A CONTINENT REVEALED

The European Geotraverse

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The northeasterly strike direction of the Maghrebide fold-thrust belt in Tunisia still reflects the geometry of the Tethyan collision suture prior to the opening of the Tyrrhenian sea. To varying degrees, the accomplished folds are lengthened along their axes and offset on strike-slip faults. These features of Pliocene and younger age (Castany 1951, Burollet 1967) are consistent with extensional tectonics.

Europe-Africa plate boundary

Near the EGT in Tunisia, global plate models (Le Pichon 1968, Livermore and Smith 1985) suggest a convergence rate between Africa and Europe of 10 to 12 mma^{-1} since the early Miocene, and a finite convergence of about 250 km. The Atlas structure suggests that much or all of this convergence has been absorbed in a detachment at the base of the sediment series. It also suggests that this implied subduction zone should involve a 200 km long NW-dipping, lower plate slab of African crust and lithosphere. Two possible sites for the surface trace of this suture are the Sardinia Channel and the south front of the Algerian Meseta block, but neither site is confirmed by a Benioff zone. There is no documented step in the southward Moho slope which could be interpreted as a suture.

The distribution of relative (residual) P-wave velocities as obtained by the seismic tomography method (Spakman 1990b) is consistent with a north-dipping subduction slab at the site of the Europe–Africa convergent plate boundary. Tomography data also suggest that the plate boundary is somehow linked to the Saharan and Tunisian Atlas and not to the Numidian thrusts and the Sardinia channel (Figure 6-33). These data suggest that the residue of a subducted slab may be represented by a deeper high-velocity body detached from the lithosphere. Extensional tectonics, suggested by crustal structure in the Sardinia Channel, is coeval with Atlasian compression and with early stages of the Tyrrhenian opening. By its setting in the upper plate of a subduction zone, this extension would qualify as a site of backarc spreading.

6.7 RECENT TECTONICS OF THE MEDITERRANEAN

D. Roeder and P. Scandone

What are the driving forces of the swirling field of orogenic belts and centres of localised extension so typical of the Mediterranean? Much of the pattern originates from the convergence between Africa and Europe (Argand 1924), evidence of which is available to us as dated plate motion paths at a resolution of 10–100 km (Dewey *et al.* 1989, Roeder 1989a). However, the topographic, paleogeographic, and geodynamic evidence suggests a more complex story (Smith 1971). Its understanding requires a look at the dynamics of compressional belts with extending and collapsing hinterlands. It requires a look at ridge push, slab pull, and sublithospheric convection patterns, and it requires a scrutiny of planetary mantle flow patterns, the subject of much of Chapter 7.

Fundamental to an understanding, however, is a knowledge of the evidence. Mediterranean tectonics is at present producing local ocean basins deeper than 3 km. It is also producing active volcanoes, seismic zones, and countless human tragedies, such as the destruction of the Minoan culture (1450 BC), of Pompeii and Herculaneum (AD79), Messina (1906), Friuli (1976), the Napoli area (1983), and Kalamata (1986) to name just a few. The system is geologically set for future disasters, although in critical areas anywhere between Budapest, Algiers, and Kurdistan we cannot easily predict their location.

A majority of orogenic belts in the Mediterranean segment of the Tethyan suture zone display a combination of convex-outward curvature, thrusting on the external side, and radial extension on the internal side (see Figure 7-6). This Mediterranean geometry is known in other parts of our planet as divergent arc (Dewey 1980) or subduction with backarc spreading (Le Pichon *et al.* 1973).

6.7.1 MEDITERRANEAN PLATE TECTONICS AND LITHOSPHERIC DYNAMICS

Our kinematic understanding of the Mediterranean became quantitative after the Mesozoic opening of the Atlantic was charted with magnetic surveys and JOIDES drilling results (Pitman and Talwani 1972). The restoration by Eulerian vector addition follows the techniques proposed by Wilson (1965) and by McKenzie and Morgan (1969). Smith (1971) showed that Mediterranean kinematics requires local plate vectors in addition to the Europe–Africa convergence. The first plate tectonic basin study of the Mediterranean (Dewey *et al.* 1973) used the global Europe–Africa convergence vectors, restorations of extensional tectonics, and available estimates of bulk strain in the fold-thrust belts. Newer attempts use improved databases. References can be found in Ziegler (1988), and improved numerical plate paths have been presented by Dewey *et al.* (1989).

Our dynamic understanding of the Mediterranean lithosphere follows the work by Le Pichon (1983) and his students, and it uses the concepts of local indentation and lateral escape flow (Pavoni 1961, Tapponnier 1977). We explain the fields of extension behind orogens as topographically determined plateau collapse (Dalmayrac and Molnar 1981, Dewey *et al.* 1986, Molnar and Lyon-Caen 1988). Advanced stages after orogenic collapse are more abundant and more obvious in the Mediterranean and are driven by the stress fields of lateral density changes between crust, lithosphere and asthenosphere (Bott and Kuznir 1979), including ridge push (Solomon and Sleep 1974, Turcotte 1983) and slab pull.

In slow or decaying convergence, subducted slabs are affected by a kinematic process called hinge retreat (Molnar and Atwater 1978) which leads to mushrooming asthenosphere and backarc spreading. The growth of these local spreading sites is governed by the geometry of the subduction zone and by points of indentation between continents. An account of areas lost and gained in convergence and backarc spreading does not reveal a direct relationship to the convergence rate. Rather, it suggests that the overall supply of fresh asthenosphere is controlled by the Europe–Africa convergence. The overall polarity of the Mediterranean orogens, their local convergence vectors, the dip direction of their subduction slabs, and their age succession perhaps suggests the effect of a global pattern, but there is no clear evidence for this.

Continental convergence

Figure 6-34 shows a collection of published plate paths of Alpine convergence since early Oligocene obtained by vector addition through time between the three plates of North America, Europe and Africa, and by vector addition between the Adriatic microplate and Africa. To illustrate the earlier effects of the Atlantic opening, Figure 6-34 also shows a plate path of Iberia relative to Africa since the Dogger age of the earliest mid-Atlantic ocean floor.

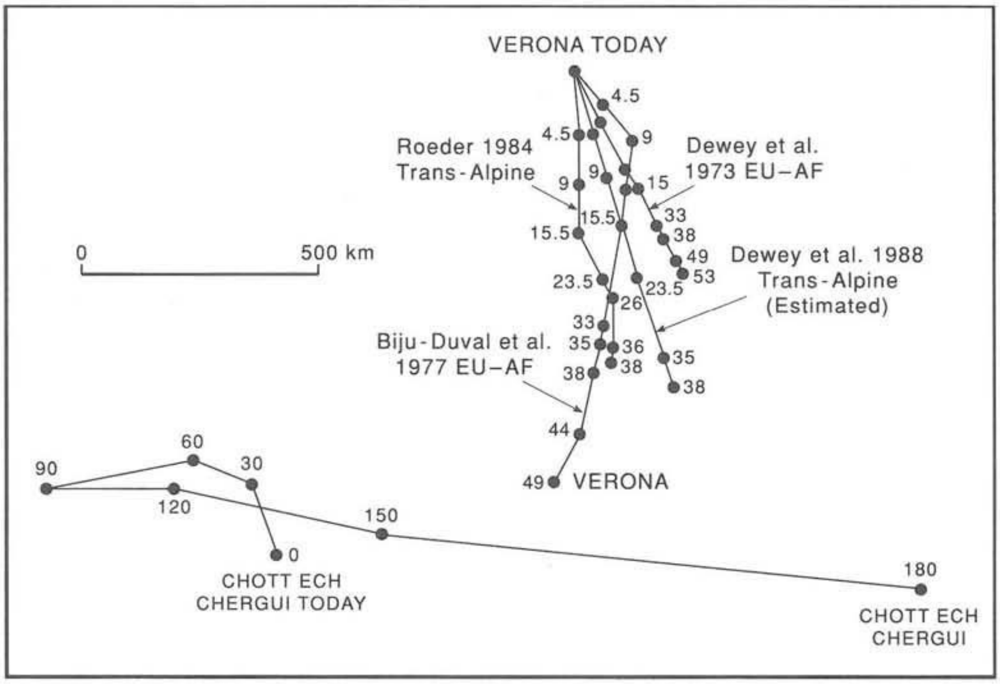

Figure 6-34. Collection of trans-Alpine plate paths projected into a vector through Verona just east of the EGT trace, redrawn after Roeder (1989a). Also shown is a plate path of Iberia relative to Africa redrawn after Roeder (1990). This path shows plate movements associated with the opening of the mid-Atlantic since the Middle Jurassic. Numbers are ages in Ma.

The trans-Alpine paths, just east of the EGT, are oriented northerly or northwesterly, and they show between 340 and 500 km of finite convergence, or between 9 and 13 mma^{-1} since collision. The convergence pattern has the shape of a spherical pie wedge. Finite and present rates increase eastward from the Eulerian pole at the Azores triple junction to nearly 50 mma^{-1} in southeast Asia (Le Pichon 1968). On a great circle through the Aegean–Hellenic system, the present rate is given as 7–10 mma^{-1} (Le Pichon 1983).

Local vectors of convergence and extension

Convergence parallel to the continental path is suggested by thrust fronts on both flanks of the Alps, on the Carpathian north front, in mountainous North Africa, and on a short segment of the Hellenide and Tauride south front. At more than half of the frontage of Mediterranean orogens, however, the thrust vectors deviate by more than 45° from the continental path. In several areas, the fold-thrust belts are accompanied by backarc extension, often at higher rates than the continental convergence. At the Hellenic trench, for example, the plate vector approaches zero (Le Pichon 1983). The vector addition becomes even more complex where the foreland of the compound orogenic belt moves independently, such as in the Apennines (Patacca and Scandone 1989).

Data for plate reconstructions in the western Mediterranean are scarce and lead to subjective and widely differing results. For individual events of backarc spreading, finite microplate paths are based on interpretively dated oceanic magnetic anomalies and on stratigraphic data at the basin flanks (Cohen 1980) as illustrated in Figure 6-35. This type of local microplate path is defined relative to the continental plate on the backside of the spreading site. The associated convergence vector at the front of the spreading site can be measured only by structural analysis of the fold-thrust belt.

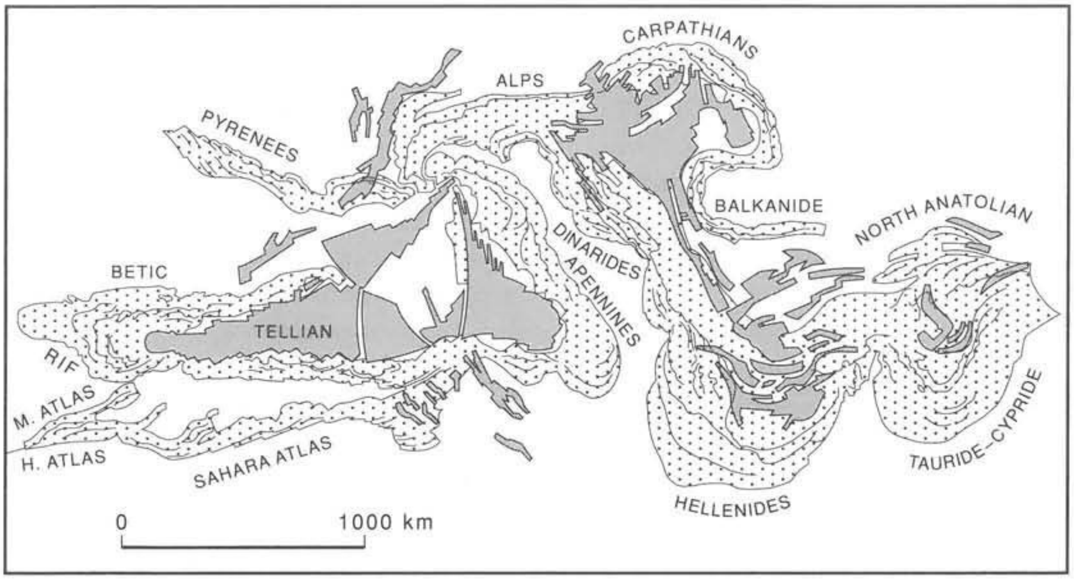

Figure 6-35. Map of Mediterranean orogens (stippled) in Recent plate configuration, showing Neogene sites of extension and/or backarc spreading (dark shading), largely based on Boccaletti et al. (1982). Coastlines are not shown.

Path of Adriatic microplate

The common foreland of Apennines, Southern Alps, and Dinarides, measuring roughly 1000 km by 700 km, the Adriatic microplate is a cratonic fragment of the early Mesozoic passive extensional margin of the Tethys embayment between Europe and Africa. Since the Oligocene, this fragment became partially and increasingly wedged between the converging continents. Since the Miocene it has been indenting Europe in a hard collision and has, we assume, been transform-sliding along, or converging across, Ionian oceanic crust connected to the African plate. Its plate path is critical to its role of indenter and to the dynamics of Apennine–Tyrrhenian backarc spreading. However, its plate path cannot be determined because it has only convergent and transforming plate boundaries. Any record of early divergence from African plate terranes is now obliterated by burial beneath the thick Ionian sediments.

Quantitative attempts at determining the paths of the Adriatic microplate and other fragments have been based on paleomagnetic data suggesting counterclockwise rotations against the global magnetic field, of the Apennine foreland in SE Italy and of the island chain of Corsica and Sardinia (Vandenberg and Zijderveld 1982). Livermore and Smith (1985) have obtained a path by determining the rotation required to restore the Apennine foreland from its present known location and orientation to its assumed initial location and orientation. Anticlockwise rotation of west Mediterranean terranes around a pole near the indenter point of the Adriatic terrane has left behind a trail of wedge-shaped backarc basins. In this conceptual model, the pivot and indenter point are assumed to show some movement relative to Africa and to move northwestward, almost along the continental path and almost at the Europe-Africa convergence rate, relative to Europe.

6.7.2 THRUST-BELT ARCHITECTURE

Convergent plate boundaries do not support vector addition because subduction destroys the record of the plate path. Fold-thrust belts, however, can supply estimates of bulk strain (Dahlstrom 1970, Mitra 1986). In fold-thrust belts, the strain consists of detachment and overlap of the supracrustal sediments along a ramped low-angle thrust fault, of detached folds, and of imbrication within the detached supracrust. Commonly, fold-thrust belts can be restored to 200% of the present, tectonised width. Foredeep fill is generated by erosion of the thrust sheets, and it can date the emplacement of marker tectonic units. Bulk-strain data and emplacement ages can produce a dated strain path of convergence. Figure 6-36 illustrates this with an E–W cross section from Corsica to Adria.

The physics of fold-thrust belts is determined by the equilibrium between the strength of the crust and the traction at the detached sole. This relationship can be quantified as thrust-wedge dynamics (Dahlen *et al.* 1984 and many others) by using Chapple's concept of 'critical taper'. In a new context, the thrust-wedge dynamics provides inroads into the puzzle of synorogenic extension. It should be possible ideally to define a convergent plate path by the strain path obtained from a fold-thrust belt. However, two additional tectonic processes limit this use of fold-thrust belts.

First, additional subduction zones may dislocate the basal detachment of a fully developed fold-thrust belt. This has been postulated in the north-central Apennine (Royden and Karner 1984) by the flexural geometry of the Pliocene foredeep fill. The implied crustal break is also visible in seismic data, and it extends southwards into the Tyrrhenian Benioff zone. It adds an unknown distance to the plate path.

Secondly, the lithosphere may detach in the weak middle crust and may create plate convergence without supracrustal expression. An intracrustal blind thrust is suggested by seismic data and seismicity data along the Alpine north front (Mueller *et al.* 1980). It may also be developed at the contact between the Adriatic microplate and the Ionian oceanic crust SE of Sicily. Intracrustal detachment is also needed to explain the effect of Tyrrhenian extension in the Tunisian Atlas.

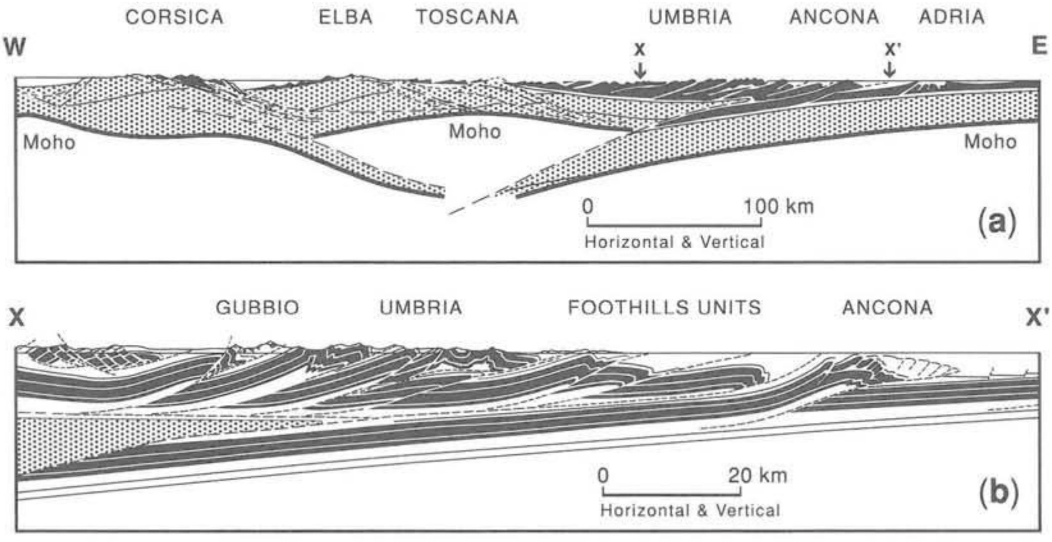

Figure 6-36. (a) Structure cross section of the Umbrian segment of the Apennines showing a fold-thrust belt detaching and imbricating a sediment series of Triassic to Miocene age. Redrawn after Roeder (1991).

(b) Detail of cross section (a) between points X and X'.

6.7.3 PLATEAU COLLAPSE AND ASTHENOSPHERE PUSH

Fold-thrust belts moving under conditions of critical taper accumulate hinterland terrains coalescing into a large, high, and extending orogenic plateau. In the model of Molnar and Lyon-Caen (1988), extensional tectonics limits the height of the plateau to 4–5 km, depending on the rheology, the rate of convergence, and the amount of erosion. In a steady-state process, the orogenic plateau grows laterally as a field of extensional shards, and its marginal fold-thrust belt advances over its foreland. The growing crustal root leads to isostatic uplift, to more top extension, and to thermal softening of the deeply buried felsic root terranes.

Figure 3-36 is a cross section of the Umbrian segment of the Apennines east of Corsica. Flexure of the foreland basement is documented in offshore seismic data and in the geometry of the Pliocene foredeep fill. It suggests an intracrustal Benioff zone piggybacking the internal (west) part of the fold-thrust belt and its crustal substratum. This subduction is also documented in a seismically mapped Moho offset and in the Calabrian Benioff zone.

Plateau collapse is driven by crustal convergence and controlled by the density contrast between felsic or supracrustal rocks and air, that is, by the same mechanism that controls the critical taper of fold-thrust wedges. This mechanism does not work in the Mediterranean sites of extension shown in Figure 6-35, because the sites are topographically lower than the forelands of their fringing fold-thrust belts. However, at sites of thin lithosphere, there is a lateral density gradient of the high-risen mantle asthenosphere adjacent to thicker lithosphere and crust with felsic and supracrustal orogenic rocks. The compressive stress exerted by this gradient easily exceeds the basal traction required to move a fold-thrust belt and build up or maintain its topography (Bott and Kuznir 1979, Turcotte 1983, Le Pichon 1983, Chapter 7.1). Because this force also exists in the flanks of oceanic spreading ridges, it is sometimes referred to as ridge push (Solomon and Sleep 1974).

Despite their fundamental difference, both forms of synorogenic extension occur in the same geological domain. At least some of the Mediterranean extensional sites may have started as collapsing orogenic plateaus. For example, the largely compressional and high-rising Western Alps grade continually into the low and largely extensional Pannonian basin. Where is the switch from deep mantle and plateau collapse to shallow mantle and ridge push?

This problem is not solved, but it leads to considering two more aspects of the lithosphere, namely metamorphic core complexes and indentation or slip-line tectonics.

Metamorphic core complexes

Extension of lithosphere is accompanied by thinning and by buoyant rise of its layers. Sharply localised fields of extension and tectonic uplift have been described in much detail from the North American cordillera as metamorphic core complexes (Crittenden *et al.* 1980, Coney 1980). Metamorphic haloes accompanied by extensional tectonics are known from several Mediterranean sites of orogeny-related extension, such as in the northern Apennines (Carmignani and Kligfield 1990), in the Pelagonian massif in the Hellenides (Le Pichon 1983), in the Menderes massif in the Taurides of Turkey (Dewey *et al.* 1986), and in the Pannonian site (Horvath, pers. comm.). There may be more core complexes to discover.

The metamorphism in some of the Mediterranean sites has been interpreted as burial beneath excessive overburden (Carmignani and Kligfield 1990) or as unspecified orogenic metamorphism (Jacobshagen 1986). In its modern form (Roeder 1989a,c and others), however, the Cordilleran model directly relates the metamorphism to the extension. As

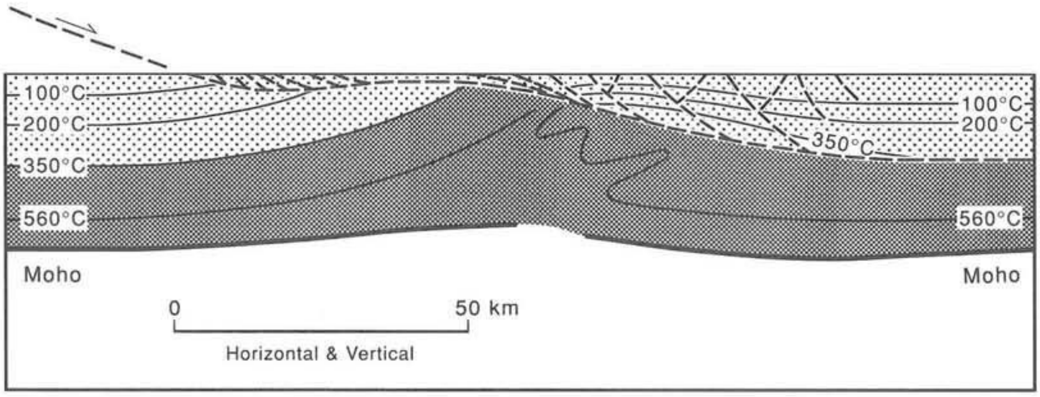

Figure 6-37. Schematic cross section of a metamorphic core complex showing the thermal effects of large-scale extension by trans-crustal low-angle normal faulting. The thermal structure, with selected isotherms, is shown ideally at the instant of accomplished extension. Stipple: elastically reacting upper crust, shading: viscous lower crust. The hangingwall is extended and warped. The footwall is flexed elastically by buoyant upwarp. Its hot basal area is in contact with shallow parts of the hangingwall. Redrawn after Roeder (1989a).

shown in Figure 6-37, the setting for this model includes a three-layer lithosphere, an average or less than Barrovian geothermal gradient, and major extensional bulk strain. Dip slip in the order of 100 km along a trans-crustal or trans-lithospheric low-angle detachment is accompanied by smaller-scale extension in the hangingwall and by buoyant and viscous uplift of the footwall. The buoyant rise of the footwall compensates for the load of the tectonically removed overburden. The upper-crustal or lithospheric parts of the footwall are upbent elastically. The viscous underpinning rises in the shape of a pillow. This setting can achieve amphibolite-grade metamorphism in the thinned hangingwall at the surface. It can also generate conduits for granitic melts into the hangingwall, but it cannot migmatise hangingwall rocks.

The temperature at the top of the buoyant viscous pillow is that of the lower crust cooled during the removal of the hangingwall. If the viscous uplift is 15 km or more, and if its strain rate exceeds the cooling rate, amphibolite-grade metamorphism and granitic wet-melt conditions (Dallmeyer *et al.* 1986, Snoke and Miller 1987, and others) can be generated in the shallow and extended hangingwall rocks. At an average dip of 15° on the detachment, this uplift requires 60 km of extension, and one-dimensional thermal modelling suggests that the extension must take place at rates well above 10 mma^{-1}.

Depending on the size of the extension and on its duration, the lithospheric effects of extension by rising viscous pillows may include rise of the Moho, flattening of orogenic Moho roots, rise of the thermal top of the asthenosphere, and incipient sea-floor spreading. The sites of extension in the lower lithosphere may be located directly below the stretched upper crust or may be found elsewhere in the system transferred laterally along detachments in the viscous lower crust.

Indentation and escape

During a collision, crustal edges with promontories and embayments will generate regional orogenic complications (Dewey and Burke 1973). During the subduction of a jagged crustal edge, the lithosphere beneath crustal promontories will sink more slowly than the lithosphere beneath the embayments. Unless the lithosphere detaches from the crust, it

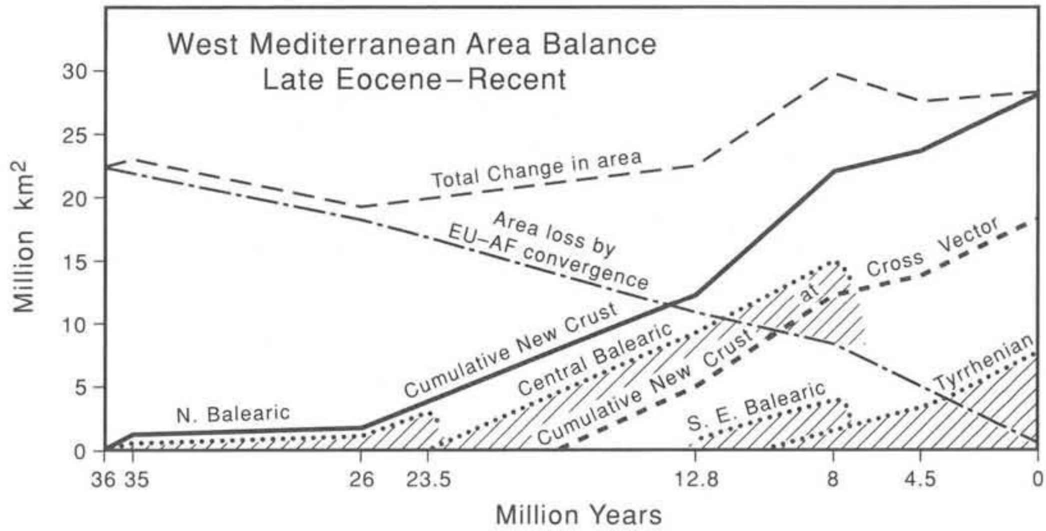

Figure 6-38. Graph of area in km² changing through time, depicting gains and loss of area during the Oligocene and Neogene evolution of the western Mediterranean.

will react with hinge advance, and it will localise compressional belts. In slow convergence, the crustal embayments will sink faster, perform extensional hinge retreat, and serve to localise the sites of backarc spreading.

Tapponnier (1977) has used an analogy from metal shaping by indenting or extrusion. This mechanism explains how material is pushed away from sites of indentation, and how it fills the voids opening at the sites of backarc spreading. As previously indicated by Pavoni (1961), the escaping material moves along slip lines, that is, faults of predictable orientation and strain. Tapponnier sees Mediterranean tectonics as a field of escape sites between the indenting Adriatic microplate and Asia minor. Material from the compressed Alps is escaping into the opening Pannonian basin accompanied by strike slip on the Insubric line. The tectonic boundary between Alps and Apennines may be a slip line. The southern edges of the west Mediterranean backarc basins require strike slip in their strain geometry. All tectonic mechanisms discussed serve to generate thickness variations in the Mediterranean lithosphere: subduction and convergent stacking, extension and viscous pillowing, indentation, and escape. Although thermal thickness variations are ephemeral, compositional thickness variations will survive the final freezing of the convergence. In the following chapter, these ideas are given quantitative expression.

6.7.4 EXPANDING BACKARC BASINS

The west Mediterranean series of backarc basins suggests that continental convergence displaces asthenosphere at depth, as depicted in Figure 4-19, and induces it to ascend in a stringer of mantle diapirs or viscous pillows. This can be derived from a west Mediterranean plate model since the early Oligocene (Cohen 1980, Livermore and Smith 1985).

Figure 6-38 is a graph showing cumulative changes in area through time. It shows the convergence as a loss in area of 0.3×10^6 km² in 26 Ma. It also shows the growth of new sea floor along spreading directions parallel to the convergence vector, and it shows the growth of backarc basins spreading normal to the convergence. The areas compared are confined to the western Mediterranean; the backarc basins do not include the Pannonian, Aegean, and

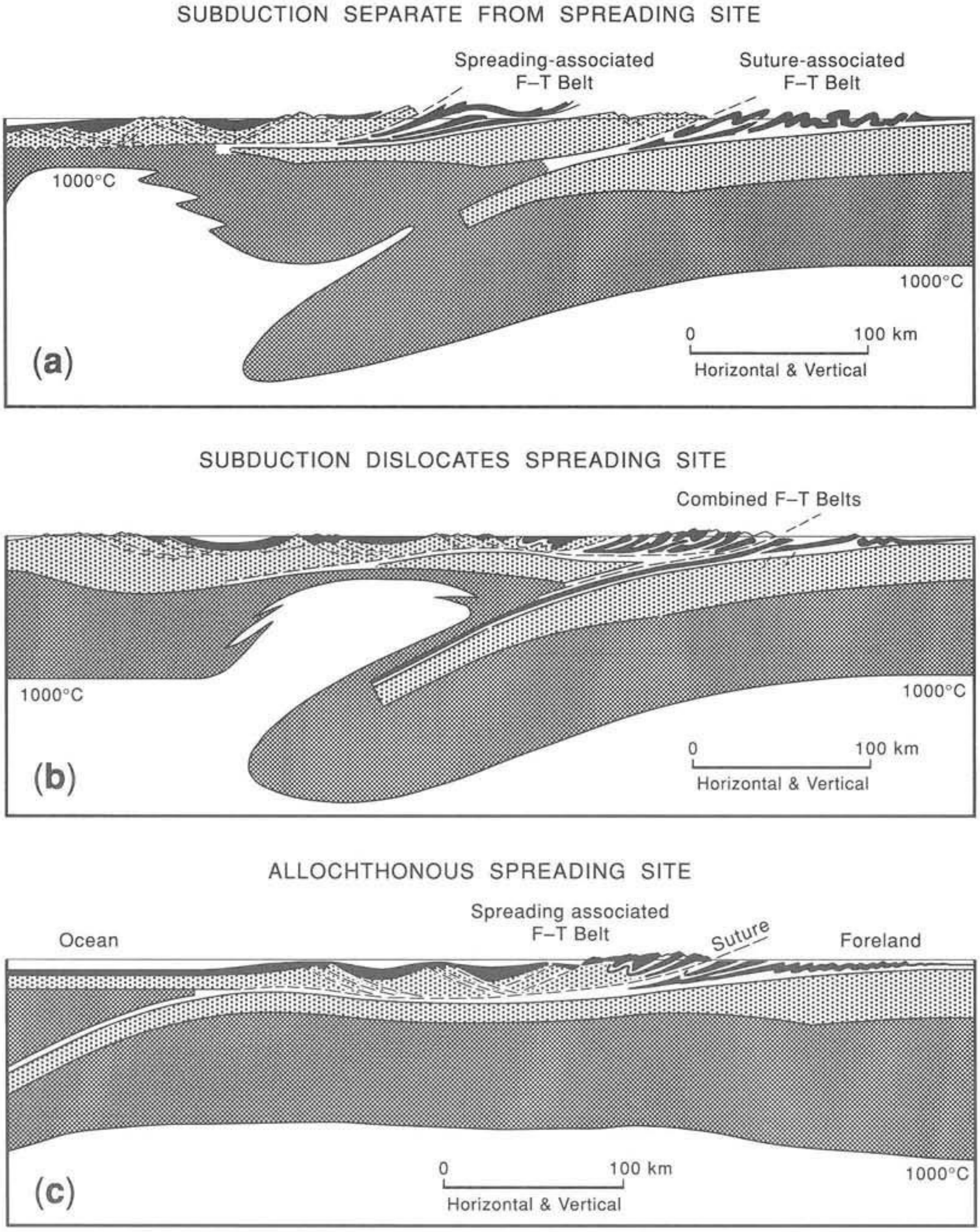

Figure 6-39. Sketch cross sections showing three varieties of orogenic belts common in the Mediterranean.

(a) Two fold-thrust belts geographically separate. The internal one is associated with an extensional site and has a shallowing basal detachment at the base of severely thinned crust, possibly overlying a lithospheric thermal high. The external one is the surface expression of a plate boundary. Its hangingwall block is a high and inclined crustal slab in outcrop or at shallow depth, referred to as median crustal slab. Its footwall block is an elastically deflected segment of crust with pre-Alpine thinning. This variety is best realised in the Tello-Rifian and Atlasian systems of Algeria.

(b) Two convergent crustal contacts with supracrustal fold-thrust belts. The external contact dislocates the internal detachment, piggybacking its fold-thrust belt. The site of subduction may not be evident in the thrust architecture, but it can be recognised by the geometry of the foredeep fill, by Moho offsets, and by Benioff zones. At the internal fold-thrust belt, extended crust is thrust

Alboran sites of extension. The graph suggests that in the western Mediterranean, extension exceeded convergence by about 50%. About half of the excess spreading is absorbed in west-to-east subduction at the Sardo–Corsican front and at the Calabro–Panormide front. One third of the remaining excess spreading is absorbed by Apennine thrusting and possible subduction within the Adriatic microplate. The remaining fifth is absorbed in the Dinaride foothills. Hence the net area change by post-collisional tectonics in the western Mediterranean is about zero. The four individual spreading events follow each other like successive explosions from the NE to the SW, building up to a climax during the Balearic event and declining through the Tyrrhenian event.

6.7.5 MEDITERRANEAN OROGENIC CYCLES

For the study of Mediterranean tectonics with cross sections, it is appropriate to separate fold-thrust belts at the edges of extensional sites from fold-thrust belts in the forelands of lithospheric subduction zones. This cannot be done everywhere. In the Apennines, both systems overlap but have been separated through elastic-load studies (Royden and Karner 1984). The Hellenide site is dominated by subduction of the retreating hinge, and the extension-related system is carried in piggyback fashion (Le Pichon 1983). In the Atlas of North Africa, the subduction system generates the Sahara Atlas and the High Atlas; it is geographically well separated from the Tellorifian fold-thrust belt along the edge of the Alboran and Balearic extension sites. In a series of sketches (Figure 6-39a,b,c) we show the possible geometric relationships between the two associated types of compressional belts. Of the three illustrated combinations, two are common and well documented. A third type (Figure 6-39c) is theoretically possible and was suggested a long time ago (Andrieux *et al.* 1971), but it is not yet acceptably documented. Transitional forms between the sketched types are also possible and in part documented.

Alpine–Mediterranean orogenic belts undergo a life cycle of subduction, collision, topographic buildup, extensional collapse, and backarc spreading. The sequence of stages is determined by the tectonics of the converging lithosphere and the flowing asthenosphere, but the time spent in any of the stages is not fixed. We have identified three dynamic settings in which the lithospheric tectonics shapes the orogenic cycle: collision, decay of convergence, and hinge retreat.

over the median crustal slab. Incipient backarc spreading is suggested by crustal extension and by a lithospheric thermal high. This tectonic configuration is common, such as in the northern Apennines, Hellenides, and possibly in the Dinarides, Taurides, and northern Carpathians.

(c) Shallow internal thrust system which has overridden and covered the external subduction system over a distance of several hundred km. The detachment is most likely located in the viscous lower crust of the hangingwall slab. In the footwall, the crustal type is probably thinned-continental or oceanic; it may display a Moho at a normal continental depth and may mask the thrust overlap. The existence of this variety is suggested by paleogeographic reconstructions and by mismatches between subsidence and apparent crustal thickness; it implies complex map-view geometry and highly efficient mechanisms of indentation and escape. An example of this type is the Alboran Sea. In this diagram, the Alboran Sea would be seen from the north, with the open Atlantic on the right edge. Another possible example is the Pannonian basin with the Balkanide Benioff zone.

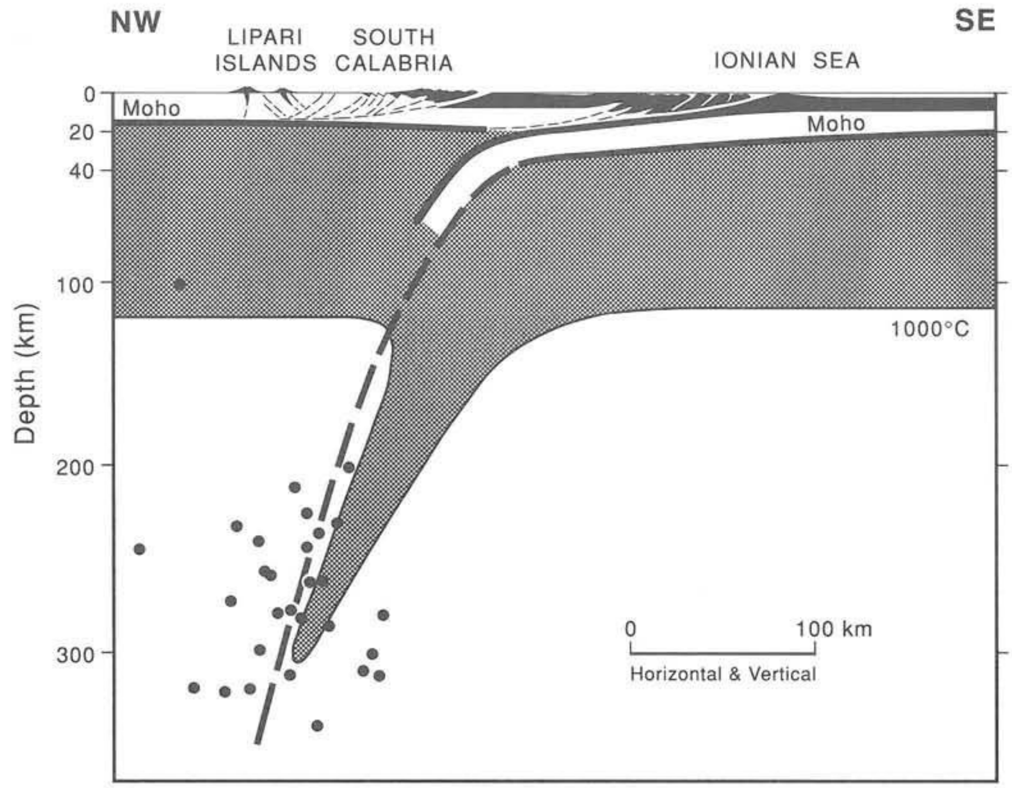

Figure 6-40. Cross section of Calabrian arc and Tyrrhenian Benioff zone, redesigned after Giese et al. (1982), with seismic foci from Ogniben (1969). Shaded area: lithosphere cooler than 1000°C.

Collision or indentation are needed to build the topographic elevation which initiates extension. Hinge advance in a subduction slab can also set an orogen into a compressional mode, but there are no Mediterranean examples of this situation.

Decay of convergence and/or hinge retreat are needed to change the setting from plateau buildup to crustal thinning and eventual sea-floor spreading. In the Neogene to Recent Mediterranean tectonic setting, both of these factors are present. Figure 6-40 shows a cross section of the Calabrian arc and Tyrrhenian Benioff zone. The sharp bend in the Ionian-Sea slab is constrained by the elastic-flexure parameter of the crust in the Ionian Sea and by the location of the Benioff zone. This cross section across the Calabrian arc and Tyrrhenian Sea serves to document the Mediterranean type of piggyback thrust belt with hinge retreat.In several Mediterranean models (Alvarez 1976, Biju-Duval *et al.* 1977, Livermore and Smith 1985), the age of the crust being subducted beneath the Apennines and the Aegean arc increases toward the trapped oceanic crust of the Ionic sea. Figure 6-41 illustrates the piggyback structure of the entire Apennines fold-thrust belt. Its location precludes that it was generated by the Tyrrhenian subduction. We interpret it as generated by Tyrrhenian backarc spreading.

In the Mediterranean, the predominance of west-dipping subduction slabs, of east-vergent fold-thrust belts, and of eastward migration of spreading sites is remarkable. Possible explanations can be based on global or on more local, specifically Mediterranean arguments but, based on available data, neither global nor local explanations are conclusive. Doglioni and collaborators suggest a westward directed toroidal shear motion between the lithosphere and deeper mantle realms (Doglioni 1990, Ricard *et al.* 1991). Some of their supportive geological observations had been used earlier to support an easterly mantle flow (Nelson and

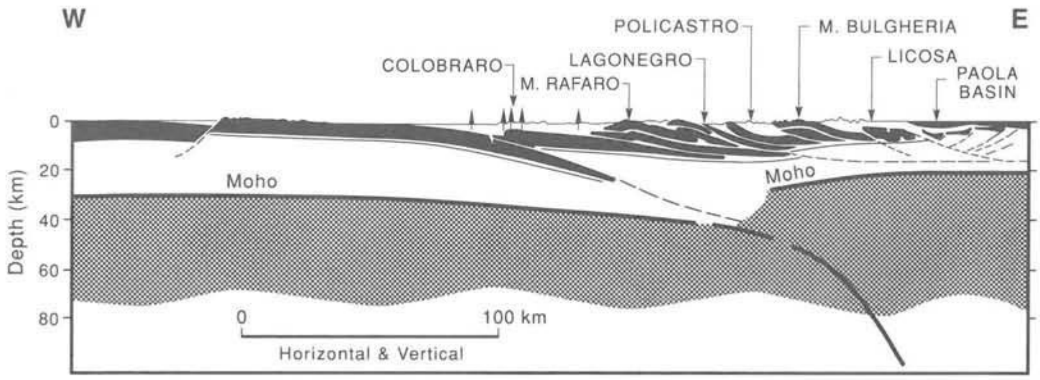

Figure 6-41. Structure cross section of southern Apennines just north of Calabria, simplified after Roeder (1984), and projected into lithospheric and Benioff geometry of Calabrian–Tyrrhenian arc as in Figure 6-40.

Temple 1972). The global hotspot configuration (reviewed by Duncan and Richards 1991) suggests a pattern of mantle convection unaffected by toroidal and global shear. It also suggests a slow toroidal displacement between the core-inducted magnetic field and the hotspot reference system; this torus is oriented roughly normal to the Earth's spin axis. A more local interpretation of Mediterranean mantle tectonics is imposed by a convecting mantle without toroidal shear. Local tectonic elements can still be explained by shallow mantle convection and lithospheric thickening (Channell and Mareschal 1989). Regional elements include the eastward increasing age of Tethyan oceanic crust. In maintaining the eastward hinge retreat, this shallow feature has predetermined much of the Mediterranean tectonic pattern. The density gradient between shallow mantle and deep continental crust (Bott and Kuznir 1979, Turcotte 1983, Le Pichon 1983) can also generate a tectonic polarity, with Tethyan mantle predominant to the east, and deep continental crust to the northwest and the south. This point is addressed again in Chapter 7.2.4.

6.7.6 VARISCAN ELEMENTS IN MEDITERRANEAN TECTONICS

Mediterranean tectonics clearly shows that orogenic loops and sites of extension and compression originate together and depend on each other. It also shows that some tectonic features outlast their orogenic cycle and help to predetermine the plate pattern of the subsequent cycle. It is therefore worth briefly re-examining Late Paleozoic orogens in the western Mediterranean area and their remnants of extensional basin fill, which possibly signal post-collisional plateau collapse. In the modern Mediterranean configuration, there are outward-vergent loops around sites of extension and inward-vergent loops around subducted or overrridden terranes. Based on our present understanding, extension at outward-vergent loops is synkinematic with orogeny. Extension may soon reach the stage of sea-floor spreading and may disperse the orogenic fringe toward distant shores. At inward-vergent loops, extension can only start as plateau collapse after terminal collision of the orogenic fringes, helped by the subducted foreland terrane being converted into a soft, buoyant orogenic root. If plateau collapse extends to the base of the crust and is not blocked by intraplate compression, the root can serve as the focus for the next generation of extension. Figure 6-42 shows the late-orogenic pattern of Atlantic Variscan belts as prepared by Ziegler (1988) and projected onto the pre-Atlantic plate configuration suggested by Livermore and

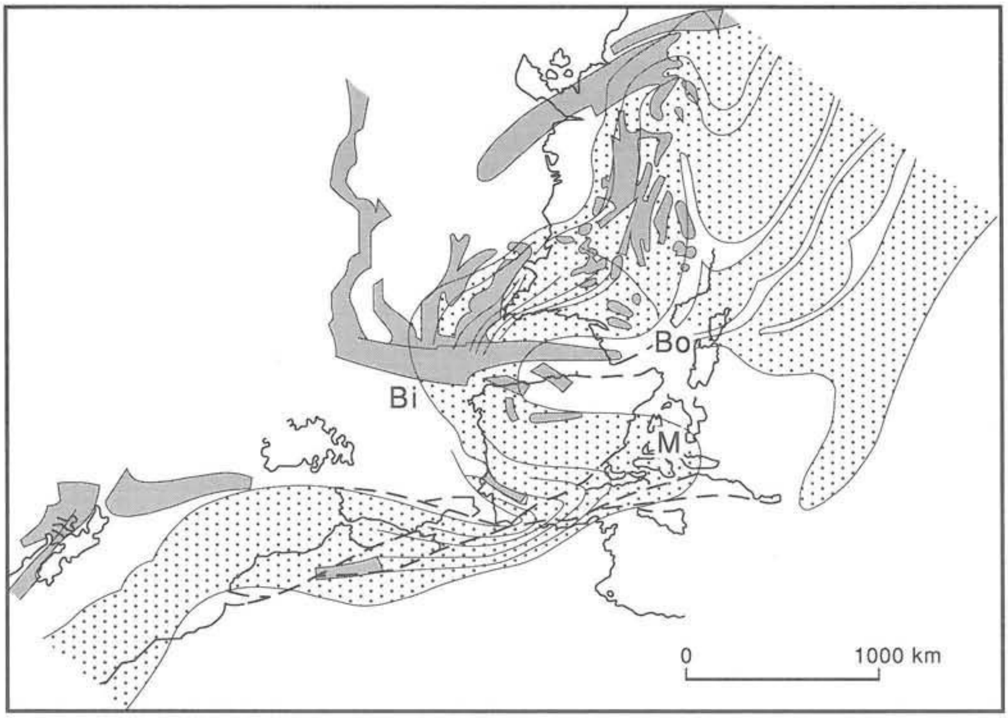

Figure 6-42. Sketch map showing the Atlantic Variscan architecture (Ziegler 1988) as projected into the plate restoration at 180 Ma by Livermore and Smith (1985). Stipple: main orogenic belts, M: Meseta loop, Bi: Biscay–Asturia loop, Bo: Bohemian loop. Dark shading: late- and post-orogenic rift sediments, possibly indicating plateau collapse.

Smith (1985). Figure 6-42 also shows Stephanian and Permian clastic units (Ziegler 1988) which are candidates for plateau collapse sediments. The Appalachian–Variscan system is shown to be bivergent. Its loops have a dual polarity and are therefore all capable of evolving towards plateau collapse. On the African–European side, the east-convex Meseta loop has no documented Late Paleozoic debris but forms the site of the Atlantic–Tethys group of transform faults of Jurassic age. The west-convex Biscay–Asturia loop is inward-vergent viewed from the east and contains considerable Stephanian and Rotliegend debris. It also serves as a transform system terminating the Faroe–Rockall rift (Ziegler 1988). The Bohemian loop is outward-vergent and its plateau collapse may have been part of the extensional field covering the Variscan heartland. As well as the well-mapped Rotliegend troughs, late Paleozoic metamorphic core complexes could still be discovered there. However, this part of the Variscan system did maintain its cratonic coherence after the Variscan orogeny. Additional Variscan areas with possible plateau collapse may have predetermined the pattern of Alpine–Penninic basins.