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Looping the loop; geotectonics of the Alpine-Mediterranean region

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Παντα ρει . . . .
(Heraclites)
It is all a matter of rheology.

Abstract

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The Alpine-Mediterranean system consists of an African and a European orogenic belt which – with a few discontinuities – extend in great loops and bends between the Atlantic Ocean and the Middle East. The evolution and deformation of these belts or so-called oroclines is closely coupled to the relative movements of the African and European plates during the Mesozoic and Cenozoic. Oroclinal bending presupposes the evolution of relatively thick and rigid ribbon continents marginal to the African and European cratons. These ribbons are considered to have formed within the passive continental margins on either side of the Mediterranean while this seaway opened in a transtensional setting, starting in the Triassic.

While the ribbons contorted the sedimentary basins in between were squeezed up and out in diapiric fashion to later be thrust (obducted) over their peripheries and to be emplaced as sedimentary and ophiolitic nappes. Orogenesis finished with the isostatic uplift of the oroclines and the subsidence of the deflated and denudated diapirs.

Introduction

In 1958 Carey presented a global tectonic scheme, which he developed as he stated based on the following assumptions: 'That geotectonics is simpler than we thought, that the gross structures stare naked from the world physical chart, and that things really are what they look like'.

Thus observing what he saw in the Mediterranean region, he reconstructed a pre-Alpine configuration by simply straightening the loops and bends (oroclines) in the entire orogenic belt. Unwinding the loops produced two deformed zones marginal to the European and African cratons that faced each other across the Tethys (Fig. 1).

Later, in 1976 Carey improved his model considerably, especially for the western Mediterranean. However, he failed to resolve the apparent incompatibility of clockwise rotations of Dinarides and Hellenides and the sinistral shear between Europe and Africa, perhaps because he did not consistently follow through the implications of his own basic assumptions. Also, the exact relation in time and space between oroclinal bending and crustal shortening remains unresolved in Carey's scheme.

Nevertheless, his ideas on oroclinical bending are significant in that they emphasize the deformation of an orogen in plan as opposed to the deformation in cross section. The latter is almost exclusively the domain or dimension in which structural geologists are interested, although the recognition of shear appears to be gaining respectability. As will be shown oroclinal bending, i.e. deformation around vertical axes, is an important element in the geotectonic evolution of the Alpine-Mediterranean region.

To make oroclinal bending acceptable, however, we must explain first how marginal, elongate, narrow, more or less straight belts are formed, which subsequently become partly or wholly detached from their neighbouring craton, to then be bent into the required shape. Carey does not seem to have been unduly worried by that problem and appears to have taken the obstacle in his stride.

There is of course one environment – but that certainly was not obvious 25 years ago – where the chances for the formation of ribbon continents are excellent and that is the passive continental margin during a particular phase of its evolution.

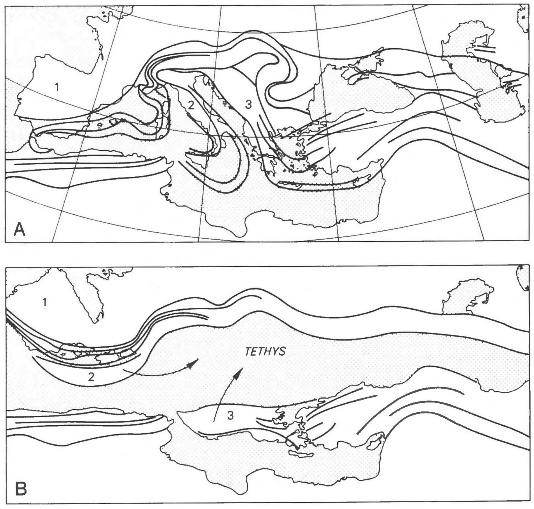

Fig. 1. (simplified after Carey 1958)
A. The present tectonic framework of the Mediterranean region. 1. Iberian block; 2. Sicilian-Apennine block; 3. Dinaride-Hellenide block.
B. The Mediterranean oroclines straightened without lateral shift. Arrows indicate the ensuing rotations of individual oroclincal segments, which are incompatible with unidirectional transcurrent movement between Europe and Africa.

Passive continental margins and the formation of Ribbon Continents

Ten years ago I proposed a model for the evolution of the Labrador Sea (Van der Linden 1975a) that is valid for the evolution of other passive or so-called Atlantic type continental margins. The model was kept simple by assuming that, prior to break-up, the original continental lithosphere is homogeneous in composition and structure in a horizontal sense. Thus, as shown in Fig. 2, a symmetrical configuration could develop that consists of a central oceanic basin that is flanked on either side by a continental fragment – i.e. a ribbon continent – and a marginal basin, the latter overlying attenuated continental crust. As stated, as a model for the evolution of passive continental margins it is simple and a perhaps more realistic, not so symmetrical and more complex picture is given by Ziegler (1978) in his reconstruction of the Mesozoic North Atlantic region. There also, in the fragmentation of Laurasia, several longer and shorter, irregularly shaped continental blocks can be seen that became detached while the break-up proceeded. However simple, the Labrador configuration (Fig. 2) is remarkably similar to that of the Mediterranean system prior to oroclinical bending à la Carey (Fig. 1). Furthermore, it is at least interesting to also note the almost one to one similarity between the palinspastic section through the Liguria-Piedmont Ocean i.e. through the Tethys in the late Jurassic (Kimmeridgian) by Laubscher & Bernoulli (1977) and the Labrador Sea model.

In this paper the Labrador Sea model is used as a first approximation to explain the subsequent overall geotectonic evolution of the Alpine-Mediterranean system. Once the general scheme is established it should be relatively easy to then modify it, starting with a more complex configuration that will lead to a more realistic evolutionary model.

Long and narrow ridges are not uncommon in the ocean. Apart from the mid ocean ridges, some fracture zones, the island and volcanic arcs in the Pacific, which are alle *oceanic* features, there is another group of elongated elevations which are recognized as marginal *continental* fragments or for which such an origin is a likely possibility. In this

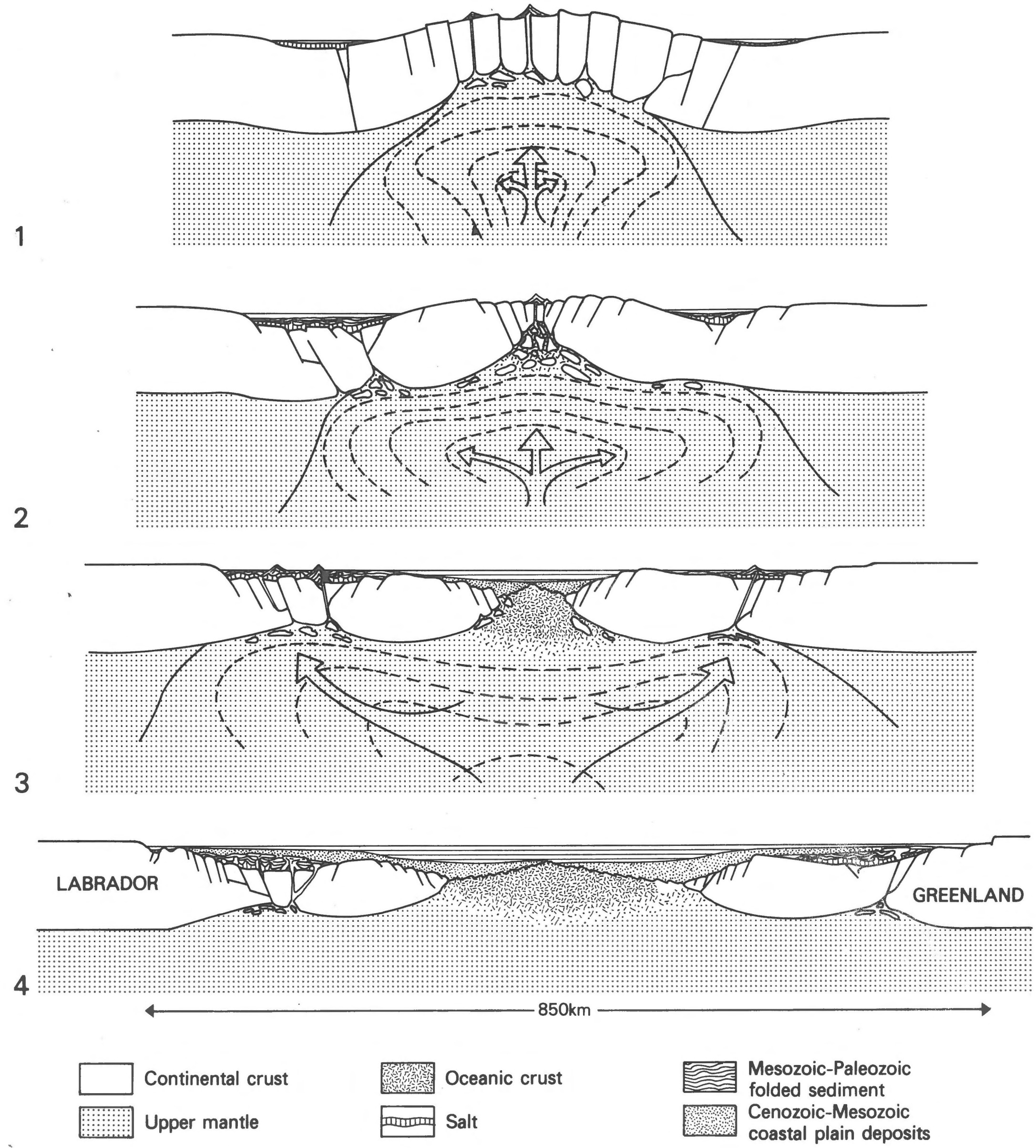

Fig. 2. Crustal model of the formation of the Labrador Sea in four stages. Vert. ex. approx. 3 ×. Isotherms (dashed) and flow directions (arrows) marked in upper mantle. Phase 1: Paleozoic? Uplift of crust over thermally expanding and ascending 'asthenolith'. Formation of marginal foredeeps. Accumulation of evaporites. Phase 2: Early and Middle Mesozoic? Sediment loading in marginal basins causes further crustal downwarp. Magmatic stoping at crust-mantle interface. Phase 3: Early Cretaceous. Inversion of relief through collapse of 'astenolith' blister. Beginning of sea-floor spreading. Phase 4 Eocene to Recent. Sea-floor spreading stopped in the Eocene. Accumulation of coastal plain sediments since the Albian. Margin subsidence continues. (After Van der Linden 1975a).

category we find the Lomonosov Ridge in the Arctic, the Lord Howe Rise and Norfolk Ridge in the SW Pacific, the Kerguelen Plateau in the Indian Ocean and of course Baja California. In the past I have argued for a similar origin for the core of the basalt-covered Madeira-Tore Rise and Walvis Ridge in the western Atlantic (Van der Linden 1979, 1980) and good other candidates for such an origin are the Chacos-Lacadive and Ninetyeast Ridges in the Indian Ocean (see also Udintsev & Koreneva 1982).

All these continental fragments are or have been situated marginal to large continental plates from which they became detached during rift and subsequent drift phases in the desintegration of Gondwana and Laurasia. The marginal basins in between continent and ribbon-shaped continental fragment are underlain by attenuated continental lithosphere and in cases when attenuation proceeded sufficiently far by oceanic crust (i.e. solidified asthenosphere). In the latter cases, as for instance in the Tasman Sea, a central spreading ridge may have developed and the spreading history of such basins can be read from sea-floor spreading magnetic anomalies.

Let us now return to the Mediterranean.

Structure of the Alpine-Mediterranean area

Fig. 3 shows the present structural framework of the Mediterranean region. Apart from displaying the tectonic grain and the directions of nappe transport a distinction is made in that figure between a European and an African fold belt.

In the European belt we find from west to east the Betides or Betic Cordillera, the Balearic Islands, the Sardinia-Corsica block, the Western Alps, the Northern (Calcareous) Alps, the Carpathians, the Transsylvanian Alps, the Balkan Mountains and their continuation on either side of the Black Sea into the Caucasus and via Pontides into the Lesser or Minor Caucasus. The African belt extends from the Rif in Morocco through the Maghrebides (Tellian Atlas), the Sicilian-Calabrian Arc into the Apennines. From there via the Southern Alps it continues through the Dinarides, Hellenides, the Hellenic Island Arc (Crete) and Taurides (including Cyprus) into the Zagros Mountains.

The continuity per belt as shown in Fig. 3 is generally agreed upon on the basis of lithological, stratigraphic and structural affinities (e.g. Alvarez 1976; Biju-Duval & Montadert 1977; Smith & Woodcock 1982) and is virtually identical to Staub's (1924) now classical interpretation. Also shown in fig. 3 are other Alpine chains such as the Pyrenees, the Iberian Cordillera on the European side and the High Atlas, the Saharian Atlas and the Syrian foldbelt or Palmyre chain on the African side. The connections of these subsidiary belts to the main orogens are not precisely defined. They are here considered 'peripheral' orogenic belts, the result of compression of relatively small peripheral basins.

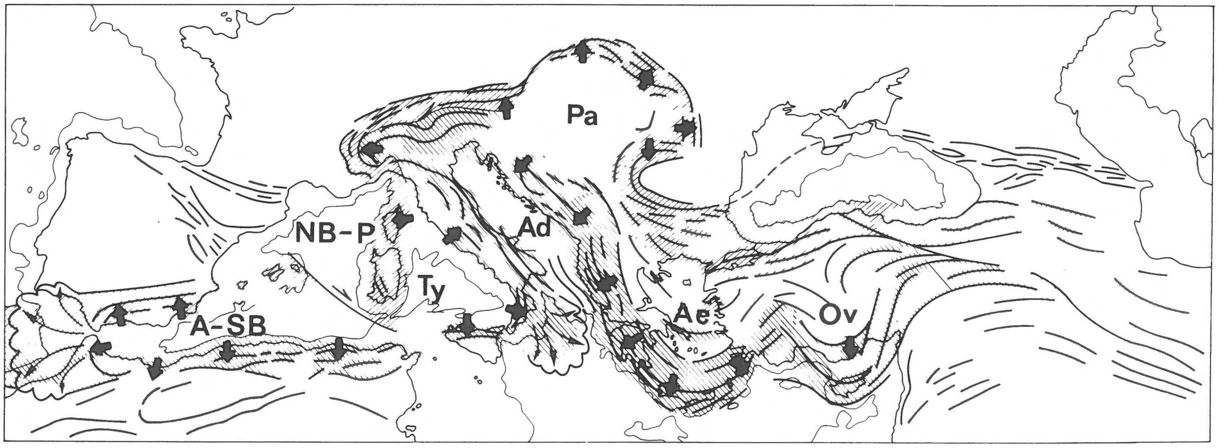

Fig. 3. Structural map of the Mediterranean region showing the gross tectonic grain and the direction of the nappe thrusting (heavy short arrows). For the western Alboran Sea and the Calabrian Arc thin arrows indicate the direction of massive slumps or gravity slides. SW-NE hachuring marks the 'European', NW-SE hachuring the 'African' orogenic belt (after UNESCO 1980, and Horvath & Berckhemer 1982). A-SB = Alboran-South Balearic Basin; NB-P = North Balearic-Provençal Basin; Ty = Tyrrhenian Basin; Ad = Adriatic Basin; Pa = Pannonian Basin; Ae = Aegean Basin; Ov = Ova Basin.

This brings us to a difference between the central part of the Mediterranean region and the western and eastern regions. The central part has a relatively open character in which two fairly wide oroclinal belts are cleanly separated by oceanic or semi-oceanic basins. The western and eastern regions are much more 'tight' and there narrow but more numerous foldbelts are separated by continental plateaus. This difference reflects fundamental aspects of the formation history of the entire region that will be considered later.

To understand the evolution of the general structure one has to take into account the shape of the oroclines, the direction of nappe transport, the subduction of African plate fragments under the Calabrian and Hellenic Arcs and also the perhaps somewhat conjectural strike-slip dislocation between Balearic Islands and Sardinia. The existence of such a strike-slip movement is substantiated by a NW-SE trending discontinuity in the magnetic anomaly pattern along that lineament (Galdeano & Rossignol 1977) and is recognised by Durand-Delga & Fontboté (1980) as 'accident Paul Fallot.' An important clue to help unravel the tectonic history furthermore is found in the nature of the intramontane regions (labelled in Fig. 3) in between the oroclinal belts, their differences from basin to basin in structure, depth and age.

The Alboran – S. Balearic Basin is underlain by strongly attenuated, 10-14 km thick continental crust over a low-velocity Moho and displays high heat flow (90 mW/m^2). It is a young basin, probably of Miocene (Burdigalian) age, i.e. of about the same age as the outward-facing nappes that surround it on the Spanish and Moroccan side. In the central basin there are a number of volcanic, predominantly andesitic seamounts, others are apparent from pronounced magnetic anomalies. Large nappes and gravity slides that appear to have their root in the Alboran Sea extend westward into the Atlantic. An interpreted NE-SW cross-section (Fig. 4, after Biju Duval & Montadert, 1977) shows the Alboran Sea in relation with the surrounding orogenic belts.

The N. Balearic – Provençal Basin is underlain by attenuated continental crust which in comparison is somewhat thicker than the crust under the Alboran Basin. Its age is presumably Oligocene-Miocene. As in the Alboran Basin seamounts abound in the basin, especially in the western part.

The Tyrrhenian Basin is underlain by a number of narrow, deep, sediment-filled grabens marginal to Corsica and Sardinia. The basin widens and deepens towards the southeast. There, under a thin cover of Cenozoic sediments we find a crustal thickness of 8 km which may point to oceanic conditions. Seismic velocities in the crust remain low, however, which would make it 'very' attenuated continental crust. Magmatic intrusions and seamounts are common and appear to merge with the volcanics of Sicily and western Italy. As the age decreases from Miocene to Recent the petrology of this magmatic province changes from tholeiitic in the deep Tyrrhenian Basin via calc-alkaline (Eolian Islands) to andesitic-phonolitic for the Pontine Islands and the Roman-Tuscany province of Italy. Heat flow in the Tyrrhenian Basin is extremely high (119 mW/m^2 in the central part). Extension and subsidence of the basin began in Tortonian, perhaps already Burdigalian time. Marine conditions prevailed in Upper Miocene (pre-Messinian)

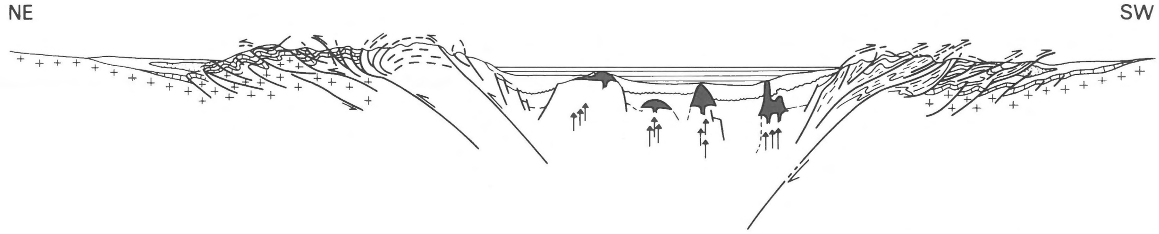

Fig. 4. Geological section through the Alboran Sea and adjacent Spanish and African margins (after Biju-Duval & Montadert, 1977). Length of section about 600 km.

times, distension took place mainly during the Upper Pliocene, and the basin foundered rapidly in the last couple of million years.

In the north, between Corsica and Tuscany Morelli et al. (1977) showed a deep crustal section in which the continental Corsican crust is seen to dip under Tuscany. To explain this phenomenon Giese et al. (1978) invoke a rather complex scheme of 'antithetical' continent-continent collision whereby the upper Tuscany crust is peeled of its basement to be thrust over the Corsican plate, while the basement is thrust under (delamination or flake tectonics).

Earthquake epicentres assemble into a vague, discontinuous Benioff zone. The earthquakes are commonly interpreted to mark a subduction zone that dips NNW from the Calabrian Arc under the Tyrrhenian Sea. Hypocentres cluster in the upper 50 km and in a zone between about 220 and 280 km deep and there is a marked seismic gap between 110 and 220 km (Ritsema 1979; see also Udias 1982).

The Adriatic Basin is underlain by continental crust that is covered by an 8-10 km, perhaps even thicker pile of posttectonic undisturbed sediments at the base of which Triassic salt is found. Refraction profiles by Morelli et al. (1969) show a sediment-filled trough marginal to Italy, that is formed by the thrust of the Apennines over westward dipping Adriatic basement. It is possible that this trough forms the northern most shallow expression of the Calabrian subduction zone.

The Pannonian Basin together with the *Transsylvanian Basin* are considered to have been part since the Late Jurassic – Early Cretaceous of the Alpine Carpathian Flysch basin and to be underlain by attenuated continental crust. Closing of that basin began during the late Cretaceous. At the end of the Oligocence the Flysch Basin was very much reduced in area through compressive folding and thrusting with major parts of the Pannonian region emergent. Linear subsidence occurred since the late Miocene at the rate of 150-250 m/Ma. The Pannonian Basin has a 25-30 m thick crust over an apparently normal Moho. At a depth of 50-60 km. however, mantle velocities of 8.0-8.2 km/s drop to values around 7.7-7.8 km/s. The surrounding Carpathians and Dinarides in contrast are underlain by 40-65 km thick crust with indications for crustal doubling under the Outer Carpathians. Heat flow is very high (80-130 mW/m^2) under the central Pannonian Basin, but low (50 mW/m^2) in the periphery (Vienna and Transsylvanian basins). Mio-Pliocene calc-alkaline magmatism, i.e. contemporaneous with subsidence, cannot be explained according Lexa & Konecny (1974) as the result of island arc type volcanism. Instead they argue strongly for a model in which volcanism is related to diapiric uprising of the asthenosphere.

The Aegean Basin is underlain by sialic crust that varies in thickness from a minimum of 20 km under the Cretan Sea to 32 km in the Cyclades. Upper mantle velocities are low again both under the Aegean Sea and the Hellenides. It seems that the crust was appreciably thicker in Lower to Middle Miocene times (about 50 km). According to Schuiling (1972) subsequent uplift and denudation would have brought the originally 10-20 km deep metamorphic zone to the surface. A seismic reflector in the S. Aegean Sea, which is interpreted as a mid-Tortonian erosion level, is in support of the existence of an upper Miocene subaerially exposed Aegean landmass, which since subsided below sea level. A tensional regime appears to have prevailed in the Aegean region since the Upper Middle Miocene and to be connected with northward subduction of an oceanic plate beneath the Hellenic Arc (backarc spreading). The envelope of earthquakes defines – in some places better than in others – a Benioff zone down to a depth of 180 km. First motion studies mark a northeasterly convergence direction with predominantly strike-slip movements along the Pliny and Strabo trenches (S and SE of Crete). Heat flow in the Aegean region is high and upper Pliocene – Pleistocene calk-alkaline basalts and rhyolites abound in the southern Aegean.

In between Pontides, immediately south of the Black Sea, and Taurides the extensional *Ova province* of central Anatolia occupies a position that may be considered the structural equivalent –

although not as advanced – of the Aegean Basin, i.e. a basin formed as the result of back-arc spreading, in this case behind the Cyprus Arc (Sengör & Canitez 1982).

From deep seismic soundings in the eastern Mediterranean Makris (1981) concluded that the deep basins in the Levantine and Ionian Sea are underlain by oceanic crust (Vp=6.7 km/s) whereby the uplifted Mediterranean Ridge is considered continental (Vp=6.0 km/s).

According to Smith & Woodcock (1982) the eastern Mediterranean is still very problematic and is considered to be either a remnant Mesozoic ocean basin south of the Paleotethys or a subsided portion of thin continental crust. Subsidence according to Biju-Duval & Montadert (1977) would have begun in Triassic or Liassic times and the ensuing seaway would have separated the Aegean-Anatolian region from Africa since the Mesozoic.

Ryan et al. (1982) interpreted the Mediterranean Ridge to be an accretionary sedimentary wedge in an island arc setting of which the evelation may have been enhanced by expansion of Aptian overpressured mud or shales in its core.

Finally, the so-called *peri-Adriatic zone,* where the suture between the European and African oroclines is particularly tight, i.e. where they are not separated by intramontane basins, is marked by narrow ophiolite belts. In the Western Alps Jurassic ophiolites closely follow the Ivrea-Insubric-Giudicaria-Pusteria lineament and in between Dinarides and the Carpathian arc upper Cretaceous ophiolites mark the Vardar zone. South of and parallel to the latter there is a second ophiolite belt through Pindos and Othris. The Vardar zone continues eastward, although in a more chaotic, less linear fashion, through central Anatolia into Iran. Similarly, the southern belt extends through the Antalya complex of SW Turkey to Cyprus and Hatay.

The ophiolites are considered by many authors to be allochtonous components from a former ocean that were emplaced in an island arc setting, to be tectonized and metamorphosed in Jurassic to upper Cretaceous time. Coleman (1984), however, argued for obduction of young interarc or small ocean basin slabs, still hot and thus light while detached, a mechanism that would agree with the Lexa & Konecny (1974) Pannonian Basin model. Abyssal, old and cold, oceanic crust in contrast would have been subducted, not accreted and obducted, and would have left no trace at the surface. Vlaar & Cloetingh (1984) also recognised ophiolite nappes as segments of young oceanic crust emplaced on top of Alpine passive margin sediments. The required gravitational instability they visualized through uplift and doming of the crust over buoyant rising mantle material which in turn they considered the result of underthrusting of oceanic ridge segments under continental platform lithosphere when basin closure took place (see also Dewey, 1976; Cloetingh et al. 1984).

For a more elaborate overview of the Alpine Mediterranean structure especially the contributions by Biju-Duval & Montadert (1977), Horvath & Berckhemer (1982) and Smith & Woodcock (1982) should be consulted, which are essentially but selectively summarized in this section.

Plate kinematics

Global tectonic theory ordains that formation of mountain chains – and that includes oroclinal bending – is the result of plate interaction; for the Mediterranean the interaction of the African and European plates. The relative motion between Africa and Europe can be determined largely from the magnetically established sea-floor spreading isochrons in the Atlantic Ocean. From about 200 Ma to 165 Ma ago, i.e. from upper Triassic to middle Jurassic time, South America together with Africa started to separate from the combined North American-Eurasian landmass. From the paleogeography of Gondwana it follows that during that period dextral transcurrent motion must have taken place in the region of the future central Atlantic Ocean accompanied by extension in the western Tethys (i.e. the Mediterranean region; see also Van der Voo & French 1974). Since sea-floor spreading and continental drift are preceded by an attenuation or rifting phase distension between Europe and Africa very likely began in the Paleozoic (perhaps in the wake of the closure of the Iapetus Ocean). There is indeed ample evidence for tensional movement in early to middle Triasic time, rifting and associated volcanism in the western Mediterranean (cf. Bernoulli & Lemoine 1980, and other papers in Aubouin et al. 1980).

The early opening of the western Tethys in a transtensional fashion would be the time when ribbon continents (à la Labrador Sea) began to form along the European and African margins.

Sea-floor spreading and relatively rapid opening of the central Atlantic Ocean started 165 Ma ago while the North Atlantic remained closed. This resulted in predominantly sinistral transcurrent motion in the Tethys region and presumably also along the axis of the future Labrador Sea, Biscay Bay and Pyrenees. Shortly thereafter, still in Jurassic time, the Labrador Sea and Biscay Bay opened, coupled via a triple junction NW off Spain to spreading in the central Atlantic (Van der Linden 1975b). This caused the anti-clockwise rotation of the Iberian peninsula relative to stable Europe.

In late Cretaceous time the motion slowed down by the first major collision, that of Arabia and Anatolia with Eurasia. The opening of the northern Atlantic, beginning also in late Cretaceous time, caused Europe to later (68-76 Ma ago) catch up on Africa which gave rise to dextral movements in the Mediterranean. Finally, about 44 Ma ago, from the middle Eocene onward, Africa and Europe collided, pivoting around a rotational pole near Morocco, which meant largest compression in the eastern Mediterranean. In time the rotational pole moved west into the Atlantic in the direction of the Azores and with this scissor-like closing of the western Tethys compression increased in the entire region.

In summary, Africa appears to have moved relative to Europe in such a way that the stress regime in the Mediterranean gradually changed form transtensional during the Mesozoic, via transpressional to virtually pure compressional in the last 44 Ma, i.e. since anomaly 24 time. Dates and kinematics of this section are largely from Biju-Duval et al. (1977) and the relative movement of Africa is schematically shown in Fig. 5.

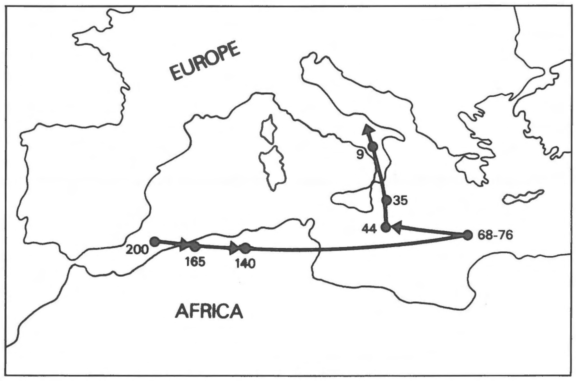

Fig. 5. Movement of Africa relative to Europe since the early Mesozoic derived from magnetic anomalies in the Atlantic Ocean. Numbers represent ages in millions of years. (after Biju-Duval et al. 1977 and Channell et al. 1979).

Looping the loop

'Un bon dessin vaut mieux qu'un long discours'
Napoleon Bonaparte

If we take Carey's (1958) assumptions (see Introduction) at face value then the formation of the Alpine foldbelts, including oroclinal bending, should be readily apparent from their shape (things are really what they look like). On that basis and as a first approximation Fig. 6 is constructed which presents a reasonable and realistic scheme of the events that characterize the kinematics of the core of the Alpine-Mediterranean system from late Cretaceous (A) to the present (D), the latter as depicted a simplified version of Fig. 3. In Fig. 6D a distinction is made between the open-shaded Aegean and Ova regions, now being formed in a back-arc spreading mode, and the other densely--shaded Mediterranean basins. No distinction has been made between subaerial (Pannonian Basin) and submarine basins; they are all areas underlain by attenuated subsided crust.

Going back in time and unwinding the oroclines (as per Carey recipe) produces intermediate stages C and B and starting-stage A. That stage, about 165 Ma ago, thus shows the configuration of two

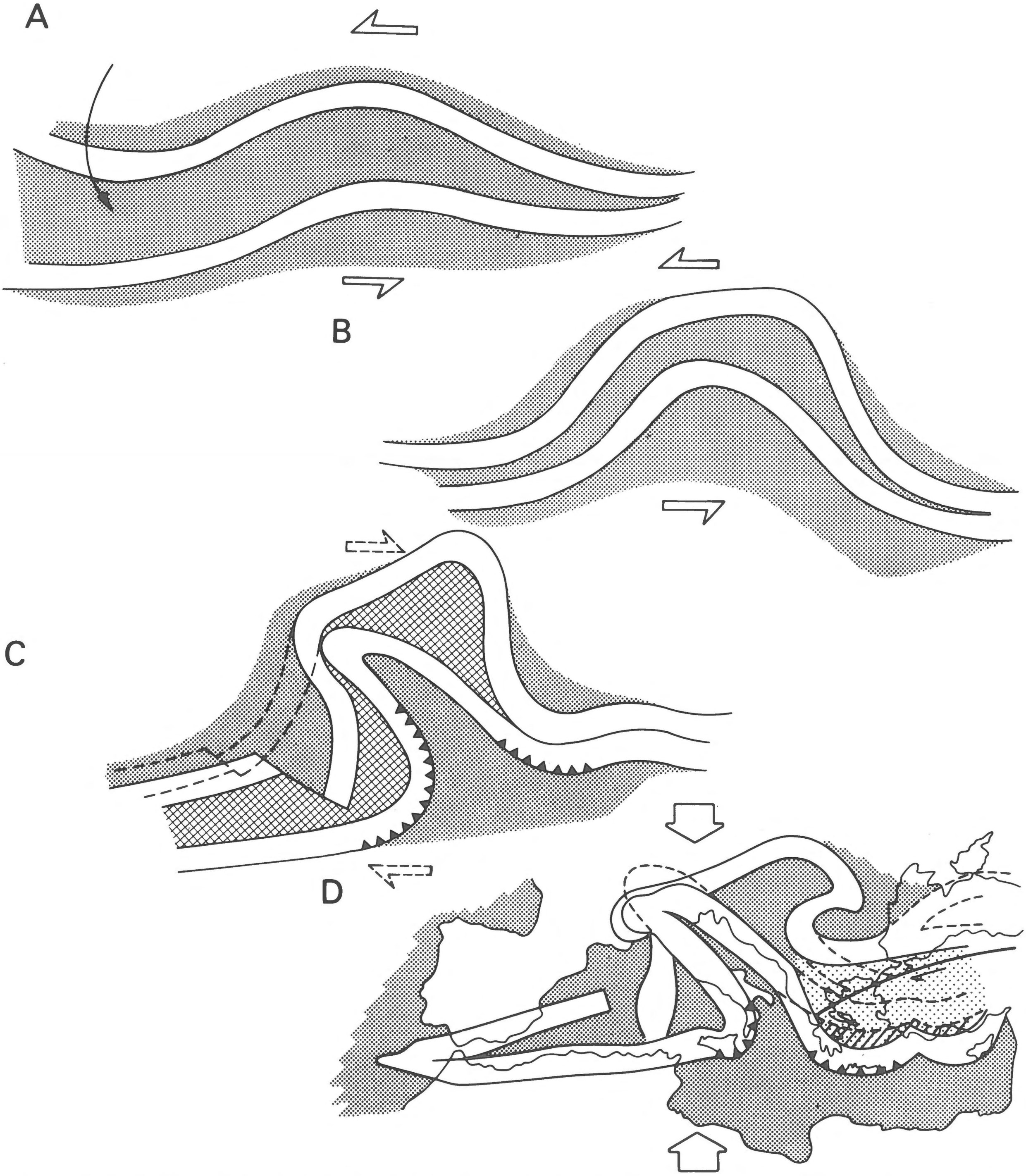

Fig. 6. Looping the loop. Scheme of oroclinal bending of ribbon continents (blank) and opening and subsequent closing of intermediate type basins (stippled) in between the European and African cratons (blank) in four stages. Stages A and B from 200? to 76 Ma; Stage C from 76 to 44 Ma; stage D the present configuration (i.e. a simplified version of Fig. 3). Crosshatching in stage C marks the areas (Alboran Sea, Tyrrhenian Sea and Pannonian Basins) that were emergent, shedding their nappes and sediments towards the periphery. Open stippling in stage D marks the Aegean and Ova regions which formed in a back-arc spreading mode behind the Hellenic-Cyprian subduction zone. Open arrows mark the relative movement of the African and European continents.

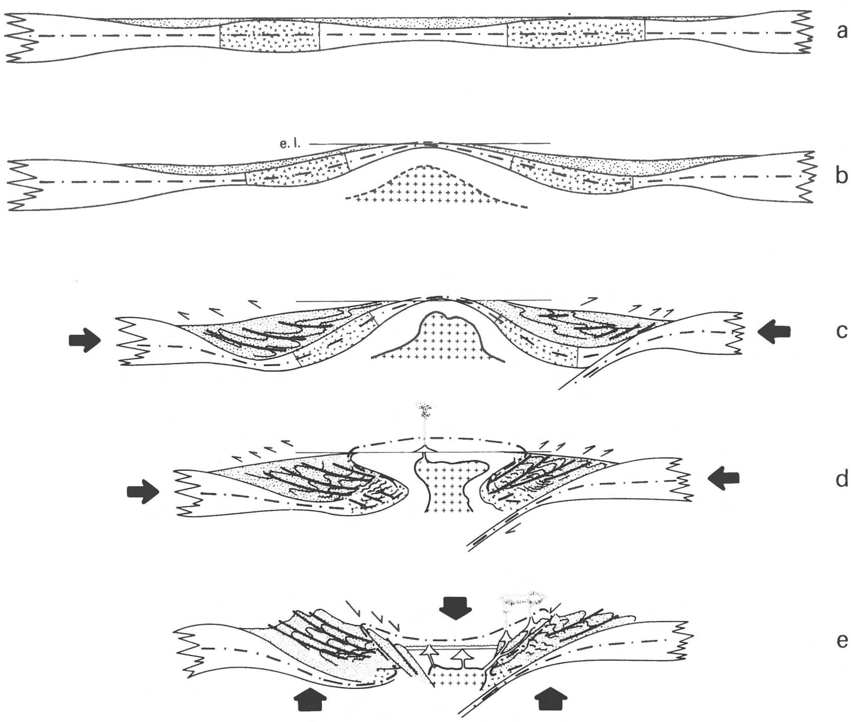

Fig. 7. Schematic cross-sections showing the evolution of the central Mediterranean basins and ranges in five steps. Ribbon continents (to either side of the central dome or basin), i.e. lithosphere relatively modestly attenuated, are marked to enable easy correlation with Fig. 6. Note that starting stage a is the simplified equivalent of a stage about halfway between phases 2 and 3 in Fig. 2. Stage e is the interpreted equivalent of Fig. 4. Subduction as shown on the right hand side of stages c, d and e is only a variation on the subsidence shown on the left hand side. Thin horizontal line (marked e.l. in b) represents the erosion level. The changing shape of the dash-dotted median line depicts the general deformation.

passive continental margins, each flanked for simplicity's sake by a singular ribbon continent. In Fig. 6C the cross-hatched areas mark the Alboran – S. Balearic, the Tyrrhenian and Pannonian 'Basins' which were emergent during the middle Tertiary. From these 'highs' nappes were shed and thrust towards and over their low-lying peripheries, to later collapse and subside to their present levels. The present geological structure and the subsidence histories of these basins support an evolution that resembles the earlier mentioned Lexa & Konecny (1974) diapir model, diapiric uprising and later collapse of the asthenosphere, for the evolution of the Pannonian Basin. This is best illustrated in cross-section, as in Fig. 7. The initial stage here is the equivalent of a stage in between phase 2 and 3 as depicted in the Labrador Sea evolution model (Fig. 2), not yet or hardly having reached the seafloor spreading configuration. While the margins interact and the oroclines are bent the intermediate region is warped, whereby attenuation of crust and lithosphere proceeds even, as shown, in regions now under compression.

The central region in between ribbons (Figs 6 and 7) which later evolves into Alboran – S. Balearic, Tyrrhenian and Pannonian basins, bulges over an asthenosphere blister or diapir. While it rises it is denudated by gravity (nappe) sliding and erosion, both effective mechanisms for heat transfer. In contrast the marginal basins buckle down and subside, aided by increasing loads of sediment. Later, when oroclinal bending and thus compression stops, isostatic adjustment will invert the relief; the central part again will become a basin while its margins will rise to attain Alpine height. Subsidence of the central area in fact already begins as soon as the thermal anomaly looses its protective isolating cover, which until then hampered rapid heat dissipation.

The reasons for the initial upwarp of the central basins and a downwarp for the marginal basins are not quite clear, they are, however, probably related to a difference in heat dissipation, which would have been relatively high for the central basin where attenuation proceeded farthest. It is important in this respect to also realise that those places where the ribbon continents are or were in close contact with each other (i.e. where there is no central basin) are exactly the places where ophiolites are abundant. In comparison with Fig. 7 those places would correspond with a central asthenosphere dome that was squeezed out and caught in between encroaching ribbons; later collapse and subsidence of that dome would be hampered, the ophiolites being carried on a light substratum.

Volcanism and the emplacement of ophiolites within orogenic belts are thus here considered in the light of a diapiric model (Fig. 7), rather than dogmatically as evidence of the closure of an oceanic basin in a exclusive island arc subduction fashion. In the present model there is ample room for attenuation and the breakthrough of asthenosphere and asthenosphere derivatives, both in the passive margin evolution stage and during the bulging and squeezing of intermediate basins between closing oroclines. This no doubt also agrees better with Coleman's (1984) earlier mentioned preference for emplacement of ophiolites by obduction of young, still hot oceanic slabs. It is interesting furthermore to realize that Bogdanov (1984) distinguished two different mechanisms for ophiolite emplacement in fold belts, one for Pacific-type fold belts, the other, the Ural-type or intercontinental orogeny, the latter thus according to the present model.

The asymmetry in the later stages of Fig. 7 is only a variation on the general theme, whereby on one side downbuckling and subsidence passes into underplating and subduction. Such is the case for the southern Adriatic Basin whereby the attenuated continental or quasi-oceanic region in between Apennines and Hellenides is thrust under the Calabrian Arc. This subduction in turn contributes to further attenuation under the Tyrrhenian Basin, which thus changes into a back-arc basin. For the northern Adria and Po Basin the loop is virtually closed and at depth either no attenuated crust is left, or originally attenuated crust in both margins is buckled and crumpled in the collision, thereby restoring the crustal thickness to continental proportions.

In the eastern Mediterranean the subsidence of the Levantine basins also changed into subduction under the Hellenic Arc, whereby the original emergent Aegean Sea (and the Anatolian Ova Basin) changed into back-arc basins. A similar situation may have existed in the western Alboran Sea for some time as testified by two isolated deep earthquakes south of Spain (1954 and 1973) which may be considered in evidence of the motion of a detached lithosphere fragment that deeply sank into the upper mantle.

The so-called Gibraltar Arc is not an orocline and the name (arc) is misleading; the arc-like structure is only the result of the squeeze-out and spill-over of the Alboran Sea diapir in all but easterly directions.

The evolution of the system as reconstructed by unwinding oroclines is generally confirmed, at least for the younger movements, by independent paleomagnetic evidence (cf. Van den Berg & Zijderveld 1982). We have to realise that the paleomagnetically established movements of obducted ophiolite complexes and of sedimentary nappes (i.e. material derived from basins in between the ribbons) can only reflect the movement of the ribbons after tectonic emplacement of this basin material on top of the ribbons. Before such time the basin emerged on top of the rising asthenosphere, vertical movement thus, independent of the horizontal rotations of the ribbons. In some cases, prior to diapirism, basin material may even have been dragged around in directions opposite to those of the ribbons (see Fig. 8). As can be seen in Fig. 7 the ribbons now lie deeply buried under thick piles of deformed sediment and are out of reach of our sampling tools.

Thus, as yet there is no independent paleomagnetic evidence for the early motions of the ribbon continents. Furthermore, even for the younger movements there are a few smaller structural elements which pose problems. According to paleomagnetic evidence a few elements seem to have been moving differently, not at the same rate or through the same rotational angle as found in the

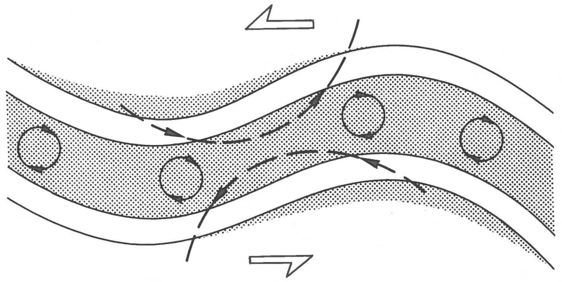

Fig. 8. Sketch showing how basins and their sedimentary fill in between ribbons may have been dragged around in directions opposite to the rotations of the ribbons prior to nappe emplacement and obduction.

regional trend, some elements may appear to have been moving even in the wrong direction. In the (broad) context of this paper such deviations are noise on an otherwise clear and consistent signal. Some discrepancies between the reconstructions and the paleomagnetic results are readily explained by accepting less coherence per belt than shown in Fig. 3, a configuration more like the upper Cretaceous Atlantic paleogeography of Ziegler (1978). Such an arrangement would allow for more independent movement of individual smaller blocks. Also, of course, movements of surficial nappes and thrustblocks are influenced by paleorelief and are often uncoupled from the movement of deep autochtonous basement by decollement zones, shales and evaporites.

As stated, Figs 6 and 7 are only first approximations and can only claim to portray the general picture. To further develop the model and add detail it will be necessary to change the embodiment of the evolution model for passive margins. Once more, coherence of ribbon continents will vary from place to place and it is therefore realistic to expect deviations from the general model, deviations that are related to discontinuities, fractures and offsets in those crustal elements.

There is one aspect in which the present model differs substantially from several other evolutionary models for the Alpine system and that is the behaviour, the movement and changing configuration of the peri-Adriatic block. Several authors (Biju-Duval et al. 1977; Hsü 1982; Van den Berg 1978) considered this so-called Apulian block (Apulia for short) or African promontory (promontoire africain), to consist of one rigid block, which as a whole rotated anti-clockwise and became detached from Africa. The present model requires an Adriatic suture, an anti-clockwise rotation of Italy, but a clockwise rotation of the Dinarides and Hellenides (cf. Fig. 6). Evidence for the present interpretation comes, at least in part, from the subduction of the southern Adriatic in a northnorthwesterly direction under the Calabrian Arc as testified by the associated somewhat crooked Benioff Zone. More so, the evidence comes from the paleomagnetic results of Laj et al. (1982) and Horner & Freeman (1983) for western Greece, which indeed support a clockwise rotation of eastern Apulia. Of course, there also, the rotations of the Hellenic and Dinaric nappes do not reflect the early rotations of autochtonous basement (the core of the ribbon continents) and they may deviate in part from the younger regional rotations. The paleomagnetic results should therefore be handled with as much care as those from the western limb of the peri-Adriatic promontory which all support anti-clockwise rotations. Both series of results, however, consistently fit the present model; they do not fit the one-block Apulia scheme. The present model also agrees with the evolution of the Hellenic arc and trench system advocated by Le Pichon & Angelier (1979) who challenged earlier 'established' paleomagnetic interpretations.

As mentioned before, in contrast to the open central part of the Alpine-Mediterranean complex the tight western and eastern zones do not show much in the form of loops and bends. The reason for that is a difference in the Paleozoic attenuation pattern between both regions; i.e. by comparison attenuation in the west and east was spread over more, shallower and narrower rifts and basins. Consequently during subsequent African-European transpressive interaction the outer regions sheared along major dislocations such as the S. Pyrenean Fault, the Betic Fault Zone, the S. Atlas Fault and the Anatolian Fault systems, while in the peri-Adriatic region loops could move relatively freely. This is a rather essential difference; after all, the loops would not loop if movement between ribbons and adjacent margins was not constricted in the outer region. In that case the entire Mediterranean would be sheared.

A simple demonstration will make this clear. If one takes one or two strips of pliable material, e.g. leather belts, in both hands and brings the hands together the strips will only bend, buckle or loop, if held tightly. Held loosely, they would slide through the hands.

Conclusions

The model developed intimately ties orogeny to plate interaction. Attenuation, rifting and (some) sea-floor spreading, transcurrent movement, collision subduction, volcanism, nappe and ophiolite emplacement in the Mediterranean all play a part in a logical sequence that reflects the opening of the Atlantic Ocean and the relative movements of the African and European plates. In contrast to now 'classic' orogenic models the emphasis is shifted from head-on collision to oblique collision in a transpressive regime, whereby the transcurrent component induces not just shear but also oroclinal bending. (Since the outline of plates is usually irregular, head-on collisions between plate margins, or parts of plate margins, are the exception rather than the rule). In many orogens, old and new, more and more transcurrent movements are recognised and thus the model is likely attractive for other 'Alpine' foldbelts for which oroclinal bending and diapiric (over)-thrusting and obduction thus far have not been considered.

As shown, the present model in essence is a modification of Carey's (1958) tectonic framework of the Mediterranean region. It also is a modification of allochtonous, exotic or suspect terrain configurations as are presently in vogue for Alpine-type foldbelts such as the NW American Cordillera (Ben Avraham et al. 1981; Jones et al. 1982) and the Appalachians (Williams & Hatcher 1982). In that scheme neighbouring terrains break apart, go through more or less incoherent, independent movements to be reassembled and trapped ultimately in the present chaotic configuration. The individual history of such terrains has been traced paleomagnetically. Allochtonous terrain configurations as a matter of fact have been recognised in the Alpine-Mediterranean region long before the terminology was introduced. Argand (1924) may perhaps be considered founding father of the Alpine fragmentationists and more recent advocates of such models are Scandone (1975), Hsü et al. (1977) and Frisch (1978, 1980). The allochtonous terrain model can, however, relatively easily be modified into the model presented here by allowing more coherence between exotic elements and permittng a less rigid tectonic scheme.

There is only one model known to me – other than the Carey (1958) one – that considers the formation of the Alpine loops in a semi-mobilistic way. Both Tapponnier (1977) and Caire (1979) proposed a schema whereby, in the case of the collision of Europe and Africa a rigid promontory of one side (Africa) is driven or pushed into a pliable other side (Europe), which in consequence was deformed into foldbelts that drape and spill around and over the 'punch'.

An important aspect of the present oroclinal-diapiric model is that it substantiates evolutionary schemes for Alpine-type orogenies that are logical sequels to earlier phases of attenuation; the Alpine foldbelt could not have formed were it not for a Paleozoic-Mesozoic attenuation phase which produced two passive continental margins, thus conditioning the region for orogenesis during the later transpressive phase. Otherwise stated, collision of two continental plates without attenuated margins, after some shuddering, shaking and cracking, would quickly come to a grinding halt without much to show in effect thereof.

Finally, the Alpine foldbelt is not a Pacific-type orogeny, an assemblage of accretionary wedges in a compressed island arc – volcanic arc – subduction complex. Although subduction played a moderate role (Calabrian, Hellenic and Carpathian Arcs, Udias 1982), obduction of attenuated basins over adjacent (ribbon) continental margins appears to have been of much greater importance and to be in better agreement with known lithology, stratigraphy and structure.

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