations.

In order to contribute to bridge this gap, those phyllitic complexes in which Alpine overprints do not occur (as deduced from the literature) were taken into consideration for a petrographic study of selected samples, focusing the attention on the estimate of the pressure character of the Variscan metamorphism.

2. The Methodological Approach

370 samples of metapelites from 6 greenschist facies complexes were taken into consideration for geothermometric and geobarometric purposes. The temperature estimates were based on the petrogenetic grids, while the pressure estimates were based on the muscovite composition as deduced from the measure of the *b* cell parameter.

Petrographic analyses were only aimed at ascertaining, within the 370 selected rock samples: (i) the mineral compatibilities for geothermometric purposes, and (ii) the occurrence of compositional situations suitable for the geobarometric classification according to the specific method used.

The *b* cell dimension of potassic white micas from metapelites of appropriate composition and metamorphic grade has been widely used for geobarometric estimate in low-grade metamorphic terrains, which very often do not have minerals or mineral assemblages useful for monitoring pressure. The analytical procedure and barometric criteria were those proposed by Sassi (1972) and Sassi & Scolari (1974), and further developed by Guidotti & Sassi (1976, 1986). They are particularly suitable in the lower temperature part of the greenschist facies, and may also be applied successfully to very low-grade rocks (i.e. sub-greenschist facies: Padan et al., 1982).

For obtaining the *b* values, the muscovite (060) spacing was measured directly on rock slices cut perpendicularly to the rock foliation, following the specific procedure suggested by Sassi & Scolari (1974). Added metallic silicon was used as standard. The analytical error is $\pm$ 0.002 Å.

The rock samples were selected in order to fulfill, as far as possible, the compositional constraints suggested by the above-mentioned authors. In particular, graphite and/or ilmenite occurs systematically, assuring that f_{O_2} was buffered at low values.

Some caution has been suggested on the use of this method by Frey (1987) and Essene (1989); however, Guidotti & Sassi (1976, 1986) previously gave rigorous answers to these doubts, and Frey (1988) also recently used this method successfully.

3. The Results

The following greenschist facies complexes have been taken into consideration: 1) the Thurntal Phyllitic Complex; 2) the Kreuzeck Phyllitic Complex; 3) the Goldeck Phyllitic Complex; 4) the Gailtal Phyllitic Complex; 5) the Steinach Phyllitic Complex; 6) the Innsbruck Phyllitic Complex. Their respective location is shown in Fig. 1.

Each of these greenschist facies complexes is considered below from two points of view: i) a geological outline, including the main lithostratigraphic sequence, as deduced from the literature; ii) the analytical results obtained from selected rock samples, specifically: the main petrographic characters of the quartzphyllites, the values of the muscovite cell parameter, and relative estimates of the T and P values of the Variscan metamorphism.

3.1 *The Thurntaler Phyllitic Complex (ThC)*

Geological outlines. The Thurntaler Phyllitic Complex (ThC for short) mainly consists of quartzphyllites and fine-grained micaschists, in which minor metabasites ("prasinites"), acid metavolcanics and metavolcanoclastics, and rare marble and quartzite lenses are interlayered, mainly close to the southern and the northern boundary (Sassi & Zanferrari, 1972; Heinisch & Schmidt, 1976, 1984). This complex lies as a synform on amphibolite facies rocks certainly not affected by Alpine metamorphic overprints (Borsi et. al., 1973, 1978).

Analytical results. The microstructural evolution recorded in the ThC quartzphyllites is relatively simple. The axial plane foliation S_2 prevails over both the almost obliterated foliation S_1 and the local, discontinuous, postcrystalline S_3.

The common mineral assemblage in the quartzphyllites is Ab + Ms + Bt + Chl + Qtz ± Grt ± Ep, typically belonging to the medium to high temperature part of the greenschist facies.

Garnet composition indicates relatively high pyralspite content. The typical bell-shaped pattern of Mn zoning (Fig. 2a) suggests that these garnets crystallized during a single metamorphic event. This pattern significantly differs from that detected in the garnet crystals from the underlying kyanite-staurolite metapelites, in which a rim can be distinguished from a core (due to the opposite trend of the chemical contents: Fig. 2d, e, f), suggesting a two-stage development. Such a situation may indicate that ThC did not share the whole metamorphic history of the underlying amphibolite basement (Sassi & Zanferrari, 1972).

The mineral compatibilities suggest a temperature range of about 450-500°C (Figs. 3, 4).

As concerns pressure conditions, 27 new measurements of the muscovite cell parameter were carried out, obtaining an average value of 8.989 Å (s=0.004). Taking also into consideration the previously published 30 *b* data (Sassi, 1972), a data base of 57 *b* values is available for ThC. The related statistics are shown in Fig. 5a and Table 1. Following the barometric criteria proposed by Sassi (1972), Sassi & Scolari (1974) and Guidotti & Sassi (1986), the pressure values of the Variscan metamorphism recorded in ThC phyllites are relatively low.

3.2 *The Kreuzeck Phyllitic Complex (KrC)*

Geological outline. The Kreuzeck Phyllitic Complex (KrC for short) also consists of prevailing quartzphyllites (Lahusen, 1972; Reimann & Stumpfl, 1981), with minor intercalation of metabasites and rare marble lenses. This metasedimentary sequence overlies a higher-grade metamorphic basement, which mainly consists of paragneisses and garnet micaschists. According to Exner (1955, 1956), in some localities there is a gradual transition from the higher to the lower metamorphic sequence, but in other localities this boundary is tectonic. This situation and the strong similarities of the deformation style and petrographic features between the eastern part of KrC and the western part of GoC (see below) emphasize a straight correlation between the two complexes.

Analytical results. The metamorphic features and the evolution recorded in the KrC quartzphyllites are very similar to those described for the ThC metapelites. An axial plane foliation S_2 is the prevailing planar anisotropy, because S_1 is largely obliterated and S_3 only locally occurs.

The common mineral assemblage in the KrC metapelites is identical to that in the ThC (Fig. 3), suggesting T values in the range 450-500°C approx (Fig. 4). The garnet composition and zoning are also similar (Fig. 2b), suggesting that the metamorphic crystallization of KrC also developed during a single event.

The muscovite *b* cell parameter was measured in 60 samples of suitable composition. The average value is identical, within the limits of the analytical error, to that obtained from the ThC metapelites (Fig. 5b and Tab. 1). Therefore, the pressure values turn out to be low.

3.3 *The Goldeck Phyllitic Complex (GoC)*

Geological outlines. The main geological features of the Goldeck Phyllitic Complex (GoC) are so similar to those of the ThC and KrC, that in several synthetic maps of the Eastern Alps KrC and GoC are represented unitarily. It is a Paleozoic sequence (a poor conodont fauna of lower Devonian age has been found: Heinz, 1987) affected by a greenschist facies metamorphism of Variscan age (radiometric age data in Brewer, 1969).

The underlying, higher grade, metamorphic substratum consists of a garnet-bearing metapsammitic-pelitic sequence in which acidic orthogneisses, amphibolites and marbles are interlayered. Staurolite occurs in the central part of this basement (Deutsch, 1977). The lack of this index mineral in the eastern part of this basement, where garnet typically

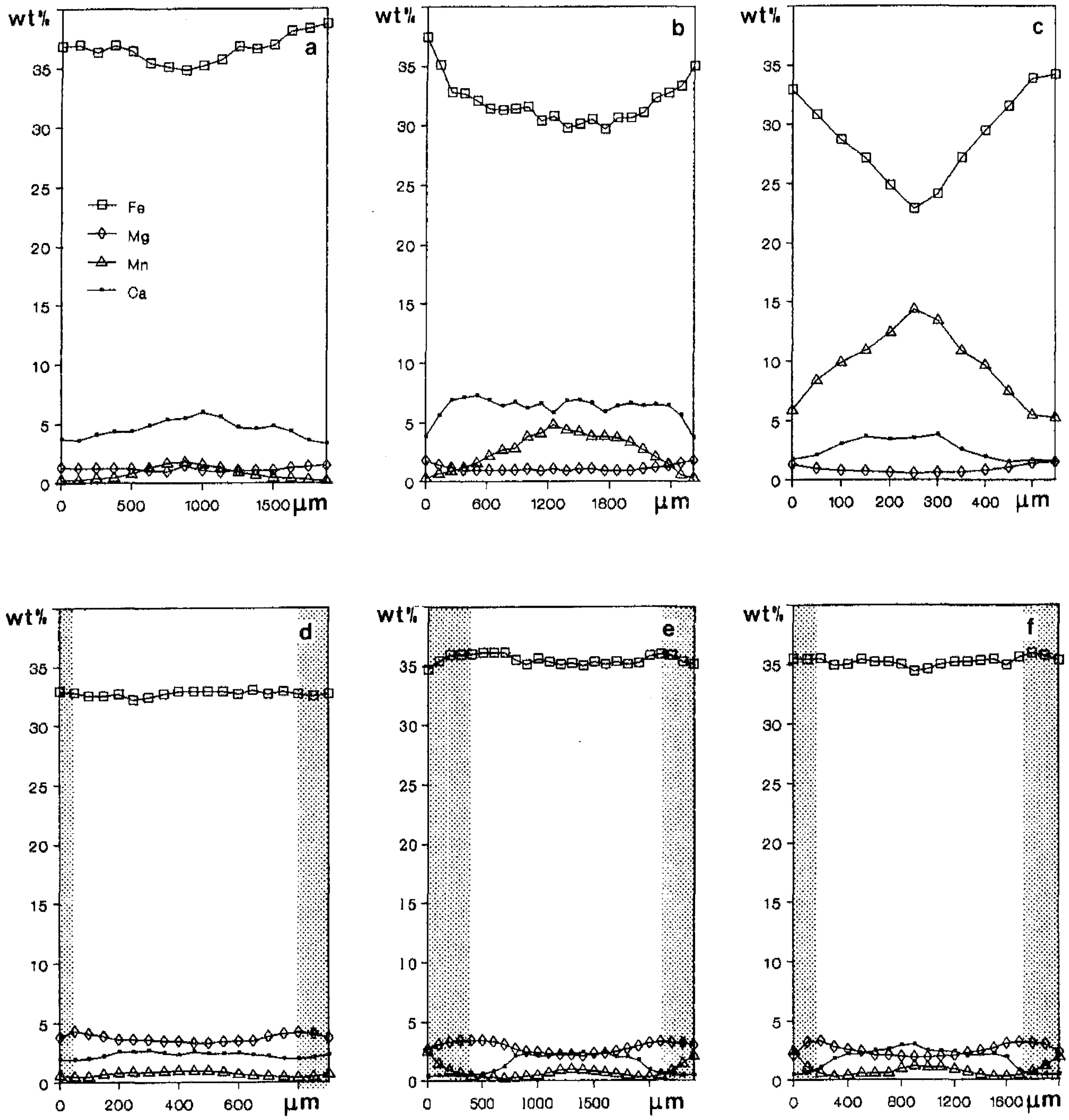

Fig.2 - Patterns of the chemical zoning of garnet crystals from three phyllitic complexes (a: ThC; b: KrC; c: StC) and the amphibolite-facies basement of ThC (d, e, f). A rim can be distinguished in the latter three crystals (d, e, f), due to the change of the chemical trends in them in comparison with the inner part of the crystals.

occurs (Heinz, 1987), may be due to bulk compositional constraints rather than to lower metamorphic grade.

A "transition zone" has been described within the lower part of GoC (Deutsch, 1977; Heinz, 1987). It has been interpreted as an intermediate metamorphic zone between the overlying, lower-grade part of GoC, and the underlying higher-grade substratum, within a unitary metamorphic zoneography which is assumed to link all rock sequences (Deutsch, 1977). However, according to Heinz (1987), the transition zone is characterized by intense folding and related crenulation and, although not representing a thrust plane, it separates two different metamorphic domains - a deep and a shallow level - both interpreted as parts of the same crustal unit juxtaposed each other due to post-metamorphic crustal shortening.

As concerns lithostratigraphy, a mafic metavolcanic unit separates an underlying sequence of quartzphyllites and metagreywackes from an overlying sequence of quartzphyllites, metacherts and calcite-dolomite marbles (the latter being of probable, late Devonian age).

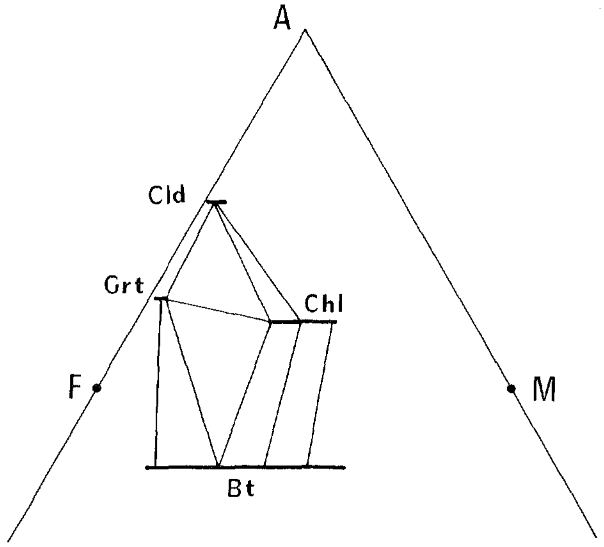

Fig.3 - AFM mineral compatibilities observed in the studied rock samples from all phyllitic complexes taken into consideration.

The whole GoC sequence is affected by tectonic complexities: a three-fold repetition of the lithostratigraphic sequence has been hypothetized by Deutsch (1977) in the central Goldeck region, and related to intrasequential thrusting.

Analytical results. Mineral compatibilities (Fig. 3) and microstructural features in the phyllites are identical to those previously described for ThC and KrC. The temperature conditions may be estimated in the range 450-500°C (Fig. 4).

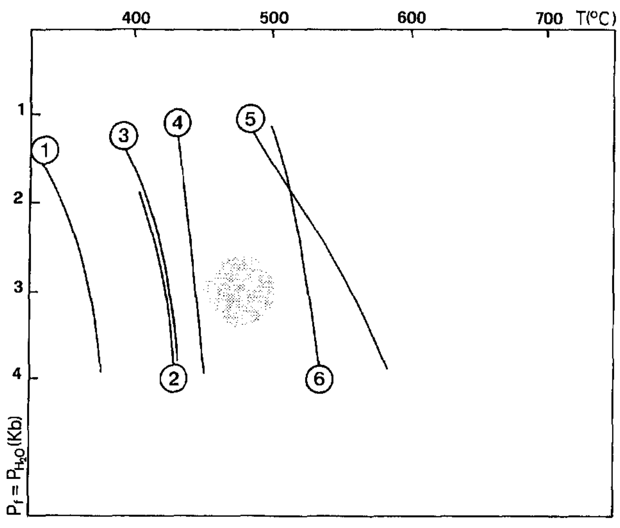

Fig.4 - Petrogenetic grid and temperature estimates at the assumed value P=3 Kb (*symbolic grey area*) concerning the studied rock samples from all phyllitic complexes taken into consideration. *Curve* 1: dehydration of kaolinite (Thompson, 1971); *curve* 2: Chl+Qtz →Sps (Hsu, 1968); *curve* 3: dehydration of pyrophyllite (Kerrick, 1968); *curve* 4: Stp+Ms → Bt+Ms (Winkler, 1976); *curve* 5: Pg+Qtz → Al_2SiO_5+Ab (Chatterjee, 1972); *curve* 6: Ctd-out (Spear & Cheney, 1989). Symbols as in Kretz (1983).

Table 1 and Fig. 5c shows the statistics of 58 measurements of *b* cell dimension of muscovite in Goldeck phyllites. The analytical results and the corresponding barometric indication are identical to those previously shown for ThC and KrC, within the range of the analytical error.

3.4 *The Gailtal Phyllitic Complex (GaC)*

Geological outlines. The Gailtal Phyllitic Complex (GaC) and the underlying amphibolite-facies substratum (called "lower complex" by some authors) make up the basement on which the non-metamorphic (or "anchimetamorphic": Niedermayr et. al., 1983) Permo-Mesozoic Drauzug stratigraphically lies.

Three main metamorphic complexes have been described in the Gailtal area (Becker et al., 1987, and references quoted therein):

1) a greenschist facies "upper complex", characterized by the "quartzphyllite-metabasite-layered marble" association;
2) an "intermediate complex" characterized by the "chloritoid-bearing micaschist-phyllonite-augengneiss" association;
3) an amphibolite facies "lower complex", characterized by the "staurolite-bearing micaschist-augengneiss" association, and the occurrence of kyanite (and rare sillimanite) in the Al-rich metapelites.

Several postcrystalline deformations are recorded in the Gailtal basement, mainly as mylonitic-diaphtoritic horizons within the intermediate complex. Heinisch (1987) interpreted these horizons as related to a Hercynian thrust plane, masked by a younger Variscan greenschist facies metamorphism. However, Sassi et al. (1974) and Becker et al. (1987) are inclined to consider the intermediate complex as a transitional metamorphic zone between the underlying kyanite-staurolite zone and the overlying lower-grade zones, notwithstanding the several mylonitic horizons occurring in it.

The "upper complex" includes metabasites, metacherts and banded marbles, besides the prevailing graphite-bearing quartzphyllites. The sedimentation age of this sequence is Paleozoic, as proven by conodont findings (Schönlaub, 1975).

Analytical results. Outside the mylonitic and diaphtoritic bands, the Gailtal quartzphyllites display microstructures substantially similar to the Thurntaler quartzphyllites.

The rock samples studied by Visonà (1974) were mainly referred by this author to the "Qtz + Ab + Ms + Chl subfacies". Biotite and sometimes garnet occur in the 21 rock samples we considered for the measurements of the muscovite *b* cell parameter. These assumedly higher grade samples were preferred because they could represent a

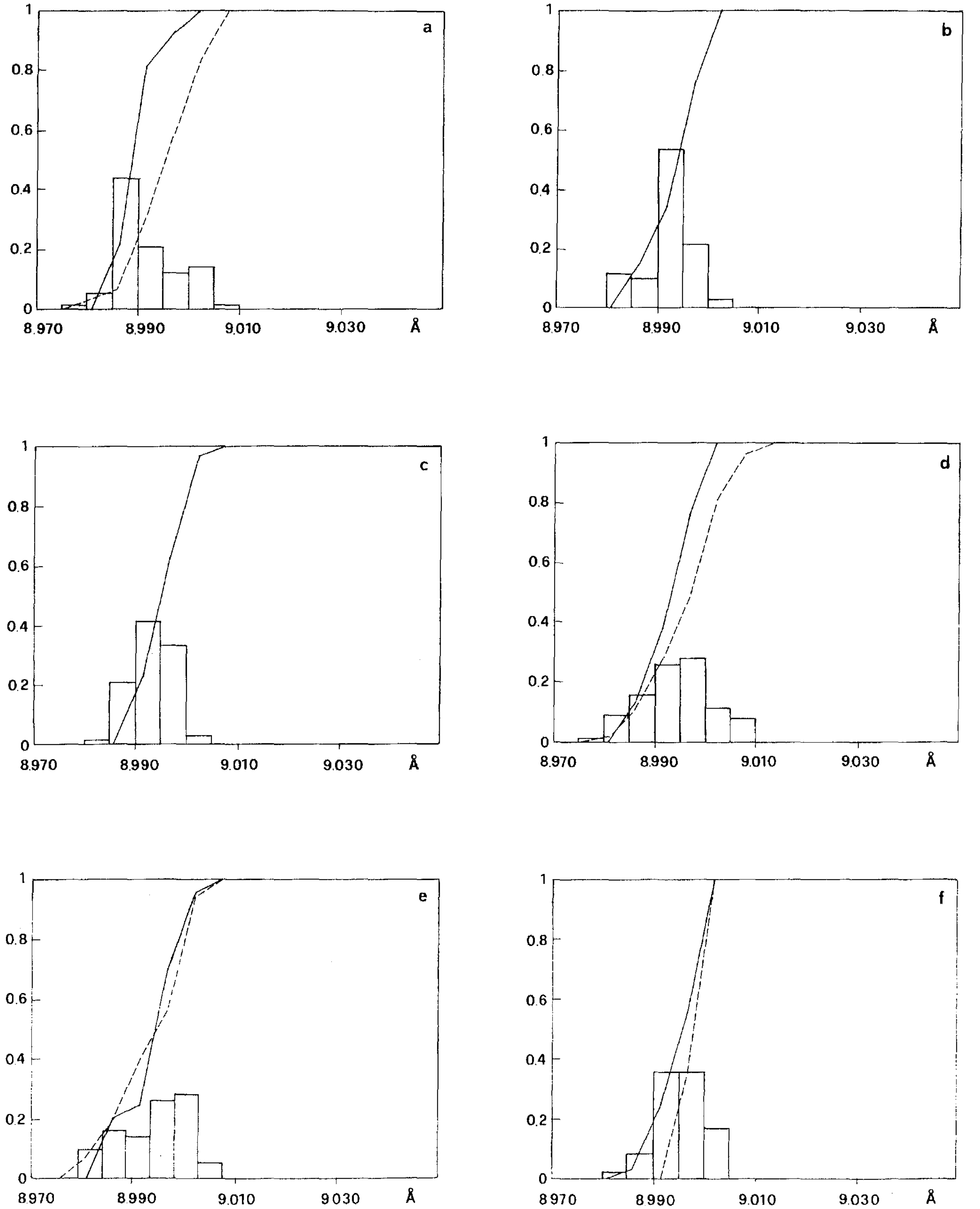

Fig.5 - Frequency distribution of the muscovite *b* cell data (in Å) concerning ThC (a), KrC (b), GoC (c), GaC (d), StC (e), InC (f). The cumulative frequency curves refer to new data (*solid line*) and to the data taken from the literature (*dashed line*). The *histograms* refer to all available *b* data from the given phyllitic complex.

Phyllitic Complex	confidence interval 95%	b_m	n	s	source of the data	Fig.
1 a Thurntaler		8,995	30	0,007	SASSI 1972	
1 b Thurntaler		8,989	27	0,004	new data	
Thurntaler (tot)	**0,003**	**8,992**	**57**	**0,006**	**all data**	5a
2 **Kreuzeck**	**0,002**	**8,993**	**60**	**0,004**	**new data**	5b
3 **Goldeck**	**0,002**	**8,995**	**58**	**0,004**	**new data**	5c
4a Gailtal		8,996	54	0,007	VISONA'1974	
4b Gailtal		8,993	21	0,005	new data	5d
Gailtal (tot)	**0,003**	**8,995**	**75**	**0,007**	**all data**	
5a Steinach		8,993	30	0,008	SASSI 1972	
5b Steinach		8,994	20	0,006	new data	5e
Steinach (tot)	**0,004**	**8,994**	**50**	**0,007**	**all data**	
6a Innsbruck		8,998	20	0,003	SASSI 1972	
6b Innsbruck		8,995	50	0,004	new data	5f
Innsbruck	**0,002**	**8,996**	**70**	**0,004**	**all data**	
TOTAL	**0,001**	**8,994**	**370**	**0,006**	ALL DATA	6

Tab. 1 - Statistics of the *b* cell parameter of muscovite from six Austridic complexes of the Eastern Alps (b_m = average value; n=number of samples; s=standard deviation).

metamorphic transition towards the amphibolite-facies basement: in such a hypothesis, their pressure conditions should be consistent with the reported occurrence of kyanite in this basement, and the muscovite *b* cell parameter in them are expected to be higher than those measured by Visonà (1974).

The mineral compatibilities (Fig. 3) indicate a temperature range of 450-500°C approx. (Fig. 4).

As concerns the estimate of the pressure condition, Fig.5d shows the frequency distribution of 75 values of the muscovite *b* cell dimension: 21 of them are new measurements while 54 are taken from the literature (Visonà, 1974). The statistics of the new and previous *b* data are shown in Table 1.

The new data do not differ significantly from those taken from the literature, within the limits of the analytical error. As a whole, they clearly indicate low pressure conditions, contrasting with the occurrence of kyanite in the amphibolite-facies basement.

3.5 *The Steinach Phyllitic Complex (StC)*

Geological outlines. This greenschist facies complex (labelled here StC for short) makes up an Alpine nappe which thrust over the so-called Brenner Mesozoikum when this latter sequence was already affected by the Cretaceous, greenschist-facies metamorphism. In fact, the Upper Carboniferous, Nösslach conglomerate is not metamorphic, and includes pebbles of the Steinach phyllites (Sassi & Menegazzo, 1971), a situation which also indicates that metamorphism in the StC is pre-Upper Carboniferous.

Quartzphyllites, black phyllites, chlorite-schists and mafic schists are the main rock components of StC (Sassi & Menegazzo, 1971). Thin lenses of marbles of various composition, including ferruginous dolomites and magnesites also occur (Frizzo & Visonà, 1981).

The postcrystalline effects related to the nappe emplacement are abundant at the bottom of the nappe, where mylonites and diaphtorites occur.

Analytical results. According to Sassi & Menegazzo (1971), the common mineral assemblage in the Steinach phyllites is: Qtz + Ms + Ab + Chl ± Bt ± Grt. Garnet has a spessartine-rich composition, and a bell-shaped, regular zoning (Fig. 2c): this zoning is significantly stronger than those shown in Figs. 2a, b probably because it was obtained from a garnet cut passing most nearly through the center of the grain. Zoned plagioclase porphyroblasts show a Ca increase in the outer rim, suggesting an increase in temperature during the latest stage of metamorphism.

The mineral compatibilities (Fig. 3) suggest a temperature range of 450-500°C approx. (Fig. 4).

Crystallization-deformation relationships indicate a sequence of stages related to prograde metamorphism which also occurs in the phyllitic pebbles within the Nösslach Conglomerate.

30 values of the muscovite *b* cell parameter were published by previous authors (Sassi, 1972; see also Sassi & Menegazzo, 1971). Their statistics are reported in Table 1, where 20 new measurements are also shown.

The frequency distribution of the whole set of 50 *b* data is shown in Fig. 5e. The metamorphism of the Steinach phyllites turns out to be of low pressure, as established by the above mentioned previuos authors.

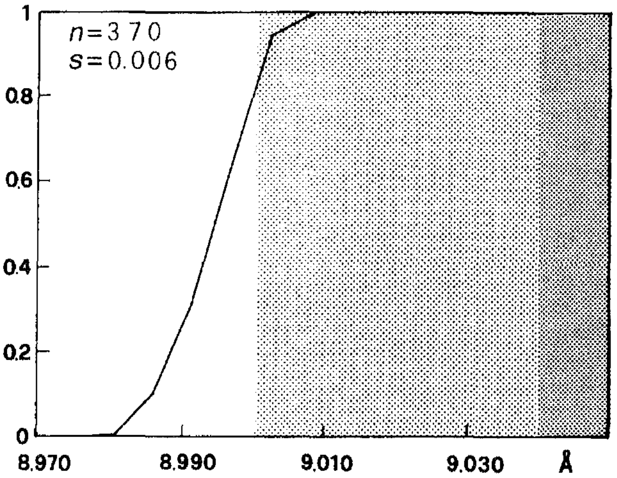

Fig.6 - Cumulative frequency curve of all 370 *b* data. The *b* boundaries between low (white), intermediate (grey) and high (dark grey) pressure fields are taken from Guidotti & Sassi (1986).

been done for the present research, obtaining results which are identical, within the range of the analytical error, to those existing in the above quoted literature.

The statistics of all data are shown in Table 1 and Fig. 5f. The low pressure character of the metamorphism recorded within the Innsbruck phyllites turns out very clearly.

3.6 *The Innsbruck Phyllitic Complex (InC)*

Geological outlines. The Innsbruck Phyllitic Complex (InC for short) largely outcrops to the north of the Tauern Window. Haditsch & Mostler (1983) subdivided it into tree sub-units.

The stratigraphically lower unit mainly consists of quartzphyllites and mafic schists, and includes acidic metavolcanics at an upper level. The intermediate unit is characterized by the occurrence of marbles and quartzites interlayered within quartzphyllites, and also includes mafic schists. The stratigraphically upper unit is characterized by black schists and dolomite/magnesite marbles: three parts have been distinguished in it, one of which mainly carbonatic, and the others pelitic and psammitic in composition, respectively.

Conodont have been found in the uppermost series (Mostler, 1973), indicating an uppermost Silurian to middle Devonian age.

Analytical results. Qtz + Ab + Chl + Ms ± Bt ± Ep is the most common mineral assemblage within the Innsbruck quartzphyllites, indicating temperatures corresponding to the medium greenschist facies (Hoschek et al., 1980), close to 450°C (Fig. 4). However, T may have been slightly higher if the lack of garnet in the rock samples we studied is due to bulk compositional rather than to thermal constrains.

Sassi (1972) published 20 values of the muscovite *b* cell parameter from biotite-free Innsbruck quartzphyllites. 50 new measurements on biotite-bearing phyllites have

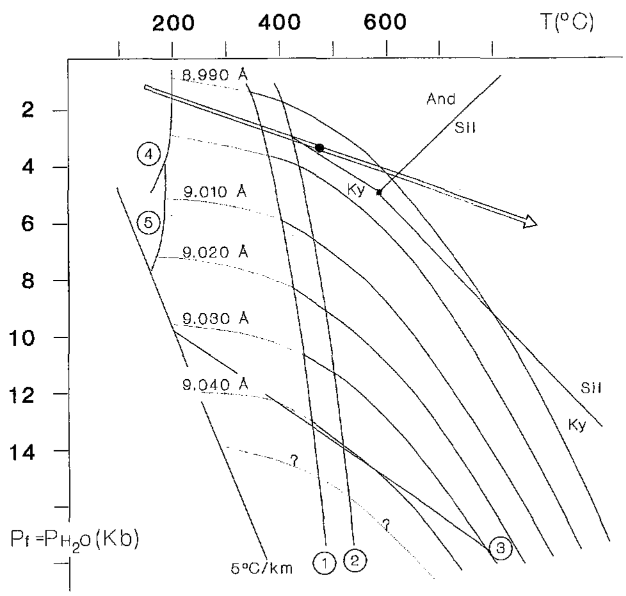

Fig.7 - Thermal gradient (arrow: about 40°C/Km) during Variscan metamorphism in the considered phyllitic complexes, as estimated on the basis of *b* and T values. The black dot, which graphically constrains the slope of the arrow, indicates the intersection between the *b* line 8.994 Å and the T range 450-500°C. Iso-*b* curves are taken from Sassi (1987), and are based on an extension to the sub-greenschist facies (dotted curves) of Guidotti & Sassi's (1986) diagram.
Curve 1: dehydration of kaolinite (Thompson, 1971); *curve* 2: dehydration of pyrophyllite (Kerrick, 1968); *curve* 3: stabilization of glaucophane (Carman & Gilbert, 1983); *curve* 4: Anl+Qtz → Ab (Thompson, 1971); *curve* 5: Hln → Lws+Qtz (Nitsch, 1968). *Al_2SiO_5 triple point*: Greenwood (1976). Symbols as in Kretz (1983).

4. Conclusions

The above presented data supply a general picture of the T, P conditions which are recorded in the considered phyllitic complexes, and are reasonably to be referred to the Variscan metamorphism. Specifically:

— the T range turns out to be 450-500°C approx. in the considered samples, but covers the whole range of the greenschist facies when the whole six complexes are considered;

— the 370 values of the muscovite *b* cell parameter fall in the narrow range 8.980-9.008 Å.

As regards the muscovite *b* cell data, it is worthy to point out from Table 1 and Fig. 5 that: (i) the standard deviation is reasonably low in every case; (ii) the 95% confidence interval is also low in each sample group; (iii) the histograms do not display a plurimodal distribution; (iv) all average *b* values obtained from each sample group are substantially identical, within the range of the analytical error. Therefore, the average *b* value calculated from all 370 *b* data significantly represents the studied parameter. This statement is supported by the relatively low standard deviation (0.006), the regular distribution of all data around the average value (Fig. 6), and the narrow confidence interval at 95% probality level ($2x1.96xs/\sqrt{n}=0.001$). Hence, the indication of low pressure conditions is unquestionable, and based on a firm statistical ground.

Therefore, the muscovite *b* cell parameter is a powerful key for correlating all these phyllitic complexes from the viewpoint of the metamorphic pressure conditions, and for considering all of them as an unique rock system which underwent the same P value during metamorphism.

Utilizing the average *b* value (8.994 Å) and the temperature range (450-500°C) estimated from the rock samples used for the measurements of the muscovite *b* cell parameter, it is possible to estimate the value of pressure and the metamorphic thermal gradient. The diagram *b* versus T proposed by Guidotti & Sassi (1986) and modified by Sassi (1987) has been used for this purpose.

As shown in Fig.7, pressure turns out to have been in the range 3-3.5 Kb, and a value of about 40°C/Km has been calculated, assuming a geobaric gradient of 270 bars/Km. We refer these values of pressure and metamorphic thermal gradient to the metamorphic peak of the Variscan event recorded in the considered phyllitic complexes.

As final consideration, we point out that this value of the thermal gradient is consistent with the Variscan andalusite ± cordierite metamorphism widespread in the amphibolite-facies Austridic basement of the Eastern Alps (e.g. Purtscheller & Sassi, 1975), and conflicts with the occurrence of kyanite in some parts of this basement. Either kyanite is pre-Variscan (as maintained by Sassi et al., 1987, and references quoted therein, consistently with radiometric age data: see review in Sassi et al., 1985), or the thermal gradient evolved from rather low (e.g. 25°C/Km) to higher values during the Variscan metamorphism, an hypothesis for which a chronological evidence is still lacking.

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Apennines and Calabrian Arc

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THE PRE-ALPINE CRYSTALLINE BASEMENT OF THE PELORITANI MOUNTAINS (SICILY): ACQUIRED KNOWLEDGES AND OPEN QUESTIONS

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Abstract

In the Peloritani Mountains (Northeastern Sicily) are represented segments of the Hercynian orogen recognized in the Calabrian-Peloritan arc, geotectonically defined by a pre-Hercynian crystalline basement (Aspromonte Unit plus Mandanici Unit) overthrusted and overturned on its Palaeozoic (Cambro-Ordovician up to Devonian-Carboniferous) volcano- sedimentary cover.

This Palaeozoic cover crops out in the form of several Alpine tectonicthrust sheets and is basically characterized by fine terrigenous meta-sedimentary rocks very rich in carbonaceous matter and detrital micas and conglomeratic channel-like layers are also present. Syn-sedimentary volcanism, represented by basic alkaline products (Ordovician,?) and late calc-alkaline basic to acidic rocks (Carboniferous,?), is suggestive of an evolution from extensional to compressional events preceding the very low monometamorphic Hercynian overprint.

Common lithologies in the Mandanici Unit are phyllites (mainly carbonaceous and/or quartzose; also iron carbonates- chlorite+albite - chloritoid- and garnet-bearing) with marble/calcschists intercalation, quartzites, metabasites and chlorite schists, grading upwards to micaschists and amphibolites.

The boundary between the Mandanici Unit and the overlying Aspromonte Unit is marked by a marble level, grading to a thick sequence of high-medium grade metamorphic graywackes (biotite paragneisses) with minor Ca-silicate felses, pelitic and mafic/acidic volcanic intercalations (kyanite-staurolite micaschists, amphibolites and augengneisses). Minor foliated and massive granitic, aplitic, and pegmatitic vein-lyke small bodies could be related to respectively ancient (eo-Caledonian/late Cadomian,?) and younger (Hercynian) late-orogenic magmatic activity.

Metamorphism in the upper units (Mandanici and Aspromonte) of the Peloritan basement developed under intermediate to low pressure conditions (early syn-tectonic crystallization of ilmenite in epimetamorphites and of kyanite in micaschists and gneisses, followed by a low-pressure prograde metamorphism and also followed by the static crystallization of andalusite, sillimanite and K-feldspar).

The peraluminous granitoids intruding the metamorphic basement are derived from crustal anatexis, according to some authors, but a I-type magma contribution is not excluded.

Key words: *Pre-Hercynian basement; paleozoic cover; stratigraphy; Peloritani Mts.; Calabrian Arc.*

1. Introduction

Reconstruction of the complex geological history of the Peloritani Mountains has been attempted several times. Ever since the end of the last century, several models and speculations about this problematic crustal sector (including Calabria), have been proposed.

The difficult reconstruction of pre-Alpine history is also mainly due to the scarcity of unaltered rocks which can be dated with accuracy, and also is due to tectonic Alpine overprint. This paper thus presents a brief update of our knowledge of Prealpine history which, from many viewpoints, is richer in events and certainly more detailed than the later Alpine history. The Alpine effects have been considered by various authors to be not always prominent, although often particularly important in determing the present structural arrangement of these ancient terrains.

The Peloritani Mountains (NE Sicily) are the southermost segment of the arc belt linking the NW-SE-trending Apennines s.s. with the E-W-trending Sicilian Maghrebide chain. According to some authors (Amodio Morelli et al. 1976; Bonardi & Giunta, 1982; Scandone, 1979, 1982), the Peloritani Mountains represent a composite body of Alpine (l.s.) age with obvious signs of a pre-existing Hercynian structure (Ferla et al.,1983).

In any case, complex structures with Lower Miocene south-verging folds (Ogniben, 1960; Haccard et al. 1972; Scandone et al. 1974) completely or partially obliterated the older structural setting. Numerous tectonic slices of Alpine age cut crosswise the original ancient basement and its presumed cover (Ferla,1974b).

The tectonic overlap of crystalline units of different metamorphic grade and the interposed Meso-Cenozoic sedimentary covers (Ogniben, 1960; Truillet, 1968; Atzori & Vezzani, 1974) are the most evident features of the Peloritani structure with Alpine nappes. The tectonic overlap happened before the Aquitanian-Langhian turbiditic formation of Stilo- Capo d'Orlando (Bonardi et al.,1980)

The Peloritani basement essentially includes:

a) A Palaeozoic volcano-sedimentary sequence, weakly metamorphosed, i.e. the South-Peloritani Complex of Ferla (1974b; 1978), or the Crystalline Basement of the lower Peloritani Units of Atzori & Ferla(1979);

b) A pre-Hercynian basement, i.e. the North-Peloritani Complex of Ferla (1974b; 1978), dismembered into the Alpine units of Mandanici and Aspromonte; this basement mainly consists of meta-igneous and metasedimentary rocks of low to high metamorphic grade, migmatites, and some late- to post-orogenic magmatic rocks (295-270 Ma).

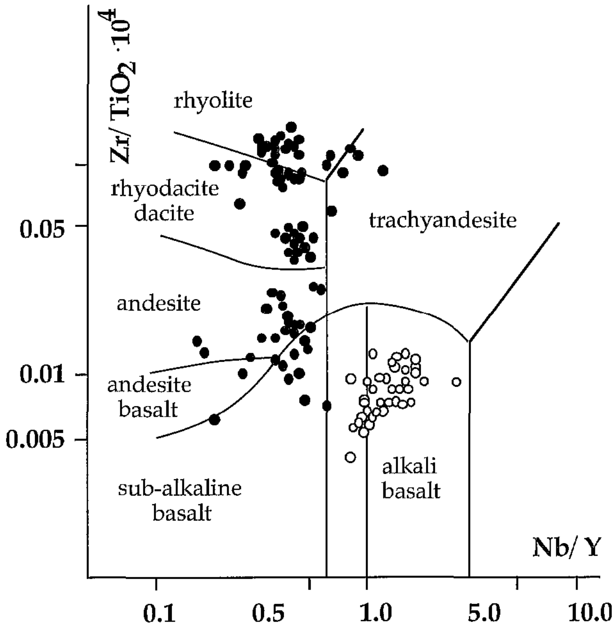

Fig. 1 - Zr/(TiO_2 x 10^4) vs. Nb/Y plot (Winchester & Floyd 1976) relative to Hercynian meta-volcanites from Peloritani Mountains. Meta-alkalibasalts (n = 35; Ferla, 1978; Ferla, & Azzaro, 1978b) are present but also orogenic meta-volcanics wich range from basalts to rhyolites (n = 81; Atzori, Ferla and Lo Giudice, 1982).

2. Crystalline Basement of the Lower Peloritani Units

This basement includes a meta-sedimentary and meta-volcanic sequence dated from Cambro-Ordovician to Carboniferous (Majesté-Menjoulas et al., 1986). Sedimentation has been essentially terrigenous, very rich in detrital micas and sometimes carbonatic material (up to 10%). Conglomeratic layers with clasts, coming from the erosion of the adjacent old crystalline basement (Ferla, 1974a), are particularly frequent in the Cambro-Ordovician, Siluro-Devonian and Carboniferous sequences.

Synsedimentary volcanic episodes (Fig. 1) include both alkali-basalts and basaltic- andesitic- dacitic calc-alkaline sequences, with associated rhyolitic-rhyodacitic products due to local melting and crustal contamination during the ascent of the calc-alkaline magmas (Ferla, 1978; Atzori et al, 1978; 1979; Ferla & Azzaro, 1978c).

These described magmatic events are older than the Hercynian metamorphic cycle. Their igneous association has, thus, been interpreted as representative of a geotectonic cycle changing from extensional conditions at least of Middle Devonian age, as suggested by the basic alkaline volcanism, to compressive conditions as suggested by the orogenic magmatism of early-Hercynian.

The metamorphic grade shown by the Palaeozoic volcano- sedimentary sequence of the Peloritani Mts. ranges from anchimetamorphism to the low-grade greenschist facies (chlorite zone)($T \leq 350°C$; $P \geq 2Kb$). Geothermometric estimates based on the $\Delta^{13}C_{(sid-gr)}$ in presumably Silurian graphitic layers, point to $T \approx 300°C$ (Censi & Ferla, 1989). Some data on the minerals present in these metamorphic conditions are shown in Table 1.

The structural features, according to Pezzino (1982), of these rocks are dominated by three folding phases: 1) an early open isoclinal fold phase with transposition of the sedimentary S_0 and formation of a pervasive axial-plane schistosity (S_1), connected with the only evident episode of metamorphic recrystallization. This event is marked by the growth of lens like chorite aggregate and of fibrous quartz in the pressure shadows around detrital quartz and pyrite crystals (Ferla,1978). 2) a local deformation, with microfolds, and with the appearance of a crenulation cleavage (S_2); 3) a late deformation, characterized by non-cylindrical, centimetric to metric folds, of similar and chevron types.

A gently westward dipping approximately E-W interpherence lineation is also linked to the early fold phase in the Peloritani Mts. (Ferla & Lucido 1973).

2.1. *Stratigraphic Sequence*

Reconstruction of the stratigraphic sequence in the southern Peloritani metamorphites is particularly difficult, due to both the presence of metamorphic folds and the effects of the Alpine tectonic dismembering: these effects cannot always be definitely distinguished from older structural characters, whenever the rare interposition of Meso-Cenozoic sedimentary cover is missing.

As regards to the age of these rocks, for more than a century various authors have recognized analogies between the Peloritani phyllites and similar Palaeozoic rocks from other Italian, i.e. Sardinia and eastern Alps, and European

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Fig.1 has been erroneously reproduced several times in this paper, instead of Figs. 2 to 7. The latter figures are correctly shown in this page.

Muscovite	$d_{002} \geq 9.985$ Å $b_0 = 9.000 \div 9.007$ Å 2M/(1Md+2M) ≤ 75% I. K. 0.18°÷0.40° Δ(2θ°)
Chlorite	Fe/(Fe+Mg) =0.6÷0.8
Graphite	d1A(+d2) (Landis) $\delta^{13}C$ =-26 ÷ -29 ‰ (PDB-1)
Siderite	$\delta^{13}C$ ≈- 9.36 ‰ $\delta^{18}O$ ≈-11.05 ‰ (PDB-1)
Pyrophyllite	rare
Paragonite	+ quartz
Calcite	$\delta^{13}C$ ≈+ 0.20 ‰ $\delta^{18}O$ ≈- 8.41 ‰
Actinolite Albite Titanite Epidote	

Tab. 1 - Crystalchemical and mineralogical features in the epimetamorphites of the lower units of the Peloritani Mountains (after Atzori, 1970; Ferla, 1974,1978, 1983; Ferla, & Lucido, 1971, 1972, 1973; Censi & Ferla, 1989; Pezzino, 1982).

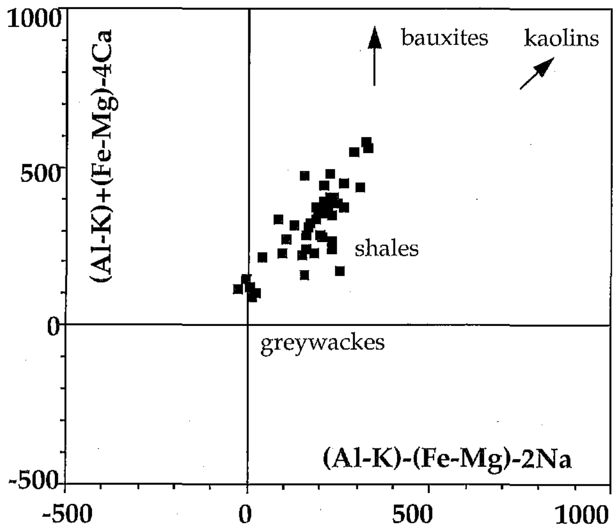


Fig.2 - (Al-K)+(Fe-Mg)-4(Ca) vs. (Al-K)-(Fe-Mg)-2Na after De La Roche (1978) for the phyllites of the Mandanici Unit from the Peloritani Mountains. Al-minerals (paragonite, chloritoid, garnet, staurolite) are common in these rocks which are derived from metamorphism of clay-like sediments.

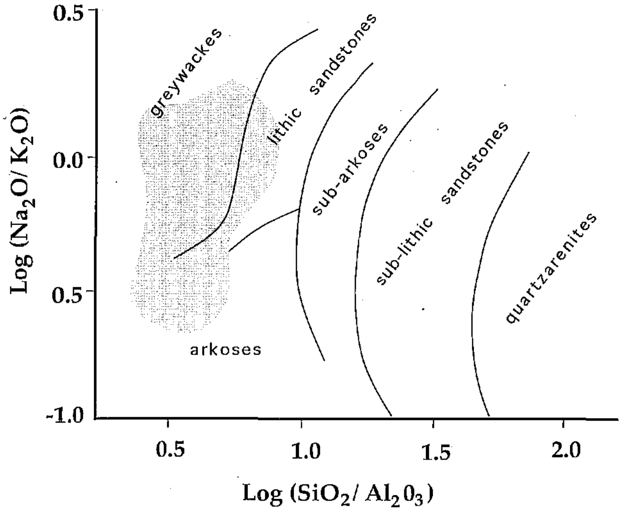


Fig.3 - Log (Na_2O/K_2O) vs. Log(SiO_2/Al_2O_3) after Pettijohn et al. (1972) for the biotite paragneisses from the Peloritani Mountains. These rocks are derived from metamorphism of flysch-like sediments (graywackes having a limited pelitic tendency).

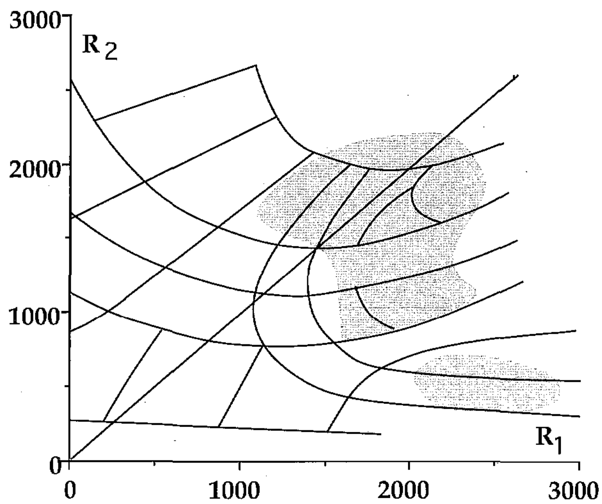


Fig.4. - R1 vs. R2 plot after De La Roche et al. (1980) for the amphibolites and augen gneisses of the upper units from the Peloritani Mountains. These rocks have been interpreted as old volcaniclastic products chemically near to active continental margin magmatism. R1=4Si-11(Na+K)-2(Fe+Ti); R2=6Ca+2Mg+Al.

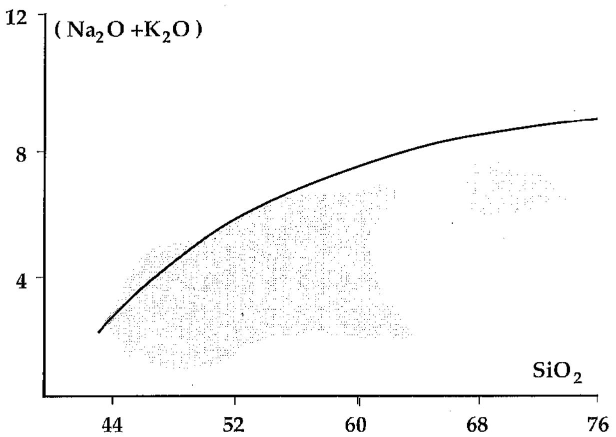


Fig.5 - (Na_2O+K_2O) vs. SiO_2 for the essentially sub-alkaline amphibolites and augen gneisses from the Peloritani Mountains.

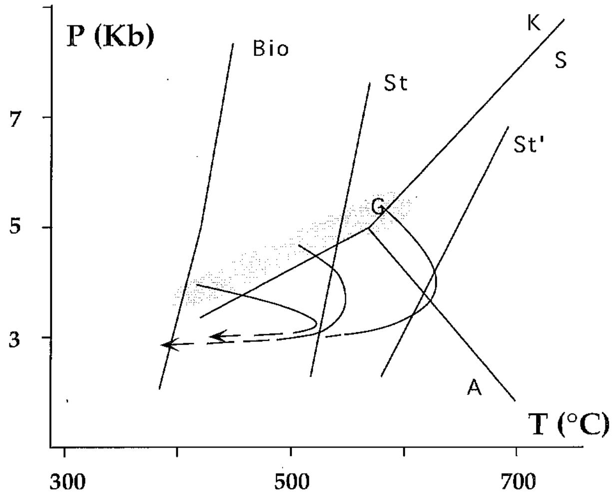
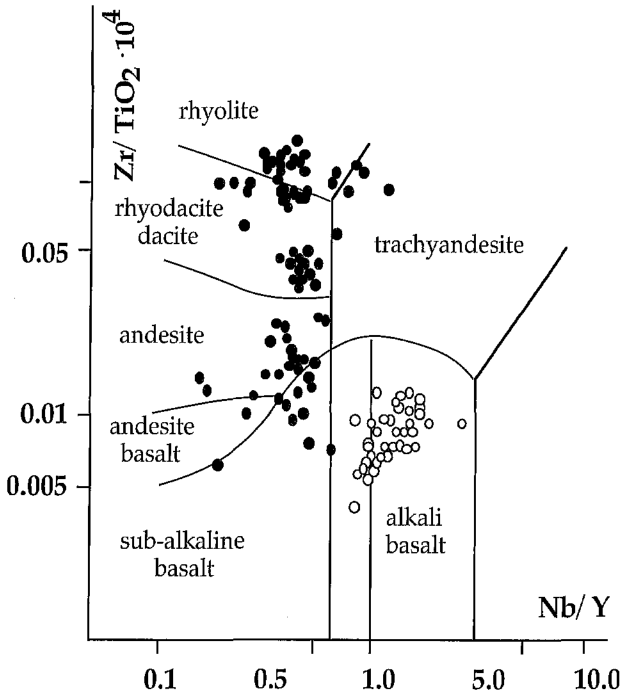


Fig.6 - P-T evolution of upper metamorphic units from the Peloritani Mountains. The spotted area represents the P-T range of the prograde eo-Hercynian or pre-Hercynian metamorphism deduced from mineral assemblages (Atzori & D'Amico; Ferla, 1974, 1978). The arrowed curves indicate the sin-S_2 and post-S_2 P-T paths with final retrograde episodes deduced also applying different geothermometers and geobarometers on various rocks (Ioppolo and Puglisi, 1988). G= Al_2SiO_5 triple point after Greenwood (1976); bio=biotite-in (Winkler, 1967); staurolite-in (Hoschek 1969); st'=staurolite+quartz out (Triboulet & Audren, 1985); K=kianite; S=sillimanite; A=andalusite.



Tab. 1 - Crystalchemical and mineralogical features in the epimetamorphites of the lower units of the Peloritani Mountains (after Atzori, 1970; Ferla, 1974,1978, 1983; Ferla, & Lucido, 1971, 1972, 1973; Censi & Ferla, 1989; Pezzino, 1982).

regions. Seguenza (1871) believed that the graphite-phyllites were Carboniferous in age; Cortese et al. (1886) believe that these rocks were Archaean or Silurian in age. Truillet (1968) identified a marker horizon composed of "andesites", polychrome schists, rhyolites, and rhyolitic tuffs, limestone with remnants of *Tentaculites*; Truillet also recognized the very low metamorphic grade - at the boundary with diagenesis - in the overlying layers. Lardeaux & Truillet (1971) discovered traces of Devonian *Dacriconarides* in the very low grade metamorphites with a non-metamorphosed Mesozoic cover.

On the basis of the composition and the cristallinity index of the white mica in the graphitic phyllites, Ferla & Lucido (1971) identified a relatively low-grade metamorphism which does not exceed the biotite isograd. The metamorphic grade decreases upwards from phyllites and graphitic slates to quartzites (with preserved rounded clasts) and up to slates and metagraywackes. Original kaolinite- and illite-rich layers, now pyrophyllite- and sericite-phyllites, may also locally be found (Ferla & Lucido, 1972).

In the sequences overlying the *Tentaculites* layers of Floresta, Ferla (1974a) recognized the occurrence of metaconglomerates and metasiltites with traces of chert clasts and acidic volcanites, as well as old phyllites and micaschists, marbles, gneisses and pegmatites. These old clast originate from a pre-Hercynian basement. According to the above author, this basement is today represented by the north Peloritani portions dismembered in the highest Alpine units of the chain (Mandanici and Aspromonte Units).

According to Atzori & Vezzani (1974), the pre-Alpine sequence was an old succession from pelitic to psammitic sediments (bottom) to arkoses and acidic volcanoclastites (top).

Bonardi et al. (1976) recognized in one of the lower Peloritani units dark-grey, rarely greenish, pelitic schists with intercalations of graphitic schists and marbles and then meta-arenites (often quartzites) alternating with the phyllites in the lower part and prevailing upwards.

Ferla (1978) e Ferla & Azzaro, (1978b,c), proposed a stratigraphic reconstruction based on estimates of metamorphic grade, on the assumption that metamorphic grade increases with the original thickness of the rock sequences. The oldest magmatic rocks, associated with higher-grade metamorphites as far as the Devonian *Tentaculites* limestones are alkaline basalts typical of rifting environments. The apparently higher layers, with porphyroids passing laterally to a calc-alkaline succession (basalts, andesites and dacites) seem to be younger, probably Carboniferous. According to Ferla (1978), some metabasites associated with the graphitic phyllites of S. Marco (Novara di Sicilia) were transitional basalts.

On the basis of K/Ar determination, Zuppetta et al. (1984) attributed the calc-alkaline metabasites to Triassic magmatism; nevertheless these data were considered unreliables by Ferla (1984).

Considering the pebbles of rhyolites in the metaconglomerate associated with the Tentaculites limestones, Ferla et al. (1983) attributed the acidic intrusions in the pre-Hercynian metamorphic basement, accompanied by ignimbritic effusions in the cover, to a pre-Devonian (Ordovician?) magmatic event. In their opinion, these igneous rocks were converted into augen-gneisses during the Hercynian orogeny.

Majesté-Menjoulas et al. (1986), Bouillin et al. (1987) found traces of Cambro- Ordovician fossils at Melia, near Taormina and tentatively attributed the overlying meta alkali-basalts to the Ordovician. In the Fiumara d'Agrò, near the village of Mitta, the same authors, on the basis of black colour, correlated the graphitic phyllites to the Silurian and the marbles to the *Tentaculites* marbles. However, according to Pezzino (1982) in this area some of these rocks belong to the Mandanici Unit. These rocks constituted the less metamorphosed part of succession and the marbles should correspond to the "Marmi II" (Censi & Ferla, 1983), of the pre-Hercynian metamorphic basement.

The upper layers of the Palaeozoic sequence, overlying the *Tentaculites* marbles, include metasandstones and metaconglomerates attributed, by lithologie, to the Carboniferous "Culm" facies of the European Hercynian chain (Majesté-Menjoulas et al., 1986).

The authors also provide a geological sketch of the Peloritani Mts. in which "phyllites and the Peloritani Palaeozoic" are unified. However, Majesté-Menjoulas et al. (1986) recognize that the "Peloritani phyllites" show features different from the "Palaeozoic" and that, by analogy with the Algerian Kabylia, they should be pre Upper Cambrian, "socle cadomien ou éo-calédonien". These phyllites belong to the Mandanici Unit auct., and are attributed by Ferla (1974, 1983) to the Caledonian.

In the light of new stratigraphic and paleontologic data, Acquafredda et al. (1988;1991) attributed to the Ordovician and to Carboniferous the andesitic magmatism often accompanied by persilicic rhyolites. In the authors' opinion, the thermal effects, often overprinting the Calabro-Peloritani metamorphics, are related to crustal melting caused by the ascent of subcrustal magmas. However, these authors do not explain why this static metamorphic recrystallization of Ordovician age was not affected by the penetrative Hercynian deformation which must have involved both meta-andesites and meta-rhyolites.

3. Crystalline Basement of the Upper Peloritani Units

The Upper Peloritani Units the Mandanici Unit and the Aspromonte Unit (auctorum). These Alpine unit still preserve large portion of a pre-Alpine crystalline basement, perhaps already overturned in the Hercynian (Ferla et al., 1983) . The underlying Mandanici Unit is mainly formed by epizonal to mesozonal metamorphics and relics of a non-metamorphosed Mesozoic sedimentary cover (Ferla, 1983, Ferla et al. 1991); the Aspromonte Unit consists of mesozonal to catazonal metamorphics and granitic rocks. There are minor tectonic slices inside these two major subdivisions.

According to Ferla (1974b, 1978, 1983), the basements of the last two Alpine unit constituted different parts of a single old basement with polymetamorphic rocks. The oldest metamorphic cristallization is pre-Hercynian, probably Caledonian, and displays synkinematic features (among other things the blastesis of ilmenite in a planar S_1 foliation in the phyllites, and of kyanite and/or staurolite in the micaschists and gneisses). During the Hercynian event, an evident S_2 foliation developed in the north Peloritani Complex roughly parallel to the original bedding S_0. A late Hercynian thermal event produced a regional static metamorphic recrystallization and was accompained by the intrusion of large granitoid stocks of the Calabrian arc, such as that preserved at Capo Rasocolmo and Capo d'Orlando.

Instead, Atzori et al., (1984a) postulated an entirely Hercynian polyphase metamorphic history. This interpretation is in agreement with some structural observations, wich indicate the presence of Hercynian deformational phases similar to those described by Pezzino (1982) in the monometamorphic rocks of the lower Peloritani units.

Metamorphic sequences, apparently similar to that observed in the Peloritani ones, are well-known in Calabria and the Algerian Kabylia. (Bouillin et al. 1984,1987; Majesté-Menjoulas et al. 1986; Vai & Cocozza, 1986). Cambro-Ordovician *Acritarchs* commonly occur in very low-grade metamorphic rocks (= Peloritani Palaeozoic) affected by Hercynian metamorphism only: therefore, the formation of the "Peloritani phyllites", composed of more or less graphite-rich schists, should be pre Upper-Cambrian in age and belong to "the upper part of a Cadomian or early Caledonian basement which are remobilized during the Hercynian orogenesis" (Majesté-Menjoulas et al. 1986).

Dueé' (1969), Bouillin et al. (1984;1987) believe that a still older gneissic basement of Precambrian age may be identified. Schenk (1980,1984,1988) traced a complex tectono- metamorphic history of the granulitc Calabrian crust, which ends at about 295 Ma ago.

Lastly, Omenetto et al. (1987; 1988) point out interesting similarities with other sectors of the European Hercynides and in particular with the Cambrian of the Montagne Noire in the area of Montedron, of Minervois etc., i.e. "similar Cambrian sequences and global metal spectrum (W,F,B,Pb,Zn), presence of tourmaline-scheelite and scheelite-black schist ores, common F-bearing scheelite skarns linked to Hercynian granites".

3.1. *Mandanici Unit*

The Mandanici Unit rock types are mainly of low and medium grade metamorphism: often graphitic phyllites (Fig.2) with intercalations of marbles and calcschists, quartzites, chloritic and amphibolic schists. These rock types are characterized by an evolution, including both synkinematic and postkinematic recrystallization: the latter was static, of varying intensity, and ranges from low to high grade: andalusite, sillimanite and cordierite in pelitic-schists, or epidote, wollastonite, pyroxene, garnet, in calcschists. Metamorphic retrograde effects also occur locally (i.e. chlorite, white mica, albite, epidote etc.,) the youngest of which are not always of clearcut Alpine age.

A tentative stratigraphic reconstruction (Ferla, 1983; Censi & Ferla, 1983; 1989; 1991; Omenetto et al., 1988) reveals at least three major horizons of carbonatic rocks intercalated in a mainly pelitic sequence rich in organic matter. The carbonatic rocks consist of limestones layers with dolomitic intercalations, locally with a silicoclastic or volcanoclastic fraction. The carbonatic rocks have $\delta^{13}C$ values compatible with a carbonate platforms environment (Censi & Ferla, 1983).

An overturned metamorphic sequence is identified by Ferla (1983): an early marble horizon (i.e. "Marmi I", or upper marble, or Tyndari marble) is associated to kyanite-staurolite micaschists. The sequence was grading to garnet

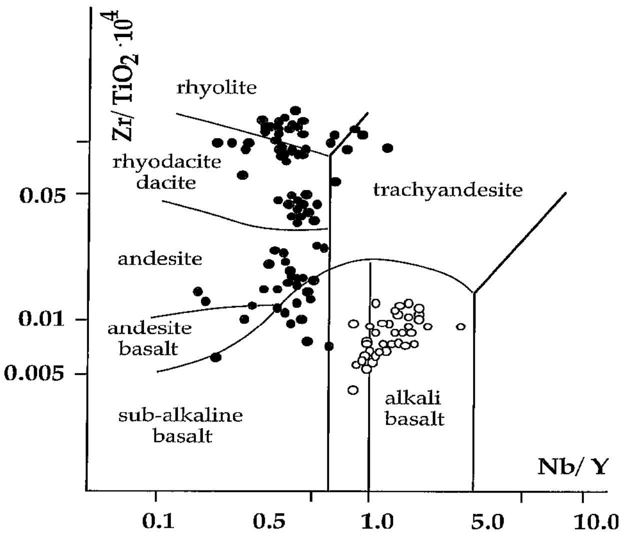


Fig.2 - (Al-K)+(Fe-Mg)-4(Ca) vs. (Al-K)-(Fe-Mg)-2Na after De La Roche (1978) for the phyllites of the Mandanici Unit from the Peloritani Mountains. Al-minerals (paragonite, chloritoid, garnet, staurolite) are common in these rocks which are derived from metamorphism of clay-like sediments.

phyllitic schists (T≥460°C); chloritoid phyllites; graphitic phyllites (T=420°C) and amphibole schists.

Another marble horizon occurs (i.e. "Marmi II", or paragonite marble, or Gioiosa Vecchia marble) (T=400°C), again with traces of graphite and locally of chlorite, titanite, etc. The succession was grading upwards to a series characterized by siderite-graphite- ankerite schists, (T=375°C) with frequent ore concentrations and Ti-rich metabasites. Lastly, marble and calcschists occurs (i.e "Marmi III", or lower marble horizon, or Giampilieri marble) wich evolve to quartz-phyllites. Between the upper and lower marbles may be recognized a number of stratabound ore-bearing horizons.

Metamorphic temperatures have been estimated according to the $\Delta^{13}C_{(carbonate\text{-}graph.)}$ geothermometer (Censi & Ferla, 1989; 1991).

According to Ferla (1983), Censi & Ferla(1983,1989) and Censi et al. (1991) an old shallow-sea sequence may thus be identified in an initial carbonatic layer . The sedimentation then evolved to a closed-sea type with euxinic features, where pelitic sediments rich in organic matter were associated with synsedimentary volcanic materials. This volcanism seems to be connected to the genesis of the associated metallic polysulfide deposits. Another limestone ad dolostone horizon occurs with traces of organic matter and a volcaniclastic fraction, later altered by the marine environment into smectites (and successively, during metamorphism, into paragonite ?). The succession was grading upward to a series characterized by Fe-carbonates with frequent volcano- sedimentary ore concentrations. A carbonate sedimentation again occurred, grading to pelitic to psammitic terrigenous deposits.

3.2. *Aspromonte Unit*

According to Ferla (1974,1983), the medium- high-grade metamorphic rocks of the Aspromonte Unit are considered as the original deeper portion of the old basement; on the contrary the Mandanici Unit is considered the upper portion of the same basement. The Aspromonte Unit is composed of a thick sequence of micaschists and paragneisses, up to migmatites. Augen-gneisses and amphibolites also occur locally.

The reconstruction of the original stratigraphic sequence can only be attempted by subtracting the polymetamorphic effects which affected these rocks. Again according to Ferla (1983), large parts of the above-mentioned sequence occur in the Mandanici or Aspromonte Units because the contact between the units cuts obliquely the old (overturned) succession. The static recrystallization also obliterates previous assemblages, making difficult a distinction between the original differences between the various sequences of the Aspromonte Unit (or Nappe).

According to the stratigraphic reconstruction, a thick flysch succession occurs, mainly consisting of graywackes (Fig.3) (Ferla, 1972; Atzori et al. 1976), with some arenitic and rarer pelite intercalations. These metapelites record the static crystallization of andalusite, cordierite and fibrolite, overprinted on the older garnet, staurolite and kyanite mineral assemblage (Ferla & Negretti, 1969). Very minor intercalations of marbles and Ca-silicatic felses with pyroxene, wollastonite, garnet, plagioclase, epidote and titanite (Atzori, 1969; Ferla 1978; Sassi, 1989), also occur.

Plagioclase-hornblende amphibolites and augen-gneisses also occur, which have been interpreted as old volcanoclastic products with chemistry similar ton that of

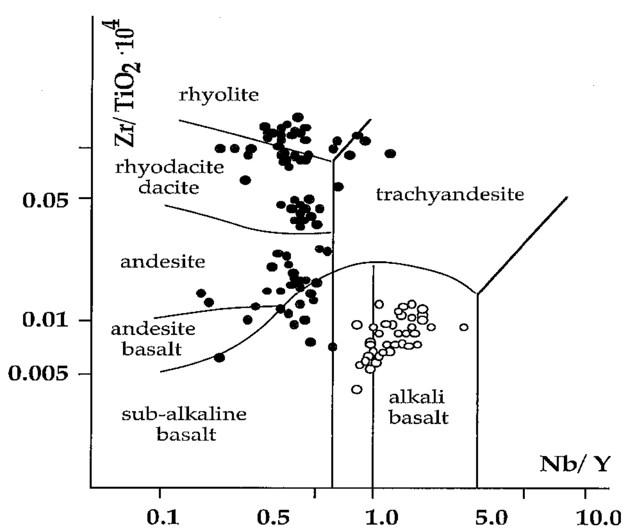


Fig.3 - Log (Na_2O/K_2O) vs. Log(SiO_2/Al_2O_3) after Pettijohn et al. (1972) for the biotite paragneisses from the Peloritani Mountains. These rocks are derived from metamorphism of flysch-like sediments (graywackes having a limited pelitic tendency).

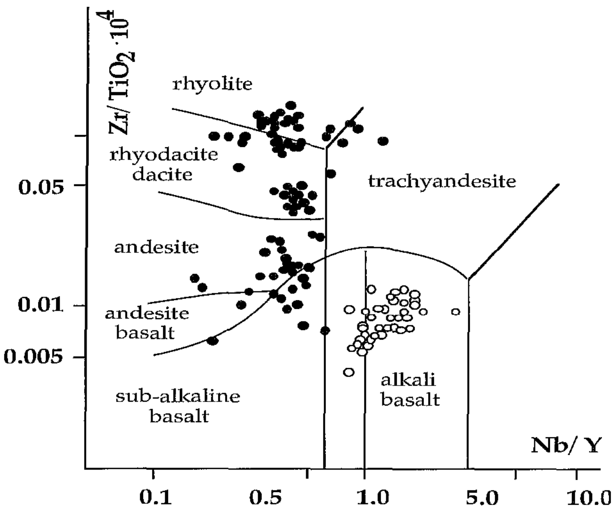


Fig.4. - R1 vs. R2 plot after De La Roche et al. (1980) for the amphibolites and augen gneisses of the upper units from the Peloritani Mountains. These rocks have been interpreted as old volcaniclastic products chemically near to active continental margin magmatism. R1=4Si-11(Na+K)-2(Fe+Ti); R2=6Ca+2Mg+Al.

island-arc or active continental margin igneous rocks (Ferla, 1978; Ferla & Azzaro, 1978a ; Atzori & Lo Giudice, 1982; Atzori et al., 1984a, 1984b;). Figs. 4 and 5 show the compositions of these rocks, which plot in essentially subalkaline magma field. also shows the chemistry. Some very basic amphibolites (tholeiitic to transitional up to alkaline) probably differ in age from the preceding ones (Ferla & Azzaro, 1978a; Atzori et al., 1988) (Fig. 4): according to first authors these amphibolites have vein position and are later in age regards the host rocks.

In the hypothesis of a single Hercynian, polyphase, metamorphic event the climax conditions of the metamorphic rocks of the Mandanici and Aspromonte units is shown in Fig. 6.

In the alternative hypothesis of a polymetamorphic history, proposed by Ferla (1978, 1983), the pre-Hercynian

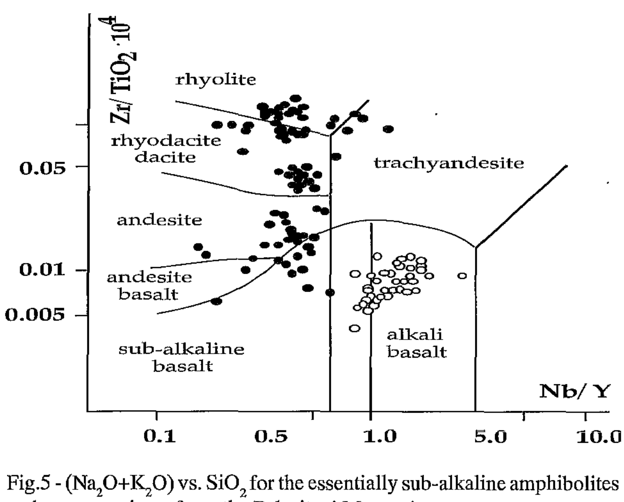


Fig.5 - (Na_2O+K_2O) vs. SiO_2 for the essentially sub-alkaline amphibolites and augen gneisses from the Peloritani Mountains.

metamorphism seems to have reached the kyanite isograd and the Hercynian metamorphism displays a higher metamorphic grade than the preceding one and slightly lower pressures. Brittle deformation then occurred, sometimes with traces of chloritization and sericitization, followed by static thermal metamorphism. The latter crystallization event took place under low pressure and with different spatial distribution regarding the former penetrative metamorphism,

The brittle deformation is considered to be linked to the uplift of the chain (Ferla et al. 1983), and different portions of it were after variously affected by heat flows from a hypothetical subduction plane. Post-metamorphic deformations put into contact rocks characterized by different metamorphic histories. These contacts are sometimes welded by the late Hercynian intrusions of monzogranites and peraluminous leucogranites dated around

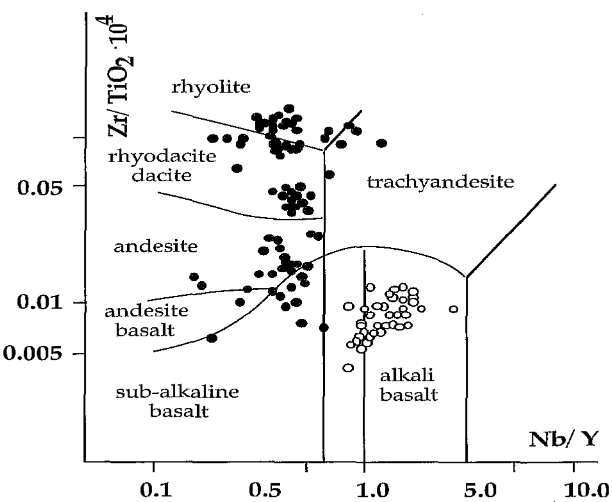


Fig.6 - P-T evolution of upper metamorphic units from the Peloritani Mountains. The spotted area represents the P-T range of the prograde eo-Hercynian or pre-Hercynian metamorphism deduced from mineral assemblages (Atzori & D'Amico; Ferla, 1974, 1978). The arrowed curves indicate the sin-S_2 and post-S_2 P-T paths with final retrograde episodes deduced also applying different geothermometers and geobarometers on various rocks (Ioppolo and Puglisi, 1988). G= Al_2SiO_5 triple point after Greenwood (1976); bio=biotite-in (Winkler, 1967); staurolite-in (Hoschek 1969); st'=staurolite+quartz out (Triboulet & Audren, 1985); K=kianite; S=sillimanite; A=andalusite.

295-270 Ma with a Rb/Sr method (D'Amico et al. 1982; Del Moro et al. 1982; Rottura et al. 1989).

As previously mentioned muscovite and biotite Rb/ Sr age determinations from paragneisses and augen-gneisses (292-262 Ma) suggest an uplift of the metamorphic basement, just after the emplacement of post-tectonic peraluminous granitoids (Atzori et al. 1990). Younger biotite Rb/Sr ages (173-137 Ma) may be interpretated as due either to Alpine overprint or alteration processes.

Lastly, in the Peloritani Mts., the Alpine metamorphic overprint has been only locally reported (Ferla & Azzaro 1978; Ferla, 1983; Ferla et al.1991; Messina et al. 1990).

4. Final Consideration and Open Problems

One of the main open question is the age of the synsedimentary orogenic volcanism in the lower units of the Peloritani Mts, considered lower Carboniferous by Ferla (1978) and Ferla & Azzaro (1978) and Ordovician or Cambro-Ordovician by Acquafredda et al. (1988, 1991), Spalletta & Vai (1989) related to the continuation of distensive regimes which had begun in the Cambrian. The Lower Carboniferous age is still uncertain, since it is based on the lower metamorphic grade of these rocks with respect to the Devonian (*Tentaculites*) and the presumed Silurian ones.

In the case of the metabasites associated to the Silurian (?) graphitic layers of S. Marco near Novara di Sicilia, Ferla (1978) recognized Ti-rich meta-volcanics with a tholeiitic to transitional basaltic composition.

The hypothesis that the Hercynian orogenesis in the Peloritani Mts. was linked to the evolution of an active Paleo-European continental margin is maintained by some authors; the presence of a calc-alkaline magmatism would support this hypothesis (Ferla, 1974; 1978; Ferla, et al, 1983). Such models best fit recent reconstructions of the European Hercynides, in which the Peloritani sector is considered part of the Paleo-European continental margin (Leeder, 1987; Wuster, 1988). Other hypotheses may include the involvement of sub-plates in the collision process, with considerable variations in the characteristics and timing of deformation between presently nearby sectors of the Hercynian orogen.

An alternative hypothesis is an intracontinental, entirely ensialic, evolution with partial involvement of the mantle during the Ordovician and Middle Devonian and the formation of megashear belts responsible for the Carboniferous orogenesis (Vai & Cocozza, 1986).

A further complication arises from recent results, which suggest a Panafrican event in large crustal sectors of the Mediterranean area (Peucat et al.1988; Rottura et al., 1988; Sacchi, 1989; Sassi & Zirpoli, 1989).

Furthermore, the problem of the augen-gneisses is still one of the many points to be clarified. These rocks, intercalated within medium/high-grade paragneisses, have sometimes been recognized to derive from rhyolitic volcanites and/or volcanoclastites (Atzori & Lo Giudice, 1982); the age of their ignimbritic deposition is assumed to be the same as that of the flysch like sediments (i.e., the paragneisses) in which they are interlayered. However, in other areas the augen-gneisses may also represent very old magmatic products reworked in a sedimentary environment, or plutonic masses (tectonically?) parallelized to the country rocks (Ferla, et al. 1983), or synkinematic intrusions (Andeatta,1955; Dubois & Truillet, 1966), or even magmatic bodies greatly modified by metasomatic events (Ferla e Negretti, 1969).

The complex chemistry of the augen-gneisses may certainly be attributed to several causes, including old sedimentary mixture of various materials. Metasomatic processes, magma hybridization, etc. could also be speculatively considered.

The intrusive hypotheses may be supported by other geological, structural and geochemical data. In this case, support may be given to the hypothesis of an Ordovician basic to persilicic magmatism, where intrusion of these melts into the pre-Hercynian basement gave rise to local volcanic episodes in the cover (Ferla & Rotolo, 1991); furthermore all products of this magmatism should have been affected by a single schistosity during the Hercynian events.

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STRUCTURAL AND METAMORPHIC EVOLUTION OF THE BOCCHIGLIERO AND THE MANDATORICCIO COMPLEXES IN THE SILA NAPPE (CALABRIAN-PELORITAN ARC, SOUTHERN ITALY)

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Abstract

The Sila Massif, northern sector of the Calabrian-Peloritan Arc, is an Alpine nappe pile, which includes basement, ophiolitic and cover (carbonate) thrust sheets.

The uppermost tectonic element, the Sila Nappe, consists of a composite metamorphic basement, intruded by late-Variscan (ca. 300 Ma) granitoids (the Sila Pluton) and overlain by a Mesozoic sedimentary cover (the Longobucco Sequence).

The metamorphic basement of the Sila Nappe consists of two portions, characterized by different metamorphic grade. The higher-grade portion (Monte Gariglione Complex), which includes the rocks so far ascribed to the kinzigitic, Polia-Copanello and/or Monte Gariglione units, is mainly composed of granulites, extensively migmatized and retrogressed under amphibolite facies conditions. The lower-grade portion is a composite unit, in which two complexes may be distinguished, characterized by comparatively higher (Mandatoriccio Complex) and lower (Bocchigliero Complex) metamorphic grade. The two complexes show different metamorphic characteristics and evolutions, acquired prior to the thermal metamorphism produced by the late-Variscan granitoids.

The Bocchigliero rocks exhibit a quite simple evolution, characterized by low- to very-low grade metamorphic conditions and only two folding phases. The Mandatoriccio rocks exhibit a polyphase deformation but a mono-metamorphic evolution, mainly acquired under low-pressure amphibolite facies conditions. Because of the occurrence of Palaeozoic (Cambro-Ordovician to Carboniferous) fossils, the Bocchigliero unit must have experienced only the Variscan metamorphism. On the contrary, the Mandatoriccio Complex may be envisaged as either its early- or pre-Palaeozoic basement or as an independent, possibly coeval, Variscan nappe. The last interpretation is supported by lack of any evidence of polymetamorphism in the Mandatoriccio rocks, which show a simple clockwise P-T trajectory peculiar of the Hercyno-type metamorphism. In particular, this P-T trajectory is consistent with unroofing processes, developed during an extensional tectonic regime of probable late-Variscan age.

Key words : *metamorphic evolution, structural analysis, mineral chemistry, extensional tectonics, Calabrian-Peloritan Arc.*

1. Geologic and Tectonic Interpretation of the Sila Massif

The Sila Massif, northern sector of the Calabrian-Peloritan Arc, is a nappe pile, including basement, ophiolitic and cover (carbonate) nappes (Dubois, 1976).

The uppermost tectonic element is a thick crystalline nappe (the Sila Nappe), consisting of a composite metamorphic basement, intruded by late-Variscan (around 300 Ma, Borsi & Dubois, 1968; Sutter et al. 1988) granitoids (the Sila Pluton), and overlain by a Mesozoic sedimentary cover (the Longobucco Sequence, Magri et al. 1965).

Two contrasting tectonic model have been suggested for the setting of the Sila Massif. On one side, Dubois (1967; 1976) envisages a relatively simple tectonic model, characterized by an Alpine nappe pile, in which the uppermost unit is the Sila Nappe. This nappe consists of medium- to high-grade metamorphic rocks (Piccarreta & Zirpoli, 1969a and b; Colonna & Zanettin Lorenzoni, 1970; 1972) overlain by lower-grade metamorphic rocks and intruded, mostly along their contact, by late-Variscan granitoids. On the other side, Amodio Morelli et al. (1976) and De Vivo et al. (1978) subdivide the Sila Nappe of Dubois (1976) into three units: 1) the Polia - Copanello Unit (the Dioritic-kinzigitic Formation of some authors), consisting of medium- to high-grade metamorphic rocks; 2) the Monte Gariglione Unit, consisting of medium-grade metamorphic rocks with migmatites, intruded by late-Variscan granitoids; 3) the Longobucco-Longi-Taormina Unit, consisting of low-grade metamorphic rocks, intruded

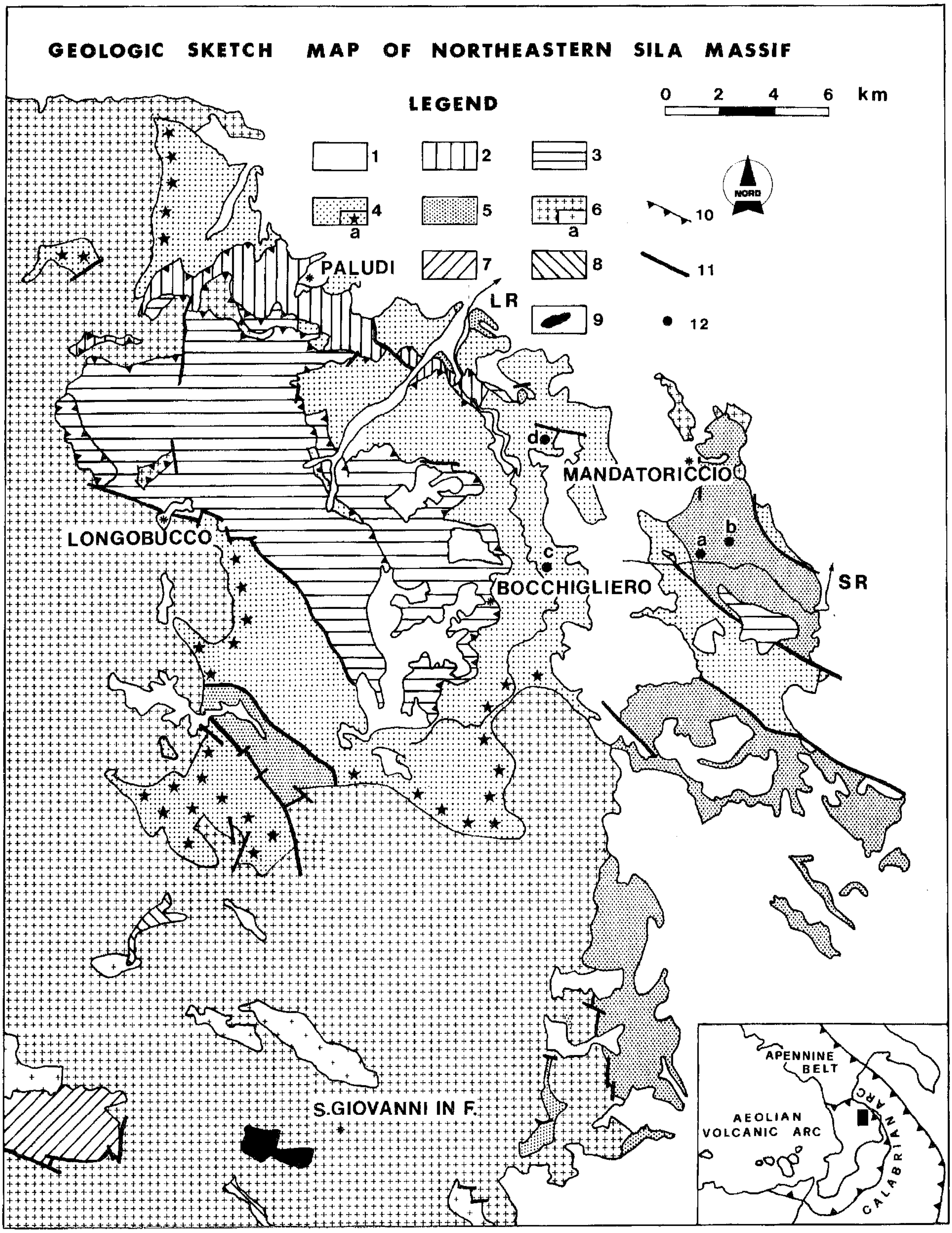

Fig. 1 - Geologic sketch map of the northeastern Sila Massif. [From the geologic Map of Calabria 1:25.000 (sheets Rossano and S. Giovanni in Fiore, (1971), Gurrieri et al. (1978), Young et al. (1986); Valdemarin (1989), and personal observations].
1 = Quaternary; 2 = Caenozoic Paludi Formation; 3 = Mesozoic Longobucco cover Sequence; 4 = Bocchigliero basement Complex, a = contact metamorphism overprint; 5 = Mandatoriccio basement Complex; 6 = Calc-alkaline and peraluminous (a) Permian granitoids; 7 = Gabbro and diorite; 8 = High-grade metamorphic rocks of the Monte Gariglione basement Complex; 9 = Metabasite and meta-ultrabasite; 10 = Thrust of "Apenninic" age; 11 = Faults ; 12 = locations of the studied rock samples. LR = Torrente Laurenzana, SR = Torrente Sappo.

as well by late-Variscan granitoids. The Longobucco Unit is considered as the lowermost thrust sheet of the northern Calabria Alpine nappe pile, tectonically emplaced below the others with opposite vergence, i.e. African instead of European. This tectonic interpretation has been further complicated by Gurrieri et al. (1978), Lorenzoni & Zanettin Lorenzoni (1979; 1983), Zanettin Lorenzoni (1982), Acquafredda et al. (1988), who consider part of the Sila Massif as a segment of a still rooted Variscan Chain.

Recent data by Barovero (1988), Valdemarin (1989) and Sutter et al. (1988) support the relatively simple tectonic setting, suggested by Dubois (1976). Therefore, the Sila Massif is here considered as an Alpine nappe pile: it consists of three main continent-derived crystalline nappes, which overly ophiolitic or ophiolitiferous nappes, resting on top of Apenninic carbonate nappes, only exposed as tectonic windows. In this interpretation, the crystalline nappes, from bottom to top, are: the Bagni phyllitic unit (Dietrich & Scandone,1972), the Castagna gneissic unit (Dubois, 1966; Colonna & Piccarreta, 1975), and the Sila Nappe (Dubois, 1970; 1976).

The Sila Nappe includes, from SW to NE:

a) - the Monte Gariglione Complex; b) - the Sila granitoids; c) - the Bocchigliero and the Mandatoriccio Complexes: d) - the Longobucco Mesozoic sedimentary cover.

The Monte Gariglione Complex includes all the crystalline rocks previously ascribed to the Polia-Copanello and the Monte Gariglione units (Dietrich et al., 1976). It consists of granulites extensively migmatized and retrogressed under amphibolite facies conditions (Dubois, 1976; Valdemarin, 1989).

The Bocchigliero Complex (De Vivo et al., 1978) consists of a low- to very-low-grade metamorphic sequence, dated palaeontologically by Bouillin et al. (1984; 1987) as Cambro-Ordovician to lower Carboniferous in age.

The Mandatoriccio Complex consists of medium-grade metamorphic rocks, characterized by a porphyroblastic texture.

The Sila granitoids, which are mainly exposed between the Monte Gariglione Complex on one side and the Bocchigliero and Mandatoriccio Complexes on the other side, developed a double contact aureole, especially well defined against the very low-grade Bocchigliero rocks (Dubois, 1976; Gurrieri et al., 1978; Barovero, 1988).

The crystalline basement of the Bocchigliero and the Mandatoriccio Complexes is unconformably overlain by the Triassic to Liassic Longobucco sequence, which mainly consists of silicoclastic turbidites (Magri et al., 1965; Sturani, 1973; Lanzafame & Tortorici, 1980; Young et al., 1986).

Given these different and conflicting interpretations of the geology and the tectonic subdivisions of the Sila Massif, this paper deals with the northeastern Sila, a key area for the whole Massif, where the two tectonic units, defined by De Vivo et al. (1978) are exposed. A detailed structural and petrographic study of these two units was performed, in order to understand their tectonic relationships and to reconstruct the two relevant P-T paths, essential for any geodynamic interpretation.

2. Geologic and Tectonic Setting of the Bocchigliero and the Mandatoriccio Complexes.

The presence of two different tectonic units in the lower-grade metamorphics of the eastern Sila was reported for the first time by De Vivo et al. (1978). These authors described in the pre-Mesozoic crystalline basement of the Longobucco Unit two tectonic units: a lower one, consisting of phyllites with interlayered Devonian meta-limestone and meta-greywake, and an upper one, composed of higher-grade paragneiss, micaschist and amphibolite. These two units were named by Gurrieri et al. (1978) Bocchigliero and Mandatoriccio units, respectively.

The two lithostratigraphic Bocchigliero and Mandatoriccio Complexes are exposed in the northeastern Sila (Fig. 1). The present contact between the two complexes usually consists in a fault, or is complicated by the east-verging thrusts of the Oligo-Miocene Apenninic tectonic phase. Only on the right side of Torrente Laurenzana, about 4 km south of the village of Mandatoriccio, the original tectonic contact appears to be preserved. This subhorizontal contact indicates that the Bocchigliero Complex is overlying the Mandatoriccio Complex. The age of this contact is most likely Variscan, owing to the lack of low-grade mylonites and brittle deformation, typical of the Alpine tectonic contacts in the Sila Massif.

The Mandatoriccio Complex consists of amphibolite-facies metapelite, minor orthogneiss and rare marble and metabasic rocks, intruded by late-Variscan granitoids. The metapelites are characterized by the presence of cm-sized porphyroblasts of staurolite, andalusite and cordierite.

The Bocchigliero Complex, on the contrary, exhibits a low- to very low-grade metamorphic characteristic. Primary sedimentary features are frequently preserved and remnants of Cambrian to Devonian acritarchs were discovered in several localities (Bouillin et al., 1984; 1987). This complex consists of slate, meta-sandstone and meta-conglomerate, interbedded with meta-volcanics of andesitic to rhyolitic composition (Acquafredda et al., 1988; 1991).

Both complexes were intruded by late-Variscan (ca. 300 Ma) epiplutonic granitoids (Borsi & Dubois, 1968; Sutter et al. 1988), which developed widespread thermal metamorphic aureoles. The contact metamorphism, especially marked in the Bocchigliero metapelite, produced the growth of poikiloblastic andalusite and cordierite, very similar to the porphyroblasts of the same mineral phases regionally developed in the Mandatoriccio Complex.

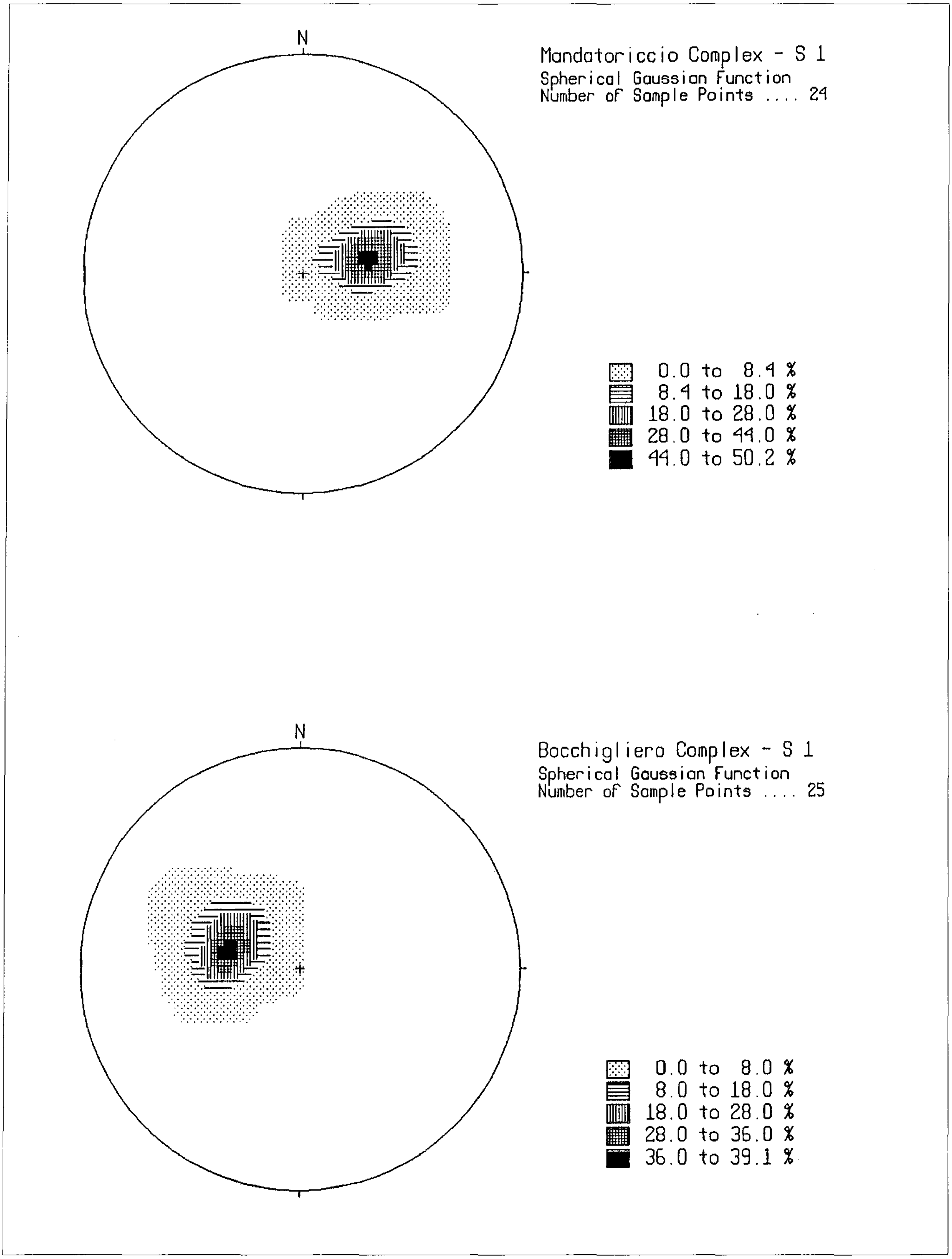

Fig. 2 - Structural data for the main foliation poles in the Mandatoriccio (a) and the Bocchigliero (b) Complexes. Schmidt equal area net.

3. Petrography and Microstructure

3.1 The Mandatoriccio Complex

For the petrographic and mineral chemistry studies, representative metapelite and orthogneiss samples were collected out of the late-Variscan thermal metamorphic aureole in the valley of the Torrente Sappo (a in Fig. 1) and about 1 km north of the village of Mandatoriccio (b in Fig.1).

The largely prevailing micaschist and paragneiss consist of quartz, muscovite, biotite, chlorite, of porphyroblastic, albite, staurolite, andalusite, cordierite, and of accessory ilmenite and titanite. The local occurrence of fibrolitic sillimanite and fine-grained garnet was also reported by Lorenzoni & Zanettin Lorenzoni (1979).

Both micaschist and paragneiss exhibit a well developed foliation (S1), defined by the preferred orientation of sheet silicates, ilmenite laths and lens-like polycrystalline quartz aggregates. The grain size is heterogeneous for the presence of millimetric to centimetric porphyroblasts of andalusite, cordierite, staurolite and albite.

The main foliation S1 usually has a gentle attitude, quite consistent all over the unit, as shown by the poles of Fig. 2a, where a strong maximum, trending to the east, is defined. The coeval fold axes b1 are rather spread in the NW and SE quadrant, whereas their dip values are usually lower than 20°.

The main regional foliation results from the transposition of a previous foliation. This is suggested by the presence of:

1) - relict millimetric intrafolial folds;

2) - bimodal distribution of the sheet silicates;

3) - syn-kinematic albite porphyroblasts with a sigmoidal S_i (cf. Bell et al., 1986), defined by very-fine-grained trails of quartz, muscovite and opaques, joining with S_e (S1) (Fig. 3);

4) - cleavage domains with pervasive shearing deformation and domains with microlithons, where simple shear and pure shear prevailed, respectively. The folding phase F1 developed under non-coaxial strain.

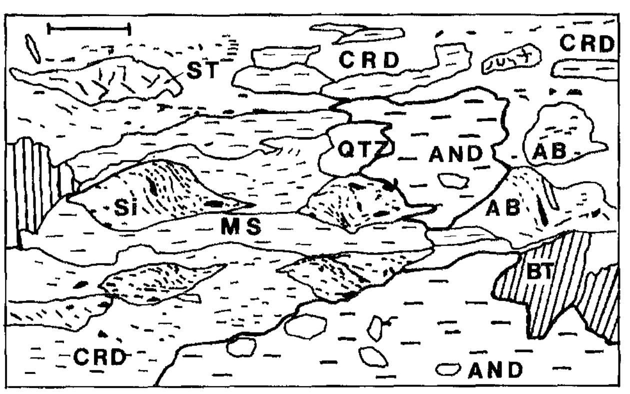

Fig. 3 - Syn-kinematic albite porphyroblast characterized by a sigmoidal Si foliation defined by very fine trails of Qtz, Ms and Ilm. Bar scale : 0.5 mm.

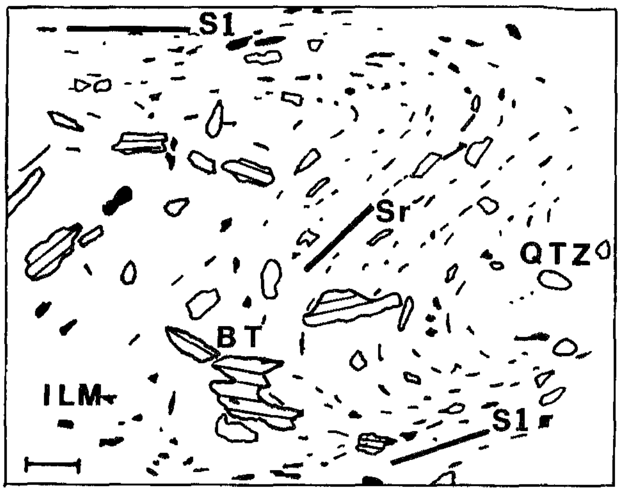

Fig. 4 - Relics of isoclinal folds preserved within an andalusite porphyroblast. S1 = main regional foliation, Sr = relict tectonic foliation transposed by F1. Bar scale : 0.01 mm.

S1 was overprinted by a second folding phase (F2), characterized by open folds and a crenulation cleavage, lacking axial plane foliation. The axes b2 are subhorizontal and N-S trending, while the axial planes are steep and NNW dipping. Minerals in the folds hinges b2 show intracrystalline deformation.

The folding phase F2 is associated to microfractures. Along these fractures, water was introduced into the rock, producing a significant mineral retrogression. This suggests that F2 occurred at shallow crustal levels, when dilatation processes, induced by tectonic decompression, were active.

In metapelite, chlorite occurs either as flakes with preferred orientation, included in the andalusite porphyroblasts (Chl I), or as late stage product of the biotite alteration (Chl II).

On the other hand, part of the biotite appears to have developed mimetically after Chl I. Most biotite, however, grew as randomly oriented post-kinematic flakes, cutting across the main foliation S1. The presence of biotite inclusions in the andalusite and cordierite porphyroblasts suggests that its growth started before their development (cf. Vernon, 1978).

Relics of intrafolial isoclinal folds are preserved within andalusite and cordierite porphyroblasts, which grew post-kinematically across S1 (Fig. 4).

Andalusite and cordierite are in equilibrium, but a slightly earlier andalusite growth is suggested by its local inclusion in cordierite porphyroblasts.

On the contrary, staurolite is older, because included in, and partly replaced by, the andalusite porphyroblasts (Fig. 5). However, staurolite as well is post-kinematic with respect to the main foliation S1. Staurolite is locally zoned, with the colour intensity closely related to the Ti content. This character, since long known and described, appears to

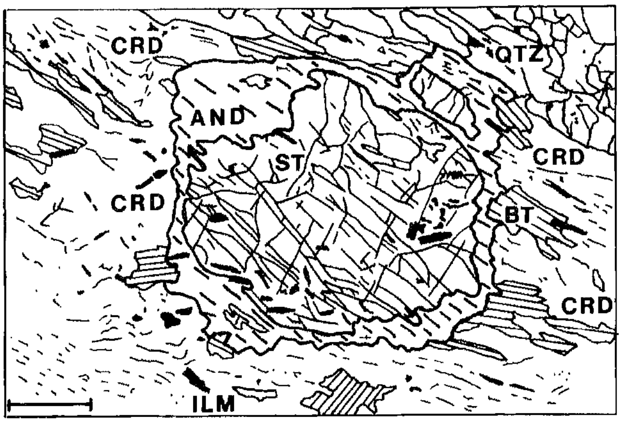

Fig. 5 - Staurolite relic, replaced by coronitic andalusite, included in a cordierite poikiloblast. Bar scale : 0.5 mm.

be typical of staurolite from regional metamorphism characterized by very high geothermal gradient (Tracy, 1986).

Andalusite is pinkish and shows sector and/or oscillatory zoning. It alters to a very fine grained felt of sericitic white mica.

Cordierite is mostly replaced by a very fine grained aggregate of sheet silicates, where potassic white mica prevails over chlorite. In some altered rock samples, cordierite exhibits a complex repeated twinning: microprobe analyses show a variable, unusually high, potassium content, suggestive of a very early stage of cordierite alteration to K-white mica.

The tectono-metamorphic evolution, as inferred from the above described microstructural relations, is summarized in Fig. 6a.

The orthogneiss, with microaugen structure, is more massive and consists of quartz, plagioclase (30% An), biotite, K-feldspar, minor muscovite, and accessory zircon, apatite and opaque ores. The orthogneiss exhibits a crenulation cleavage associated to tight folds which deform a pre-existing tectonic foliation. This pervasive deformation is accompanied by a general mineral re-crystallization, which extensively obliterates the original igneous fabric. The K-feldspar augen are igneous relics, which include small euhedral crystals of the magmatic plagioclase. On the contrary, the igneous quartz is completely converted to lens-like aggregates of syn-kinematically recrystallized grains. The igneous biotite is replaced by polycrystalline aggregates of pale-brown biotite II and titanite.

3.2 *The Bocchigliero Complex*

Samples were collected close to the bridge on the Torrente Laurenzana and along the Basilicò Valley (Fig. 1, c and d), i.e. in two areas that escaped the thermal metamorphism produced by the Sila pluton intrusion.

The most significant Bocchigliero rock type is a slate, with the main foliation gently dipping towards SE (Fig. 2b). The slate mainly consists of quartz, phengitic muscovite, albite, anatase and black organic matter (graphitoid). Locally, chlorite and/or calcite also occur. The slate exhibits a well developed platy foliation, defined by the preferred orientation of muscovite and graphitoid films. The tectonic nature of the main foliation is suggested by the occurrence of rare intrafolial isoclinal folds which transpose the original sedimentary compositional layering. The layering consists of phyllitic beds, locally enriched in organic matter, alternating with meta-arenite and meta-rudite layers rich in quartz, feldspar and carbonate.

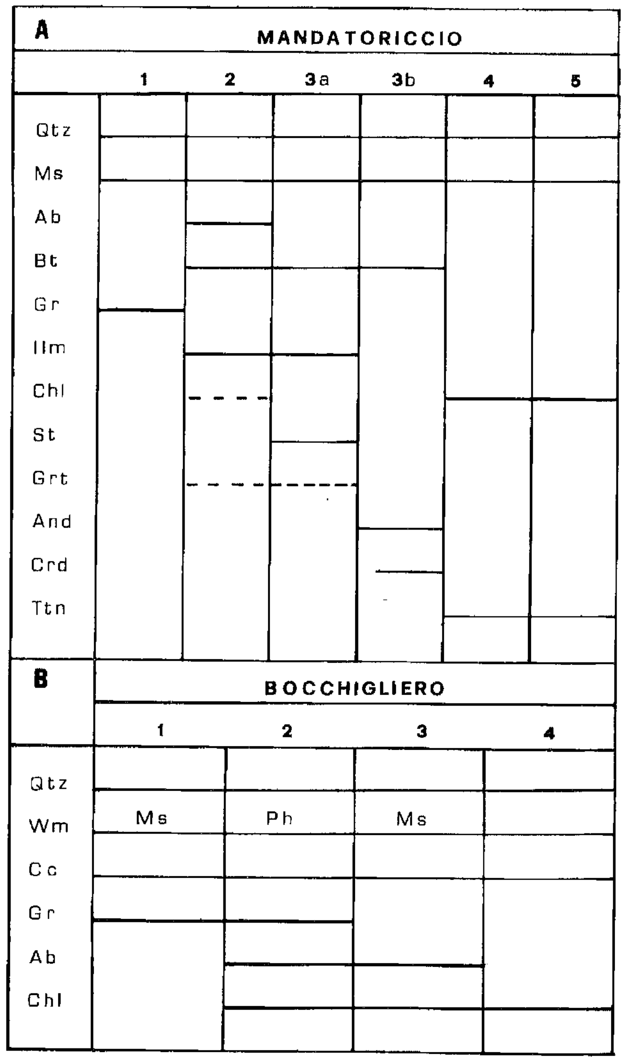

Fig. 6 - Relations between mineral growth and folding phases in the Mandatoriccio and the Bocchigliero Complexes. The thermal metamorphic minerals are neglected.

A: Mandatoriccio : 1 = earliest relict mineral assemblage; 2 = syn-kinematic mineral assemblage connected with the main transposition foliation (S1); 3a and 3b = postkinematic higher-P and lower-P mineral assemblages, respectively; 4 = synkinematic mineral assemblage connected with the late crenulation cleavage (F2); 5 = very late retrogressive minerals of possible Alpine age. B: Bocchigliero : 1 = detrital minerals; 2 = syn-kinematic mineral assemblage connected with the main foliation; 3 = synkinematic mineral assemblage connected with late kinks; 4 = very late retrogressive minerals of possible Alpine age.

In meta-arenite and meta-rudite, the original sedimentary fabric is better preserved and the coarser-grained detrital micas can be easily distinguished from the finer-grained metamorphic ones.

The low- to very-low grade metamorphism suggested by the mineral assemblage is consistent with the type of deformation, which is mainly controlled by pressure solution mechanisms.

The main foliation is deformed by a system of dm- to m-spaced kinks with sub-horizontal, NE-trending, axes and steep axial planes dipping both to NW and SE.

The moderate retrogression associated to this deformation phase produced albite, chlorite and sericitic white mica. Thus, the Bocchigliero slates exhibit a planar composite fabric, defined by the low angle intersection of the sedimentary Ss with the main tectonic foliation. The tectono-metamorphic evolution of the Bocchigliero Complex is summarized in Fig. 6b.

4. Mineral Chemistry

The analyses were performed with a LINK-EDS equipped Cambridge SEM. Natural silicates and oxides were used as standards.

The accelerating voltage was 15 Kw and the integration time 50 s. Each analysis is the average of at least tree points. All analyses were recalculated using the NORM program by Ulmer (1986). The mineral compositions are expressed as atoms per formula unit (p.f.u.). Unless otherwise specified, the mineral symbols are according to Kretz (1983). The location of the analysed samples is shown in Fig. 1 (a, b, c and d), whereas the composition of representative minerals is reported in Tables 1, 2 and 3. White micas (Tab. 1) and biotites (Tab. 3, an. 1-6) were calculated on the basis of 12 O, chlorites (Tab. 2) on the basis of 11 O and staurolites (Tab. 3, an. 7-9) on the basis of 47 O.

Dioctahedral micas

The paragonitic molecule, always lower than 23 wt %, averages 7 wt % for the Bocchigliero Complex and 20 wt % for the Mandatoriccio Complex. The margaritic molecule is lacking.

The Ti content is lower than 0.06 p.f.u., while the value of the Na/(Na + K) ratio is lower than 0.23. These characteristics are in agreement with those reported in the literature for medium- to low-grade white micas (Guidotti, 1984, with references).

As a whole, potassic white micas from the Mandatoriccio Complex are quite homogeneous low-celadonite muscovites with Si ranging from 3.02 to 3.12 atoms p.f.u. (cf. Fig. 7 a and 7 b).

In particular, there are not significant differences

	1	2	3	4	5	6	7	8	9	10	11	12
SiO2	46.47	48.39	46.67	48.99	48.74	48.25	46.64	46.85	46.35	46.37	46.63	47.13
TiO2	.61	.00	.67	1.15	1.09	.00	.33	.50	.71	.00	.00	.00
Al2O3	30.56	32.56	30.70	27.21	27.76	32.97	36.84	36.48	35.41	35.45	35.65	35.33
Fe2O3	5.08	1.49	5.31	2.71	2.31	.93	.50	1.01	1.09	.54	1.28	.00
FeO	.02	.03	.00	1.26	1.50	.33	.00	.00	.57	.43	.00	.90
MgO	1.39	1.68	1.63	2.64	2.58	1.38	.37	.37	.45	.61	.47	.56
Na2O	.90	.49	.82	.00	.53	.44	1.48	1.40	1.05	.41	.00	.38
K2O	9.92	10.56	10.01	11.32	10.91	10.45	8.91	8.76	9.89	10.97	10.61	11.05
H2O	4.44	4.51	4.48	4.44	4.45	4.50	4.56	4.57	4.52	4.49	4.51	4.51
Total	99.39	99.71	100.29	99.72	99.86	99.25	99.63	99.94	100.04	99.27	99.15	99.86
Si	3.1380	3.2113	3.1189	3.3045	3.2819	3.2116	3.0673	3.0711	3.0702	3.0973	3.0981	3.1297
Ti	.0310	.0000	.0337	.0583	.0552	.0000	.0163	.0246	.0354	.0000	.0000	.0000
Al	2.4320	2.5465	2.4179	2.1630	2.2029	2.5863	2.8553	2.8182	2.7642	2.7906	2.7914	2.7649
Fe3	.2581	.0742	.2671	.1376	.1168	0.466	.0247	.0499	.0545	.0273	.0639	.0000
Fe2	.0011	.0019	.0000	.0711	.0842	.0186	.0000	.0000	.0314	.0241	.0000	.0500
Mg	.1399	.1662	.1624	.2655	.2590	.1369	.0363	.0362	.0444	.0607	.0466	.0554
Na	.1178	.0630	.1062	.0000	.0692	.0568	.1887	.1779	.1348	.0531	.0000	.0489
K	.8543	.8937	.8531	.9738	.9369	.8871	.7473	.7323	.8355	.9345	.8990	.9358
OH	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000
xMg(FeII+)	.992	.989	1.000	.789	.755	.881	1.000	1.000	.586	.716	1.000	.526
xMg(Fe tot)	.351	.686	.378	.560	.563	.678	.594	.420	.341	.542	.422	.526
Al (IV)	.862	.789	.881	.695	.718	.788	.933	.929	.930	.903	.902	.870
Al (VI)	1.570	1.758	1.537	1.468	1.485	1.798	1.923	1.889	1.834	1.888	1.890	1.895
Charge Def.	.000	.000	.050	.000	.000	.000	.017	.030	.000	.000	.049	-.009

Tab. 1 - Representative microprobe analyses and atomic proportions on the basis of 12 ox. of detrital (No. 1 - 3) and metamorphic (No. 4 - 6) white micas from the Bocchigliero Complex and of medium- (No. 7 - 9) and low-grade (No. 10 - 12) muscovite from the Mandatoriccio Complex.

	1	2	3	4	5	6	7	8	9
SiO2	24.69	24.66	25.31	23.39	22.57	23.23	25.63	25.20	24.37
Al2O3	21.99	21.53	21.93	23.33	22.94	22.54	21.52	21.75	22.16
FeO	31.03	30.41	30.52	34.24	34.67	33.86	29.24	26.29	30.73
MnO	.44	.00	.00	.00	.00	.00	.00	.00	.41
MgO	11.48	11.88	11.59	8.87	8.86	9.51	12.19	11.71	11.54
H2O	11.27	11.17	11.26	11.11	11.00	11.06	11.33	11.22	11.23
	-------	-------	-------	-------	-------	-------	-------	-------	-------
Total	100.90	99.65	100.61	100.94	100.04	100.20	100.81	99.96	100.44
Si	5.2510	5.2934	5.3878	5.0443	4.9170	5.0344	5.4225	5.3852	5.1999
Al	5.5116	5.4465	5.5016	5.9294	5.8897	5.7569	5.3657	5.4776	5.5723
Fe2	5.5183	5.4584	5.4326	6.1745	6.3158	6.1361	5.1729	5.2338	5.4828
Mn	.0793	.0000	.0000	.0000	.0000	.0000	.0000	.0000	.0000
Mg	3.6399	3.8017	3.6781	2.8518	2.8776	3.0726	3.8449	3.7306	3.6709
OH	16.0000	16.0000	16.0000	16.0000	16.0000	16.0000	16.0000	16.0000	16.0000
xMg (Fe tot)	.397	.411	.404	.316	.313	.334	.426	.416	.401

Tab. 2 - Representative microprobe analyses and atomic proportions on the basis of 11 ox. of chlorites from the Bocchigliero (No. 1 - 3) and the Mandatoriccio (No. 4 - 6 = Chl I; No. 7 - 9 = Chl II) Complexes.

between white micas defining the main regional foliation and the younger white micas developed after andalusite (Fig. 7a).

The white micas from the Bocchigliero Complex are more heterogeneous. White micas defining the metamorphic foliation are phengitic (Fig. 7a) with Si contents ranging from 3.15 to 3.25 atoms p.f.u. (Fig. 7b). The phengitic character of the Bocchigliero white micas is in agreement with the results of the b_0 study by Dietrich et al. (1976).

On the contrary, the white mica flakes, recognized as detrital on the ground of microscopic examination, show a compositional range from muscovite to phengite.

Most likely this wide range is the consequence of incomplete metamorphic re-equilibration of original detrital muscovites.

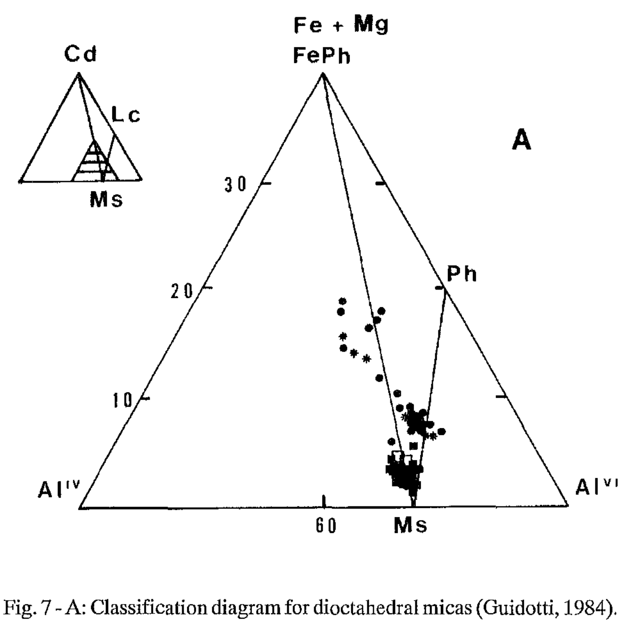

Fig. 7 - A: Classification diagram for dioctahedral micas (Guidotti, 1984). Cd = celadonite, Ms = muscovite, Lc = leucophyllite, Ph = phengite, FePh = Fe-phengite.

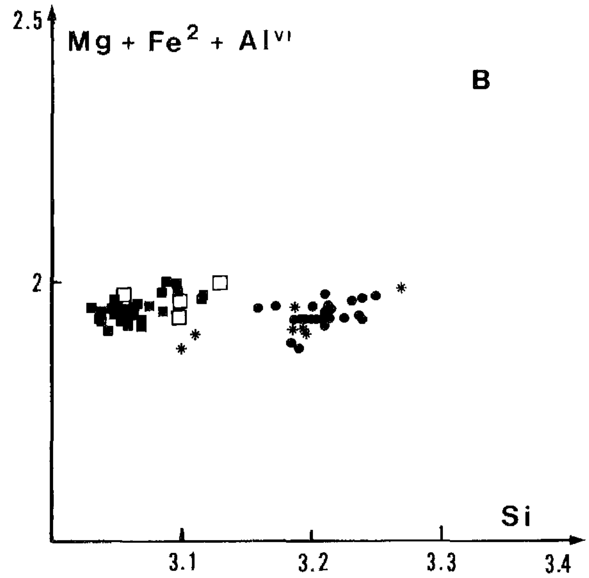

Fig. 7 - B: Si content variation of dioctahedral micas vs. the sum of octhedral cations. Two different clusters show muscovitic (squares) and phengitic (circle) compositions, respectively. Mandatoriccio Complex : Filled squares = white micas defining the main foliation, open squares = sericitic white micas developed after andalusite. Bocchigliero Complex: Filled circles = metamorphic white micas defining the main foliation (S1), asterisks = detrital white micas with variable amounts of overprinted celadonitic substitution.

	1	2	3	4	5	6	7	8	9
SiO2	34.52	36.59	35.91	35.75	33.80	34.87	27.58	26.56	27.16
TiO2	4.31	1.93	2.22	2.28	1.84	1.74	.55	.44	.00
Al2O3	20.21	20.41	20.60	20.84	20.48	20.17	55.87	56.81	55.77
FeO	22.95	21.78	21.99	22.54	23.20	23.10	12.41	12.92	13.00
MnO	.00	.00	.00	.00	.00	.00	.00	.00	.00
MgO	7.24	6.93	7.05	7.08	6.95	7.20	1.25	1.34	1.28
Na2O	.00	.00	.00	.44	.00	.00	.39	.00	.00
K2O	7.66	8.63	9.00	8.06	9.21	8.72	.00	.00	.00
H2O	3.92	3.88	3.88	3.91	3.81	3.85	2.13	2.13	2.11
	------	------	------	------	------	------	------	------	------
Total	100.81	100.15	100.65	100.90	99.29	99.65	100.18	100.20	99.40
	=====	=====	=====	=====	=====	=====	=====	=====	=====
Si	2.6391	2.8259	2.7698	2.7372	2.6560	2.7122	7.7448	7.4717	7.7064
Ti	.2478	.1121	.1288	.1313	.1087	.1018	.1161	.0931	.0000
Al	1.8209	1.8576	1.8725	1.8804	1.8966	1.8488	18.4895	18.8341	18.6488
Fe2	1.4671	1.4065	1.4183	1.4430	1.5244	1.5024	2.9140	3.0392	3.1034
Mg	.8252	.7979	.8107	.8081	.8142	.8349	.5233	.5620	.5620
Na	.0000	.0000	.0000	.0653	.0000	.0000	.2123	.0000	.0000
K	.7468	.8500	.8853	.7870	.9230	.8650	.0000	.0000	.0000
OH	2.0000	2.0000	2.0000	2.0000	2.0000	2.0000	4.0000	4.0000	4.0000
xMg(FeII+)	.360	.362	.364	.359	.348	.357			
xMg(Fetot)	.360	.362	.364	.359	.348	.357	.152	.156	.149
Al (IV)	1.361	1.174	1.230	1.263	1.344	1.288			
Al (VI)	.460	.683	.642	.618	.553	.561			

Tab. 3 - Representative microprobe analyses and atomic proportions of biotite (basis 12 ox.; No. 1 to 6) and staurolite (basis 47 ox.; No. 7 to 9) from the Mandatoriccio Complex.

Chlorites

In the Mandatoriccio Complex, the two chlorite generations show significantly different compositions. Chlorite I is relatively homogeneous and plots at the boundary between corundophyllite and ripidolite (Fig. 8), with Si ranging from 4.90 and 5.30 and X_{Fe} from 3.7 to 4.2. On the contrary, chlorite II exhibits a wider compositional range and plots from corundophyllite to pycnochlorite through the ripidolite field (Fig.8), with Si ranging from 5.00 to 5.70 and X_{Fe} from 4.3 to 5.0. In the Bocchigliero Complex, all analyzed chlorites are ripidolites with compositions similar to that of the chlorite II of the Mandatoriccio Complex (Fig. 8). All chlorites have octahedral vacancies as elsewhere observed by Holdaway et al. (1988).

Trioctahedral micas

The Mandatoriccio biotites are homogeneous and plot as a cluster, bounded by X_{Mg} values ranging from 0.3 to 0.5 and Al^{VI} values ranging from 0.3 to 0.7 atoms p.f.u., respectively (Fig. 9). The Ti content is lower than 0.15 atoms p.f.u. and Na is very low and always lower than 0.05 atoms p.f.u.. The Ti and Al^{VI} values are significant, because the analyzed biotites occur in assemblages, where the two elements are buffered by ilmenite (or titanite) and Al-silicate, respectively. The (Na + K) value is lower than 1.000 atom p.f.u., such as shown also by many biotites from similar metamorphic conditions (Guidotti, 1984).

Because octahedral cations are systematically lower (from 2.80 to 2.96) than the theoretic value of 3.00 atoms p.f.u., octahedral vacancies most likely occur in the Mandatoriccio Complex biotites.

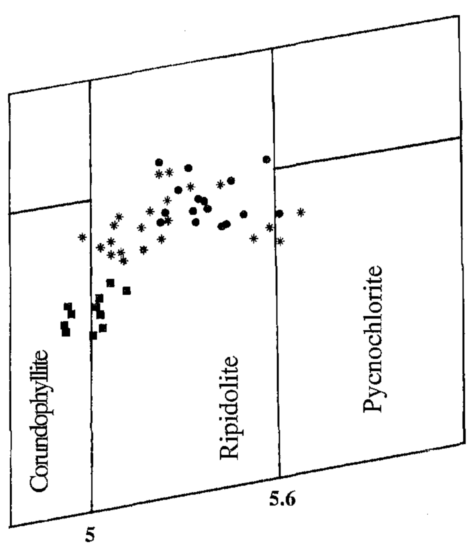

Fig. 8 - Classification diagram for chlorites (Hey, 1954). Mandatoriccio Complex: full square = Chl I, asterisk = Chl II. Bocchigliero Complex = filled circle.

In conclusion, the analyzed biotites show chemical composition consistent with that of similar two-micas bearing metapelites, re-equilibrated under medium-grade metamorphic conditions (cf. Guidotti, 1984).

Staurolite

Both fresh crystals and crystals partly replaced by andalusite were analysed, either included in cordierite or not. Except for Ti, staurolite is homogeneous with a very low X_{Mg} value, ranging from 0.13 to 0.17. The Ti content,

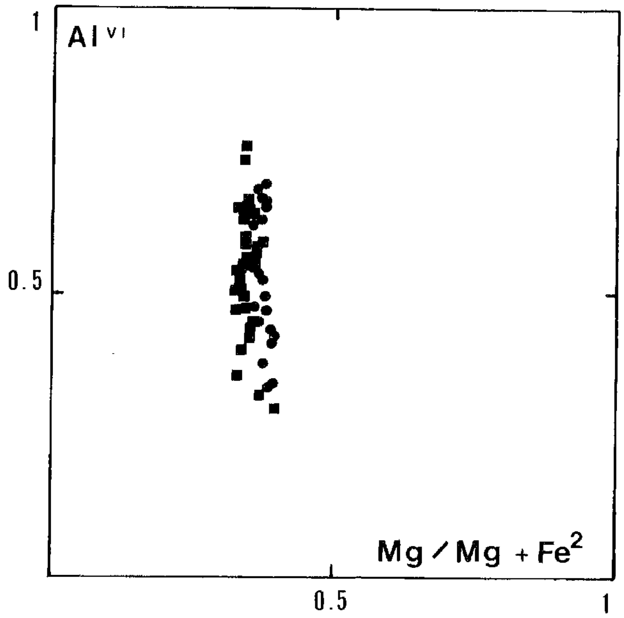

Fig. 9 - Classification diagram for biotites (Guidotti, 1984). Full squares = Mandatoriccio Complex; full circles = Bocchigliero Complex.

ranging from 0.170 down to values below detection, is proportional to the colour intensity. The Si content, always lower than 8.000, implies the presence of small amounts of Al^{IV}. Zn was never detected.

5. Metamorphic Evolution

A comparison between the two complexes clearly indicates a significantly different tectonic and metamorphic evolution, as to both the number of deformation phases and the metamorphic grade.

5.1 *The Mandatoriccio Complex*

The evolution of the Mandatoriccio Complex consists of at least three metamorphic stages, corresponding to different P - T conditions. Using the main tectonic foliation (S1) as a reference chronologic marker and applying the microstructural criteria, an earliest syn-kinematic mineral association was recognized, which consists of quartz + albite + chlorite I + muscovite (Si contents < 3.15 atoms p.f.u.) + ilmenite. The P - T conditions of this first event may be inferred from the equilibrium curve of the reaction Cld + Qtz = St + H_2O and from the stability fields of Ms, Ab and Ilm. As evident from the petrogenetic grid of Fig. 10, P < 4-5 kb and T < 500°-550° C may be envisaged for the oldest mineral assemblage, which corrisponds to metamorphic conditions lower than the upper greenschist facies.

This greenshist facies event is followed by a better constrained evolution, characterized by the successive development of the mineral assemblages St + Grt and And + Crd + Bt, respectively. This relative chronology is clearly shown by the occurrence of staurolite corroded and replaced by andalusite (Fig. 5). This decompressional evolution occurred under static conditions, because the St, And, Crd and Bt porphyroblasts cut across the main foliation and because S_i is consistent with S_e. The two successive assemblages indicate a progressive decompression at approximately constant medium T. The thermal peak at 550°to 600°C, suggested by the stability of Ms + Qtz and the lack of partial melting, was reached during the decompression event at P lower than 4 Kbar, as constrained by the coexistence of And + Crd. As a consequence, the geothermal gradient corresponding to the second event, was quite high (ca. 50° C/km). The record of this low P - medium T metamorphic event is quite uniform all over the Mandatoriccio Complex.

The metamorphic evolution of the Mandatoriccio Complex ended with a regional pervasive retrogression under low P and T conditions. The most significant hydration reactions, enhanced by the presence of the phase F2 folds, are: Crd + H_2O = Ms + Qtz + Chl II and And + Qtz + H_2O = Ms. At the same time, the biotite was replaced by Chl II and ilmenite by titanite.

5.2 *The Bocchigliero Complex*

The simpler metamorphic evolution of the Bocchigliero Complex is characterized by the low-grade syn-kinematic assemblage Qtz + Ab + Chl + phengitic mica + anatase, followed by a very-low-grade non-pervasive retrogression. During this retrogression, the phengitic white mica is replaced by sericitic muscovite.

The local preservation of acritarchs (Bouillin et al., 1984) suggests that during the metamorphic evolution the temperature never exceeded approximately 300°- 350° C. These temperatures are in agreement with the petrologic constraints. For such temperatures, the celadonitic content of phengitic white micas (Si = 3.25 atoms p.f.u.) indicates P in excess of 4 - 4.5 kbar, (Massone & Schreyer, 1987, Fig. 10).

Therefore, at the metamorphic climax the Bocchigliero Complex was characterized by a geothermal gradient significantly lower than that inferred for the Mandatoriccio Complex.

6. Discussion and Conclusion

As to the Sila Massif different opinions still exist about the number of tectonic units, their mutual geometric

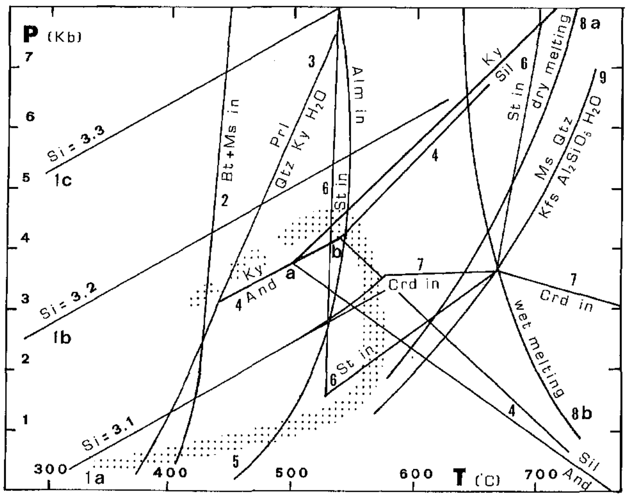

Fig. 10 - Petrogenetic grid for the Bocchigliero and the Mandatoriccio Complexes, northeastern Sila Massif (Calabrian-Peloritan Arc). The P-T path inferred for theMandatoriccio Complex is dotted.
1a, 1b, 1c = curves for the maximum stability of phengitic white micas with Si contents of 3.1, 3.2 and 3.3 atoms p.f.u., respectively (Massone & Schreyer, 1987); 2 = Nitsch (1970); 3 = Holland & Powell (1990); 4: a = Holdaway (1971), b = Bohlen et al. (1991); 5 = Spear & Cheney (1989); 6 = Ganguly (1972); 7 = Richardson (1968); 8 a and b = Thompson (1982); 9 = Chatterjee & Johannes (1974).

relations, and the age of emplacement (Alpine vs. Variscan).

In agreement with Dubois (1976), the Sila Massif is here considered as an Alpine nappe pile, whose uppermost portion is a single composite thrust, the Sila Nappe. This nappe consists both the higher-grade (kinzigitic or Polia-Copanello and Monte Gariglione Complexes of some authors) and lower-grade metamorphic rocks intruded by late-Variscan granitoids (the Sila pluton). The lower-grade metamorphic rocks belong to the Mandatoriccio and Bocchigliero Complexes, characterized by different tectono-metamorphic evolutions. Though the present contacts between the Mandatoriccio and the Bocchigliero Complexes are mostly marked by faults, in few localities (see map of Fig. 1) the Bocchigliero rocks appear to overly the Mandatoriccio rocks. These geometric relations between the two complexes are also supported by the presence of the Mesozoic Longobucco cover only on the Bocchigliero Complex and by the occurrence, in the lowermost sedimentary rocks of the Longobucco Sequence, of reworked fragments of the Bocchigliero rocks.

The presence in the Bocchigliero Complex of detrital muscovite and albite and pebbles of slates (Acquafredda et al., 1988) indicates that the sediments derive from the erosion of an older low-to very-low-grade metamorphic basement. The age of this metamorphic basement must be older than Cambrian, and possibly Cadomian, as suggested by Bouillin et al. (1984). On the other hand, the simple tectonic and metamorphic evolution, exhibited by the Bocchigliero rocks, suggests that it underwent only part of the common Variscan tectono-metamorphic history, either because it was situated at shallow crustal levels during the orogenic evolution or because it was located in a marginal position with respect to the main Variscan chain. This is in agreement with the Rb/Sr radiometric ages of 328 - 330 Ma determined by Acquafredda et al. (1991) on metavolcanics of the Bocchigliero Complex.

On the contrary, the Mandatoriccio Complex exhibits a polyphase tectonic and metamorphic evolution, characterized by at least two transposition folding phases and by an orogenic pervasive porphyroblastic mineral growth under low-pressure amphibolite facies conditions.

The mono- or poly-metamorphic nature of this complex is difficult to ascertain.

The Qtz + Ms + Ilm assemblage and the fine grain-size of the minerals, defining the relict tectonic foliation preserved in the And and Crd poikiloblasts, imply that this early tectonic foliation formed under consistent low-grade conditions. Therefore, it is probable that the Mandatoriccio Complex as well was affected by only one, although polyphase, tectono-metamorphic cycle. If this is the case, also the age of the Mandatoriccio metamorphism is Variscan. In this view, the clockwise P-T path shown by the Mandatoriccio Complex should be consistent with the Hercyno-type metamorphism described in central and southern Europe (Pin & Vielzeuf, 1983; von Raumer, 1988).

Therefore, the Mandatoriccio and the Bocchigliero Complexes show a tectono-metamorphic evolution, characterized by different P-T paths and structural patterns indicative of two different tectonic histories. In particular, the metamorphic climax conditions were attained in the two complexes at similar pressures (around 4 - 4.5 kbar), but at significantly different temperatures (300° - 350° C for the Bocchigliero Complex and ca. 600° C for the Mandatoriccio Complex), indicative of two distinct thermal gradients and, therefore, geodynamic environments. This suggests that the two metamorphic complexes were tectonically juxtaposed towards the end of the Variscan orogeny, but earlier than the emplacement of the granitoids of the Sila pluton. Alternatively, the Mandatoriccio Complex might be considered as the Cadomian metamorphic basement of the Bocchigliero Complex, as suggested by Bouillin et al. (1984). However, the evidence of a simple clockwise P-T path without remnants of distinct mineral assemblages, is in favour of a monocyclic evolution. In this interpretation, the real metamorphic basement of both the Mandatoriccio and the Bocchigliero Complexes is the Monte Gariglione Complex, in which the granulitic metamorphism may be Cadomian (or even older) and the amphibolite facies retrogression Variscan in age. Therefore, the age of 280-290 Ma obtained by Schenk (1980) in the Serre Massif and interpreted as evidence of a Variscan age for the granulitic metamorphism must be considered as cooling ages.

The actual P-T path that a rock follows is the result of a complex interplay of tectonic events and heat flow (Spear et al., 1984). In particular, retrograde P-T trajectories provide important constraints on the tectonic processes operative during orogeneses. For this reason, the compositional data of the minerals have been combined with the petrological constraints inferred from metamorphic parageneses to construct a relatively complete pre-Alpine P-T path for the Mandatoriccio and Bocchigliero Complexes.

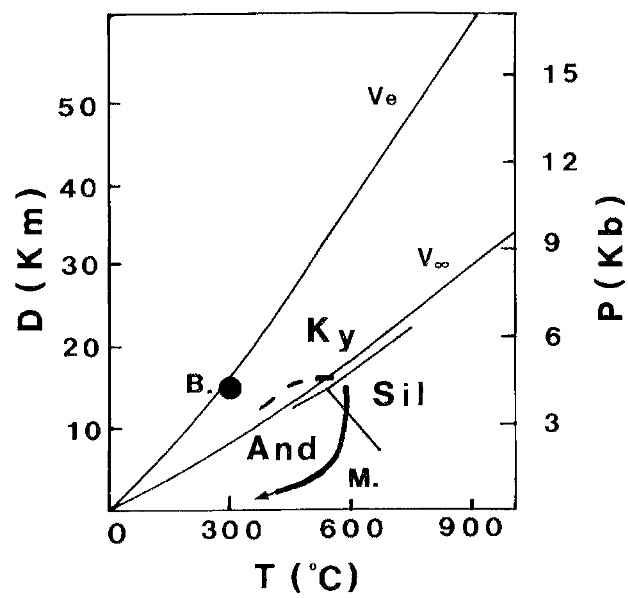

Fig. 11 - Variscan P-T path for the Mandatoriccio Complex (M.). For comparison, the metamorphic climax of the Bocchigliero Complex is also shown (full circle labelled B.). The V ∞ curve represents the undisturbed geotherm before thrusting, whereas the V ∞ curve is the steady-state geotherm that would be obtained after thrusting, if erosion was not to occur (from Thompson & England, 1984). The Al-silicates triple point is after Bohlen et al. (1991).

The climax metamorphic conditions of the Bocchigliero Complex plot close to the undisturbed geotherm of Thompson & England (1984) (V_e in Fig. 11), whereas the decompressional trajectory shown by the Mandatoriccio Complex plots below the maximally relaxed geotherm (V ∞ in Fig 11), namely in the field characterized by an extensional tectonic regime. Moreover, the P - T path of the Mandatoriccio Complex postdating the baric peak consists of an initial almost isothermal decompression followed by a marked cooling, necessarily coupled with a weak decompression. This path implies that the Mandatoriccio Complex decompressional path largely occurred at a relatively high-T and that cooling towards a more normal gradient did not occur untill the rocks were quite close to the surface.

In recent years several authors have shown the relationship between low-P high-T metamorphism and extensional tectonics, resulting in a thinning of the continental crust (Selverstone et al., 1983; Wickham & Oxburg, 1983; Sandiford & Powell, 1986; Thompson & Ridley, 1987; Gardien et al., 1990; Lardeaux & Spalla, 1990; Pinardon et al., 1990; Gibson, 1991). In the case of the Mandatoriccio Complex, the exhumation process must have occurred at a much faster rate than is usually achieved by erosion. This means that also these rocks might have been subjected to tectonic unroofing during an extensional phase.

It is difficult to date and to locate this event in the geodynamic evolution of the Calabrian-Peloritan Arc, because the the only available geochronological datum on the studied metamorphic rocks was produced by Borsi & Dubois (1968). They obtained a Rb/Sr age of 282 Ma on a metapelite of the Mandatoriccio Complex. If this age can be considered representative of the Mandatoriccio regional metamorphism, the low-P amphibolitic event can be assigned to the late-Variscan continental rifting.

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TOURMALINITES FROM THE TRIASSIC VERRUCANO OF THE NORTHERN APENNINES, ITALY

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Abstract

Tourmalinites are often present as pebbles within the basal metaconglomerate of the Middle-Upper Triassic Verrucano sequence of Tuscany. Two types were distinguished: 1) Acidic volcanites in which tourmaline fills vugs, fractures and replace pumices and phenocrysts; 2) Quartz-wacke (tourmalinized matrix up to 20-60 %). Type 1) has schorl: dravite ratio of 44:56 to 60:40 and uvite: dravite ratio of 12:88 to 37:63; type 2) shows schorl: dravite ratio of 44:56 to 48:52 and uvite: dravite ratio of 9:91 to 27:73. Only alkali-defect substitution $Na_x + Mg_y = Al_y + \square_x$ is widespread. Type 1) was produced by late magmatic fluids in volcanites; type 2) is of subacqueous continental-littoral exhalative origin.

Key words: *Northern Apennines, Tuscany, Trias, Verrucano, Tourmalinites.*

1. Introduction

Tourmalinites are widespread as pebbles within a quartzose metaconglomerate representing the lowermost level of the Middle Triassic Verrucano sequence of Tuscany ("Anageniti grossolane" member of the "Formazione della Verruca": Rau & Tongiorgi, 1974), the basal clastic deposits of the Alpine sedimentary sequence. Petrographic and microprobe analyses were carried out on several samples of tourmalinites collected in most of the Verrucano outcrops of Tuscany (from the Apuan Alps to the Monte Argentario; location in Fig. 1) to define their composition and genesis. These data are showed and discussed below.

2. Textural and Petrographic Data

The microscopic analyses reveal two main lithotypes among the tourmalinite clasts:

1 - Hematite-rich recrystallized acidic volcanites with abundant secondary tourmaline. These glassy rhyolitic ignimbrites and lava flows are related to the last extensional phase (Saalian phase, Permian in age) of the Hercynian Orogeny.

2 - Coarse to fine grained, often well sorted, quartz-wacke with a tourmalinized matrix, sometimes with hematite pigment.

In the first type of tourmalinite blue to green and green to greenish brown secondary tourmaline appears

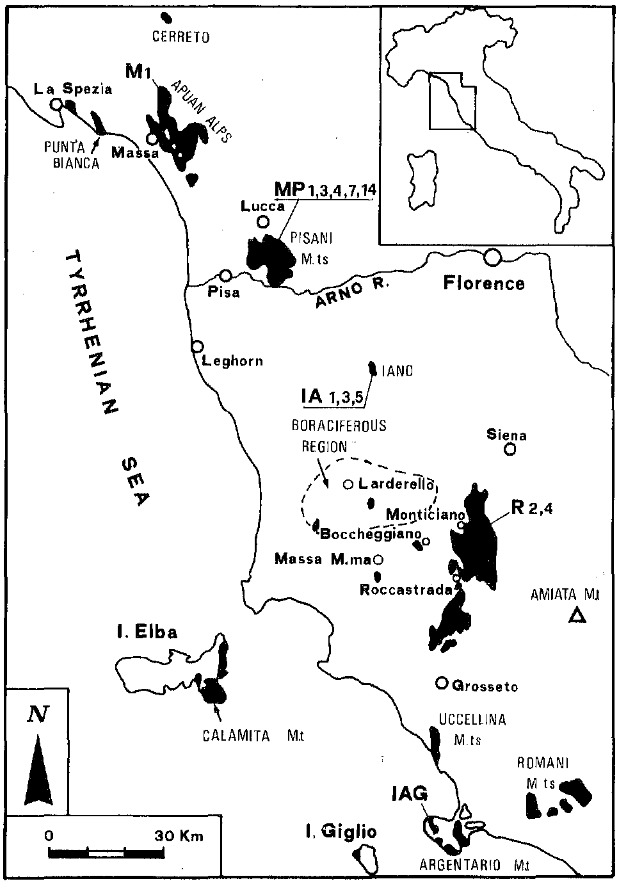

Fig. 1 - Regional distribution of the Paleozoic and Verrucano sequences and sample locations.

locally as fan-shaped aggregates, but generally as xenoblastic micro- to crypto-crystalline patches filling vugs and fractures and replacing phenocrysts, pumices and most of the groundmass. Tourmaline mineralization show a strongly variable distribution of colour and pleochroism: sharp contacts between patchy aggregates with greenish blue to blue colour or greenish brown to brown colour. Fading zoning was locally observed. The tourmalinization is a clearly late process that almost completely obliterates the original textures.

In the second type of tourmalinites the matrix is almost entirely made up of generally microcrystalline xenoblastic bluish-green to greenish-brown tourmaline locally evolving into idioblastic elongated prisms. Colour and pleochroism show uniform and homogeneous distribution, without zoning. In many cases the tourmaline-rich matrix partially replaces clast rims giving a sinuous or irregular pattern to the border of the detrital quartz.

These late processes between matrix and clast increased the volume percent of the matrix up to the present day value of 20 to 60 %.

In both types of tourmalinites, secondary cavities and veins are often observed and filled by quartz and/or idioblastic short prism or needles of tourmaline, with a grain size greater than that of the surrounding groundmass tourmaline.

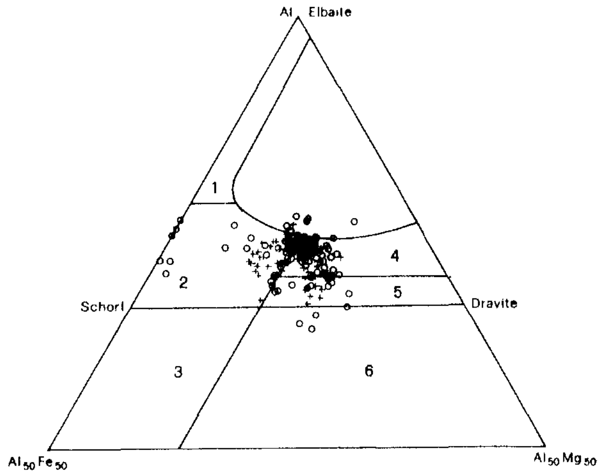

Fig. 2 - Al-Fe_{tot}-Mg diagram for Tuscan tourmalinites. Field numbers as in Henry & Guidotti (1985): 1) Li-rich granitoid pegmatites and aplites; 2) Li-poor granitoids and their associated pegmatites and aplites; 3) Fe^{3+}-rich quartz-tourmaline rocks (hydrothermally altered granites); 4) metapelites and metapsammites coexisting with an Al-saturating phase; 5) metapelites and metapsammites not coexisting with an Al-saturating phase; 6) Fe^{3+}-rich quartz-tourmaline rocks, calc-silicate rocks, and metapelites. End members in molecular proportions.
Crosses: type 1 (volcanic exhalative); circles: type 2 (sedimentary exhalative).

3. Analytical Techniques and Results

Chemical analyses were made by means of a Cambridge Geoscan twin-spectrometer WD electron microprobe operated at 15 kV and 15 nA current (Faraday cup). Reference standards were diopside for Si, Mg, and Ca; jadeite for Na and Al; orthoclase for K; rutile for Ti; and pure metals for Mn and Fe. X-ray data were corrected using a standard ZAF procedure coded by one of the authors (G.C.) for an on-line microcomputer. Since boron was not determined with the electron microprobe, the stoichiometry was based on an oxygen number of 24.5 (this assumption is supported by the very low fluorine content which never exceeds 0.15 wt% corresponding to about 0.1 atomic aboundance in the formula).

The analyses reveal no remarkable differences between the two groups of tourmalinites (as shown in Fig. 1, Fig. 2 and in Tab. 1).

Both types are schorl-dravitic and closely approach an average composition of schorl 50%-dravite 50%. In particular Type 1) shows schorlite: dravite ratio of 44:56 to 60:40 and uvite: dravite ratio of 12:88 to 37:63; type 2) shows a schorl: dravite ratio generally within the range 44:56 to 48:52 with sporadic values up to 59:41 or pure schorlitic compositions. The uvite: dravite ratio is variable between 9:91 and 27:73.

Both types show a little but strongly variable replacement of Si by Al in tetrahedral site, full occupancy of Y site, remarkable vacancy of X site (averages of 0.693 and 0.689 for type 1 and 2 respectively) well correlated with Al^{VI} (average of 0.421 and 0.429 for type 1 and 2). The latter feature indicates that alkali defect substitution $Na_x + Mg_y = Al_y + \square_x$ (site substitution (7) of Henry & Guidotti 1985) is the most important replacement. Finally

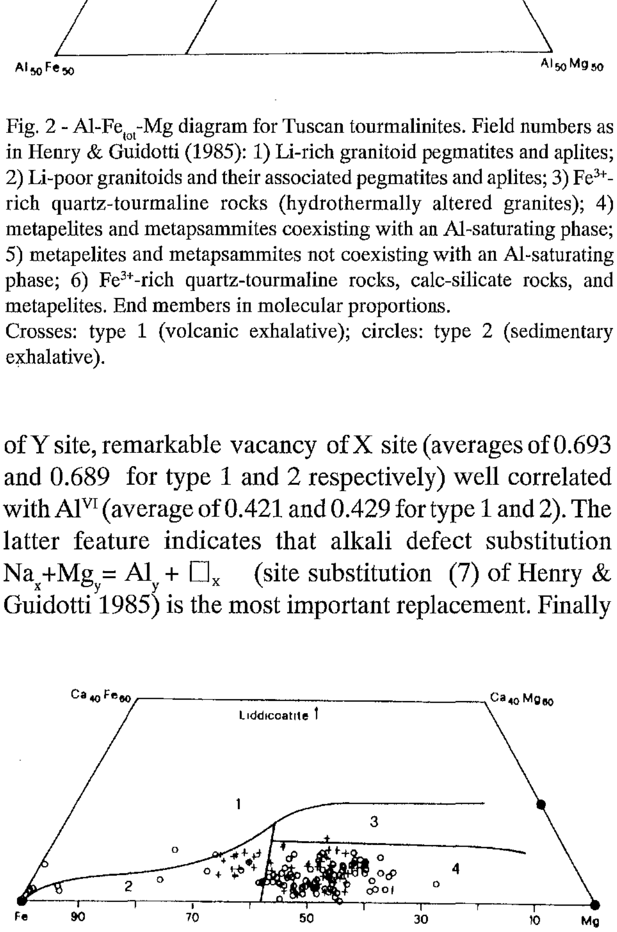

Fig. 3 - Ca-Fe_{tot}-Mg diagram for Tuscan tourmalinites. Field numbers from Henry & Guidotti (1985): 1) & 2) as in Fig. 2; 3) Ca-rich metapelites and metapsammites, and calc-silicate rocks; 4) Ca-poor metapelites, metapsammites, and quartz-tourmaline rocks. End members in molecular proportions.

in both types some spot analyses have Al_2O_3 high and slightly lower or greater than SiO_2 content, FeO below 7-9% and MnO in the range 0.1-0.3% (see sample MP 1.5 in Tab. 1). All these features, taken as a whole, indicate the presence of an elbaite component.

The tourmalines here described are markedly enriched in Fe and show a wider range of Fe/Mg ratio in comparison with tourmalinites of the Bottino mining district studied by Benvenuti et al., 1989, 1991). The latter

	IA 1.3 (S)	MP 1.5 S)	IA 3.2 (V)	MP 7.3 V)	R 2.4 (S)
SiO_2	35.98	34.93	34.96	33.46	35.12
AlO_3	34.39	33.55	32.01	33.22	33.65
TiO_2	0.28	0.43	0.69	0.44	0.04
FeO	7.00	8.55	9.05	12.50	15.48
MnO	0.09	0.24	0.03	0.12	0.06
MgO	6.14	5.10	5.94	3.43	0.06
CaO	0.31	0.62	1.77	1.47	0.18
Na_2O	2.08	1.47	1.41	1.25	1.41
K_2O	0.00	0.02	0.00	0.06	0.00
TOTAL	86.27	84.91	85.86	85.95	86.00
STRUCTURAL FORMULA ON THE BASIS OF 24.5 OXYGENS					
Si	5.85	5.83	5.82	5.66	5.96
Al^{IV}	0.15	0.17	0.18	0.34	0.04
Al^{VI}	6.00	6.00	6.00	6.00	6.00
Ti	0.03	0.05	0.09	0.06	0.01
Al	0.44	0.43	0.10	0.28	0.69
Fe^{+2}	0.95	1.19	1.26	1.77	2.20
Mn	0.01	0.03	0.01	0.02	0.01
Mg	1.49	1.27	1.47	0.87	0.02
YTOTAL	2.92	2.97	2.93	3.00	2.93
Ca	0.05	0.11	0.32	0.27	0.03
Na	0.66	0.47	0.45	0.41	0.46
K	0.00	0.01	0.00	0.01	0.00
XTOTAL	0.71	0.59	0.77	0.69	0.49

Tab. 1 - selected electron microprobe analyses (wt %) of the Tuscan Verrucano tourmalinites, (V)= volcanic; (S)= sedimentary.

show Fe/Fe+Mg ratios ranging from 33 to 40 with sporadic values up to 45 and 52; tourmalines of this paper on the contrary have a Fe/Fe+Mg ratio generally higher than 44 and sometimes surprisingly in the sedimentary type can reach almost pure schorlitic compositions.

4. Origin of the Tourmalinites

As regards the genesis of tourmalinites, textural features, mineral assemblages of the rock, chemical compositions of tourmaline and type of occurrence may help to elucidate the different origins of the two types of Tuscan tourmalinites.

As summarized by Appel (1984) tourmalinites occur within:

1) granitic rocks and their differentiates (pure schorlitic);
2) placer deposits;
3) evaporitic or sabkha environment (pure dravitic);
4) subacqueous-exhalative chemical sediments.

Tourmaline in type 1) (Permian volcanites) was deposited by late hydrothermal fluids within slowly cooling acidic volcanites.

For tourmaline of type 2) a sedimentary-evaporitic origin (e.g. Moine et al., 1981; Déchomets, 1983; Appel, 1984; Brown & Ayuso, 1985; Chown, 1986) can be ruled out for the following reasons: 1) tourmaline never shows pure dravitic or dravitic-uvitic compositions and anhydrite and scapolite are missing. 2) marine carbonate levels are sporadic and evaporites unknown in the Permian sequences of Tuscany. A subacqueous exhalative origin in a continental-litoral environment can easily be proposed for type 2) tourmalinites considering the overlap of the compositional fields between type 1) and 2) (see Fig. 1, 2) and the sporadic appearance of pure schorlitic and elbaite-bearing composition in some spot analyses of type 2) tourmalinites.

The appearance of such extreme compositions reveal an almost complete prevalence of the magmatic-component over the Mg-rich marine and/or fresh water-component within the subaqueous hydrothermal system.

In the triangular diagram Al-Fe-Mg most of the Tuscan tourmalinites fall in the fields 2, 4, 5 of Henry and Guidotti (1985) corresponding respectively to Li-poor granitoids and their differentiates and to metasedimentary rocks. Some spots fall in field 6 corresponding mainly to Fe^{3+} - rich quartz-tourmaline rocks.

Summarizing all the Tuscan tourmalinites, with sporadic exceptions, show compositions intermediate between schorl and dravite, generally considered typical of a subacqueous exhalative origin (e.g. Plimer, 1983; Appel, 1984).

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SEISMOTECTONIC IDENTITY OF THE SOUTHERN ADRIATIC AREA

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Abstract

The Adriatic microplate is considered in literature as a block relatively undeformed and aseismic with unitary dynamical behaviour. Nevertheless such model is inadequate to justify the recent seismicity of the Adriatic basin. Recent mesostructural analyses are showing a continuous strain boundary all around the Southern Adriatic platform. Several structures were recognized continuously outcropping from the Southern Dalmatian coastline (Kotor zone) to Split-Sibenik area and, across the Adriatic isles, until the Tremiti Islands and the Gargano-Murge regions. Data point out a centripetal trend of the Dinaric and Apenninic units, delimited to the north by ductile and brittle strain belts. These belts seem to set apart the southern block and to allow the release from the northern structures. The agreement between the geological-structural and seismological data allows to identify the Southern Adriatic block as an indipendent crustal structure, divided from the northern one by strike-slip faults. The Southern Adriatic block is able to condition the structural evolution of the neighbouring areas.

Key Words: *Adriatic Sea, Southern Italy, Geodynamics, Seismology, Geodynamics*

1. Introduction

Several microplates are singled out in the Mediterranean area (McKenzie, 1972), following the complex collisional history between the African and European plates. The Adriatic microplate is an elongated continental block including the Po Plain and the Adriatic basin, surrounded by a series of strongly deformed belts. Its collision with the European plate gave rise to the uplift of the peri-Adriatic chains: the Alpine and Dinaric-Hellenic systems during Cretaceous-Eocene, the Western Alps during Eocene-Miocene and the Apenninic chain starting from Upper Miocene.

Among others, the Adriatic microplate, whose margins were identified on the basis of the epicentres distribution (Lort, 1971), has been interpreted also recently as a unitary and nearly aseismic block (e.g. Mantovani et al., 1985; Anderson & Jackson, 1987; Royden et al., 1987). Neverthless this model does not justify either the newly collected structural data in the area (Funiciello et al., 1991) or the recent seismicity localized by the Istituto Nazionale di Geofisica (ING, 1990) in the Central Adriatic basin.

2. Geodynamical Setting

Structural and geological data show that the Southern Adriatic block can be interpreted as homogeneous and brittle (Fig.1) (Funiciello et al., 1988; Montone & Funiciello, 1989), with generally low tectonized areas in its internal part (Salento, Murge) and highly deformed units outcropping along the borders. To the south-west the transition from the Adriatic to the Tyrrhenian domain is marked by a thrust chain (Apennines) derived from the shortening of the sedimentary wedge of a passive margin. This is related to the sinking of the Adriatic lithosphere with its entire thickness (Moretti & Royden, 1988). Several regional geological cross sections (Mostardini & Merlini, 1986), carried out with the contribution of seismic soundings for oil research, confirm this model.

The upper crust structural setting results from a series of mainly in-sequence thrusts according to ramp-flat duplex geometries. Several detachment layers have been recognized. The sole thrust was identified in the Triassic units, as suggested by the age of the oldest involved lithologies (evaporitic sequences) and by their rheology (Ogniben et al., 1975). The development of this chain has been started in Late Tortonian, as suggested by the closing age of the westernmost sedimentary cycles (Bigi et al., 1989).

The central part of the Apenninic chain is structured by NW-SE trending thrust and normal (not shown in Fig. 1) faults that indicate a NE-SW shortening; N-S right-lateral strike-slip faults and NW-SE left-lateral strike-slip faults

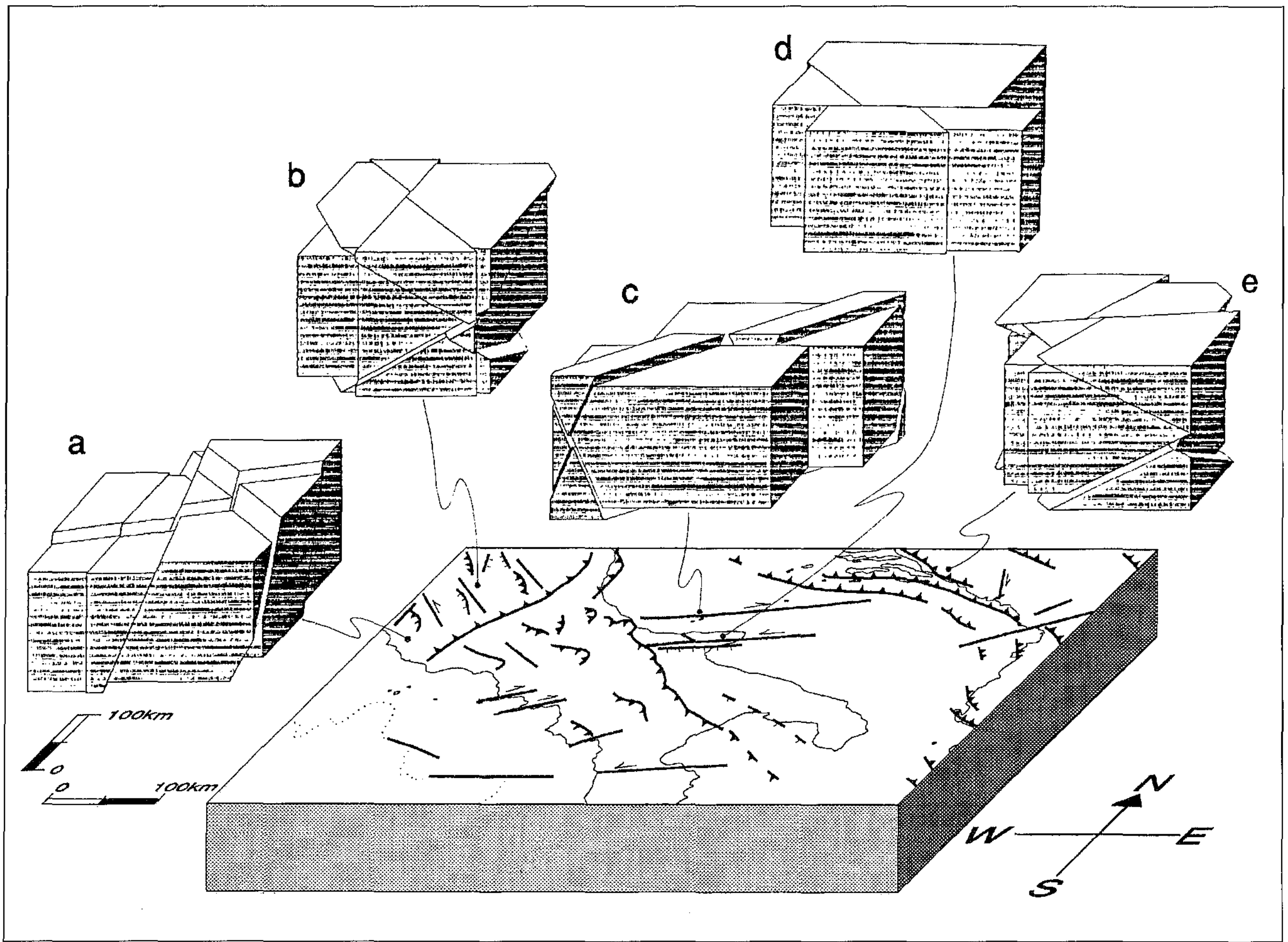

Fig. 1 - Summary of the results from structural data. Blocks show the main fault systems. a) Tyrrhenian margin: it is interested mainly by NW-SE and then N 10°W and N 70°E normal faulting; N-S striking right lateral strike-slip faults with moderate displacement seem to dissect all other features. b) Apenninic chain: this sector is structured by NW-SE trending thrust and normal (not shown) faults that indicate a NE-SW shortening; N-S right lateral strike-slip faults and NW-SE left lateral strike-slip faults dissect all other deformations. c) Tremiti Islands: push-up structure related to E-W right lateral strike-slip faulting. d) Gargano Promontory: NW-SE strike-slip and vertical (not shown) faults; E-W Mattinata fault clearly cut them; E-W normal faulting (not shown) seems to represent the last event. e) Dinarides: NW-SE thrust faults together with a wrench system indicate a NE-SW shortening.

dissect all other deformations (Boccaletti et al.,1982; Alfonsi et al., 1991a; 1991b; Mattei & Miccadei, 1991; Montone & Salvini, 1991). Geological and structural data from the northern margin were collected in the Gargano Promontory and Tremiti Islands. In the Tremiti Islands the deformations can be related to the presence of a push-up structure while in the Gargano Promontory the structural analysis evidenced NW-SE left-lateral strike-slip and vertical faults (not shown in Fig. 1). A regional E-W left-lateral strike-slip faulting clearly displaces all of them. Eventually an E-W faulting (not shown in Fig. 1) seems to represent the last event (Funiciello et al., 1988).

The north-eastern boundary of the Adriatic microplate is also characterized by a thrust chain (Dinarides). The age of deformation is generally considered older than the Apenninic one, while recent offshore data show the involving of terrigenous cycles of Upper Cenozoic age. The strength of the lithologies in the Dinaric units is higher than that of the Apenninic domain, as confirmed by mesostructural data. Fault analysis at outcrop scale pointed out the presence of NW-SE thrust faults together with a wrench system that both indicate a NE-SW shortening. The high percent of wrench tectonic deformations suggests a structural style related to a lithospheric thickening that does not involve subduction.

3. Seismic Activity

The historical earthquakes occurred in the Adriatic Sea are poorly documented and often mislocated. About a hundred earthquakes are known to be occurred in this area from 1000 to 1985 (Postpischl, 1985; ING, 1990), in particular in the Gargano offshore. Starting from 1975 the Seismic Catalogue of the Istituto Nazionale di Geofisica can be considered microseismic, i.e. every hypocentral location is based on the seismogram analysis instead of macroseismic considerations.

We particularly focused on the Central Adriatic basin, where an interesting seismic activity occurred during the last years. In January 1986 a seismic sequence occurred about 50 km to the north of the Tremiti Islands (main shock Mb=4.2). After about one year of quiescence, another sequence has been detected to the east-southeast of Gargano (April 1988, main shock Mb=5.3) and in October 1989 a moderate activity occurred again in the Tremiti Islands (main shock Mb=4.7) until 1990. In figure 2 the 1986-1990 seismic data with 3.0 minimum magnitude threshold and the Centroid-Moment Tensor focal solutions (Favali et al., 1990) relate the main structural features.

The focal solutions are consistent with the regional stress field that is mainly compressive in the eastern side and extensional in the Apennines (e.g. Udias, 1980; D'Argenio, 1988).

4. Discussion

The proposed geodynamical models of the Adriatic microplate as unique, undeformed and aseismic should be improved. The information derived from seismology, structural geology and geophysics singles out at least two domains separated by regional discontinuities at the Gargano latitudes.

The seismic activity occurred in the last years has pointed out the progressive activation of structures, which define the margins of a lithospheric block with unitary geodynamical behaviour. The boundaries of this block are the Southern Apennines to the south-west, the Dinarides-Hellenides to the north-east and the middle Adriatic deformation belt to the north (Mele et al., 1990).

The seismic energy release is different between the two parts of the Adriatic area. In the southern one high magnitude earthquakes (Mb>=6.0) are quite frequent while in the northern part they occur only in the Friuli-Carnia zone, where the compressive interaction between the Northern Adriatic and the European plates is acting (Slejko et al., 1987). In the southern part we modelled the interaction of the Adriatic lithosphere with both the Apennines and the Dinarides-Hellenides, with strong deformations of the sedimentary cover (fig. 3).

In the western sector, a westward lithospheric subduction is coherent with the deep seismicity in Eastern Tyrrhenian coastline (Giardini and Velonà, 1991), while

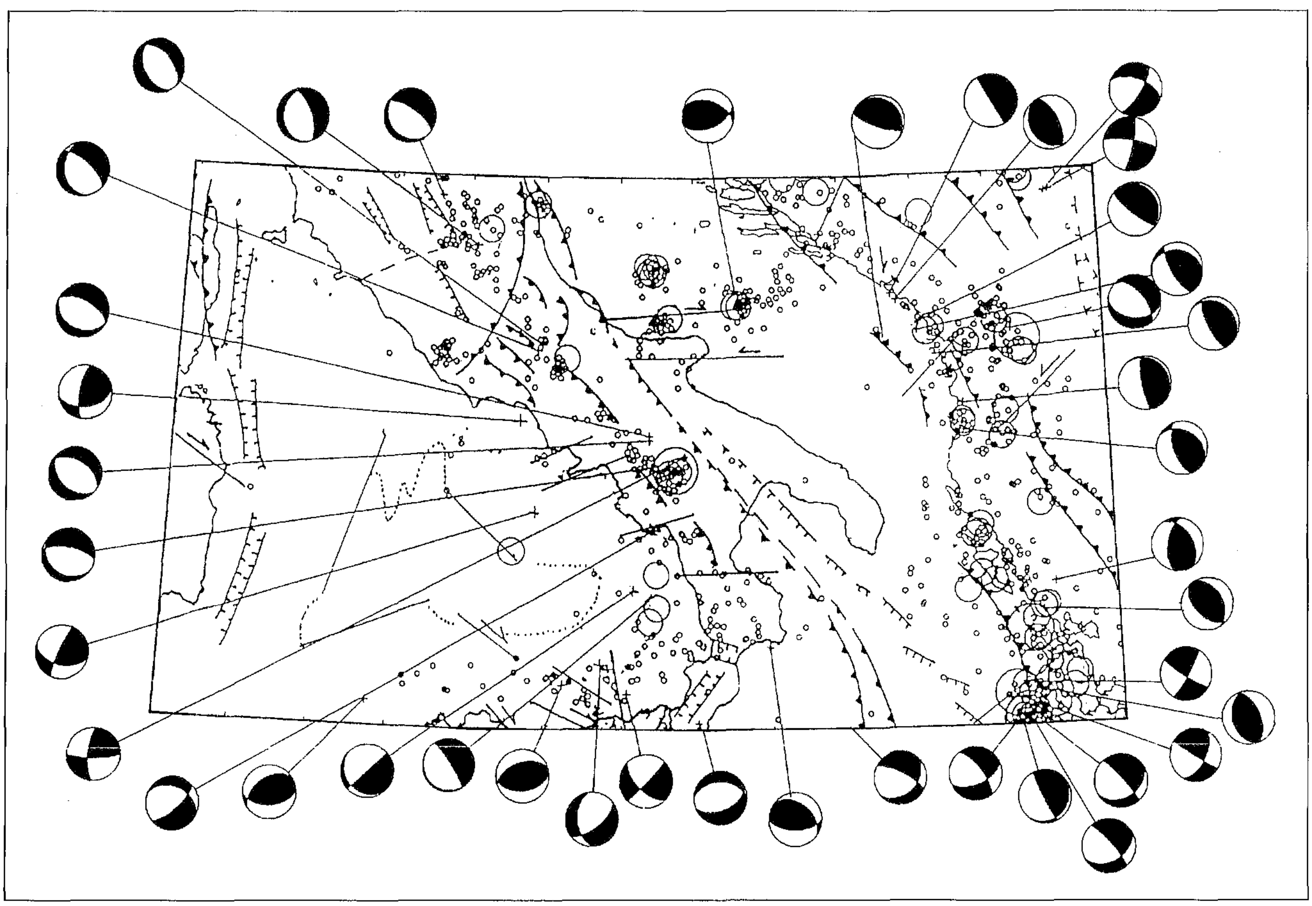

Fig. 2 - 1986-1990 seismic activity (M>=3.0), CMT focal solutions and main tectonic features.

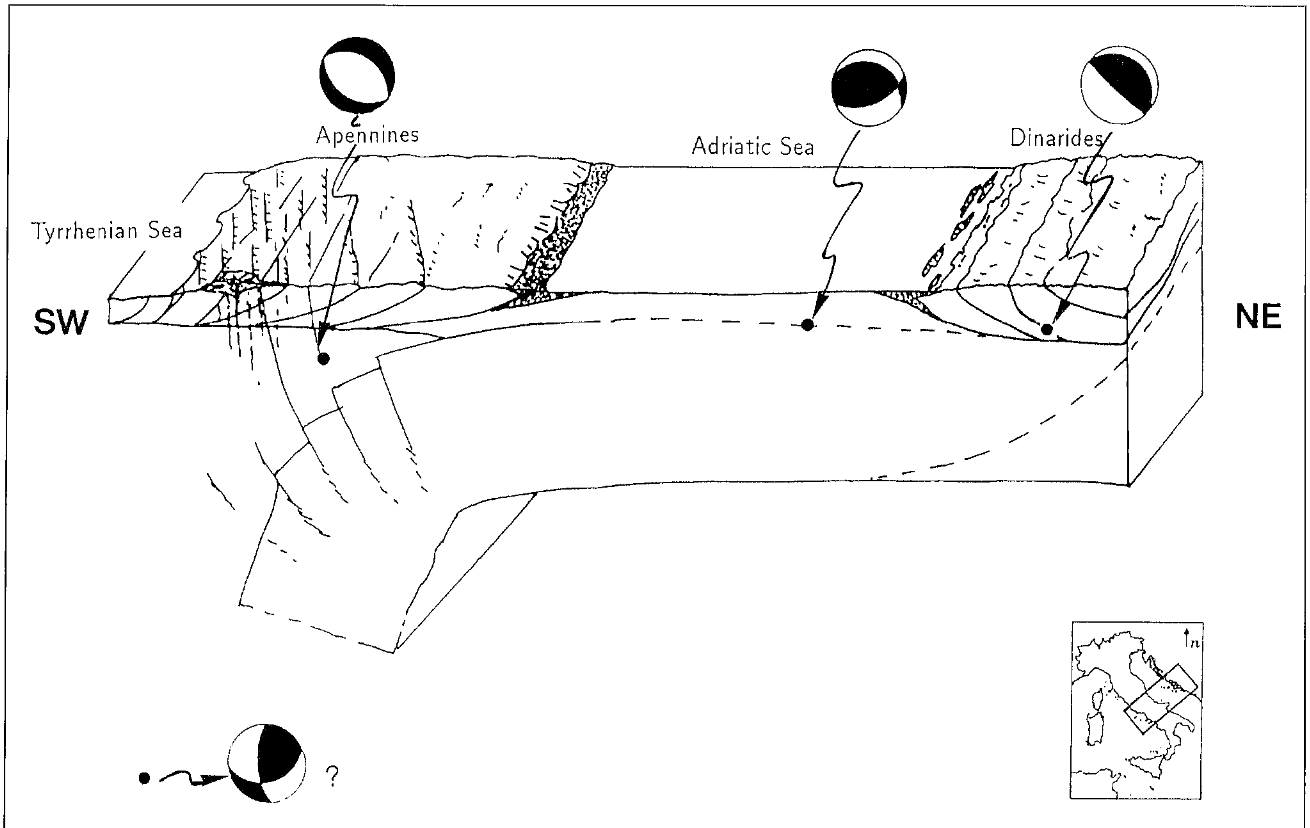

Fig. 3 - Block-diagram showing the interaction between the Adriatic lithosphere and the surrounding chains.

along the eastern margin the upper crust interaction is marked by low angle compressive tectonics. This essentially determined a crustal thickening with the preservation of a dynamical coupling between the basament and the sedimentary cover. This coupling is suggested by the substantial corrispondence between the deep deformations, inferred from the seismological data (see fig. 2), and the superficial ones (see fig. 1). The western margin, on the contrary, presents a clear decoupling between the deep extensional deformations and the intense shortening in the upper crust units.

In the Southern Apennines a sinking lithosphere collapse in the hinge zone could be the seismogenic mechanism, as inferred from the typical focal depths ranging 10 to 30 km. In the eastern margin earthquakes occur near the outermost thrust, in the western one they are located farther whithin the chain.

5. Conclusion

Our new data no longer justify a geodynamical model considering the Adriatic microplate as a unique, undeformed and internally aseismic block. At least two lithospheric domains are identified. The transition zone corresponds to an active deformation belt crossing the Adriatic Sea at the Gargano latitudes, characterized by regional strike-slip discontinuities.

In the southern block, the Dinaric-Hellenic chain marks a crustal thickening that represents its north-eastern margin. A sinking process below the Apennine chain identifies its south-western boundary. The aforementioned transitional zone represents the northern margin of this block, whereas its southern edge is questionable. Some Authors (D'Ingeo et al., 1980; Anderson & Jackson, 1987) tentatively recognized it with the Cephalonia active structure or with the escarpment toward the Ionian Sea.

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PRELIMINARY DATA ON THE ILLITE CRYSTALLINITY IN PELITIC ROCKS OF THE ALPINE UNITS OF SOUTHERN TUSCANY, ITALY

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Abstract

A systematic survey of the illite crystallinity was carried out on all the tectonic units outcropping in southern Tuscany and western Umbria. All these units show values of °Δ2Θ ranging from 0.66 to 0,20 (high grade diagenesis to epimetamorphic grade) and a complex regional distribution pattern without clear vertical and horizontal zonations as those observed north of the Arno river. Only in the Tuscan-Umbrian units, as a whole, in spite of the anomalous Larderello geothermal area, a gradual eastward decrease in the illite crystallinity is present. The relative low values of the illite crystallinity (in respect to ones of northern Tuscany) and the oblique contacts between the isopleth contour lines of the °Δ2Θ and the boundaries of the Tuscan-Umbrian units, could be mostly determined by a post-tectonic increase of the geothermal gradient related to lithospheric thinning, mantle swelling and shallow emplacement of acidic magmas in Southern Tuscany.

Key words: *Northern Apennines, Southern Tuscany, Alpine Units, Low Grade Metamorphism, Illite Crystallinity.*

1. Introduction

In the last twenty years many studies were undertaken on the diagenetic and very low grade metamorphic processes that affected most of the units of the Northern Apennines.

In particular studies on the coalification of the plant debris (Reutter et al., 1980) and on the illite crystallinity (I.C.) (Venturelli & Frey, 1977; Cerrina Feroni et al. 1980, 1983, 1985; Bonazzi et al. 1982, 1984) revealed that some tectonic units (e.g. the Tuscan Nappe), previously considered as "non metamorphic" show a weak metamorphism from the anchimetamorphic to the low epimetamorphic grade. Moreover the regional distribution patterns of the different parameters showed a decrease of the metamorphic grade either in a horizontal direction from west to east for each unit or in a vertical direction from the lowermost to the uppermost units of the tectonic pile.

All the studies quoted above have considered up to now only the Tuscany north of the Arno river. The poor number of data for the central-south Tuscany has suggested the opportunity to plan a systematic survey on the I.C. for the tectonic units south of the Arno river.

This paper is a preliminary report on the first data obtained by this research.

2. Geological Outline

The Northern Apennines are a complex tectonic pile of units, with a NE-E vergence, built up during the polyphased Alpine orogeny (27-10 Ma, Kligfield et al., 1986). The uppermost tectonic units, that belong to the Ligurian Oceanic Realm (western Tethys), were thrusted on the lowermost.

Tuscan-Umbrian units, derived from the African continental margin or Adria microplate (Sestini, 1970; Dallan Nardi & Nardi, 1972; Giannini et al., 1972; Giannini & Lazzarotto, 1975; Boccaletti et al., 1980, 1982; Carmignani et al., 1980; Burgassi et al., 1983).

In southern Tuscany (Fig. 1), three main Ligurian complexes were distinguished from top to bottom of the tectonic pile:

1) Ophiolitic Complex (Upper Jurassic-Lower Cretaceous) consisting of an ophiolitic basement overlain by the "Diaspri di Monte Alpe" Fm (jaspers), the "Calcari a Calpionelle" Fm (limestone) and the "Argille a Palombini" Fm (Shales with limestone intercalations).

2) Cretaceous-Paleogene Flysch Complex ("Flysch a Elmintoidi" Auct.).

These units scraped off from their oceanic substratum, are made up of calcareous-marly turbiditic sequences with sandstone and shale intercalations. Two main sequences are recognizable in the studied areas:

—The Cretaceous-Paleocene "Unità di Monteverdi Marittimo" (correspondig to the "Flysch di Monte Caio", north of the Arno river), in the north-western part of the investigated area (between Leghorn,Campiglia and Roccastrada).

—The Cretaceous-Eocene "Unità di Santa Fiora"

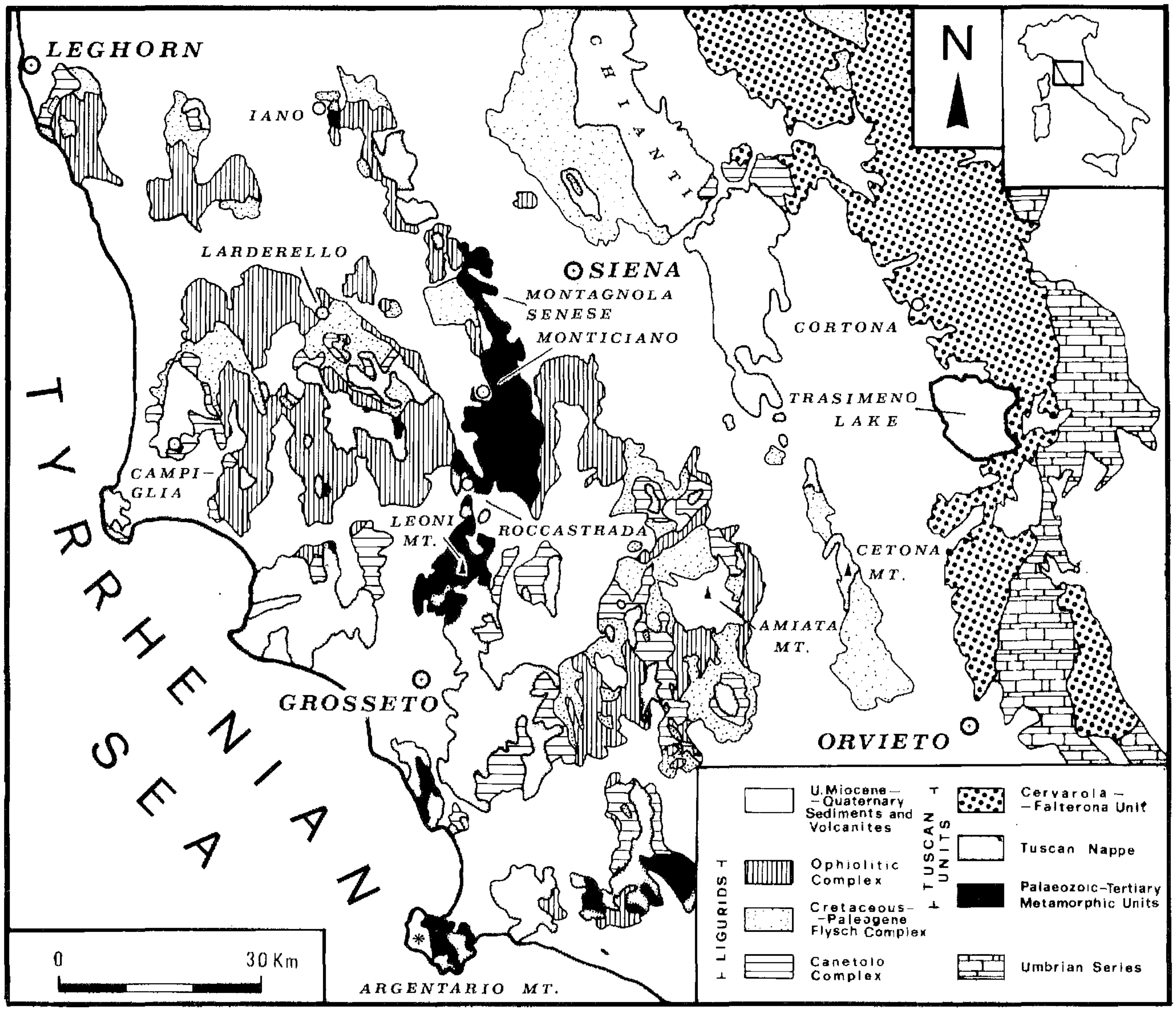

Fig. 1 - Geological sketch map of the southern Tuscany and western Umbria.(The asterisk in the westernmost Argentario Mt. area is referred to Mesozoic Australpine and Oceanic HP-LT methamorphic Units: Decandia et al., 1980).

(equivalent to the Calvana Supergroup, outcropping within the Chianti region, near Florence), in the Monte Amiata area.

The overlapping of the Unità di Monteverdi Marittimo on to the Unità di Santa Fiora / Calvana Supergroup was observed in the Amiata and Chianti regions (Bortolotti et al., 1970; Giannini et al., 1972).

3) Canetolo complex (Paleocene-Eocene) belonging to a transition zone between the Oceanic and Continental domains is made up of shales with calcareous and minor arenaceous intercalations ("Argille e Calcari" Fm).

The Tuscan domain includes the Paleozoic to Tertiary metamorphic units (e.g. Monticiano-Roccastrada Unit Auct.), the Tuscan Nappe (Upper Trias-Lower Miocene) and the Cervarola-Falterona Unit (Middle Eocene-Middle Miocene). The Tuscan Nappe tectonically overlies the metamorphic units in western Tuscany and eastward overrodes the Cervarola-Falterona Unit; the latter was thrusted onto the Umbrian series (Upper Trias-Tortonian) in a more external position.

The type sequence of the Tuscan-Umbrian Domain can be briefly summarized as follows from top to bottom:

— Siliciclastic turbiditic sediments from west to east: a) "Macigno" sandstone (Middle/Upper Oligocene-Lower Miocene) and "Arenarie del M. Cervarola-M. Falterona" (Lower to Middle Miocene) in the Tuscan Domain; b) "Marnoso-arenacea" (Middle to Upper Miocene) in the Umbrian Domain.

— Upper Cretaceous to Lower Miocene pelagic shaly-marly sediments with limestone and carbonate levels ("Scisti policromi" Group) in the Tuscan units, and calcareous-marly sediments "Scaglia Umbra", and shaly-marly sediments "Scaglia Cinerea", in the Umbrian series.

— pelagic carbonate-marly and siliceous sediments (Middle Lias to Upper Cretaceous);

— shallow marine carbonate sediments (Upper Trias-Lower Lias);

— continental to coastal siliciclastic sediments ("Verrucano" Auct.) Middle-Upper Trias);metamorphic Paleozoic to Precambrian (?) basement.

3. Sampling and Analytical Techniques

Among the different lithotypes of the quoted units only the pelitic shaly or shaly-marly rocks were sampled.

For each of the choosen outcrops 2 to 5 samples were collected with different stratigraphic position. Care was also taken to exclude the outcrops with clear evidence of weathering, hydrothermal alteration or faulting.

The geological map used for sampling (fig. 1) was obtained by integrating the "Carta Geologica d'Italia" 1:100.000 with several more detailed geological maps (see references of the geological maps).

In the different complexes or units the following formations (or Groups) were investigated:

1)"Argille a Palombini" Fm (Lower Cretaceous) from the Ophiolitic Complex (35 samples).

2)"Monteverdi Marittimo" Fm (Upper Cretaceous-Paleocene) (25 samples) and "Santa Fiora" Fm (Upper Cretaceous) (10 samples) from the Cretaceous-Paleogene Flysch Complex.

3)"Argille e calcari" Fm (Paleocene-Eocene) from the Canetolo Complex 35 samples).

4)"Argilliti di Brolio" Fm (Upper Cretaceous-Oligocene) from the Tuscan Nappe (60 samples) and "Scisti Varicolori" Fm (Upper Cretaceous?/Middle Eocene-Middle Miocene) from the Cervarola- Falterona Unit (15 samples), both belonging to the Scisti Policromi Group.

5)"Scaglia Cinerea" Fm (Upper Eocene-Oligocene) from the Umbrian series (10 samples).

6)"Pietralata" Fm (Cretaceous-Eocene) in the Tuscan metamorphic units. These rocks correspond to the "Scisti Sericitici varicolori" Fm in the Alpi Apuane and Monti Pisani areas and are the metamorphic equivalent of the "Scisti Policromi" Group (5 samples).

7)"Verrucano" Group (Middle-Upper Trias) from the Tuscan metamorphic units (Iano to Monte Argentario areas).

3.1 *Analytical Methods*

Samples were dried and crushed using a disc mill for about 10 sec. and sieved through a 0,5 mm sieve. The powder was then suspended in deionized water. The acqueous suspension containing the < 2 μm grain-size fraction was pipetted onto glass slides and dried at 80° C, in order to obtain highly oriented preparates. Glycolation of samples was performed at least for 12 h at room temperature.

I.C. was determined by means of a Rigaku Geiger flex difractometer of the Cagliari University on the <2.0 air dried μm clay fractions. Operating conditions were: Cuka Ni filtered radiation; divergence slit = 0.5°; receiving slit = 0.3°; scatter slit 1°, goniometer rotation speed = 0.5°/min. The crystallinity values were calibrated using standards kindly provided by Prof. H.C. Hunziker, Lausanne University, and calibrated against Kübler's standards. The error in a single measurement of I.C. was estimated to be 0.02°.

The diagenetic/anchimetamorphic (°Δ2Θ = 0.42°) and anchimetamorphic/epimetamorphic grade (°Δ2Θ = 0.25°) boundaries of Kübler (1984) were assumed.

In previous works on the Northern Apennines I.C. was defined by both the Kübler index of 1968 (half width in mm of 10 Å peak) and the Kübler index of 1975 (half width of 10 Å peak in units of °Δ2Θ). All these data were obtained using the Philips PW 1793 difractometer of the Pisa University. In order to minimize the interlaboratory bias, a comparison was made between data obtained on the same 15 samples using both the Rigaku diffractometer of the Cagliari University and the Philips diffractometer of the Pisa University. No significant difference between the two sets of data was observed for the range: 0.20-0.60 of °Δ2Θ.

4. Illite Crystallinity Maps

The °Δ2Θ indices for each complex are reported in the geological maps of figure 2 to 6.

Each index is the average of the different samples collected in the same oucrop.

The following values were obtained for the different complexes:

1)Ophiolitic complex (Argille a Palombini) (Fig. 2): on a regional scale the values fall in the range 0.43-0.67 max 0.71 (S= 0.09) and 0.90 in hydrothermalized areas near Massa Marittima-Serrabotini and CampigliaMarittima areas.A significant decrease of the °Δ2Θ values0.34-0.46)was observed in a restricted area roughly coincident with the Larderello-Travale geothermal area.

2) Cretaceous-Paleogene Flysch complex (Fig. 3): the Monteverdi Marittimo Fm (Larderello-Travale region) show values in the range 0.44-0.53 of °Δ2Θ with sporadic values of 0.69 (s =0.06) for °Δ2Θ in areas with hydrothermal alteration in the Larderello region. Lower values (0.37-0.42 of °Δ2Θ) were obtained for the Santa Fiora Fm in the Monte Amiata and Monte Cetona areas. Values ranging from 0.57 (s 0.07) to 0.68 (s =0.09) were found in altered outcrops of Monte Amiata.

3) Canetolo Complex (Fig. 4): the °Δ2Θ values are in the range 0.28-0.57values up to 0.81-0.84 were obtained near the granitic outcrop ofCampiglia Marittima.

4) Tuscan Nappe (Fig. 5): the following three main groups of values,concentrated in roughly parallel N-S trending bands, were identified from west to east: a)

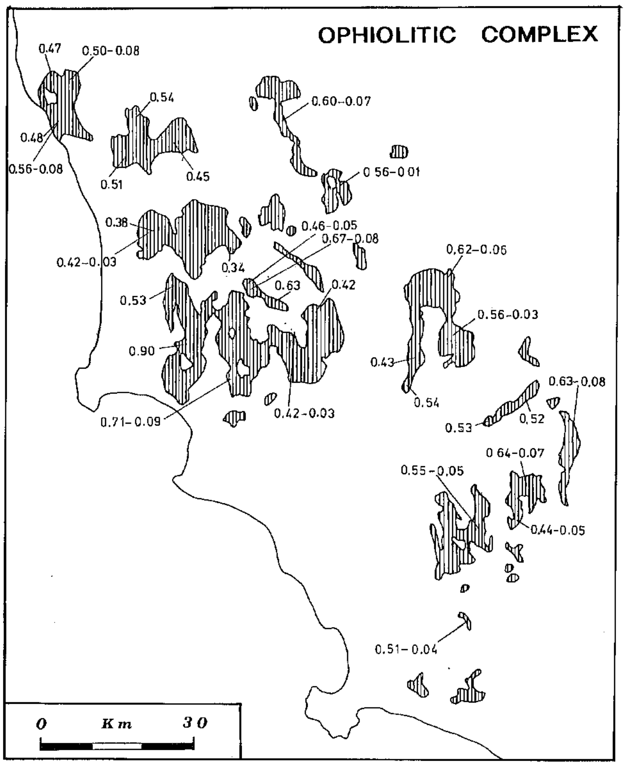

Fig. 2 - Distribution of the I.C. values in the Ophiolitic Complex. Standard deviation (S) is reported when more than two values are available.

Casciana-Campiglia-Larderello region: 0.36-0.59 of °Δ2Θ; b) Monti del Chianti and the western side of the Monte Amiata area: 0.29-0.42 of °Δ2Θ with the only exception of

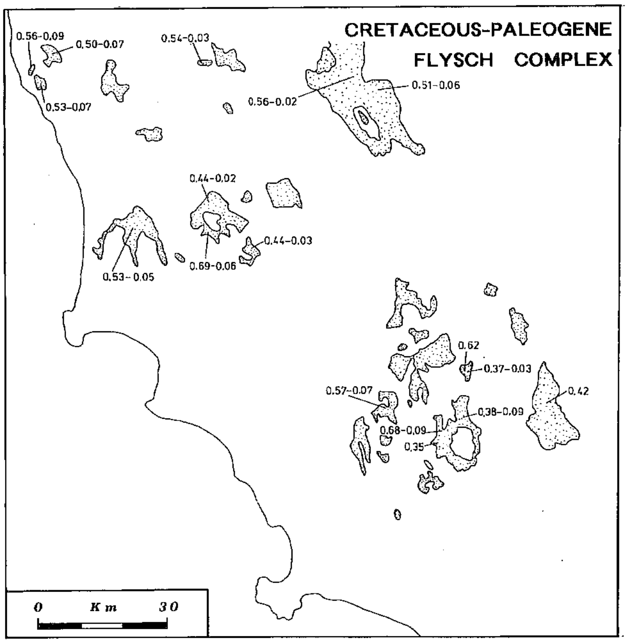

Fig. 3 - Distribution of the I.C. values in the Createaceous- Paleogene Flysch Complex. Standard deviation as for fig. 2.

0.49; c) the eastern side of the Monte Amiata area and Monte Cetona area: 0.47-0.74 of °Δ2Θ; the highest values indicate hydrothermally altered outcrops.

5) Cervarola-Falterona Unit (Fig. 5): in the Scisti Varicolori three average values identified a range of 0.47-0.61 for °Δ2Θ, while two average values are strikingly as low as 0.32 and 0.37. The latter were obtained on samples very close to overthrust surfaces.

6) Umbrian series (Fig. 5): in the Scaglia Cinerea Fm values of 0.45-0.60 of °Δ2Θ were found.

7)Tuscan metamorphic Units (Fig. 6): the Pietralata Fm, in the Montagnola Senese area, yielded the lowermost values among all the data of this paper (0.17 of °Δ2Θ), but in the Fontalcinaldo mining area (west of the Monticiano-Roccastrada ridge) the same rocks show a range of 0.28-0.39 of °Δ2Θ almost coincident with those of the Verrucano phyllites (0.36-0.40 of °Δ2Θ) found by Franceschelli et al (1990).

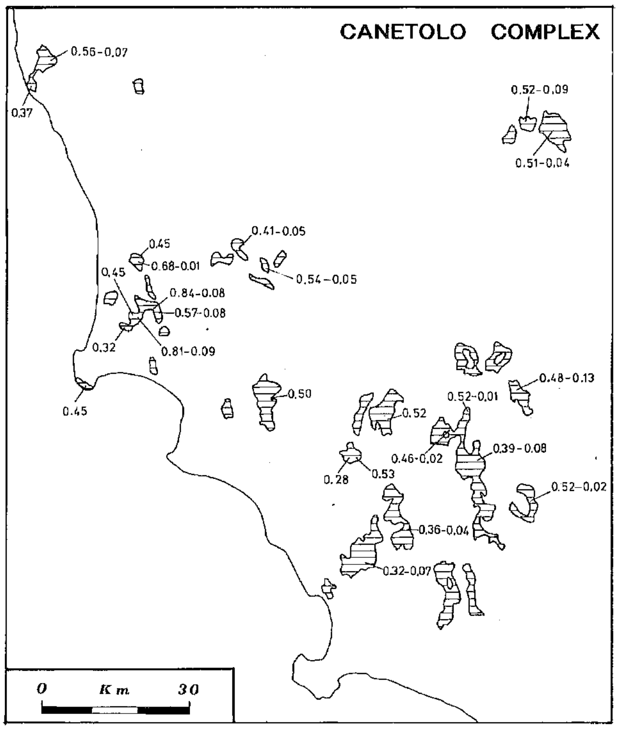

Fig. 4 - Distribution of the I.C. values in the Canetolo Complex. Standard deviation as for fig. 2.

5. Discussion

A preliminary comparison between the different sets of data in the present paper and those reported in literature for the same tectonic units, north of the Arno river, gives the following main results:

1) The illite crystallinity is strongly enhanced south of the Arno river. For example the Canetolo Complex shows ranges of 0.42-0.85 of °Δ2Θ (Cerrina Feroni et al., 1985)

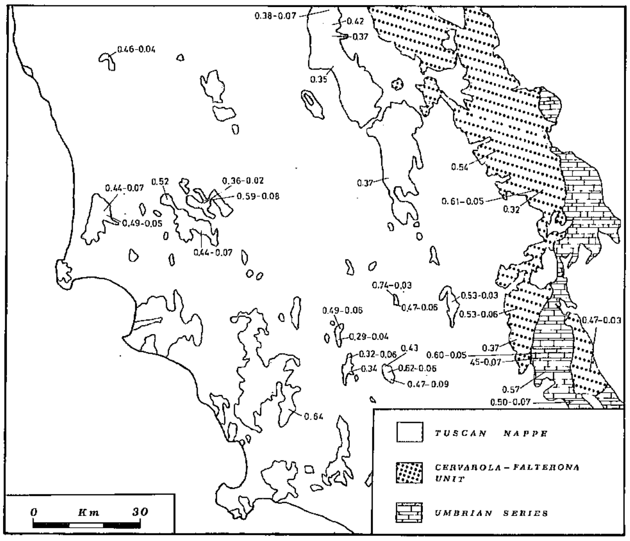

Fig. 5 - Distribution of the I.C. values in the Tuscan and Umbrian Units. Standard deviation as for fig. 2.

and of 0.36-0.53 of °Δ2Θ (this paper) north and south of the Arno river respectively.

The °Δ2Θ values of each complex, in south Tuscany, strongly decrease and show a more restricted range in comparison with those of north Tuscany. All the tectonic units, in south Tuscany, show °Δ2Θ values in the range 0.66-0,13 (highest grade diagenesis to low epimetamorphic grade).

2) In north Tuscany a general decrease of metamorphic grade was observed from the lowermost Units (low epimetamorphic grade) to the uppermost ones (diagenesis): moreover within each complex or unit (e.g. Tuscan Nappe in Cerrina Feroni et al., 1983) a clear regional zonation was revealed by an increase of the °Δ2Θ values and of the degree of coalification from west to east (Cerrina Feroni et al. 1983; Reutter at al. 1980).

In south Tuscany remarkable overlapping of the I.C. ranges were obtained for the different complexes or units produces a less clear vertical zonation of I.C. within the tectonic pile. The range of °Δ2Θ values in south Tuscany, taken as a whole, is by far narrower than the corresponding range observed in north Tuscany.

By analogy each complex or unit, south of the Arno river, does not generally show a clear horizontal zonation as that observed north of the Arno river. Only the Tuscan and Umbrian units, as a whole, in spite of the anomalous high values of the Larderello region, show a gradual eastward I.C. decrease.

Another remarkable feature is the lack of correspondence between the trend of the different Tuscan-Umbrian units (Tuscan Nappe, Cervarola-Falterona Unit and Umbrian series) and the regional distribution pattern of I.C. indices found for the same units with oblique contacts between the N-S trending isopleth contour lines of the °Δ2Θ values and the NW-SE trending boundaries among the different units.

The remarkable I.C. increase and hence of metamorphic grade observed from north to south Tuscany may be ascribed to a corresponding significant increase of the geothermal gradient generated by: A) lithospheric thinning (Mongelli et al., 1991); B) swelling of a soft mantle (Moho depth of 20-22 km below the Grosseto area, Boccaletti et al., 1985; Vp=7.9 - 7.4 km/sec. in the mantle, Letz et al., 1978; Ginzburg et al., 1986); C) widespread occurrence of still hot and slowly cooling shallow batholithic acidic intrusive bodies. All these phenomena produced also rise and thickening of the isotherms, that homogenized the metamorphism in the uppermost levels of the crust, flattening and reducing the variability of the PT conditions among the different units.

The oblique contacts between the isopleth contour lines of the Δ2Θ values for the Tuscan Nappe and the boundaries of the different units emphasize that the present day regional distribution pattern of the I.C. indices, partly related to the emplacement of the different tectonic units and to burial depth, is mostly determined by the late orogenic phenomena previously described, such as lithospheric thinning, mantle swelling and shallow emplacement of acidic magmas. By analogy a lack of systematic correspondence between the I.C. isopleths and

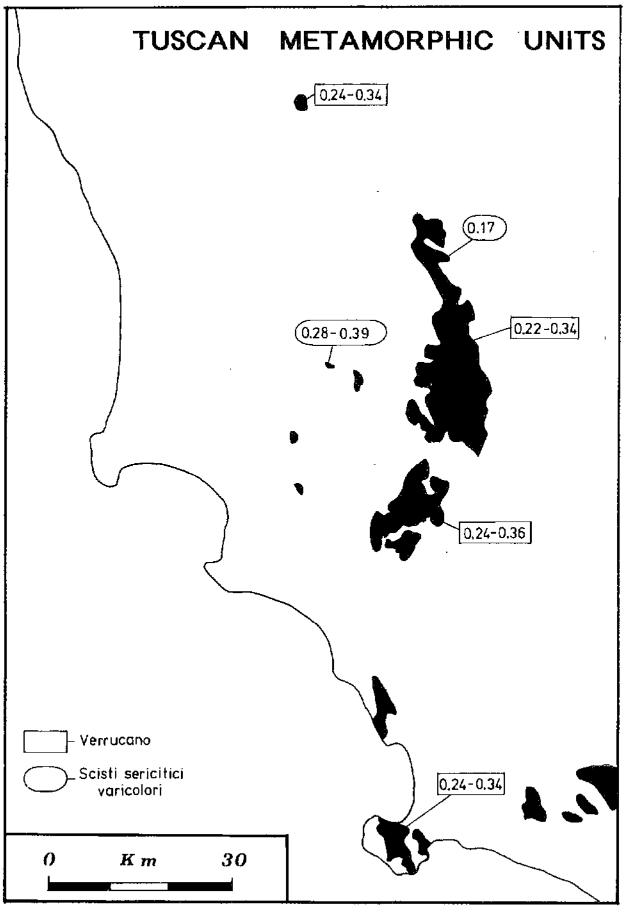

Fig. 6 - Range of the I.C. values in the Alpine Tuscan Metamorphic sequences (after Franceschelli et al., 1990).

the contacts among the different lithostratigraphic units was observed in south-western Gaspe', Quebec, Canada by Duba & Williams-Jones (1983). The same authors attributed the "larger zone of anomalously low illite crystallinity indices" to reheating produced by "buried igneous masses of batholithic proportions".

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ALPINE METAMORPHIC OVERPRINT IN THE CRYSTALLINE BASEMENT OF THE ASPROMONTE UNIT (CALABRIAN - PELORITAN ARC - SOUTHERN ITALY)

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Abstract

A tectono-metamorphic overprint of Alpine age affected the Aspromonte Unit (Southern Sector of the Calabrian Peloritan Arc), re-equilibrating a Variscan crystalline basement made up of amphibolite-facies metamorphic rocks intruded by syn- and post-tectonic plutonites.

The overprint, which is strong in the eastern portion of the Aspromonte Massif and weaker in the remaining zone, developed along centimeter- to kilometer-thick shear bands.

In the strongly re-equilibrated area both extensively and partly recrystallized Variscan rocks can be distinguished. In the extensively re-equilibrated rocks a pervasive deformation produced a roughly horizontal foliation accompanied by a reduction in grain-size and by an almost complete recrystallization of the Variscan mineral assemblages. In the partly re-equilibrated rocks, only a partial metamorphic recrystallization occurred. The neoblastic minerals may be ascribed to at least two different events or stages of a single metamorphic event.

In the weakly re-equilibrated area a cataclastic to mylonitic deformation led to the development of few very fine grain-sized neoblastic minerals ascribed to both recognized events.

Both in the partly and extensively recrystallized rocks, the coexistence of relict and neoblastic minerals is confirmed by the sharply bimodal chemical composition of the most important mineral phases.

The Variscan minerals point to low-P medium- to high-grade amphibolite-facies conditions. The Alpine minerals of the first event (Fe- and/or Ca-rich almandine, kyanite, phengite and paragonite, albite, Fe-Mg-chloritoid, zoisite/clinozoisite, hastingsite to edenite and K-rich pargasite, Mg-Fe-ripidolite) suggest greenschist-facies conditions of a Barrovian type metamorphism. The Alpine minerals of the second event (Fe-phlogopite-annite, oligoclase) suggest higher temperatures, and lower pressures than those of the first one.

Variscan and Alpine parageneses have been projected in the A-F-M and Na_2O-AF_2O_3-CaO-FMO diagrams.

Geothermobarometric calculations gave 5 to 8 kbar and 500 ± 20° C for the first event; higher T and slightly lower P were inferred for the second one.

In the weakly re-equilibrated area, the same two Alpine events developed under lower P-T conditions.

Key words: *Calabrian-Peloritan Arc, Italy; Aspromonte Unit; Alpine metamorphic overprint; Mineral chemistry; Geothermobarometers.*

1. Introduction

This paper describes the tectono-metamorphic evolution of the polyphase Alpine overprint discovered in the Aspromonte Unit (Bonardi et al., 1984a). The P-T conditions of the Alpine events have been deduced from the chemical compositions of the newly-formed minerals from partly and extensively re-equilibrated rocks.

2. Previous Investigations

The Aspromonte Unit, extending from the Aspromonte Massif (Calabria) to the Peloritani Mts. (Sicily), outcrops in the Southern Sector of the Calabrian-Peloritan Arc (Fig. 1).

Several petrological studies, mostly of local interest, have been devoted to this unit (Messina et al., 1990 with ref.; Bonardi et al., 1991 with ref.).

In the first extensive regional paper on the Aspromonte Massif, Bonardi et al. (1979) recognized an Alpine unit, which consists of a Variscan basement made up of augen gneiss, micaschist, paragneiss, minor amphibolite and marble, intruded by the Villa San Giovanni, Punta d'Atò and Delianuova late-Variscan peraluminous granitoids. The strong cataclastic overprint, which occurs all over the crystalline basement, was attributed to Alpine tectonic

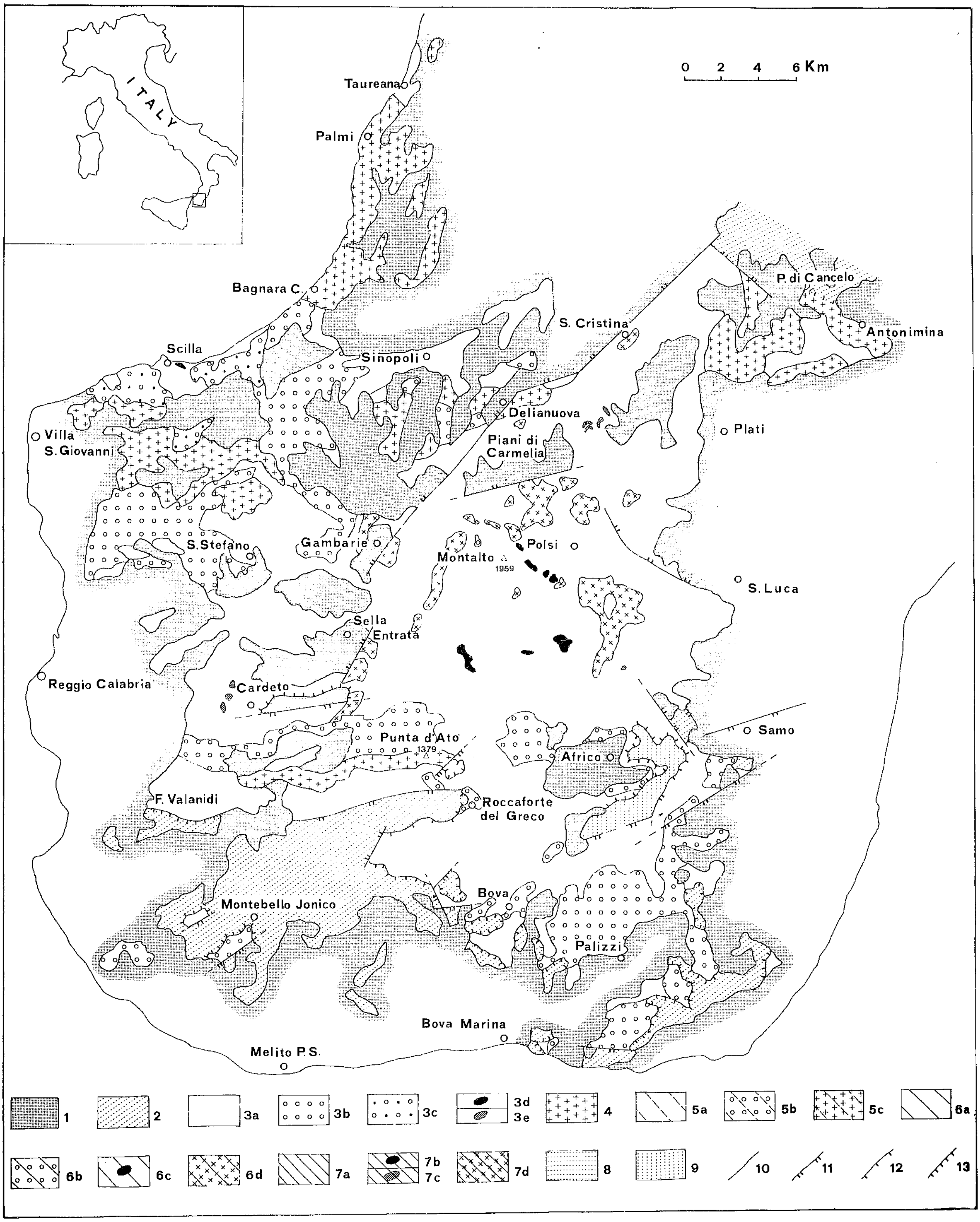


Fig. 1 - Geological sketch map of the Aspromonte Unit. Legend - 1 = Mesozoic to Cenozoic sedimentary cover. 2 = Stilo Unit. 3 to 7 = **ASPROMONTE UNIT**: 3 = **Variscan metamorphic rocks**: 3a = qtz + olig/ande + bi + sil + gt ± ms ± Kf ± cord gneiss, micaschist and fels, with interlayered amphibolite, meta-ultramafite, meta-pyroxenite and marble; 3b = Kf + qtz + olig/ande + bi ± ms ± sil augen gneiss; 3c = qtz + olig + Kf + bi ± ms metagranites associated with augen gneiss; 3d = hor + labr/byt ± gt ± qtz ± Kf ± cumm ± bi amphibolite s.l.; 3e = cc ± di ± hor ± cumm ± pl ± phl ± ms ± qtz marble s.l. (continued in the next page) Fig. 1 - (continued from the previous page) 4 = **Late-Variscan plutonic rocks**: two mica + Al-silicate-bearing leucotonalite to monzogranite; 5 = **Variscan rocks weakly Alpine overprinted**: 5a = cataclastic to mylonitic rocks as 3a with newly-formed ser + gr-bl amph + gr-bl bi

events. The above unit, at first informally called the *Intermediate Unit* because of its geometrical position, was later renamed the ***Aspromonte Unit*** by Bonardi et al. (1982) as it corresponds to the Aspromonte Nappe defined by Ogniben (1973)

Inside the Aspromonte Unit, Lorenzoni et al. (1980) and Lorenzoni & Zanettin Lorenzoni (1982) distinguished three Variscan tectonic units which they considered to be the uppermost portion of a "rooted Hercynian Chain".

Pezzino & Puglisi (1980), and Ioppolo et al. (1982) distinguished two juxtaposed, *upper* and *lower*, tectonic units in the northeastern area of the Aspromonte Unit. The upper unit (Ioppolo et al., 1983a; Ioppolo & Puglisi, 1986-87a) consists of medium to high grade metamorphic rocks characterized by T = 650-550° C and P = 3 kbar in a postkinematic event. The lower unit (Ioppolo et al., 1983b; Ioppolo & Puglisi, 1986-87b) is made up of low to medium grade metamorphic rocks (with prograde albite-epidote amphibolite-facies, garnet zone), characterized by P = 5-8 kbar and T = 500 ± 50°C in the Polsi area and by T = 470 ± 60°C and P = 7 ± 1 kbar in the Delianuova area. The contact between the two units is marked by a mylonitic band of unknown age, characterized by the post-tectonic development of chloritoid, chlorite and staurolite (Ioppolo et al., 1983a).

Bonardi et al. (1984a) reaffirmed the existence of a single medium-high grade metamorphic thrust sheet in the Aspromonte Unit, on the grounds of an extensive petrographic study. They also described the presence of a greenschist-facies metamorphic re-equilibration, probably Alpine in age, which overprints the whole lower unit and part of Pezzino & Puglisi's (1980) upper unit. This overprint, recognized over a large area around Montalto, deformed both magmatic and metamorphic Variscan rocks and developed during at least two different events: the first characterized by P = 0.5 ± 0.1 GPa and T = 500 ± 30° C, the second by P a bit lower and T higher than the first event.

Preliminary Rb/Sr data (Bonardi et al., 1987), on separated micas from overprinted rocks of the Aspromonte Unit, suggest that the Variscan metamorphism peaked at or before 330 Ma, whereas the second Alpine event developed at about 25-30 Ma.

Pezzino et al. (1990) re-emphasized the existence of both lower and upper units, however, now, the authors indicate that the MP prograde metamorphism of the lower unit is compatible with the conditions of an Alpine event. Along the contact, both units have been affected by a high greenschist- to low amphibolite-facies blastic event, after their emplacement. The overthrusting occurred after the MP metamorphic event that affected the lower unit.

A structural study (Platt & Compagnoni, 1990) of the re-equilibrated rocks in the Aspromonte Unit indicates that a moderate ductile deformation preceded, and in some areas accompanied, the relatively high pressure earlier Alpine stage. An intense ductile deformation, responsible for a mylonitic foliation, a strong N-S stretching lineation and several shear bands, accompanied the lower pressure second Alpine stage. The earlier Alpine deformation may have been related to a crustal thickening stage, the second to the uplift and exhumation of the rocks during late Oligocene or early Miocene times.

In the Peloritani section of the Aspromonte Unit, Messina et al. (1990) have recently discovered a cataclastic to mylonitic deformation similar to the Alpine overprint described in the Aspromonte Massif by Bonardi et al. (1984a; 1987). West of Messina, over a zone of about 3 km long and 2 km wide (the Badiazza Valley), the Variscan rocks have been strongly overprinted by a greenschist-facies metamorphism, which may be attributed to at least two different events or two stages of a single metamorphic event. The first event characterized by moderate HP minerals suggesting T = 480-550° C and P = 0.4-0.7 GPa; the second event characterized by lower P minerals suggesting T > 550° C and P < 0.4 GPa. Around the above zone, a weaker overprint consists of a cataclastic deformation accompanied by the development of very fine-grained minerals ascribed to both Alpine events, suggesting lower P-T conditions with respect to the strongly overprinted area.

3. Geological Outlines

The Aspromonte Unit in Calabria occupies much of the massif of the same name (Fig. 1). It extends from Taureana-Antonimina in the north, to Bova Marina in the south (Bonardi et al., 1979). The Passo di Cancelo to Antonimina line marks the northeastern tectonic contact between the Aspromonte Unit and that part of the overlying Stilo Unit which outcrops in the Serre Massif and consists of low to medium grade phyllites with granite intrusives (Bonardi et al., 1984b). The Fiumara Valanidi-Roccaforte del Greco fault, marks the southwestern tectonic contact with that part of the overlying Stilo Unit which outcrops in

± gtII;. 5b = cataclastic to mylonitic rocks as 3b with newly-formed ser + gr-bl bi; 5c = cataclastic to mylonitic Variscan syntectonic plutonites: biotite-amphibole-bearing tonalite to granodiorite with newly-formed ser + gr-bl amph + gr-bl bi ± gtII. 6 and 7 = **Variscan rocks strongly Alpine overprinted.** 6 = **partly re- equilibrated rocks**: 6a = sil ± gt micaschist, gneiss and fels partly re-equilibrated to ky + gtII + wmII ± clt ± chl + biII micaschist; 6b = augen gneiss partly re-equilibrated to gtII + wmII + biII + chl orthogneiss; 6c = amphibolite partly re-equilibrated to amphII + gtII + wmII + chl + biII amphibolite; 6d = late-Variscan peraluminous plutonic rocks partly re- equilibrated to ky + gtII + wmII + amph + biII + chl orthogneiss. 7 = **extensively re-equilibrated rocks**: 7a = gt ± staur ± and gneissic micaschist extensively re-equilibrated to two mica + gtII + ab/olig + amph + Kf gneissic micaschist; 7b = meta-hornblendite extensively re-equilibrated to amphII ± gtII ± chl ± biII meta-hornblendite; 7c = marble extensively re-equilibrated to cc + wmII + amphII + chl + biII marble; 7d = late-Variscan peraluminous plutonic rocks extensively re-equilibrated to gtII + wmII + biII + chl leuco-orthogneiss. 8 = Mandanici Unit. 9 = Africo Unit. 10 = Stratigraphic boundary. 11 = Direct fault. 12 = Inverse fault. 13 = Overthrust. Mineral abbreviation list in appendix the Aspromonte Massif and consists of low to medium grade metamorphic rocks (Crisci et al., 1983). The Aspromonte Unit is also exposed in the Stilo Unit tectonic windows of Montebello Ionico. The two tectonic windows of Cardeto and Africo Vecchio - Casalnuovo occur inside the Aspromonte Unit. In these windows low-grade metamorphic rocks belonging to the underlying Mandanici (Bonardi et al., 1980) and Africo Units (Bonardi et al., 1979) outcrop respectively.

The Aspromonte Unit, which is devoid of Meso Cenozoic sedimentary cover, consists of a Variscan crystalline basement made up of heterogeneous low to upper amphibolite-facies metamorphic rocks intruded by syn- to post-tectonic plutonic rocks.

The metamorphic complex outcrops discontinuously owing to the presence of a Tertiary to Quaternary sedimentary cover, of numerous faults and of plutonic bodies. The metamorphic rocks consist of banded paragneiss and micaschist, with minor interlayered amphibolite, marble and Ca-silicate fels. Several bodies of augen gneiss and subordinate metagranitoids are also present.

Meter to kilometer-thick layers of **paragneiss** and intercalated **micaschist** are the most widespread rocks. They show a wide range of fabric, grain-size and mineralogy.

Plurikilometer-sized **augen gneiss** bodies outcrop (going from north to south) in the Bagnara-Gambarie, S. Stefano, Montalto, Punta d'Atò, Roccaforte del Greco, Bova, Palizzi and Montebello Ionico areas. Numerous large-scale isoclinal folds form an interlayering of paraderived metamorphic rocks and these augen gneiss bodies, which are probably derived from pre-Variscan porphyritic granitic intrusions into an already metamorphic basement. This hypothesis is supported by the presence of several concordant decimeter to hectometer-thick **leucocratic gneisses**, abundant inside the augen gneiss and widespread in the other metamorphic rocks, which show pre-Variscan features (myrmekitic texture, zoned plagioclase) typical of primary aplo-pegmatitic or microgranitic dykes.

Metagranitic bodies, heterogeneous in grain-size and composition, are in some places spatially associated with augen gneiss.

Amphibolite and **meta-hornblendite** are commonly found as lenticular bodies of up to a hectometer in size; rare **meta-ultramafite** is also present. Meter-thick lenses or decimeter-thick layers of **marble** and **Ca-silicate fels** are found in some areas.

A small **migmatitic complex** is exposed in the western area of the unit, at Scilla.

The Variscan intrusive bodies are: i) Syntectonic inequigranular medium to very coarse-grained biotite ± amphibole-bearing tonalites to granodiorites outcropping in the Palmi-Bagnara area; ii) Post-tectonic, varying in grain-size, two mica Al-silicate-bearing leucotonalites to leucomonzogranites, which outcrop in the Villa S. Giovanni, Antonimina, Delianuova, S. Cristina, Gambarie, Montalto, Polsi, Sella Entrata and Punta d'Atò areas. These plutonic stocks consist of many distinct intrusions. The whole crystalline basement is crosscut by a network of aplitic, pegmatitic, microgranitic and felsitic dykes, which are the latest intrusions. The intrusive contact between plutonites and metamorphic rocks is characterized by the widespread occurrence of granitic and aplitic dykes, which form 50% of the rock, and constitute layered migmatites.

The Alpine overprint developed along centimeter- to kilometer-thick shear bands. It is strong in the eastern portion of the Aspromonte Massif, and becomes weaker in the outer zone. In the remaining portion of the unit the Alpine overprint exhibits a widespread cataclastic to mylonitic deformation.

In the **stronger re-equilibrated area**, which is exposed from north of S. Cristina to Sella Entrata (Fig. 1), both partly recrystallized rocks and extensively recrystallized rock types have been distinguished. *In the extensively re-equilibrated rocks,* which outcrop in the S. Cristina - Delianuova - Piani di Carmelia - Polsi - S. Luca area, a ductile pervasive deformation produced a roughly horizontal foliation, accompanied by a reduction in grain-size and an almost complete recrystallization of the Variscan amphibolite-facies rocks. These appear to have been converted to phyllonites. The Variscan granitic bodies and aplo-pegmatitic dykes have been transposed and have become leucocratic orthogneisses. Inside these strongly re-equilibrated rocks, apparently preserved coarse-grained meta-hornblendite and amphibolite boudins are widespread. *In the partly recrystallized rocks,* a non pervasive deformation was accompanied by a partial metamorphic recrystallization.

In the **weaker re-equilibrated area**, which is exposed between Taureana and Sinopoli to the north, and between Sella Entrata and Bova to the south (Fig. 1), a mylonitic deformation, responsible for a slight recrystallization, has occurred.

In the western area, from Scilla to Punta d'Atò , the Alpine deformation produced cataclastic effects, but a stronger recrystallization is found in localized cm- to m-thick shear zones.

It is very difficult to recognize a detailed Variscan metamorphic zoning in the above sequence, owing to the widespread Alpine overprint and to the presence of numerous faults. Field findings, confirmed by petrographic and geochemical data, which also take into account the relict assemblages in the Alpine re-equilibrated rocks, indicate that: i) relatively deep crust metamorphic rocks (biotite + andesine to labradorite + garnet + prismatic sillimanite + K-feldspar + cordierite upper amphibolite-facies metapelites) outcrop in the northwestern area, around Taureana; ii) relatively high crust metamorphic rocks (biotite + oligoclase + muscovite + staurolite + garnet ± K-feldspar lower amphibolite-facies metapelites) outcrop in the eastern area between S. Cristina and Polsi; iii) **in the remaining portion**

of the Aspromonte Unit medium amphibolite-facies metamorphic rocks (biotite + oligoclase to andesine + fibrolitic and/or prismatic sillimanite + muscovite + garnet ± K-feldspar ± cordierite metapelites) and a medium to high grade migmatitic complex (oligoclase to labradorite + biotite + K-feldspar + muscovite metapelites) occur.

4. Petrography

4.1 *Variscan rocks*

Paragneiss, micaschist and subordinate **fels** show a pervasive deformation phase originating a main foliation (Dvn), locally crenulated or cut by a second deformation phase. These rocks exhibit a wide range of grain-size (fine to very coarse; equi- to inequigranular), structure (massive to foliated), and texture (xenoblastic to polygonal and lepidoblastic to diablastic or porphyroblastic).

Quartz + plagioclase + biotite are the common mineral phases to which sillimanite ± garnet, and localized K-feldspar ± muscovite ± cordierite ± staurolite ± andalusite are associated. Their tectono-metamorphic history is characterized by a synkinematic (dimensional preferred orientation due to dynamic recrystallization) to postkinematic (blastic texture) crystallization of quartz + plagioclase + biotite ± sillimanite ± muscovite. Staurolite and cordierite are present in some areas. The former is found as: i) idioblastic crystals; ii) grains included in garnet; iii) corroded relicts inside andalusite; the latter is mostly found as crystals, sometimes intergrown with fibrolitic sillimanite. The postkinematic centimeter-thick porphyroblastic andalusite, which includes all the above-mentioned minerals, grew at the expense of sillimanite or staurolite. Large muscovite flakes after sillimanite can be considered as post-kinematic. Apatite, rutile, zircon, tourmaline, magnetite, ilmenite, graphite and rare titanite are the accessory phases.

Augen gneiss is very coarse grained and characterized by up to 4 cm porphyroclastic K-feldspar surrounded by lepidoblastic to diablastic biotite in a granoblastic to polygonal matrix, made up of quartz + oligoclase + K-feldspar + subordinate muscovite + rare fibrolitic sillimanite. Magmatic zoned plagioclase, biotite and quartz relicts are included in the porphyroclasts of K-feldspar. Apatite, zircon and opaque ores are the most common accessories.

Metagranitic bodies and **dykes** show the same mineralogical assemblages as the augen gneiss but a very heterogeneous grain-size and bi/pl and bi/Kf ratios. They range in composition from tonalite to monzogranite.

Many different types of **amphibolite** have been recognized. They show massive to banded structure, medium to coarse grain-size and a prevalently blastic fabric. They consist of hornblende + andesine to bytownite ± diopside ± garnet ± biotite ± quartz. Titanite, magnetite, ilmenite and apatite are the accessory minerals. The most widespread are: i) granoblastic *garnet-bearing amphibolite* made up of green-brown hornblende + garnet + labradorite ± cummingtonite ± rare K-feldspar and quartz; ii) *poikiloblastic amphibolite* consisting of andesine to labradorite + green-brown to green hornblende ± quartz; iii) *biotite-bearing amphibolite* made up of lepidoblastic to diablastic biotite + nematoblastic green hornblende-rich bands alternating with granular to polygonal andesine + quartz ± garnet-rich bands; iv) *gneissic amphibolite* consisting of andesine + quartz + green hornblende + biotite ± K-feldspar ± garnet.

Among the ultramafic rocks, massive coarse-grained **meta-hornblendite** prevails; it consists of brownish-green hornblende (up to 90%), and magnetite + apatite ± bytownite ± garnet. Subordinate medium-coarse grained massive **meta-pyroxenite** and banded **gneissic pyroxenite**, made up of diopside ± green hornblende ± bytownite ± garnet ± magnetite ± titanite, occur. Rare **meta-peridotite**, which shows an olivine ± clinopyroxene + pale-brown amphibole assemblage, is also present.

Impure **marble** and **Ca-silicate fels** consist of calcite + dolomite + Ca-rich plagioclase + phlogopite ± localized muscovite or diopside ± hornblende ± K-feldspar.

4.2. *Alpine re-equilibrated rocks*

4.2.1 *Stronger re-equilibrated area*

A - Mostly recrystallized rocks, exposed between S. Cristina and S. Luca.

Extensively re-equilibrated gneissic micaschist is prevalent. Its tectono-metamorphic evolution is summarized in Table I. Porphyroclastic Alm-rich garnet (Gt I), oligoclase and rare staurolite are the relict minerals of the Variscan amphibolite-facies rocks i.e. quartz + plagioclase + biotite + muscovite + garnet ± K-feldspar ± staurolite ± andalusite gneiss and micaschist. *A first Alpine deformation phase* (Da1) was responsible for a new pervasive foliation, accompanied by an almost complete recrystallization of the Variscan minerals to give a new assemblage, including quartz, sodic plagioclase + epidote, white mica (Wm II), Alm-rich garnet (Gt II), and greenish-brown amphibole and chlorite, both of which grew at the expense of relict biotite. In some places, sericite pseudomorphed porphyroblasts of Variscan andalusite. *A second deformation phase* (Da2), which prevalently did not obliterate the first one, gave the main foliation (Sm = S1 + S2). It was accompanied by the growth of oligoclase around albite, red-brown biotite (Bi II) around white mica II, quartz, white mica III and epidote. *A third deformation phase* (Da3) was responsible for the growth of the same minerals as those of the second one. *A fourth deformation phase* (DA4), developed in connection with submillimetric shear planes, only created a centimeter-spaced crenulation. Late veins of adularia and chlorite are also widespread.

Extensively re-equilibrated leuco-orthogneiss, which consists of overprinted granitic bodies and aplopegmatite dykes, shows the same Alpine tectono-metamorphic evolution as the above rock types (Table II) and exhibits a metamorphic foliation which can be ascribed to the second deformation phase (Sm = S1+ S2). This is suggested by the bimodal distribution of white mica (Wm II and III) and minor chlorite, which indicates the transposition of a previous foliation, and by the presence of elongated garnet II (grown at the expense of biotite I) rotated on this foliation. The main foliation is defined by thin layers of muscovite rimmed by minor biotite (Bi II), alternating with thicker layers consisting of polygonal granoblastic aggregates of quarz, albite rimmed by oligoclase and K-feldspar. In the overprinted granitic bodies, porphyroclastic feldspars are the only relict phases, whereas in the overprinted dykes, porphyroclastic garnets (Gt I) and/or grey-green tourmaline (Tou I), constantly rimmed by garnet II and blue tourmaline (Tou II), respectively, are also present.

In the extensively re-equilibrated meta-horneblendite (Table III) the first deformation phase (Da1) produced a subgranular recrystallization of the original centimeter-sized brown hornblende phenoblasts (Amph I) into neoblastic bluish-green amphibole (Amph II) + ilmenite, and also often led to the growth of neoblastic garnet (Gt II) after relict amphibole and/or plagioclase. The second deformation phase (Da2) formed shear planes (or, in some areas, a pervasive foliation), along which both pale-green amphibole (Amph III) and orange biotite recrystallized at the expense of amphibole II.

B - Partly recrystallized rocks

Partly re-equilibrated micaschist is widely prevalent. Its evolution, characterized by three tectono-metamorphic events, is summarized in Table IV. The Variscan rocks were medium-coarse grained quartz + biotite + plagioclase + fibrolitic or prismatic sillimanite + garnet ± muscovite ± K- feldspar gneiss, micaschist and fels. The rocks show the different stages of the Alpine tectono-metamorphic overprint and consist of submillimeter thick quartz + plagioclase + garnet bands alternating with layers of prevailing micas + Al-silicates. The first two overprinting events (Da1 and Da2), which were responsible for a reduction in grain-size, were accompanied by the development of either a crenulation or a new foliation, which are neither pervasive nor ubiquitous. The metamorphic overprint appears to be essentially static and relevant reactions mainly pseudomorphic or coronitic. A third event, ascribed to the fourth deformation phase (Da4) of the extensively re-equilibrated rock types, developed in connection with submillimetric shear planes. All the amphibolite-facies minerals may have remained as relict phases. In the most recrystallized rocks, the feldspars and sillimanite were pseudomorphed by newly-formed minerals. Quartz, micas, garnet and accessory phases (zircon, monazite, apatite, graphite and opaque ores), on the other hand, are common Variscan relicts. The neoblastic minerals which grew in the first metamorphic stage are:

White mica (Wm II) as: i) small flakes developed at the expense of large relict muscovite ii) small crystal after biotite; iii) sericite after sillimanite; iv) sericite pseudomorphing plagioclase; v) fine-grained flake associations pseudomorphing feldspars; vi) flakes intergrown with quartz, after K-feldspar + sillimanite relicts.

Quartz, which defines wavy ribbon-like layers of recrystallized granoblastic aggregates with indented to polygonal boundaries.

Plagioclase, partly retrogressed to granoblastic aggregates of more sodic plagioclase + zoisite ± sericite. It was formed by recrystallization of coarser-grained blasts, which are still preserved in places. In the most deformed rocks the plagioclase is commonly replaced by a sericite felt.

K-feldspar, which is only found as cataclastic relict grains.

Sillimanite, which has been pseudomorphed by a fine-grained aggregate of kyanite ± sericite. Relict fibrolite, both as inclusions in the muscovite porphyroclasts and as needles intergrown with relict biotite, and/or relict prismatic sillimanite, associated with biotite and defining the Variscan foliation, rarely occur.

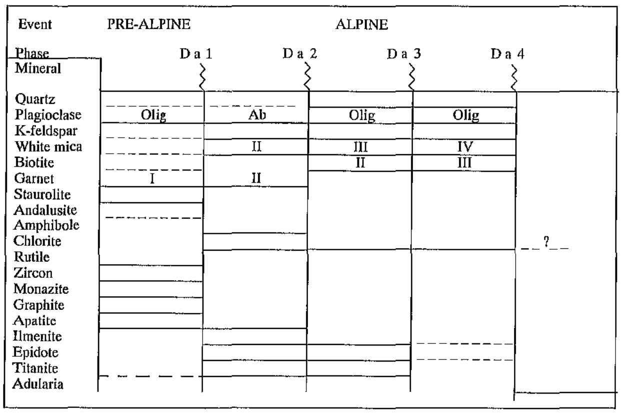
Ilmenite inclusion-rich *Garnet II*, which progressively replaced relict biotite, and mimetically acquired its tabular habit. Inclusion-free garnet II, which developed after plagioclase. The newly-formed crystals occur as clusters, as single grains or as rims overgrowing, in a thin corona, the relict porphyroclastic garnet (Gt I).

Chloritoid and/or chlorite found as porphyroblasts up to 4 cm in size and as needles intergrown with aggregates of sericite, both replacing the original sillimanite + biotite + garnet domains.

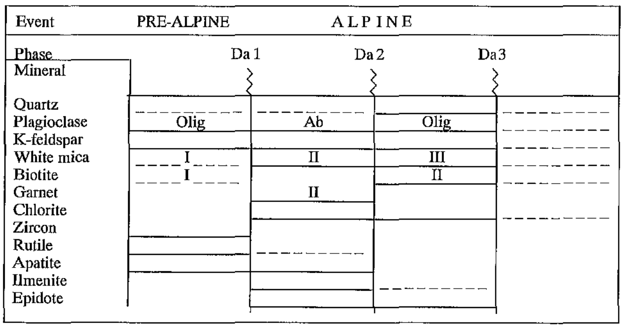
Colour-free *epidote* aggregates, showing anomalous blue interference colours which inherited their REE and radioactive elements, centripetally replacing Variscan monazite.

All the above-mentioned newly-formed minerals show a synkinematic to postkinematic growth during the first metamorphic recrystallization event. The second overprinting event generally produced a partial recrystallization of much finer grain-sized decussate *biotite (Bi II)* and white mica (Wm III) flakes at the expense of relict micas.

Partly re-equilibrated orthogneiss (Table V) still shows its original magmatic fabric. The relict minerals are numerous: large perthitic K-feldspar, sericitized and/or saussuritized plagioclase, micas, apatite, zircon and rare allanite. The Alpine gneissic structure is defined by a non-pervasive foliation, marked by white mica II and epidote. Albite,, garnet II, bluish-green or grey-green amphibole (Amph II) after plagioclase and biotite, kyanite needles replacing magmatic sillimanite, are the other newly-formed crystals belonging to the first Alpine event. Small decussate



Tab. I - Alpine tectono-metamorphic evolution of Variscan garnet ± staurolite ± andalusite *gneissic micaschist extensively re - equilibrated to* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar *gneissic micaschist.*



Tab. II - Alpine tectono-metamorphic evolution of *late-Variscan peraluminous plutonic rocks extensively re-equilibrated to* garnet II + white mica II + chlorite + biotite II *leuco-orthogneiss*

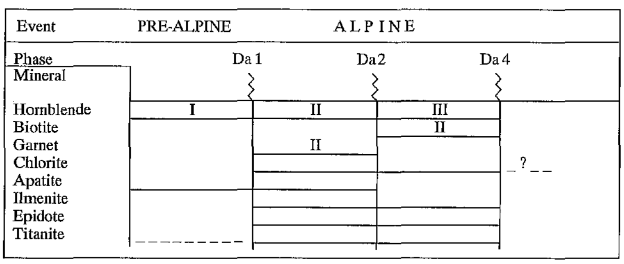


Table III - Alpine tectono-metamorphic evolution of *Variscan meta-hornblendite extensively re-equilibrated to* amphibole II ± garnetII ± chlorite ± biotite II *meta-hornblendite.*

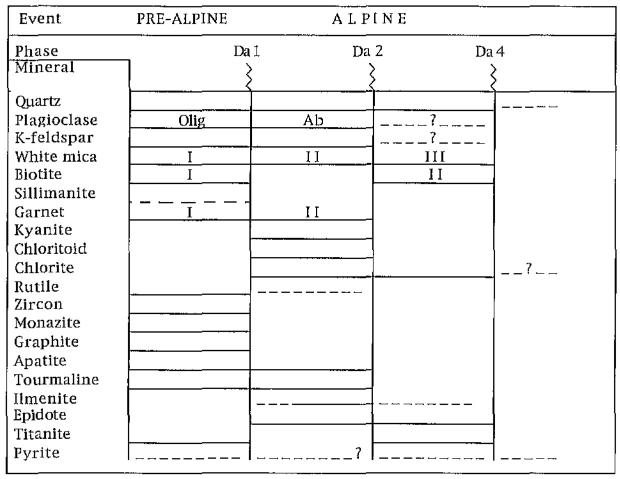


Table IV - Alpine tectono-metamorphic evolution of Variscan sillimanite ± garnet *gneiss, micaschist and fels partly re-equilibrated to* kyanite + garnet II + white mica II ± chloritoid ± chlorite ± biotite II *micaschist.*

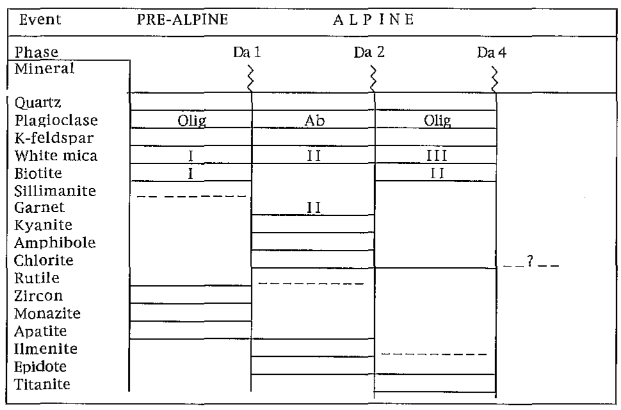


Table V - Alpine tectono-metamorphic evolution of *late-Variscan peraluminous plutonic rocks partly re-equilibrated to* kyanite + garnetII + white mica II + amphibole + chlorite + biotite II *orthogneiss.*

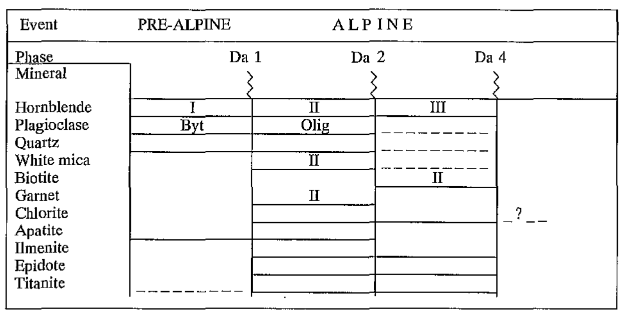


Table VI - Alpine tectono-metamorphic evolution of *Variscan amphibolite partly re-equilibrated to* amphibole II + garnet II + white mica II + chlorite + biotite II *amphibolite*

biotite flakes and oligoclase rims around albite are evidence of a second Alpine metamorphic phase.

Partly re-equilibrated amphibolite (Table VI). The relict amphibole shows its original brown-green colour and only at the rims recrystallized into very small pale-blue amphibole. In the most deformed types anorthite was replaced by oligoclase + zoisite and/or by small garnet (Gt II) and/or by paragonite. Rare red-orange biotite (Bi II), growing after amphibole II, is evidence of the second event.

4.2.2 *Weaker re-equilibrated area*

The area with weaker Alpine overprint extends around the more strongly overprinted zone (Fig. 1) and is characterized by a *cataclastic* to *mylonitic* deformation. In this fabric, quartz recrystallization is accompanied by the development of very fine-grained white mica, blue-green amphibole, epidote and bluish-green to green biotite and by the deformation and retrogression of plagioclase. Overprint relations are not clear, but, as evidenced by the tectono-metamorphic evolution of the strongly re-equilibrated rocks, bluish-green amphibole and part of the epidote and of the white mica grew during the first Alpine event, bluish-green biotite and part of the epidote and white mica during the second one.

5. Mineral Chemistry

The chemical composition of the relict and newly-formed mineral phases from representative *six extensively* and *eight partly Alpine overprinted rocks* (see APPENDIX, list of samples analysed) was determined by electron microprobe, using Bence & Albee (1968) corrections. Natural silicates were used as standards.

5.1 *Garnet*

The presence of two garnet generations in the *extensively and partly re-equilibrated micaschist* was confirmed by microprobe analyses (Tab. VII; Fig. 2). Both relict and neoblastic garnets are rich in almandine component: the former shows the highest Mn content, low Ca and Fe^{2+} scattering and almost similar Mg values; the latter exhibits a wide range of Ca and Fe^{2+}, which are

	a B1827				b B 1914	c B 1747	d B 1907				e B 1918
	□	+	•	o	o	o	□	+	•	o	o
SiO_2	37.38	37.06	37.50	37.03	38.31	38.44	37.26	37.50	37.40	37.27	37.38
TiO_2	0.12	0.32	0.11	0.15	0.04	0.05	0.00	0.01	0.04	0.08	0.10
Al_2O_3	21.89	21.86	21.80	22.02	21.85	22.02	21.94	22.19	22.33	22.20	22.36
Cr_2O_3	0.06	0.03	0.01	0.04	0.00	0.01	0.06	0.00	0.01	0.03	0.00
FeO	26.08	28.15	29.95	29.50	23.04	21.92	33.45	34.90	35.60	35.63	28.81
MnO	4.41	1.92	0.66	0.80	0.31	2.78	2.30	1.19	0.59	0.69	1.16
MgO	1.12	1.33	1.84	1.56	0.78	1.82	2.81	2.69	2.93	3.07	2.83
CaO	9.87	10.01	8.92	9.57	16.74	14.66	3.10	2.74	2.25	2.20	8.00
Na_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.93	100.68	100.79	100.67	101.07	101.70	100.92	101.22	101.15	101.17	100.64
Cations per 12 oxygens											
Si	2.957	2.936	2.965	2.931	2.983	2.970	2.963	2.976	2.968	2.957	2.941
Al^{IV}	0.043	0.064	0.035	0.069	0.017	0.030	0.037	0.024	0.032	0.043	0.059
	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000
Al^{VI}	1.999	1.978	1.996	1.986	1.989	1.975	2.020	2.052	2.057	2.033	2.015
Ti	0.007	0.019	0.007	0.009	0.002	0.003	0.000	0.001	0.002	0.005	0.006
	2.006	1.997	2.003	1.995	1.991	1.978	2.020	2.053	2.059	2.038	2.021
Mg	0.132	0.157	0.217	0.184	0.091	0.210	0.333	0.318	0.347	0.363	0.332
Fe^{2+}	1.700	1.820	1.955	1.891	1.478	1.368	2.211	2.316	2.363	2.364	1.863
Mn	0.296	0.129	0.044	0.054	0.020	0.182	0.155	0.080	0.040	0.046	0.077
Cr	0.004	0.002	0.001	0.003	0.000	0.001	0.004	0.000	0.001	0.002	0.000
Ca	0.837	0.850	0.755	0.812	1.397	1.213	0.264	0.233	0.191	0.187	0.674
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	2.969	2.958	2.972	2.944	2.986	2.974	2.967	2.947	2.942	2.962	2.946
Alm	57.3	61.3	65.7	64.2	49.5	46.0	74.6	78.6	80.4	79.8	63.2
Grs	28.1	28.8	25.5	27.6	46.8	40.8	8.7	7.9	6.5	6.2	22.9
Sps	10.0	4.4	1.5	1.8	0.7	6.1	5.2	2.7	1.3	1.6	2.6
Prp	4.5	5.3	7.3	6.3	3.0	7.1	11.2	10.8	11.8	12.3	11.3

Extensively (a) (b) (c) and partly (d) and (e) re-equilibrated rock types; see list of samples analysed.
Symbols as in Figs. 2 and 3.
Tructural formulae calculated folowing Rickwood's (1968) method.

Table VII - Representative micropobe analyses of garnet clearly dependent on the composition of the microdomain in which garnet crystallized. Generally garnet with the highest Fe^{2+} developed from relict biotite, whereas garnet with the highest Ca content formed after plagioclase. Garnet grown at the expense of both the above relict minerals show scattered intermediate values. The different Ca/Fe^{2+} ratio of relict garnet (Gt I) from the two petrographic groups is related to a different original chemical composition.

In the Mn-Fe^{2+}-Ca diagram from the extensively re-equilibrated gneissic micaschist (Fig. 2a) the relict garnet (Gt I) shows Mn and Ca decreasing from core to rim and Fe^{2+} lower than that of Gt II. The different Ca/Fe^{2+} ratio shown by the neoblastic garnet suggests that the grains overgrowing the relict garnet developed only after Variscan biotite, whereas the isolated crystals grew at the expense of relict biotite and plagioclase. In the Mn-Fe^{2+}-Ca diagrams from the partly re-equilibrated micaschist (Fig. 2b and c) garnet II data also show different Ca/Fe^{2+} ratios depending on their structural site: in b and c diagrams are plotted garnets of biotite-rich and plagioclase-rich Variscan micaschists, respectively.

The Mn-Mg-Fe^{2+} diagrams (Fig. 2) show a narrow Mg compositional range for the relict and newly-formed garnets.

In the *extensively* and *partly re-equilibrated orthogneiss* and *meta-hornblendite* (Tab. VII; Fig. 3) the newly-formed garnets, developed after plagioclase and biotite in the orthogneiss or after plagioclase and amphibole in the hornblendite, are almandine-rich with the highest Ca content.

The mineral chemistry of the Alpine garnet, in

	a		b	c			d	
	B1827		B 1914	B1957			B1918	
	○	◊	○	•	+	○	•	+
SiO_2	46.74	46.30	47.92	45.08	44.56	46.32	45.72	47.19
TiO_2	0.47	0.10	0.52	0.70	0.13	0.40	0.74	0.54
Al_2O_3	32.90	38.88	29.28	36.20	39.82	35.07	35.74	32.34
Cr_2O_3	0.05	0.05	0.00	0.03	0.04	0.06	0.00	0.00
FeO	1.56	0.77	3.98	1.15	0.53	1.41	1.40	2.06
MnO	0.02	0.00	0.04	0.02	0.01	0.00	0.03	0.00
MgO	1.82	0.30	1.91	0.59	0.22	1.04	0.69	1.52
CaO	0.01	0.36	0.01	0.00	0.75	0.00	0.01	0.04
Na_2O	1.33	6.87	0.25	0.86	5.71	1.50	0.49	0.17
K_2O	9.34	1.11	11.52	10.30	2.47	9.05	10.63	11.09
Total	94.24	94.74	95.43	94.93	94.24	94.85	95.45	94.95
Cations per 11 oxygens								
Si	3.118	2.976	3.229	3.000	2.893	3.063	3.027	3.156
Al^{IV}	0.882	1.024	0.771	1.000	1.107	0.937	0.973	0.844
	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Al^{VI}	1.705	1.922	1.555	1.840	1.941	1.797	1.816	1.706
Ti	0.024	0.005	0.026	0.035	0.006	0.020	0.037	0.027
Cr	0.003	0.003	0.000	0.002	0.002	0.003	0.000	0.000
Fe^{3+}	0.087	0.041	0.224	0.064	0.029	0.078	0.078	0.115
Fe^{2+}	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Mn	0.001	0.000	0.002	0.001	0.001	0.000	0.002	0.000
Mg	0.181	0.029	0.192	0.059	0.021	0.102	0.068	0.152
	2.001	2.000	1.999	2.001	2.000	2.000	2.001	2.000
Ca	0.001	0.025	0.001	0.000	0.052	0.000	0.001	0.003
Na	0.172	0.856	0.033	0.111	0.719	0.192	0.063	0.022
K	0.795	0.091	0.990	0.874	0.205	0.763	0.898	0.946
	0.968	0.972	1.024	0.985	0.976	0.955	0.962	0.971

Extensively (a) and (b)and partly (c) and (d) re-equilibrated rock types; see list of samples analysed. Symbols as in Figs. 4 and 5. Structural formulae calculated following Laird & Albee's (1981) method.

Tab. VIII - Rapresentative microprobe analyses of white mica

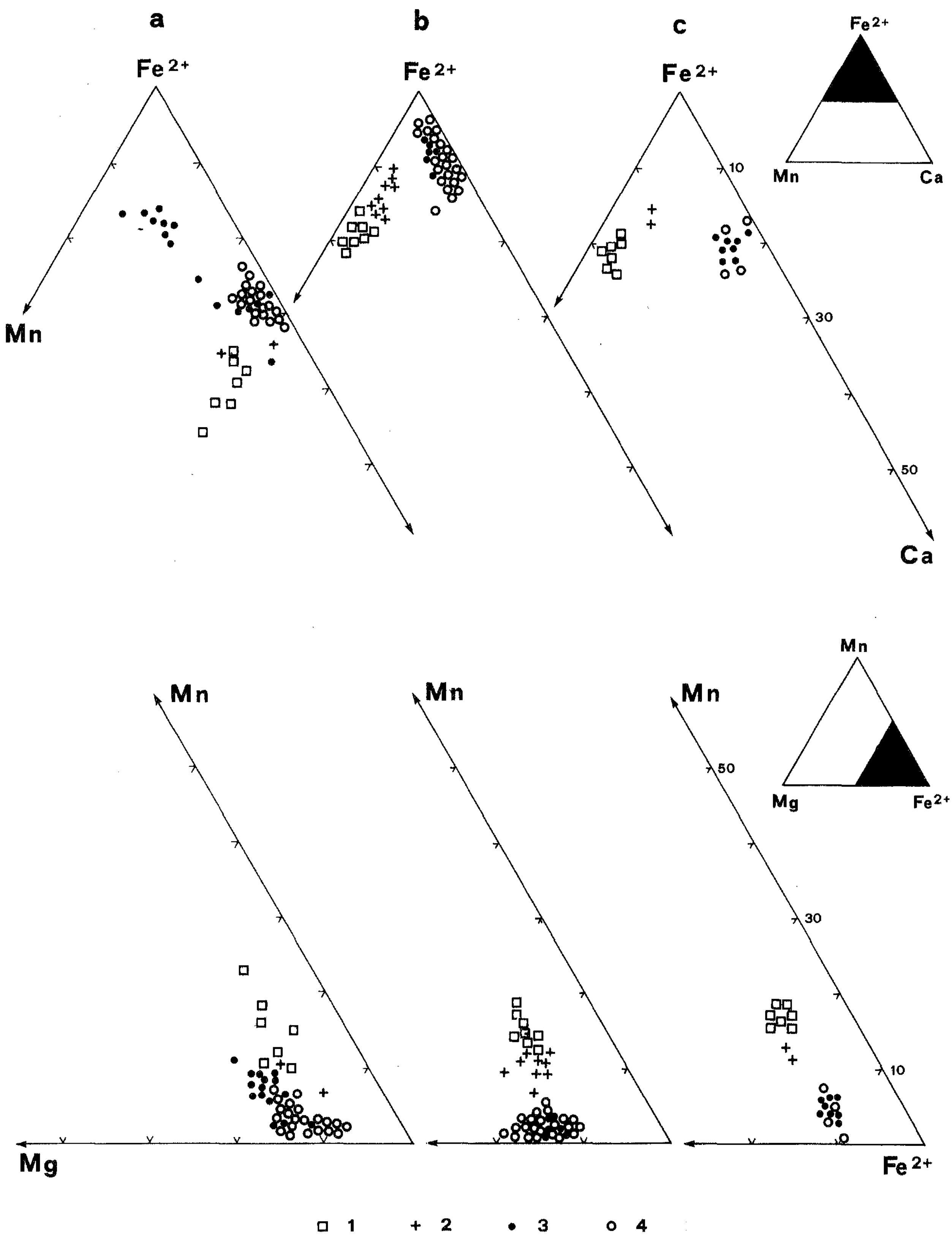


Fig. 2 - Fe^{2+}-Mn-Ca and Mn-Mg-Fe^{2+} diagrams showing the compositional variations in the **relict (Gt I)** and **neoblastic (Gt II) garnets** from:
a) *Variscan* garnet ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar gneissic *micaschist.*
b and c) *Variscan* sillimanite ± garnet *gneiss, micaschist* (b) and *fels* (c) *partly re-equilibrated to Alpine* kyanite + garnet II + white mica II ± chloritoid ± *chlorite* + biotite II *micaschist.*
— **Garnet** I: 1 = relict core; 2 = intermediate zone. **Garnet II**: 3 = neoblastic corona overgrowing relict garnet; 4 = single isolated idioblastic crystals.

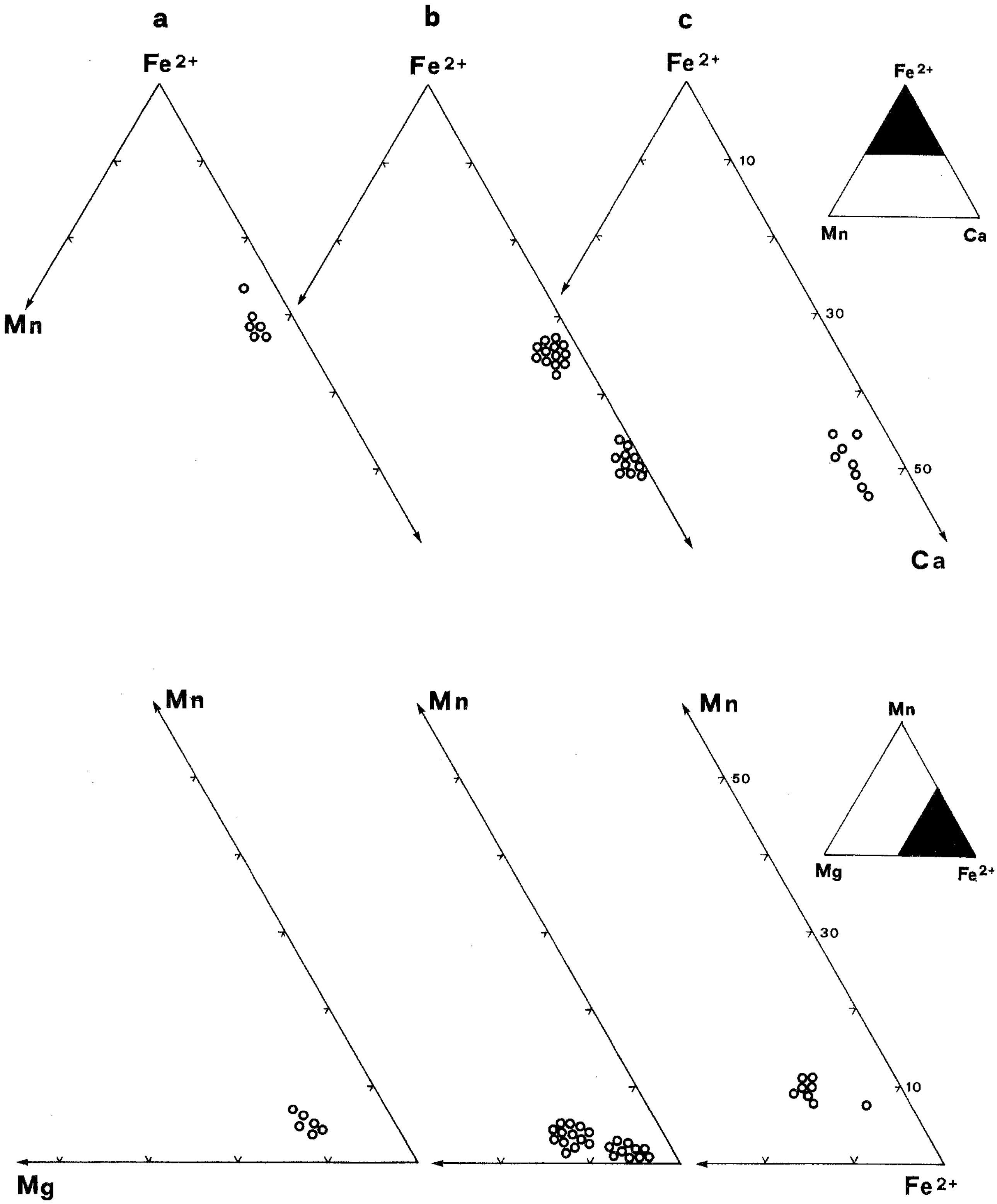


Fig. 3 - Fe^{2+}-Mn-Ca and Mn-Mg-Fe^{2+} diagrams showing the chemical composition of the **neoblastic garnet (Gt II)** from:
a) *Late-Variscan peraluminous plutonic rocks extensively re-equilibrated to Alpine* garnet II + white mica II + biotite II + chlorite *leuco-orthogneiss.*
b) *Late-Variscan peraluminous plutonic rocks partly re-equilibrated to Alpine* kyanite + garnet II + white mica II + amphibole + biotite II + chlorite *orthogneiss.*
c) *Variscan meta-hornblendite extensively re-equilibrated to Alpine* amphibole II ± garnet II ± chlorite ± biotite II *meta-hornblendite.*
— **Neoblastic garnet (Gt II)**: single isolated idioblastic crystals growing at the expense of relict magmatic plagioclase and biotite (a and b) and of relict amphibole (c).

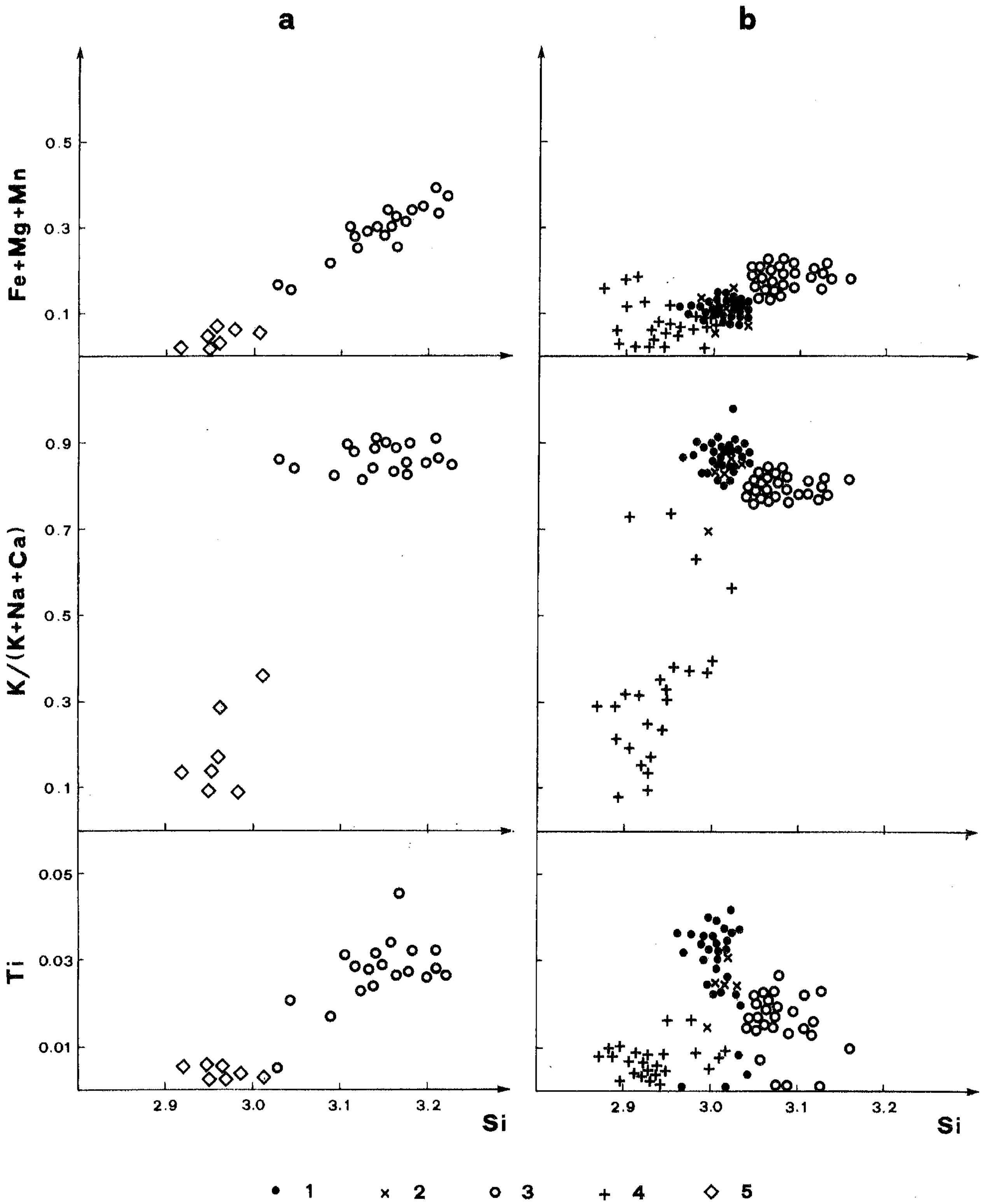


Fig. 4 - Si vs. Ti, Si vs. K/(K + Na + Ca) and Si vs. Fe + Mg + Mn diagrams showing the chemical composition of **relict (WmI) and neoblastic (Wm II, III, IV) white mica** from:

a) *Variscan* garnet ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar *gneissic micaschist.* **Neoblastic white mica**: 3 = small phengite flakes after relict biotite and white mica **(Wm II, III and IV)**; 5 = paragonite flakes after Variscan plagioclase **(Wm II, III and IV)**.

b) *Variscan* sillimanite ± garnet gneiss, *micaschist and fels partly re-equilibrated to Alpine* kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II *micaschist.*

— **Relict white mica**: 1 = large flakes (Wm I). **Neoblastic white mica**: 2 = sericite K-muscovite pseudomorphs after relict muscovite and K-feldspar **(Wm II)**; 3 = small neoblastic phengite flakes developed after biotite **(Wm II and III)**; 4 paragonitic sericite after Variscan plagioclase **(Wm II)**.

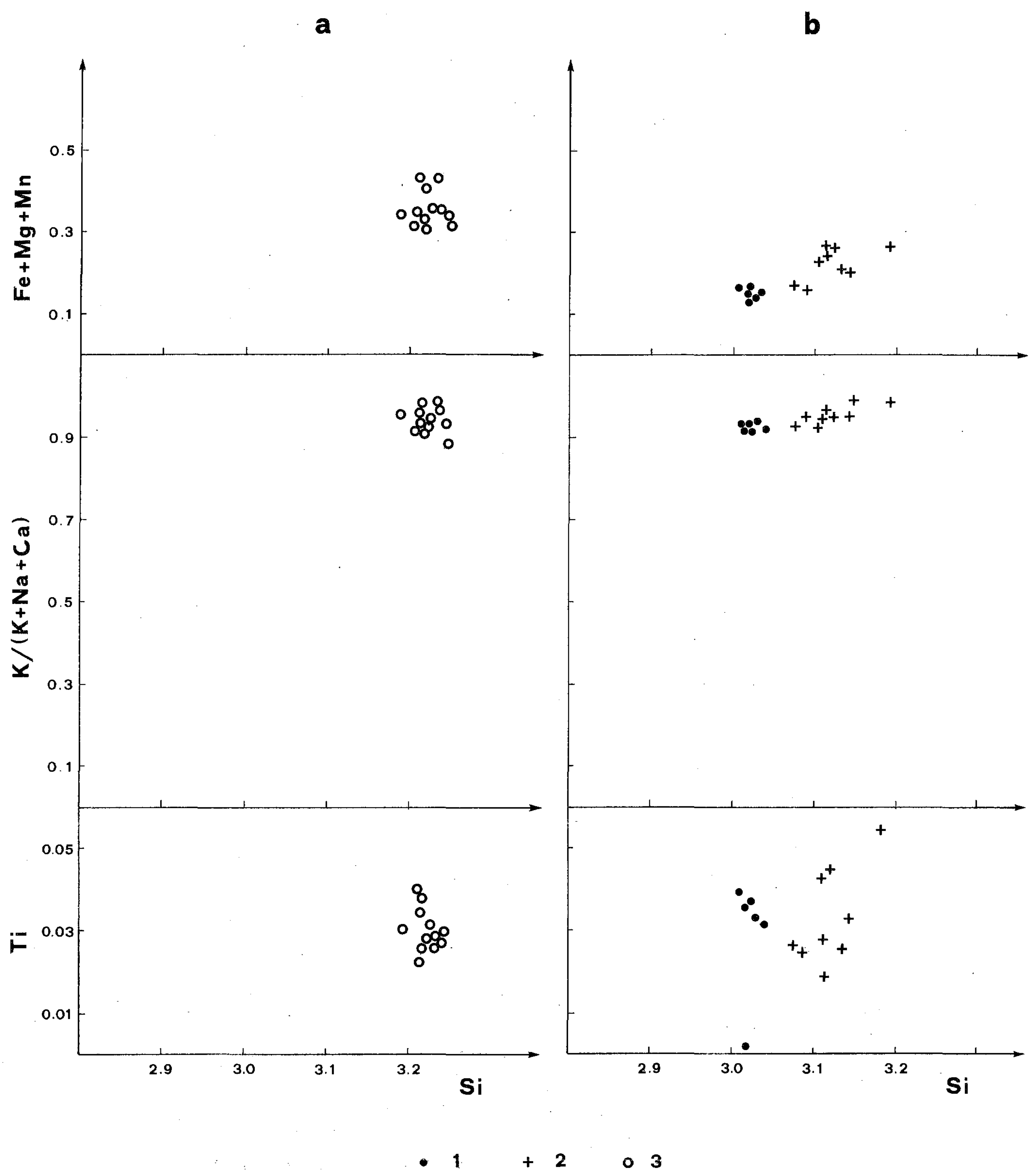


Fig. 5 - Si vs. Ti, Si vs. K/(K + Na + Ca), Si vs. Fe + Mg + Mn diagrams showing the chemical composition of **relict (Wm I) and neoblastic (Wm II, III, IV) white mica** from:

a) *Late-Variscan peraluminous plutonic rocks extensively re-equilibrated to Alpine* garnet II + white mica II + biotite II + chlorite *leuco-orthogneiss.* — **Neoblastic white mica**: 3 = phengite flakes (**Wm III and IV**) after magmatic micas.

b) *Late-Variscan peraluminous plutonic rocks partly re-equilibrated to Alpine* kyanite + garnet II + white mica II + amphibole + biotite II + chlorite *orthogneiss.* —**Relict white mica**: 1 = large magmatic muscovite (**Wm I**). Neoblastic white mica: 2 = phengite sericite (**Wm II**) after relict magmatic micas.

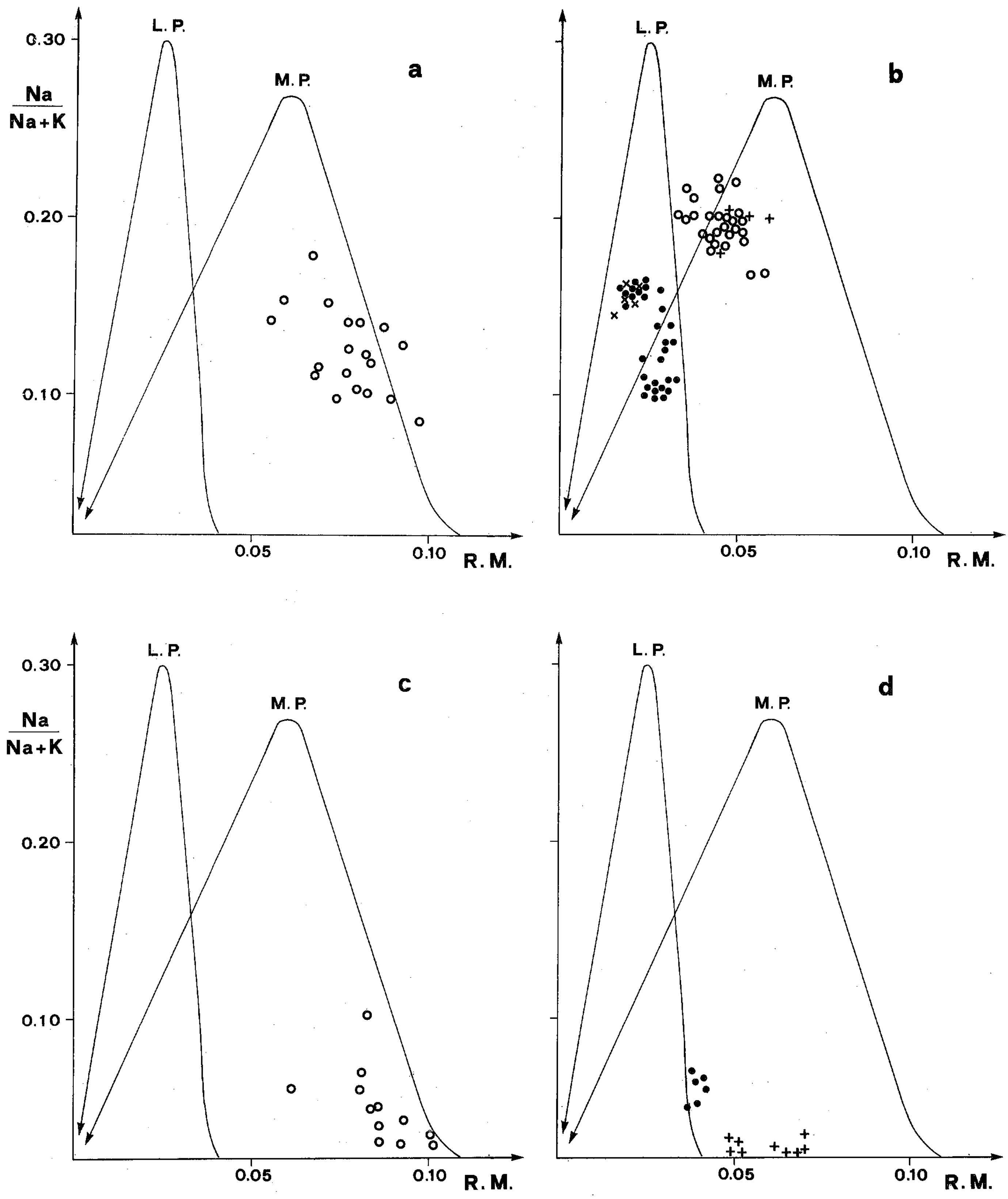


Fig. 6 - RM vs. Na/(Na + K) diagram (Guidotti & Sassi, 1976) showing the relation between the compositional variations of the Variscan and Alpine white micas and the temperature increase at different baric conditions in the *extensively* (a) and *partly* (b) *re-equilibrated micaschist* and in the *extensively* (c) and *partly* (d) *re-equilibrated orthogneiss*. Symbols as in Figs. 4 and 5. In this diagram the paragonite cannot be represented.

	a		b				c	
	B 2164		B1957		B 1905		B 1918	
	○	○	●	○	●	○	●	○
SiO_2	38.20	37.64	36.08	35.31	36.66	36.17	34.79	35.61
TiO_2	1.87	2.11	1.56	1.40	1.44	1.56	2.37	4.37
Al_2O_3	18.56	19.51	19.21	19.68	18.57	18.67	19.48	16.70
Cr_2O_3	0.04	0.07	0.01	0.00	0.02	0.00	0.00	0.00
FeO	16.10	15.76	18.37	20.20	16.84	18.95	22.35	22.07
MnO	0.14	0.14	0.04	0.06	0.05	0.02	0.08	0.07
MgO	11.54	11.03	10.88	9.11	11.29	10.23	8.05	7.13
CaO	0.04	0.02	0.00	0.05	0.00	0.02	0.06	0.10
Na_2O	0.15	0.11	0.14	0.16	0.28	0.21	0.86	0.05
K_2O	9.82	10.03	8.89	9.08	9.42	8.96	8.42	9.27
Total	96.46	96.42	95.18	95.05	94.57	94.79	96.46	95.37
Cations per 11 oxygens								
Si	2.790	2.748	2.717	2.687	2.752	2.739	2.645	2.739
Al^{IV}	1.210	1.252	1.283	1.313	1.248	1.261	1.355	1.261
	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Al^{VI}	0.388	0.427	0.423	0.452	0.395	0.407	0.391	0.253
Ti	0.103	0.116	0.088	0.080	0.081	0.089	0.136	0.253
Cr	0.002	0.004	0.001	0.000	0.001	0.000	0.000	0.000
Fe^{3+}	0.188	0.200	0.023	0.079	0.142	0.071	0.011	0.075
Fe^{2+}	0.796	0.762	1.134	1.206	0.916	1.130	1.410	1.345
Mn	0.009	0.009	0.003	0.004	0.003	0.001	0.005	0.005
Mg	1.256	1.200	1.221	1.033	1.263	1.155	0.912	0.817
	2.742	2.718	2.893	2.854	2.801	2.853	2.865	2.748
Ca	0.003	0.002	0.000	0.004	0.000	0.001	0.005	0.008
Na	0.021	0.016	0.020	0.024	0.041	0.031	0.127	0.007
K	0.915	0.934	0.854	0.881	0.902	0.866	0.817	0.910
	0.939	0.952	0.874	0.909	0.943	0.898	0.949	0.925

ExtensIvely (a) and partly (b) and (c) re-equilibrated rock types; see list of samples analysed. Symbols as in Figs. 7 to 9. Structural formulae calculated following Dymek's (1983) method.

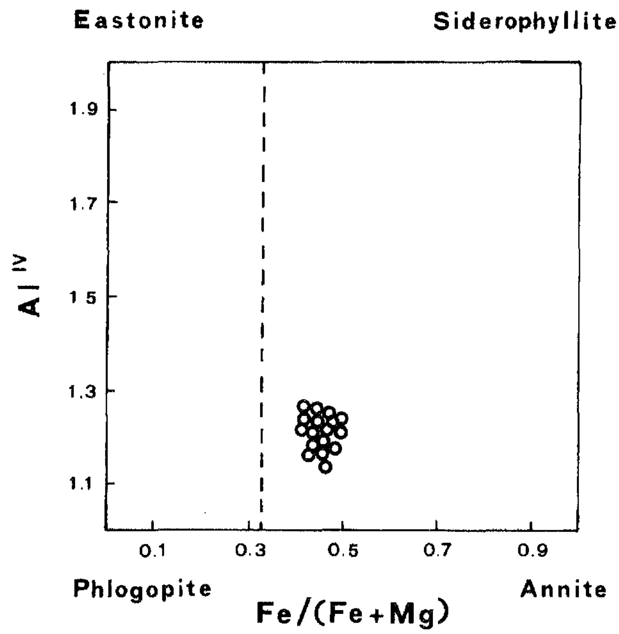
Tab. IX - Representative microprobe analyses of biotite

agreement with the results of Messina et al. (1990) for the overprinted area of the Peloritani Mountains, indicates that **the relatively high-P Alpine tectono-metamorphic event produced Fe and/or Ca-rich almandine in the Aspromonte Unit.**

5.2 *White mica*

In the *extensively re-equilibrated gneissic micaschist* both phengite and paragonite developed during the three Alpine events (Wm II) (Wm III) (Wm IV) (Tab. VIII; Fig. 4a). It is very difficult to distinguish the different structural sites of these newly-formed minerals, because of their pervasive re-equilibration. The phengite shows a high K/(K+Na+Ca) ratio (0.8-0.93), noticeable Si (3.1-3.2) and Fe+Mg+Mn (0.1 to 0.4) contents, and a Ti value (0.02 to 0.05) which increases with increasing celadonitic content. The paragonite shows low Fe+Mg+Mn and Ti contents (< 0.1 and < 0.01, respectively) and a K/(K+Na+Ca) ratio ranging from 0.1 to 0.4.

In the *partly re-equilibrated micaschist* (Tab. VIII; Fig. 4b) the white mica show different compositions according to the structural types: the large relict flakes and the neoblastic white mica types i), iii) and v) (see chapter 4.2.1) are muscovites with a scattered Ti content (0.0 to 0.045); the newly-formed flakes grown at the expense of biotite (type ii) are low phengitic muscovites (Si = 3.03 to 3.17; Fe+Mg+Mn = 0.1 to 0.27); the neoblastic sericite replacing plagioclase (type iv) is paragonite with a K/



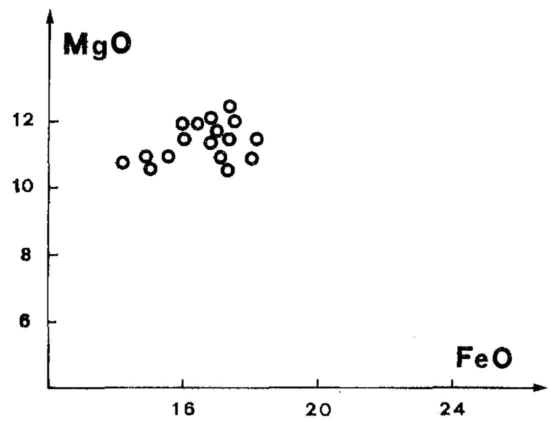


Fig. 7 - Fe/(Fe + Mg) vs. Al^{IV} classification diagram and FeO/MgO diagram of **neoblastic biotite (Bi II, III)** from *Variscan* garnet ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar gneissic *micaschist*. The dashed line divides the phlogopite from the biotite compositional field. —**Neoblastic biotite:** = oriented or decussate red-brown flakes **(Bi II and III)**.

(K+Na+Ca) ratio ranging from 0.1 to 0.75. Some analyses of newly-formed sericite show intermediate values between muscovite and paragonite, suggesting the existence of a submicroscopic intergrowth of the two white micas.

In the *extensively re-equilibrated orthogneiss* (Tab. VIII; Fig. 5a) the newly-formed flakes (Wm II) are homogeneous phengite with the highest Si (3.25) and Fe+Mg+Mn (0.3 to 0.45) contents.

In the *partly re-equilibrated orthogneiss* (Tab. VIII; Fig. 5b) all the analysed white micas are muscovites with a low celadonitic substitution. Neoblastic white micas are more phengitic (Si = 3.07 to 3.18; Fe+Mg+Mn = 0.1 to 0.28) than the relict ones (Si = 3.05; Fe+Mg+Mn = 0.1 to 0.2). The varying Ti content of the newly-formed sericite was inherited from the parental phases: in fact the highest (0.055) and the lowest (0.018) Ti contents are given by the white micas which developed after relict biotite and muscovite, respectively.

In the RM vs. Na/(Na+K) diagram (Guidotti & Sassi, 1976) (Fig. 6), which interprets the composition of the white mica in terms of the pressure conditions, it is clear that the Alpine and the Variscan white micas crystallized at medium and low P, respectively.

	a		b		
	B 1827		B1918		
	o	o	o	o	o
SiO_2	46.24	45.61	42.31	37.55	47.03
TiO_2	0.31	0.00	0.08	0.77	0.39
Al_2O_3	13.99	14.69	19.40	19.17	17.39
Cr_2O_3	0.00	0.03	0.01	0.01	0.00
FeO	13.90	13.64	17.24	20.82	17.51
MnO	0.14	0.08	0.13	0.17	0.15
MgO	11.31	10.67	2.91	3.69	3.01
CaO	7.66	8.80	8.48	9.87	8.11
Na_2O	2.51	2.67	1.28	1.43	1.18
K_2O	0.35	0.33	5.89	3.27	3.05
Total	96.41	96.52	97.73	96.75	97.82
Cations per 23 oxygens					
Si	6.565	6.564	6.539	5.826	7.022
Al^{IV}	1.435	1.436	1.461	2.174	0.978
	8.000	8.000	8.000	8.000	8.000
Al^{VI}	0.907	1.056	2.074	1.332	2.082
Fe^{3+}	1.378	0.858	0.000	0.303	0.000
Ti	0.033	0.000	0.009	0.090	0.044
Cr	0.000	0.003	0.001	0.001	0.000
Mg	2.393	2.288	0.670	0.853	0.669
Fe^{2+}	0.273	0.784	2.228	2.398	2.186
Mn	0.017	0.010	0.017	0.022	0.018
	5.001	4.999	4.999	4.999	4.999
Ca	1.165	1.356	1.404	1.640	1.297
Na	0.691	0.644	0.384	0.360	0.342
	1.856	2.000	1.788	2.000	1.639
Na	0.000	0.100	0.000	0.071	0.000
K	0.063	0.062	1.161	0.647	0.580
	0.063	0.162	1.161	0.718	0.580
$Mg/Mg+Fe^{2+}$	0.898	0.745	0.231	0.262	0.669

Extensively (a) and partly (b) re-equilibrated rock types; see list of samples analysed. Symbols as in Fig. 10. Structural formulae calculated following Robinson's et al. (1982) method.

Tab. XA - Representative microprobe analyses of amphibole.

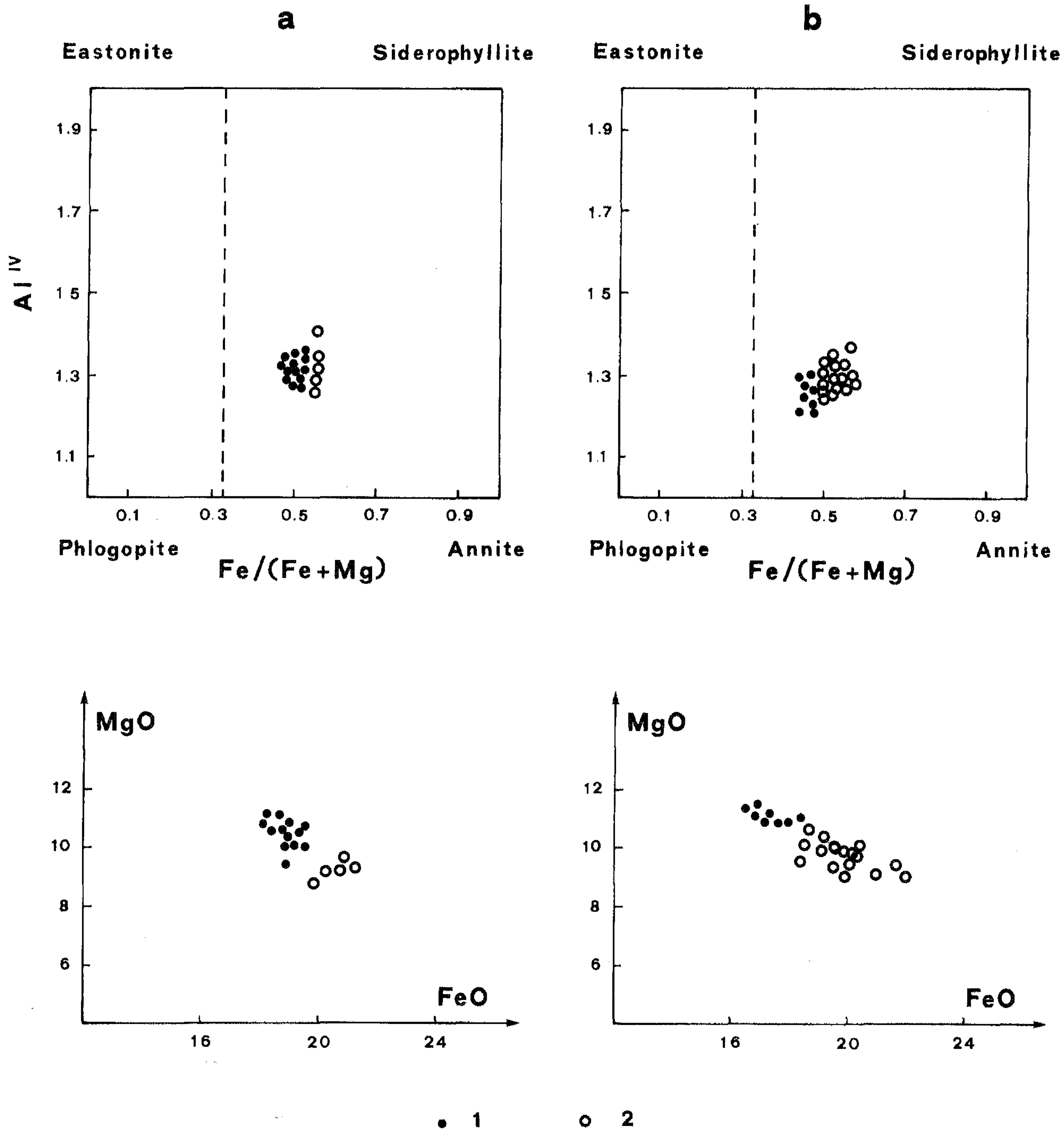


Fig. 8 - Fe/(Fe+ Mg) vs. Al^{IV} classification diagram and FeO/ MgO diagram of **relict (Bi I) and neoblastic (Bi II)biotite** from *Variscan* sillimanite+ garnet *gneissic micaschist* (a) *micaschist* and *fels* (b) *partly re-equilibrated to Alpine* Kyanite + garnet II + white mica II + chloritoid + chlorite + biotite II micaschist.

— **Relict biotite**: 1 = large red - brown relict flakes **(Bi I). Neoblastic biotite:** 2 = small red flakes growing around the white mica I rims or as clusters of decussate laminae **(Bi II)**.

5.3. *Biotite*

In the *extensively re-equilibrated gneissic micaschist* (Tab. IX; Fig. 7) the newly-formed biotite shows a homogeneous phlogopite-annite composition with high Ti content (0.07-0.12) and Mg/Fe ratio (1.1 to 1.3).

In the partly re-equilibrated micaschist (Tab. IX; Fig. 8a and b) all the analysed crystals exhibit an annite-phlogopite composition, but brown relict and red-brown neoblastic biotites show different Fe/Fe+Mg ratios because the main octahedral substitution of neoblastic biotite is of Fe^{2+} replacing Mg. This suggests that in these rocks the temperature decreased during the Alpine tectono-metamorphic overprint. Both relict and newly-formed

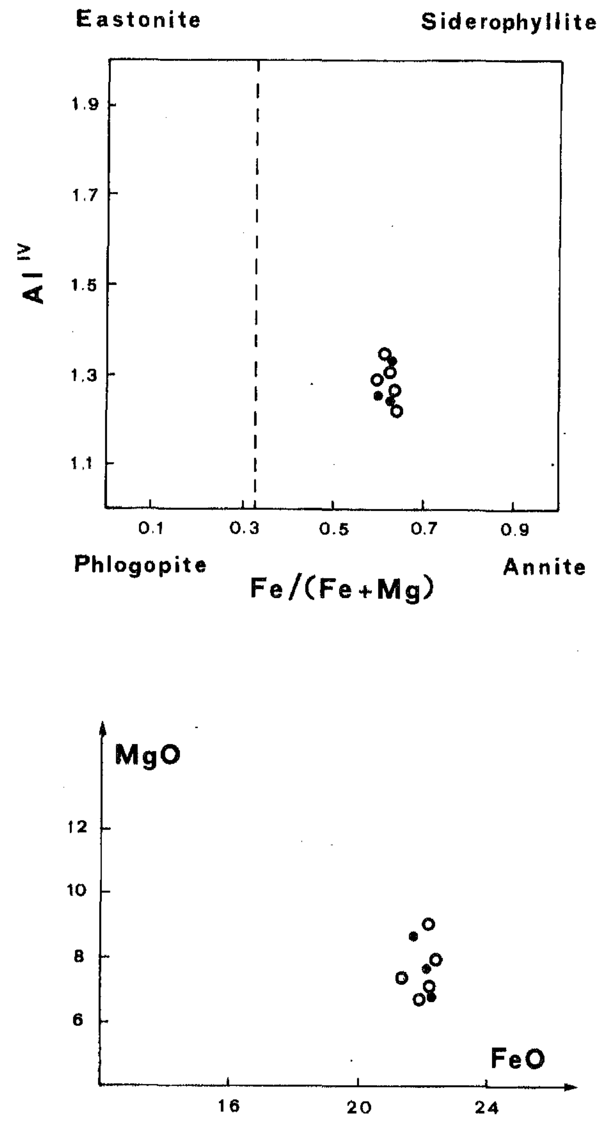


Fig. 9 - Fe/(Fe + Mg) vs. Al^{IV} classification diagram and FeO/MgO diagram of **relict (Bi I) and neoblastic (Bi II) biotite** from *late-Variscan peraluminous plutonic rocks partly re-equilibrated to Alpine* kyanite + garnet II + white mica II + amphibole + biotite II + chlorite orthogneiss. — **Relict biotite**: 1 = large magmatic flakes **(Bi I)**. **Neoblastic biotite**: 2 = very small flakes growing around white mica I or as decussate blasts **(Bi II)**.

biotites show Mg/Fe ratios and Ti contents (0.05 to 0.10) lower than those of the extensively re-equilibrated gneissic micaschist, which exhibits lower grade Variscan assemblages with respect to the partly re-equilibrated micaschist. This discrepancy is probably due to the partial recrystallization of the rock: consequently the relict biotite is only texturally preserved and does not show its original composition.

In the *partly re-equilibrated orthogneiss* (Tab. IX; Fig. 9) relict and neoblastic biotites show annite-phlogopite compositions and similar Fe/Fe+Mg ratios. This could indicate that the rare relict magmatic biotite is only structurally preserved. Bi I and Bi II show a low Mg/Fe ratio (0.5 to 0.75) and the highest Ti content (0.12 to 0.24).

All the newly-formed biotites exhibit Al and Ti contents typical of an intermediate temperature metamorphism as suggested by Dahl, 1970; Dallmeyer, 1974; Evans & Guidotti, 1966; Guidotti, 1970; Guidotti et al, 1977; Kwak, 1968.

5.4 *Amphibole*

In the *extensively re-equilibrated gneissic micaschist* (Tab. XA; Fig. 10a) the neoblastic amphibole ranges from tschermakitic to Mg-hornblende (Leake, 1978) characterized by high K (0.19 to 0.24) and low Ti (0.002 to 0.058) contents, and with Fe = 1.61-1.89; Mg = 1.96-2.46; Al = 1.96-2.46.

In the *partly re-equilibrated orthogneiss* (Tab. XA; Fig. 10b) the neoblastic amphibole shows a composition ranging from hastingsite to Fe-edenite (Leake, 1978). The amphibole grains close to plagioclase show a Si content slightly lower than that of the grains close to biotite, and are rich in K (0.15 to 0.22), Al (3.33 to 3.92), Fe (2.38 to 2.78), with Mg = 0.71-1.08 and Ti = 0.002- 0.09.

In the *extensively re-equilibrated meta-hornblendite* (Tab. XB; Fig. 11a, b and c) the chemical composition of the structurally preserved pale green porphyroclastic amphibole covers a continuous range from tschermakitic to Mg-hornblende (Ti = 0.00-0.034; Fe = 1.48-1.85; Mg =

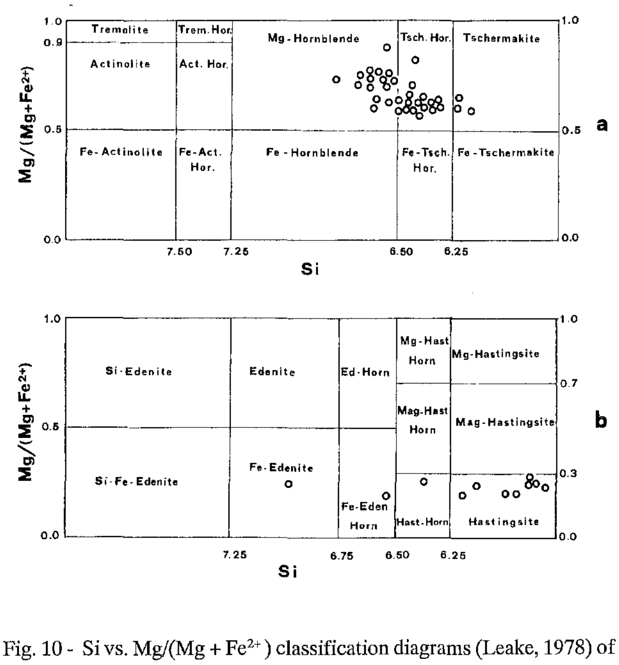


Fig. 10 - Si vs. $Mg/(Mg + Fe^{2+})$ classification diagrams (Leake, 1978) of the **neoblastic amphibole** from:

a) *Variscan garnet* ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar *gneissic micaschist*.

b) *Late-Variscan peraluminous plutonic rocks partly re-equilibrated to Alpine* kyanite + garnet II + white mica II + amphibole + biotite II + chlorite orthogneiss.

— **Neoblastic amphibole** = bluish-green and pale brown grains growing after relict biotite.

	a					b	
	B 1748		B 1747			B 1922	
	•	o	o	*	o	•	o
SiO_2	42.87	41.80	50.73	47.93	43.56	45.50	51.42
TiO_2	0.31	0.44	0.12	0.18	0.31	1.25	0.32
Al_2O_3	13.75	12.25	5.26	8.02	12.35	10.94	5.47
Cr_2O_3	0.03	0.00	0.00	0.00	0.02	0.16	0.05
FeO	15.03	13.77	11.10	12.11	13.92	13.68	9.97
MnO	0.25	0.13	0.21	0.23	0.21	0.25	0.28
MgO	10.72	12.29	15.84	13.94	11.88	11.67	15.08
CaO	11.68	12.15	12.48	12.20	12.02	11.52	14.18
Na_2O	1.53	1.72	0.76	1.14	1.80	1.50	0.79
K_2O	1.00	1.04	0.27	0.38	1.00	0.80	0.27
Total	97.17	95.59	96.77	96.13	97.07	97.07	97.83
Cations per 23 oxygens							
Si	6.326	6.281	7.320	7.031	6.443	6.669	7.494
Al^{IV}	1.674	1.719	0.680	0.969	1.557	1.131	0.506
	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al^{VI}	0.718	0.451	0.215	0.418	0.596	0.567	0.434
Fe^{3+}	0.564	0.556	0.317	0.281	0.375	0.253	0.000
Ti	0.034	0.050	0.013	0.020	0.034	0.139	0.035
Cr	0.003	0.000	0.000	0.000	0.002	0.019	0.006
Mg	2.358	2.752	3.406	3.048	2.619	2.560	3.275
Fe^{3+}	1.291	1.174	1.022	1.205	1.346	1.431	1.215
Mn	0.031	0.017	0.026	0.029	0.026	0.031	0.035
	4.999	5.000	4.999	5.001	4.998	5.000	5.000
Ca	1.847	1.956	1.929	1.917	1.905	1.817	2.000
Na	0.153	0.044	0.071	0.083	0.095	0.183	0.000
	2.000	2.000	2.000	2.000	2.000	2.000	2.000
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.214
Na	0.284	0.457	0.142	0.242	0.421	0.247	0.223
K	0.188	0.199	0.050	0.071	0.189	0.150	0.050
	0.472	0.656	0.192	0.313	0.610	0.397	0.487
$Mg/Mg+Fe^{2+}$	0.646	0.701	0.769	0.717	0.660	0.641	0.729

Extensively re-equilibrated meta-hornblendite (a) and partly re-equilibrated amphibolite (b); see list of samples analysed. Symbols as in Fig. 11. Structural formulae calculated following Robinson's et al. (1982) method.

Tab. XB - Representative microprobe analyses of amphibole

2.35-3.05; Al = 1.38-2.50) (Fig. 11a). The different compositions of the newly-formed crystals Mag-hastingsite to Mg-hastingsitic hornblende (Ti = 0.003-0.079; Fe = 1.58-1.85; Mg = 2.37-2.80; Al = 2.47-2.62) (Fig. 11b), pargasite to pargasitic hornblende (Ti = 0.038-0.054; Fe = 1.71-1.75; Mg = 2.65-2.75; Al = 2.14-2.24) (Fig. 11c), actinolitic hornblende (Ti = 0.007-0.015; Fe = 1.15-1.33; Mg = 3.40-3.77; Al = 0.07-0.92) (Fig. 11a), are related to their different structural sites.

In the *partly re-equilibrated amphibolite* (Tab. XB; Fig. 11d) the relict brown amphibole is a Mg-hornblende (Leake, 1978) (Ti = 0.005-0.17; Fe = 1.41-1.72; Mg = 2.85-3.27; Al = 1.73-1.97), whereas the newly-formed bluish-green needles have an actinolitic hornblende composition (Ti = 0.032-0.035; Fe = 1.21-1.28; Mg = 2.56-2.60; Al = 0.94-0.98).

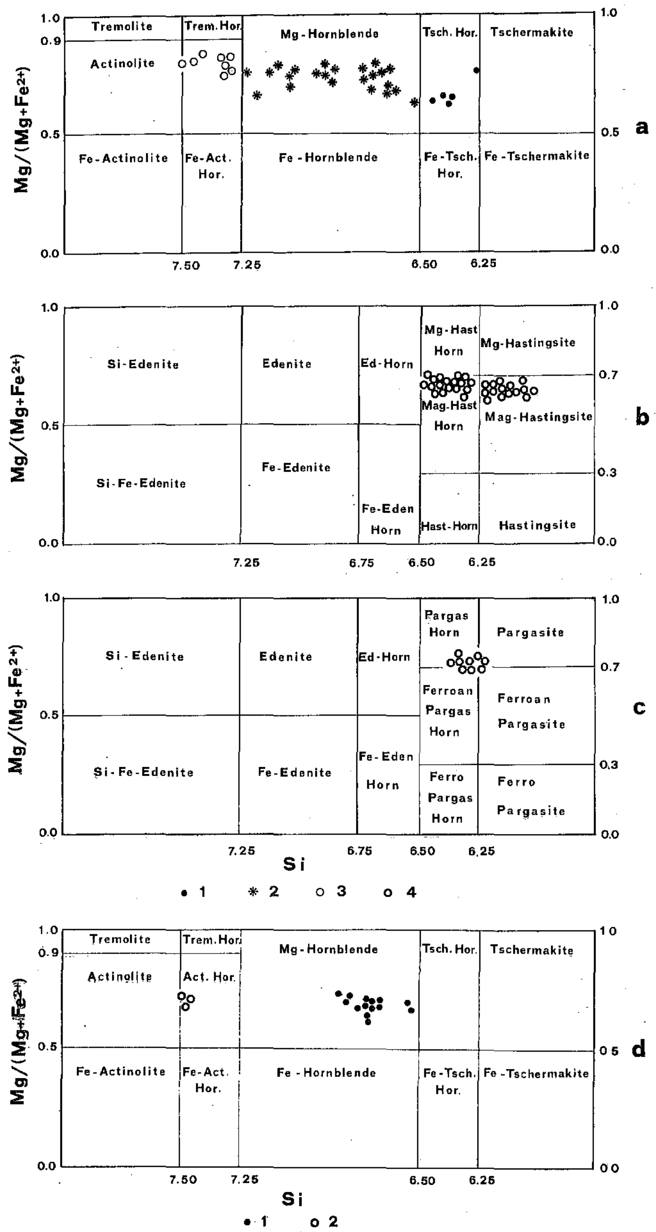


Fig. 11 - Si vs. $Mg/(Mg + Fe^{2+})$ classification diagrams (Leake, 1978) of **relict (Amph. I) and neoblastic (Amph. II, III) amphibole** from:
a, b and c) *Variscan meta-hornblendite extensively re-equilibrated to Alpine* amphibole II ± garnet II ± chlorite ± biotite II *meta-hornblendite*.
— **Relict amphibole**: 1 = structurally preserved pale-green porpyroclasts **(Amph I). Neoblastic amphibole**: 2 = pale bluish-green amphibole recrystallized with subgranular texture **(Amph II)**; 3 = small bluish- green crystals **(Amph II)**, overgrowing the large amphibole clasts, developing along shear planes (Da1); 4 = bluish-green crystals **(Amph III)** growing along shear planes (Da2).
d) *Variscan amphibolite partly re-equilibrated to Alpine* amphibole II + garnet II + white mica II + chlorite + biotite II amphibolite.
— **Relict amphibole**: 1 = brown-green crystals **(Amph I). Neoblastic amphibole**: 2 = very small pale-green needles **(Amph II)** overgrowing the amphibole I rims.

In Laird & Albee's (1981) diagrams the data points of all the newly-formed crystals analysed plot in correspondence with an intermediate pressure metamorphic

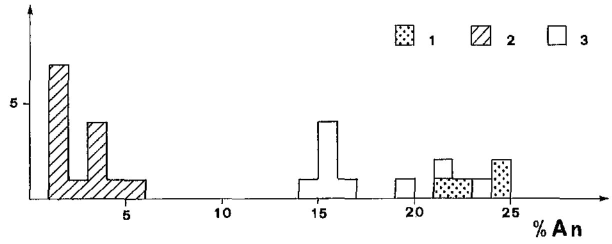


Fig. 12 - Histogram showing the An content of the **relict (Pl I) and neoblastic (Pl II and III) plagioclase** from Variscan garnet ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± *K-feldspar gneissic micaschist.*
— **Relict plagioclase**: 1 = relict crystals **(Pl I)**; **Neoblastic plagioclase**: 2 = small grains **(Pl II)**; 3 = rims overgrowing Pl II and homogeneous crystals **(Pl III)**.

series, which goes from the staurolite to garnet+albite zones.

5.5 *Plagioclase*

In the *extensively re-equilibrated gneissic micaschist* and *leuco-orthogneiss* (Tab. XI; Fig. 12) the plagioclase in small grains crystallized during the first Alpine event, is albite (Pl II with An 1 to 6), whereas the rims or single grains, grown during the second one are oligoclase (Pl III with An 14 to 24). The rare relict crystals show An 21 to 24 in the *gneissic micaschist*, and An 8 to 15 in the *leuco-orthogneiss.*

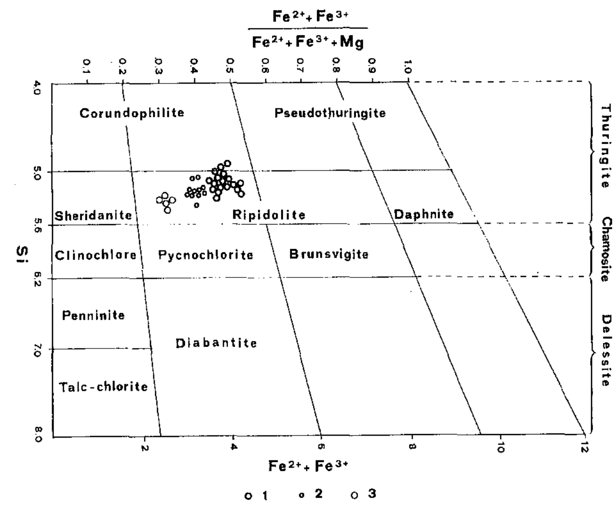


Fig. 13 - Hey's (1954) classification diagram of ***neoblastic chlorite*** from: 1 = *Variscan* sillimanite ± garnet *gneissic micaschist, micaschist and fels partly re-equilibrated to Alpine* kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II *micaschist*; 2 = *Variscan garnet* ± staurolite ± andalusite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K- feldspar *gneissic micaschist*; 3 = *Variscan meta-hornblendite extensively re-equilibrated to Alpine* amphibole II ± garnet II ± chlorite ± biotite II *meta-hornblendite.*

	a		b			c	d
	B 2164		B 1914			B 1957	B 1918
	Ac	Ar	Ar	Ac	Ar	Vr	Vr
SiO_2	69.10	62.72	66.00	66.80	66.16	63.41	63.79
TiO_2	0.01	0.00	0.00	0.02	0.00	0.01	0.00
Al_2O_3	20.19	23.60	21.70	21.08	21.27	22.48	22.85
Cr_2O_3	0.00	0.03	0.00	0.00	0.00	0.00	0.00
FeO	0.03	0.12	0.00	0.03	0.00	0.00	0.01
MnO	0.00	0.03	0.01	0.00	0.00	0.00	0.01
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.71	4.56	2.45	1.53	1.79	3.95	3.79
Na_2O	10.93	8.79	9.91	10.23	10.31	9.30	9.13
K_2O	0.04	0.10	0.10	0.21	0.11	0.07	0.27
Total	101.01	99.95	100.17	99.90	99.64	99.22	99.85
Cations per 8 oxygens							
Si	3.006	2.781	2.906	2.944	2.920	2.824	2.826
Al	1.035	1.234	1.126	1.095	1.107	1.180	1.194
Fe^{2+}	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Cr	0.000	0.001	0.000	0.000	0.000	0.0()0	0.000
Mn	0.000	0.001	0.000	0.000	0.000	0.000	0.000
Na	0.922	0.756	0.846	0.874	0.882	0.803	0.784
Ca	0.033	0.217	0.116	0.072	0.085	0.188	0.180
K	0.002	0.006	0.006	0.012	0.006	0.004	0.015
An	3.46	22.15	11.95	7.54	8.70	18.93	18.35
Ab	96.31	77.27	87.47	91.23	90.67	80.67	80.07
Or	0.23	0.58	0.58	1.23	0.63	0.40	1.58

Extensively (a) and (b) and partly (c) re-equilibrated rock types; see list of samples analysed.
Vr = Variscan relicts; Ac = Alpine cores; Ar = Alpine rims.

Tab. XI - Representative microprobe analyses of plagioclase

In the *partly re-equilibrated* rocks, plagioclase exhibits a wide compositional range in accordance with petrographic observations, which show an incomplete recrystallization of the original plagioclase into zoisite + a more albitic end-member.

In the *amphibolite*, the plagioclase ranges from An 75 to An 25. In the *micaschist*, Variscan plagioclase is commonly pseudomorphed by paragonitic sericite, the newly-formed grains cluster around An 5. Relict crystals with An 28 to 14 are also present. In the *orthogneiss*, plagioclase ranges from An 18 in the Variscan magmatic relict to An 5 in the Alpine grains. Very rare oligoclase rims (An 18) are also present.

5.6 *K-Feldspar*

In the *extensively re-equilibrated gneissic micaschist* the newly-formed K-feldspar exhibits a significant Ab substitution (Ab 22.9-29.2) and a low Ca content (An 0.8-1.4).

In the *partly re-equilibrated micaschist* the relict potassium feldspar shows very low albite substitution (Ab 1.0) and practically no Ca (An 0.1).

In the *extensively and partly re-equilibrated orthogneiss* the neoblastic crystals show Or = 93.0-94.7 with An = 5.2-7.0 and Or = 88.9-91.4 with An = 7.9-10.3, respectively.

5.7 *Chlorite*

All the newly-formed analysed crystals (Tab. XII) from the *extensively* and *partly re-equilibrated micaschist* and from the *meta-hornblendite*, show relatively Al-rich and Fe-poor ripidolitic compositions (Fig. 13). They form three clusters, which show very small differences in Si content and Fe/Mg ratio, related to the bulk rock chemistry.

The chemistry of the above Alpine chlorite is typical

	a	b	c
	B 1827	B 1748	B1907
	o	O	O
SiO_2	25.29	26.37	24.00
TiO_2	0.10	0.03	0.12
Al_2O_3	22.22	21.22	22.07
Cr_2O_3	0.08	0.03	0.00
FeO	21.32	17.70	26.39
MnO	0.14	0.16	0.08
MgO	17.85	20.84	13.89
CaO	0.04	0.00	0.00
Na_2O	0.00	0.00	0.00
K_2O	0.00	0.00	0.00
Total	87.04	86.35	86.55
Cations per 12 oxygens			
Si	2.628	2.707	2.582
Al^{IV}	1.372	1.293	1.418
	4.000	4.000	4.000
Al^{VI}	1.351	1.274	1.381
Ti	0.008	0.002	0.010
Cr	0.007	0.002	0.000
Fe^{3+}	0.000	0.012	0.017
Fe^{2+}	1.853	1.507	2.357
Mn	0.012	0.014	0.007
Mg	2.765	3.188	2.227
Ca	0.004	0.000	0.000
Na	0.000	0.000	0.000
K	0.000	0.000	0.000
	6.000	5.999	5.999

Extensively (a) and (b) and partly (c) re-equilibrated rock types; see list of samples analysed. Symbols as in Fig. 13. Structural formulae calculated following Laird & Albee's (1981) method.

Tab. XII - Representative microprobe analyses of chlorite

of Laird & Albee's (1981) intermediate grade of metamorphism.

5.8 *Other minerals*

Chloritoid - In the *partly re-equilibrated micaschist,* chloritoid porphyroblasts show the Mg/Fe ratio ranging from 0.23 to 0.35.

Epidote - In the *extensively re-equilibrated gneissic micaschist,* epidote generally shows compositions halfway between Al- and Fe-epidote, and rims richer in Fe content; Al-clinozoisite and Al-zoisite grains are also present.

In the *extensively re-equilibrated leuco-orthogneiss,* the Alpine epidote is Fe-clinozoisite and Fe-zoisite with the Fe content increasing from core to rim.

Ilmenite - The varying Mn/Fe ratio depends on the structural site and on the original rock chemistry. It ranges from 0.054 to 0.060 in the *extensively re-equilibrated gneissic micaschist,* where the newly-formed ilmenite is widespread; from 0.024 to 0.029 in the *partly re-equilibrated micaschist,* where both intergrown ilmenite and white mica II developed at the expense of biotite; from 0.082 to 0.087 in the *partly re-equilibrated amphibolite,* where neoblastic ilmenite developed after Variscan amphibole.

Tourmaline - In the *extensively re-equilibrated leuco-orthogneiss,* microprobe analyses confirm the existence of two different tourmaline generations. The neoblastic rim shows Mg/Fe ratio lower than the relict core.

Staurolite - In the *extensively re-equilibrated gneissic micaschist* the relict grains are very rich in the Fe-end member with Fe/(Fe+Mg) ratio of 0.635.

6. P-T Conditions

The chemical compositions of the Variscan relicts and the Alpine re-equilibrated mineral phases indicate that the Alpine overprint occurred at T lower than the Variscan metamorphism. The occurrence among the newly-formed minerals of kyanite and/or phengite and/or Ti-poor hornblende (tsch-Mg-parg-hast-hornblende) and/or zoisite indicate relatively high pressure conditions; the presence of Fe-Ca-rich almandine + albite/oligoclase + Mg-Fe-ripidolite, on the other hand, indicates low to intermediate temperatures.

The chemical compositions of the Variscan relicts and of the newly-formed Alpine mineral phases which crystallized from *partly re-equilibrated micaschist* during the first Alpine event, have been projected from K-feldspar in A-F-M diagrams (Thompson, 1957; Fig. 14a and b), together with the bulk rock chemistry.

In spite of the petrographic evidence according to which **high-grade sillimanite + Mg-Mn-almandine + Mg-Fe-biotite** constitute a **Variscan A-F-M paragenesis,** the rock composition plots, approximately, on the sillimanite-biotite tie-line (Fig. 14a). This is probably due to the shift towards FeO-rich compositions of the seemingly preserved relict biotite, which, although preserving its original shape and textural position, has been partly rehomogenized. **The Alpine parageneses of the first event** may have consisted of either **kyanite + Mg-Fe-ripidolite + Fe-Mg-chloritoid** or **Fe-Ca-almandine + Mg-Fe-ripidolite + Fe-Mg-chloritoid,** (both of which are accompanied by phengite + quartz) (Fig. 14b). Only in the first paragenesis does the whole rock plot inside the tie-lines. Petrographic evidence, however, indicates that kyanite pseudomorphed sillimanite, whereas garnet, chlorite and chloritoid grew as porphyroblasts at the same time. Both mineral assemblages are typical of the garnet zone of the greenschist-facies.

The A-F-M diagram cannot be used to discuss the Variscan and the Alpine parageneses of the *extensively re-*

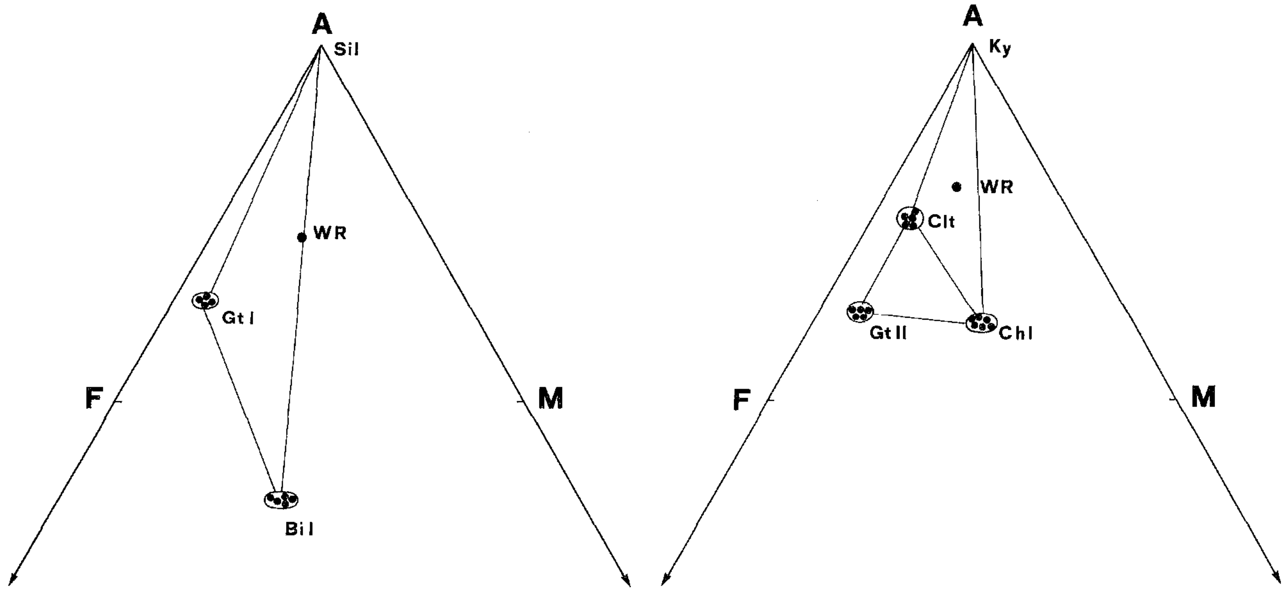


Fig. 14 - A-F-M diagram (Thompson, 1957) showing the **paragenetic mineral phases**: a) sillimanite + garnet I + biotite I of the **Variscan amphibolite-facies event** in the *Variscan* sillimanite ± garnet gneiss, *micaschist and fels partly re-equilibrated to Alpine* kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II *micaschist.* b) kyanite + garnet II + chlorite and garnet II + chlorite + chloritoid of the **first Alpine event** in the Variscan sillimanite ± garnet *gneiss, micaschist* and *fels partly re-equilibrated to Alpine* kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II *micaschist.*

equilibrated gneissic micaschist. In fact, the Variscan biotite was completely rehomogenized during the last two Alpine events, and the first Alpine paragenesis is characterized by the lack of Al-rich mineral phases such as kyanite and chloritoid. The Na_2O-(AF_2O_3)-CaO-FMO tetrahedron is more suitable for projecting the **Fe-Ca-almandine + albite + Tsch-Mg-hornblende** (+ phengite + epidote + quartz + Mg-Fe-ripidolite) **paragenesis of the first Alpine event** in the *extensively re-equilibrated gneissic micaschist* (Fig. 15).

The different Alpine parageneses of the second event, shown by the different rock types studied (Tabs. I to VI), cannot be discussed using the chemiographic diagrams.

To better clarify the P-T conditions of the Alpine events, suitable geothermobarometers were used.

Plyushina's (1982) plagioclase-amphibole geothermobarometer gave T = 460-470° C and P = 7-8 kbar for the first overprinting event in the *extensively re-equilibrated gneissic micaschist.*

Perchuck's et al. (1985) geothermometers, based on partition of Mg between garnet and amphibole and of Ca between plagioclase and amphibole, for Gt II - Amph II and Pl II - Amph II pairs, respectively, gave temperatures ranging from 400 to 460 and 400 to 500°C for the first overprinting event in the *extensively re-equilibrated gneissic micaschist,* and from 480 to 500°C in the *partly re-equilibrated orthogneiss.*

The phengite geobarometer (Massonne & Schreyer, 1987), when applied to neoblastic white micas of different textural sites, gave, in both overprinting events, for T = 500°C, minimum pressures ranging from 3.6 to 4.5 kbar for the *partly re-equilibrated micaschist,* and from 4 to 6 kbar for the *extensively re-equilibrated gneissic micaschist* and *leuco-orthogneiss.*

A more restricted range of P-T conditions can be inferred from mineral compatibilities and stability fields.

— The first overprinting event.

In the area of strong re-equilibration both **i)** Fe-Ca-almandine + albite + Al-Fe-epidote + tscher-Mg-hornblende + Mg-Fe-ripidolite assemblage and **ii)** kyanite + Fe-Ca-almandine + Mg-Fe-ripidolite + Fe-Mg-chloritoid assemblage indicate **P > 4 kbar and T ranging from 460 to 540° C**. The lowermost baric limit, conditioned by the presence of kyanite and by the 3.2 Si value in phengite (Massonne & Schreyer, 1987), is 5-6 kbar. The upper thermal limit is constrained by the absence of neoblastic staurolite and/or oligoclase. In fact, at P = 4 kbar and at T > 540°C, in the rocks with the mineral assemblage **i)**, an An-richer plagioclase (An > 17) should coexist with hornblende (Winkler, 1976); in pelitic rocks with the mineral assemblage ii), Mg-Fe-chlorite and chloritoid should disappear with the appearance of staurolite, when muscovite and quartz are present (staurolite-in isograd band in Winkler, 1976). There are no features suggestive of the upper baric limit, but it clearly falls below Maresch's (1977) upper stability limit of glaucophane. The lower thermic limit is defined by the presence of the hornblende + An<17 plagioclase + epidote assemblage and by the presence of almandine-rich garnet. The almandine + chlorite + white mica assemblage, typical of the garnet zone (low-grade almandine), remains stable at the beginning of medium-grade metamorphism until it

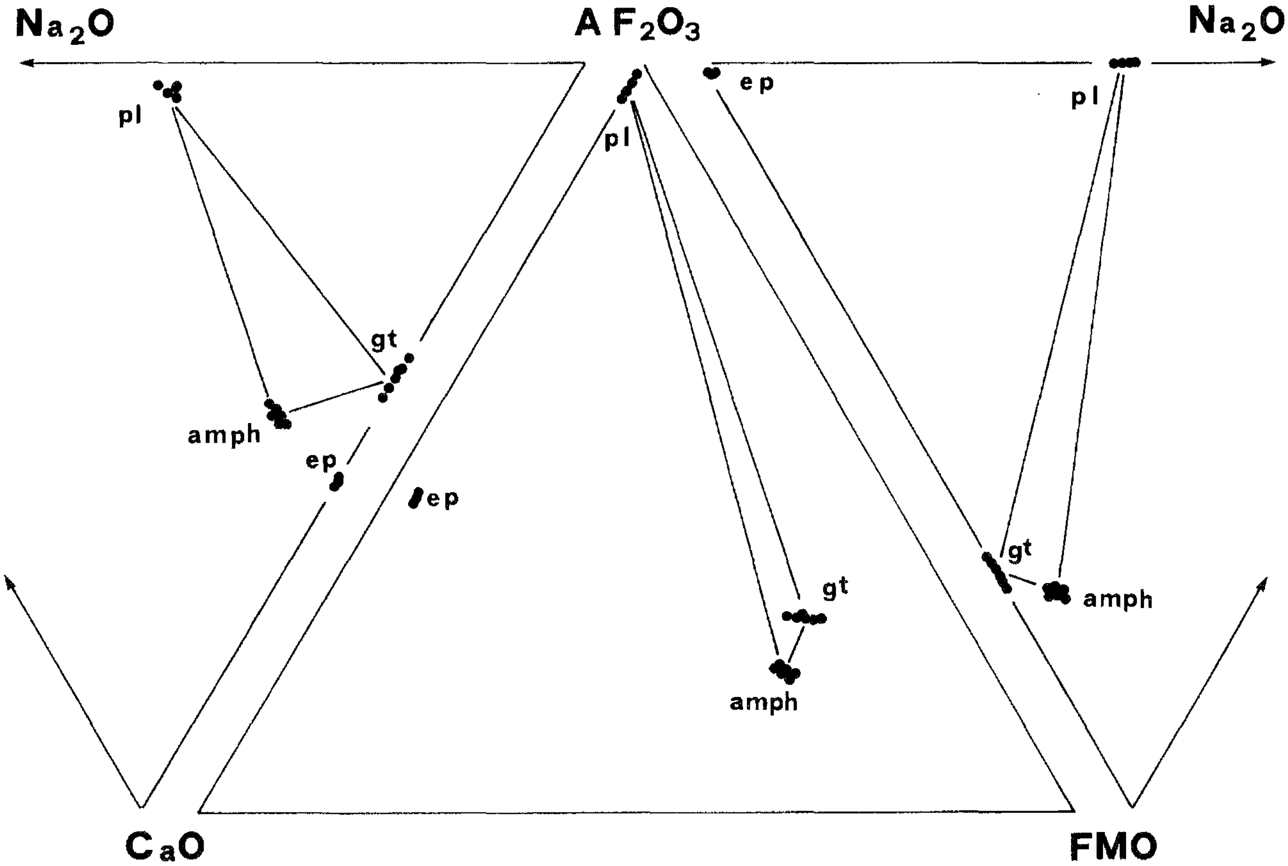


Fig. 15 - Na_2O-AF_2O_3(Al_2O_3+Fe_2O_3)-CaO-FMO(FeO+MgO+MnO) tetrahedron (Laird, 1980) showing the **paragenetic mineral phases** - albite + epidote + amphibole + garnet II - of the **first Alpine event** in the *Variscan* garnet ± staurolite *gneissic micaschist extensively re-equilibrated to Alpine* two mica + garnet II + albite/oligoclase ± amphibole ± K-feldspar *gneissic micaschist.*

breaks down to staurolite + biotite. The appearance of almandine-rich garnet does not indicate very precisely P-T conditions, but gives a wide range of pressures in the high temperature low-grade metamorphism. The absence of biotite in the first Alpine event could be due to both the relatively high pressure conditions which favoured the growth of garnet, and to the presence of chloritoid, which precludes biotite when associated with almandine + chlorite.

Upper greenschist-facies, with P = 6 ± 2 kbar and T = 500 ± 30° C, equivalent to those of the Barrovian metamorphism garnet zone, can therefore be inferred for the first Alpine overprinting event in the Aspromonte Unit (Fig. 16).

In the area of weaker re-equilibration the same overprinting event must have developed under lower P-T conditions, as suggested by the widespread presence of sericite, Fe-rich epidote, blue-green amphibole and rare garnet II.

— The second overprinting event

In the area of stronger re-equilibration the appearance of An 14-24 plagioclase and the disappearance of chloritoid, in the metapelitic rocks, indicate temperatures higher than that of the oligoclase-in isograd, thus T > 550°C. The lack of kyanite and/or of garnet, on the other hand, suggests pressures lower than those of the first overprinting event.

In the area of weaker re-equilibration, the metamorphic grade of the second event also gradually decreases, as suggested by the presence of the bluish-green to green biotite and of sericite.

7. Discussion and Conclusions

The chemistry of the mineral phases taken from both extensively and partly re-equilibrated rocks, belonging to the strongly overprinted area of the Aspromonte Unit, shows differences between the Variscan relict and the newly-formed Alpine minerals. When compared with the former, the latter show:

— a lower plagioclase An content;

— a lower Mg content in the biotite;

— a Mn decrease and a Fe and Ca increase in the garnet.

The mineral chemistry also shows newly-formed phases such as K- and Al-rich amphibole, phengite and paragonite, Mg-Fe-chloritoid and Mg-Fe- chlorite.

Since differences in whole rock chemistry do not exist between the non or little cataclastic and the partly re-equilibrated micaschists, both of which outcrop at centimeter distances, an Alpine isochemical metamorphism can be inferred for the partly re-equilibrated rock types.

In conclusion, our new field, petrographic and mineral chemistry data in the Aspromonte Unit confirm the presence of a single Alpine tectonic unit overprinted by two different polyphase tectono-metamorphic events: the older one, Variscan in age, is characterized by polyphase low-P amphibolite-facies conditions, and the younger one, Alpine in age, is characterized by a greenschist-facies metamorphism, evolving from medium- to relatively low-P conditions, though with different intensity. Anyhow, all over the extensively re-equilibrated area, remnants of pre-Alpine amphibolite-facies minerals, such as garnet, oligoclase, hornblende, staurolite or andalusite pseudomorphed by white mica, are also widespread. This polymetamorphic evidence is also supported by the bimodal distribution of mineral chemistry (see chapter 5). The Alpine age of these rocks, indicated by a preliminary radiometric study (Bonardi et al., 1987), is confirmed by Messina et al. (1991): Rb/Sr data on biotite and muscovite separates, indicate that the Variscan metamorphism peaked at, or before, 314 Ma, whereas the Alpine overprint developed at about 28-24 Ma.

According to the above results, there is no petrographic and mineralogical evidence to support the existence of two Alpine units of different grades inside Bonardi's et al. (1979; 1982) Aspromonte Unit as suggested by Pezzino & Puglisi (1980) and Pezzino et al. (1990). In fact, the greenschist-facies parageneses of Pezzino & Puglisi's (1980) lower unit in the northeastern area of the Aspromonte Unit (see chapter 2), have to be considered the result of a retrogressive Alpine re-equilibration of Variscan amphibolite-facies rocks. The mineral chemistry data of this unit in the Delianuova and Polsi zones (Ioppolo et al., 1983b; Ioppolo & Puglisi, 1986-87b; Pezzino et al., 1990) are quite similar to the data presented in this study relative to the extensively re-equilibrated areas, (from S. Cristina-Delianuova to Polsi-S. Luca; Fig. 1), but they have been misinterpreted. In fact, lacking identification of Variscan relict mineral phases, led the above authors to interpret the metamorphism of the lower unit as a prograde garnet zone medium-grade metamorphism. The non-recognition of the two Alpine events, subsequent to a pre-Alpine metamorphism, also led the above authors to apply geothermobarometers to non stable mineral pairs with consequently unreliable P-T results. Furthermore, the mylonitic rocks characterized by chloritoid, Mg-Fe-chlorite, epidote and staurolite, which according to Ioppolo et al. (1983a) mark the tectonic contact between Pezzino & Puglisi's (1980) upper and lower units, in the present study are interpreted as Variscan felses partly re-equilibrated to Alpine schists, very widespread in the Alpine strongly overprinted area and not localized in a single well defined horizon. These rocks locally also exhibit Alpine kyanite pseudomorphing sillimanite. Neoblastic staurolite has never been recognized; its presence, however, would not modify the P-T conditions defined for the first Alpine tectono-metamorphic event.

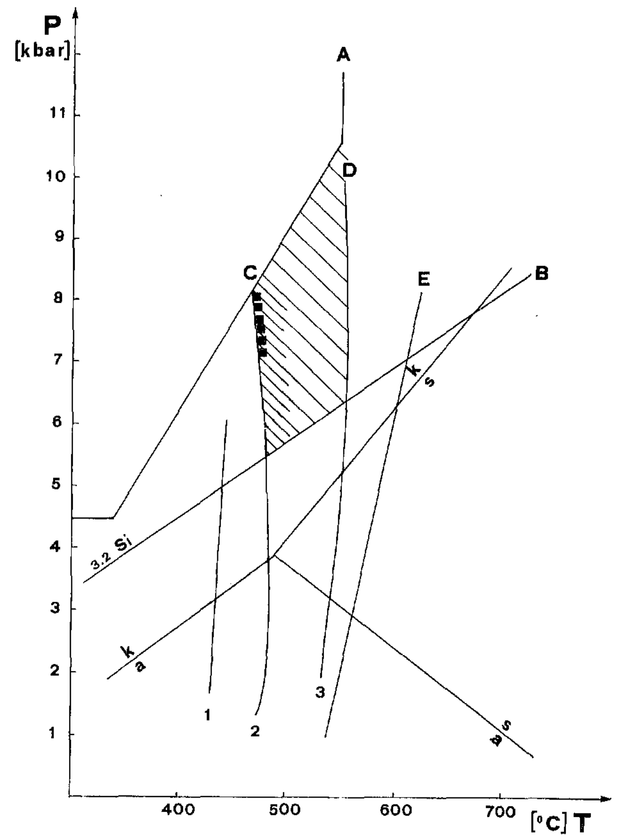


Fig. 16 - Inferred P-T conditions for the first Alpine overprinting event in the strongly re-equilibrated rocks from the Aspromonte Unit. Squares: P-T values for the extensively re-equilibrated gneissic micaschist (geothermobarometer of Plyushina, 1982). 1: chloritoid-in (pyrophyllite + Fe-rich chlorite = chloritoid + quartz + H_2O; Frey, 1972). 2 = almandine-in (Fe-chlorite + quartz = almandine-rich garnet; Hsu, 1968). 3 = staurolite-in (chloritoid + kyanite = staurolite + quartz; Hoschek, 1967). A = Maresch's (1977) stability limit of natural glaucophane. B = Si values for the muscovite phengite series (Massonne & Schreyer, 1987). C = hornblende + albite + epidote-in isograd (Winkler, 1976). D = oligoclase-in isograd (Winkler, 1976). E = hornblende + An >17 plagioclase-in isograd (Winkler, 1976). Stability fields of the Al_2SiO_5 polymorphs after Holdaway (1971): k = kyanite, s = sillimanite, a = andalusite.

The Aspromonte Unit exhibits the same Alpine tectono-metamorphic evolution in both Calabria and Sicily. In the overprinted area of the Peloritani Mountains, however, the extensively re-equilibrated rocks also exhibit kyanite, in addition to the same mineral assemblages which characterize the Calabrian extensively re-equilibrated gneissic micaschists. This suggests that also the upper amphibolite-facies Variscan sillimanite + garnet gneiss and micaschist were extensively and pervasively Alpine overprinted in a portion of the Aspromonte Unit.

The presence of a strong Alpine overprint in the Aspromonte Unit emphasizes the importance of the Alpine orogeny in the Southern Sector of the Calabrian-Peloritan Arc.

Appendix

List of samples analysed:

— **B 1827 and 2164**: Variscan garnet ± staurolite ± andalusite gneissic micaschists **extensively re-equilibrated** to Alpine two mica + garnet II + albite/oligoclase + amphibole + K-feldspar **gneissic micaschists.**

B 1827 collected at approximately 1 km from Polsi, along the Gambarie - Polsi road.

B 2164 collected at S. Cristina d'Aspromonte, km 45.900, S.S. (National road) 112.

— **B 1907 and 1957**: Variscan sillimanite ± garnet micaschist and gneissic micaschist, respectively, **partly re-equilibrated** to Alpine kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II **micaschists.**

B 1907 and 1957 collected at Acqua a Vace, S.S. 183.

— **B 935 and 1905**: Variscan sillimanite ± garnet felses **partly re-equilibrated** to Alpine kyanite + garnet II + white mica II ± chloritoid ± chlorite + biotite II **micaschists.**

B 935 collected at Croce di Romeo, S.S. 183.

B 1905 collected at Acqua a Vace, S.S. 183.

— **B 1914 and 2122A**: late-Variscan peraluminous plutonites **extensively re-equilibrated** to Alpine garnet II + white mica II + biotite II + chlorite **leuco-orthogneisses.**

B 1914 collected at S. Cristina d'Aspromonte, km 45.000, S.S. 112.

B 2122 collected between Piminoro and Platì, km 57.700, S.S. 112.

— **B 1918 and 1919**: late-Variscan peraluminous plutonites **partly re-equilibrated** to Alpine kyanite + garnet II + white mica II + amphibole + biotite II + chlorite **orthogneiss.**

B 1918 and 1919 collected at S. Cristina d'Aspromonte, km 45.900, S.S. 112.

— **B 1747 and 1748**: Variscan meta-hornblendite **extensively re-equilibrated** to Alpine amphibole II ± garnet II ± chlorite ± biotite II **meta-hornblendite.**

B 1747 and 1748 collected at km 52.000, S.S. 112.

— **B 1922 and 2167** Variscan amphibolite **partly re-equilibrated** to amphibole II + garnet II + white mica II + chlorite + biotite II **amphibolite.**

B 1922 collected along the Montalto-Croce di Romeo road, about 1 km before the junction with the SS. 183.

B 2167 collected at about 2 km before Platí, S.S. 112.

Abbreviation list:

ab = albite, amph = amphibole, and = andalusite, ande = andesine, bi = biotite, byt = bytownite, cc = carbonates, chl = chlorite, clt = chloritoid, cord = cordierite, cumm = cummingtonite, di = diopside, ep = epidote, gr-bl = green-blue, gt = garnet, hor = hornblende, Kf = K-feldspar, ky = kyanite, labr = labradorite, ms = muscovite, olig = oligoclase, phl = phlogopite, pl = plagioclase, qtz = quartz, ser = sericite, sil = sillimanite, staur = staurolite, wm = white mica.

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NATURE AND EVOLUTION OF CRYSTALLINE BASEMENT BENEATH THE HYBLEAN PLATEAU (SICILY) INFERRED FROM XENOLITHS IN VOLCANIC ROCKS

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Abstract

A xenolith suite recently found in some volcanic rocks from Hyblean Plateau (Southern Sicily) has provided insights into the nature of the crystalline basement. It is covered by a Permo-Caenozoic sedimentary and volcanic sequence whose thickness is estimated about 7 Km. Mafic granulites form the dominant lithology of the above basement. They show granoblastic polygonal texture (both coarse and fine-grained). They are formed of plagioclase feldspar, coexisting Ca-poor and calcic pyroxenes, Al-spinel ± accessories. Mineral and whole-rock chemistry indicate basaltic parentage for Hyblean granulites. They were equilibrated in the pressure - temperature range of the spinel-gabbro field (CMAS system) in most cases. Some coarse-grained lithotypes show complex and abundant reaction textures between minerals, recording a relative rapid uplifting of the Plateau. More rare nodules of anorthosites, quite similar to those from world-wide anorogenic massifs, also occur in the Hyblean volcanics.

All above-mentioned nodule suite is consistent with a lower continental-crust origin. Extremely rare granitoid microxenoliths and no nodules of greenshist and amphibolite facies rocks have been found in the Hyblean volcanics. This fact suggests that the middle and upper part of the basement have been severely thinned either by long-standing exposure and erosion or, more likely, laminated by "magmatic underplating" during continental rifting episodes.

Key words: *Sicily, volcanic rocks, xenoliths, mafic granulites, anorthosites.*

1. Introduction

Deep-seated xenoliths from volcanic rocks may provide useful information on the lithology of inaccessible crust and mantle, enhancing geophysical data in most cases. The state of art on *crustal* xenoliths from the Hyblean Plateau is reported here, attempting to give some contributions to the knowledge of the crystalline basement of southernmost part of Italy.

The Hyblean Plateau represents the emerged northern part of the Malta Platform which belongs to the African Plate. The Plateau consists of a platform-type sedimentary sequence with intercalated volcanic horizons. The age of the exposed terrains ranges from Upper Cretaceous to Lower Pleistocene. Oil drills have reached Upper Triassic limestones and volcanics about 5 Km below the sea level. An additional 2 Km thick succession of Triassic carbonates and Permo-Triassic clastic sediments is deduced from geophysical data. All 7 Km thick sedimentary and volcanic sequence lays upon a strongly magnetic crystalline basement (Bianchi et al., 1987). The main tectonic features of Hyblean Plateau consist of two fault systems. The first trends NE-SW, leading to the formation of a structural depression (The Gela-Catania Forethrough) at the northern and western margins of the Plateau. The second fault system trends NNW-SSE. It truncates the Plateau along the Ionian coast (Ragusa-Malta Escarpment) separating the shelf zone from the deep Ionian basin. (Carbone et al., 1982, 1985).

Submarine and subaerial alkali-basalt lava flows of Upper Cretaceous age occur in the southernmost zone of the Plateau (Capo Passero and Pachino areas) (Amore et al., 1988). Volcanics of Miocene age, outcropping in the central-eastern zone, consist of alkaline-mafic diatremes and associated pyroclastic-flow deposits, with subordinate lava flows. Plio-Pleistocenic alkaline and tholeiite basaltic flows, related to a fissural-type activity, outcrop in the northeastern part of the Plateau (Fig. 1).

A complex suite of deep-seated nodules was recently found in some Miocenic diatremes and Quaternary alkalibasalt lava flows: mantle derived xenoliths consist of spinel-bearing peridotites (mostly harzburgites) and several types of spinel-bearing websterites. Feldspar-bearing xenoliths, though less abundant than the formers also occur. They consist of mafic granulites, subordinate anorthosites and very rare granitoids (Scribano 1987). They conform the minimum criteria for the recognition of lower crustal origin (Griffin & O'Reilly, 1987): 1) similar

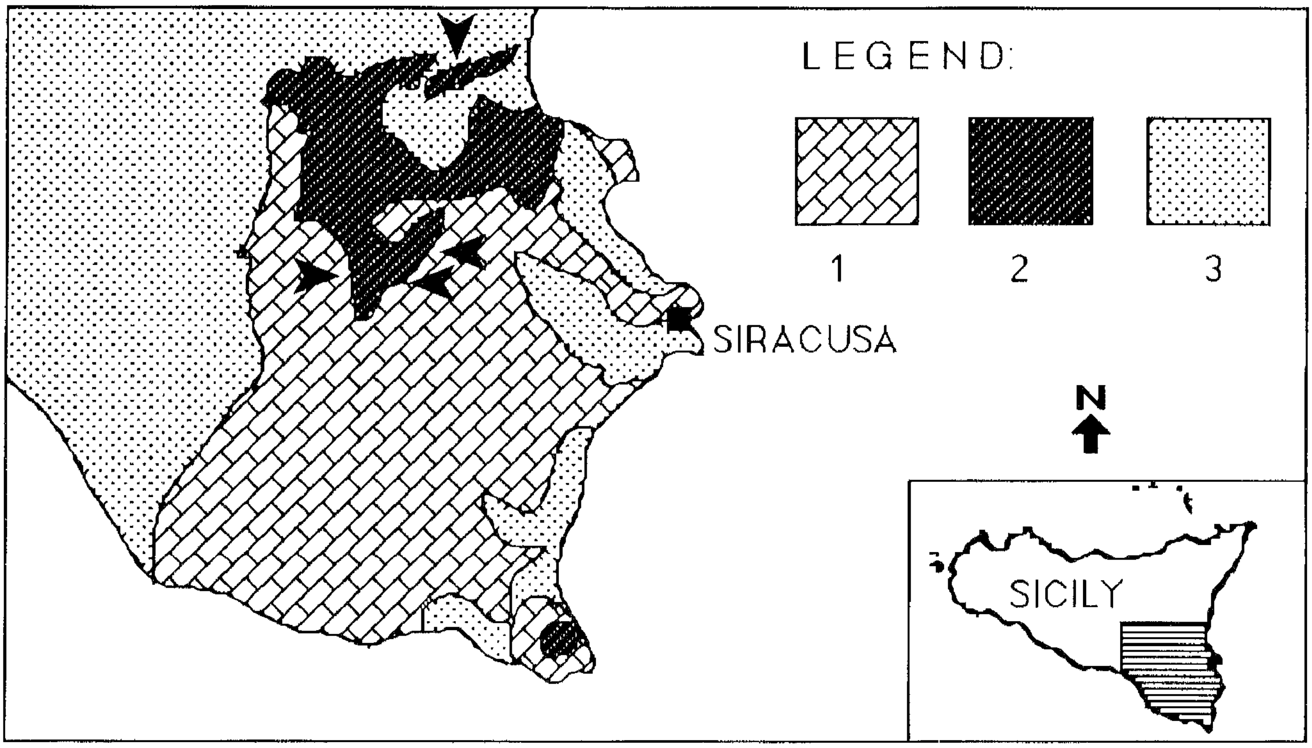

Fig. 1 - Geological sketch map of Southeastern Sicily (following Carbone et al., 1985). Explanation to the legend: 1) Meso-Caenozoic sedimentary rocks (mostly platform-type carbonates). 2) Volcanic rocks. 3) Quaternary nappes and alluvional cover. Arrows point to the xenolith-rich outcrops.

rock types do not occur as adjacent outcrops. 2) Pressure-Temperature estimates are consistent with lower crust conditions (Scribano, 1988) and available geophysical data. Though mantle-xenoliths occur in both diatremes and lava flows, crustal xenoliths are more frequent into the former in most cases.This fact may depend on the eruptive style, since the "fluidization" process, invoked for diatreme emplacement (Cloos, 1941), implies elevated ascenting velocity and relative low eruptive temperature (Clement, 1975; Dawson, 1980). It prevents deep-seated xenoliths from significant interaction with the host magma.

2. Two-Pyroxene Granulite Nodules

Hyblean mafic granulites show fine- (0.3-1 mm) to coarse- (1-3 mm) granoblastic polygonal texture, with plagioclase feldspar, coexisting Ca-poor and Ca-rich pyroxenes, green Al-spinel and accessories (ores, ilmenite, rare amphybole) (Fig. 2). Trails of very fine fluid inclusions, consisting of almost pure CO_2 (Clocchiatti & Wilmarth, 1992, personal comm.), are ubiquitous in the above-mentioned minerals.

Plagioclase grains show 120º junctions. They are either untwinned or poorly twinned, according to the pericline law in most cases (Fig. 5). Undulose extinction occurs in some coarse grains. Pyroxenes show both curvilinear and straight grain boundaries. Exsolute blades of spinel and Ca-poor from Ca-rich pyroxene are ubiquitous. Dark- green, vermiform pleonaste occur within Ca-rich and Ca-poor pyroxene grains. More rarely it forms discrete grains between the coarser silicates.

The modal percentage of plagioclase with respect to the pyroxenes vary from nodule to nodule, though they show intermediate to elevated colour indexes. (C. I. = 45 - 90). Gneissic banding is quite common in fine-grained types. The coarse types show characteristic alternating and partially juxtaposing lenses of pyroxenes (+ spinel) and plagioclase ("lenticular banding") in some cases (Fig. 2 c). Modal variations in nodules may reflect composition unhomogenities in their source geological bodies (e.g. banding), though they may depend on the size of the nodule only, for the coarse-grained types.

Summary of mineral chemistry of Hyblean mafic granulites is reported in Tables 2-5. Each mineral type shows relative large composition interval, though above variations are very minor, even absent, in a given specimen. Whole-nodule chemistry reflects the modal and mineral-chemistry variations (Tab. 1). It must be remarked the very low content of relatively incompatible elements (K_2O, P_2O_5, TiO_2) in all cases (Scribano, 1988).

Reaction textures between minerals are ubiquitous in Hyblean granulite nodules, though extremely abundant in some coarse-grained, banded lithotypes (hence called "coronitic metagabbros"). Reaction textures may be divided into three types : 1) *Coronas of Ca-rich pyroxene.* They consists of 15-30 μ thick collars of clinopyroxene occurring between plagioclase and Ca-poor pyroxene grains. The latter is largely replaced by a reddish mat of serpentine -like minerals and Fe-oxydes micrograins in most cases. The above coronas appear as opacitic rims under optical microscope. 2) *Ca-poor pyroxene and spinel symplectites.* They consist of vermicular intergrowths of Ca-poor pyroxene and pleonaste replacing a previous orthopyroxene grain. 3) *Ca-rich pyroxene and spinel keliphytes.* They form finely branched fibres of spinel and pyroxene between pyroxene and plagioclase grains, protruding towards the latter in form of juxtaposing fans.

All above reaction textures may coexist in some places forming a three-shells corona: relative coarse vermicular orthopyroxene and spinel form the inner shell, coated by fibrous keliphytes and the clinopyroxene collars. Relic core of orthopyroxene or/and its alteration products (see above) may also occur (Fig.8). One or more shells of the above sequence are missing in some cases. In particular, the clinopyroxene collars may appear discontinuous gradually passing to a fine granular intergrowth of clinopyroxene and anorthite-rich plagioclase. Mineral chemistry differences between unreacted and reacted pyroxenes are not elevated, though significant variations in aluminium content: clinopyroxene collars exhibit highest Al content (up to 7 wt%). On the contrary, pyroxenes from symplectites and keliphytes show lower Al content than coexisting unreacted ones. Plagioclase grains show significant Ca-decreasing (from 70 to 20 mol% of anorthite content) towards the above-mentioned clinopyroxene collars. Minor patches of almost pure anorthite have been found between breakdown products of orthopyroxene as shown in Fig. 8 b.

3. Anorthosite Nodules

The most common type of anorthosite xenoliths occurring in the Hyblean volcanics show granoblastic texture, which varies from inequigranular interlobate to equigranular polygonal (Moore, 1970) even at the nodule scale. Both equant and plate grains occur, the former being the dominant type. All plagioclase grains display polisynthetic twinning according to the albite law and a combination of the pericline and albite laws (Fig. 5). Kink bands and other mechanical distortions give the crystal a shadowing extinction in some cases. Micrograins of green spinel are often enclosed into the plagioclase grains. The rounded shape (drop-like in most cases) of above spinels is often obliterate by opacitic blades of alteration products irradiating from the grain-boundaries towards the enclosing feldspar. Very rare clinopyroxene, interstitial to the plagioclase grains also occur in our anorthosites. Plagioclase shows labradorite composition (An $_{63-65}$). Composition zoning occurs only in plagioclase grains hosting spinel: the anorthite content of the former dramatically increases towards the latter, reaching values of about An_{92} nearby the contact surface between the two minerals. On the contrary, spinel does not show any significant composition zoning. It is a green pleonaste, with 58 - 60 wt% in Al_2O_3.

Above-mentioned anorthosite nodules are cross-

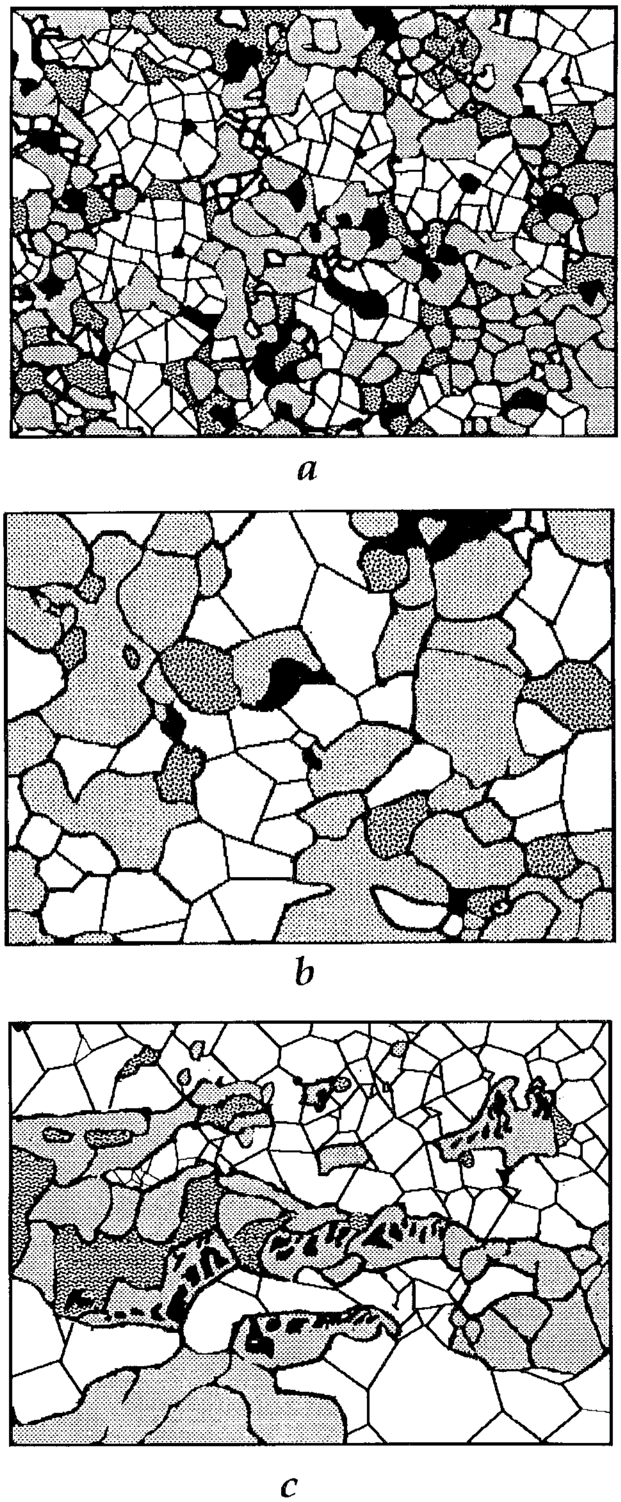

Fig. 2 - Characteristic textures of Hyblean mafic granulite nodules (Draws after microphotograph) *a:* fine-equant type; *b:* coarse-equant type; *c:* coarse type showing "lenticular banding" (for full explanation see the text). Blank: plagioclase, fine stipped: Ca-rich pyroxene, coarse stipped: Ca-poor pyroxene, black: Al- spinel. Base edges of all figures represent 8 mm.

wt%	Fine Type	Coarse-Type	Ultramafic
SiO_2	48.3 -49.7	49.0 -50. 3	50.2 -52
Al_2O_3	14.5 -25.7	16.6 -21.6	6.5 - 8.2
FeOt	3.9 -10.9	4.1 - 6.5	10.6 -12
MgO	4.6 - 9.2	8.5 -11.3	23.3 -25.2
CaO	12.6 -14.5	11.2 -17.3	4.3 - 6.5
Na_2O	1.7 - 2.9	1.4 - 3.3	0.3 - 0.5
K_2O	0.04 - 0.06	0.03- 0.15	0.03
TiO_2	0.3 - 0.6	0.3 - 0.4	0.4
P_2O_5	0.02 - 0.07	0.04- 0.08	0.03

Tab. 1- Ranges of whole-rock chemistry (major elements) of Hyblean mafic granulite nodules. Chemical analyses from Scribano, 1988 and unpublished data.

cutted by ultramafic veins in some cases. Width of the veins varies between 0.5 and 2 cm. Some nodules show several, subparallel websterite veinlets.

The above-mentioned websterite displays well developed polygonal texture, with coarse -equant clinopyroxene grains (75 vol%) with 120º junctions, Ca-poor pyroxene and spinel.

Orthopyroxene occurs as relative small rounded grains, often at the periphery of the coarser clinopyroxene. Ca-poor pyroxene also forms spindle-shaped lamellae within the Ca-rich host. It displays polisynthetic twinning. Relative coarse grains of green spinel occur within the pyroxene grains nearby the boundary surface between the websterite veins and the host rock. In some cases spinel appears as elongated grains between pyroxene and plagioclase, just marking the above-mentioned surface. It must be remarked that an anorthite (An_{90-92}) collar always occurs between pyroxene and spinel. It appears corroded and plunged into patches of calcite, anorthite and sieved clinopyroxene micrograins in some cases.

Wt%	Coarse Type	Fine Type	Ultramafic	WEB
Al_2O_3	6 - 9	4 - 7.5	7 - 7.5	6 -7
FeO tot	4 - 7.5	7.5- 9	4 - 4.5	5.5- 6
MgO	13 - 14	12 -14	13 -14	12 -14
CaO	20 - 22	20 -22	22 -22.5	22 -22.5
Na_2O	0.6- 1.5	0.7- 1	1 - 1.5	0.8- 1
Mg/(Mg+Fe tot)	0.76-0.85	0.71-0.78	0.84-0.86	0.79-0.81
Ca/(Ca+Mg)	0.53-0.55	0.53-0.55	0.54	0.55-0.56

Tab. 2 - Representative mineral chemistry ranges of clinopyroxenes from coarse, fine and ultramafic Hyblean granulites and websterite veins (WEB) into the anorthosites (After Scribano, 1988. WDS microprobe analyses).

Wt%	Coarse Type	Fine Type	Ultramafic	WEB
Al_2O_3	4 - 6.5	2 - 5	4 - 5	4.5- 6
FeO tot	11 -17	19 - 21	12 -13	16.5-17.5
MgO	25 -29	12 -14	28 - 29	24.5-26
CaO	0.3 - 0.6	0.4- 0.7	0.5 - 0.6	0.3- 0.5
Mg/(Mg+Fe tot)	0.68-0.78	0.65-0.70	0.80-0.82	0.72-0.8

Tab.3 - Representative composition intervals of orthopyroxenes from coarse, fine and ultramafic Hyblean granulites and websterite veins into the anorthosites (WEB) (After Scribano, 1988. WDS analyses).

Clinopyroxene is an Al-rich augite, with mg* = 0.79- 0.80; its Al_2O_3 content drops down from about 7 wt% to 4 wt% (ca.) in zones nearby the spinel xenoblasts . The Ca-poor pyroxene is slightly richer in Fe (mg* = 0.73) than the coexisting Ca-rich one. Spinel is a dark-green pleonaste showing some chromium content (Cr_2O_3= 2 - 4 wt%).

Wt%	Coarse Type	Fine Type	Ultramafic	WEB
Al_2O_3	46 -61	46 -61	56 -59	56.0-60
FeO tot	16 -25	29 -40	18 -19	25 -28
MgO	13 -17	9 -11	17 -18	11 -15
Cr_2O_3	0 - 2.5	0 - 0.5	5 - 7	0.7- 4.3
Mg/(Mg+Fe tot)	0.46-0.65	0.32-0.47	0.62-0.65	0.46-0.49

Tab. 4 - Representative composition intervals of spinels from Hyblean granulites (coarse-type, fine-type, ultramafic) and websterite veins into anorthosites (WEB) (After Scribano, 1988. WDS analyses).

In addition, it must be remarked that the anorthite content of plagioclase grains adjacent to the websterite veins increases up to An_{92}. Another type of anorthosite xenoliths occuring in the Quaternary lavas only, consists

	An (mol%)
Coarse-type	40 - 75
Fine-type	54 - 58
Ultramafic	58 - 65
Anorthosite	63 - 66

Tab. 5 - Representative composition intervals of plagioclases (anorthite mol%) from Hyblean mafic granulite and anorthosite nodules (After Scribano, 1988 and unpublished data. WDS and EDS analyses).

coarse (up to 3 cm) elongated An_{30} plagioclase grains with cumulate texture.

This plagioclase is polisynthetically twinned according to the albite law, with very fine lamellae. It is homogeneous, contains abundant tiny apatite grains and shows complex deformation features.

The contact surfaces between plagioclase grains and the host lava are curvilinear owing to resorption reaction. Isolated plagioclase xenocrysts showing the same composition and morphology features as the previous cumulates may represent disrupted parts of the latter.

Extremely rare felsic microxenoliths have been found in some quaternary alkalibasalts from northern part of the Plateau. They consist of a few, large crystals of quartz with undulose extinction, showing polygonalization at the crystal edges.

4. Discussion

Granoblastic texture, high degree of unmixing between Ca-rich and Ca-poor pyroxenes (Fig. 3 c) and twinning style of the plagioclases (Fig. 5; Suwa, 1979) suggest that Hyblean granulite nodules have undergone protract sub-solidus reactions. This fact is also supported by the aluminous, tschermakitic nature of clinopyroxene (Fig. 3 b) and chemistry of Ca-poor pyroxenes. They plot into the metamorphic field of Ca/Ca+Mg+Fe) vs Fe/

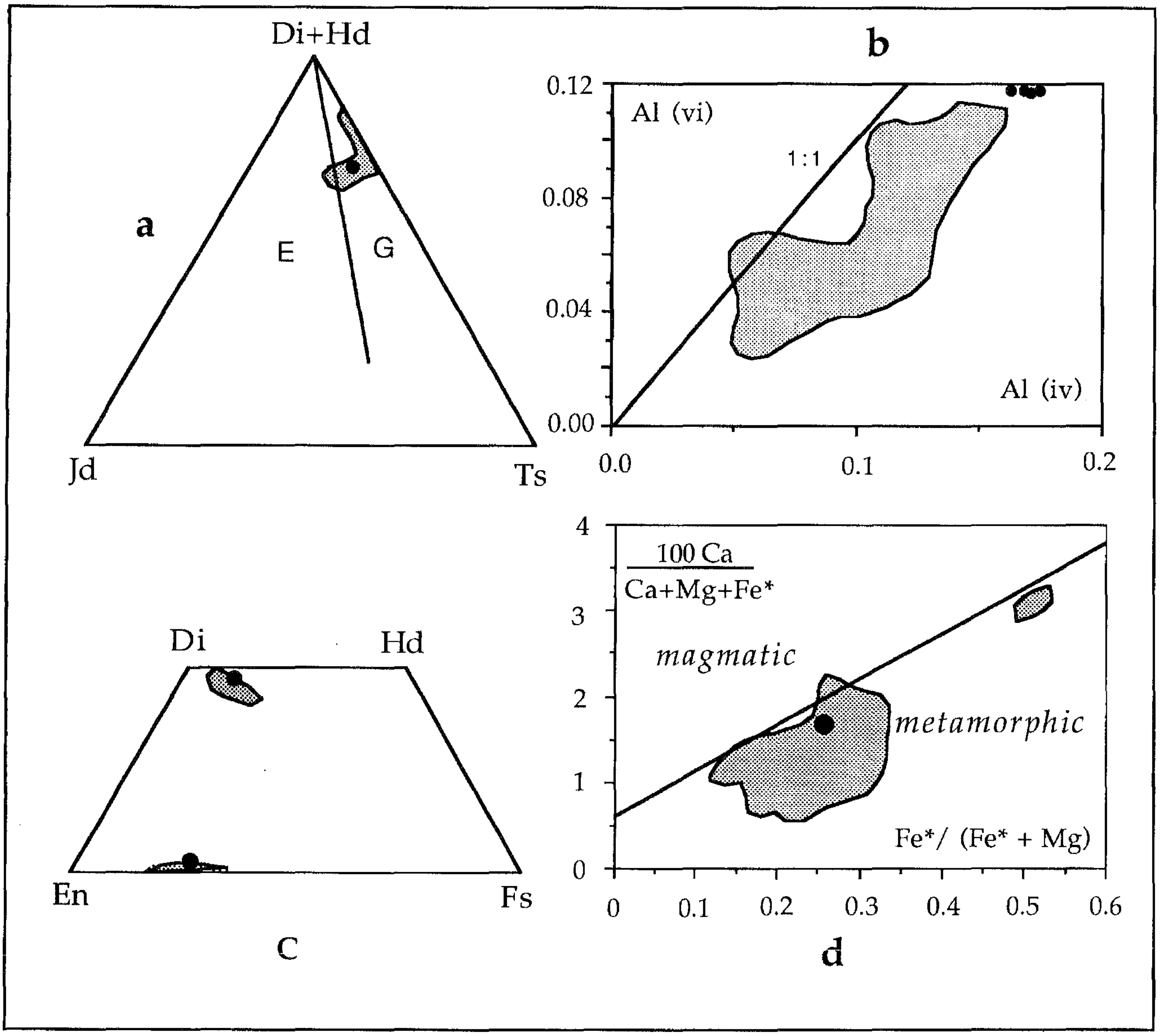

Fig. 3 - Diagrams summarizing mineral chemistry characters of pyroxenes from Hyblean granulite nodules (grey areas) and websterite veins into the anorthosites (black dots) (after Scribano, 1988); *a:* eclogite (E) and granulite (G) fields in the Jadeite (Jd)-Tschermakite (Ts) and diopside+Hedembergite (Di+Hd) diagram, following Lowering & White, 1969); *b:* plots of Al(vi) vs. Al (iv) (atoms in formula) for clinopyroxenes; *c:* plots into QUAD diagram; *d:* magmatic and metamorphic fields for orthopyroxenes after Rietmaijer, 1983: most of the pyroxenes plot into the metamorphic field.

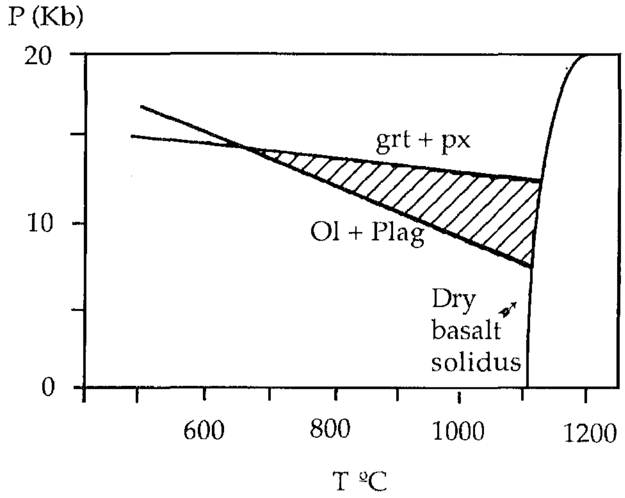

Fig. 4 - Pressure - temperature space of Hyblean mafic granulites between the Olivine-*out* curve (Ol + Plag), the garnet-*in* (grt + px) and the dry-basalt solidus, following Herzberg, 1978, Kushiro & Yoder, 1966; Irving & Green, 1970). (For more explanations, see the text).

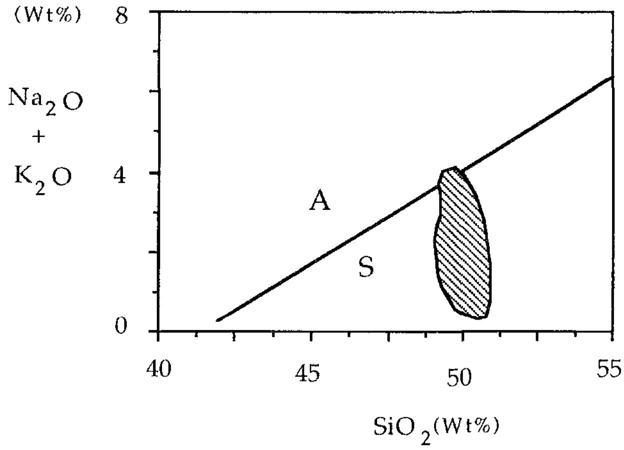

Fig. 6 - Total alkalies vs silica diagram for Hyblean mafic granulite nodules (whole nodules analyses after Scribano, 1988) and unpublished data. Alkaline (A) and subalkaline (S) fields after MacDonald & Katsura (1964).

(Fe+Mg) diagram, following Rietmaijer (1983) (Fig. 3d). Above reported mineral assemblage and chemistry suggest that equilibrium pressure - temperature space of Hyblean two-pyroxene granulites is defined by three intersecting curves. Two of them represent loci of the following reactions: 1) olivine + plagioclase = aluminous pyroxenes + spinel (olivine out); 2) aluminous pyroxenes + spinel + plagioclase = garnet + low-Al pyroxenes (garnet in). The third curve (3) represents dry basalt solidus (Fig. 4). Reaction (1) is constrained between the points: P= 0.7 GPa , T=800 ºC and P=0.8 Gpa, T= 1150 ºC; (2) between P= 0.6 GPa, T=750 ºC and P= 1.3 GPa, T= 1180 ºC (Kushiro & Yoder, 1966; Irving & Green, 1970; Green & Ringwood, 1966; Herzberg, 1978). Additional detailed pressure - temperature estimates for particular specimens are hindered by uncertainties of conventional geo-thermo-barometers for above-mentioned mineral assemblages.

Whole nodule chemistry (see above) may indicate basaltic protoliths, with subalkaline affinity in most cases (Fig. 6). In addition, the Al_2O_3-CaO-MgO diagram (Fig. 7) suggests a cumulate nature of our rocks. It is consistent with low K_2O, P_2O_5, TiO_2 contents and the banded texture of some specimens, probably reproducing a previous magmatic layering. On the contrary, the abundance of Al-spinel in most Hyblean granulites would instead indicate the silica undersaturated, alkaline character of protoliths. The low content in relatively incompatible elements of our rocks might also indicate depletion after partial melting episodes. All above self-inconsistent statements prove the complexity

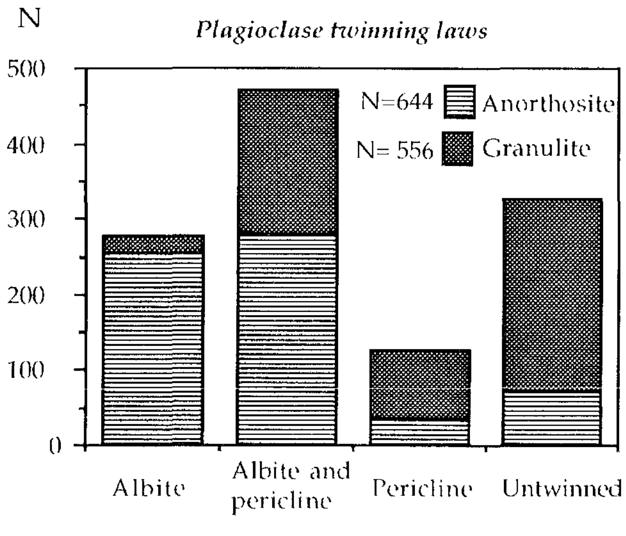

Fig. 5 - Twinning styles of plagioclases from granulite and anorthosite nodules. 644 plagioclase grains for anorthosites and 556 for granulites have been plotted.

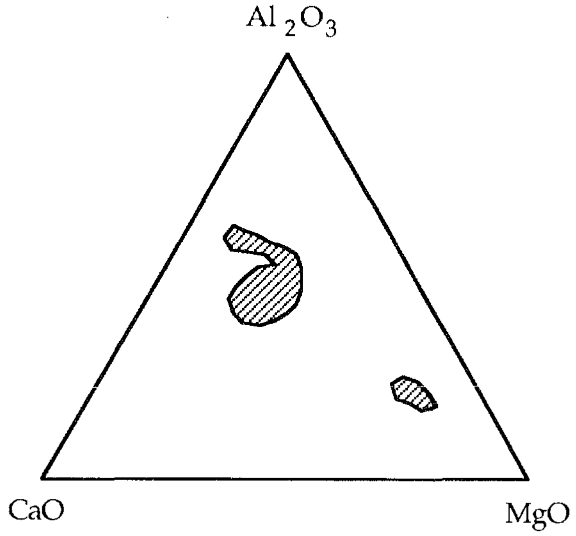

Fig. 7 - Plots of Hyblean mafic and ultramafic granulites (stripped areas) into the MgO-Al_2O_3-CaO (wt%) diagram.

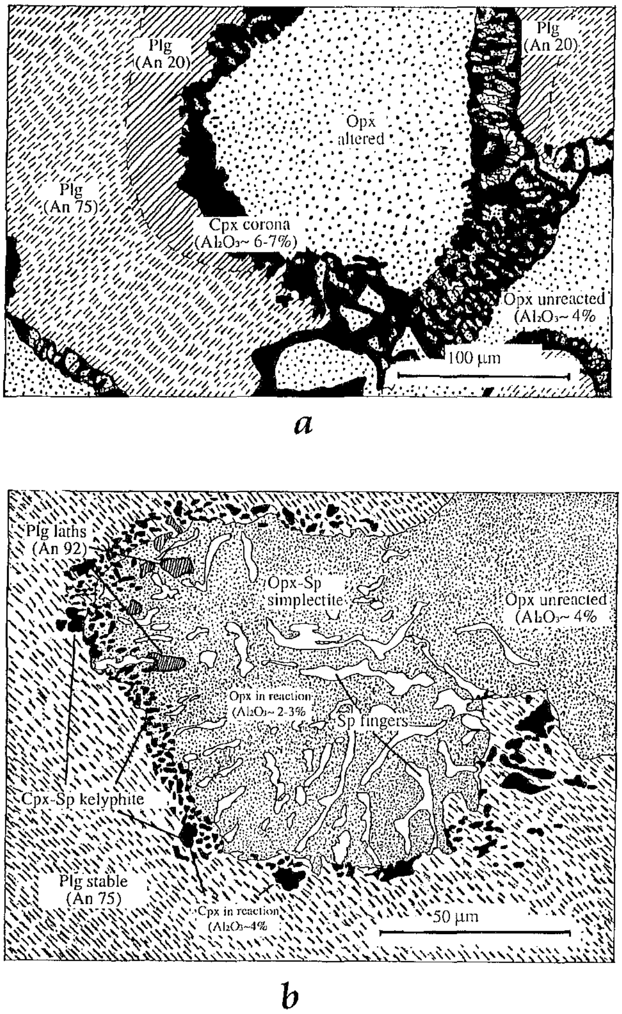

Fig. 8 - Sketches after SEM microphotographs of representative reaction textures from coronitic metagabbros. a)- *Type 1* clinopyroxene collar around orthopyroxene (now strongly altered) at the contact with large plagioclase grain. Note the composition zoning of the latter towards the corona. b)- *Type 2* keliphyte rim (clinopyroxene + spinel) between orthopyroxene-spinel coarse synplectite and equilibrium plagioclase. Note the occurrence of reaction anorthite-rich plagioclase nearby the keliphite. For full explanation see the text.

of the long-standing debated problem on the origin of lower-crustal mafic rocks (e. g. Griffin & O'Reilly, 1987 and therein references; Dawson, 1980).

About the reaction textures occurring in some coarse-banded granulite nodules (coronitic metagabbros), a three-step reaction may be suggested: in response to an increase in temperature, a diffusion-controlled reaction between Ca-poor pyroxene and plagioclase yields the clinopyroxene coronas (Messiga, & Bettini, 1990 Grant, 1988). A following drop in pressure with concomitant undercooling may favour the orthopyroxene exsolution reaction under near isochemical conditions due to the previous clinopyroxene collars. They prevent ionic exchange with the adjacent plagioclase grains. A further increase in undercooling degree may involve the clinopyroxene corona in the reaction, yielding the clinopyroxene and spinel keliphytes (Fig. 8 b).

The postulated pressure decreasing likely enhances the fluid phases activity into the granulite system, as it appears from the abundance of fluid inclusions in corona bearing minerals. Looking into more detail, the reaction producing clinopyroxene collars may be written: orthopyroxene + plagioclase I (calcic) = clinopyroxene + spinel + plagioclase II (less calcic). On the other hand, the pressure decreasing may involve exsolution of Al-spinel from pyroxene: orthopyroxene I (aluminous) = orthopyroxene II (Al -poor) + spinel. Even clinopyroxene collar may loss aluminium in form of Tschermak's molecule: clinopyroxene I (tscermakite-rich) = clinopyroxene II (Tschermakite poor) + Ca-rich plagioclase + spinel.

Texture and mineral chemistry of Hyblean anorthosite nodules closely resemble lithotypes from world-wide exposed massifs, whose magmatic origin is generally accepted (e.g. Isachsen, 1969 and therein references; Emslie, 1978; Ashwal, 1982; Morse, 1982; McLelland, 1989; Phinney et al., 1988). In addition, the composition of plagioclases of our anorthosite might give rough information on the age of the postulated source massif by comparison with information from world-wide massif anorthosites: it is known that plagioclases from Pre-Cambrian massifs are more anorthite-rich (bitownite - anorthite) than those (andesine - labradorite) from younger Proterozoic massifs

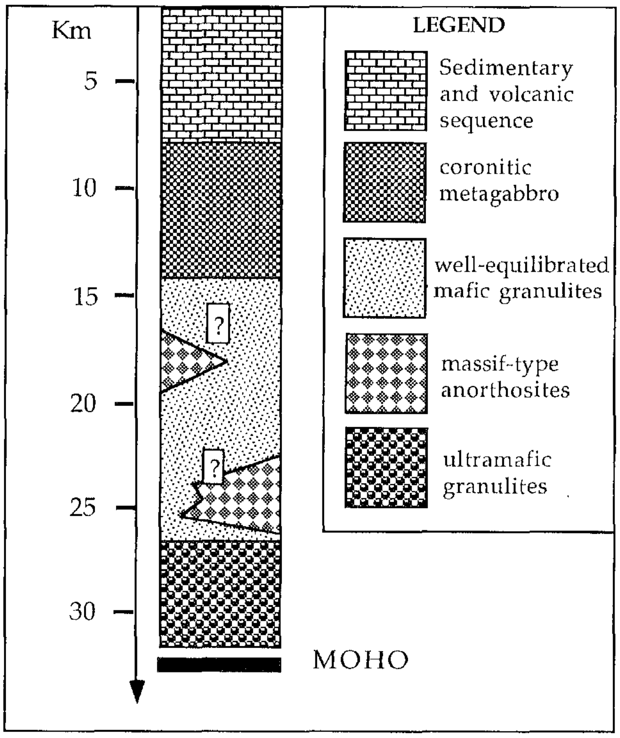

Fig. 9 - Preliminary hypothesis of stratigraphy of Hyblean crystalline basement inferred from the study of xenoliths. Relic lenses of granitoids may occur at any level of the crust, especially in the upper part. Lenses of anorthosites have been arbitrary located into the mafic granulite space. Depth of the base of the sedimentary sequence after Bianchi et al., 1987. Depth of the Moho discontinuity after Nicolich, 1986.

(Phinney et al. 1988).

The magmatic origin of websterite veins which occur in our granoblastic anorthosites is very likely, though it is evident that also the former has undergone complete recrystallization at relative low temperature: this is shown by coarse polygonal texture and high degree of unmixing between Ca-rich and Ca-poor pyroxenes: the latter occur even as granule-type exsoluted, which indicate a very long time of sub-solidus annealing (e. g. Nabelek et al., 1987). The Al (vi)/Al(iv) ratio of above-mentioned calcic pyroxenes is also consistent with a granulite-facies equilibrium (Fig. 3 b; cf. Aoki & Shiba, 1973).

Some petrographic features, as well as the occurrence of stretched spinel grains along the boundaries between the host anorthosite and veins, might indicate that textural equilibrium has been attained between the two rock types.

The occurrence of above-mentioned spinel with anorthite collars may indicate a sub-solidus reaction which involves expulsion of Tschermak's molecules from pyroxene. A number of diffusion-controlled reactions are suitable, as well as the following (Griffin, 1971) :

$$CaAl_2Si_2O_6 + MgAl_2SiO_6 = CaAl_2Si_2O_8 + MgAl_2O_4$$

On the other hand, the occurrence of above spinel only nearby the boundary surface with host plagioclase, which shows composition zoning (from An_{62} to An_{92}: see above), may suggest that the increasing in Al and Ca in the plagioclase occurred before sub-solidus recrystallization, owing to the local temperature increasing during the vein injection.

Though it is difficult to obtain the whole-rock chemistry of websterite veins because of their relative small size, some analytical results by integrating modal data and mineral chemistry suggest that they are far out any suitable magmatic liquid, being instead consistent with an Al-rich, subcalcic augite composition.

This fact may suggest a crystal/liquid segregation process: "flow differentiation" has been first invoked for the origin of websterite veins found in mantle-peridotite nodules (Irving, 1980) and is believed to represent a major mechanism of polybaric fractionation of ascending magmas within narrow conduits: for slow flowing rates, liquidus crystals would be continuously "plated" on conduit walls as the magma (now fractionated by removal of the crystals) moved through.

It is known from exposed massifs (for example those outcropping in Eastern North America) that the typical anorthosite suite contains minor facies with larger proportions of mafic silicates, whose extreme concentration produces minor ultramafic rocks.

These occur either as ultramafic layers or dikes (and veins) into the host anorthosites, having been interpreted as late-stage differentiates from that magma which have undergone extensive fractionation of plagioclase (e.g. McLelland, 1989).

Some Authors also suggest that the websterite dikes into the anorthosites represent remobilized cumulates (Emslie, 1978). Unfortunately we have not enough data to define if the above-mentioned ultramafic veins derive from liquids comagmatic with the host anorthosites or they represent injection of unrelated magmas.

The close similarity between above-mentioned nodules and some anorthosites from world-wide exposed massifs may give additional information on the nature of Hyblean crystalline basement. In fact, the anorogenic setting of most anorthosite massifs has been already demonstrated (e.g. McLelland, 1989).

The coarse, cumulate-textured anorthosites may represent either magmatic relics of above postulated massif or separated igneous bodies. The policrystalline quartz microxenoliths may represent fragments of granitoids.

5. Conclusions

The above-mentioned nodule suite may give useful information on the nature of the Hyblean crystalline basement, providing that 1) the xenoliths represent a complete sampling of Hyblean crystalline basement 2) the collected specimens well represent the crustal -nodule suite from the Hyblean volcanics. The first assumption seem suitable at least for the Miocenic diatremes (see above) where mantle xenoliths, crystalline-basement xenoliths and lithologies from all the sedimentary sequence coexist. About the second statement, Griffin & O'Reilly (1987) remark that " Field sampling of xenolith suites may be strongly biased by the idiosyncrasies of individual collectors, including failure to recognise certain types of samples as significant...": Though carefully sampling of Hyblean xenoliths has been carried out since 1986, additional discovers are still possible. It gives a preliminary character to the present and other reports for more and more years in future.

Comparison of our results with published models of continental crust either from exposed cross-sections or geophysical data (e. g. Salisbury & Fountain , 1990; for the Italian lithosphere see Boriani et al., 1987), may suggest the following considerations:

1) the felsic upper crust is missing or exceeding thinned in the Hyblean area. Hence the top of the crystalline basement, directly supporting the sedimentary sequence, probably consists of disequilibrium lower-crustal materials (coronitic metagabbros). Thin granitoid lenses, representing relics of the older "salic" crust, are probably intercalated within the mafic rocks. The postulated crustal thinning may depend either on long standing uplifting and erosion of the basement or on long-duration basaltic underplating

process, which can laminate and erode the upper crust from below, by repeated stratiform magmatic intrusions. No indications of above processes are clearly envisaged in the nodules. Though disequilibrium textures occurring in some coarse-grained mafic granulites indicate a rapid uplifting of the basement, it seems unlikely to refer them to the upward movements responsible for the postulated, severe shaving of the crust. On the contrary, the coronitic metagabbros may record an increasing of the geotherm and an uplifting of the Plateau during the lower and Middle Miocene, pre-dating the Upper Miocenic volcanic activity which brought the xenoliths to the surface.

2) The stratiform attitude of all crustal lithologies (different granulite types, anorthosites) represented by nodules is very likely. Though no geobarometers can give information to draw a precise stratigraphy of all rock types it seem suitable that coronitic metagabbros occur at the top of the sequence and ultramafic granulites stay at the bottom, nearby the Moho discontinuity. It is posed at a depth of 28-35 Km beneath the Plateau on the basis of seismical data and geotectonic correlations (Boriani et al., 1987).

3) Though only 15 nodules of anorthosites have been found in the Hyblean volcanics, it is very likely that an anorthosite massif, or a relic of that, occurs somewhere beneath the Plateau. The very well equilibrated metamorphic texture of above-mentioned anorthosites and their websterite diklets suggest that these rocks never passed through the spinel-gabbro field boundaries, probably representing the most mature and oldest rocks of Hyblean crystalline basement.

To conclude, a 20 - 25 Km sequence of intermediate to mafic magmatic rocks, equilibrated (or re-equilibrated) under granulite facies, form the most part of the crystalline basement beneath south eastern sector of Hyblean Plateau (Fig. 9). These lithologies are typical of anorogenic lower continental crust. The upper "felsic" crust is unsignificant in this area. This fact is consistent with recent magnetometric results suggesting the occurrence of a strongly magnetic "mafic" basement just at the base of the sedimentary sequence (Bianchi et al., 1987). Additional investigations on the Hyblean crustal nodules are still in progress to give further support to the above-mentioned hypothesis.

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GEOPHYSICAL PRESENCE OF A DEEP-SEATED "GRANITIC" STOCK IN THE MASSAMARITTIMA MINING DISTRICT (GROSSETO, SOUTHERN TUSCANY): METALLOGENETIC IMPLICATIONS

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Abstract

A series of geophysical surveys has been completed over a 250 sq. kms area in Southern Tuscany.

This work has been carried out on behalf of the Italian Governement in the framework of the national mineral inventory and research program.

The aim of the geophysical survey was the ground follow-up of some intense magnetic anomalies detected by an airbone survey performed over this area in 1980.

A deep-seated, anomalous and seismically transparent low-density body, which has a low magnetic susceptibility with respect to country rocks, has been localized. Geophysical ground surveys (magnetic and gravimetric) were supported by a reinterpretation of seismic surveys, carried out by the ENEL over the adjacent Larderello-Monterotondo geothermal field. This body may be interpreted as a probable igneous intrusive of low specific gravity, such as a granite, whose roof lies 1,300-2,500 m below the surface and within the basement schists.

The existence in the mining district of southern Tuscany of a shallow igneous intrusive, with Niccioleta mine directly above it and the Boccheggiano and Fontalcinaldo pyrite deposits and more satellite bodies at its outer limits, supports the interpretation that the Tuscany deposits occur in a close spatial correlation with the granitic bodies. This existence also confirms the importance of these intrusives with regard to the local metallogenesis and calls for a review of the current genetic models of the Tuscan mining district.

Key words: *ore-deposits, mineralization, tectono-magmatic evolution, Alpine orogenesis, southern Tuscany.*

1. Introduction

First geophysical informations about the hypothesis that an intrusive body might exist in the area just north of Massa Marittima were provided by an airborne magnetometric survey performed during the first stage of the basic research in southern Tuscany, in 1979-80, with the financial support of the Italian Government (Ministry of Industry and Trade).

The comparison between the results of this work and the mining knowledge pointed out a fairly good areal correspondance between some positive magnetic anomalies and the known mining zones (Boccheggiano, Fontalcinaldo, Niccioleta, Fenice Capanne, Gavorrano, Punta Calamita, Monte Argentario) (RIMIN, 1980; LEONARDELLI & STEA, 1983).

It was noted in particular that the positive anomalies in the Boccheggiano-Massa Marittima area showed a sub-circular distribution all around a negative magnetic core, similarly to what had previously been checked in other areas (e.g. Elba). Hypothesis have been made that this pattern could be linked to "ore-bodies" around an intrusive stock.

Some of these anomalous areas (i.e. Fenice Capanne, Fontalcinaldo and Boccheggiano) were later examined with multimethodologic and detailed geophysical prospectings (magnetometry, gravimetry, electrical sounding, seismic prospecting) for mining research. The most recent phases of this research program concerned an area, which is about 250 sq. km wide, between Monterotondo Marittimo and Montieri in the north, and Massa Marittima and Roccatederighi in the south.

In this region, the geophysical work consisted of detailed magnetometric prospecting (24,000 measure stations), a regional gravimetric survey (about 1,000 stations), more than 200 V.E.S. and other, more particular operations. The results of this work by RIMIN are treated in the final report of the research (1990 a), and in a note by BENEO & DUPRAT (1990, in press).

With the permission of ENEL (Italian National Electricity Board), some seismic lines, previously carried out for geothermic research in the Monterotondo-Larderello-Travale area by ENEL, were also examined. They provided precious indications for improving the interpretation of the deep structure of the Massa Marittima region.

2. Results of the Geophysical Prospecting

The field magnetometric survey reveals in detail a large magnetic minimum situated north of Massa Marittima. This area of magnetic minimum is surrounded by a series of positive anomalies: Massa Marittima, SE of Monterotondo Marittimo, south of Prata, north of Tatti. Inside the magnetic minimum area, other positive anomalies exist, among which the most evident is situated SE of Niccioleta (Fig. 1).

At the same time the gravimetric surwey displays on the Bouguer anomalies map (calculated for a density of 2.4 g/cm^3; Fig. 2) a large and negative anomaly which corresponds to the magnetometric minimum.

Southward, westward and northward this gravimetric minimum is bounded by marked gradients leading to positive anomalies (west of Massa Marittima, area between Tatti and Prata, east of Monterotondo Marittimo), which represent the probable boundaries of the "causative body" and may be interpreted as lines of tectonic significance.

The analysis of the ENEL seismic lines of Monterotondo-Massa Marittima-Montieri region reveals a deep-seated body with a seismic response different from that of the surrounding rocks. In fact, it seismically behaves like an isotropic formation, i.e. it is fully transparent and no reflection surface may be distinguished within it. Its western boundary, recognized on the basis of the seismic profiles, fits the steep gradients of the magnetic and gravimetric map fairly well. Its northern boundary is controlled by the ENE-WSW "LAR-26" seismic profile through the area (Fig. 3).

Thus, north of Massa Marittima, a superposition of (1) a large and marked negative magnetic anomaly, (2) a strong gravimetric minimum and (3) a seismic pattern referable to a deep-seated isotropic rock-body can be recognized. According to geophysicists, such a combination of features may only be related to the presence of a salt dome or a "granitic" intrusion.

The regional geological frame and the inferred depth of the "causative body" exclude the first solution and indicate the second one as the only acceptable. In fact, the geophysically estimated depth of the roof of this rock-body ranges from 1,300 and 2,500 m below the topography, depending on the zone where it is calculated and the kind of geophysical information. Anyway, even if the depth is shallower, the rock-body should lie inside the metamorphic basement formation whose geology excludes a possible presence of salt domes.

The width of the upper part of the "granitic" body, redrawn on the basis of geophysical input, is shown in Fig. 3. The roof of the intrusion must be at a higher altitude (i.e. at a minor depth) below Niccioleta, reaching progressively lower levels toward east. The eastern boundary of the diapir is hardly recognizable because of a lack of seismic data. Its western boundary seems well marked by a probable north-south striking tectonic feature, while the nature of the northern and southern boundaries is not so clear.

The existence of an intrusive body below the Massa Marittima region was already proposed by MARINELLI (1967, 1983), and is strengthened by the following geological and minero-petrographical evidences:

— during the preparation of the drainage gallery in the old Boccheggiano mine, at about "Pozzo Ballarino" stope (south of Boccheggiano), the tunnel crossed a porphyritic granite dike which could not be dated because of a strong and deep metasomatism (MAZZUOLI, 1967);

— the n. 319 drill-hole (RIMIN archives), just NW of the Campiano mine, crossed a 16-meter-thick pyroxene-garnet- and epidote-bearing skarn with minor hematite and sulfides, lying in an alternation of anhydrite and dolomite levels in the hangingwall of the Boccheggiano fault;

— in the Campiano mine, silicatic lithotypes deriving from dolomitic-anhydritic levels are associated with the ore-body and consist of chlorite, epidote, clinozoisite, garnet, amphibole, carbonate, albite and serpentine; pyrrhotine, ilvaite and sulfides joined with the skarn were observed in part of the ore deposit (GREGORIO et al., 1980; MASOTTI & FAVILLI, 1987);

— in the Niccioleta mine, the n° 264 internal drill-hole which started at "level 103" (i.e. at 106 m above sea level) crossed an almost 700-meter-thick formation of phyllite, often epidotized and containing levels of pyroxene-garnet- and amphibole-bearing skarn; below the level of 450 m of the drill-hole biotite, plagioclase porphyroblasts, abundant amphibole, pyroxene and sometimes also brucite were present, i.e. a cornubianitic paragenesis typical for a thermometamorphic facies;

— at Niccioleta, the ore-bodies consist of massive pyrite stocks joined with skarn silicates within the sulfide-carbonate lenses interbedded in the basement sequences; the silicates are almost totally represented by hedenbergite which is more or less changed into amphibole and andradite in the lowest parts, while garnet exists at higher levels (TANELLI, 1977; LATTANZI & TANELLI, 1985).

3. Discussion

The presence of an intrusive body below the Niccioleta-Boccheggiano area raises again the long debated question of the relationships between Tuscan "granitic" magmatites and Fe sulfide and oxide massive ore deposits.

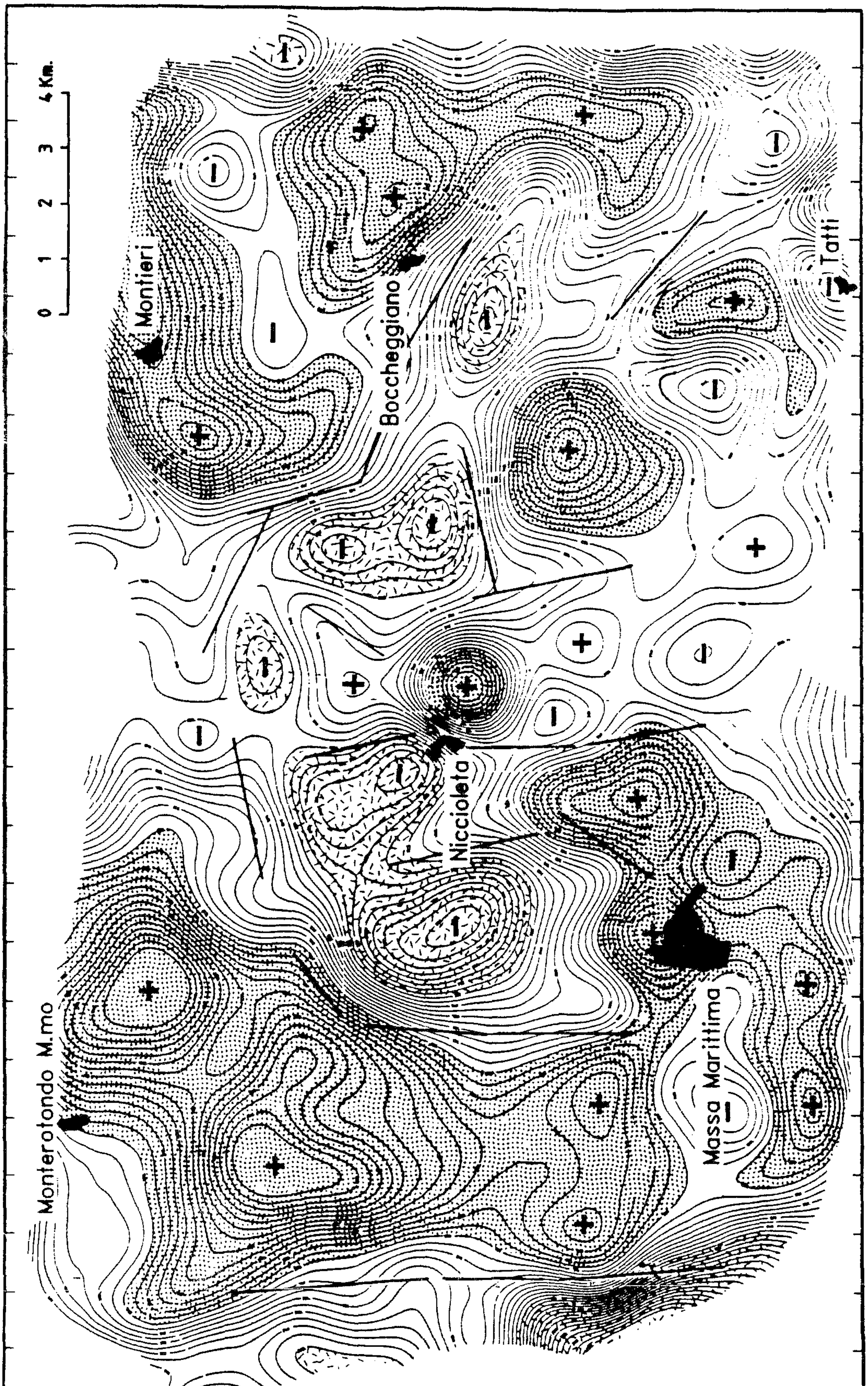

Fig. 1 - Filtered ($\lambda > 5$ km) and pole reducted total magnetic field map (RIMIN, 1990, modified).

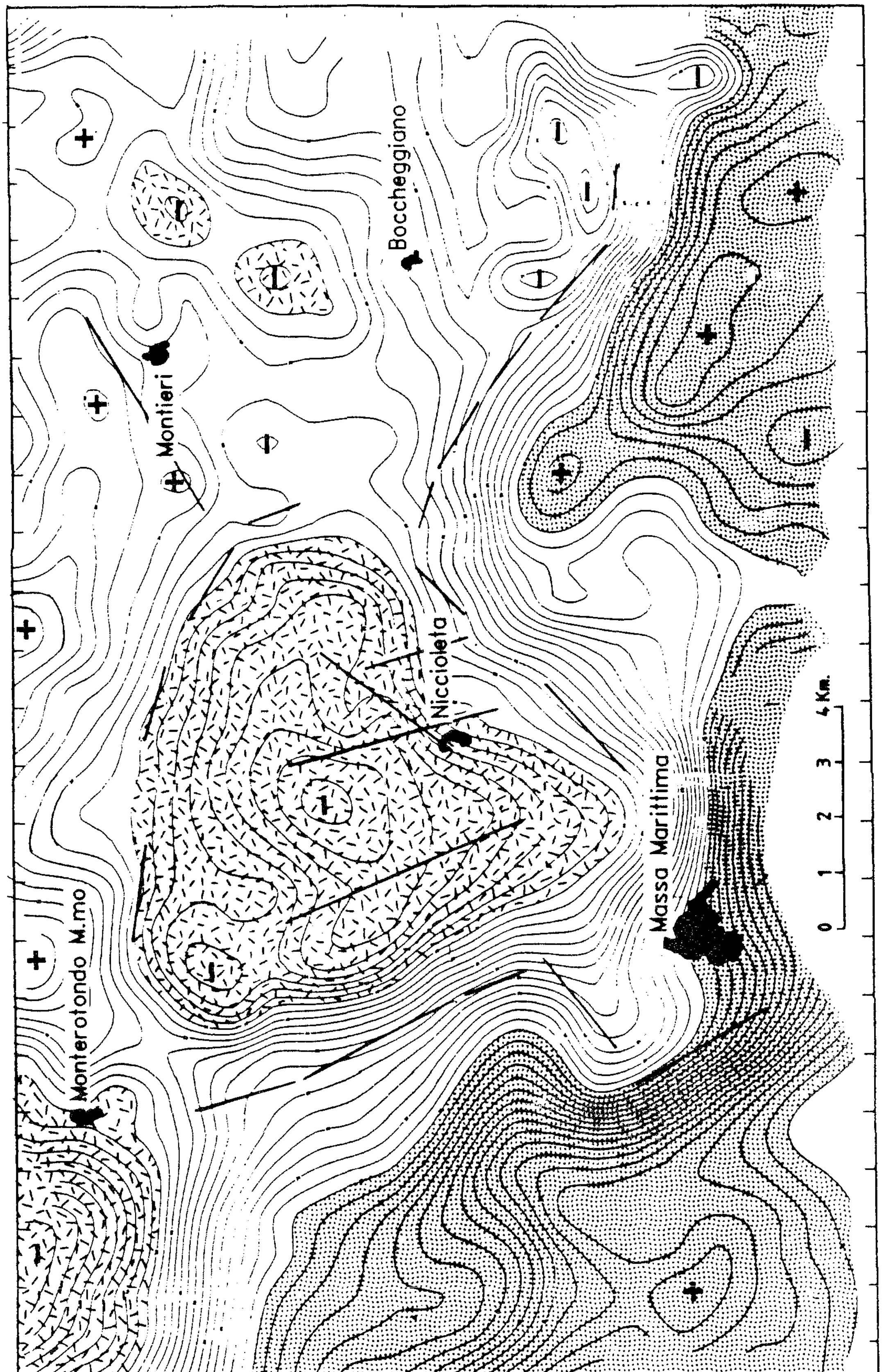

Fig. 2 - Bouguer anomalies map (density = 2,4 gr/cm^3) (RIMIN, 1990, modified)

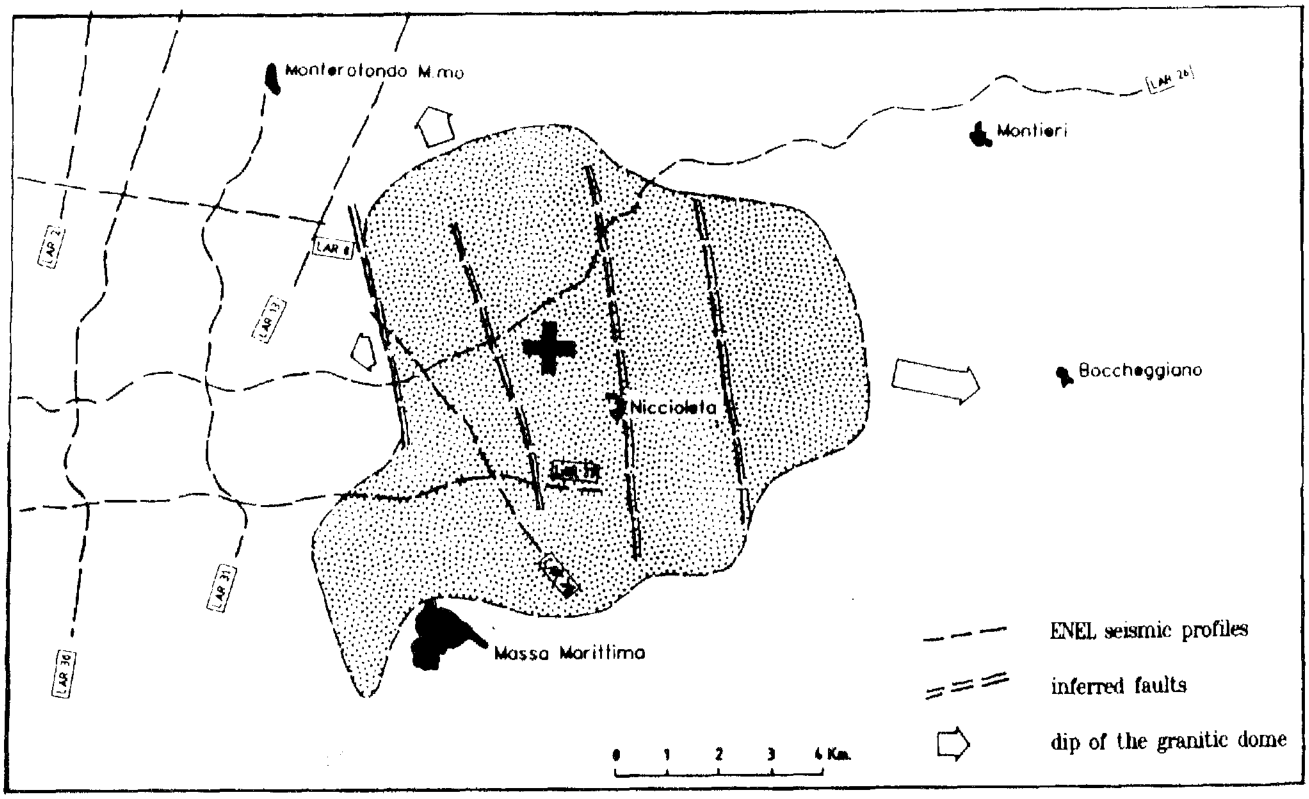

Fig. 3 - Schematic geophysical recostruction of the top (+) of the "granitic" dome (RIMIN, 1990, modified).

All the researchers who have enquired into the origin of the Maremma ore deposits and have provided many data on their structural, mineralogical-textural and geochemical features, acknowledge the importance of the Apenninic tectono-magmatic event during the ore evolution, but their opinions about the metal provenance and the processes that led to their concentration are discordant.

Here we only want to give a schematic exposition of the two main opinions, and refer the reader to BRALIA et al. (1979), TANELLI (1983), MARINELLI (1983), TANELLI & LATTANZI (1986), LATTANZI et al. (1987) and DECHOMETS (1987) for further details and references.

According to the first opinion, every type of ore deposits (Fe oxides, pyrite, sulfides, antimony, mercury) of southern Tuscany derived from the Apenninic tectono-magmatic event.

The second opinion considers that the most important pyrite and Fe oxide deposits of southern Tuscany "..., at least as pre-concentrations, predate the Apenninic event and were formed in sedimentary and/or hydrothermal-sedimentary environments of Triassic and/or Paleozoic age" (TANELLI & LATTANZI, 1986); these mineralizations were partly reworked by the Apenninic tectono-magmatic action.

Certainly, the new geophysical data and the above mentioned geological and mining considerations are not in itself conclusive enough to support a metallogenetic theory rather than others. Nevertheless they strengthen an interdependence of tectonics, magmatic activity and mineralizations and their space relationships.

The first consideration is that the association of granites with mineralizations (at least the massive oxide and sulfide ones) seems less and less casual at the regional scale. Such a feature is shown by almost all the known ore deposits: Elba, the surroundings of Campiglia, Gavorrano, Castel di Pietra, to which Niccioleta and Boccheggiano-Campiano are now to be added.

The above association may be also inferred for the ore deposits of Fenice Capanne and P. Mortaio (Monte Argentario) (RIMIN, 1972; TANELLI, 1977; MARINELLI, 1983).

The magmatism-mineralization link is also evident if we consider that the common association and the paragenesis of all the southern Tuscany mineralizations indicate a thermal decrease toward the east (Fig. 4), in accordance with a general increasing of the intrusion depths in the same direction, due to the progressive deepening of the crust-mantle boundary (LAVECCHIA & STOPPA, 1989 a, b).

But it is even more interesting to compare the areal distribution of magmatic and mineralization phenomena to

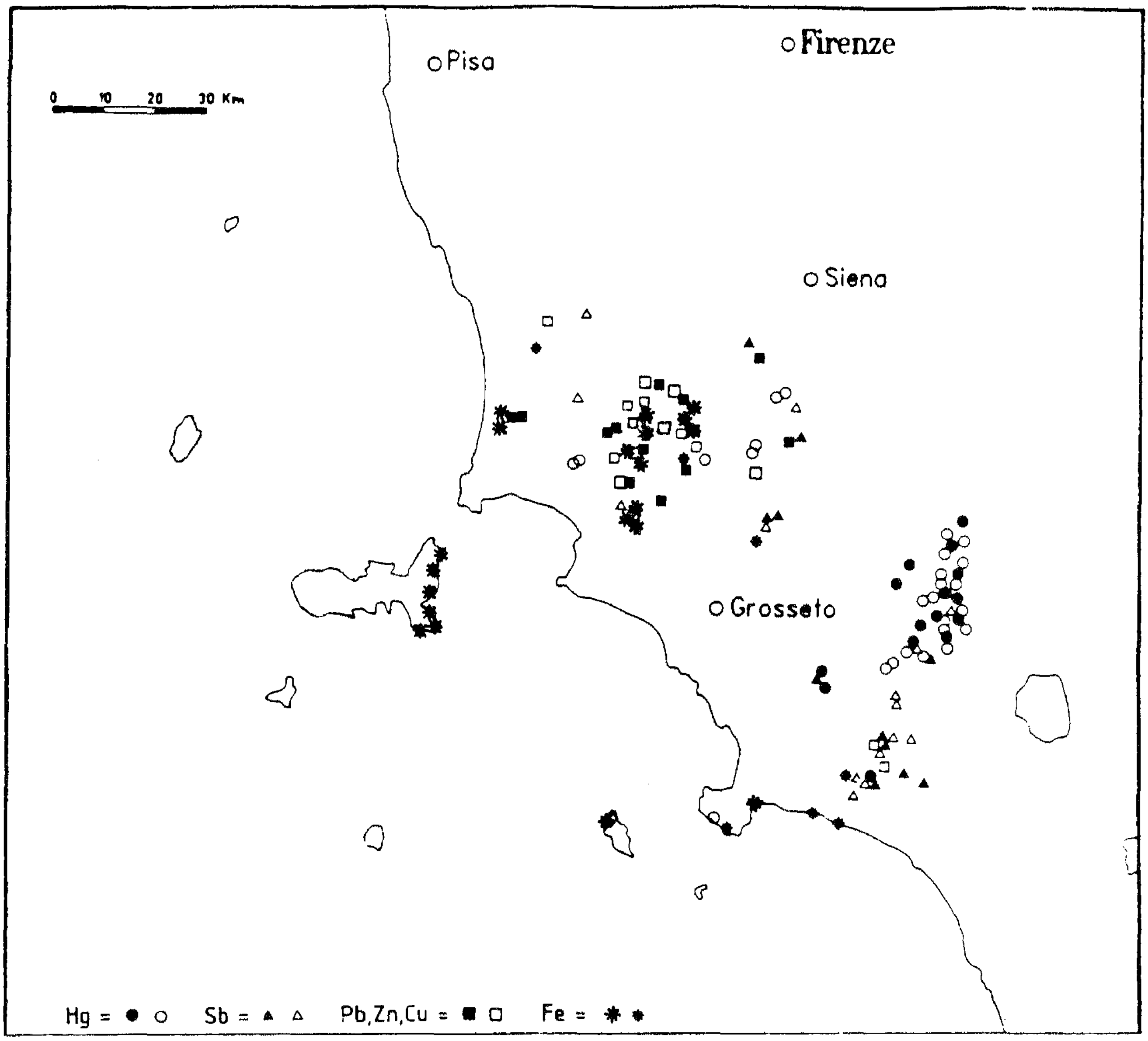

Fig.4 - Location of the main mineral occurrences in Southern Tuscany. Full and open symbols represent mines and research/traces, respective

the trends of the late-Alpine extensional tectonics. This tectonics was generated by a general geodynamic regime in which the Tyrrhenian area has been depicted as a back-arc basin, or the result of an active or passive rifting process.

At any rate, the opening of the Tyrrhenian area must have occurred since the Middle-Upper Miocene, during the action of a regional strain regime with an east-west maximum extension and the development of north-south striking rift systems.

According to LAVECCHIA & STOPPA (1989a,b), in the Tyrrhenian area, the rifting must have led to "... the development of north-south striking semigraben ..." coeval with a counter-clockwise rotation of the Adriatic block around a pole situated north of Elba.

During the advance of extension and rotation, the longitudinal direction of the Tuscan graben progressively changed from north-south to NNW-SSE, up to NW-SE, and shifted toward the external portions of the chain getting younger and younger eastward and northward (GIGLIA, 1974).

The Tuscanid "granites", commonly related to anatexis phenomena connected with extensional tectonics, are therefore affected and controlled by the geometric features developed during that deformation regime. The outcropping granites, those which are crossed by drillings and those which are inferred on the basis of a geophysical approach (Fig. 5), are generally distributed along evident north-south trends.

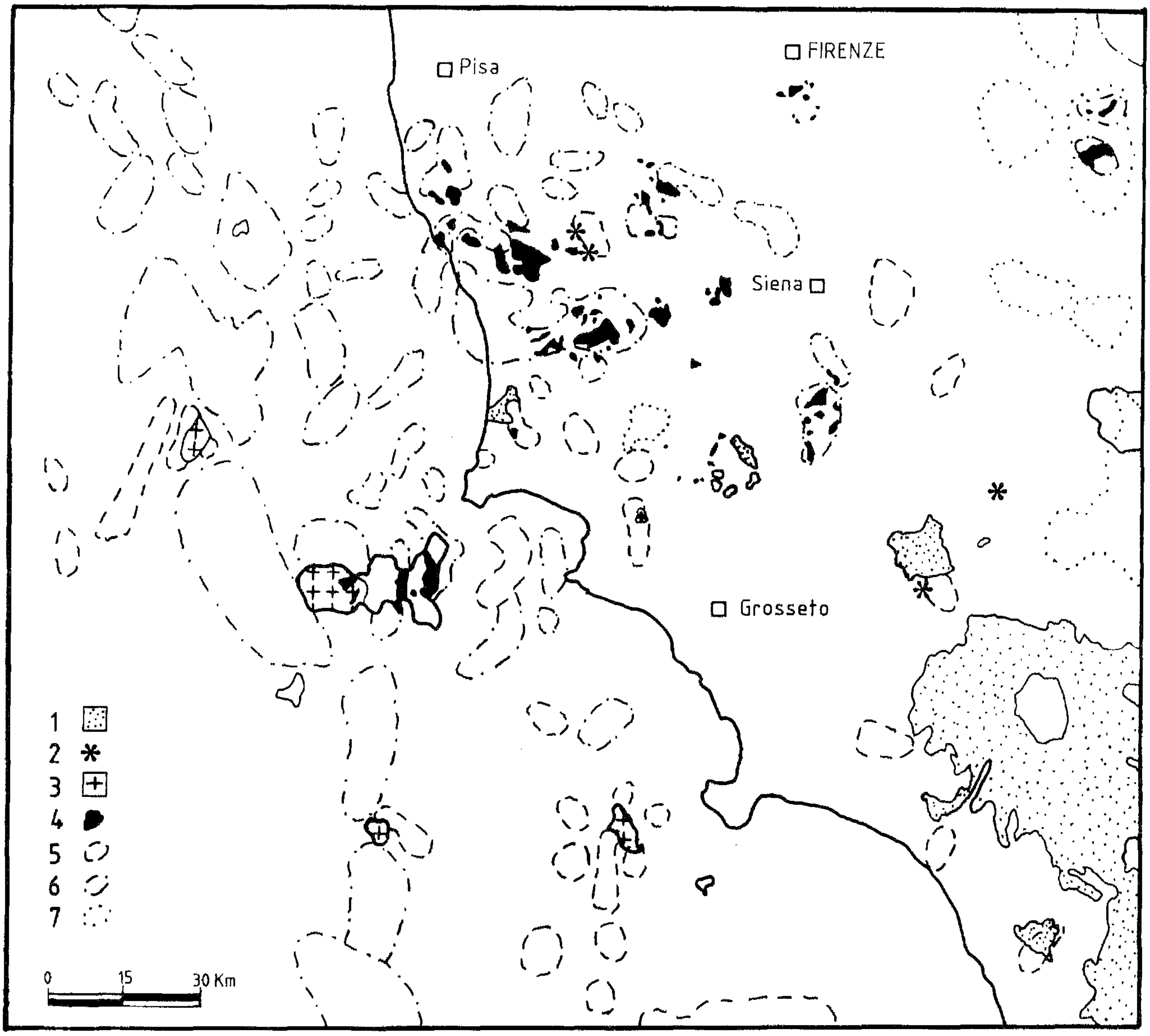

Fig. 5 - Map of main intrasedimentary magnetic bodies (after CASSANO et al., 1986 modified)
1. acidic subvulcanite outcrops; 2. basic subvulcanite outcrops; 3. granodiorite and granite outcrops; 4. ophiolite oucrops; 5. hidden bodies of Neogenic - Quaternary volcanite and plutonite (magnetic suscettivity = 100->1800 uem cgs x 10^{-3}); 6. hidden bodies of ophiolites s.l. (magnetic suscettivity = 1000->6000 uem cgs x 10^{-3}); 7. hidden bodies of "Appennino Tosco-Romagnolo" volcanite and plutonite (magnetic suscettivity = 100->500 uem cgs x 10^{-3});

It is apparent that the same north-south tectonic framework also influenced the distribution of the main ore deposits of southern Tuscany (Fig. 4), at a regional and sometimes local scale, independently from the stratigraphic-structural appartenance of the host units of the mineralizations (basement metamorphic sequences, Tuscan Nappe sequence, Ligurian Units).

In fact, at Elba and in the Campiglia, Manciano, Monte Amiata and mainly Massa Marittima areas, the ore deposits outline north-south and NNW-SSE trends which are locally associated to faults, tectonic highs and depressions with the same orientation. An evident example is the meridian mineralization belt extending from Gavorrano-Castel di Pietra toward Fenice Capanne-P. al Montone, up to Niccioleta.

Even if we try to connect the main tectono-morphological allignments recognized on the LANDSAT images (BOCCALETTI et al., 1977; BEMPORAD et al., 1986; RIMIN, 1980, 1990b) with the distribution of the ore deposits and the geochemical anomalies (RIMIN, 1985, 1990b), it appears clearly that the regional distribution of the mineralization phenomena is not casual (Fig. 6).

In detail, it can be seen in Fig. 4, 5 and 6 that in the Massa Marittima, Manciano and Monte Amiata areas the ore deposits (and joined mineralization phenomena) are

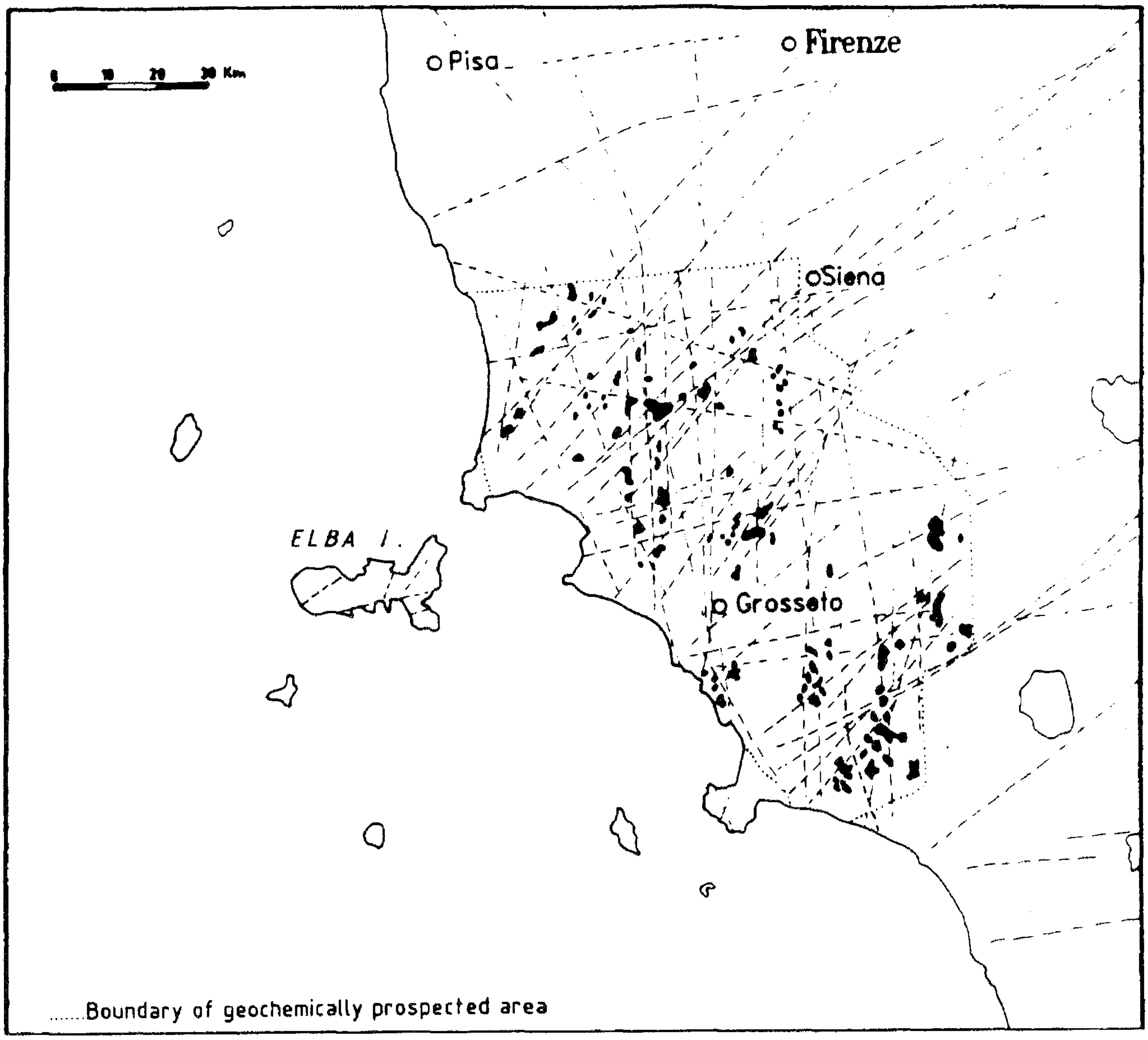

Fig. 6 - Main anomalous geochemical areas of Southern Tuscany and tectono-morphological lineament sistems from LANDSAT images.

especially concentrated around the intersections between meridian Apenninic and transversal anti-Apenninic tectonic structures. A good correspondance of the geochemical-mineralogical trends with the north-south and/or NNW-SSE lineation systems is also outstanding.

In conclusion, we retain that the spatial distribution of the mineralizations known in southern Tuscany was controlled, at least at a regional scale, by the Tuscan crustal extension which gave rise to almost meridian rift systems; the same tectonic system controlled the spatial distribution of granitic intrusions of Neogenic-Quaternary age.

These phenomena must have been more striking in the intersection areas between Apenninic and anti-Apenninic tectonic features, which could be a heritage of the Oligocene-Miocene compression event.

The emplacement of the intrusions must have been accompanied and/or followed (MARINELLI, 1983; DECHOMETS, 1986; LAVECCHIA & STOPPA, 1989 a,b) by a regional uplift of southern Tuscany generating particular structures. The sub-circular Boccheggiano fault which, referring to the geometry of the pyrite body of the Campiano mine, seems to cut off and re-activate some pre-existing north-south and NE-SW structures (Fig. 7), could be considered as an example of such a tectonic picture.

4. Conclusions

Starting from geophysical and mining evidences of the existence of a Neogene-Quaternary "granitic" intrusion in the Niccioleta-Boccheggiano region, this short note is

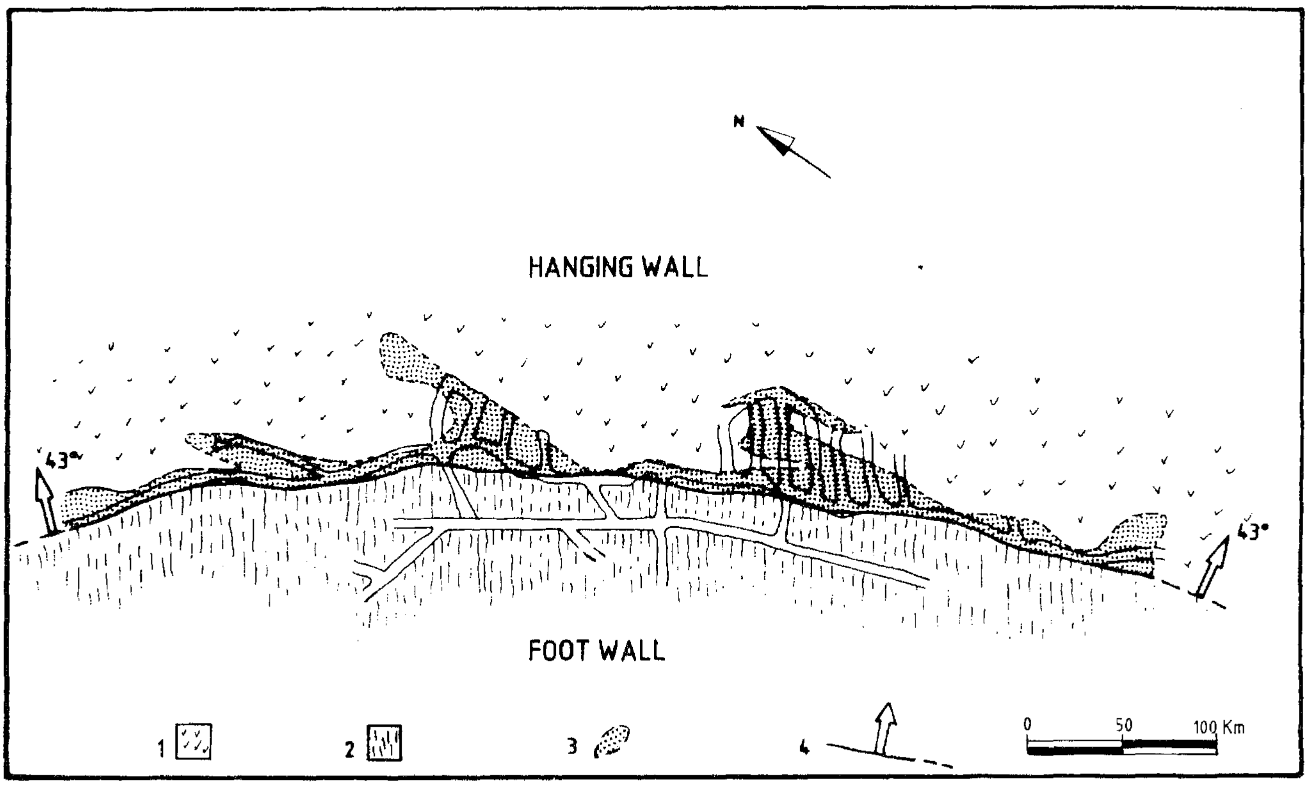

Fig. 7 - Sketch map of -140 u.s.l. level of Campiano mine (after MASOTTI & FAVILLI, 1987, modif.) 1 - Burano anhydritic Formation, Late Trias; 2 - Boccheggiano phylladic Formation, Trias/Paleozoic; 3 - pyrite ore-body; 4 - Boccheggiano Fault.

aimed to demonstrate the relationships that seem to link the regional distribution of mineralizations, the late-Alpine extensional structures and the "granitic" intrusions in southern Tuscany.

For this purpose, the recently acquired knowledges on the geostructural and ore deposit framework of the region, and the results of field research by RIMIN were taken into due consideration.

The outlined regional picture seems to reveal that the mineralizations were controlled by the meridian extensional tectonics, the anti-Apenninic tectonics and the emplacement of the magmatic intrusions. The last show a non casual relationship with the ores.

We may state that these features are substantial and basic indicators in the mining research. On the other hand, their regional consistency needs further investigation, and they must match the equally important observations on single mineralization bodies, in order to draw metallogenetic models based on as many data as possible.

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GLOBAL TECTONICS

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PERIODIC ALTERNATION OF TECTONIC REGIMES DURING THE CENOZOIC: A GLOBAL PERSPECTIVE

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Abstract

Two diverse geological, paleontological and climatic regimes appear to have alternated at the global scale during the Cenozoic era. They can both be grouped into five second-order cycles, which occurred with a mean periodicity of about 15 Ma. The first regime appears to have been characterized by phases of mafic activity and regional crustal subsidence, accompanied by widespread marine transgression and voluminous filling of basins. The period ended with phases of transpressive orogeny. This "systaltic" regime was distinguished by climatic amelioration and a predominantly normal geomagnetic field and was followed by a "diastaltic" regime, which was characterized by crustal uplift (including continental elevation), denudation and peneplanation, felsic volcanism and intrusion. This second "continental" period was also distinguished by global sea level fall, climatic deterioration (which was capable of resulting in ice ages), and reversal of the geomagnetic field.

It is held that these regimes represent stages in the geotectonic development of our planet, with its pulsing rhythm of lithospheric expansion and contraction, as reflected in the geomagnetic "clock", which has acted as a sort of metronome for the global "tectonic orchestra". The change from one regime to the other tends to be marked by a profound kinematic reorganization and by biotic crises.

Key words: *Tectonic regime, contraction, expansion, subsidence, elevation*

1. Introduction

Geological and mineralogical evidence indicates that during the Cenozoic numerous geological variables underwent concomitant first-order changes. Increasing elevation of the continents was accompanied by a decline in global temperatures associated with progressive deterioration of the world's climates. The increasing rates of uplift and erosion of mountain systems must have had a significant impact upon chemical weathering rates and the transport of dissolved material to the seas and oceans. According to Raymo et al. (1988), the global increase in river discharges must have caused an increase in carbonate sedimentation, together with the extension of average calcite compensation depth and a decrease in the amount of isotopic organic carbon present in biogenic sediments. All of these phenomena have been detected in the Cenozoic record.

The "geocratic" and regressive tendencies of the Cenozoic contrast with the "thalassocratic" aspect of the Mesozoic, in which the submergence of continental areas predominated, along with substantial transgressions affecting many coastal areas around the world. The climax of the marine invasions that occurred during the Triassic, Jurassic and Lower Cretaceous was reached during the Turonian, 91 Ma ago, when mean sea level was 250 m higher than at present (Haq et al., 1987). We may therefore identify a great first-order megacycle, which lasted about 250 Ma and which consisted of a transgressive hemicycle during the Mesozoic followed by a regressive one during the Cenozoic.

These long-term episodes, of first-order magnitude and global extent, appear to be regulated by a great terrestrial mechanism, which we may term the "geomagnetic clock". The beginning of the megacycle since the Trias was followed by reduction in the percentage of reversed polarity (Creer, 1975), which remained below 50% until the Upper Cretaceous. From this point onwards, and throughout the Cenozoic, there has been an increase in the frequency of geomagnetic reversals. According to Mazaud et al. (1983), over the last 100 Ma the mean frequency of reversal appears to have been subject to a linearly increasing trend upon which a rhythmic fluctuation is superimposed.

The objective of this paper is to delineate the second-order gobal pulsations in geological, climatic and biological phenomena which are superimposed on the great Cenozoic megacycle. This attempt has resulted in subdivision of the period in question using a scheme that, however, must be considered tentative until further refinements have been made and verification has taken place. This temporal scheme is characterized at the global scale by two types of alternating regime, related respectively to continental submergence and emergence.

2. Times of Magnetic Quiescence: Periods of Continental Submergence and Warm Climates

These are the times in which mafic activity and crustal foundering are extensive and marine transgression is widespread. Subsidence affects both the basins on the continental margins and those which are intracratonic. Before acquiring a regional character, foundering is preceded by intense mafic volcanism and intrusion, consisting of voluminous outpourings of flood basalts.

The clearest example of regional subsidence preceded by mafic magmatism is that of the formation of back-arc basins. These develop within the context of compressive deformation and crustal shortening. The principal episodes of regional subsidence associated with back-arcs occurred during the Early Miocene (for example, the South China, Parece-Vela, Shikoku and Japan Sea basins: see Klein, 1985; Miyashiro, 1986; Weissel, 1981) and Early Pliocene (e.g., the Mariana Trough, Bismarck Basin, Lau Basin, Havre Trough and Tyrrhenian Sea Basin). Other episodes of back-arc regional subsidence took place in the Late Cretaceous and Early Paleocene (e.g., the Tasman Basin and Coral Sea Basin) and in the Late Eocene (e.g., the Western Philippine Basin and New Hebrides Basin). According to data obtained from DSDP sites (see Klein, 1985), it appears that the initial phases of back-arc basin sedimentation occurred during the Paleocene (e.g., the Coral Sea Basin), Priabonian (e.g., the Western Philippine Basin), the Aquitanian (e.g., the Shikoku and Parece-Vela basins) and the Early Pliocene (e.g., the Mariana Trough). The periods in question were characterized by pulses of compressive deformation which gave rise to orogeny in continental areas.

The peaks of orogenic deformation and crustal shortening were reached at the following times before present: 53 Ma (the Laramide phase), 42 Ma (the Pyrenean phase and meso-Alps), 17 Ma (the Styrian phase and Himalayas), 7 Ma (the Apennine phase) and 5 Ma (the Attican phase and Jura Mountains). According to Schwann (1980, 1984), the Alpine-type compressional pulses at 42, 38 and 17 Ma correspond to remarkably similar forms of global events.

These brief observations suggest that the periods dominated by regional subsidence under marine conditions and the outpouring of flood basalts (oceanic basement) are times of dominant crustal compression culminating in the formation of Alpine-type fold belts. In other words, in the context of the scheme of plate tectonics, the main sea-floor spreading events would correspond to phases of compressional stress in the lithosphere and not, as is usually thought, to those of tensional stress. It follows, moreover, that rises in sea level should be indicator of compressional tectonicevents.

Brunn (1986) offered a possible explanation for the generation of great masses of flood basalt material. Sudden decompression following the opening of a crustal rift on a mid-oceanic ridge (i.e., the creation of a transcurrent zone) causes the expansion of volatiles, and the fluidification and rising of mafic magmas. The fundamental problem is one of knowing whether the oceanic ridge could essentially have been caused by tangential compressional stresses. We should note in this respect that Kent (1977) regarded the regional subsidence of the basin margins to be of the geosynclinal type, forming part of a system of simple "miogeosynclinal subsidence". Furthermore, Beck, & Lehner (1975) considered the Mid-Atlantic Ridge to be a recent "geanticline" of global proportions, with flanks that drop off below the elongated marginal geosynclines (i.e., the Atlantic continental margins).

The phases of general submergence are followed by periods of global climatic amelioration. Warm conditions are indicated, for example, by peaks in the abundance of silica (in associations of chert) and by the presence of broad shelf areas composed of large amounts of marine limestone, calcareous shale and dolomite. The warm-climate floras and faunas of these periods are characterized by high biotic diversity and major advances in evolution. The world-wide presence of comparatively warm climates is confirmed by low oxygen-isotope values obtained for the ocean basins (see, for example, Kennett, 1983; Kennett & von der Borch, 1985; Miller & Kent, 1987).

That the Cenozoic climatic optimum was reached during the Early Eocene (see Fig. 1 of Miller & Kent, 198) is demonstrated by, among other phenomena, extensive precipitation of carbonates on shallow continental shelves covered by warm seas, which led to the widespread development of nummulitic limestones. A second climax of warm conditions, with smaller dimensions than the first, occurred about 17 Ma before present in the Late Burdigalian (Kennett & von der Borch, 1985). In summary, the periods during which sea level generally rises are marked by phases of climatic amelioration, high biotic diversity and major evolutionary progress.

During the periods of continental submergence the frequency of geomagnetic reversals tends to decline, as is shown very clearly by the curve presented by Mazaud et al. (1983). Diminution of the number of reversals to approximately one per Ma occurred during the Cenozoic at the end of the Lower Eocene in correspondence with the maximum sea level attained (Fig. 1). Other decreases in the frequency of reversals (in the Upper Eocene, Burdigalian and Lower Pliocene) also correspond to relative rises in sea level in the curve published by Haq et al. (1987). Thus, the phases of transgression occurred during periods of predominantly normal polarity. Vogt (1975) related periods of infrequent reversals with weakness in the magnetic field and smoothness in the fluid core convection (i.e., few perturbations at the interface between core and mantle). Smooth and more laminar convection is a result of the relative coolness of the core with respect to the mantle

(Sheridan, 1987).

Although they are incomplete and rather simplistic in character, the few geological data which are presented here appear to show that the periods of great tectono-magmatic activity which created the oceanic basins were dominated by a powerful horizontal compressive stress field which acted within the crust. The compressive stresses were both intensive and extensive enough to cause large-scale shearing (see Hast, 1969). Flood basalts were extruded through deep shear fractures.The result was the occurrence of immense vertical shear belts associated with phases of lateral expansion of the network of mafic intrusions and basaltic flows (representing "oceanic basement", or "layer 2"), both within and on top of an old continental crust that had been "orogenized" (Wezel, 1992). Small oceanic basins, Atlantic-type continental margins, rift lineaments and fold belts are crustal structures that the writer regards as "mobile megashear belts" (Wezel, 1988, 1992), meaning deep, thermally weakened zones in which the lithosphere has undergone intense tectonic deformation.

The global stress field in the Earth's crust has changed its direction and magnitude during geological times. For example, Bergerat's (1987) analysis of brittle deformation in the European platform during Cenozoic times has indicated that there were four palaeostress fields (see Fig. 1):

(1) N-S compression during the Late Eocene;

(2) E-W extension during the Oligocene;

(3) NE-SW compression of Early Miocene age; and

(4) NW-SE and NNW-SSE compressions dating from the Late Miocene and thereafter.

Every change in the orientation of stresses caused the mobile shear zones to be reactivated, thus imparting to them a complex and polycyclic geological history. Their long process of evolution has depended on many factors besides simple tension and compression. Hence the lithosphere appears to be traversed by wide zones of vertical shearing, which are very ancient and which undergo periodic reactivation. This is manifest in the form of horizontal movements in which shear stresses are released as the Earth contracts under the influence of polar compression (Wezel, 1988, 1992).

The writer has previously drawn attention to a remarkable analogy between the phenomena described above and the circulation of the Earth's atmosphere (Wezel, 1988, 1992). The formation of back-arc marginal basins occurs by a series of stages which resembles those in the life-cycle of a frontal cyclone. This "cyclogenesis" creates an incipient shear zone which undergoes perturbations, giving rise first to 'waves' and then to 'vortices'. As shown in Fig. 9 of Wezel (1992), the cross-section through cold and warm fronts in a mature cyclone recalls the section through a back-arc marginal basin delimited by the old Alpine-type zones and the younger Benioff zones.

3. Times of Disturbed Magnetism: Periods of Continental Emergence and Cold Climates

During the Cenozoic, regions of widely differing tectonic settings have undergone periods of large-scale epeirogenic uplift and subsequent erosion. The highest topographic elevations have been created during the Quaternary, and rates of uplift and erosion have increased substantially in the Himalayas since the Early Pliocene. In the stratigraphy of the molasse deposits of the Upper Siwalik Group along the southern margin of the Himalayan range conglomeratic fuvial facies first appeared about 2.5 Ma ago (Raynolds & Johnson, 1985). On the Tibetan Plateau, Late Miocene-Early Pliocene subtropical pollen assemblages indicative of elevations below 1.5 km are found today at altitudes in excess of 4 km (see Raymo et al., 1988). According to Gansser (1982), it seems reasonable to assume that the Himalayas have been uplifted more than 4 km since the beginning of the Pleistocene. Similar major phases of mountain uplift occurred in Plio-Pleistocene times in the Andes and their Altiplano zones (Allmendinger, 1986; Benjamin et al., 1987) and in the East African plateaus (Smith, 1982). The principal Cenozoic uplift of southern Africa (covering about 1200 m) began at the end of the Pliocene and continued into the Quaternary (Siesser & Dingle, 1981).

The essential point is that these phases of considerable vertical uplift occurred more or less simultaneously in different parts of the world, creating much of the land that is currently elevated in various continents. It is also important to bear in mind that such uplift took place during a geological epoch that was not marked by significant compressional or strike-slip movements (Gansser, 1982). Gansser (op. cit.) considered a peak in metamorphism and related acid plutonism to be the catalysts of the most important morphogenetic event. Morphogenesis was, however, not limited to the Pio-Pleistocene, but can be assumed to have occurred during other periods of the Cenozoic.

King (1983) hypothesized that phases of elevation were active over whole continents at the end of both the Oligocene and the Miocene, and at the Pliocene-Pleistocene boundary. The importance of post-paroxysmal uplift during the Oligocene in both the Himalayas and the Alps is indicated by the almost synchronous initiation of thick molassic sedimentation about 35 Ma ago (see Trümpy, 1980; Reynolds et al., 1983). The Eocene-Oligocene boundary marks the point of change from flysch to molasse, representing the point of transition from compression and subsidence to extension and uplift. A second pulse of uplift occurred during the Middle Miocene, as is indicated in the Alps and Himalayas by the influx of coarse clastic sediments representing molassic deposits. Uplift during the Oligocene produced a relief (in the Bergell alpine massif and the Trans-Himalayas) which exceeded the present height of about 6 km (Gansser, 1982). In the

Himalayas, the amounts of uplift which occurred during the Middle Miocene and Plio-Pleistocene are vividly illustrated by the presence of molasse deposits about 18 km thick, comprising the Murree, Siwalik and Karewa formations and the Indo-Gangetic alluvium (Mehta, 1980).

Concerning the recent uplift of ancient mountain chains (such as the Caledonides of Scandinavia, the Appalachians and the Ural Mountains) which has been regarded as enigmatic, it seems difficult to invoke isostatic rebound (Ayrton, 1987). One may add that global epeirogeny, raising granitoid plutons to elevations which surpass those of the present-day mountain ranges, could well increase the Earth's radius (Wezel, 1988).

The timespans covering topographic elevation, high relief and continental denudation correspond to periods of global fall in sea level and matching increase in land area, representing phases of "continentalization". In the eustatic curves of Haq et al. (1987), the minima of sea level appear in the Late Pliocene and Early Pleistocene, near the Serravallian-Tortonian boundary (10 Ma ago), in the mid-Oligocene (30 Ma ago) and in the Late Paleocene (58 Ma ago). The overall falls in sea level coincide with times of widespread hiatuses in world oceanic sedimentation (e.g., Moore et al., 1978; Vail et al., 1980). They also coincide with periods of global cooling, as shown by the oxygen isotopes derived from marine sediments, which have provided strong evidence of major cooling events, in decreasing order of importance, at about 0.9, 2.5, 10, 13-16, 24-25, 31-32 and 36-37 Ma (see Kennett, 1983; Kennett & von der Borch, 1985; Miller & Kent, 1987). The entire Cenozoic has been distinguished by climatic deterioration culminating in ice ages.

During these periods the geomagnetic field was characterized by an increase in the number of reversals (reaching a maximum of six per million years at about 9 Ma ago: see Fig. 1). As mentioned above, this may indicate high degrees of turbulence in core convection, that the core has been hotter than the mantle, and that the magnetic field has strengthened (see, for example, Sheridan, 1987). According to Aparin & Vedenkov (1975), the increase in frequency of geomagnetic reversals may have resulted from growth of the Earth's core.

At the end of this process, the biotic assemblages were characterized by acceleration of the rate of extinctions, decreases in the diversity of genera and restriction of the geographical location of species. Sepkoski (1986) recognized mass extinction events in the Late Eocene and Middle Miocene. Large, shallow-water foraminifera became extinct, deep-sea ostracods altered in character and new foraminifera developed.

Terrigenous sedimentation occurred in reduced shelf areas and there was a large-scale transfer of carbonates to deep waters. In back-arc basins, debris flow deposits and submarine fan systems tended to accumulate during times of the maximum tectonic uplift of their source areas (Klein & Lee, 1984, Table 1). With reference to the prevailing depositional systems, these periods of uplift were as follows: the Middle Eocene (e.g., the Daito Basin), the Late Oligocene (e.g., the West Philippine Sea), the Middle Miocene (e.g., the South Fiji Basin) and the Late Pleistocene (e.g., the Sea of Japan). In general, higher rates of turbidite deposition preserved in the sedimentary record correlate with high rates of uplift of their source areas.

In summary, the regime of continental emergence and regression described above is characterized by phases of vertical upheaval, horizontal extension, intrusion, regional metamorphism, granitization, a high frequency of geomagnetic reversals, increases in the solid load of rivers and deterioration in climate.

4. Towards a Synthesis of Current Knowledge

The Cenozoic has a polyphase tectono-metamorphic history, characterized by recurrent, world-wide pulses of compressive deformation, separated by periods during which compressive stresses fell off and were replaced by extension. The pulses of crustal compression seemed to affect the entire globe, creating mobile belts which appeared in the form of megashear zones (Wezel, 1988, 1992). These gave rise to continental fracturing, tectonic "megabrecciation", rapid subsidence of basins, sinking into the mantle through a vortex motion of sections of continental crust, high rates of accumulation of sediments, crustal shortening of hundreds of kilometres, high-pressure metamorphism, genesis of a vertical orogenic root, folding and thrusting. Global compression was very active during the Late Eocene, Early Miocene and Early Pliocene.

Studies of the reorientation over time of the Earth's systems of stress (e.g., Bergerat, 1987: Fig. 1) suggest that the compression tended to be orientated approximately N-S, NW-SE or NE-SW. This led Wezel (1988) to propose a new alternative hypothesis of global dynamics, in which these tectonic deformations are the result of impulses of polar flattening and equatorial bulging, which may have been produced by an increase in the Earth's rate of spin. Cyclic changes in speeds of rotation would explain parallel variations in the state of the geomagnetic field: in fact, rotational changes will generate changes in the magnetic field (see, for example, Mörner, 1984).

These "revolutionary" periods dominated by compression alternate with time intervals characterized by substantial topographic uplift and the genesis of morphological surfaces defined by concordant summits and, moreover, by the occurrence of broad, elevated plateaux. The existence of uplifted highlands is corroborated by the presence of coarse conglomerates and by increases in the solid load of rivers. During these phases of global uplift, high-pressure rocks (such as eclogites) re-emerge at the surface: thus, rocks of continental origin that acquired their high-pressure metamorphism at depths of up to 100 km have been recognized in several parts of the Alps

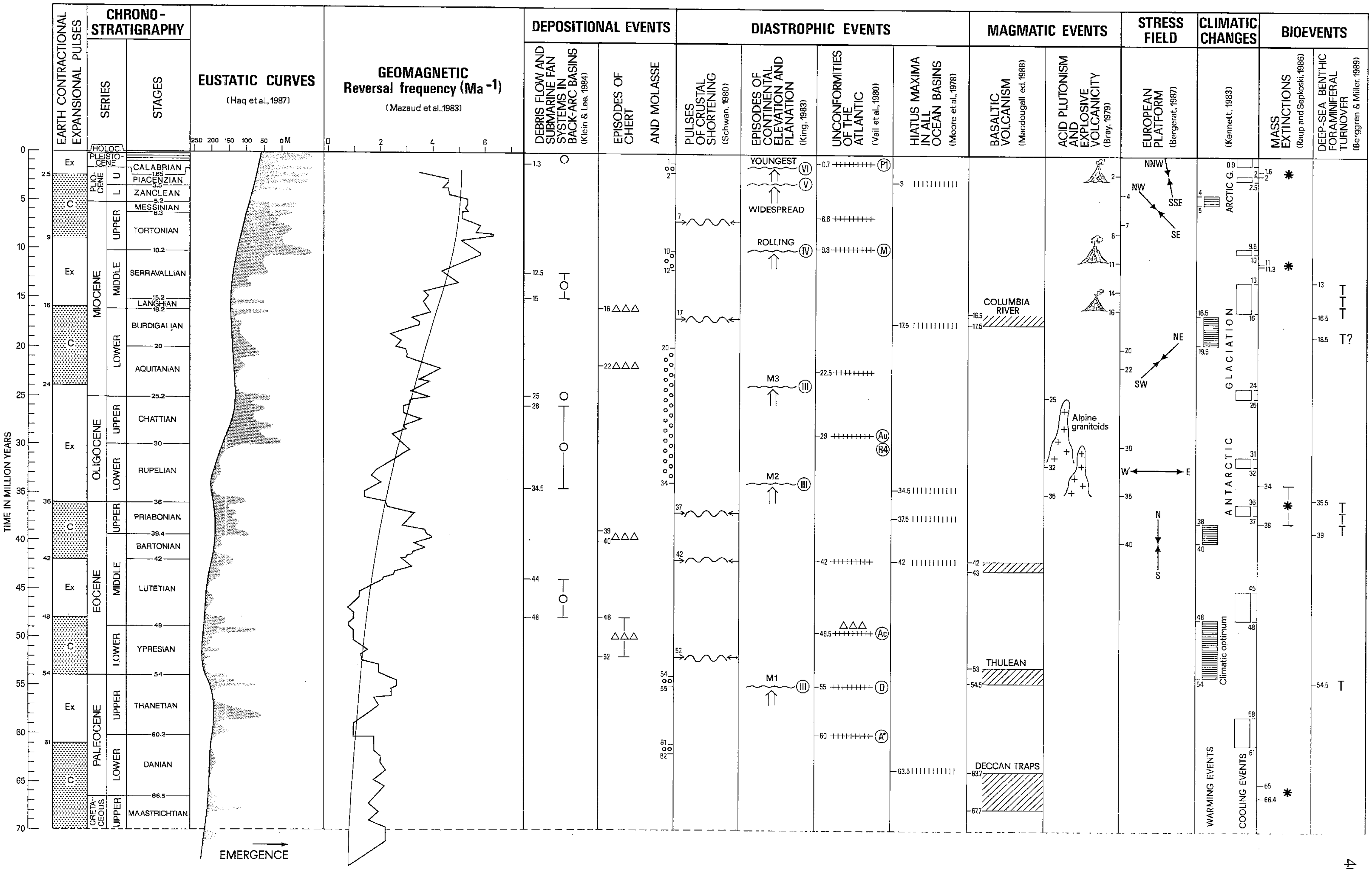

Fig. 1 - Synoptic timetable of principal depositional, diastrophic, magmatic, intraplate stress, climatic and biotic events during the Cenozoic history. Such correlative events are in phase with the proposed global-regional episodes of crustal contraction (C) and crustal expansion (Ex), allusive of a global rhythm of the mantle convection. Note the suggestion of correlation between global sea-level fluctuations and geomagnetic reversal frequency.

(Laubscher, 1988, 1990).

Tectonic deformation is essentially extensional in form and is accompanied by granitic intrusions. Rapid, generalized uplift occurred in a substantial way in the Oligocene, the mid-Miocene and the Pleistocene. In the case of the Alps, Laubscher (1988) identified an "Oligocene lull" before a "Miocene revolution", characterized by extension and acid intrusion. Thus, hot rocks undergo rapid vertical elevation during periods dominated, not by compressional or significant strike-slip movements, but by post-compressional pressure release and extension. The writer believes that expansion in the mantle is caused by a diminution of the polar flattening of the Earth resulting from a decrease in the rate of rotation. It seems unlikely that isostatic uplift is a significant cause of the elevation of mountains, considering the enigmatic case of the recent uplift of ancient mountain chains, such as the Caledonian orogen of Scandinavia, the Appalachians and the Ural Mountains (see Ayrton, 1987).

Global regression of the seas, substantial general elevation of the continents and deposition of detrital molasse are elements which can be explained by mantle upwellings accompanied by heating of the lithosphere and the crust. When heated from below by the widespread thermal anomaly in the mantle, continental crust will produce melts of the granite family (Fyfe, 1987). The associated heat fronts cause regional metamorphism, anatexis and "granitization". Below many plateau-like uplifts the mantle possesses anomalous features (lowered seismic velocities and increased attenuation of seismic waves) which can be interpreted as resulting from the upward migration of columns of hot material (see Zorin & Florensov, 1979). Portions of heated material of the anomalous mantle may penetrate and intrude into the crust giving rise to volcanicity and local anomalies in heat flow. In short, the process of morphogenic uplift seems to result from zonal upwelling in the asthenosphere. In contrast, basin subsidence and mountain belts overlie zones of downwelling convective flow in the mantle (Molnar, 1988).

The history of the Cenozoic appears to consist of alternating periods of continental "deflation" and buckling, interspersed with those of 'inflation' and tension, accompanied by plutonism and explosive volcanism. In the case of back-arc basins, subsidence events are sometimes marked by a cyclonic-type "down-sucking" which can be ascribed to powerful lateral compression. As the compressional stresses are global in extent, one is led to hypothesize (Wezel, 1988, 1992) that global contraction results from polar flattening which may be a function of acceleration of the Earth's spin rate. The ensuing increase in tectonic stresses may produce rapid regional (geosynclinal) subsidence and the accumulation of sediments in mobile megashear belts.

Periods of rapid "sea-floor spreading" correlate well with those of tectogenesis and fast rates of subsidence in cratonic basins (see Bally, 1980). Thus, they correspond to time intervals in which the lithosphere has undergone regional compression rather than tensional stresses (see De Rito et al., 1983). Sea level changes also indicate world-wide tectonic events. Compressional pulses cause syndiastrophic flysch sequences to be deposited, buckling to take place during large-scale folding, and plastic flow to occur as a result of tectogenesis. Mafic volcanism and intrusion occur episodically at times of crustal foundering. The phases of mafic activity coincide approximately with periods of the accelerated subsidence of basins, as manifested by widespread marine transgression or by the vigorous accumulation of terrigenous sediments in ocean basins. Flood basalts mark the early stages of regional (geosynclinal) subsidence and foundering of the Atlantic-type continental margins (Jordi & Lehner, 1973). Tethyan ophiolite terrains constitute the basement of synorogenic flysch sequences. Studies of their chemistry and mineralogy suggest that many classic ophiolites are more closely related to an arc setting than a mid-ocean ridge one (see McCulloch & Cameron, 1983). In short, the mobile belt associations, such as ophiolites and flysch, are here considered products of regional compression and shear.

The creation of fold belts is followed by fundamental changes in terrestrial dynamics and in the tectonic framework governing the distribution and behaviour of tectonic regimes. In post-paroxysmal times, general uplift and erosion take the place of subsidence and sedimentation, crustal extension replaces compression, "granitization" is substituted for "basaltization", the global retreat of the seas replaces world-wide transgression and the accretion of continental crust takes the place of its destruction. Granitoids and molasse replace ophiolites and flysch. The general vertical uplift of continents with respect to sea level results from a wide-spread thermal event which introduces heat into the lithosphere and thus causes metamorphism, acid plutonism and volcanism. Rapid, large-scale uplift of this kind occurs contemporaneously in continental areas characterized by very varied tectonic settings (e.g. the Andes, Colorado Plateau, East African plateaux and Himalayas). This global vertical tectonism implies that continental inflation occurs and was possibly caused by deceleration of the Earth's rotation (Wezel, 1988). Pulsatory expansion would thus result from the release of pressure during anorogenic, quiescent (or morphogenetic) periods.

Uplift forces major changes to occur in the atmospheric circulation, with a profound impact upon temperature, precipitation and wind patterns at the Earth's surface. Such changes in terrestrial and marine climates favour a general process of cooling. Experiments on uplift conducted by Ruddiman & Kutzbach (1990) show that topographic elevation is an important factor in climatic deterioration and the occurrence of ice ages.

5. Some Observations on Synchronous Fluctuations in the Ocean Basins

The global fluctuations discussed above are clearly manifest in the oceanic areas, as well as the continents. In the context of the plate tectonic theory, phases of terrestrial expansion should correspond to periods of slow spreading. Times of slowdown in spreading rate were considered to occur during the Paleocene, Oligocene, Middle Miocene and Upper Pliocene-Lower Pleistocene (see Cande et al., 1988, for a South Atlantic example). The intervals of slow spreading also correspond to increases in the density of lineaments in fracture zones, to a greater relief in topographic profiles and to evidence of significant crustal elevation. We may thus cite the example of uplift during the mid Eocene in the Rio Grand Rise (Barker & Carlston, 1983). During the Cenozoic the South Atlantic has been characterized by a general diminution in the rate of spreading, accompanied by an overall increase in the density of fracture zones (see Cande et al., 1988), which accords well with the current regressive tendency of our planet.

The general trend towards expansion and regression is pulsatory in character and is interrupted by periods of global contraction and polar flattening. In oceanic areas these latter correspond to times of increase in the rate of spreading. Such increases occurred in the South Atlantic, in the late Middle Eocene-Upper Eocene, the Lower Miocene and Upper Miocene-Lower Pliocene (Cande et al., 1988). In other words, temporal variations in spreading rates appear to be synchronous with the global pulses. Correlations therefore exist between highstands of sea level and episodes of fast spreading, and between lowstands and slow spreading rates.

This phenomenon of pulsation in the entire terrestrial body is also reflected over time in the curve of average global calcite compensation depth (CCD) during the Cenozoic (Delaney & Boyle, 1988). This shows a general tendency towards deepening interrupted by peaks of shallowing around 40 Ma and 20 Ma. The principal peaks of deepening (about 30 Ma and at present) correspond to phases of global expansion. Since the Lower Pliocene the CCD has deepened by approximately 350 m.

It should also be borne in mind that erosional events (hiatuses) are contemporaneous in both shallow-water and deep-water depositional settings. For example, four such hiatuses occurred during the Paleogene (near the Lower-Middle Eocene boundary, within the Middle Eocene, in the early Upper Eocene and within the Oligocene) of the New Jersey coastal plain (according to data from the ACGS4 borehole). Miller et al. (1990) correlated these with similar hiatuses that have been observed in the deep-sea record at DSDP sites. Such inter-regional synchrony reinforces the hypothesis that a global phenomenon is at work. But the fundamental problem is one of understanding the causal mechanism which produces erosion of the deep-sea sediments, that is, of whether it is a question of eustasy (global fall in sea level) or basin tectonics (uplift). Some of the stratigraphic gaps found in deep water sediments correspond with either regression or tectonic events on land. For example, hiatuses were most abundant with respect to all ocean basins during the Upper Eocene (Moore et al., 1978) in correspondence with the world-wide Pyrenean pulse of crustal shortening (Schwann, 1980). Another maximum in the frequency of hiatuses found in pelagic sediments of the world's oceans occurred at the same time as the Styrian phase of the Himalayan orogeny (about 17 Ma ago, during the Late Burdigalian). Moore et al. (1978) described two other maxima at approximately 63.5 Ma (Lower Paleocene) and about 3 Ma (during the mid-Pliocene), which correspond to phases of regional tectonic compression on land. It thus seems that during the Cenozoic the occurrence of major hiatuses in deep-sea sediments has had tectonic causes.

It is, moreover, essential to bear in mind the new interpretations of the Earth's crust offered by Choi (1987) and Choi et al. (1990). According to various lines of evidence (dredgings, DSDP drilling holes, seismic stratigraphy and palaeogeographic data), Choi believes that the so-called "oceanic crust" (particularly "layer 3") occurring under the deep abyssal plains of the northwestern Pacific is composed mostly of continental rocks. The continental lithologies, which range in age from Precambrian to Mesozoic, consist of orthoquartzites, gneisses, granodiorites, gabbros and crystalline slates. The true composition of the present "oceanic layer 3" is considered to be submergent remnants of a large Palaeozoic-Mesozoic landmass on the rim of the present Pacific basin. Choi et al. (1990) have terme this the "Great Oyashio Palaeoland". This emergent landmass fed sediments into the Tethys Sea throughout the Paleozoic and Mesozoic, including conglomeratic clasts of Permian granodiorite, Proterozoic orthoquartzite and Carboniferous-Permian limestones. It started to submerge in Late Jurassic and Cretaceous times after basalts had been extensively outpoured. One must note that the Early Cretaceous magnetic anomalies are located in correspondence with this ancient landmass. This casts doubt on the validity of using the magnetic anomalies to determine the age and rate of spreading.

Such data call into question the very distinction between what is termed oceanic and continental crust. According to Orlenek (1986; *fide* Dickins et al., 1992), the view that there are two types of crust - continental and oceanic - has no factual basis. In point of fact, Choi's interpretations lead to the conclusion that oceanic basalts cover foundered remnants of continental blocks and are not simply new sections of crust resulting directly and entirely from sea-floor spreading.

6. Conclusions

All geological, climatic and biotic phenomena may be considered as manifestations of a continuous oscillation between two tendencies in the Earth's crust, namely contraction and expansion. During the course of the Cenozoic there were five second-order cyclic patterns of crustal deflation and buckling followed by inflation and tension, all of which has had a mean periodicity of about 14 Ma. Such pulsations, which are identified in continental areas, can also be picked out in oceanic regions.

The revolutionary "systaltic phases" of the cycles are characterized by episodes of general submergence and collapse of sectors of the continental crust, preceded and accompanied by the outpouring of enormous volumes of ophiolites and flood basalts, such as the Deccan traps and Thulean basalts (Fig. 1). After the mafic outpouring, the continental sector began to founder and to be engulfed under the influence of strong tectonic compression. Extremes of horizontal pressure produced folding and thrusting, created vertical deep root zones and resulted in high-pressure metamorphism (for example, during the Late Eocene Pyrenean and Middle Miocene Styrian phases of crustal shortening). These compressional periods of orogenic restructuring were also distinguished by general rises in sea level, climatic amelioration and a geomagnetic field of predominantly normal orientation. They appear to have resulted from the existence of a compressive global stress field, produced by flattening of the Earth's polar areas under the duress of acceleration in its rate of spin (Wezel, 1988, 1992).

Reorganization and reorientation of the global crustal stresses occurs at a turning point when the peak of compression is reached. Pulses of compressional deformation are replaced by "diastaltic impulses" of global decompression and relaxation. These phases are marked by granitoid intrusions, destruction of the deep roots, uplift of regions of land, dominant erosion and extensional tectonics. The importance and magnitude of these processes of uplift is attested by the re-emergence at the surface of eclogitic continental rocks formed at depths of about 100 km. Apart from causing extensive erosion (associated with the deposition of molasse) and tectonically-driven denudation (by gravity slides), the topographic uplift of wide sectors of the globe resulted in the formation of major planation and other morphological surfaces. These periods of "continentalization" (for example, that which occurred during the Oligocene) are characterized by global falls in sea level, deterioration in world climate and finally in ice ages and reversal of the geomagnetic field. The general process of decompression can be related to reversal of the process of polar flattening as a result of deceleration in the Earth's rate of spin (Wezel, 1988, 1992).

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ACTIVE TECTONICS IN THE SOUTHERN APENNINES: RELATIONSHIPS BETWEEN COVER GEOMETRIES AND BASEMENT STRUCTURE. A HYPOTHESIS FOR A GEODYNAMIC MODEL

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Abstract

The aim of this paper is to give an explanation of the basic outlines of the basement of the Southern Apennines on the grounds of geological and geomorphological surface observations, mainly combined with seismic and seismological data. All these data point out the role of the active tectonics at different structural levels in the Quaternary deformational history.

Finally, the authors formulate a hypothesis for a geodynamic model, based on some postulates and on the analysis of constraints on a regional scale.

The South-Apennine chain is an african-verging nappe edifice built on during the Neogene tectogenesis. Structural units result mainly by the deformation of palaeogeographic domains belonging to the African continental margin. All those terrains are strongly affected by Quaternary tectonics as well.

In order to formulate a geodynamic model for the active tectonics in that part of the chain, an application of the concept of regional stress field inversion is proposed and discussed.

Key Words: *Active Tectonics, Geodynamics, Southern Apennines.*

1. Regional Outlines

Geological basement does not outcrop in the Southern Apennines. Only its lowest terrigenous cover ("Verrucano") has been reached by drilling in the foreland zone. As a matter of fact, the basement in Southern Italy is covered by sedimentary rocks, 8 to 15 km thick. The crystalline nappes originated by basement peeling, outcropping in the Calabrian Arc (AMODIO MORELLI et al., 1976), are indeed rootless and not related to the present deep structure. Therefore the crustal framework of the south-apenninic chain is investigated normally by means of classic geophysical methods (CORRADO et al., 1974; MORELLI et al., 1975; CASSANO et al., 1986; MOSTARDINI & MERLINI, 1986).

In this paper we try to explain the basic outlines of the basement of the Southern Apennines on the grounds of geological and geomorphological surface observations, mainly combined with seismic and seismological data, pointing out the role of the active tectonics at different structural levels in the Quaternary deformational history. In the end, we formulate a hypothesis for a geodynamic model, based on the analysis of constraints on regional scale.

The south-apenninic chain is an african-verging nappe edifice built on during neogenic tectogenesis (fig. 1). Structural units result prevalently by the deformation of palaeogeographic domains belonging to the African continental margin and, in a subordinate amount, forming the covers of Tethyan oceanic floors.

Detachment and emplacement of all those terrains occur in the Miocene, Pliocene and Lower Pleistocene (ORTOLANI, 1978; D'ARGENIO et al., 1986; BALDUZZI et al., 1982; CASNEDI et al., 1982).

During the Pleistocene, the post-nappe tectonic history of the chain leads to the genesis of an internal Tyrrhenian belt, strongly subsiding, and of an external apenninic sector, roughly parallel to the first one, affected by progressive uplift from west toward east, with offsets about of 1000-2000 meters. In the same period, counter-apenninic regional transcurrent faults are generated (INCORONATO et al., 1985; ORTOLANI & PAGLIUCA, 1988). Quaternary tectonic evolution of the Southern Apennines is characterized by a common and marked uplift controlled by two main longitudinal structures.

The first element is located along the eastern margin of the chain, while the second one follows the alignment Praia-Sapri-Mignano Montelungo. Moreover, the chain presents a central fracture zone which axially lies according to the alignment Crati Valley-Alta Val d'Agri-

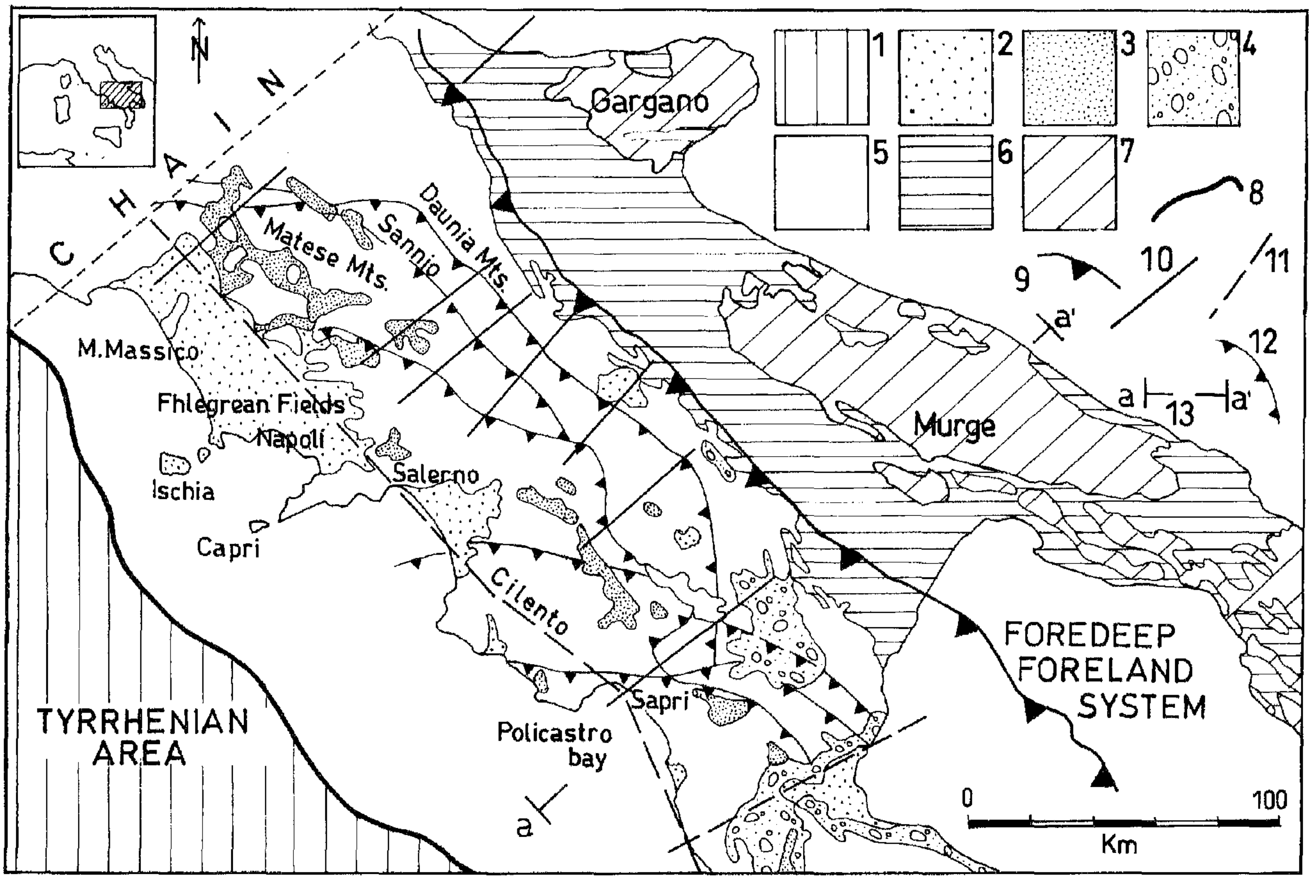

Fig. 1 - Structural sketch map of Southern Italy. Legend: 1. Area with Tyrrhenian-type crust; 2. Coastal plains with Quaternary clastic rocks and volcanics; 3. Intra-apenninic Pleistocene basins arranged along the central fracture zone of the chain; 4. Pliocene basins; 5. Meso-Cenozoic tectonic units; 6. Foredeep Plio-Quaternary terrains; 7. Foreland Meso-Cenozoic formations; 8. Tyrrhenian area-chain boundary; 9. Buried front of the nappes; 10. Counter-apenninic regional faults; 11. Praia-Sapri-Mignano Montelungo (longitudinal) alignment and Sangineto (trasversal) line; 12. Mio-Pliocene thrusts; 13. Trace of the cross-section (fig. 4).

Lioni-Boiano-Alta Val di Sangro (fig. 1).

During the uplift phase, the counter-apenninic brittle structures are reactivated on a regional scale, bordering on sectors with differential vertical displacement. Sometimes the faults of transversal systems show left-lateral strike-slip movements followed by inversion of horizontal displacement.

The alignment Matese-Ofanto and Sele high valleys-Alta Val d'Agri marks the zone of the Southern Apennines which presents the highest values of seismic activity.

2. Tectonic Framework

The south-apenninic sector bordered by Volturno and Noce valleys is divided into active kinematic blocks, separated by tectonic lineaments which have longitudinal and transversal directions as regards to the axis of the chain (fig. 1).

The main transversal (counter-apenninic) faults recognized in that region are the "Roccamonfina-Isernia", "Benevento-Buonalbergo", "Parolise-Grottaminarda", "Bagnoli Irpino-T.te Calaggio", "S.Fele-Vulture", "Satriano-Tito-Pietragalla" and "Golfo di Policastro-S.Arcangelo" lines (ORTOLANI, 1974; INCORONATO et al., 1985). Some of these lineaments cut all the chain - as the "Roccamonfina-Isernia", "Benevento-Buonalbergo", "S.Fele-Vulture" and "G. di Policastro-S.Arcangelo" lines - while other faults trend toward east from the central fracture zone (fig. 1).

Relationships among counter-apenninic regional faults, Quaternary volcanism and seismicity, indicate that the first ones reflect on the surface the existence of important crustal discontinuities (ORTOLANI & PAGLIUCA, 1987; DI GIROLAMO et al., 1988; ORTOLANI et al., 1990).

The interactions between axial and transversal active faults have produced locally structural patterns showing transtensive or transpressive deformational systems. In this way, pull-apart lacustrine basins and morphological anomalies due to the rising of pressure

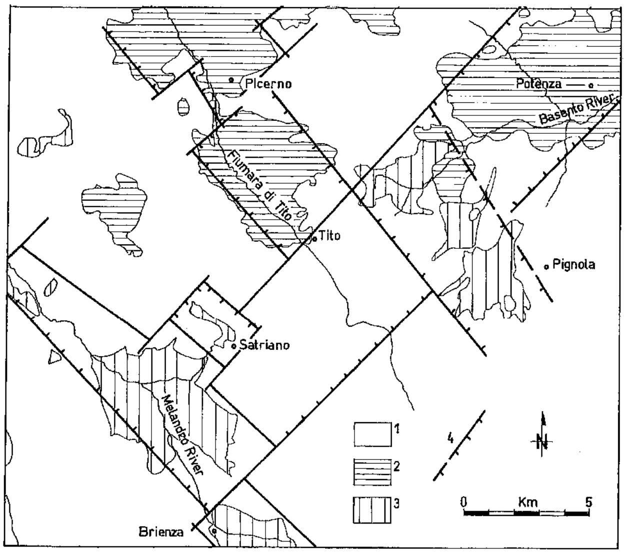

Fig. 2 - Apenninic and counter-apenninic faults in the area between Brienza and Potenza (Lucania Apennine). Legend: 1. Meso-Cenozoic bedrock; 2. Pliocene basins; 3. Quaternary basins; 4. faults.

ridges have been produced during the Quaternary (fig. 2).

Pleistocene axial-trending structural associations are characterized by listric antithetic couples of faults, emplaced in correspondence of deep discontinuities, as shown by recent earthquakes (BOSCHI et al., 1990). Yet, the faults of the cover units don't represent the formal prosecution of the basement structures toward the superficial terrains, but only the mechanical response of the upper structural levels to the deep deformation (fig. 3).

This behaviour could be achieved by means of crustal bending with great curvature radius, sharply truncated by the geosuture marked at the surface by the chain-foredeep boundary. In a bidimensional model, the neutral line of the folding could lie under the top of the basement rocks. Therefore, the upper part of gently curved crust (sedimentary covers and basement top) would be affected by extensional features consistent with the general compressional stress field acting in the deep crust (this is just a possible hypothesis, since no compressional features younger than the Emilian have been noticed up to now in the Southern Apennines).

When the tectonic collapse of the basement top occurs, the transmission of the geometric pattern at the shallower levels is achieved by the interposition of structural horizons which accommodate plastically the deformation.

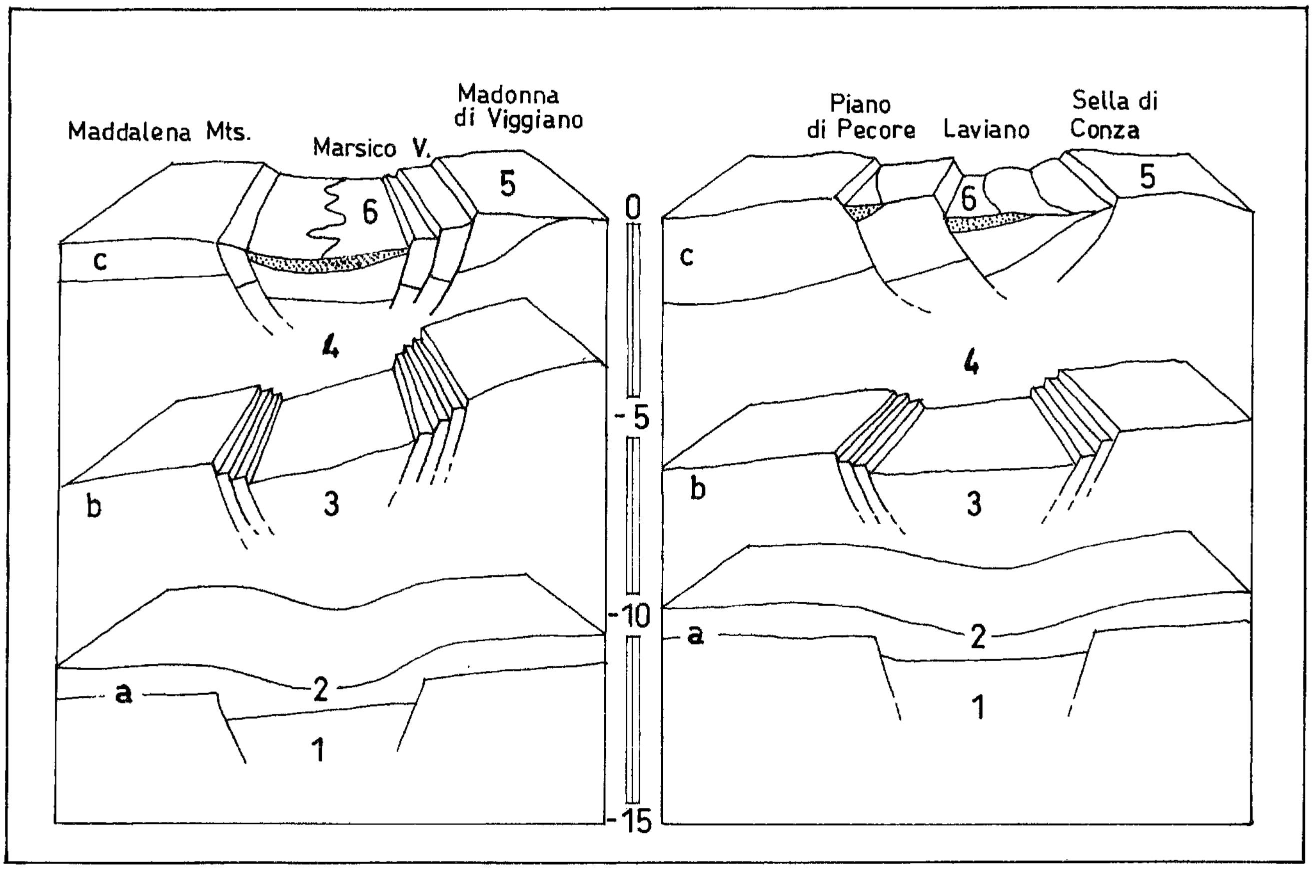

Fig. 3 - Block-diagrams showing inferred relationships between cover and basement structures (the scale is expressed in kilometer). Legend: a, b and c. Brittle levels of the tectonic multilayer; 1. cristalline-metamorphic basement; 2. Triassic evaporitic horizon; 3. Lower carbonate unit; 4. Lagonegro Basin pelagic sequence; 5. Upper carbonate unit; 6. Quaternary clastic terrains.

3. Plio-Quaternary Structural Evolution Related to the Basement History

In order to restore main deformational history which affected the Southern Apennines in the Middle Pliocene and Quaternary, a reinterpretation of gravimetric, aeromagnetic, seismic and seismological data (CORRADO et al., 1974; MORELLI et al., 1975; C.N.R., 1985; CASSANO et al., 1986; MOSTARDINI & MERLINI, 1986) is given by a new survey of geological and geomorphological features. This study allows to point out the basic relationships existing between superficial and deep deformations.

The results are partly shown in a main geological section across the chain, located along an area of particular tectonic significance (fig. 4), showing the duplication of the basement.

Since the early Pleistocene (Emilian stage), the following homogeneous zones formed the first order regional framework of Southern Italy (from internal to external domains, i.e. from west toward east):

— *Tyrrhenian area,* delimited eastward by the Pontine Islands Palinuro Seamount alignment, characterized by a thin cover made up prevalently of Tortonian to Holocene sediments supported by crust of continental type; the crustal thickness may vary from 8 to 12 km and the basement top is located at a depth of 5-6 km.

— *Chain,* delimited eastward by the Termoli - Bradano alignment, characterized by continental crust with a thickness of about 25-35 km; the basement top is located at a depth which varies from 14-15 km to 8-9 km (going from west to east) and it supports a very thick sedimentary cover made up of several structural units transported and stacked during Miocene, Pliocene and Pleistocene tectonic phases, covered in turn by both marine and continental Quaternary parautochthonous terrains.

—*Foredeep-foreland system,* delimited eastward by the Gargano - Murge offshore, characterized by continental crust about 30 km thick, whose basement top is located at a depth which varies from 14-15 km (in correspondence with the foredeep) to 8-9 km (in corrispondence with the foreland) and which supports a thick autochthonous sedimentary cover made up prevalently of Meso-Cenozoic carbonates.

Starting from the Upper Pliocene (?) - Lower Pleistocene, while the nappes keep on moving toward the external sectors, roughly NW-SE trending morpho-structural zones break off the former tectonic elements, often modifing their attitudes.

Along the Tyrrhenian margin of the chain are located active subsiding depressions, filled by marine and continental deposits whose thickness changes going from NW toward SE, from 1000 m (Garigliano Plain) to 4000-5000 m (Campania Plain, with at least 50 % of volcanics), to about 2000 m (Sele Plain - Salerno Bay), to about 1000 m (Policastro Bay). Those depressions are interrupted by transversal heights (Aurunci, Massico, Lattari and Cilento Mountains) which at the beginning of Pleistocene reached altitudes varing from about 400 to 1500 m. Along the chain zone, a progressive uplift occurred first along the western sectors and then along the eastern ones. The accentuation of the foredeep (Bradano Valley) continued untill, during

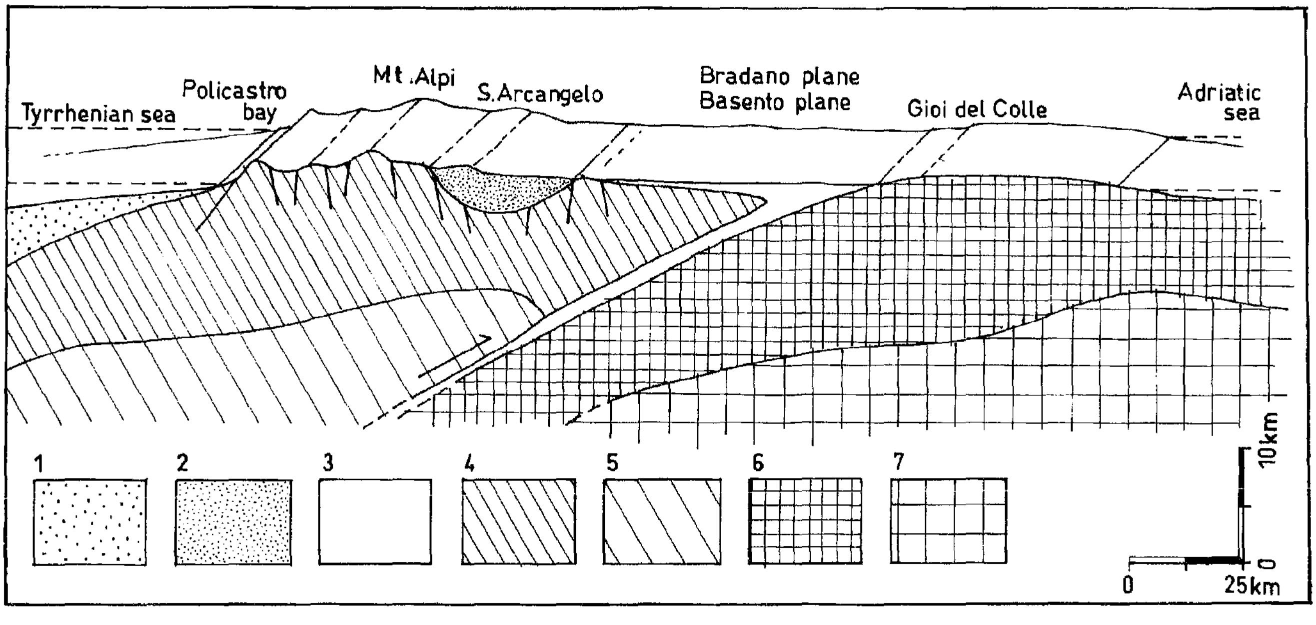

Fig. 4 - Geological cross-section on a regional scale (for location see fig. 1). Legend: 1. Tortonian to present marine sediments; 2. Pleistocene terrains of S.Arcangelo Basin; 3. Plio-Quaternary terrains of the Bradano foredeep; 4. Cover units of the chain; 5. Basement of the chain; 6. Cover units of the foreland; 7. Basement of the foreland.

Lower Pleistocene, its sediments were lifted too, in particular along the most internal (i.e. western) margin, assuming a regional dip toward NE.

Evidences of the recent evolution are well documented by geomorphological, sedimentary and seismological characteristics determined by apenninic and counter-apenninic structures, which also partly re-establish Pliocene tectonic elements (e.g. "Benevento-Buonalbergo" line, as shown by ORTOLANI, 1974).

The faults of the Tyrrhenian margin of the chain have considerable offsets (about 4000-5000 m), displacing Meso-Cenozoic bedrock and permitting sedimentary and volcanic hoards mentioned above. As a matter of fact, during Quaternary, a potassic volcanism with shoshonitic affinity took place, with several eruptive centres arranged along those tectonic lineaments. ORTOLANI & PAGLIUCA (1984, 1987, 1988) supposed for that internal sector a megasynform crustal structure. According to the Authors, the transversal hights would represent deformation of the sedimentary cover only. The differentiated uplift of the chain from west to east was achieved by means of other dominant apenninic faults, separating the western block from the more recently lifted eastern one: that zone is conjectured to be a megantiform crustal structure, broken in the hinge zone. On the surface, along the projection of the crustal hinge zone, several intra-apenninic Pleistocene basins (Boiano, Benevento, Lioni, Agri high valley, Mercure) was formed. Their sedimentation seem to be conditioned by strike-slip faults, whose control is available in the trend of the chain water-parting as well.

Data from regional cross-sections display situations of both agreement and discordance between crystalline basement and sedimentary cover structures, during Quaternary deformation.

The global interpretation results from a comparison among the present tectonic setting and the palaeostructural one at the beginning of Pleistocene. Structural patterns with important geodynamic implications, particularly along the Tyrrhenian zone and the most external sector of the chain, are discussed in the next paragraph. At the Tyrrhenian margin, the direct relationships between crustal deformations and volcanism (see for example the Campania Plain) and several discordances between cover and basement framework (i.e. structural high zones at the topographic surface in correspondence of flat profiles of the basement, as observed for Lattari and Massico Mounts) have been inferred (cf. geophysical data in CASSANO et al., 1986).

Several indications, as arrangement and geochemical characters of the Pleistocene to Holocene volcanism of the Tyrrhenian margin, seismicity patterns recorded in the zone of the chain and uplift rates of Quaternary deposits, suggest the persistence of compressional tectonics in the Southern Apennines responsible for folded structures of the cover, supposed to exist in the basement too ("duplex").

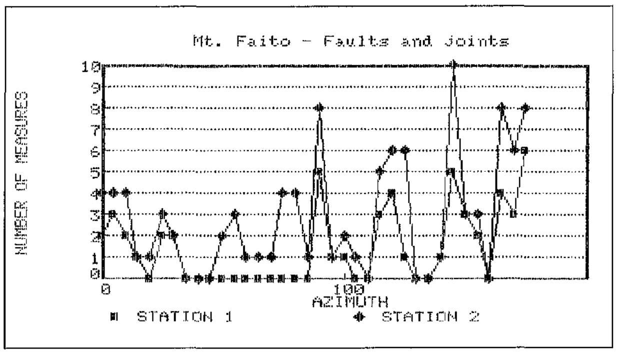

Fig. 5 - An example of brittle pattern (minor faults and joints) from Mt. Faito (Lattari Mts., Sorrento Peninsula), collected on Cretaceous limestones. Several systems are quite recognizable.

4. Geodynamic Model

In order to formulate a geodynamic model for the active tectonics in the Southern Apennines, we take in account the following constraints and postulates:

1) the regional active faults have longitudinal and transversal trends as regards to the axis of the chain, forming among them an angle about of 90°;

2) the two systems are singenetic, or at least sinkinematic, and they show components of recent strike-slip movements;

3) the active stress field works on anisotropic bulks, characterized at least by four basic domains of ancient tectonic lineaments (fig. 5);

4) the active structures in the shallower cover units are genetically related to deep strain fields generated by crustal bending, of which neutral surface lies under the basement top (fig. 3);

5) the faults with meridian and Tyrrhenian trends don't show indications of recent seismic activity.

For simplicity one can consider that the axial and transversal faults are oriented N45W and N45E respectively.

From the points 1 and 2 it results that the two systems are comparable at a conjugate couple with $2\Phi = 90°$, limit-condition for $\Phi = 0$ (absence of internal friction); this theoric state approximates to real condition because of the anisotropy previously acquired (point 3), that involves a decrease of the shear strength of the medium. It follows that the maximum principal stress of the tensorial fields on crustal scale (point 4) can occupy indifferently the bisector position in one of the two complementary dihedral and therefore it can lie in the space with meridian (N-S) or Tyrrhenian (E-W) azimuthal orientation (fig. 6).

Several remarks on geodynamics of the Mediterranean area (BOCCALETTI & DAINELLI, 1982) lead to consider the first configuration as the most probable.

Since from the deformation physics point of view, the second possibility is equivalent to the first one, one can conclude that, in the case of close modular values of $\sigma1$ and $\sigma3$, they could be exchanged through the time.

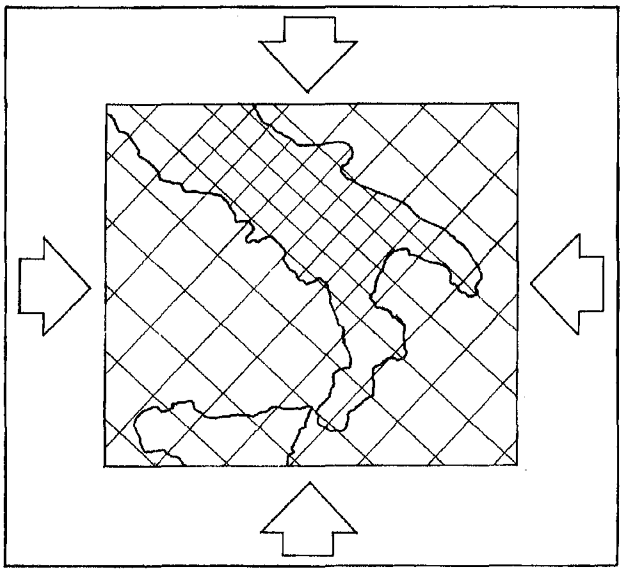

Fig. 6 - Stress-strain inferred model for the Southern Apennines. The grid represents ideally the main fault systems, forming an angle of 90°. The couple of arrows may represent both maximum and minimum principal stress, changing over the time.

This condition is generally reached in high confining pressure systems. In that case it follows that the intermediate principal stress, which has constant vertical attitude, may vary among very close σ values, involving a nearly spherical stress tensor. This fact suggests a progressive isotropization of the crustal stress field as conseguence of a likely attenuation of the thermodynamic upset state in the upper mantle.

During the passage from a N-S trending $\sigma 1$ axis to an E-W one it may be that the vertical principal stress becomes temporarily the maximum one: that is the dynamic condition for the "neotectonic" events which have provoked just the uplift of the blocks. Field observations suggest an Upper Pleistocene age for the latest tectonic inversion in Southern Apennines.

The ancient meridian and Tyirrhenian lineaments don't play a determinative role in the recent tectonics (point 5) since the relative planes contain the maximum and minimum principal stresses of the field hypothesized in the model, showing in this way the only non-active space configuration in brittle deformation. However, the N-S and E-W trending faults may follow passively the kinematics of the NW-SE and NE-SW trending structures, being included in an interconnective framework inherited by the former tetctonic phases.

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EXHUMING AGE AND STRUCTURAL SETTING OF LOWER CRUST WEDGES OUTCROPPING IN ITALY

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Abstract

Lower crust wedges and slices are known in the western Alps (Southalpine Ivrea-Verbano Zone and Austroalpine II Dioritico-kinzigitica Zone), in northern Calabria (Dioritico-kinzigitica Polia-Copanello Unit of Catena Costiera and Sila nappe system) and in north-eastern Sardinia (partly retrocessed granulite slices within the amphibolite-facies metamorphics of the Axial Zone). The two first areas show evidence of strong Hercynian structuring and crustal consolidation followed by weak (Southalpine Ivrea Zone) to severe (Austroalpine and Calabria units) involvement in the Alpine restructuring. The third area was preserved from Alpine orogeny, thus showing a standard case for comparison. The first exhuming age is definitely Hercynian (Devono-Dinantian ?) for the Sardinia slices, and almost probably Hercynian (Devono-Dinantian) for the Ivrea Zone too. Less reliable are the geological age constraints for the Calabride and Austroalpine lower crust wedges. Newertheless, they are usually considered as having been first exhumed through the Hercynian structuring. Theoretically, deep crustal to upper mantle segments can be raised to the surface by means of thrust ramps, up-thrusts, transpressive positive flower-structures, isostatic uplift, etc. Conditions favouring this process include thinned crust, extensional margin settings and crustal denudation. The Italian outcrops of lower crust listed above are discussed in terms of present structural setting and pertinent geological evolution. A point is made strongly relating such events to strike-slip motions during orogeny.

Key Words: *Hercynian, Alpine, orogeny, Alps, Sardinia, Calabria*

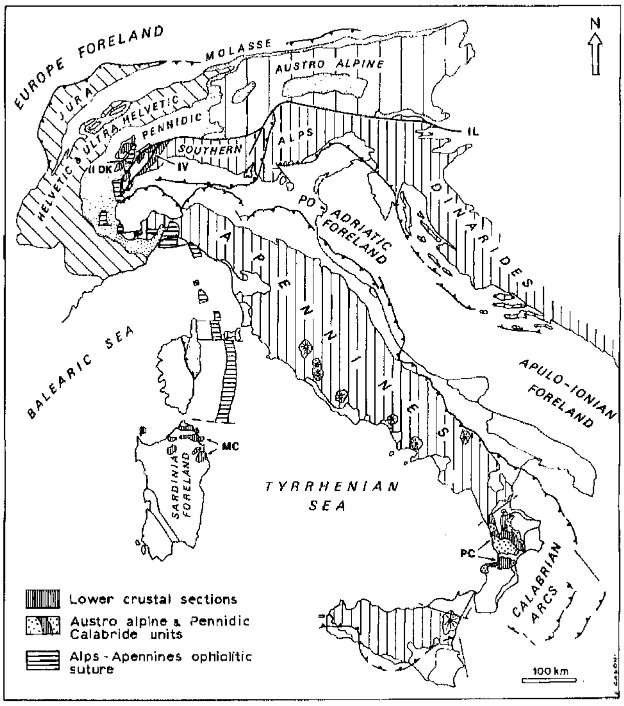

Fig. 1. Schematic structural map of Italy with location of the lower crustal sections dealt with in this paper. (MC) Migmatite complex of NE Sardinia, (II DK) 2nd Diorito-kinzigitic Zone (Austroalpine, western Alps), (PC) Polia-Copanello Unit of N Calabria, (IV) Ivrea-Verbano Zone (Southern Alps), (IL) Insubric Lineament.

Gravitational pull of the dense, cold oceanic crust and lithosphere downdip the Benioff subduction zone accounts for the classic plate tectonic motions along with thermal uprise at mid oceanic ridges since the foundation of the theory. About two decades later, very similar gravitational sinking of the dense, cold lower crust and lithosphere, decoupled from its warmer (Taylor and McLennan, 1985) and light upper part, downdip the Ampferer "Verschluckung" zone, is increasingly accepted. Why should then slabs or remnants of such deep crustal elements attain the surface or upper crustal levels? The question is particularly intriguing as far as the thick and deep seated lower crust is concerned. This fact is mirrored by the relative abundance at the surface of both subducted (HP-LT) (Platt, 1986) and obducted (LP-LT) oceanic to ophiolitic material as compared with exhumed lower crustal to upper mantle sections.

Altogether, despite the increasing number of recent findings (Fountain and Salisbury, 1981), the frequence and absolute extent of exposed lower crustal sections is still so limited that this has to be considered as a low-probability event. However, what are the boundary conditions allowing or even triggering such rare event to occur?

Aim of this paper is 1) to describe synoptically the most important Italian lower crust outcrops, some of which have not been taken into consideration even in the most recent reviews (Windley, 1981; Fountain and Salisbury, 1981); 2) to discuss the problems involved with their formation and obduction ages; 3) to evaluate their European versus African affinity and their Hercynian versus Alpine exhuming age; 4) to comment upon the regional clustering and the possible recurrence in space and time of lower crustal obductions; and finally 5) to discuss the main geologic processes favouring and eventually producing such uncommon settings.

1. Outcropping Lower Crustal Sections

1.1 *The Piedmont Western Alps*

Lower crustal wedges and slices occur in both the Southalpine Ivrea-Verbano (or I "Diorito-kinzigitica") Zone and the Austroalpine II "Diorito-kinzigitica" Zone (Figs. 1-3). Lower crustal rocks of the Eastern Alps are not considered in this review.

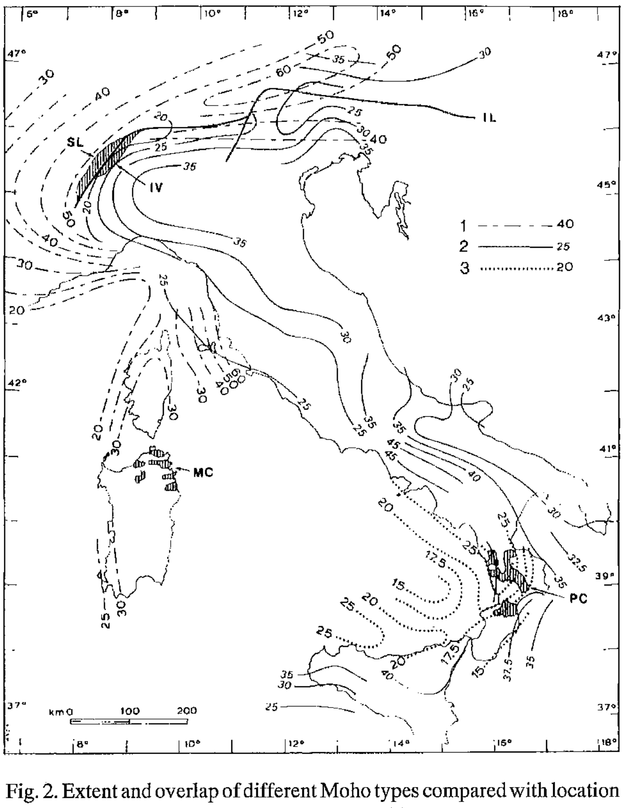

Fig. 2. Extent and overlap of different Moho types compared with location of the Italian lower crustal sections discussed in text.
(MC) Migmatite complex of N. Sardinia, (SL) Sesia-Lanzo, (PC) Polia-Copanello Unit of N Calabria, (IV) Ivrea-Verbano. (1) European, (2) African, (3) Tyrrhenian Moho. (IL) Insubric Lineament. Moho data (depths in km) after Cassinis et al. (1984).

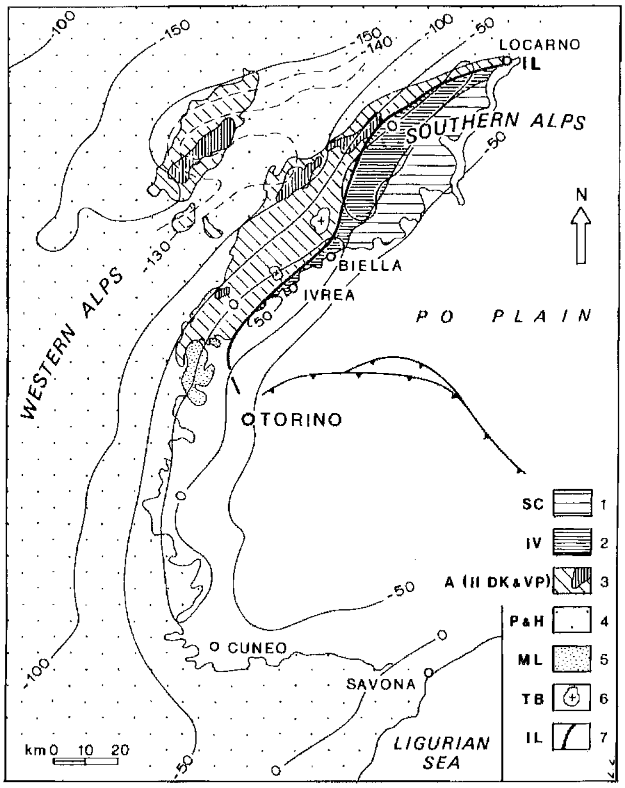

Fig. 3. Bouger anomaly map of the western Alpine Arc compared with distribution of IV and II DK lower crustal sections. (SC) Strona-Ceneri Zone, (IV) Ivrea-Verbano Zone, (A) Austroalpine Units ((II DK) Diorito-kinzigitic Zone, (VP) Valpelline "series"), (P&H) Pennidic and Helvetic Units, (ML) Lanzo Massif, (TB) Traversella and Biella plutons, (IL) Insubric Lineament. Notice the low gravity relief produced by the flat lying Valpelline outcrop. (Modified after Lanza, 1987).

1.1.1 *Southalpine Domain*

The Ivrea-Verbano Zone contains the best exposed and world-wide famous section of exhumed crustal rocks (Drake and Maxwell, 1981). The lower crust to upper mantle transition is also represented there by mafic and ultramafic rocks (e.g. sub-continental mantle peridotites and crustal layered gabbros) (Boriani and Sacchi, 1973; Boriani et al., 1977; Bigioggero et al., 1979; Shervais, 1979; Kruhl and Voll, 1979; Hunziker and Zingg, 1980; Rivalenti et al., 1981; Zingg, 1983; Schmid et al., 1987). Altogether, the granulite-facies rocks crop out for about 700 square km, with a crustal thickness of about 10 km.

The country rocks (kinzigitic Fm of the Authors) are mainly metapelites plus marbles, metarenites and minor metabasites, deposited sometime between 700 and 480 Ma (Hunziker and Zingg, 1980), corresponding to late Precambrian-early Ordovician times.

A first, pre-Hercynian, low-grade metamorphic event is possibly recorded by the date of 478 ± 20 Ma (Hunziker and Zingg, 1980; Boriani et al., 1983). The main granulitic-facies metamorphism should be slightly older than 325 Ma (Boriani et al., 1983) with re-equilibration conditions of about 700 °C and 9 kb when the sequence was

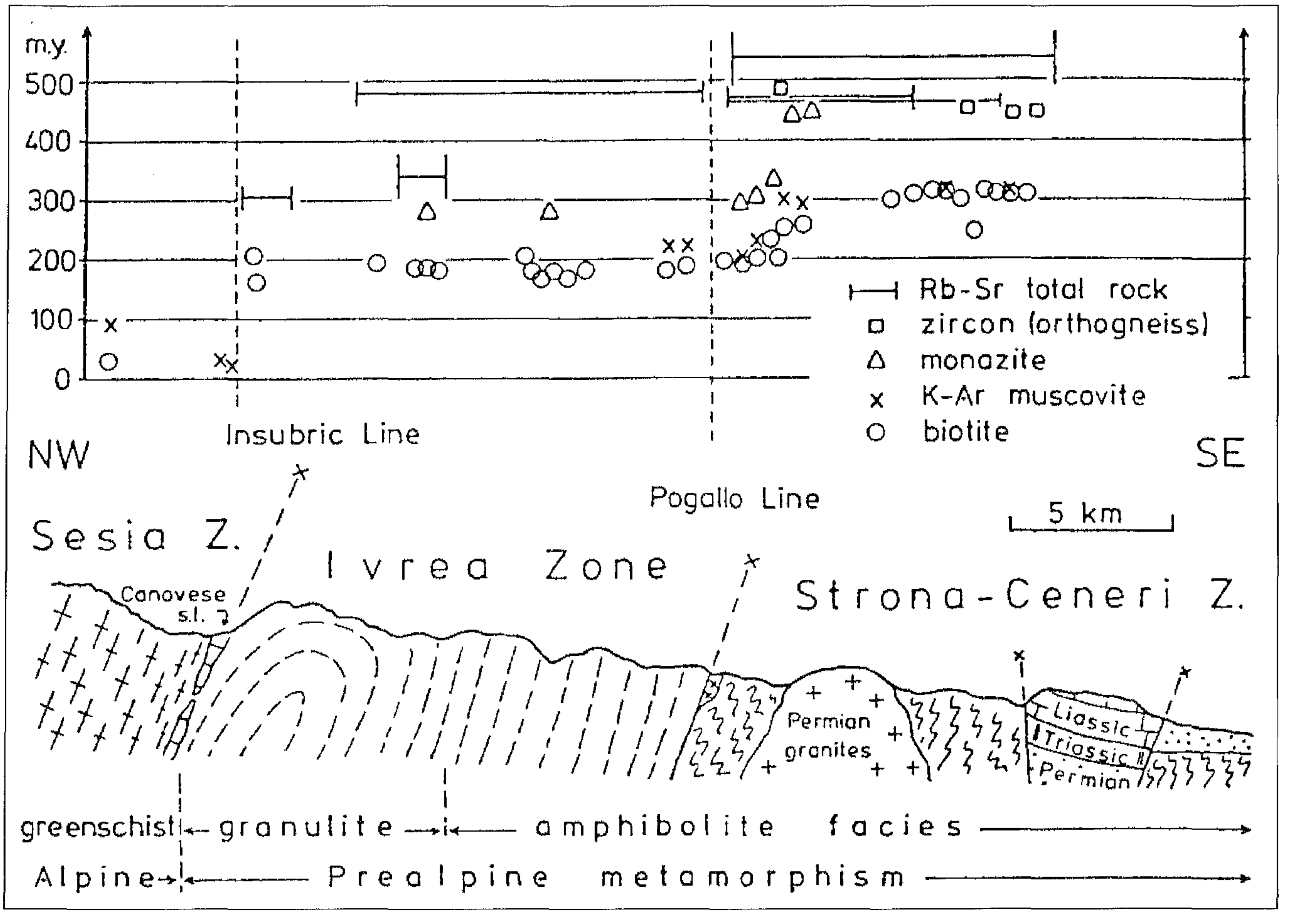

Fig. 4. Simplified composite geologic section across the Ivrea-Verbano Zone (Ossola and Sesia valleys) with distribution of radiometric dates. (After Zingg, 1983).

still flat lying (Hunziker and Zingg, 1980) or after the vertical upthrusting of the sequence (Schmid and Wood, 1976) and following the intrusion of the layered stratiform complex.

This complex consists of an eastward facing sequence of ultramafic (spinel-pyroxenite-clinopyroxenite-peridotite), mafic (norite-gabbro-anorthosite) and "dioritic" rocks. There is no agreement about the metamorphism of the complex. Many authors favour an intrusion following the main deformation or even the peak of regional metamorphism. According to Rivalenti et al. (1981) the conditions of metamorphic re-equilibration of the complex were 850-950 °C and 8-9 kb and partial re-crystallization took place during cooling. The regional metamorphism of the Ivrea-Verbano Zone is gradational to that of the adjacent Strona-Ceneri Zone, which postdates the Ordovician granites (466 ± 5 Ma, Boriani et al., 1983) and is therefore Hercynian.

Different peridotite bodies are associated with the stratiform complex close to the Insubric Line (Baldissero, Balmuccia, and Finero, at least partly). They are interpreted as slices of sub-continental mantle tectonically emplaced in the lower crust at 20-30 km depth (Shervais, 1979; Rivalenti et al.1981). Both the country rocks and the stratiform complex have subvertical dip with a major antiform to the west (Fig. 4) and few subvertical open folds overprinting earlier ductile deformations (Boriani and Sacchi, 1973; Boriani et al., 1977). The Ivrea-Verbano Zone as a whole is confined by the Insubric (Canavese) Line to the NW and the Pogallo Line (Boriani, 1970) to the SE. The Pogallo Line marks a late Hercynian thrust surface along which the Ivrea-Verbano overrode the Strona-Ceneri Zone. Similar important late Hercynian shortening phases are documented further eastward in the western Southalpine basement by anomalous superposition of units having different metamorphic grade. The direction of tectonic transport is toward the SE (Boriani, 1976; Boriani and Mottana, 1978). Summarizing, in the broader area of "Massiccio dei Laghi", three main Hercynian deformation phases have been separated: a main, isoclinal folding older than 325 Ma, a second open refolding, and a third phase of localized folding with vertical axes ("Schlingen-Tektonik"), brittle shortening and strike-slip motions (Vai et al., 1984; Vai and Cocozza, 1986). Some of the Hercynian shear planes were rectivated in Alpine time (Bocchio et al., 1980; Schmid et al., 1987).

Some authors have raised doubt about a main Hercynian age of deformation (or even of exhumation) as far as the Ivrea-Verbano Zone is concerned. They refer mainly i) to the age and nature of the Pogallo Line, inferred to be a Liassic deep-crustal low-angle normal fault (Hodges and Fountain, 1984; Schmidt at al., 1987); ii) to the age of exhumation of the Ivrea-Verbano Zone, inferred to be Triassic to Liassic (Hunziker and Zingg, 1980; Schmid et al., 1987); and iii) to the age of verticalization (up-turn) of the same Zone, inferred to be related to the early Neogene Insubric phase (Dal Piaz et al., 1972; Martinotti and Hunziker, 1987; Schmid et al., 1987). All these interpretation relay upon the different thermal history of the Ivrea-Verbano versus StronaCeneri Zones (Hunziker and Zingg, 1980; Schmid et al., 1987).

This point leads directly to one of the the main concern of this paper, that is the exhuming age of the Ivrea lower crust.

For discussing this point we need to take into account some regional boundary conditions, and to make comparison with other lower crustal sections outcropping in the central Mediterranean area.

1. The Southern Alps provide a fine picture of a complete Hercynian metamorphic zoneography showing a consistent tectonic polarity from the NNW to the SSE, passing from the granulite - facies metamorphism of the

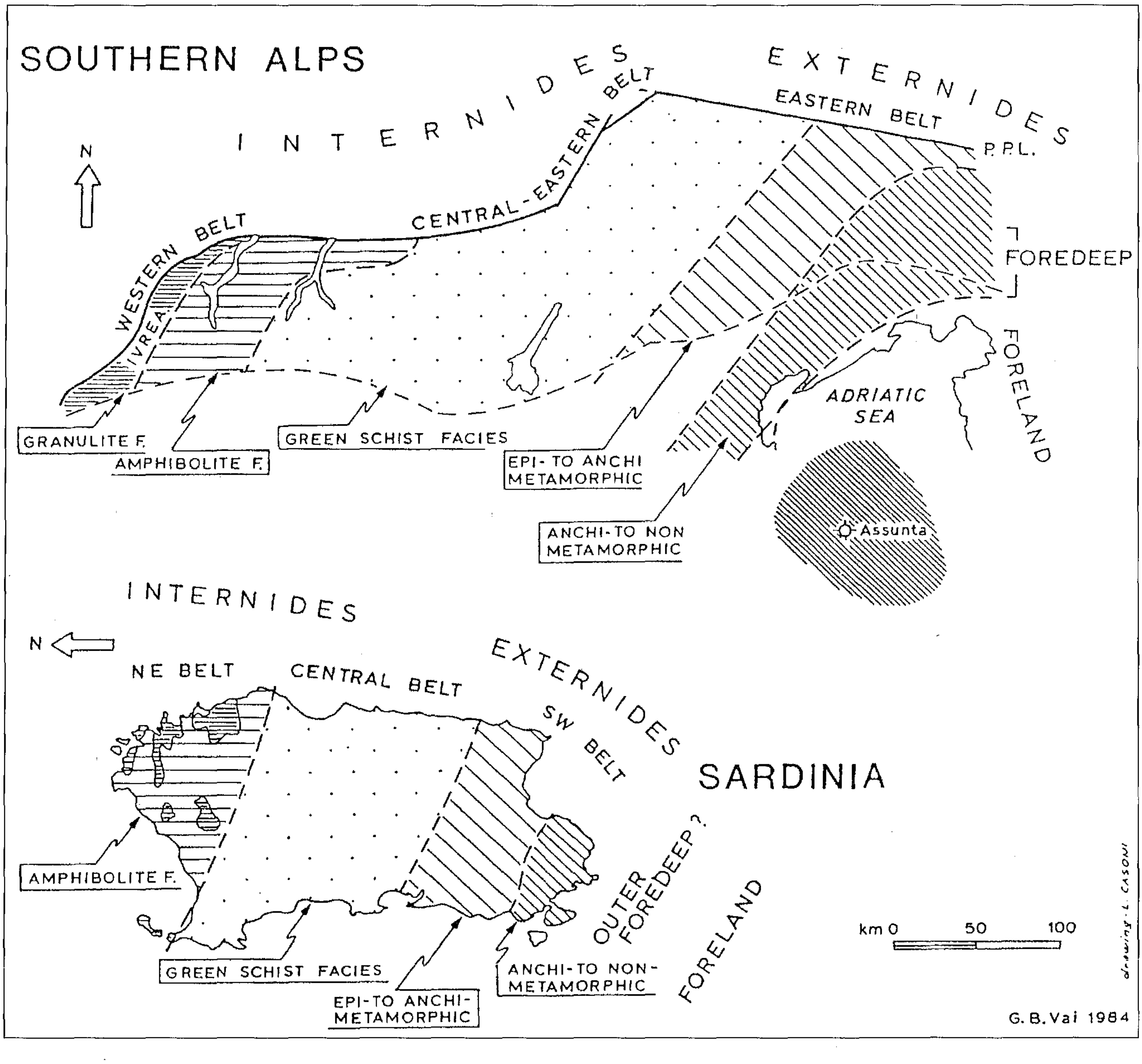

Fig. 5. Comparative tectono-metamorphic maps of the Variscan segments in the Southern Alps and Sardinia. The tentative palinspastic restoration of the Southalpine chain in pre-Alpine times was performed applying Alpine shortening factors of 2 in Lombardy, 1.3 in the Dolomites and 3 in the Carnic Alps. (P.P.L.) Palaeo Periadriatic Lineament. (Modified after Vai and Cocozza, 1986). Dense horizontal hatching in N Sardinia marks areas with granulite-facies relics.

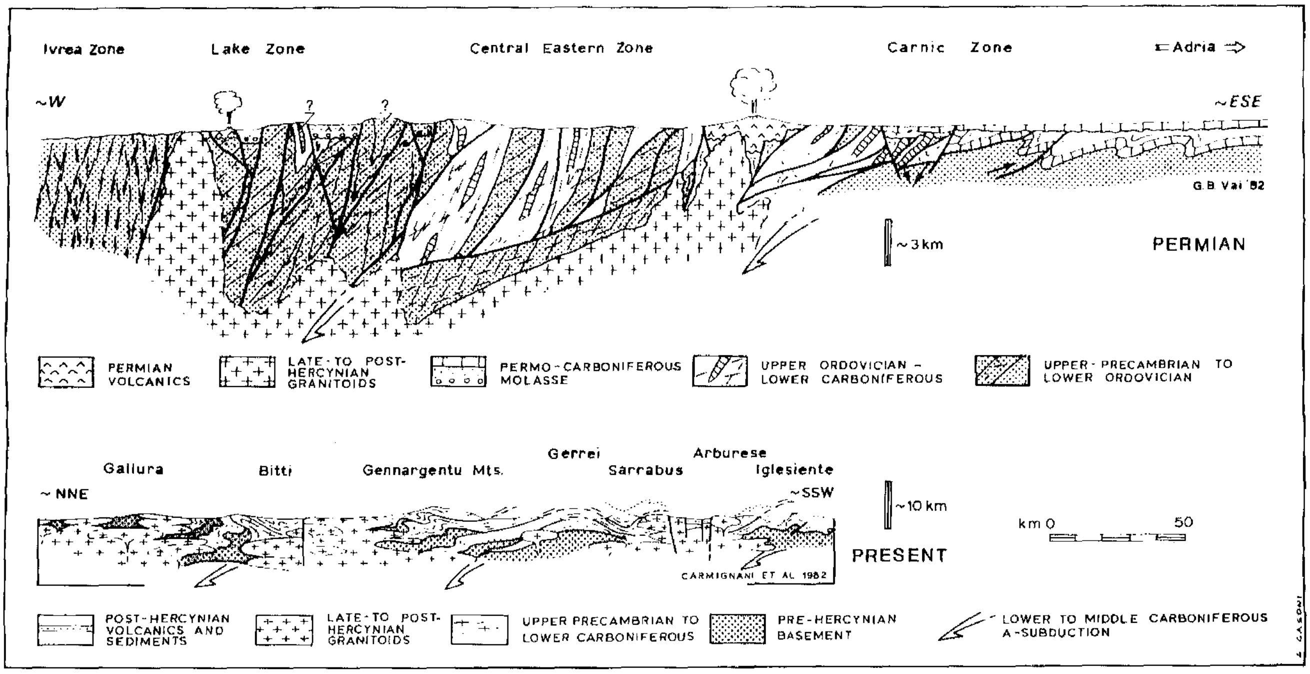

Fig. 6. Schematic geologic cross sections across the Southern Alps and Sardinia (modified after Vai et al., 1984 and Carmignani et al., 1982). Same horizontal scale. Maximum vertical thickness represented in both sections is about 13 km. Notice that the upper section is a palinspastic restoration at Permian times.

Ivrea Zone , the amphibolite and green schist-facies metamorphism of the Orobic and Venetian Southern Alps, through the anchi and non-metamorphic imbricate thrusts of the Carnic Alps foredeep to the foreland in the northern Adriatic Sea (Vai et al., 1984; Vai and Cocozza, 1986).Now, such a picture can be easily restored palinspastically, in spite of the late Hercynian (NS trending, eastward verging) and Alpine (EW trending, southward verging) shortenings, and can be compared with the coeval Hercynian belt of Sardinia which has not been affected by Alpine orogeny (Figs. 5, 6).

2. Each metamorphic facies of this Hercynian chain segment (except the Ivrea-Verbano Zone, apparently) is still sealed by non-metamorphic, unconformable late Carboniferous to early Permian sediments and volcanics. Only in places the Hercynian basement appears to have been intruded by post-tectonic granitoids along preferred meridian belts (Vai et al., 1984; Vai and Cocozza, 1986).

However, some Permian intrusions follow the limit between the Ivrea-Verbano and the Strona-Ceneri and, most important, Permo-Mesozoic cover rocks are found in the fault-rock band of the Canavese Line (Schmid et al., 1987 and older authors). These Permo-Mesozoic cover rocks include Permo-Scythian quartz-rich micaceous clastics, Triassic (Norian) dolomites and early Jurassic dark shales and siliceous limestones (Schmid et al., 1987), i.e. exactly the same sequence which can be found more or less continuously along the entire Insubric Lineament as far as the Gailtal and Karawanken Lines (Sassi et al., 1974; Zanferrari, 1976; Castellarin and Vai, 1982; see also the "wedged Permo-Mesozoic synclines" of the older authors).

These sediments represent a condensed sequence draping a Permo-Triassic structural high which splitted the Mid Triassic rift basin into two (Southalpine and Austroalpine) branches. This structural high included the northern Palaeocarnic Range, the Comelico and Pustertal phyllites, the Edolo and Morbegno schists and gneisses, and the Strona-Ceneri plus Ivrea-Verbano Zone. The Canavese Permo-Mesozoic sediments show Alpine green schist-facies metamorphism with dolomites preserved in boudins and other mylonitized lithologies: they are imbricated with Ivrea-derived mylonites of the same Canavese Line. Similar Permo-Scythian coarse clastic sediments with green schist-facies metamorphism are also found in different parts of the Insubric Lineament (e.g. Passo di Pennes, Alto Adige area). From the above data, it seems obvious to derive that the Permo-Mesozoic metasediments along the Canavese Line "represent the cover of the Southern Alps" (as Schmid et al., 1987 do) including, however, first of all the Ivrea-Verbano Zone, which is the part of the Southern Alps relevant to this discussion. The similarity in depositional and metamorphic characters of the Permo-Mesozoic condensed sequence rocks (referred above), bordering the Ivrea-Verbano Zone (along the Canavese Line) and the remaining part of the Southalpine Hercynian belt (along the Tonale, Judicaria and Gailtal Lines) (emphasizing the ancestral role of the Insubric Lineament, as suggested by Castellarin, 1982 and Castellarin and Vai, 1982), is a strong argument against a major differential uplift of late Triassic (or even Liassic) age limited to the Ivrea-Verbano Zone Zone. In fact a similar uplift would have affected the remaining entire Southalpine Hercynian belt. As such a late Triassic uplift is neither assumed nor documented in the whole Southalpine belt, the uplift of the Ivrea-Verbano Zone itself and the age of the Pogallo Line and related mylonites are assumed to be late Hercynian (Boriani et al., 1982).

3. The interpretation of the Pogallo Line as an intermediate to deep Liassic low-angle extensional fault (Hodges and Fountain, 1984; Schmid et al., 1987),though consistent with the Jurassic evolution of the Southern Alps as a passive continental margin (Laubscher and Bernouilli, 1977; Bally et al., 1981; Winterer and Bosellini, 1981), raises some problems:

i) nobody knows the thickness of the Canavese to Austroalpine Liassic crust because of the Alpine shortening;

ii) the Liassic continental margin should be expected much further W at the outer edge of the Austroalpine domain;

iii) one would expect major normal listric faults dipping toward the ocean (W to NW), as for the present Tyrrhenian and Biscay margins, and as documented by the substantial westward decrease in width of the Jurassic shallow water platforms (Castellarin, 1982, fig. 7);

iv) one should expect the maximum Liassic synrift sediment thickness to be found just over or close to the Pogallo site (actually a relative high) instead of the Lombardian basin;

v) the major problem, however, is the assumption (quite naif but strictly needed for internal consistency) of a Liassic flat lying Ivrea to Strona-Ceneri interface (and related regional metamorphic foliation) which could have acted as a possible low-angle fault plane. The different tectonic stage and the major angular unconformity between the Strona-Ceneri Zone and its almost flat lying Permo-Triassic cover (Fig. 4, after Zingg, 1983), however, require a pre-Permian tectogenic and orogenic event, and cannot just be explained by a simple Liassic low- angle extensional fault paralleling a still undeformed metamorphic facies boundary.

The need of trying to demonstrate a late Alpine age of the Ivrea Zone upturn, as well as of the open folding of the Ivrea foliation (Schmid et al., 1987) is strictly related to the Liassic low-angle normal fault assumption.

From the above discussion, we still maintain that a late Hercynian age of the Ivrea Zone uplift and of the Pogallo Line fits well the most available geological data. The different thermal history of the Ivrea-Verbano Zone as compared with the Strona-Ceneri Zone (as inferred by Hunziker and Zingg, 1980; Schmid et al., 1987), if reliable, may have a different meaning. I suggest that differential re-heating and cooling as a function of distance and gradient from discrete diffusion centers related to thermal events (like the Permian, Mid-Triassic and early Jurassic rifts) should be carefully tested (as already suggested by Ferrara and Innocenti, 1974). Otherwise, we should assume this Ivrea-Verbano-like obduction mechanism by extensional faulting for many Alpine "Verrucano" conglomerates (e.g. Passo di Pennes, central Alps) showing mid-Triassic and/ or Liassic cooling ages. A simple check of the thermal event hypothesis could be attained testing the thermal history of Austroalpine crystalline segments that have still on top their Permo-Triassic cover. If a Liassic (or Triassic) age is obtained, than the thermal event model will be proven.

To sum up, the Ivrea-Verbano Zone appears as a minor-sized wedge of lower crustal origin with some additional upper mantle slices overprinted by granulitic facies metamorphism of Hercynian age and raised to or near the surface in late Hercynian times in a setting and by a process which will be discussed later.

The gravimetric, magnetic and crustal features of the Ivrea area have been described and commented in a number of papers in the two last decades, and the reader is referred to a recent review by Lanza (1987), where the main possible relationships between the geological Ivrea Zone and the geophysical "Ivrea body" are critically discussed.

The present structural setting is represented on Fig. 4, whereas Fig. 6 gives a reasonable picture of the late-Hercynian Permian setting. A possible schematic structural evolution from late Precambrian is suggested on Fig. 10. It appears that both kinematic and genetic decoupling of the Ivrea Zone from the "Ivrea body" is favoured (see below).

1.1.2 *Austroalpine Domain*

A second, less known section of deeper continental crust crops out in the over 350 sq km II "Diorito-kinzigitica" Zone (Franchi, 1905) and Valpelline "series" (II DK and VPS), which form parts of the Austroalpine nappe system (Sesia-Lanzo Zone and Dent Blanche Nappe, Dal Piaz et al., 1972) (Fig. 3). This section runs a few km NW of and almost parallel to the IV, on the opposite side of the Insubric Lineament. Unlike the Southern Alps, the Austroalpine system underwent severe deformation and metamorphism during the Cretaceous to Eocene collision (eo-Alpine and meso-Alpine events) and the later neo-Alpine event. However, in the geometrically lower unit of the Dent Blanche and Sesia Zone the Alpine metamorphic overprint is prevailing, whereas in the upper one (which contains the II DK and the Valpelline "series") the pre-Alpine metamorphism is only partly affected by Alpine events (Martinotti and Hunziker, 1987, cum bibl.).

Part of the Sesia Zone and Dent Blanche Hercynian crust was intruded by late Hercynian granitoids (e.g. Mucrone), exposed and draped by Permo-Mesozoic cover sediments (e.g. Mt. Dolin, Canavese Zone, etc.). However, no diagnostic primary contact seems preserved and none of these sediments sealing the intrusions refer to the II DK outcrops.

The DK is composed of Al-rich granulitic paragneisses, dolomitic silicate marbles, granulitic-facies metabasites and rare ultramafic masses with peridotitic assemblages preserved.

It is not surprising that both II DK and IV show similar mica cooling ages, between 240 and 140 Ma (Martinotti and Hunziker, 1987), as these units are close to each other and the DK represents the highest part of the Sesia Zone. This could suggest a similar late Hercynian exumation age. However, geological data are much less constrainig than for the IV (see Dal Piaz et al., 1972). Some events, as the strong eo-Alpine subduction-obduction process, would even suggest a possible mechanism of exhumation or re-exhumation for the II DK.

It is of special interest to note that the Sesia-Lanzo Zone fits the strong gravimetric anomaly of the geophysical "Ivrea body" much better than the IV does (see Fig. 3). It was already noted that both the IV and the Canavese Line are obliquely intersected by the Bouger anomaly strip which extends southward to the Pennidic Dora-Maira Massif (Lanza, 1987). Therefore, if a connection between the "Ivrea body" and the outcropping structural units is to be assumed, not only the IV (older Authors) but also the II DK should be taken into account, not leaving a part the possible connection with the external Pennidic European basement (Menard and Thouvenot, 1984; Lanza, 1987), or with the Pennidic Piedmont oceanic lithosphere slices (Lanza, 1987).

1.1.3 *Geological Affinity of the Geophysical "Ivrea Body"*

No geological constraint implies the same age for the exhumation of the IV and the emplacement of the "Ivrea body". It would appear rather strange, in fact, to assume a unique Hercynian age (as done until recently, except for

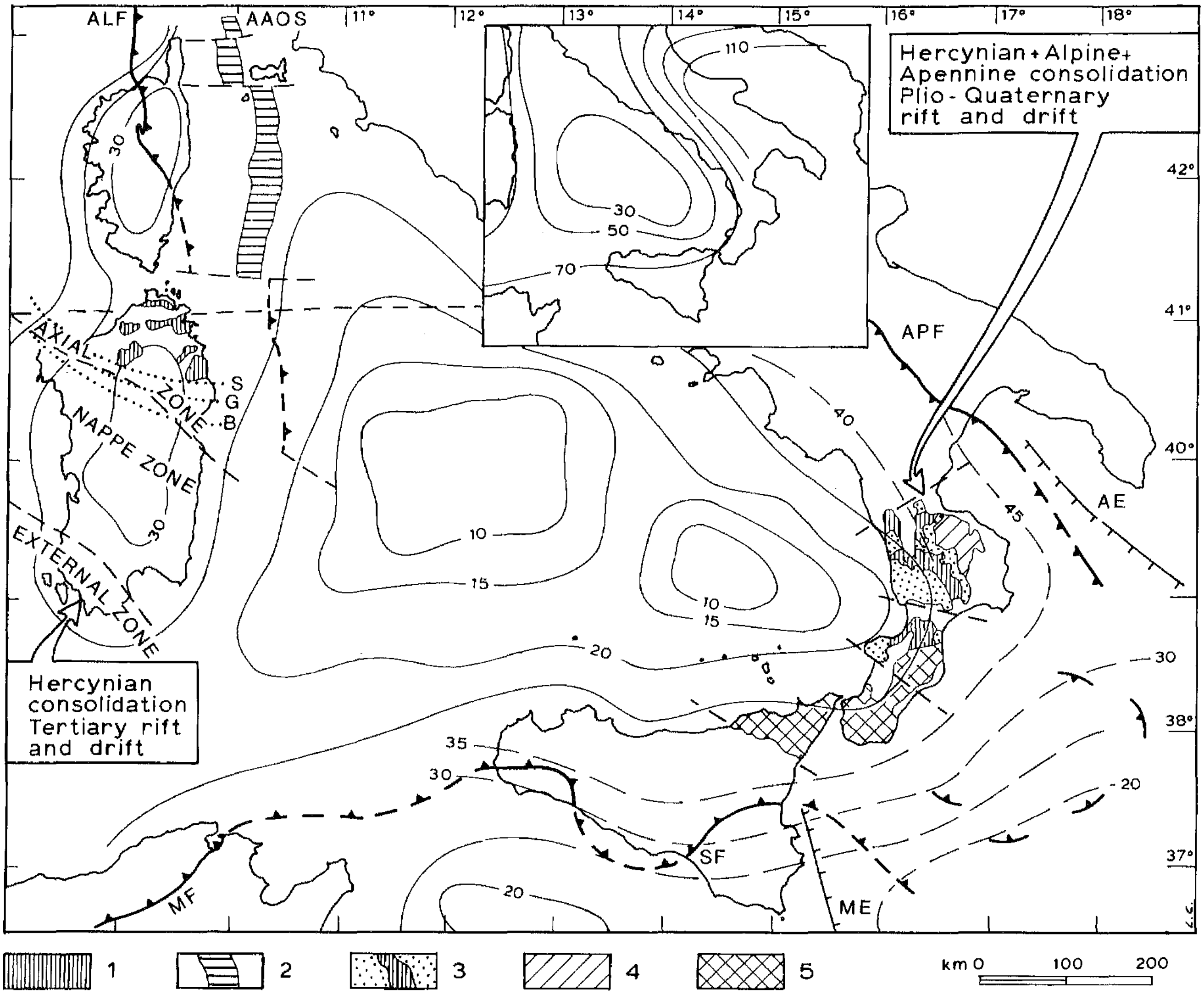

Fig. 7. Lower crustal sections outcropping in Sardinia and Calabria related to the main Hercynian, Alpine and Apennine tectonic frames and compared with the Moho depth (km below sea level).
(AAOS) Alps-Apennines ophiolitic suture, (AE) Apulia escarpment, (ALF) Alpine front, (APF) Apennine front, (ME) Malta-Siracusa escarpment, (SF) Sicilian front, (MF) Maghrebian front. (1) lower crust outcrops, (2) ophiolitic suture, (3) Westalpine Calabride element, (4) "Insubric"? Calabride element, (5) Kabylian Calabride element. Moho depth in the Ionian-Apulia area is shown as dashed lines. Thickness of the lithosphere (km) after Panza et al. (1980) in the inset. (Sources: Selli, 1981; Carmignani et al., 1982; Scandone, 1982; Cassano et al., 1986; Sartori et al., 1987; Rehault et al., 1987).

Lower crustal section	Sedimentation &/or magmatic emplacement	Granulite facies metamorph.	Late tect. intr.	Cover age	Exuming age	Alpine HP metamor.	Moho depths Km	Lithosph. thickness Km	Crust & lithosph. doubling	Gravity anomaly	Magnetic anomaly
MC Eur.Foreland N Sardinia	458$\pm$31Ma ------------	pre-Herc. eo-Herc.	yes	P-T	Hercynian	no	30	80	no	no	yes high
PC ?Austroalpine Calabria	~450 Ma ------------	~295 Ma	?yes		?eo-Alpine	yes	?20 40	90-110	?yes	no	yes low
II DK Austroalpine W Alps		pre-Alp.	yes	?P-M Olig.	?Hercynian eo- to me- so-Alpine	yes	<5 40-50	?	yes	yes very high	yes high
IV Southalpine W Alps	<700	~325 Ma ?478 $\pm$ 20 Ma	no	?P-M	?late Herc. ??neo-Alp.	no	<5 40-50	?	yes	yes very high	yes high
"Ivrea body"	?	?	?		eo- to me- so-Alpine	?	<5 40-50	?	yes	yes	yes

Tab. 1 - Main aspects of the outcropping Italian lower crustal sections discussed in text. Some features of the buried geopysics "Ivrea body" are also reported for comparison. (P) Permian, (T) Triassic, (M) Mesozoic.

Dal Piaz et al., 1972) in an area showing a major development of Alpine tectogenesis or viceversa to require a single Alpine age in an area of major Hercynian deformation (Castellarin and Vai, 1981; Vai et al., 1984; Vai and Cocozza, 1986). This is even more relevant when the strong similarity in geologic evolution of the Alpine and Hercynian Wilson cycles in the circum-Mediterranean area is taken into account (Vai and Spalletta, 1982) (see Fig. 10).

On the other hand, the main argument against a Hercynian age, i.e. the difficulty in conceiving a long-lasting large-size structure in isostatic disequilibrium, though claimed in the recent past, has a few practical consequences. As a matter of fact, we know many geologic and geophysical bodies resting in isostatic disequilibrium since millions of years. A good example is the large vertical Moho offset connected with the Hercynian Faille de Bray (Cazes et al., 1986) which survives since at least 300 Ma, or the preserved gravity anomalies of many lower crustal sections dating back to Precambrian and pan-African ages (Fountain and Salisbury, 1981, fig. 3). Assuming the youngest possible ages for the "Ivrea body" (rougly 60 to 20 Ma) we have to face the problem of a difference of many orders of magnitude as compared with the duration of classic isostatic processes (e.g. 10 to 100 thousand years for glacial rebound). These long lasting anti-gravitational settings have to be explained in some different way. A possible one could be an elastic lithosphere supporting a dense body of size less than the lithosphere thickness. In fact, topographic mass anomalies with a wavelength less than about 100 km are not compensated even in cases where the thickness of the elastic lithosphere is about 6 km (Turcotte and Schubert, 1982). The flexure of the elastic lithosphere would in fact support the uncompensated mass anomalies. An other solution could be that of a thin dense body supported by a larger mass of soft buoyant material (such a condition is likely to have occurred in zones of continental collision). In this case, however, no gravity anomaly would be observed (as it happens for Calabria and Sardinia, see Tab. 1).

Following these general statements, we prefer to keep the "Ivrea body" distinct from each of the outcropping structural units to which it can be related and to assume an eo-Alpine emplacement age (Dal Piaz et al., 1972) and a neo-Alpine back-thrust for its near-surface tip (following the kinematic proposed by Laubscher, 1984, and partly by Schmid et al., 1987).

The IV, which is the zone topographically closest to the "Ivrea body", was most likely upthrusted in late-Hercynian times (Boriani et al., 1982; Vai et al., 1984).

The II DK, together with the Sesia-Lanzo Zone, is kinematically, structurally and magnetically closest to the "Ivrea body", and was possible first exhumed in Hercynian time. However, it was subducted during the eo-Alpine event and re-exhumed before the deposition of the Oligocene volcanoclastic cover and the intrusion of the coeval Traversella granitoid (Martinotti and Hunziker, 1987).

The oceanic mantle slices (e.g. Lanzo Massif, which shows physical properties closest to the "Ivrea body", Lanza, 1987) appear less obviously related to the contoured surface anomaly (Fig. 3), but could fit better at depth the same anomaly.

Wedges of European (Helvetic) crust-mantle transition are less likely related to the "Ivrea body" for both geometric and kinematic reasons.

To sum up, it appears that upthrust of lower crustal segments occurred at expenses of the Hercynian basement of the Austroalpine to Southalpine domains of the present western Alps in both Hercynian and Alpine times. Alternatively, the magnetic anomaly connected to the "Ivrea body" could be related possibly to upthrust of the oceanic lithosphere in Alpine time.

1.2 *Northern Calabria*

Lower crustal sections outcrop for about 1,500 sq km as a part of the eo-Alpine chain fragment, which was thrust in toto over the Apennines (Figs. 1, 7, 8) during early Miocene time (Scandone, 1980; 1982).They are composed of Hercynian metamorphics, partly intruded by late Hercynian granitoids and affected by Alpine metamorphism. They are, therefore, correlated with the Austroalpine units of the western Alps and referred to as Polia-Copanello unit ("Dioritico-kinzigitica" of older Authors). They consist of biotite-garnet gneisses with sillimanite and cordierite, metabasites and subordinate peridotites and marbles in the Calabria Coastal range (Scandone, 1982) and Sila (Bonardi et al., 1982), up to several hundred m thick, with Hercynian granulite to amphibolite facies metamorphism and Alpine overprint of moderate high pressure (lawsonite). In the northern Serre, the granulite-facies rocks reach a thickness of 7 km and form a homocline dipping to the SE. It consists of a lower layered granulite-pyriclasite unit with minor amphibolites and ultramafics and an upper metapelite with meta-monzogabbronorite and marble unit (Schenck, 1980; 1981; Maccarrone et al., 1982). Late Hercynian quartz diorites, tonalites and granites intruded at higher crustal levels overlie in Alpine tectonic contact the previous sequence. Granulite facies metamorphism occurred near 800°C and 8 kb (about 25-30 km) in Hercynian times, with a minimum age of 296 $\pm$ 2 Ma. The granulites were then rapidly uplifted 5 to 7 km higher (1.5 to 2 kb), where they slowly cooled until 110 Ma and were finally uplifted at the surface during Oligocene to Miocene times (Schenck, 1980); this is not supported, however, by radiometric data. An early Cretaceous uplift is preferred by Borsi et al. (1976) to explain the coeval biotite dates.

Lack of Permian sediments and Mesozoic cover, which is apparently limited to the Bagni unit (Bonardi et al., 1982), favours a final exhumation of the Calabrian lower crust in eo-Alpine times. This is not contradicted by the thermal evolution (Schenck, 1980), which is very similar to that of the Austroalpine Sesia-Lanzo Zone. However, as discussed above, the thermal evolution of the Calabrian lower crustal sections too could be controlled mainly by late Hercynian to early Alpine rift pulses and shear zones rather than by simple post-metamorphic cooling.

It is worth mentioning the resemblances between Calabria and the Moldanubicum (Central European Hercynian belt) where granulite bodies with characteristic granulites, pyriclasites and ultramafics are linked with anatectic paragneiss series (Schenck, 1980). This is true also for N Sardinia (see below) and might support a former connection of the three blocks in Hercynian times (Vai and Cocozza, 1986).

No special gravity anomaly seems to be linked to the Calabrian lower crustal outcrops. The magnetic anomaly map shows only a very faint elongated high, matching the Sila Piccola granulites. The first simple explanation might be that the lower crustal sections of Calabria (and the associated ophiolite and Pennidic units of Westalpine affinity, Scandone, 1982) represent totally unrooted, thin-skinned tectonic elements thrust upon the Apennine crust.

The crustal structure of the Tyrrhenian side of Calabria and Sicily is quite complex, showing a sharp jump from 20-25 km deep marginal Tyrrhenian crust to the 40-45 km deep south Apennine-Sicilian crust. Even more complex appears the overlap of the two different crusts across the Serre and Aspromonte (Figs. 2, 7, and Schwarz, 1978, fig 2). This setting is quite unlike the classic lithospheric doubling of the Ivrea-Sesia area.

If the Calabrian lithospheric doubling could have been produced by the eo-Alpine collisional event (as might be supported by the subducted SE Tyrrhenian, "fossil", lithospheric slab (Scandone, 1980, 1982), the present setting is more related to the Plio-Quaternary rift and spreading which have produced the oceanic part of the Tyrrhenian Sea (Sartori, 1987), that implies a quick, progressive oceanization of the Apennine crust underlying Calabria.

The present structural setting of the Calabria lower crustal sections is represented on Fig. 8A. The relatively thin and flat-lying lower crustal sheet detached from its roots explains the lack of gravity anomaly and the very low magnetic one. Fig. 8B shows how, after the eo-Alpine Sardinia-vergent obduction of the Polia-Copanello lower crust, Africa-vergent (back) thrusting translated the Alpine pile of nappes onto the Apennine domains (Scandone, 1980). Continued Neogene Apennine thrusting (Scandone, 1982) and Plio-Quaternary Tyrrhenian rifting and spreading (Sartori, 1987), accompanied by the tremendous Quaternary uplift of Calabria (Ghisetti and Vezzani, 1979), brought to the present setting. However, the place of the Tyrrhenian suture (i.e. the former continuation of the Alpine ophiolite belt from Corsica to northern Calabria across the present Tyrrhenian Sea), the African versus European affinity of the Calabrian lower crust, the original position of the Longobucco unit (E or W of the Tethys ocean) and the place or places of the Apulian A-subduction are still matter of debate, starting from Alvarez et al. (194) through Bouillin et al. (1988).

1.3 *Northern Sardinia*

Fragments of a lower crustal section crops out all across NE Sardinia and S Corsica (Figs. 1, 5, 7, 9). They have been described as migmatite complex (MC) and consist of migmatites, calc-silicate masses and amphibolites with relics of granulite facies parageneses and rarely

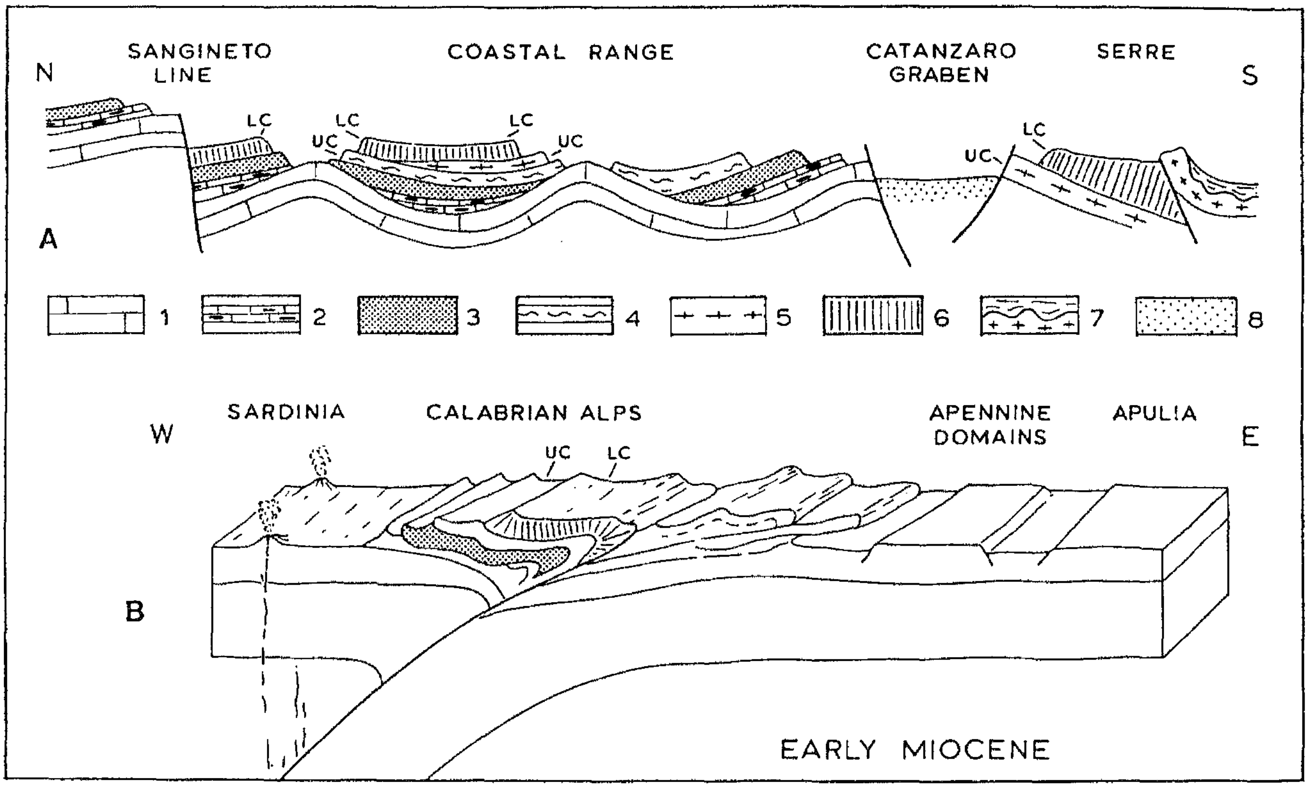

Fig. 8. A: Present geometric relationships among the Calabria Alpine units in the Coastal Range and the Serre. (1) Apennine units, (2) Frido Unit, (3) Ophiolitic units, (4) Bagni Unit, (5) Castagna Unit, (6) Polia-Copanello diorito-kinzigitic Unit, (7) Stilo Unit, (8) Neogene to Quaternary Graben fill. (LC) lower crust, (UC) upper crust.
B: Palinspastic cross section from Sardinia to Apulia showing lower crust (hatching) exhumed upon upper crust and ophiolites (dark) in eo-Alpine times and involved in the later Apennine thrusting with opposite vergence. (Slightly modified after Scandone, 1982).

eclogites partially retrogressed.

Granulites and eclogites occur as small bodies within the migmatitic rocks. They may represent the relics of a pre-Hercynian basement (Ghezzo et al., 1982). Anyway, these outcrops appear as two main WNW-ESE trending belts wedged within the Hercynian axial zone of NE Sardinia (plus part of Corsica). They were exhumed in late Hercynian times (a Permian cover with volcanics being present in Nurra and Anglona) and intruded by late Hercynian granitoids.

The gravity map (Bouger anomaly) with values ranging from 0 to 20-30 mgal shows no remarkable feature except for a faint trend of the isoanomalies to parallel the NW-SE trending main Hercynian structural axes. Also the Moho depth map (Fig. 7) suggests a slightly attenuated crustal thickness (25-30 km) common for those blocks which underwent Hercynian crustal consolidation and were located close to the Tethys realm or involved in the Alpine belt.

Much more characterized is the magnetic anomaly map (AGIP, 1981; Galdeano and Rossignol, 1977). Leaving a part large anomalies related to Tertiary and Quaternary high-susceptibility volcanic and subvolcanic bodies, which, however, parallel the Hercynian structural axes, it is interesting to remark the two strong elongated positive anomalies of NE Sardinia and SW Corsica. They fit quite well the outcrops of the granulite and eclogite bearing Palaeozoic migmatitic complex (Fig. 9), if the regional dip is taken into account, even though a couple of subcircular anomalies shown in the NW Sardinia rift and offshore might suggest the concurrence of recent volcanic bodies.

A geologic section across the northeastward dipping Hercynian belt of Sardinia is shown in Fig. 6. The relatively subordinate extent of the lower crust remains at the surface, as compared with the granitoids, might have been underestimated and could increase considerably even at shallow and intermediate depth (Carmignani et al., 1982). This might explain how larger amphibolite and eclogite masses at depth could heavily contribute to the relevant magnetic anomaly, whereas a check on the outcropping granitoids by Lanza provided only low susceptibility values (Carmignani, pers. com., 1988).

2. Comparison

Table 1 summarizes the main features of some Italian lower crust outcrops. Sedimentation or magmatic emplacement of the protoliths span from late Precambrian (about 700 Ma) to Ordovician (about 450 Ma) with oldest inherited zircon ages of about 1.9 Ga. Hercynian ages for the granulite-facies metamorphism are largely preferred, but pre-Hercynian ones are also suggested; they include

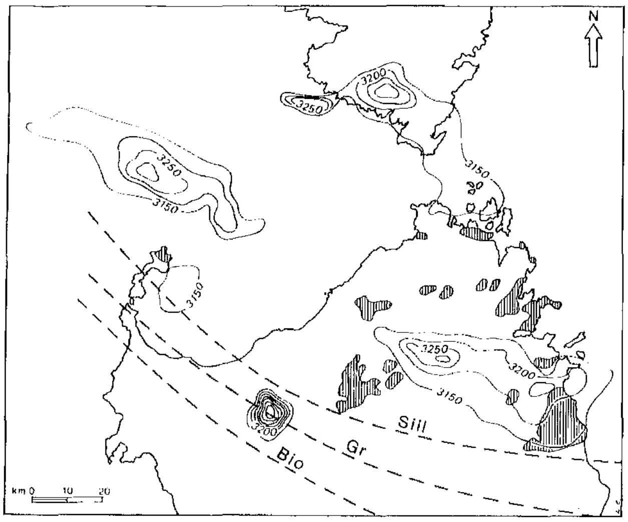

Fig. 9. Main positive magnetic anomalies of NE Sardinia compared with the migmatite complex outcrops (including partly retrogressed granulite-facies rocks). Contour interval: 50 nT.
(Sources: Galdeano and Rossignol, 1977; AGIP, 1981; Ghezzo and Orsini, 1982; Ghezzo et al., 1982; Elter et al., 1985).

both Ordovician anorogenic "Caledonian" (Vai, 1975; Ghezzo et al., 1982) and even Cadomian (in the Ceneri Zone, Hamet and Albarède, 1973) dates.

Late tectonic intrusions, following the regional granulite metamorphism, are possibly (Calabria) to definitely present in all the examined section, except for the Ivrea Zone. This might be linked to the quick and very late tectonic exhuming of the Ivrea Zone as compared with the slower and possibly two-phase exhuming of the remaining lower crustal sections (e.g. Calabria, Schenk, 1980).

Lithospheric conditions (of both crust and lid) are sharply different for lower crustal sections preserved from (Sardinia) or deeply involved in the Alpine diastrophism (Panza et al., 1980). A further difference separates Ivrea-Verbano and II Diorito-kinzigitica from the Calabria lower crustal sections, as the two first zones crop on top of a major lithosphere doubling, whereas the third shows a complex superimposition of a new oceanic lithosphere over an old continental one. This is reflected in the gravity anomaly distribution, with Sardinia and Calabria poorly characterized, whereas Ivrea-Verbano and II Diorito-kinzigitica are marked by strong gravity anomaly.

The magnetic anomaly distribution points out to the totally unrooted setting of the Calabria lower crustal sections as opposed to the partly rooted setting of the remaining ones.

Finally, the exhuming age appears to be definitely Hercynian in Sardinia, almost likely late Hercynian for the Ivrea-Verbano, uncertain (Hercynian or/and eo- to meso-Alpine) for the II Diorito-kinzigitica, and possibly eo-Alpine for Calabria. One might therefore suggest Sardinia as a test site for checking the different interpretation of the Ivrea-Verbano (plus II Diorito-kinzigitica and Calabria) thermal history (thermal overprint by Permo-Mesozoic rifts, Alpine diastrophic rejuvenation, slow cooling) some of which are markedly contrasting with basic geological data, as discussed above.

3. Palaeotectonic Restoration

All the examined lower crustal sections show Hercynian (to possibly pre-Hercynian) granulite-facies metamorphism, and part of them Hercynian exhumation. Therefore, a palaeotectonic tracing back of their geological evolution would be useful, even though very difficult (e.g. von Raumer, 1987).

In developing a tectonic model previously suggested (Vai, 1980a, 1980b; Vai and Cocozza, 1986) I will summarize the geologic evolution of two critical, well studied sections, the Ivrea-Verbano and the II Diorito-kinzigitica, by means of serial, schematic palinspastic cross sections (Fig. 10). A very close evolution can be suggested in Calabria and, except for the Alpine diastrophism, Sardinia too.

(1) Following a Cadomian-Baikalian-Panafrican crustal consolidation in the latest Precambrian, the south European circum-Mediterranean area underwent repeated oblique riftings through most of the early to middle Palaeozoic. Two mainly transtensional (but locally also transpressional) peaks have been recognized. The first was detected in the Upper Cambrian to Lower Ordovician rocks of Sardinia (Vai, 1982), the Helvetic (von Raumer, 1987) to Pennidic realms, the Southern Alps (Vai, 1982,1984) and many other south European and north African areas. The effects of this rifting in the infra-lithospheric layers are expressed by the Ordovician magmatism, frequently scattered over large areas, and especially the so-called Ordovician thermal or thermo-metamorphic, "anorogenic", "Caledonian" event, mainly recognized by radiometric dating (Vai, 1975).

The second rifting peak took place in an already attenuated lithosphere, at the Ordovician-Silurian boundary and in the basal Silurian; it was accompanied by widespread alkaline magmatism (Sardinia, Helvetic realm, central and western Europe, etc.). This event correlates with a well-known, world-wide, climatic (melting of the late Ordovician ice cap) and geotectonic (fast early Silurian spreading rate) transgression (Vai, 1982, 1984; Vai and Cocozza, 1986)

A third, even stronger transform rifting peak was recognized in the externides to foredeep parts of the Southern Alps,Montagne Noire, Pyrenees, Catalunia, Sardinia, Moroccan Meseta, etc. during late Devonian to early Silesian times. It is matched by tectogenic conditions in the internal zones of the same areas, with HP-metamorphism, collision, thrusting and magmatism. This is especially true for a long segment extending from the future western Southalpine (Vai and Cocozza, 1986) to the Helvetic realms (von Raumer, 1987). I suggest this is

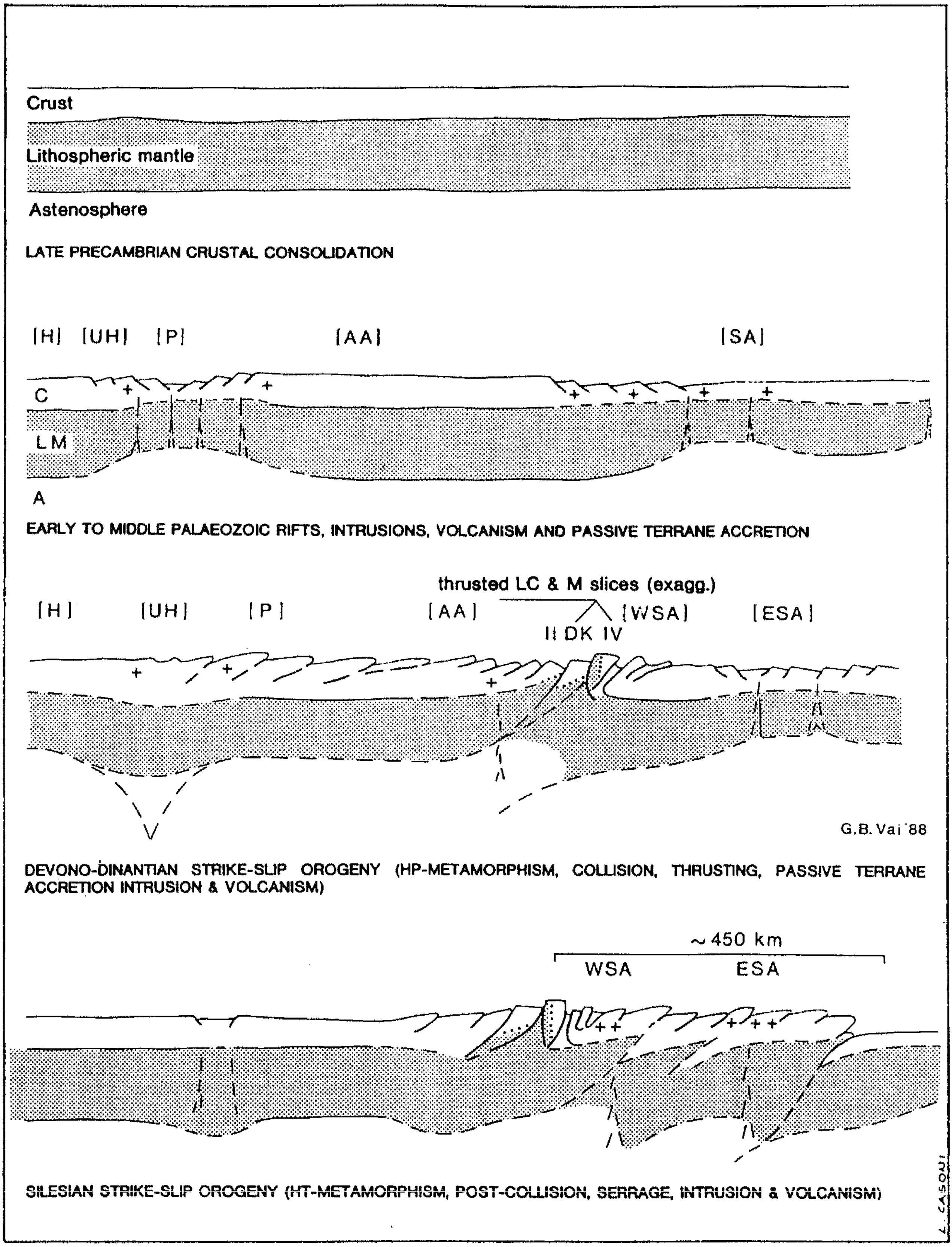

Fig. 10. Conceptual serial palinspastic sections showing repeated anomalous development of the Wilson cycle in Hercynian and Alpine times across the Alps (see text). (H) Helvetic, (UH) Ultrahelvetic, (P) Pennidic, (AA) Austroalpine, (SA) Southalpine, (WSA) western Southern Alps, (ESA) eastern Southern Alps, (LC & M) lower crust and mantle, (II DK) Diorito-kinzigitica, (IV) Ivrea- Verbano, (NP) north Pennidic, (CP) central Pennidic, (LPO) Liguria-Piedmont ocean, (S) Sesia Zone, (L) Lombardo Basin, (T) Trento Platform, (B) Belluno Basin, (F) Friuli Platform, (J) Julian Basin; (P) Permian, (T) Triassic, (IL) Insubric Lineament. The two dotted segments mark the Ivrea-Verbano and the II Diorito-Kinzigitica Zones.

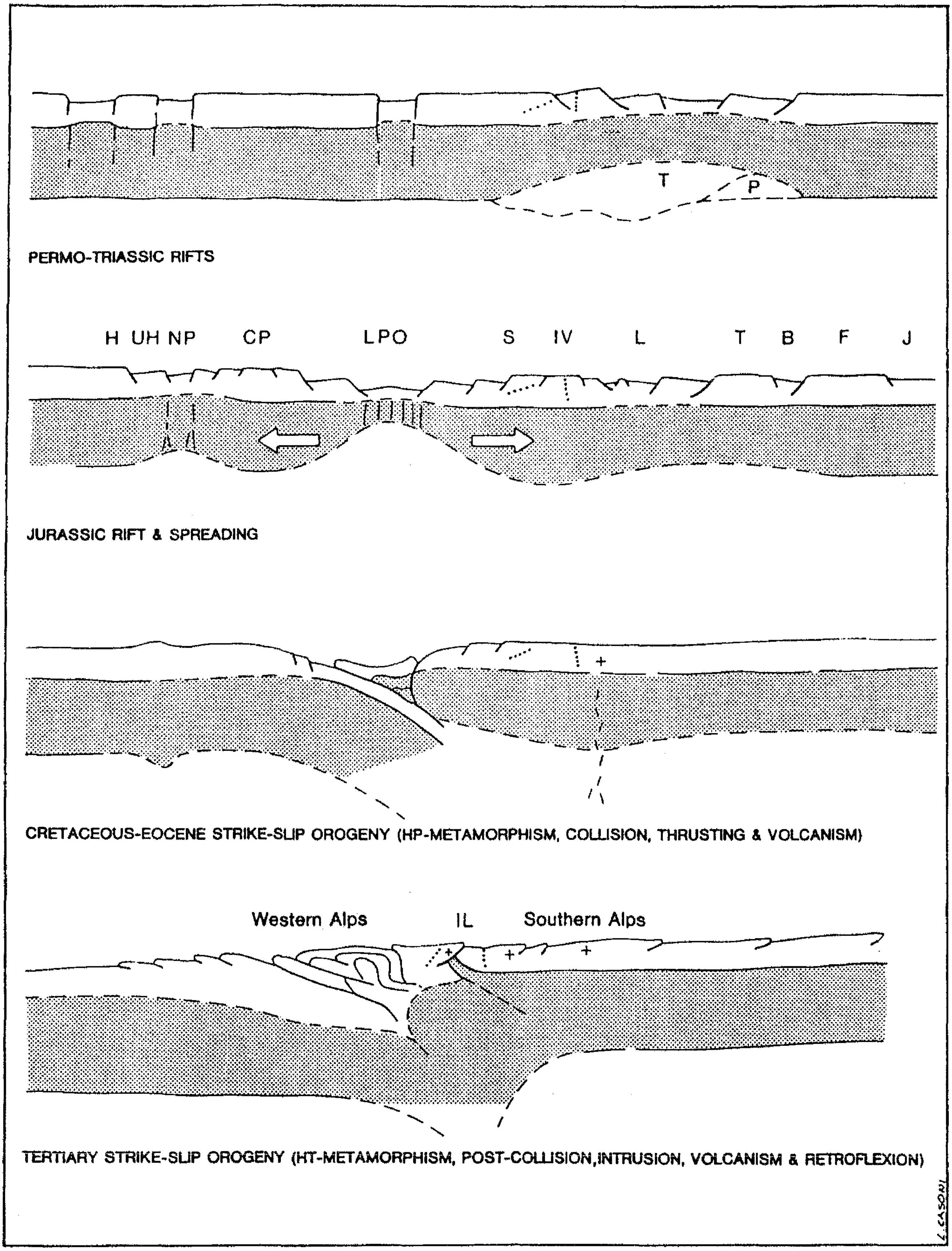

Fig. 10 - continued).

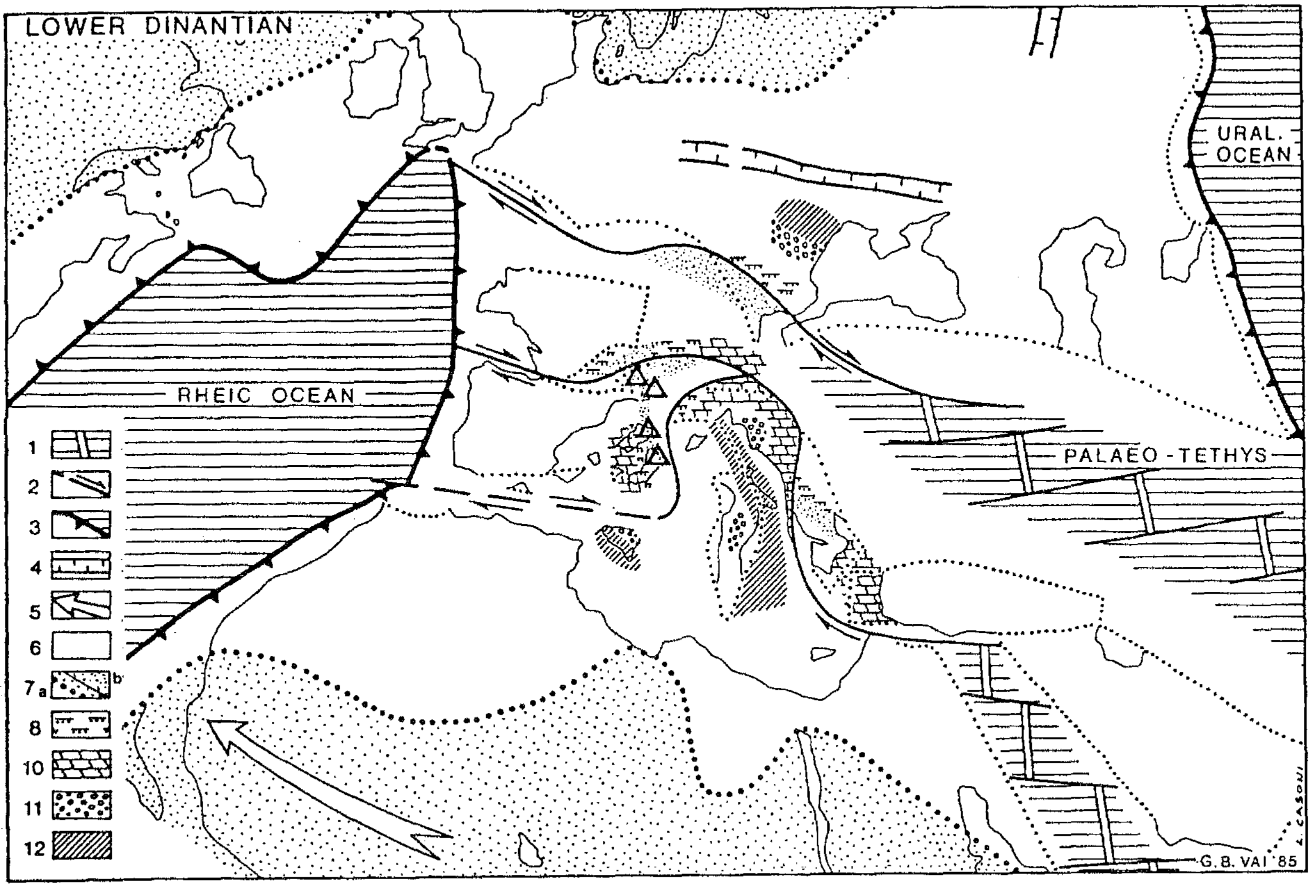

Fig. 11. Physiographic environment and tectofacies distribution in the circum-Mediterranean areas and inferred geodynamic setting during the early Dinantian. Open triangles mark the location of the four lower crustal sections described in the text. (1) Oceans and spreading axes, (2) main transform and wrench faults (transpression and transtension zones not differentiated), (3) B-subduction zones, (4) main rifts, (5) main rotational vector, (6) epicontinental seas, (7) major land areas with coast lines (a) and minor emergent areas (b), (8-12) depositional environments and tectofacies within 6, (8) Flysch basins, (10) pelagic carbonate basins, (11) foredeep basins, (12) foreland areas.

the time when the first exhumation for both IV and IIDK took place.

Passive to transform terrane accretion may have taken place during and between these three rifts at both E and W oceanic sides (Palaeotethys and Protoatlantic Iapetus to Rheic oceans respectively) of the main ensialic Palaeozoic Europe. The most conspicuous examples of this process are the bretonian "Armorica" and the southalpine "Uralia" blocks drifting east- and westward respectively (Perroud et al., 1984; Vai, 1976; Vai and Cocozza, 1986).

The major dextral shear, located along the Palaeo-Insubric Lineament (Vai, 1976; Castellarin and Vai, 1981) and responsible for the late Devonian drift of the Uralian province fragment of the Carnic Alps and external Dinarides ("Uralia"), was suggested as able to built a strike-slip orogen with frontal collision in the present Pennidic-Helvetic realms and prograde oblique "collisions" at the northern margin of the present Southern Alps (Vai, 1980a, 1980b; Vai et al., 1984).

The Hercynian crustal consolidation achieved at the Permo-Carboniferous boundary closed the poorly developed Wilsonian Hercynian cycle started with the late Precambrian in central and southern Europe.

(2) The newly formed Hercynian crust of Europe, however, was not very stable and started its first disintegration phase along a dextral transform faults system transecting the whole European orogen during the Permian (Arthaud and Matte, 1977; Ziegler, 1982, 1984). A new series of repeated oblique rifts involved the European and, particularly, the future Tethyan realm in Permian, Middle Triassic and Liassic times (up to the connected Tethys spreading), suggesting a close comparison with the previous cycle in the same area (Vai, 1982, 1984; Vai and Spalletta, 1982).

The bearing of such rifts and of connected local, short-lived Mid- Triassic (Indosinian) orogenic conditions (Castellarin et al., 1988) is of primary importance for the thermal history of previously obducted lower crustal sections, especially those located close to major or minor rifting and spreading centers. A common feature of this fragmentation network was the relatively small size of the net mesh. This might explain the exceedingly high frequency of Permian, Mid-Triassic and Jurassic dates in virtually every Tethyan area where suitable rock types and textures were available.

The Jurassic to present evolution of the Alps and related areas has been convincingly described by many specialists to which the reader is referred to (e.g. Dal Piaz et al., 1972; Laubscher and Bernouilli, 1977, 1982). I will only remark that, unlike the IV, the IIDK underwent at least partial subduction with HP metamorphism during the eo-Alpine event (an early stage of this subduction is represented on Fig. 10) and was obducted again at the end of this event and/or during the following one. The not completely accomplished overthrusting and exhuming of the "Ivrea body" seem to be related to the eo-Alpine collision, as the "Ivrea body" appears to be truncated at the top by the neo-Alpine back-thrust of the Canavese Line (Schmid et al., 1987).

4. Discussion and Implications

Some final remarks can be focused on the basis of the above data and their interpretation.

(1) A long ranging persistence of lower crustal blocks some tens of km long and up to few thousand sq km in uncompensated isostatic setting has to be admitted as a common geologic process. Minimum survival age for such settings span from 30 to 300 Ma in the studied area, but exceeds the Phanerozoic elsewhere. The first easy explanation refers to the flexural support of an elastic lithosphere loaded by mass anomaly with a wavelength less than about 100 km (Turcotte and Schubert, 1982).

(2) The common geotectonic features characterizing the lower crustal sections examined include i) streching, rifting, rotational block faulting and formation of thinned, warm, partly buoyant continental lithosphere quite close to spreading axes, followed rapidly by ii) B-type subduction and subsequently by A-type subduction with opposite vergence. The inversion tectonics (e.g. low-angle normal faulting followed by subduction events on the same place) provides the ideal condition for squeezing lower continental crust or even mantle wedges at the surface. However, some additional features can be recognized in the study area. Exhumation of lower continental crust is not ubiquitous along this setting but appears to be localized at or close to the intersection with transform faults, fracture zones and

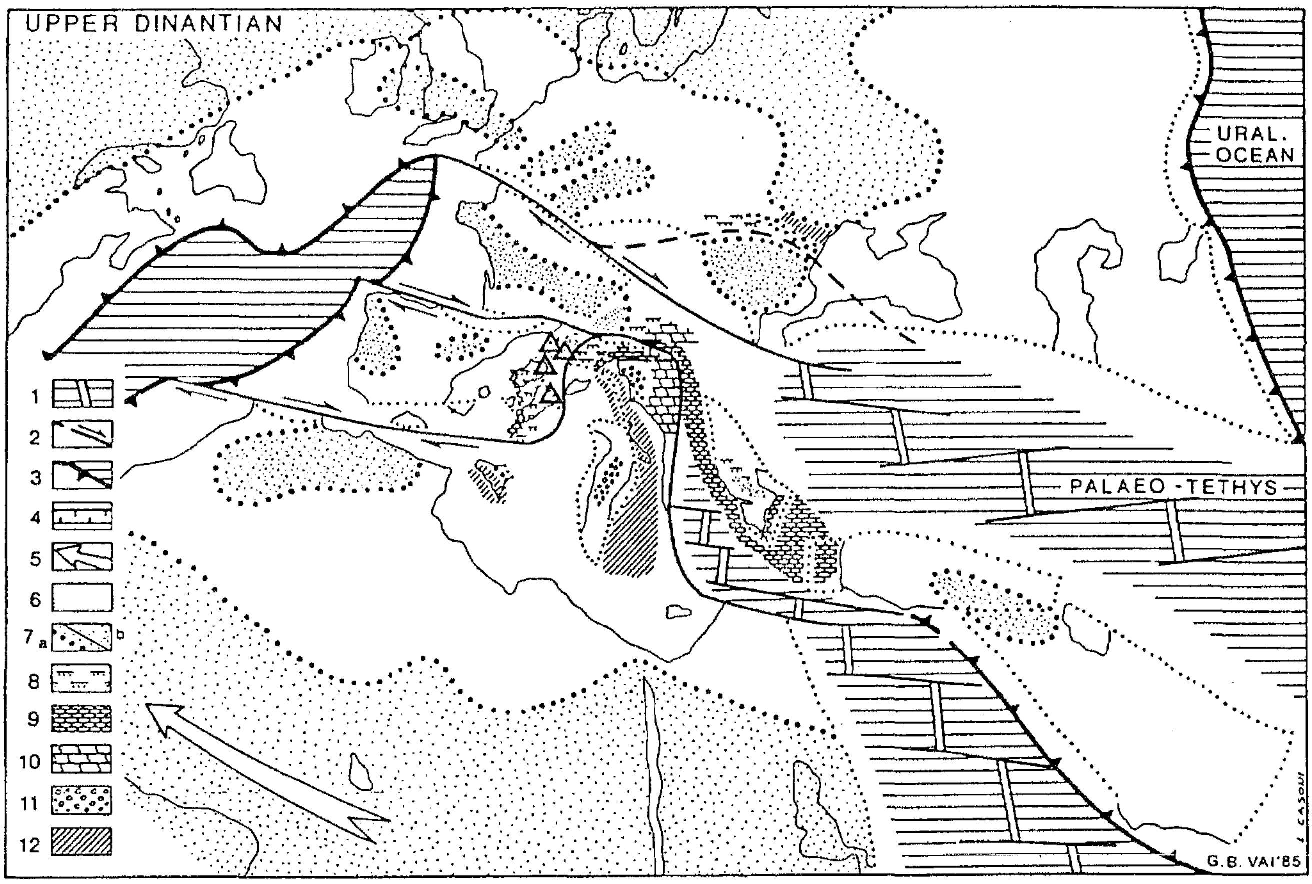

Fig. 12. Physiographic environment and tectofacies distribution in the circum-Mediterranean areas and inferred geodynamic setting during the late Dinantian. Open triangles mark the location of the four lower crustal sections described in the text; at this time they were still waiting or just beginning exhumation. (1) Oceans and spreading axes, (2) main transform and wrench faults (transpression and transtension zones not differentiated), (3) B-subduction zones, (4) main rifts, (5) main rotational vector, (6) epicontinental seas, (7) major land areas with coast lines (a) and minor emergent areas (b), (8-12) depositional environments and tectofacies within 6, (8) Flysch basins, (9) radiolarite basins, (10) pelagic carbonate basins, (11) foredeep basins, (12) foreland areas.

oblique rift belts (Figs. 11-12). Lithospheric inversion tectonics by antithetic compressional shear after lithospheric normal listric faulting (Fig. 10) and oblique convergence (with resulting vertical movements and transpressive squeezing out of discrete blocks) (Vai and Cocozza, 1986) appear as the most effective way (after low-angle extensional crustal denudation) by which lower crustal slabs can rise at the surface, because of the incremental sqeezing effect provided by transpression.

(3) The internal consistency of a Neogene obduction of the Ivrea-Verbano Zone is based on the assumption that the Pogallo Line, instead of a late Hercynian thrust (Boriani, 1970; Boriani et al., 1983), was first a Liassic low-angle extensional fault (Hodges and Fountain, 1984; Schmid et al., 1987). This assumption would require no Hercynian tectogenesis in the Southern Alps, which was, however, very severe indeed, involving shallow and deep crustal levels (Castellarin and Vai, 1981; Vai and Cocozza, 1986). Furthermore, it would require difficult explanations for the Mesozoic (Castellarin, 1972; Bosellini, 1973) and late Palaeozoic (Vai, 1980a) N-S trending facies belts in an area of E-W trending Alpine structural axes; the N-S trend in fact is interpreted as an ancestral feature inherited from the Hercynian structure (Castellarin and Vai, 1981, 1982; Vai et al. 1984). Keeping this last view, an extensional Liassic reactivation of the late Hercynian Pogallo thrust would be possible or even welcome, but it would not explain the exhumation of the Ivrea Zone on the basis of the available data. With different position and geometry, however, such an extensional fault would be relevant for the subsequent blind thrust of the geophysical "Ivrea body".

(4) Quite remarkable in the area is the recurrence in time of completely comparable settings during the Hercynian and the Alpine cycles. However, of unique importance is the recurrence of exhumations in almost the same place, the IV and the IIDK, if our reconstruction is correct. This is in good agreement with the model suggested above, and the overall geological evolution recognized in the circum-Mediterranean area, where similar processes occurred repeatedly over almost the same areas (e.g. Figs. 10-12). This remark is supported, at broader scale, by the quite similar location of major suture and spreading zones inside and around the present Atlantic Ocean (at least a first Proto-Atlantic (Iapetus), a second Rheic, and a third present day North Atlantic).

This fact has a general implication, which is, however, rather a part from the scope of this paper. It might actually explain the apparently contradictory statemet that lower continental crust is forming continuously, at any times and in very different tectonic settings (Fountain and Salisbury, 1981; Windley, 1981) and that the growth of continental crust was nearly complete (70-90% of its present mass) by 2.5 Ga (Taylor and McLennan, 1985). Continuous recycling and metamorphism of the lower continental crust in a variety of tectonic settings since its first Precambrian segregation is easier understood by the recurrence on the same place of lower crustal exhuming at different times, as recognized above for the central Mediterranean area.

Remark on Terminology and Ackonowledgements

The term "Hercynian" was preferred to "Variscan" throughout this paper in agreement with the use of Ziegler, 1984.

Actually, both terms have been originally used for given structural directions. Later on, Suess first in 1885 used the term "Variscan" for the part of the old orogenic system runningfrom the Central Massif to Silesia, and Bertrand first in 1892 used the term "Hercynian" for the whole late Palaeozoic orogenic system of northern Europe.

The priorities reported above favour the use of "Variscan" as a regional (western and central Europe) and of "Hercynian" as a wider term (including the portions of the late Palaeozoic orogen drifted to the other side of the Atlantic, those reactivated by the Alpine orogeny in the circum-Mediterranean area, plus the northwest African belt), as already suggested by Ziegler, 1984.

Criticism and proposal made by Stille in 1924 are less relevant as compared with the priorities above or with the consolidated, nearly synonymic use of both terms over the last century.

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EXTENDED ABSTRACTS

Contributions to the Geology of Italy
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THE GEOLOGY OF SERIE DEI LAGHI: A SUMMARY

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Key words: *Southalpine basement, Hercynian orogeny, Ordovician plutonism, Turbidites*

1. Introduction

The westernmost part of the Southalpine basement (Massiccio dei Laghi, Novarese, 1929) consists of two units: the Ivrea Verbano Zone (IVZ) and the Serie dei Laghi (SdL).

The Ivrea Verbano Zone, or Ivrea Zone, consists of two formations, a Mafic Complex and the Kinzigite Formation (metapelites, marble and meta-arenites, interlayered with amphibolites), ranging from upper amphibolite to granulite facies metamorphism. Mantle tectonites are associated with mafic plutonic rocks and granulite facies metasediments. This association has brought to the interpretation that IVZ is a section through upper mantle and deep crust (Voshage et al., 1990).

The adjacent SdL is the low amphibolite facies metamorphic product of a Lower Paleozoic plutonic complex and of a sedimentary sequence of unknown age with pelites (Scisti dei Laghi = SL), and arenaceous-conglomeratic deposits (Strona Ceneri Zone = SCZ). At and near the boundary between SdL and IVZ, mafic and intermediate dykes (Appinites) and granite plutons (Graniti dei Laghi) of lower Permian age occur (Boriani et al., 1988) (Fig. 1).

2. Serie dei laghi (SdL)

2.1. *Scisti dei laghi (SL)*

SL consist of micaschists and paragneisses with porphyroblasts of garnet, staurolite and kyanite.

2.2. *Strona Ceneri Zone (SCZ)*

In the SCZ three subunits can be recognized: Cenerigneisses, Gneiss Minuti and Strona Ceneri Border Zone.

2.2.1. *Strona Ceneri Border Zone (SCBZ)*

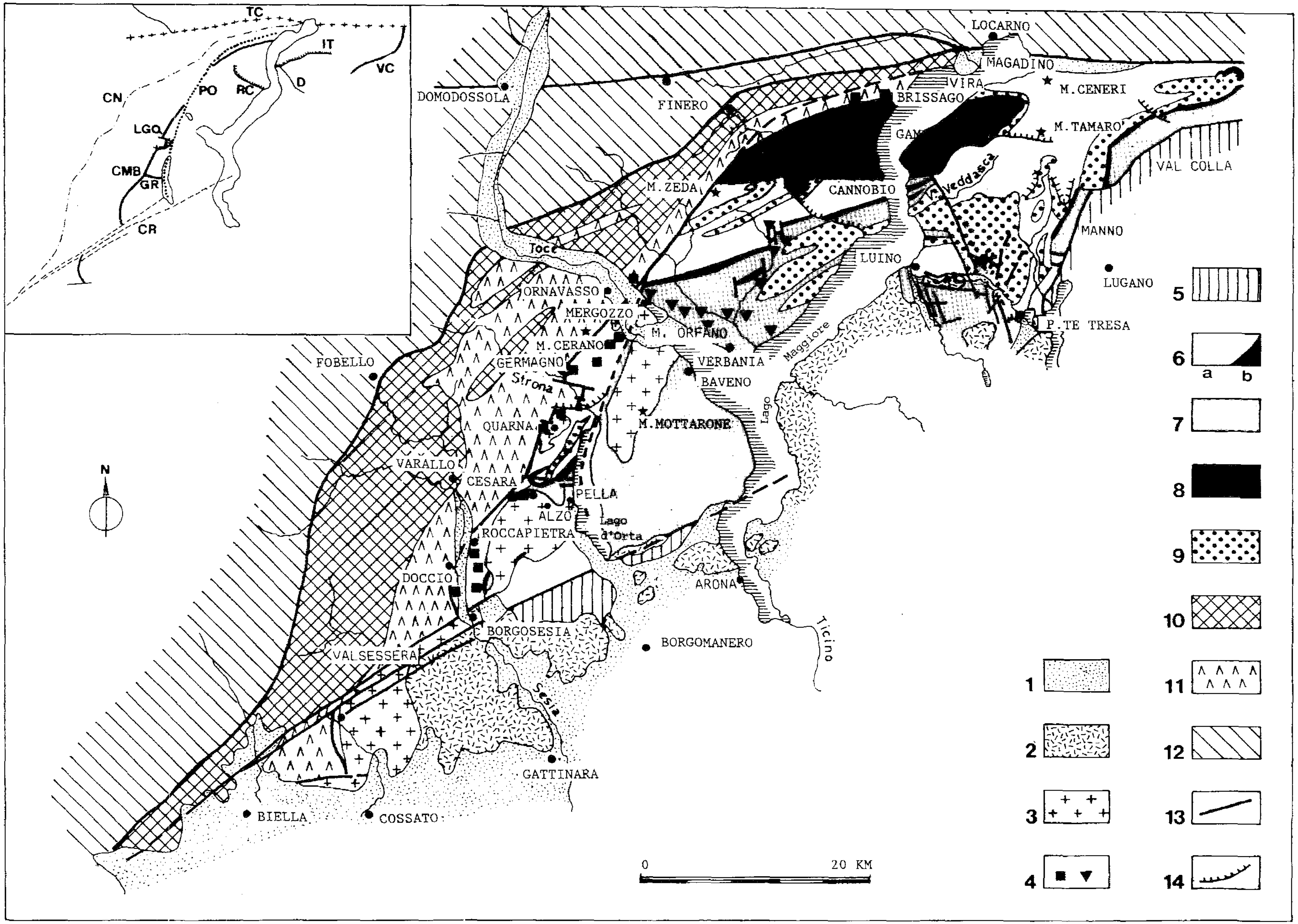
A fairly continuous horizon of amphibolites striking SW - NE marks the boundary between SL and SCZ from Lago d'Orta to Lago Maggiore (Boriani et al., 1977). The amphibolites are strongly banded and heterogeneous both in grain size and composition, and grade along and across strike into thin layered metasandstones and metapelites. In Sottoceneri, along Val Vedeggio, this horizon is segmented and bent northwards; a rather thick layer of amphibolites is also found in Val Isone. In the northernmost part of SdL, near the Insubric Line, amphibolites are also present though less abundant: rare, small lenses of garnetiferous amphibolites (Buletti, 1983), eclogitic after Borghi (1989), with tholeitic chemical character, occur in an interlayered metapelitic and meta-arenaceous sequence. Small lenses of metagabbros, pyroxenites and serpentinites are associated with all the amphibolites.

The distribution of all these rock types varies from place to place, thus indicating a different source of the material. These observations support the idea that this association may represent the metamorphic product of a turbidite, bearing clasts and olistolites of ophiolitic material. These rocks are referred to as the Strona Ceneri Border Zone in analogy with the Highland Border Complex in the Grampian Highlands of Scotland (Fettes et al., 1986). The Scottish Complex comprises a mixture of spilitic lavas, black phyllites, cherts, limestones and greywakes and is thought to be part of a dismembered oceanic crustal sequence.

W of lago Maggiore the amphibolites contain large K-feldspar porphyroclasts (Boriani and Giobbi Mancini, 1972; Giobbi Origoni et al., 1982/83).

2.2.2. *Cenerigneisses*

They are coarse metasandstones to metaconglomerates with abundant pebbles of quartz and Al-rich clasts. They contain decimetric zoned Ca-silicate nodules (Baechlin, 1937) that recall the ellipsoidal dolomite concretions of many shallow-water arenitic deposits like those of Monterey Formation in California (Durham, 1974).
Cenerigneisses display a very peculiar glomeroblastic microstructure, with porphyroclasts of plagioclase and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micas, both surrounded by polycristalline aggregates of the same minerals; tiny garnet and kyanite and/or sillimanite crystals often occur in mica-rich aggregates. K-feldspar porphyroclasts are present in the coarse-grained varieties.

2.2.3. *Gneiss Minuti (Hornfelsgneise after Reinhard, 1964)*

They are the finest-grained sediments in the Strona Ceneri. Their mineral assemblage is quartz, plagioclase, biotite, muscovite ± hornblende. They contain thin Ca-silicate lenses similar to those of Cenerigneisses. Foliated meta-aplites and metapegmatites or thin veinlets of quartz-feldspathic material, mainly discordant with the sedimentary foliation, occur throughout this subunit.

2.3. *Orthogneisses*

Elongated lenses of orthogneisses, up to 1 km thick and several km long, are abundant near the boundary between SL and SCZ (in the SCBZ). These rocks were defined (Reinhard, 1964) "Gesteine von Orthogneis typus" since they were suspected to be mainly granitized paragneisses (the intermediate type was called "Mischgneise"). They actually are (Boriani et al., 1982/83, Boriani et al., 1991) Ordovician intrusives - mostly granodiorite. Varieties with large augens of K-feldspar are found in the SCZ. Metaaplites and metapegmatites occur both in the orthogneiss and in the host rocks. Schlieren, mafic enclaves and xenolithes, among which the eclogitic amphibolites of Gambarogno, locally occur. The age of intrusion (Rb-Sr, WR) is 468 ± 5 with I.R. = 0.7086. The Sr initial ratio and the Pb isotopic data indicate a subcrustal origin and a process of mixing with crustal components (Boriani et al., 1991).

3. Structural Setting

E of Lago Maggiore the Serie dei Laghi is separated from the Alpine metamorphic units by the Insubric Line. W of the lake the boundary between SdL and the adjacent IVZ is represented by the CMB - Pogallo Fault System (Boriani et al., 1990a), whose meaning and age is highly debated.

The Cremosina - Val Colla Fault System could be considered the southern boundary of SdL, provided that the Val Colla Zone, that lies south, is really something different from SdL (Boriani et al., 1974; Heitzman and Oppizzi, 1991). This lineament is of late Carboniferous-Permian age and was reactivated in late-Alpine times (Boriani and Sacchi, 1974). The structural setting of Massiccio dei Laghi is very complicated and not yet well understood. E of Lago Maggiore Baechlin (1937) distinguished three lithological and structural domains: a) Noerdliche Injectionszone, b) Paragneiszwischenzone, c) Suedliche Injectionzone. Baechlin explained this sequence with the presence of a syncline with fold axis steeply plunging SW, whose hinge lies in the Monte Ceneri Pass area. The northern limb of this fold is in turn folded on nearly vertical axes (Schlingenbau) and locally overthrust on the intermediate part. In his geological map Reinhard (1964) accepts Baechlin's interpretation. In Malcantone (Graeter, 1951) a series of vertical faults and overthrusts divide the SdL in blocks showing different orientations (Schollen-tectonik).

Boriani et al. (1977) support the ideas of the Swiss authors extending the interpretation of the Schlingenbau to the vertical axis folds of Val Cannobina (Monte Riga Unit). Another vertical axis fold mapped W of Lago d'Orta, near Cesara, at the contact between IVZ and SdL along the CMB (Burlini and Caironi, 1988) is also attributed to the Schlingenbau. Since this style of folding seems to be present in the SdL near the CMB Line, its origin is tentatively interpreted as large scale drag folding along that fault (Boriani et al., 1990b).

New mapping in Val Cannobina and M. Ceneri regions, strengthens the opinion that the main structure of SdL is a large syncline, with hinge at M. Camoghè in Val d'Isone, whose northern limb was further refolded on vertical axes. However from our data the axis of the M. Ceneri - Camoghè fold plunges 50 SSE instead of SW as indicated by Baechlin. An alternative interpretation is that the M. Camoghè fold may represent one of the folds of the Schlingen zone.

Brittle overprints, mainly of Alpine age, further complicate the geology of this area, like the Riale di Cannero, Gambarogno, M. Bigorio and Alpe del Tiglio thrusts, the M. Tamaro transfer zone and many other accidents, in part described by Schumacher (1990). The discussion of the brittle tectonics in the frame of the Alpine

Fig. 1 - (previous page) Geological sketch-map of Massiccio dei Laghi (modified after Boriani et al., 1990).
1 - Pliocene and Quaternary. 2 - Volcanic and sedimentary Permo-Mesozoic cover. 3 - Late-Hercynian granites ("Graniti dei Laghi"). 4 - Calcalkaline mafic stocks and dykes: full squares = "Appinites"; full triangles = other mafic dykes.
VAL COLLA ZONE: 5 - Schists, phyllonites, epidote-amphibolites, "Gneiss Chiari".
SERIE DEI LAGHI: 6 - Strona Ceneri Zone (a: paragneisses, including Cenerigneisses and Gneiss minuti; b: metabasites and subordinate ultramafites of the Strona-Ceneri Border Zone). 7 - Scisti dei Laghi (micaschists, paragneisses). 8-Monte Riga and Gambarogno Zone: Strona-Ceneri and Scisti dei Laghi rocks with complex deformation. 9 - Orthogneisses.
IVREA VERBANO ZONE: 10 - Basic rocks, mainly in granulite facies, including some ultramafites and subordinate metasediments. 11 - Kinzigites (pelitic and semipelitic, high-grade metasediments, with minor marble and amphibolite intercalations).
12 - ALPINE DOMAIN. 13 - FAULTS: CN = Canavese; TC = Tonale-Centovalli; CMB = Cossato-Mergozzo-Brissago; FC = Falmenta-Cannobio; L,G,Q = Val Lessa, Germagno, Quarna; Gr = Grottaccio; PO = Pogallo-Lago d'Orta; CR = Cremosina; D = Val Dumentina; VC = Val Colla. 14 - Overthrusts: RC = Riale di Cannero; IT = Indemini-M. Tamaro.

tectonics of the Southern Alps is beyond the scopes of this abstract.

4. Evolution

Any evolutionary model of this geological domain is inevitably tentative and provisional for the following reasons:

a) a continuous updating of our model is necessary apace with the new findings of the field and laboratory studies that have been conducted by the authors, with only short breaks, since 1968.

b) any interpretation or evolutionary model of SdL implies consequences on the origin and evolution of the adjacent IVZ as well as on the whole Southalpine crust.

c) conversely, any new evidence arising from studies on other parts of the Southalpine crust helps understanding the evolution of SdL.

As a working hypothesis, the sedimentary protolith of SdL could be considered a late-to-postcollisional terrigenous sequence starting with pelagic shales (SL) followed by turbidites (SCBZ) involving an ophiolite belt, in turn followed by a molasse (SCZ). This sequence was intruded at a shallow depth, around 470 Ma ago by postorogenic calcalkaline magmas (Boriani et al., 1982/83).

In the SCZ rocks, S1 crosscuts sedimentary layering and igneous contacts, hence most probably the emplacement of the Ordovician intrusives occurred into non-metamorphic sediments. The finer grained sediments (the present Gneiss Minuti) contain aplite and pegmatite dykes, whilst the coarse-grained sediments (i.e. the present Cenerigneisses and the coarse mafic metatuffites), near the orthogneiss bodies are rich in K-feldspar porphyroclasts.

It seems very likely that hydrous residual melts of the plutonic bodies invaded the host rocks; the coarser grained sediments were not intruded by discrete dykes, but were impregnated to a variable extent: a good explanation for the origin of the K-feldspar bearing amphibolites and coarse Cenerigneisses.

The main regional metamorphism is Hercynian (Boriani et al.,1982/83). At least three important deformational phases can be recognized: F1 and F2 (often difficult to detect) in the lower amphibolite conditions, F3 ("schlingen" folds) in the greenschist facies. The PTt path (Boriani et al., 1990 and unpublished data), reconstructed only in some particular areas, shows an uplift-related decompression and the local reheating near the CMB line connected with the Permian mafic stocks and dykes (Appinites). In the belt of their occurrence the metamorphic host rocks show evidence of a post-kinematic lower pressure episode with local partial melting (Burlini and Caironi, 1988).

Several authors (Zingg et al. 1990 and references therein) maintain that SdL underwent an important tectonic event connected with the Triassic-Jurassic crustal thinning. The S-verging overthrusts of Indemini-M. Tamaro, M. Bar, Germagno, Cannero and many other brittle faults and transfer zones may be related to the late Alpine phases.

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STRUCTURAL ANALYSIS OF GIGLIO ISLAND PLUTON (NORTHERN TYRRHENIAN SEA)

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This paper is an abstract of a detailed structural study of the Giglio Island granitic pluton.

The intrusion of Giglio Island crops out in the Northern Tyrrhenian Sea. This little granitic body has an isotopic age (K/Ar) of 5 Ma (Borsi et al., 1967; Ferrara & Tonarini, 1985) and extends about 20 Km^2 occuping almost the whole island. On the northwestern and northern part of the island (Promontorio del Franco and Punta Torricella) some metamorphic rocks also outcrop (Lazzarotto et al., 1964).

A lot of aplitic and pegmatitic dykes crosscut the Giglio pluton.

The main rock-type consists of light grey medium grained monzogranite (Poli et al., 1991), which sometimes has little k-feldspar phenocrysts (max lenght 1-2 cm). Several small mafic enclaves and meter sized intrusions of light medium-grained leucogranite occur in the granitic body.

The structural analysis, whose aim is to reconstruct the fabric, has been carried out on meso- and microscopic scale. The fabric of the intrusive rock consists of planes and lineations, produced, during the emplacement of the intrusive body, by heterogeneous distribution of crystals and enclaves.

The data of Fig. 1 represent the main trends of the structural planes and lineations either measured directly in field or obtained from microscopic analysis of 30 oriented rock samples. This samples provided nearly 6500 measuraments of biotite and muscovite (001) planes. The Fig. 2 shows 23 circular histograms corresponding to the measure stations of the dyke and the joint systems.

The structural analysis of the Giglio pluton allowed to point out some interesting elements in order to interpret the evolution of the pluton during its emplacement:

1) plastic and brittle deformation of the rock; 2) heterogeneous distribution of the deformation; 3) trend of internal structures; 4) variation in dip of planar structures.

The pluton shows variable degrees of both plastic and brittle deformation.

Brittle deformation appear not only at the microscale, recorded usually by K-feldspar crystals, but also at the macroscale: in fact, the whole pluton appears strongly fractured. Also the concentration of the dykes is very high and their formation is probably contemporary, as is indicated by their reciprocal relationships.

The rock samples showing crystalline deformation are not homogeneously distributed, but rather concentrated in a wide longitudinal band in the western part of the pluton (Fig. 3), where the fabric is very complex. The structures, in fact, arranged in subcircular bodies, show sudden variations in dip of the structural planes. The Fig. 3 shows the distribution of rock deformation intensity (from minimum to maximum deformation) at microcrystalline scale, given by: I) large quartz grains assembled in subsferical (undeformed) aggregates; II) quarzt-aggregates becoming ellipsoidal in shape; III to V) from elongate shape to strongly elongate shape of quartz-aggregates, fractured K-feldspar crystals and folded biotite crystals.

On the contrary, the eastern zone of the pluton shows a regular planar trend (about NW-SE) of the fabric with a progressive rotation at its southern part. Because, in the northeastern area, there is almost no deformation of the rock at crystalline scale, we believe that these structures are mainly produced by primitive fluidal mechanisms.

On the basis of the above considerations, we can suppose that, when the magma was still partially melted, tipical flow structures (NW-SE trend) were developed, and subsequently western secondary magmatic emission centres were emplaced. The sudden variation in the dip of the structural planes in these subcircular bodies, associated to the strong deformation of the rock at the crystalline scale suggests local diapiric late movements of the highly crystallized subsolidus magma.

The secondary magmatic bodies, with plastic behaviour, were probably pressed toghether, creating deformations at crystalline scale. In the eastern part of the central zone of the pluton, in fact, the rotation of the general NW-SE strike to N-S seems to be related to the emplacement of the neighbouring subcircular structures.

High strain microstructures have been observed also

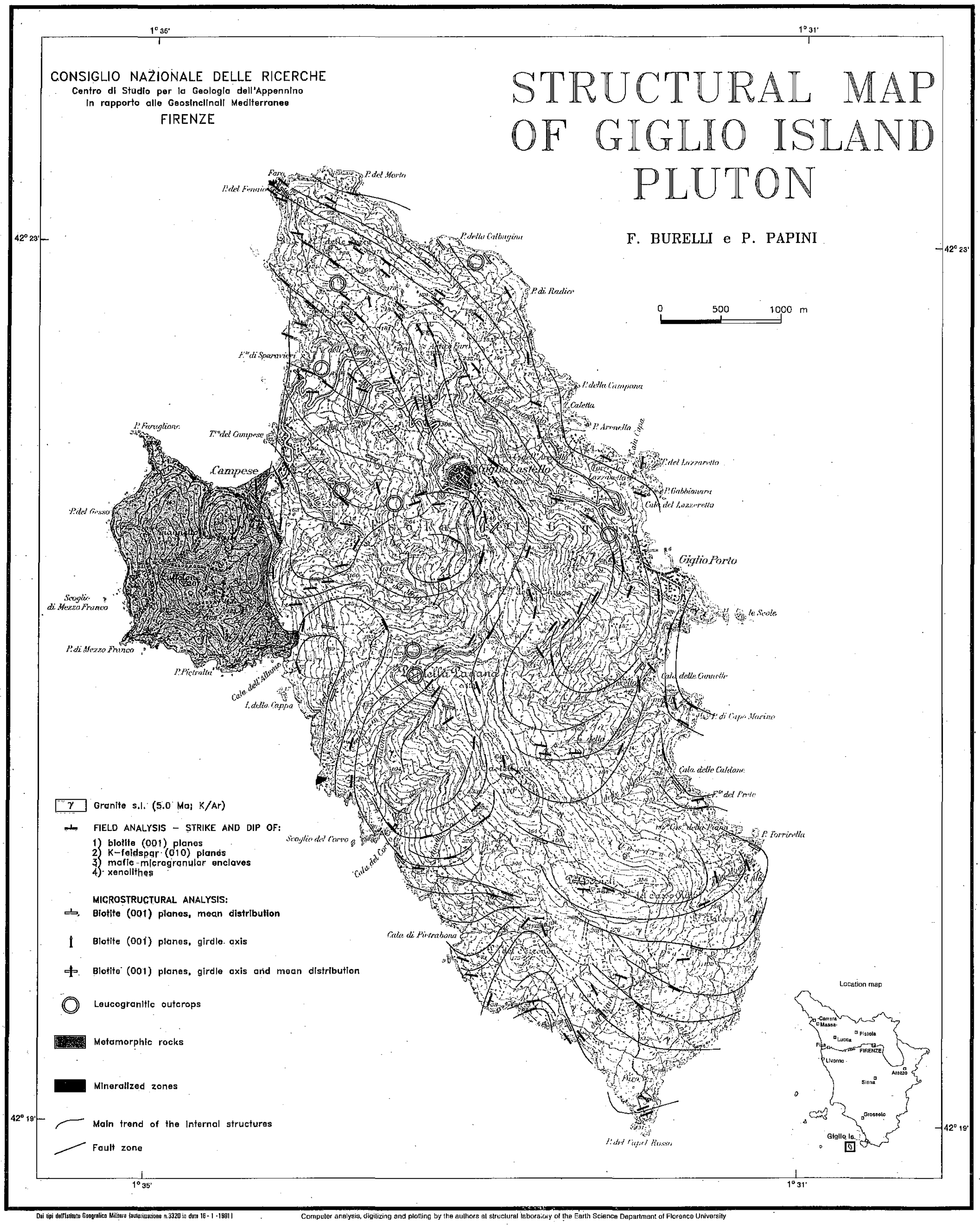

Fig. 1 - Structural Map showing the internal structures

NW of Poggio della Pagana, where the leucogranite outcrops. The microstructural analysis of this rock showed, besides a high crystalline deformation, a pronounced SW-NE isoorientation of the muscovite crystals. This orientation

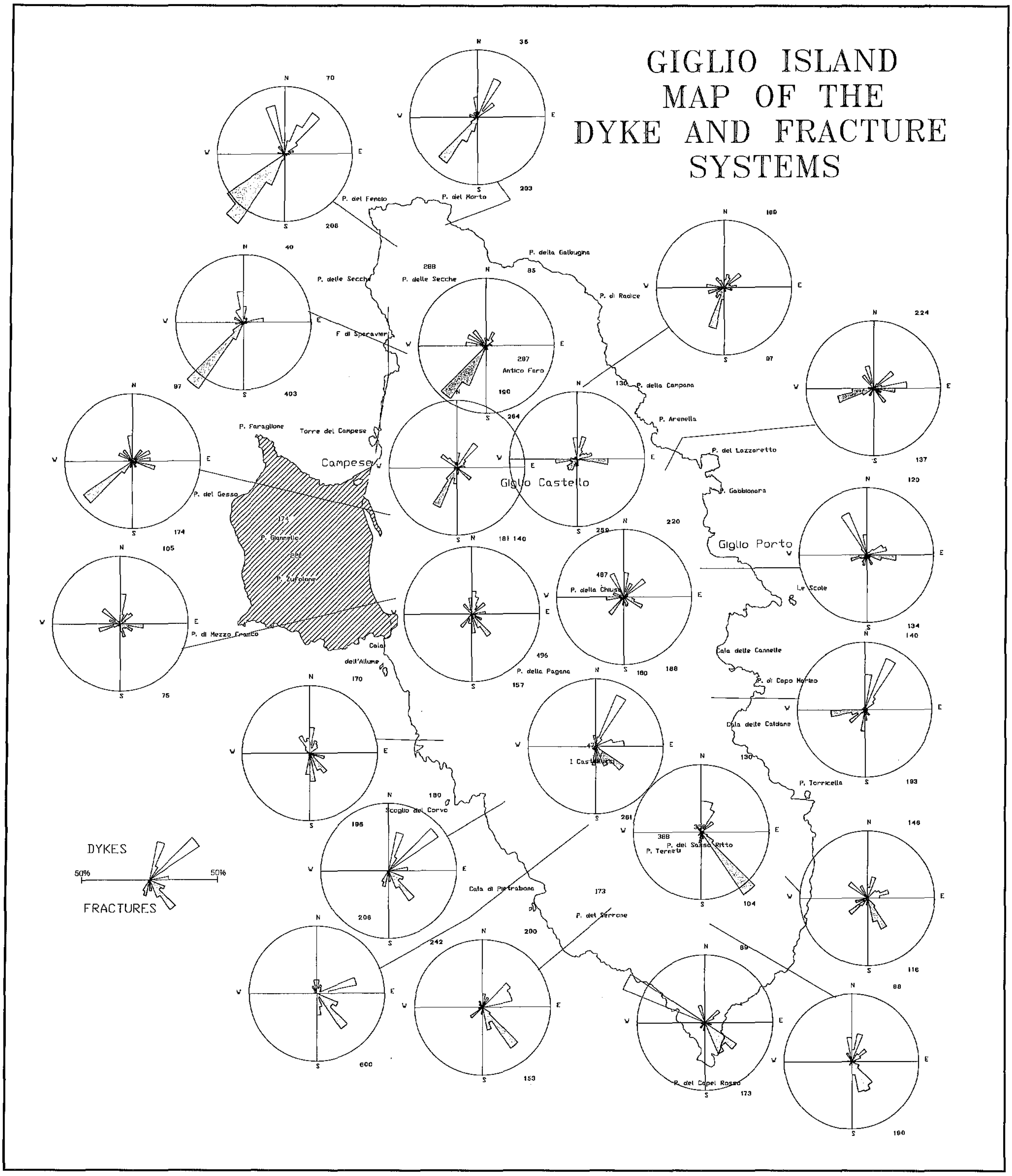

Fig. 2 - Map of dyke and fracture systems showing the measure stations (23 circular histograms; angular interval: 11° 15').

is confirmed by the mesostructures of the host granite, very evident here and perfectly concordant with the average plane resulting from the analysis of the oriented samples. Therefore, the leucogranitic mass presumably recorded, at crystalline scale, the deformations induced by the emplacement of the subcircular bodies mentioned above. Analogously, the microstructural diagram suggests that also in the zone near Giglio Castello a high overimposed strain acted and induced partial reorientation of the biotite planes.

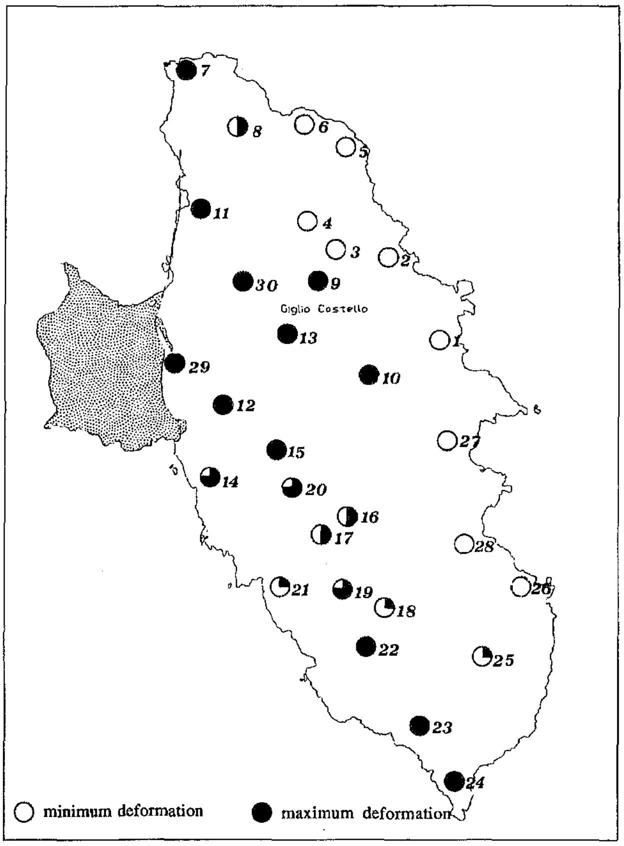

Fig. 3 - Location map of the thirty oriented rock samples with the distribution of the deformation at microcrystalline scale (circles).

We are now carrying the work further ahead by performing a statistical evaluation of the meso- and microstructural data using a appropriate computer program. By doing so, we might be able to obtain strain parameters from fabric data.

ACKNOWLEDGEMENTS

This paper is part of the research project regarding the structural analysis of the Mio-Pliocene intrusive bodies of the Tuscan Magmatic Province, under supervision of Prof. Mario Boccaletti. Funds provided by C.N.R., Centro di studio di geologia dell'Appennino e delle catene perimediterranee, publ. n° 227.

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THE EVOLUTION OF EASTERN ELBA ISLAND (SOUTHERN TUSCANY, ITALY): GEOLOGICAL, STRUCTURAL AND PETROLOGICAL DATA ON COMPLEX II.

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Eastern Elba is characterized by the superposition of several tectonic units belonging to different paleogeographic domains with different tectonic styles and metamorphic histories. According to Trevisan (1950) the units outcropping in eastern Elba between Porto Azzurro and Rio Marina can be assigned, from top to bottom, to Complex IV (Ligurid Units), Complex III (unmetamorphosed to very low-grade metamorphic Tuscan Nappe I), Complex II (low grade metamorphic Tuscan Nappe II) and Complex I (mainly andalusite-cordierite micaschist) assumed to be the local autochthonous.

Our team recently performed geological mapping and structural analysis associated with petrologic studies of the metamorphic assemblages of the three lowermost tectonic complexes.

Here we report some preliminary results on Complex II (Fig.1).

Two main geometrically superimposed units were identified in Complex II: the lower Ortano Unit and the upper Punta dell'Acquadolce Unit separated by a discontinuous and tectonically interposed "Carbonatic layer" (Fig.2). The Ortano Unit consists, from bottom to top, of light quartzite and quartzitic schist, black to dark green andalusite-cordierite micaschist; dark to light brown metarhyolite ("Porfiroidi"); light brown-grey metarkose; graphitic metasandstone; brownish and light grey quartzitic metasandstone, metasiltstone and metaconglomerate.

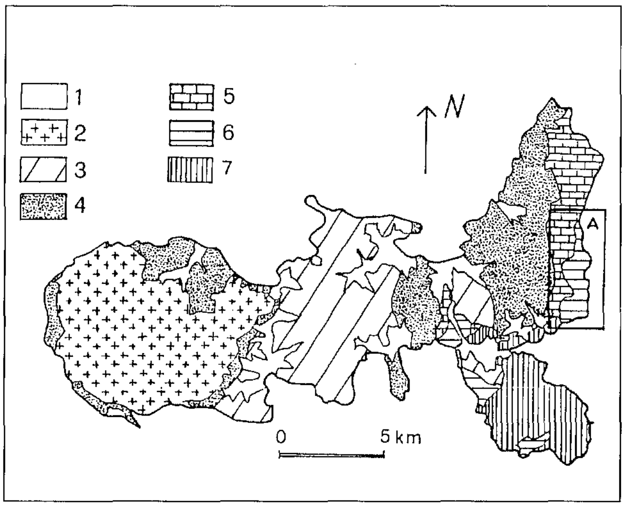

Fig.1 - Structural sketch map of Elba Island modified from Barberi et alii (1969). 1) Quaternary deposits. 2) Granitoid. 3) Complex V. 4) Complex IV. 5) Complex III. 6) Complex II. 7) Complex I. - A) Study area.

A "layer" of vacuolar metamorphosed carbonate lies with tectonic unconformity on the Ortano Unit. It is sandwiched between a breccia with prevailing phyllite elements at the bottom and marble elements at the top.

This layer is in its turn tectonically covered by the Punta dell'Acquadolce Unit consisting of white marble, greenish to grey carbonate schist, dark green-grey metapelite, and quartzo-feldspatic schist.

The Ortano Unit was in the past attributed to the Permo-Carboniferous (Trevisan, 1950; Barberi,1966) but in recent papers (Pandeli & Puxeddu,1990; Keller & Pialli,1990) it has been correlated with Sardinian Cambrian-Ordovician sequences.

Different hypotheses on the age attribution of the Punta dell'Acquadolce Unit and the "Carbonatic layer" have been proposed. Several authors consider them to belong to a Mesozoic sequence related to the "Apuan Autochthonous" (Trevisan, 1950; Perrin, 1974; Boccaletti et al., 1977; Keller & Pialli, 1990), whereas Pandeli & Puxeddu (1990), propose to compare the whole Complex II to Cambrian-Devonian sequences of Sardinia and regard the "Carbonate layer" as late Tertiary tectonic breccia.

The finding of a fossiliferous carbonatic level enabled the Punta dell'Acquadolce Unit to be dated to the upper part of the lower Cretaceous (Duranti et al., 1992).

Three folding phases (D1, D2, D3) have been recognized followed by brittle tectonics. The whole complex was later affected by contact metamorphism induced by the Porto Azzurro tertiary plutonic intrusion.

On a meso and microscale, the D1 deformation phase is recognized as a foliation transposed by later structures and it is probably related to isoclinal folds with similar

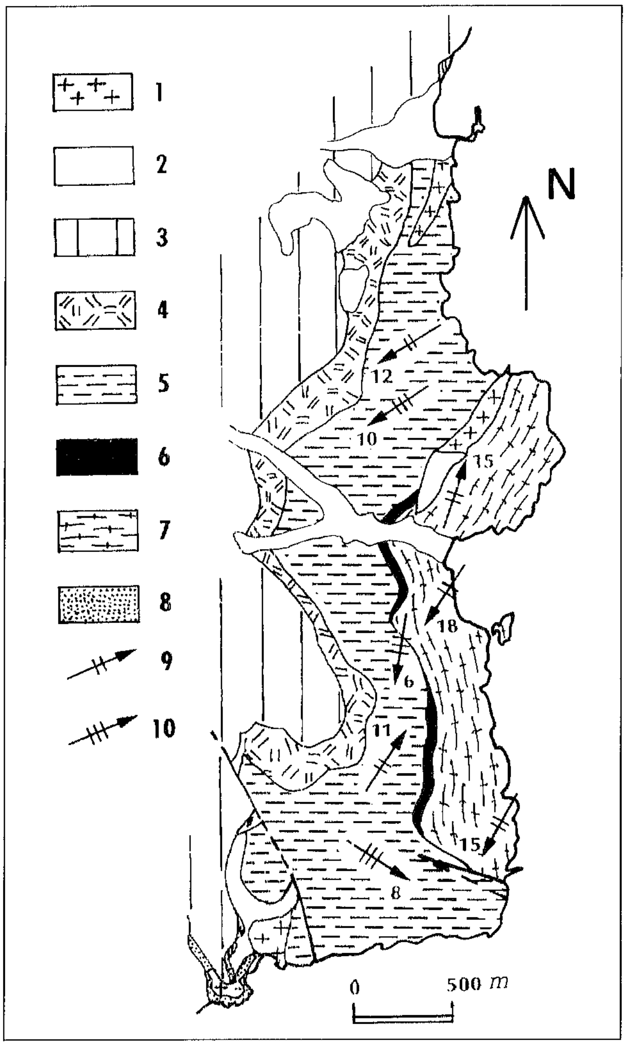

Fig.2 - Structural sketch map of Eastern Elba. 1) Ore deposits. 2) Quaternary deposits. 3) Complex III. 4) Ligurid Unit (only serpentinites). 5) Punta dell'Acquadolce Unit. 6) "Carbonatic layer". 7) Ortano Unit. 8) Mt.Calamita micaschists. 9) Second phase folding axis. 10) Third phase folding axis.

geometry. Synkinematic association of Bt (X_{Mg}=Mg/Mg+Fe≈0.45-0.50; Ti=0-0.2 a.f.u.) + Ms ± Chl develops on S1 foliation.

The D2 phase structures are the most widespread on a micro- and mesoscale. It consists of isoclinal folds with approximately similar geometry whose axial plane foliations are the main schistosity S2. The L2 mineral lineations trend N 30 plunging SW and NE and are usually subparallel to A2 axes (Fig.3).

The synkinematic growth of Ms + Bt (X_{Mg}≈0.35-0.45; Ti≈0.25-0.40 a.f.u.) on the S2 surface caracterizes the D2 phase. During the latest stages of the D2 deformation phase, several thrust surfaces develop without blastesis.

The D3 deformation phase gives rise to gentle folding of thrust planes and to asymmetric folds with steep axial plane and axial trend varying from SW in the northern to SE in the southern zone (Fig.2).

The emplacement of Porto Azzurro monzogranite (5.9±0.5 Ma Rb/Sr method, Saupè et al., 1982) post-dates the above deformation events.

The contact metamorphic effects increase from NNE (Rio Marina- P.ta dell'Acquadolce) to SSW (Capo D'Arco area). The data obtained permits the identification of three zones of increasing metamorphism in the calcschist and two in the metapelite. The zones are mineralogically characterized from the lowest to highest grade as follows:

A1) Calcite+Chlorite+Albite (calcschist)

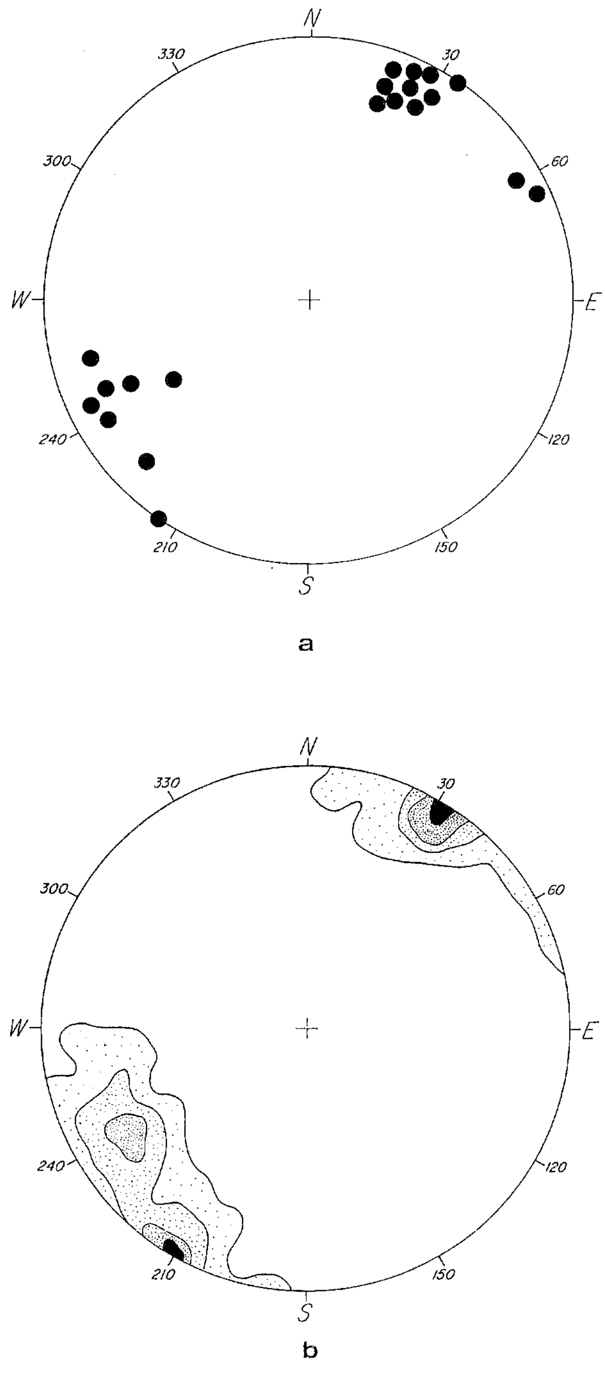

Fig.3 - a) F2 fold axis Schmidt net equiareal projection; b) L2 strechting/ mineral lineation; contours 2%-6%-10%-14%.

B1) Muscovite+Biotite± Chlorite (metapelite)

A2)Calcite+Clinoamphibole+Plagioclase (calcschist)

B2) Cordierite (X_{Mg}=0.54)+Biotite± Andalusite (metapelite)

A3) Wollastonite+Diopside (calcschist).

Metapelite associated with A3 calcschist displays the same critical mineralogical association as the B2 zone. Spessartine-rich garnet (X_{Mn}=Mn/Mg+Mn+Fe+Ca≈0.65-0.55) has been found in some quartzo-pelitic schist.

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GEOLOGICAL FRAMEWORK OF THE IANO METAMORPHIC SEQUENCE (SOUTHERN TUSCANY, ITALY).

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1. Introduction

The Paleozoic-Triassic low grade metamorphic sequence outcropping near the Iano village (Fig. 1) represents the northernmost part of the so-called "Monticiano-Roccastrada Ridge" (Costantini et Al., 1988; 1991; Conti et Al., 1991 references herein). A 1:5.000 geological mapping and detailed sedimentologic-petrographic studies on some stratigraphic sections were carried out by the Authors to define the stratigraphy and the tectonic setting of the Iano metamorphic sequences (a geological sketch map is shown in Fig. 1).

Key Words: *Southern Tuscany, Paleozoic, Triassic, Stratigraphy, Alpine Tectonics.*

2. Stratigraphic Features

The Iano "type" sequence, that shows many analogies with one described by Rau & Tongiorgi (1974) on the Pisani Mts., is summarized below (from the bottom to the top) (Fig. 2):

—"Iano Shale and Sandstone" Formation (Stephanian A-B pro parte: Vai & Francavilla, 1974): consists of parallel to lenticular grey quartzitic beds alternating with dark-grey to black phyllite. Primary sedimentary structures, such as grading, parallel to low angle cross lamination and, locally, herringbone cross-bedding are preserved in the former lithotype. Massive lenticular and erosive whitish metaconglomeratic beds, vertically passing to quartzites, are particularly frequent in the lower part of the Formation ("Borro delle Penere metaconglomerate and metasandstone" Member).The sedimentary features and the fossiliferous content (abundant plant debris and rare shallow marine organism: Savi & Meneghini, 1851; De Stefani, 1879; Vai & Francavilla, 1974) point out a deltaic-littoral environment for these metasediments.

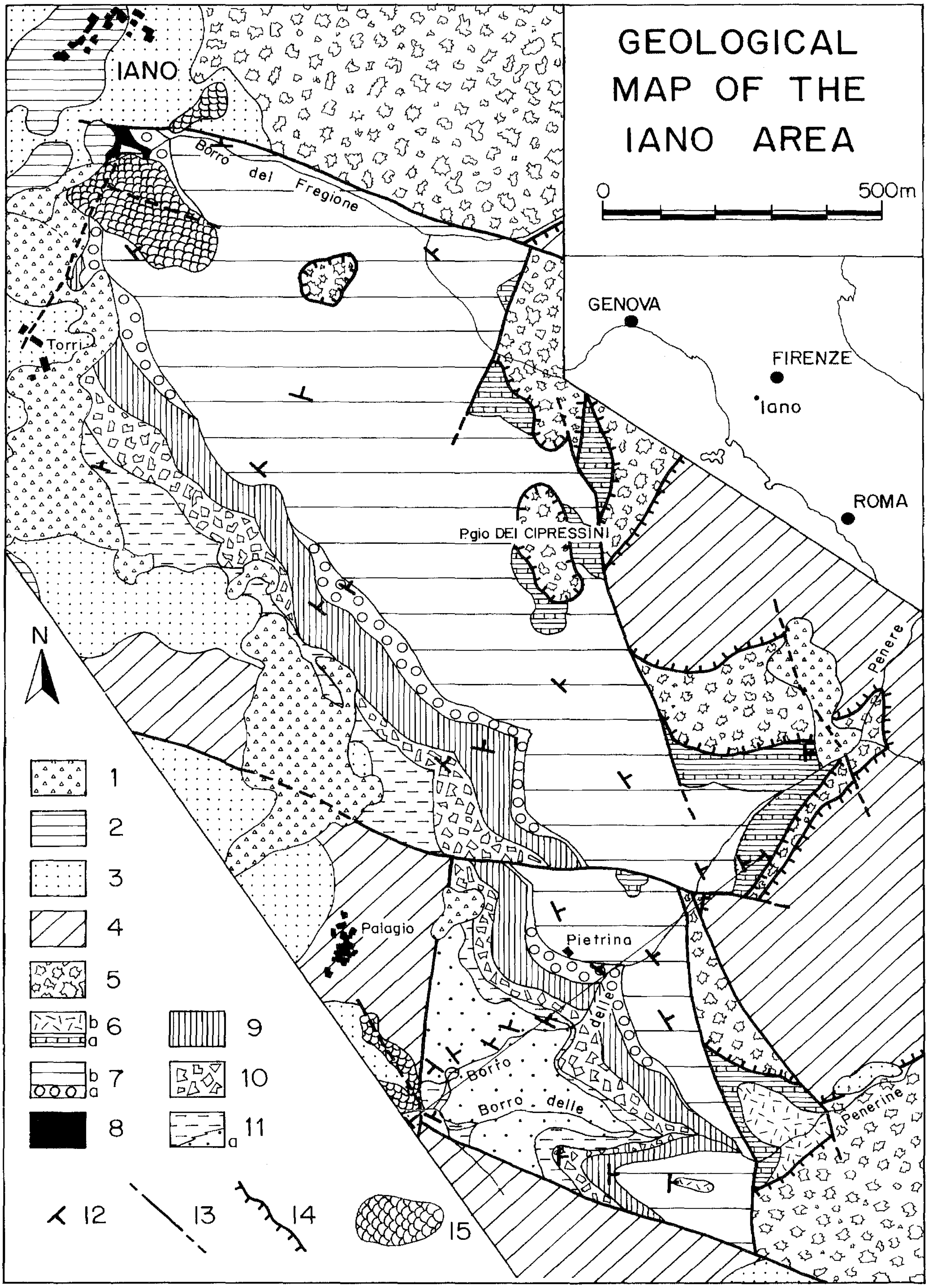
— "Torri Breccia and Conglomerate Formation" (Permian?): massive low sorted polymictic meta breccia and metaconglomerate with a hematite-rich phyllitic-matrix and abundant angular to sub-rounded hercynian clasts. The stratification is often indistinct. A continental alluvial fan environment in a sub-arid climate was hypotized for similar metasediments ("Asciano breccia and conglomerate") on the Pisani Mts. by Rau & Tongiorgi (1974).

—"Porphyric Schists" Formation (Permian?): grey - greenish phyllitic quartzose metasandstones and/or acidic metavolcanites (ignimbrites); magmatic quartz, and clasts of rhyolites and stretched pumice are widespread; a similar volcanoclastic level is also locally intercalated in the middle-upper part of the Torri Breccia and Conglomerate Formation (see Fig. 2).

— Borro del Fregione Siltite Formation (Triassic?): massive purple meta siltite have been found only in a small outcrop near the Iano village; the finding of dolomitic caliche-like nodules and the abundance of hematite pigment suggest a continental environment for these metasediments.

—"Verrucano s.s." Formation (Middle?-Late Triassic): consists of two members; the lower member (Pietrina Anagenite Member) is constituted by poorly-bedded grey - purple massive quartzose metaconglomerate beds with erosional surfaces and trough cross-stratification; the upper member (Poggio dei Cipressini Microanagenite and Phyllite Member) is made up by stratified gray pink, coarse to fine grained quartzitic metaconglomerate and metasandstone with intercalations (up to several meters thick) of purple and green phyllite with rare caliche-like nodules; finiing upward sequences and planar cross-bedding are frequently observed. A continental braided (Pietrina Anagenite) to continental coastal medium-sinuosity meandering (Poggio dei Cipressini microanagenites and phyllites) stream environments was hypothized for analogous metasediments on the Pisani Mts. (Verruca Fm.) by Rau & Tongiorgi (1974) and Tongiorgi et al. (1977).

— "Tocchi Formation" (Upper Triassic): alternation of green phyllite and yellowish impure forams-bearing intrapelmicritic carbonate beds passing upward to a massive yellowish carbonate breccia including scattered phyllitic-fragments. A continental-marine (lagoon, locally evaporitic) environment was suggested for the Tocchi Formation by Costantini et al. (1980).


GEOLOGICAL MAP OF THE IANO AREA
0
500m
IANO
Borro del Fregione
Torri
GENOVA
FIRENZE
Iano
ROMA
Pgio DEI CIPRESSINI
Penere
N
Palagio
Pietrina
Borro
Borro delle
delle
Penerine
1
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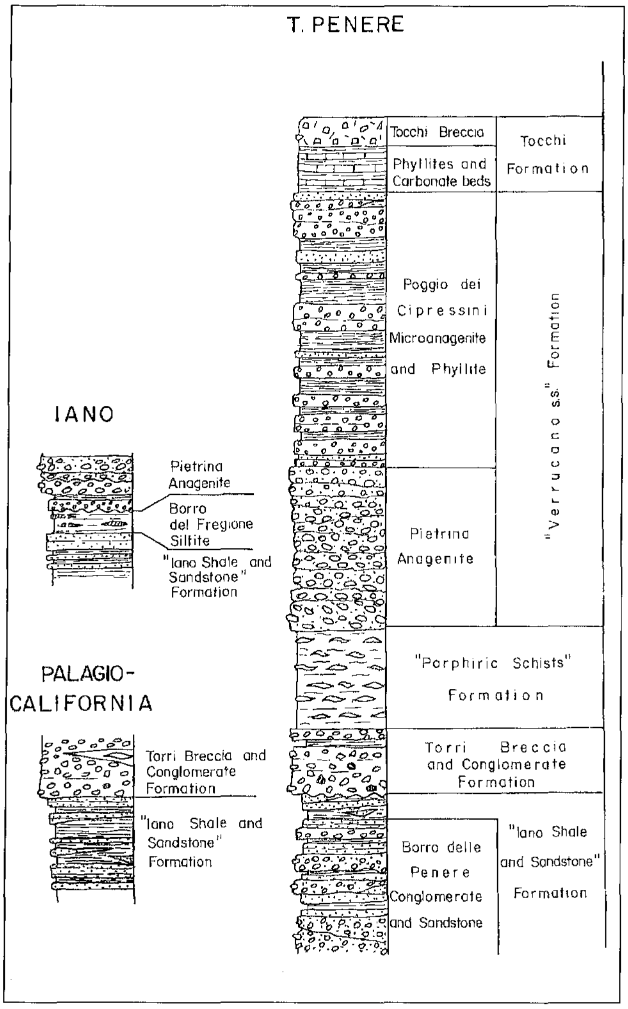

Fig. 2 - Stratigraphic coloumns of some metamorphic Paleozoic-Triassic sequences of the Iano area.

3. Metamorphic and structural relationships.

The metamorphic-structural framework of the Iano Paleozoic-Triassic sequences is characterized by the occurrence of three deformation events:

3.1 *D1 deformation event*

A penetrative S1 schistosity which makes variable angles with the S_0 stratification. On this planar anisotropy it is possible to recognize a blastesis made up by fine grained muscovite + quartz + chlorite + oxides + graphite (only in the "Iano shale and sandstone") + calcite. This mineralogical association allow us to define the metamorphic condition as low Greenschist Facies. On the xy plane (S1) rare lineations are recognizable; a N30 ÷ 50 mineralogical lineations (and/or extensive lineation L1) are instead always clearly visible. The strike of the axis belonging to the microfolds is Apenninic-trending (N120 ÷ 140) with a vergence towards North East.

3.2 *D2 deformation event*

A discontinous gradational boundaries cleavage (C2) and many mesoscopic folds are associated to the D2 deformation event. Aligments of opaques + fine grained muscovite + quartz + calcite underline the C2 planes. The strike of the D2 axes has an Apenninic direction: the relationships between D1 axes and D2 axes define a III type interference pattern (coaxial folding). In the Borro delle Penere area the D2 folds show a southwestern vergence.

3.3 *D3 deformation event*

A discontinous craks boundaries cleavage (C3) is related to the D3 event. This cleavage is generally a very discontinuous asymmetric kink. The strike of D3 axes is generally anti Apenninic - trending (N40 ÷ 80). The relationships between D2 and D3 axes allow us to define a I-II type interference pattern (Dome-Basins structures). No metamorphic association are recognizable on the C3 planar anisotropy.

4. Conclusions

The geological studies performed on the Iano metamorphic rocks point out a more complicate stratigraphic and structural features respect to ones previously described in literature (e.g. Mazzanti, 1961). In the Carboniferous sequences, a metaconglomeratic member has been distinguished (Borro delle Penere Member) and referred to distributary channels of a delta-system in which the "Iano shale and sandstone" represent the neighbouring interdistributary bays and coastal sediments. Moreover between the Carboniferous metasediments and the "Porphyric schists" a new, likely Permian, lithostratigraphic unit (Torri Breccia and Conglomerate), similar to the

Fig. 1 - (previous page) 1) Debris; 2) Travertine (Quaternary); 3) Clay, Sandy-Clay and Conglomerate (Pliocene); 4) Palombini Shales and Gabbro (Late Jurassic - Early Cretaceous); 5) Burano Anhydritic Formation (Late Triassic) and "Calcare Cavernoso"; 6) Tocchi Formation (Late Triassic): 6b - Tocchi Breccia Member, 6a - Phyllite and Carbonate beds Member; 7) "Verrucano s.s." Formation (Middle?-Late Triassic): 7b - Poggio dei Cipressini Microanagenite and Phyllite Member, 7a - Pietrina Anagenite Member; 8) Borro del Fregione Siltite (Triassic?); 9) "Porphyric Schists" Formation (Permian?); 10) Torri Breccia and Conglomerate Formation (Permian?); 11) "Iano shale and Sandstone" Formation (Late Carboniferous): 11a - Borro delle Penere Conglomerate and Sandstone Member; 12) Dip and strike of stratification; 13) Faults; 14) Thrust; 15) Landslide.

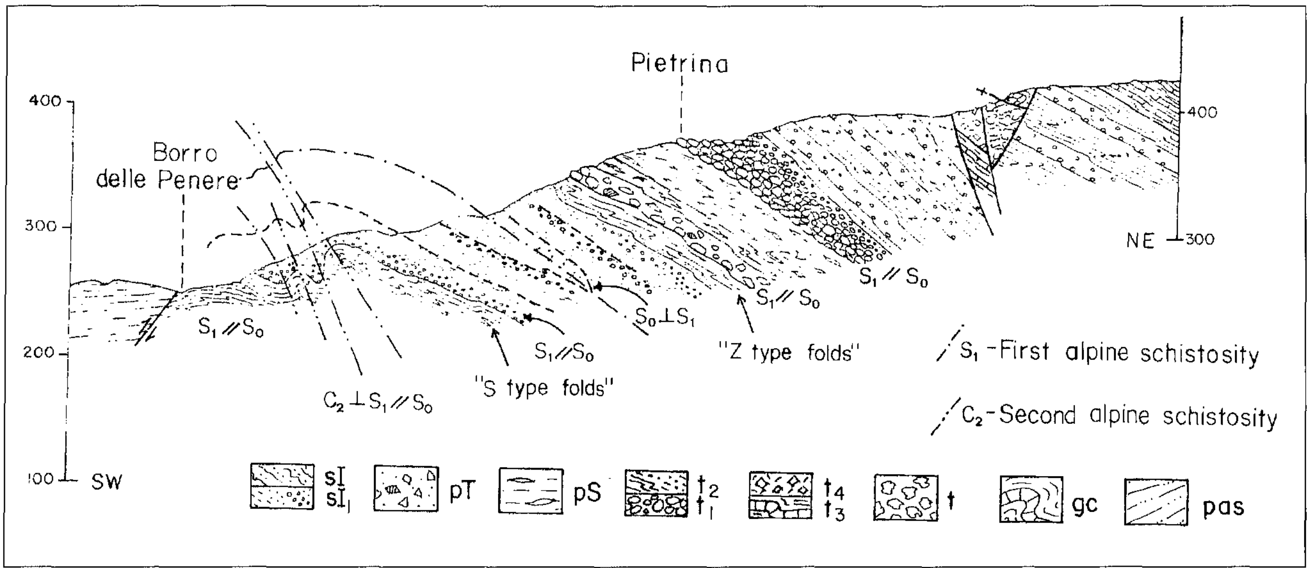

Fig.3 - Geological section across the Iano metamorphic sequence: pas - Pliocene sediments; gc - Palombini shales; t - Burano Anhydritic Formation; Tocchi Formation: t_3 - Phyllite and Carbonate beds member, t_4 - Tocchi Breccia; Verrucano s.s. Formation: t_1 - Pietrina Anagenite Member, t_2 - Poggio dei Cipressini Microanagenite and Phyllite Member; ps - "Porphyric Schists"; pT -Torri Breccia and Conglomerate; sI - "Iano Shale and Sandstone": sI_1 - Borro delle Penere Conglomerate and Sandstone Member.

"Asciano Breccia and Conglomerate" of the Pisani Mts, was defined. Also the sedimentological features of the Verrucano sequences point out a close similarity with lower part of the Pisani Mts. ones (Verruca Fm. Auct.). In the Iano Area, the Carnic sediments are instead represented, such as in most of the Southern Tuscany, by the Tocchi Fm.

The Iano Paleozoic-Triassic sequences are affected by three Alpine deformation events (D1,D2,D3). The first two syn-metamorphic in the low Greenschist facies, similar to that observed in the corrispondent sequence outcropping in the Monticiano-Roccastrada area. Howewer a structural peculiarity of the Iano sequences is the SW vergence of syn-D2 folds.

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REWORKED PALYNOMORPHS IN THE SOLANAS SANDSTONE (CENTRAL SARDINIA) AND THEIR SIGNIFICANCE FOR THE BASIN ANALYSIS

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Four eustatic regressions occurred during the deposition of the turbiditic Solanas Sandstone Formation of Central Sardinia, whose age extends from the Late Cambrian (Di Milia, 1991) to the latest Early Arenig (Tongiorgi et al., 1984; Albani, 1989).

The first event is a major eustatic regression, whose peak corresponds to the *Acerocare* trilobite Zone, near the Cambro-Ordovician boundary (ARE global regressive event of Erdtmann, 1986 = LREE event, as defined by Miller, 1984). It is considered to be of glacial origin.

Another event (Erdtmann's PRE = Miller's MBBE) which relates to a eustatic low of glacial origin was described in the earliest Tremadoc. It extends from the first appearance of the conodont *Cordylodus angulatus* to the *Clonograptus tenellus* graptolite Subzone (middle - upper part of the *Rhabdinopora flabelliformis* graptolite Zone, early Tremadoc).

A third global, probably glacial, regressive event is present near the Tremadoc-Arenig boundary (Erdtmann's "CRE"), just below the *Didymograptus extensus* graptolite Superzone, *Tetragraptus approximatus* Zone (see also Fortey, 1984 and Barnes, 1984).

Another global regression of unidentified origin (Erdtmann's "VRE") has been recognized in the latest Arenig (*Didymograptus hirundo* graptolite Zone).

Extensive debris-slides and sediment reworking from the platform toward the deep sea basins are common during the regressions phases, expecially those of glacial origin. The proximal, coarse deposits with intraformational clasts, which have been described (Minzoni, 1975; Carmignani *et al.* , 1982), in the middle part of the Solanas Formation probably are related to the two major regressions near the Cambrian-Ordovician boundary, which normally appear as a unique, very intense and long lasting event.

Actually, within the Solanas Sandstone, palynomorph assemblages corresponding to the peak of the latest Cambrian glacial regression (Erdtmann's *Acerocare* Regressive Event) and to the peak of early Tremadoc regression (Erdtmann's *Peltocare* Regressive Event) have never been recognized with certainty. That is probably to ascribe just to the large amount of reworked sediments (most of which very coarsely grained and not favourable for obtaining rich acritarch assemblages) that were discharged from the platform toward the basin, thus drowning the autochtonous microfloras, during this marked sea level minima.

Also the regressive interval at the Tremadoc-Arenig boundary is apparently not represented among the palynomorph assemblages yielded by the Solanas Sandstone: some of the above mentioned coarse deposits in the middle part of the Solanas Formation may correspond to the peak of this regression (Erdtmann's *Ceratopyge* Regressive Event).

The VRE (glacial?) event certainly corresponds to the unfossiliferous, coarse deposits (Minzoni, 1975) at the top of the Solanas Sandstone. But during the Late Arenig, tectonic mouvements (Sardic Phase) cause the end of the sedimentation and their effects are superimposed on the possible effects of the eustatic regression.

In contrast with the regressive intervals, the transgressive phases are palynologically well represented within the Solanas Sandstone: it is significant that all the rich assemblages yielded by this formation (Albani, 1989; Albani *et al.* , 1985; Di Milia, 1988, 1991; Tongiorgi *et al.*, 1984) may be referred to stratigraphical intervals corresponding to eustatic sea level highs.

The above distribution of the palynomorph assemblages (Fig.1) were to be expected, since regressive events lead to proximal, mostly coarse grained, unfossiliferous sediments, characterized by extensive intraformational reworking (paleontologically proved only in the finest lithotypes, mostly deposited at the end or at the beginning of the regressive events), whereas transgressive events lead to fine grained, slowly deposited and highly fossiliferous, distal sediments without reworkings.

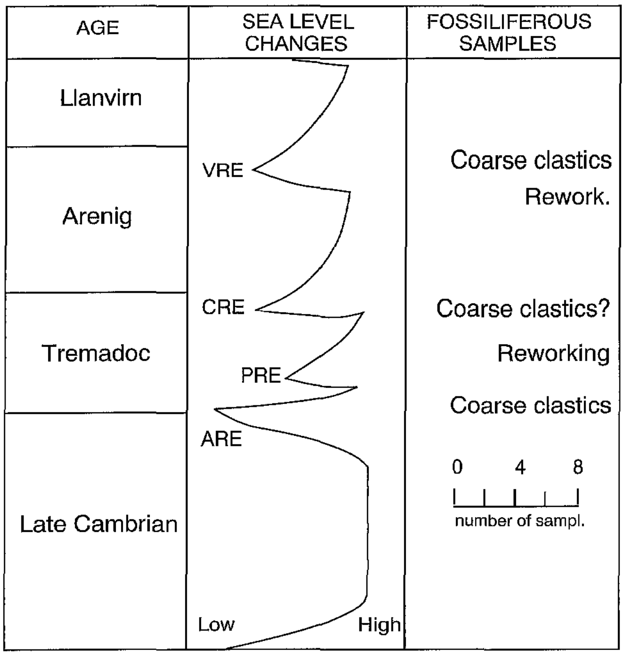

Fig. 1 - Stratigraphical distribution of the fossiliferous (acritarchs) samples yielded by the Solanas Sandstone Formation, Central Sardinia, in relation to eustacy (sea level changes from Erdtmann, 1986). In the last column, bar graph indicates the number of fossiliferous sample at each stratigraphical level. In the same column also the stratigraphical position of the coarse grained sediments and the presence of palynomorph reworking are indicated.

Consequently, it is likely that palynological dating and the actual age of the sediment do not correspond in some cases: since reworking is expected during regression phases, some palynomorph assemblages corresponding in age to transgressive intervals are probably reworked into more recent sediments, during regressions. In particular, it seems to be probable that a part of the assemblages which are palynologically referred to the transgressive interval older than the *Acerocare* trilobite Zone, are comprised in younger, upper Cambrian or lower Tremadocian sediments.

As an example, two early Tremadocian assemblages (samples S139 and S169) from the Riu Araxisi (Meana Sardo) section are here worth recording. In both samples Early Ordovician taxa (such as *Acanthodiacrodium complanatum* (Deunff) Vavrdová, 1965, *A. formosum* Górka, 1967, *A. partiale* Timofeev, 1959, *A. scytotomillei* Martin, 1973, *A.* sp. cf. *A. achrasii* Martin, 1973, *C. cuvillieri* (Deunff) Deunff, 1964, *Dasydiacrodium caudatum* Vanguestaine, 1973, *D. tremadocum* (Górka) emend. Tongiorgi, 1988, *D. tumidum* (Deunff) Tongiorgi, 1988, *Polygonium dentatum* (Timofeev) Albani, 1989, *Solisphaeridium nanum* (Deflandre) Turner, 1984, *Stelliferidium furcatum* (Deunff) emend. Deunff, Górka & Rauscher, 1974, *S. glabrum* (Martin) emend. Tongiorgi, 1988, *S. simplex* (Deunff) emend. Deunff, Górka & Rauscher 1974, *S.* sp. cf. *S. cortinulum* (Deunff) emend. Deunff, Górka & Rauscher, 1974) are associated with Cambrian taxa, such as *Leiofusa stoumonensis* Vanguestaine, 1973, *Retisphaeridium dichamerum* Staplin, Jansonius & Pocock, 1965, *Timofeevia phosphoritica* Vanguestaine, 1978 and *Veryhachium dumontii* Vanguestaine, 1973. The latter taxa indicate, as a whole, a time interval ranging between the uppermost part of the *Olenus* trilobite Zone and the middle part of the *Leptoplastus* trilobite Zone.

In conclusion, both samples are referable to the early (but not earliest) Tremadoc, i.e. to to an age corresponding to the end of the PRE regression. It also contain some representatives of a typical Late Cambrian (older than the ARE regression) assemblage.

The proportion of reworked specimens reach 10% of the total palynomorphs, at least, in both samples. But, it is reasonable to believe that the reworked taxa may be more numerous. In fact, some of the observed species, such as *Acanthodiacrodium achrasii* and *Stelliferidium cortinulum* (as cf. *cortinulum*), have a wide stratigraphical range, from the Late Cambrian to Early Ordovician.

Reworking was also identified by Albani (personal communication) within an unpublished sample from Lago Medio del Flumendosa, dated to an interval located immediately under the Arenigian *D. hirundo* graptolite Zone, i.e. at the beginning of the VRE regression.

On the basis of the exposed criteria, as shown in figure 1, regressions may be recognized within the Solanas Sandstone Formation, even if their sedimentation in a basinal environment was probably uninterrupted.

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ENCLAVES IN A PERALUMINOUS HERCYNIAN GRANITE FROM CENTRAL-EASTERN SARDINIA

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The origin and meaning of enclaves in granitic rocks is still the subject of debate, especially with regard to microgranular mafic enclaves (e.g. Chappell et al., 1987; Chen et al., 1989; Ried et al., 1983; Vernon, 1984; Didier, 1987; Zorpi et al., 1989; Poli & Tommasini, 1991). Recent research has mainly been directed to microgranular mafic enclaves found in calc-alkaline granitoids, and little has been written on the different enclave types occurring in peraluminous granitoids. In these rocks the predominant occurrence of enclaves of metamorphic origin is reported (White & Chappell, 1987; Price, 1983), but microgranular mafic enclaves are also described (Vernon, 1983; Lameyre et al., 1989).

In this paper preliminary petrographic and mineralogical data on main enclave types in a Hercynian peraluminous granite from Sardinia are documented.

The S.Basilio granite (SBG), mainly composed of monzogranite granodiorites, constitutes a peraluminous intrusion belonging to the Hercynian Sardinia-Corsica Batholith cropping out about 5 km east of Nuoro. SBG has an exposed area of about 200 sq Km, with the long axis of the intrusion oriented ENE-WSW. A weak foliation sometimes occurs (parallel to the long axis of the body) but the overall SBG is a medium-grained massive rock. A typical sample of SBG consists of quartz, plagioclase (An20-An42) sometimes oscillatory-zoned, K-feldspar, reddish biotite, cordierite, muscovite (pseudomorphing cordierite and minor primary-looking), sillimanite as fibrous mats and rare andalusite. Cordierite is irregular in shape or euhedral, and occurs both as fresh crystals and as pseudomorphs. Accessory phases include ilmenite, apatite, monazite, zircon and rare xenotime.

From a geochemical point of view SBG is characterized by SiO_2 ranging from 68.5 to 72.2 wt%, ASI (alumina saturation index) from 1.11 to 1.20, K_2O/Na_2O ratio from 1.1 to 1.3, low Rb/Sr and Rb/Ba ratios (0.30-0.62 and 0.11-0.15, respectively) and rare earth elements chondritic patterns with steep fractionation ($(La/Yb)_n$ from 28.5 to 55.6) and low negative Eu anomaly (Eu/Eu^*=0.7-0.8).

Enclaves in SBG are common and three main types can be recognized:

a) microgranular mafic enclaves;

b) mica-rich enclaves, generally schistose;

c) garnet+cordierite-bearing enclaves.

Microgranular mafic enclaves (a-type enclaves) range in size from a few centimeters to about half a meter and have a generally rounded shape. Contacts with host granite are sharp or, in more felsic enclaves, diffuse. They range in composition from tonalitic to leucotonalitic and are fine-grained equigranular or weakly porphyritic. Texture varies from "pseudo-doleritic" (Didier, 1973) with well shaped plagioclase and biotite, to microgranular polygonal, sometimes with poikylitic quartz. Microgranular mafic enclaves consist of biotite, plagioclase (An20-An64), quartz and the accessory phases, needle apatite, zircon, monazite, ilmenite and rare pyrite and chalcopyrite. Small amounts of cordierite and muscovite occur in some of the more felsic types.

Mica-rich enclaves, generally schistose, (b-type enclaves) range in size from a few centimeters up to 30cm. The shape is variable but enclaves with lenticular outline are more common. Most have schistose fabric and sharp contacts with the host granite. They consist of biotite, sillimanite (as coarse needles and also as fibrous mats), plagioclase, quartz, minor muscovite (often pseudomorphing sillimanite) and cordierite, and rare spinel (hercynite).

Garnet+cordierite-bearing enclaves (c-type enclaves) are rare rounded enclaves ranging in size from plurimillimetric to about 5cm. They consist of anhedral garnet aggregates with cordierite+biotite+quartz corona. Inclusions of a Al_2SiO_5 phase (up to 0.05mm in size), plagioclase (An22), quartz and accessory minerals (monazite and zircon) have been recognized in garnets.

All major mineral phases were analyzed by an X-ray Energy-Dispersive System (EDAX) attached to a Scanning Electron Microscope.

Biotites from a-type enclaves show higher magnesium and lower aluminium (in particular Al^{VI}) contents on the average than biotites from host granite (Figs. 1-2): XFe ranges from 0.50 to 0.61 against 0.58-0.64 and Al^{VI} (a.p.f.u. on the base of 22 oxygens) from 0.65 to 0.91 against 0.85-1.05 respectively. On the contrary biotites from b-type enclaves show similar composition to those from SBG.

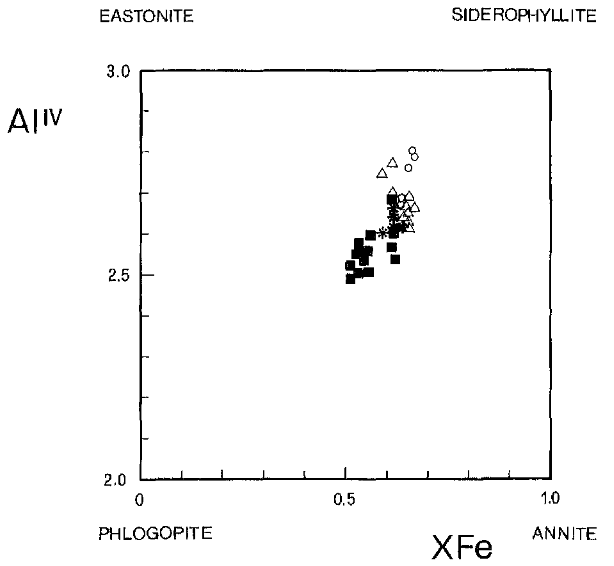

Fig.1-Biotite composition in terms of Al^{IV} vs. Fe/(Fe+Mg). Filled squares: a-type enclaves; open triangles: b-type enclaves; open circles: c-type enclaves; asterisks: S.Basilio granite.

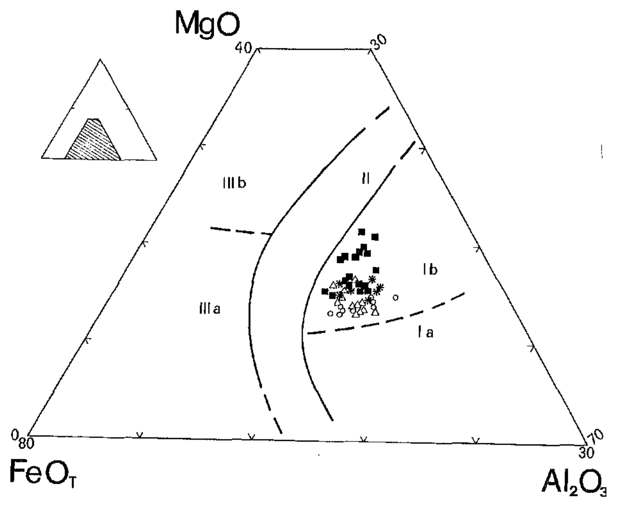

Fig.2-Ternary FeO_T-MgO-Al_2O_3 diagram for biotites (after Rossi & Chevremount, 1987). Field I: Aluminopotassique association (Ia Lymousin type, Ib Guéret type); field II: calc-alkaline association; field III: monzonitic association (IIIa Fe-potassic, IIIb Mg-potassic). Symbols as in Fig.1.

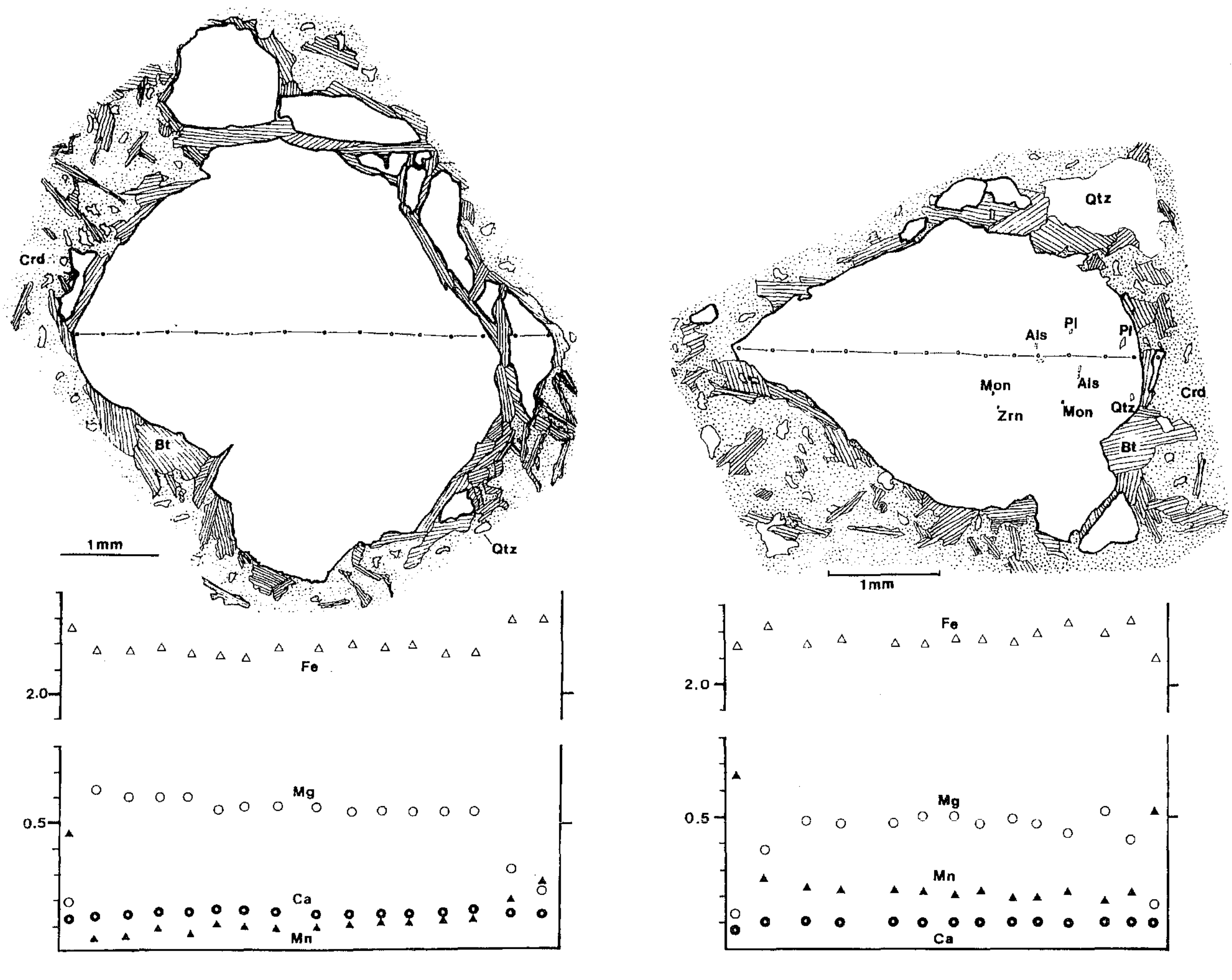

Fig.3-The zoning of garnet from two c-type enclaves. Cations calculated on the base of 12 oxygens.

Moreover biotites associated to cordierite and quartz from c-type enclaves show lower titanium contents than all analyzed biotites.

Cordierite from c-type enclaves is characterized by lower magnesium contents than cordierite from host granite (XMg ranging from 0.42 to 0.50 against 0.54-0.56 of cordierite from SBG), while rare cordierites occurring in the more felsic a-type enclaves have the same composition as cordierite from SBG.

All *garnets* analyzed from c-type enclaves are almandine-rich (XFe=0.70-0.73) and show zoning in proximity to the edges: pyrope content decreases (XMg from 0.17 to 0.04) and spessartine content increases (XMn from 0.07 to 0.22) rimwards, whereas grossularia is nearly constant (Fig. 3).

In the S.Basilio cordierite-bearing granite there are enclaves of clearly metamorphic origin and enclaves similar, from a structural and mineralogic point of view, to the well-known microgranular mafic enclaves found in calc-alkaline granitoids.

The occurrence of the latter suggests that mechanisms similar to those involved in the generation of the more common calc-alkaline suite, are involved in the genesis of peraluminous SBG.

The mica-rich enclaves may consist of more or less modified fragments of refractory lithologies; petrographic and mineralogical data suggests that they do not originate as restite complementary to the partial melt from which SBG evolved.

With regard to c-type enclaves, petrographic and mineralogical data, including the chemical zoning of garnets and SBG rare earth elements patterns, are consistent with a restitic origin of the garnet (generated by a reaction such as: Bt+Sill+Qtz+Pl=Grt+melt - Vielzeuf & Holloway, 1988) and development of Crd+Bt+Qtz corona during retrograde evolution linked to the ascent and crystallization of the granodioritic body according to a reaction similar to Crd+Bt+Qtz=Grt+melt (Vielzeuf & Holloway, 1988).

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PERALUMINOUS HERCYNIAN GRANITOIDS IN SARDINIA, CORSICA AND PROVENCE: A PRELIMINARY NOTE

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The Sardinia-Corsica Batholith and the plutonic bodies of Provence cover an area of about 13,000 sq km and intrude a pre-batholithic metamorphic basement formed during the Hercynian orogenesis. Until now the Batholith has been considered to consist of two orogenic suites (Orsini, 1976; Rossi, 1986; Rossi & Cocherie, 1991): a "*Mg-K calc-alkaline suite*" ranging from syenomonzonites to syenogranites (outcropping only in northern Corsica) and a "*composite calc-alkaline suite*" mainly composed by metaluminous intrusives ranging in composition from gabbros and some ultramafic rocks to monzogranites and biotite leucogranites (Orsini, 1980; Bralia et al., 1982).

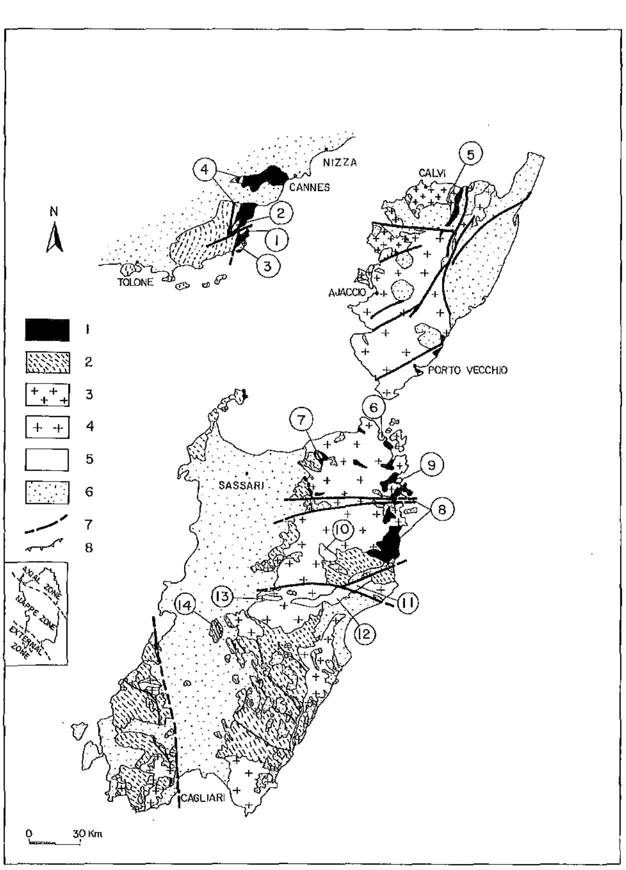

Fig. 1 - Sketch-map showing the distribution of main peraluminous intrusive rocks in Sardinia, Corsica and Provence (pre-Cenozoic drift position- after Orsini, 1980, modified). Reference numbers as in Tab.1. 1 = migmatitic complex; 2 = medium- to low-grade metamorphic complex; 3 = "Mg-K" calc-alkaline suite; 4 = calc-alkaline suite; 5 = peraluminous granitoids; 6 = Permian and post-Paleozoic volcanic rocks and sedimentary formations; 7 = faults; 8 = thrusts.

The emplacement sequence occurred in a time span of about 40 Ma (from about 320 Ma to about 280Ma) in an evolving tectonic setting postdating a major thickening stage.

The more felsic terms, sometimes garnet-bearing, of calc-alkaline suite often show a weak peraluminous character. In addition peraluminous intrusives also occur among the widespread late-Hercynian dyke system and microgranitic stocks that cross-cut the metamorphic complex and the plutonic bodies (Atzori & Traversa, 1986; Vaccaro et al., 1991). In some cases, they represent, from both field and geochemical point of view, extremely differentiated calc-alkaline magmas. The same conclusions have been achieved for Arburese cordierite-bearing leucogranites by Secchi et al. (1991).

Recent research has shown that in Sardinia some strongly peraluminous granitoids and dykes, which do not show clear field and geochemical relationship with the calc-alkaline granitoids, are also present. The main outcrops are reported in Fig. 1.

In Provence, peraluminous granitoids have previously been recognized (Vauchez & Bufalo, 1988, and references therein; Amenzou, 1988) and constitute most of the plutonic rocks (about 150 sq Km).

In Corsica, only a few peraluminous granitoids are known, among these the syntectonic Corbara Granodiorite, outcropping in the northern part of the island (Orsini, 1980; Laporte, 1987).

In Sardinia, several intrusions have been recognized over an area of about 600 sq Km.

In Tab.1, the main characteristics of these peraluminous intrusives are summarized. Data are taken from the literature and from work in progress.

These intrusives are characterized by different field, textural and mineralogical (i.e. mineral assemblages, biotite composition; Tab. 1 and Fig.2) features. They consist of dominant granodiorites and monzogranites and minor tonalites and leucogranites. The mineralogy is characterized by Al-biotite+muscovite+cordierite+andalusite±garnet and

	Intrusions	Emplacement	Age (Ma)	Lithologies	Peraluminous Minerals	Accessory Phases Radioactive minerals	Accessory Phases Opaque minerals	Enclave types	$(^{87}Sr/^{86}Sr)_i$	Ref.
PROVENCE	St. Tropez Granite	Syntectonic	-	LSG	Ms+Crd+Bt± And	Ap+Mon+Zrn +Xen	n.o.	n.o.	-	(1)
	St. Pons Le Mures Granodiorite	Syntectonic	-	GD	Bt+Ms±Crd	Ap+Zrn+Mon	Ilm	n.o.	-	(2)
	Plan de la Tour and Rouet Granite	Late-tectonic	325±10 320±15	GD-MG	Bt+Crd±Ms± And±Grt	Ap+Zrn+Mon	n.d.	Met+Mm	0.709	(4)
	Camarat Granite	Post-tectonic	287±5	MG-SG	Bt+Ms±Crd± And	Ap+Zrn±Xen	n.d.	Mm	0.713	(3)
CORSICA	Corbara Granodiorite	Syntectonic	-	T-GD-MG	Bt+Grt+Crd	Ap+Zrn+Xen+ Mon	Po+Ilm±Ccp	Met	-	(5)
SARDINIA	Tarra Padedda Granites	Syntectonic	300±7	GD-MG	Bt+Ms±And± Grt±Sil	Ap+Zrn	n.d.	-	0.7134	(7)
	Barrabisa Intrusion	Syntectonic	§ 300±4	GD-MG	Bt+Ms+Crd± And	Ap+Zrn	n.d.	Met	* 0.7074-0.7080	(6)
	Granitic Dykes within migmatites	Syn-to Post-tectonic	-	LMG-LSG	Ms+Bt±Grt	n.d.	n.d.	n.o.	-	(8)
	M. Grighini Leucogranite	Synkinematic	# 298±5	LMG	Bt+Ms±Grt	n.d.	n.d.	-	0.7135	(14)
	M. Senes Leucogranites	Late-Post-tectonic	-	LMG-LSG	Bt+Ms±Grt	Ap+Zrn	n.d.	n.o.	-	(11)
	S. Basilio-Mamoiada Granodiorite	Late-tectonic	-	GD-MG	Bt+Crd+Ms± And±Sil	Ap+Mon+Zrn ±Xen	Ilm	Mm+Met	-	(12)
	M. Nieddu-Ottana Granites	Late-tectonic	-	MG	Bt+Ms±Sil± And	Ap+Zrn+Mon	Ilm	n.o.	-	(13)
	Sos Canales Intrusive Complex	Late-tectonic	about 300	GD-MG-SG	Bt+Ms+Crd+ And±Tur±Sil	Ap+Mon+Zrn ±Xen±Ur±Aln	Ilm±Mag±Ccp	Mm+Met	-	(10)
	Rio Moronzu Leucogranite	Post-tectonic	-	AG	Ms+Grt±Bt	n.d.	n.d.	n.o.	-	(9)

§ : Muscovite + whole rock isochron; *: calculated at 310 Ma; #: internal isochron, using muscovite, plagioclase and K-feldspar. (n.o.) = not observed; (n.d.) = not determined.
AG = alkaligranite; LSG = leuco-syenogranite; LMG = leuco-monzogranite; SG = syenogranite; MG = monzogranite; GD = granodiorite; T = tonalite. Mineral symbols according to Kretz (1983); Xen = xenotime; Ur = uraninite; Met = metamorphic enclaves; Mm = microgranular mafic enclaves. Field and petrographic data from: (1), (2), (4), (5), (8), (9), (10), (11), (12), (13) and (14) unpublished data from the authors; (4) and (3) Amenzou (1988); (7) Macera et al. (1989); (6) Innocenti et al. (1991); (14) Carmignani et al. (1987). Radiometric data from: (4) and (3) Roubault et al. (1970) and Maluski (1972); (7) Macera et al. (1989); (14) Carmignani et al. (1987); (10) Di Vincenzo et al. in preparation.

Tab. 1 - Main features of some peraluminous intrusives in Sardinia, Corsica and Provence.

ilmenite and monazite as typical accessory phases. They also show different geochemical signatures (i.e. high and variable ASI values, different K_2O/Na_2O ratios and SiO_2 contents; Figs. 3-4).

The emplacement sequence is still scarcely defined. The existing radiometric data suggest that the emplacement of the peraluminous granitoids took place in Sardinia synchronously with the more abundant calcalkaline granitoids and mainly around 300 Ma (Tab.1).

On the base of field, petrographic and available radiometric data syntectonic and late- to post-tectonic intrusions can be recognized (referring to relationship among fabric, time of emplacement and regional tectono-metamorphic evolution).

— Syntectonic granitoids constitute dykes, sills and elongated plutons, generally semiconformable, ranging in composition from tonalites (rare), granodiorites to monzogranites, more or less leucocratic. They are characterized by a well developed foliation and by microstructural features partially developed under subsolidus conditions. Syntectonic granitoids are present in the high-grade basement in Provence (1-2, Fig.1) and in northern Corsica (5, Fig.1) and Sardinia (6-7-8, Fig.1). They were emplaced within previously structured migmatitic gneisses and have been deformed by the late-tectonic phases connected to the exhumation stage of the orogeny.

Some of these foliated granitoids are clearly synkinematic and connected to late-orogenic shear zones developed both in the axial and in the nappe zones (i.e. M. Grighini shear-zone; Carmignani et al., 1987).

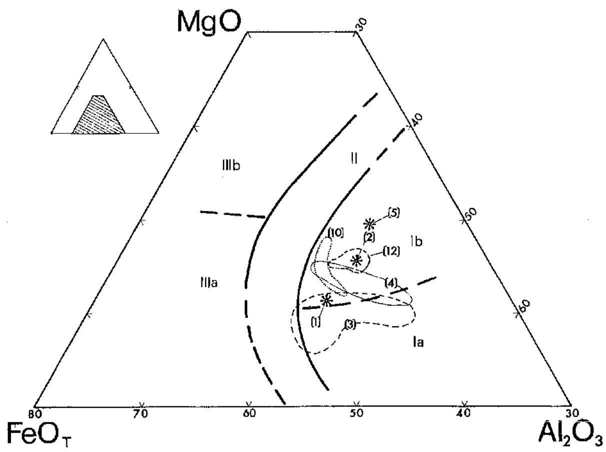

Fig. 2 - Biotite compositions in terms of FeO_T-MgO-Al_2O_3 (after Rossi & Chevremont, 1987) for some peraluminous plutons from Sardinia, Corsica and Provence. Field I: Aluminopotassique association (Ia Lymousin type, Ib Guéret type); field II: calc-alkaline association; field III: monzonitic association (IIIa Fe-potassic, IIIb Mg-potassic) Reference numbers as in Tab. 1; (4) and (3) data from Amenzou (1988).

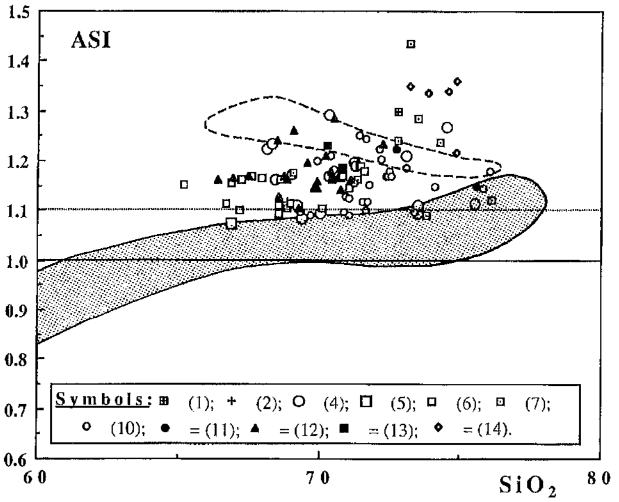

Fig. 3 - $Al_2O_3/(K_2O+Na_2O+CaO)$ molecular ratio (ASI) vs. SiO_2 wt%. Reference numbers are reported in table 1. Values higher than 1.1 are typical of S-type granites (Chappell & White, 1974). Chemical data from: (1), (2), (4), (5), (8), (9), (10), (11), (12), (13) and (14) unpublished data from the authors; (6) Innocenti, 1990; (7) Macera et al., 1989; (11) Poli et al., 1989. The field of calc-alkaline suite (dotted area, data from Bralia et al., 1982) and of S-type granites from Lachlan Fold Belt (dashed line, data from White & Chappel, 1988) are also shown.

— Late- to post-tectonic granitoids constitute stocks and large plutons ranging in composition from granodiorites to leucogranites. Moderate foliation sometimes occurs, but are often massive rocks. They are widespread all over the Sardinia-Provence area both in the high-grade and in the medium- to low-grade metamorphic zones (3-4-9-10-12-13).

Most of the peraluminous granodiorite monzogranite late- to post-tectonic plutons contain both microgranular mafic enclaves and mica-rich, often schistose enclaves (4-10-12, Fig.1; Di Vincenzo & Ghezzo, this volume), while syntectonic intrusions are generally enclave-free. In other bodies (4-6, Fig.1), rare metamorphic xenoliths are also present.

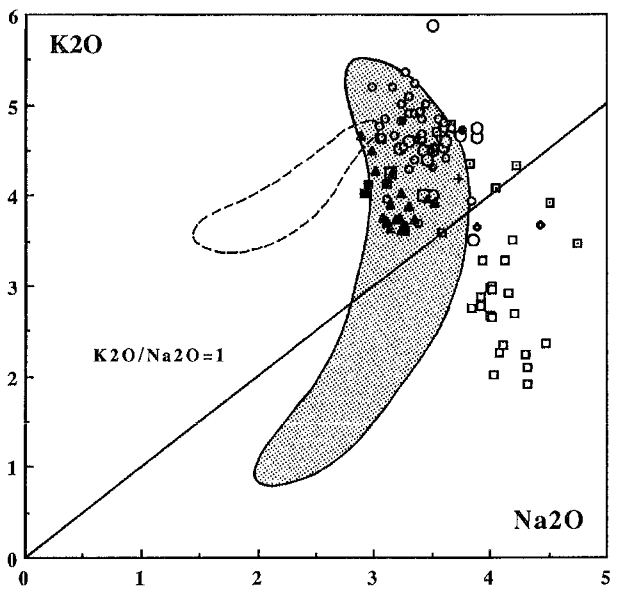

Fig. 4 - K_2O wt% vs. Na_2O wt%. Symbols and contoured fields as in fig.3.

The following conclusions may be summarized:

a) Hercynian intrusive rocks in Sardinia, Corsica and Provence constitute a composite magmatic cycle. The peraluminous granitoids, mainly characterized by Al-biotite+muscovite+cordierite+andalusite±garnet, represent a third suite that cannot be neglected.

b) The "peraluminous suite" is mainly composed of granodiorites and monzogranites and minor tonalites and leucogranites.

c) Field and radiometric data suggest that the emplacement sequence, at least in Sardinia, took place synchronously with the calc-alkaline granitoids and mainly around 300 Ma.

d) In the genesis of the peraluminous suite involvement of heterogeneous crustal rocks is required during the exhumation stage of the orogeny. Some geochemical features indicate that they cannot be considered S-type granites as defined by Chappell & White (1974): the high Na_2O content suggest involvement of dominantly quartz-feldspathic crustal rocks.

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AUGENGNEISSES FROM THE LOWER PORTION OF THE ASPROMONTE UNIT ,WESTERN PELORITANI MTS (SICILY)

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Augengneisses from the lower portion of the Aspromonte Unit - North-Western Peloritani Mts - are tectonically interposed between the underlying low-medium grade Mandanici Unit and the overhanging medium-high grade Aspromonte Unit micaschists (ky+st+grt+bt+ms) and marbles.

The augengneisses from Torre delle Ciavole (Fig.1), (kfs+plg+bt±ms+qz) show a sequence of metamorphic and deformative events different from that of other Peloritani Mts units: pre-kynematic assemblage includes euhedral oligoclase and K-feldspar cores (with cross-hatched twinning), that have been overgrown by untwinned microcline.

The latter is extremely rich of quartz inclusions, and biotite free; biotite, on the contrary, wraps around K-feldspar and defines S_1 foliation.

Small flattened paragneissic enclaves, ortho-amphibolitic boudins and pegmatitic dykes, are also present, within augengneisses.

The S_1 foliation in augengneisses coincides with the S_2 in paragneissic enclaves. Amphibolites, instead, do show only one foliation plane and are characterized by biotite rich rims, at the contact with augengneisses.

Microprobe investigations on suitable mineral pairs in textural equilibrium have given contrasting thermometric

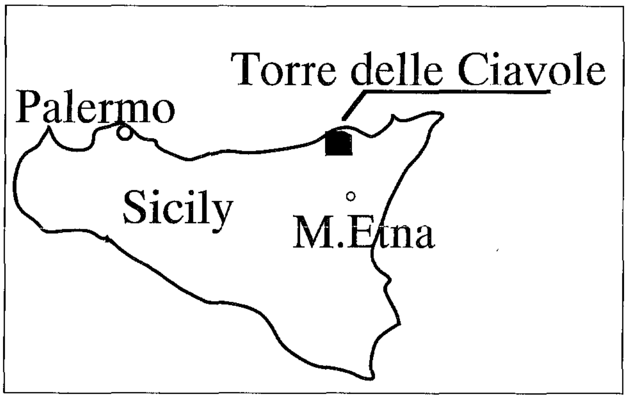

Fig.1 - Torre delle Ciavole location.

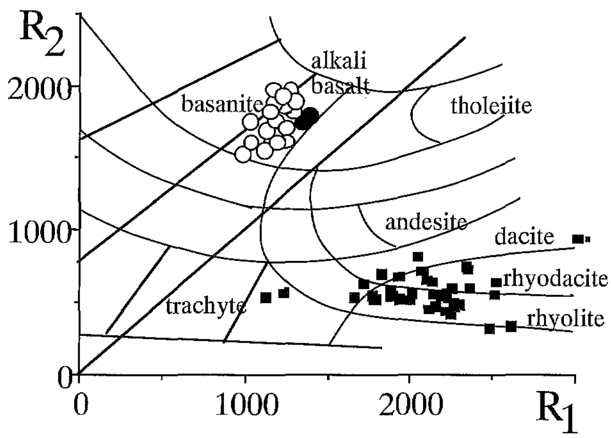

Fig. 2 - R1-R2 diagram (De La Roche,1980). Open circles : biotitic amphibolites. Filled circles : amphibolites. Filled squares: augengneisses.

values; a) the two feldspars geo-thermometer (Whitney & Stormer, 1977; Haselton et al. 1983; Price,1985) gave temperature between 330° and 410°C, probably reflecting both late stage re-equilibration and "system's opening"; calculation performed on homogeneized feldspars gave temperature close to 540°C (4 Kb); b) the Hbl-Plg geo-thermometer (Spear,1980; Plyusnina, 1982) applied to associated mafic amphibolites (hbl+bt+plg±kfs+spn) gave values ranging from 450° to 520 °C, at 4 Kb pressure (estimated on phengitic content in white micas) .

The augengneisses have a granitic-granodioritic chemical composition; the amphibolites have a alkalic basalt composition (Fig.2).

Migration of K-rich fluids, and amphibolite contamination should be proved by suitable mass balance calculation and have this aspect:

— in the amphibolites :"biotitization" of amphiboles, high K_2O content , K-feldspar blastesis, LIL elements enrichment (Fig. 3).

— in the augengneisses: syn-tectonic microcline growth over a preexisting plagioclase and/or K-feldspar .

Trace elements in augengneisses are in good agreement with syn-subductive or post-collisional derivation (Theblemont & Cabanis ,1990) or VAG (Pearce et. al . 1984) (Fig. 4), nevertheless REE are similar in augengneisses

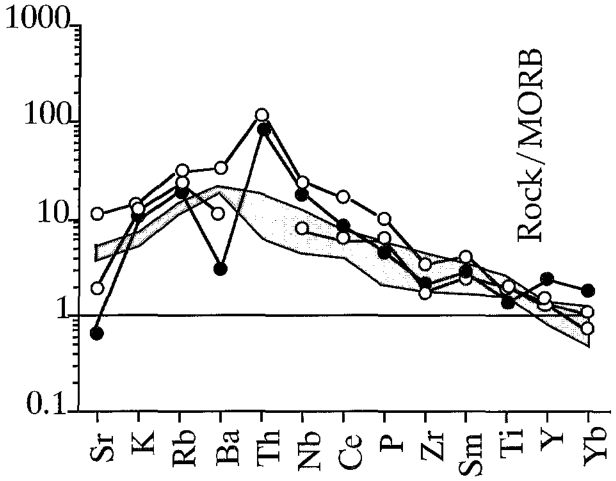

Fig 3 - MORB normalized spiderdiagram for amphibolites, MORB values from Pearce (1983), Hawaiian basalts from Wilson (1989). Open circles : biotitic amphibolites. Filled circles : amphibolite. Dotted field: Hawaiian alkalic basalts.

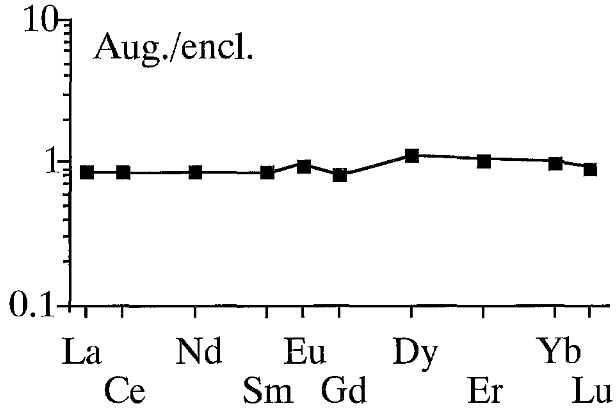

Fig. 5 Augengneisses normalized to paragneissic enclaves (mean values).

and in the paragneissic enclaves (Fig.5).

The amphibolites have a WPB geochemistry.

An S-granitic derivation should be suggested for augengneisses and is supported by outcrop geometries, textural patterns, and whole rock chemistry, ($^{87}Sr/^{86}Sr>0.720$ in the augengneisses near Messina studied by Atzori et al.,1990).

Our data support the hypothesis of various Ordovician(?) basic to persilicic magmatisms.

The subsequent intrusions of these melts into the pre-Hercynian basement could have resulted into local basic to acidic volcanic events in surface (cambro-ordovician metabasites and porphyries' clasts in post-ordovician age meta-conglomerate). Hercynian metamorphism had then originated the single foliation in these plutonics and in volcanics .

A later hydrothermal stage (chl-czo-ps-aln-cal-py) was then followed by alpine deformation (mainly brittle) that locally yielded thin cataclastic to pseudotachylitic zones.

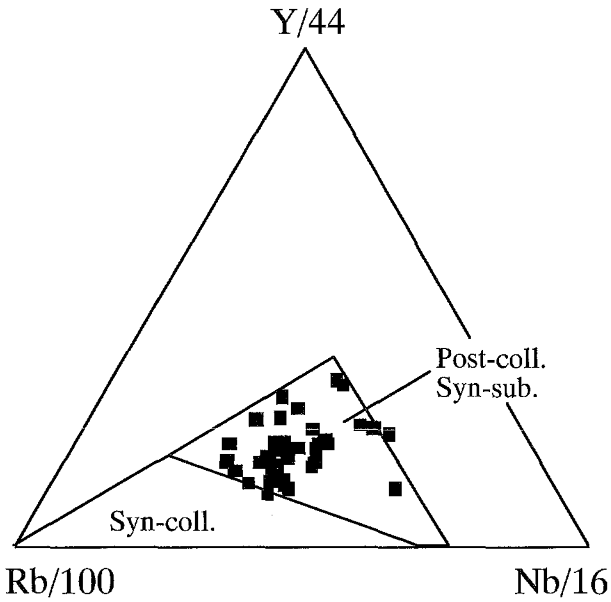

Fig.4 Augengneisses in Rb-Y-Nb graph (Theblèmont & Cabanis, 1990).

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THE GABBRO-AMPHIBOLITE COMPLEX OF CORNO BIANCO (BOLZANO, NE ITALY): AN EOVARISCAN PLUTON IN THE AUSTROALPINE OF THE EASTERN ALPS?

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ABSTRACT

The Corno Bianco amphibolites represent the metamorphosed shell of a stratified basic pluton, mainly composed of norite and olivine gabbros with minor ultramafic cumulates. The geochemical features of the magmatic rocks are similar to those of island-arc tholeiitic basalts. Comparison between these rocks and the associated amphibolites reveals metamorphic mobilization of elements usually believed to be immobile. Bearing in mind the evolution of the ötztal crystalline basement into which the pluton intruded, the age of metamorphism undergone by this pluton can only be Variscan. If this interpretation is correct, a new element may be added to the evolutionary model of the Variscan orogeny, including the formation of a Devonian-Carboniferous magmatic arc.

KEY WORDS: *Austroalpine, gabbro, Devonian, geochemistry, metamorphism.*

1. INTRODUCTION

The Austroalpine basement of the Eastern Alps contains many amphibolitic bodies which, due to their possible magmatic origin, are a great source of valuable information in the reconstruction of the geodynamic history of the Prealpine crust.

The protoliths, considered of magmatic origin, have been related to various geodynamic environments (e.g., orogenic and anorogenic) and attributed to the Caledonian event and later to the Silurian rifting (Frisch et al. 1987, Neubauer & Frisch 1988, Poli 1989). The metamorphic degree and evolution shown by these rocks varie from place to place, so that while a Prealpine polymetamorphic history has been recognised for some of them (Moghessie & Purtscheller 1986), others, e.g., the Silurian volcanites of the Graz area, are only slightly or not at all metamorphic (Fritz & Neubauer 1988).

In the Ötztal crystalline, identification of amphibolites with exclusively Variscan metamorphism is hindered by the fact that it may be of the same medium-high grade (Gregnanin & Piccirillo 1974) as the Caledonian metamorphism (Sassi et al. 1987).

2. THE GABBRO-AMPHIBOLITE COMPLEX OF CORNO BIANCO (SARNTALER ALPEN)(CBC)

Within the amphibolitic bodies occurring in the southern Ötztal, the Corno Bianco deserves particular interest known of augite and magmatic textures (Sander 1912, Briegler 1967). According to Poli (1989), the presence of magmatic mineralogy only slightly reworked by metamorphism is confirmed by the geochemical features of these rocks. Conversely, evidence of magmatic history appears to be lacking in the remaining amphibolites of the Ötztal basement.

Detailed field analysis revealed that the amphibolites described by Briegler (1967) (i.e., fine-grained schistous amphibolites with spots and/or bands) correspond to the metamorphosed shell of a moderately well-stratified basic pluton. The inner part of this intrusive body shows transition between gabbros,metagabbros with clinopyroxene relics and well-preserved magmatic fabric, and amphibolites with clear-cut, sometimes banded, metamorphic foliation. These amphibolites and metagabbros are clearly different from the amphibolites of the surrounding basement, which are closely intercalated with gneiss and some times also found as spiky inclusions in the rocks of the gabbroid complex.

The main magmatic types are: gabbros and norite gabbros (Pl 46-61% An, Cpx 55-60% En, Aug 40-45% En, Opx 60-65% En), olivine gabbros and coronite trochtolites (Pl 52-69% An, Ol 72% Fo, Cpx 45-53% En, Opx 72-78% En). Thin magmatic layering defined by alternating centimetric plagioclase and femic minerals is common.

Ultramafic layers (each layer is abaut 1 m thick), probably cu mulitic in origin, have only been found in the metamorphosed part, witch may contains actinolite alone

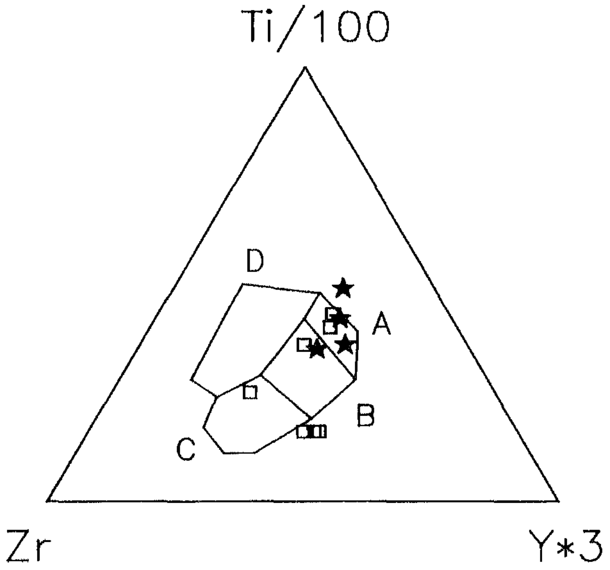

Fig. 1 - Diagram of PEARCE & CANN (1973): B=OFB, A,B=LKT,B,C=CAB, D=WPB. Stars: this work; squares: data from Poli (1989).

or serpentine with well-preserved magmatic microstructures.

3. Geochemistry

On the basis of the distribution of some elements, an sland-arc tholeiitic nature cannot be ruled out for the CBC plutonites (unmetamorphosed rocks) (Fig. 1). This interpretation agrees with the orogenic environment hypothesized by Poli (1989), on the basis of the geochemical features of the amphibolites produced by the metamorphism of these rocks. However, comparison between the composition of the magmatic rocks (this work) and their metamorphic derivatives (Poli 1989) reveal marked mobility of chemical elements during the metamorphism of the CBC. Table 1 shows some of the elements believed to be less mobile and Fig. 2 compares rare earth patterns. The mean values and variation intervals of the elements shown in Table 1 and the LREE (Fig. 2) show marked differences. The amphibolites do not even show the expected “coherent mobility” of REE (e.g., Nyström 1984), in spite of the occurrence of epidote.

Metamorphism must therefore have caused mobilization of elements believed to be only weakly mobile (excluding HREE; Fig. 2). Thus, as the CBC amphibolites do not represent the chemical features of the original magmatic rocks, they cannot be used for petrogenetic and paleo-reconstructions of the intrusive body.

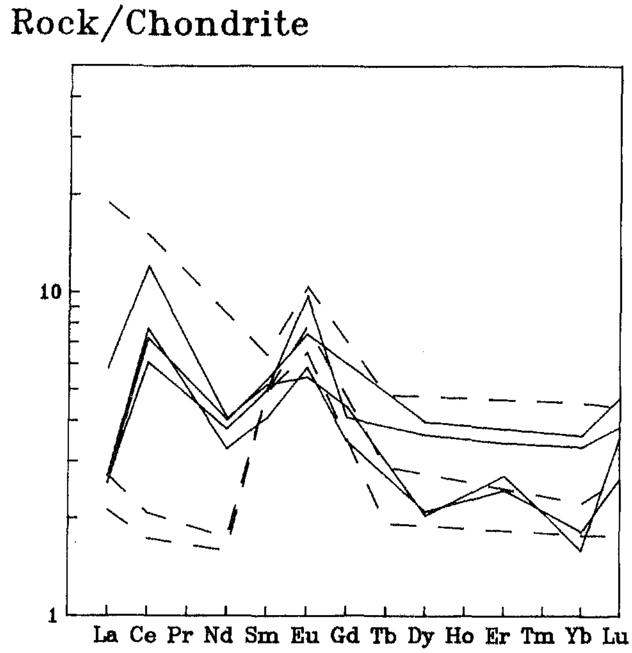

Fig. 2. - Non-metamorphosed rocks (full lines, this work) are characterized by Ce and Eu positive anomalies and tholeiite-like patterns. Metamorphosed rocks (dashed lines, from POLI 1986) show strong LREE impoverishment or enrichment. Both types of rock are characterized by Eu anomaly which reflect the cumulitic origin of pluton-forming rocks. On the other hand, the presence of some Ce-rich accessory mineral cannot be excluded to explaine the Ce anomaly. Normalizing values are after NAKAMURA (1974).

Discussion and conclusions

According to the present evolutionary model proposed by Frisch et al. 1987 and Neubauer & Frisch 1988, the southern Ötztal amphibolites (“gneiss-amphibolite association”) formed on an active continental margin or island-arc and were meta morphosed during the Caledonian event.

Regarding the metamorphic evolution of the CBC and in relation to the other amphibolites of the host basement, it should be noted that: i) the range of Rb-Sr cooling ages of the surrounding basement micas is 330-290 Ma (Thoni 1980, Del Moro et al. 1982); ii) a Prealpine polymetamorphic history has been recognized in the Ötztal amphibolites (“gneiss-amphibolite association”) (Moghessie & Purtscheller 1986); iii) weakly metamorphosed portions of the CBC contain spiky inclusions of schistous

	gabbro n=4		amphibolite n=7	
	range	x	range	x
Al_2O_3	14.4 - 16.5	15.3	14.2 - 19.9	17
TiO_2	.2 - .54	.32	.18 - 1.28	.72
P_2O_5	.07 - .14	.12	.02 - .17	.04
Y ppm	3.4 - 8.1	5.7	10 - 22	16.3
Zr ppm	6 -15	10.7	27 - 112	49.3

Tab. 1 - Comparison between some “less mobile elements” in gabbro (this work) and amphibolites (from Poli, 1989) of CBC.

amphibolites (the "gneiss-amphibolite association" which Frisch et al. (1986) consider as metamorphosed for the first time during the Caledonian). Consequently, the metamorphism of the CBC can only be Variscan.

In the Eastern Alps, the basic magmatic activity occurring between the Caledonian and the Variscan orogeneses is assigned to the Silurian and is alkaline. In the investigated sector of the Austroalpine, there is no evidence of the Silurian basic alkaline magmatism, which is located eastward, and also identified in the Southalpine (Visonà & Zanferrari 1987). In contrast with the distensive regime indicated by the Silurian basalts, the affinity of the CBC with island-arc tholeiites suggests a supra-subduction origin and thus a lithospheric compressive regime. As in the model proposed by Wurster (1988), this compression followed Silurian distension, and thus the intrusion of the CBC should be assigned to a period between the Devonian and the Early Carboniferous.

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DEDOLOMITIZATION AND RECRYSTALLIZATION OF CAMBRIAN CARBONATES OF SOUTHWESTERN SARDINIA

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1. Stratigraphy

The Lower Cambrian Gonnesa Group of SW Sardinia is subdivided from base to top into the following lithological units (cp. Bechstädt & Boni, 1989):

— The Dolomia Rigata facies of the Arcu Biasterria Member (sensu Pillola, 1990) is formed by tidal to supratidal platform carbonates (mostly fine crystalline dolomites) with stromatolitic laminated fabrics, birdseyes, oncolites, grapestones, peloids and pisolites. The small crystal size of the dolomite crystals (- 10 ¸m) and the preservation of sedimentary structures indicate an early diagenetic dolomitization, in connection with processes in a shallow marine environment.

— The Planu Sartu Member represents the slope of this platform. Finely laminated carbonates, slumped sediment bodies and debris flows are the characteristic features of this environment.

— The overlying "Grey Dolomite facies" of the Arcu Biasterria Member is characterized by middle to coarse crystalline dolomites. Algal laminations can be observed sometimes in the field. The relative coarse crystal size (- 500 ¸m) of these dolomites and the destruction of sedimentary structures by dolomitization indicates a later diagenetic dolomitization.

— The "Grey Dolomite facies" is often followed by or laterally transitional into "Black Limestones facies" of the Is Ollastus Member (Pillola, 1990), which were interpreted by Bechstädt & Boni (1989) as a flooded, isolated platform;

— The "Ceroide facies" of the Is Ollastus Member occurs at the top of the Gonnesa Group and consists of light coloured microsparitic carbonates and peloid-, ooid- and lithoclast-rich limestones. A deeper water environment was assumed by Bechstädt & Boni (1989) for the microsparitic "Ceroide facies" with the limitation, that this interpretation is hampered by the strong recrystallization of the carbonates.

2. Dedolomitization within the Arcu Biasterria Member

The studied area is situated in SW-Sardinia around the small village of Buggerru where excellent outcrops of the Gonnesa Group exist. Near to the top of the Arcu sa Cruxi fine crystalline dolomites of the Dolomia Rigata facies occur, containing more or less irregular masses (up to 10 m2) of dark grey limestones. These limestones consist of coarse crystalline calcites (with individual crystal sizes up to 1 cm), revealing a marble-like texture. The contact from these marble-like limestones to the surrounding dolomites of the Dolomia Rigata facies is characterized by a transitional zone, containing fine crystalline dolomite crystals and coarse calcite crystals.

In stained (with Alizarin-red-S) thin sections from the transitional zone, calcitic (red stained) spots can be observed within the unstained outer parts of individual dolomite rhombs. These calcitic parts luminesce under the cathode mostly yellow to orange (very rare are non luminescing parts), while the dolomite crystals luminesce red. The amount of calcitic parts within the dolomite crystals increases, approaching the marble-like limestones. Within the marble-like limestones former dolomite crystals can be recognized only by the appearance of dolomitic crystal shapes within calcites.

These features are interpreted as a result of dedolomitization from partial to complete. With increasing dedolomitization, coarse crystalline calcites (non luminescing or with a yellow-orange luminescence) occur.They replace the fine crystalline dolomitic sediment or the coarse crystalline dolomite rhombs of the ¸Grey Dolomite facies¸. Finally, complete dedolomitization is connected with the appearence of coarse crystalline, marblelike limestones containing only relics of former dolomite.

The late diagenetic (deeper burial ?) cementation sequence in pores and fissures starts with a non luminescing blocky calcite and is followed by a yellow-orange luminescing calcite. Beginning dedolomitization is connected with the formation of this non luminescing calcite, which replaces only small parts of some dolomite crystals. During the development of the yellow-orange luminescing calcite, the main stage of dedolomitization and the formation of marblelike textures took place.

3. Recrystallization within the "Black Limestones Facies" and the"Ceroide Facies.

The "Black Limestone facies" predominantly consists of carbonates with a marble-like structure and of carbonates containing coarse crystalline calcites in a fine crystalline, microsparitic calcite matrix. Under the cathode, the microsparitic matrix luminesce yellow-orange and the coarse crystalline calcites show either no luminescence or a yellow-orange luminescence. The coarse crystalline calcites are partly folded or crossed by cleavage planes. Twin laminae exhibit a kink-like structure. Similar microstructures in marbles from South-Australia were described by Vernon (1981). According to this author, the fine microsparitic matrix could represent the result of a syntectonic, dynamic recrystallization or/and a static recrystallization (i.e. grain diminution), which effect a limestone consisting of coarse calcite crystals (i.e. marble).

Marble-like structures are attributed within the Arcu Biasterria Member to a replacement of dolomite crystals by coarse crystalline calcites (dedolomitization). Sometimes relics of dolomite crystals can be observed in the carbonates of the “Black Limestone facies” as well, and some of these limestones might represent former dolomites. But the entire "Black Limestone facies" does not represent a former dolomite. Also a replacement of limestone by coarse crystalline calcites (i.e. grain growth) might be possible.

Within the “Ceroide facies”, limestones with sedimentary structures (e.g. Peloid-Grain/ Packstones) occur, as well as light coloured microsparitic carbonates with a distinct schistosity. Latter exhibit dull and yellow-orange luminescing layers. Additionally, coarse crystalline calcites are present which either do not luminesce or show a yellow- orange luminescence. These light coloured micritic carbonates are thought to be developed from carbonates with a marble-like structure and subsequent deformed by tectonic movements. The deformation seems to be more extensive than in the carbonates of the “Black Limestones facies”.

4. Conclusion

Following diagenetic events can be deduced within the Gonnesa Group:

1.- early diagenetic dolomitization within the Dolomia Rigata facies;

2.- later diagenetic dolomitization/grain growth within the “Grey Dolomite facies” and partially within the “Black Limestone facies”:

3.- a first recrystallization (grain growth) effects upper parts of the Gonnesa Group. Dolomites of the Arcu Biasterria Member (and the “Black Limestone facies”) were replaced (dedolomitized) by coarse crystalline calcites. Sedimentary limestones of the “Black Limestone facies” and the “Ceroide facies” were transformed to coarse crystalline limestones. A marble-like structure results in both cases;

4.- a second,syntectonic- dynamic and/or static recrystallization necessitates a grain diminution. This recrystallization is restricted to the uppermost parts of the Gonnesa Group (“Ceroide facies”, “Black Limestone facies”) and decreases towards the Arcu Biasterria Member.

The interpretation of the depositional environment is complicated by diagenetic alterations and is possible only in the diagenetically less altered carbonates of the Gonnesa Group.

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