Origin of the Betic-Rif mountain belt

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Abstract. In recent years, the origin of the Betic-Rif orocline has been the subject of considerable debate. Much of this debate has focused on mechanisms required to generate rapid late-orogenic extension with coeval shortening. Here we summarize the principal geological and geophysical observations and propose a model for the Miocene evolution of the Betic-Rif mountain belts, which is compatible with the evolution of the rest of the western Mediterranean. We regard palaeomagnetic data, which indicate that there have been large rotations about vertical axes, and earthquake data, which show that deep seismicity occurs beneath the Alboran Sea, to be the most significant data sets. Neither data set is satisfactorily accounted for by models which invoke convective removal or delamination of lithospheric mantle. Existing geological and geophysical observations are, however, entirely consistent with the existence of a subduction zone which rolled or peeled back until it collided with North Africa. We suggest that this ancient subducting slab consequently split into two fragments, one of which has continued to roll back, generating the Tyrrhenian Sea and forming the present-day Calabrian Arc. The other slab fragment rolled back to the west, generating the Alboran Sea and the Betic-Rif orocline during the early to middle Miocene.

Introduction

Extensional basins associated with convergent plate boundaries are common in the western Pacific Ocean and in the Mediterranean Sea [e.g., Vine and Smith, 1981]. Whilst many of the Pacific Ocean basins have clearly formed as a result of back arc extension caused by subduction zone rollback, the origin of geometrically similar basins in the western Mediterranean Sea has proved more controversial. Much of the current debate is focused on the origin of the Alboran Sea together with the surrounding Betic-Rif mountain belts, the most westerly of the Alpine mountain chains of southern Europe (Figure 1). This orocline developed during, and partly in response to, Late Mesozoic to Tertiary convergence between Africa and Iberia. In recent years, three principal tectonic models have been proposed to account for its geometry and evolution: (1) rapid westward motion of a rigid Alboran microplate, (2) subduction zone rollback, and (3) radial extensional collapse caused by rapid convective removal or delamination of lithospheric mantle.

Here we first summarize the important geological and geophysical observations that must be accounted for by any model for the evolution of the western end of the Mediterranean Sea. The limitations of models which rely on the rapid, convective removal of lithospheric mantle shall then be discussed. We argue, instead, that the most important observations are consistent with a short, steep east dipping subduction zone which peeled or rolled back at a rate of between 50 and 100 mm/yr to the west during the early Miocene until subductable oceanic lithosphere was exhausted in the vicinity of Gibraltar.

Traditionally, rocks of the Betic-Rif orogen have been divided into External Zones, Internal Zones, and Flysch nappes (Figure 2). In the Betic Cordillera, the External Zone consists of Mesozoic and Tertiary sedimentary rocks which were deposited in basinal (Subbetics) and shelf (Prebetics) environments on the Iberian margin of the Tethyan Ocean. These rocks were deformed by northwest directed, thin skinned thrusting and folding during the early to late Miocene [Garcia-Hernandez et al., 1980; Banks and Warburton, 1991; Allerton et al., 1993]. Shortening was accompanied by development of the Guadalquivir foreland basin. The Internal Zone, located farther south, consists of mountain ranges of metamorphosed Palaeozoic and Mesozoic rocks separated by Neogene intermontane basins. These metamorphic rocks were affected by Palaeogene-early Miocene penetrative deformation and regional metamorphism.

A threefold division of the Rif belt of North Africa is generally accepted [Wildi, 1983] (Figure 2). Its Internal Zone contains metamorphic rocks broadly similar to those of the Betic Cordillera, although the nomenclature differs. The intermediate Flysch Zone, informally known as the Flysch nappes, consists mainly of Early Cretaceous to early Miocene deep marine clas-
tic deposits [Wildi, 1983]. The External Zone of the Rif is made up of Mesozoic and Tertiary sedimentary rocks deposited on the North African margin. During the early Miocene, its Internal Zone was thrust onto the Flysch nappes. Continued convergence during the Miocene deformed the External Zones into a thin-skinned fold and thrust belt [Andrieux, 1971, Frizon de Lamotte, 1987, Morley, 1988]. Continental crust beneath the Alboran Sea itself consists of metamorphic rocks which are similar to those of the adjacent Internal Zones [Platt et al., 1996].

Key Observations

Coeval Extension and Shortening

The Alboran domain and adjacent Internal Zones underwent considerable extension at the same time as northwest, south, and southwest vergent thrusting occurred in the External Zones of the Betic and Rif mountains, respectively (Figure 3) [Frizon de Lamotte, 1987, García-Dueñas et al., 1988, Platt and Vissers, 1989, Banks and Warburton, 1991, Faure et al., 1991, García-Dueñas et al., 1992, Comas et al., 1992, Lonergan et al., 1994, Jaboloy et al., 1993, Watts et al., 1993, Platsman et al., 1993]. Close to Gibraltar, thrusting is directed westward [Balanyá and García-Dueñas, 1987, Platsman et al., 1993, Platt et al., 1995]. The ENE striking Internal/External Zone boundary and the Crevillente fault of the Betics together with the NE striking Jebha and Nekor faults of the Rif were originally interpreted as the northern and southern boundaries of a westward moving Alboran microplate [Andrieux et al., 1971, Leblanc and Olivier, 1984]. However, recent structural studies have shown that oblique convergence occurred along the Internal/External Zone boundary during early Miocene times [Lonergan et al., 1994, Allerton et al., 1993].
Wide-angle seismic data modeling shows that the crust varies from 40 km thickness at the Internal/External Zone boundary in the Betics to about 15 km thickness in the center of the Alboran Sea [Banda and Anosy, 1980; Torné and Banda, 1992]. Seismic reflection profiles which traverse the Alboran Sea also show evidence of large-scale normal faulting [Watts et al., 1993; Comas et al., 1992; Mauffret et al., 1992]. Analysis of Neogene subsidence using well-log information from the northern margin of the basin gives stretching factors of 1.3-1.6 [Watts et al., 1993].

There is also good evidence for extension both within the metamorphic rocks of the Internal zones [e.g., García-Dueñas et al., 1992; Jabaloy et al., 1993; Crespo-Blanc et al., 1994] and within the well-exposed Neogene basins which occur onshore Spain [e.g., Montenat et al., 1987; Mora-Grulkstadt, 1993]. The existence of large amounts of crustal extension within the metamorphosed Internal Zones is inferred from the structural juxtaposition of rocks of significantly different metamorphic grade along major extensional shear zones such as the Betic movement zone [Platt and Vissers, 1989] and the Alpujarride/Malaguide contact [Aldaya et al., 1991, Loneran and Platt, 1995].

Rapid exhumation of metamorphic rocks in the Internal Zones of the Betics has been documented using both radiometric dating and stratigraphic provenance studies [Zeck et al., 1992; Monié et al., 1991, 1994; Loneran and Mense-Rajetzky, 1994]. Exhumation and rapid cooling of metamorphic rocks occurred from earliest Miocene times to ~ 15 Ma, coeval with extension.

**Rotations**

Palaeomagnetic studies confirm that thrust sheets within the External Zones of the Betic mountains have rotated clockwise about vertical axes by up to 130 degrees post-Oligocene, probably in the Miocene [Osete et al., 1988, 1989; Platzman, 1992, Allerton et al., 1993, Platt et al., 1995]. Even larger rotations (~ 200 degrees) have been documented in Malaguide rocks on the northern margin of the Internal Zone in the eastern Betics [Allerton et al., 1993]. In the Rif, counterclockwise rotations of ~ 100 degrees have been measured [Platzman et al., 1993].

Most of the published palaeomagnetic data for the
Figure 3. Timing of Neogene events in Betic Cordillera and Alboran Sea.

Betic and Rif mountains were collected from *Ammonitico rosso* and *capas rojas* facies limestones of Upper Jurassic and Upper Cretaceous ages, respectively. Tectonic rotations must have occurred after deposition and are likely to have been associated with folding and thrusting during the early to middle Miocene [Allerton, 1994]. The Allerton et al. [1993] study was directed at magnetic palaeomagnetic rotations in a suite of rocks ranging in age from Triassic to upper Miocene located at the northern margin of the Internal Zones in the eastern Betics. Triassic, Jurassic, and upper Oligocene rocks within a thrust stack all show similar amounts of rotation (~200°), but Oligo-Miocene marls from the associated foreland basin are only rotated by 140°. This result led Allerton et al. [1993] to conclude that rotation started in the upper Oligocene and continued while foreland basin sediments were being deposited. Undefomed upper Miocene marls from the adjacent intermontane basin are not rotated, providing an upper time constraint on the rotations. If rotation started in the late Oligocene, the maximum time interval for the observed 200° rotation is ~30 Ma, giving a minimum rotation rate of ~7°/Ma. However, if we assume that rotation was contemporaneous with contractional deformation along the Internal/External Zone boundary, then rotation must have ceased by Langhian times (~15 Ma [Lonergan et al., 1994]), suggesting a rotation rate of ~13°/Ma. In the Subbetic of the eastern Betics, the maximum rotations are 130°. Here shortening commenced in the early Miocene (~22 Ma), and if we assume rotations took place between the early Miocene and the present day, the minimum rotation rate is 6°/Ma. If rotational deformation is accommodated along thrusts active in the Subbetic during the early-middle Miocene (22-15 Ma), then the rotation rates could have been as high as 18°/Ma. Rotation rates of 6°-18°/Ma are the same order of magnitude as those proposed by Kissel and Laj [1988] for parts of the Aegean region.

**Volcanism**

Volcanism has both accompanied and postdated Neogene extension: calc-alkaline, potassic, and basaltic volcanism is scattered across the eastern sector of Alboran Sea and Betic-Rif systems (Figures 2 and 3) [Bellon, 1981, Hernandez and Bellon, 1985, Hernandez et al., 1987]. The earliest igneous activity is a basaltic dyke swarm at ~22 Ma (K/Ar dating) located in the central and western Internal Zones of the Betics [Torres Roldán et al., 1986]. The earliest volcanic rocks are mostly calc-alkaline and in Spain are restricted to the Sierra de Gata-Carboneras area. In the eastern Rif, calc-alkaline volcanic rocks are more widespread, occurring in Ras Taraf, Trios Fourches, and in the western Tell near Oran. In southern Spain, the dates obtained for this volcanic suite range from 15 to 7 Ma, whereas in the Rif-Tell mountains, dates range from 13–8 Ma [Bellon, 1981, Hernandez et al., 1987]. Offshore, Alboran Island itself is a calc-alkaline volcanic edifice. Early work, quoted by Bellon [1981], suggested an age of 24±5 Ma, but more recent work by Apatico et al. [1991] dates the volcanism at 18–7 Ma, which is in more general agreement with the ages obtained for onshore suites of similar composition in both Spain and North Africa (Figure 2). Relatively large amounts of volcanism are inferred offshore from both magnetic data and dredging; these volcanic rocks are also assumed to be calc-alkaline in composition [Hernandez et al., 1987].

A second broad suite of dominantly potassium-enriched rocks with a wide variety of compositions (shoshonitic to lamproitic) were erupted in Spain between 8 and 5 Ma and in North Africa between 9 and 4 Ma. The lamproitic volcanic rocks are very widely scattered (Figure 2). In southern Spain, the youngest lavas are Plio-Quaternary alkaline basalts which were erupted near Cartagena. The youngest lavas in North Africa are also alkaline basalts, ranging in age from Messinian (5.0 Ma) to Quaternary.

In the past, both north and south dipping subduction zones have been postulated between Spain and Africa [e.g., Torres-Roldán et al., 1986, Sanz de Galdeano, 1990]. However, it is clear from Figure 2 that such subduction zones are not compatible with the spatial distribution of volcanism. More recently, French workers [Hernandez et al., 1987, de Larouzère et al., 1988, Montenat et al., 1987], noting the spatial relationship between volcanic rocks and late Neogene strike-slip fault systems in SE Spain (Carboneras-Palomares faults), have attempted to relate the volcanism to a “trans-Alboran” shear zone. Seismic reflection data in the Alboran Sea [e.g., Watts et al., 1993] provide no
evidence for a continuous shear zone offshore. The pattern of early calc-alkaline volcanism in the Alboran Sea is not controlled by the location of strike-slip faults. Existing offshore data suggest that the trend of calc-alkaline volcanic rocks crudely follows the strike of the orocline.

Seismicity

Buforn et al. [1990] summarize data for three deep earthquakes which occurred beneath Granada this century. These earthquakes testify to the existence of material which is cold and rigid enough to produce sudden releases of strain energy at considerable depths beneath the Alboran Sea. The best known and largest of these earthquakes occurred in 1954 (\(M_s = 7.1\), Figure 4) at a depth of \(\sim 630 \pm 4 \) km. The \(P\) axes for all three earthquakes dip steeply east [Chung and Kanamori, 1976, Grimson and Chen, 1986, Buforn et al., 1988, 1990]. Good station distribution throughout Europe means that the nodal plane strike of the 1954 event is well determined (\(2^\circ \pm 5^\circ\) E). Intermediate-depth and shallow seismicity occurs throughout the region, the biggest events indicating oblique convergence between Africa and Europe (Figure 4). In the Alboran Sea, intermediate events do not occur below 150 km, and there is also a smaller gap between 60 and 100 km [Seber et al., 1996].

Grimson and Chen [1986] and Buforn et al. [1990] point out that the presence of a steep, east dipping, slab is very difficult to reconcile with the Neogene and Recent plate motion vectors. These vectors are oriented north-south to northwest-southeast and have been used to invoke the existence of a north or south dipping subduction zone between North Africa and Europe. The above authors also argue that accepted plate reconstructions preclude the existence of 600 km of oceanic crust between Africa and Iberia during the Tertiary [e.g., Dewey et al., 1989].

This geometric problem has fueled more recent speculation that the deep earthquakes are evidence for a block of lithospheric mantle which has been convectively removed [Grimson and Chen, 1986]. Seber et al. [1996] suggest that the distribution of relocated intermediate-depth seismicity is consistent with active delamination. Unfortunately, their analysis ignores the existence of the deep 1954 event, which dominates the total seismic moment release within the region. Furthermore, their inference that a large zone of upwelled astheno-
The presence of an east dipping P axis within an eastward dipping Wadati-Benioff zone is entirely consistent with the existence of a N-S striking subducted slab. The large seismic gap between 150 and 600 km suggests that this cold subducted slab has now detached, in agreement with tectonic evidence that 12-15 Ma have elapsed since subduction terminated. We also note that, in fact, there is little difficulty in subducting 600 km of old oceanic lithosphere if the subduction zone rolled back from a position west of the current eastern margin of Iberia toward Gibraltar.

Timing and Plate Motions

Foreland-directed thrusting, coeval with extension, occurs on all three sides of the orocline (Figure 3). On the Spanish mainland, typical foreland-basin style subsidence occurs within the Guadalquivir basin. This basin can be traced from the Prebetic arc in the east, through Gibraltar and onto the Moroccan margin where it is called the Rharb basin. Seismic reflection data show that deformed units (the Flysch/External Zones) occur up to 250 km west of Gibraltar, probably representing the most westerly manifestation of shortening [Laït et al., 1975]. Using bathymetric data, Laït et al. [1975] also propose that shortening, manifest as olistostromes, extends up to 500 km west of Gibraltar and that Tortonian sedimentary rocks are shortened.

Since latest Tortonian times (8–9 Ma), the Africa-Eurasia plate convergence vector has changed from north-south to northwest-southeast [Dewey et al., 1989, Mazzoli and Helman, 1994]. As a consequence, dextral oblique shortening is now occurring between North Africa and Iberia. This overall motion is manifest by strike-slip faulting, local folding, thrust faulting, and structural inversion of previous normal faults which affected sediments within both the Alboran Sea and within onshore Neogene basins [Woodside and Maldonado, 1992, Watts et al., 1993, de Larouzière et al., 1988]. Since the Pliocene, the onshore Neogene basins in southern Spain have been rapidly uplifted, attesting to continued convergence between Africa and Eurasia and/or slab break-off.

Models for Generating Late Orogenic Extension

The paradox of coeval shortening and extension led Dewey [1988], Platt and Vissers [1989], and Doblas and Ogarrun [1989] to propose that the Alboran Sea and surrounding mountain chains formed by extensional collapse of a previously thickened crust and lithospheric mantle. Their work follows the earlier suggestion of Grimming and Chen [1986] who used teleseismic earthquake data to infer the existence of a detached fragment of lithospheric mantle. Several other workers have elaborated these ideas [Platzman, 1992, Platt and England, 1994, Vissers et al., 1995]. Along with the Tethyan plateau, the Betic-Rif mountain belt is now regarded as an important example of extensional collapse driven by rapid, convective removal or delamination of thickened lithospheric mantle [Houseman et al., 1981, England and Houseman, 1989]. Here we briefly review two principal mechanisms which have been used to account for the main observations in the western Mediterranean: convective removal or delamination of the lithospheric mantle and subduction zone rollback. We shall not consider models which invoke westward motion of an Alboran “microplate” within a zone of overall convergence between Africa and Europe [Andreux et al., 1971, Leblanc and Olivier, 1984]. Although this is an appealing mechanism for producing the dramatic orcinal bend about the Straits of Gibraltar, it is evident that the Alboran domain did not behave rigidly during the Miocene.

Convective Removal/Delamination Models

Using fluid dynamical arguments, Parsons and McKenzie [1977] showed that the vigorously convecting, adiabatic interior beneath the oceanic plate is separated from the mechanical boundary layer of the lithosphere (~ 80 km thick) by a thermal boundary layer (~ 50 km thick). This thermal boundary layer has a lower viscosity than the mechanical boundary layer and acts to maintain the thickness of the lithosphere at 120 km by periodically overturning [England, 1983]. Subsequently, Houseman et al. [1981] carried out a range of numerical experiments which showed that if an equilibrated lithospheric plate is instantaneously thickened, then the thermal boundary layer becomes unstable and drops off on timescales of about 10 Ma for geologically realistic Raleigh numbers. In their calculations, they systematically varied the thickness of the thermal boundary layer by 1 order of magnitude.

The principal difficulties with convective removal/delamination models are twofold. First, the time taken for delamination to occur is sensitive to the viscosity structure of the lithosphere. Buck and Toksöz [1983] have carried out a much smaller number of numerical experiments, and they conclude that only a thin one at the base of the thermal boundary layer is removed during convective overturn. This disagreement is a consequence of uncertainties in the rheological properties of the upper mantle. Furthermore, removal of lithospheric mantle is an inefficient means for generating substantial potential energy differences because the density contrast between lithospheric mantle and as-
thensphere is small ($\sim 0.13$ Mg m$^{-3}$). Second, Houseman et al. [1981] assumed that shortening of the lithosphere was both uniform and instantaneous. If, however, shortening occurs at geologically realistic strain rates, the thermal boundary layer will probably convectively overturn on a continual basis.

If convective instability of a thickened thermal boundary layer can take place sufficiently rapidly and on a large enough scale, then this model for the Betic-Rif area is appealing. However, there are two important sets of observations which are not easily explained by a combination of lithospheric thickening and radial extensional collapse. First, shortening has only been taken up on three sides of the orogen and is in no sense radial (Figures 1 and 2). It is unclear how the free eastern edge east of the Alboran Sea can be accounted for, except by inferring that extensional collapse generated the whole of the western Mediterranean region. This inference requires evidence for coeval outward directed thrusting all around the western Mediterranean Sea. Second, no serious attempt has been made to account for the large, clockwise palaeomagnetic declinations observed in the External Zones of the Betic mountains and for the comparable counterclockwise declinations within the Rif. Platman [1992] suggests that these rotations could be incorporated into an extensional collapse model, although she concludes that the palaeomagnetic data require some westward motion of the Alboran region. In summary, large vertical-axis rotations are the crucial observation which cannot easily be accounted for by convective removal or delamination. McKenzie and Jackson [1983] showed that the spreading of either a circular or elliptical blob does not generate any rigid body rotation.

**Potential Energy Calculations**

If significant convective removal occurs, it will generate large amounts of gravitational potential energy. It has long been recognized [e.g., Jeffreys, 1952, Bott and Dean, 1972] that whilst the lithosphere is in approximate isostatic equilibrium, considerable lateral variation in density means that the stress balance is not complete: horizontal deviatoric stresses can be of the order of 100 MPa. If the gravitational potential energy of a region of continental lithosphere exceeds that of its surroundings, then that region will be subjected to extensional deviatoric stresses. Regions with greater potential energy are usually at greater surface elevation, but since potential energy depends upon the density distribution within the lithosphere, elevation alone cannot be used to calculate potential energy differences. It has been shown that differences in the depth-averaged horizontal deviatoric stresses for two lithospheric columns in isostatic equilibrium are proportional to differences between their gravitational potential energies [Molnar and Lyon-Caen, 1988]. Lithospheric columns are balanced with reference to mid-oceanic ridges where it is reasonable to assume that horizontal and vertical stresses are equal (i.e., $\sigma_{xx} = \sigma_{zz}$).

Extensional collapse of the continental lithosphere is thought to be triggered by a rapid increase in the potential energy of a lithospheric column [Platt and England, 1994]. This inference is based upon simple calculations which show that if the bottom part of thickened lithospheric mantle is suddenly removed, a rapid increase in potential energy occurs (Figures 5a and 5b and appendix). This increase partly manifests itself as an increase in surface elevation. It is generally assumed that rapid increases in potential energy ($\sim 5 \times 10^{12}$ Nm$^{-1}$) will give rise to extensional collapse.

The difficulty in using potential energy calculations to argue in favor of extensional collapse models is that the continental lithosphere is clearly able to sustain large differences in potential energy without generating extension. For example, the potential energy difference between southern Africa and the surrounding old ocean basins is as large as that expected by convective removal calculations (compare Figures 5b and 5d). In fact, differences in potential energy may have no particular significance: it is more important to develop an understanding of mechanisms which allow stored potential energy to be released rather than interpret potential energy differences per se. Even if large differences in potential energy did lead to extensional collapse, it is not clear why extension should continue after thickened continental crust has been thinned and its upper surface restored to sea level, unless subsidence below sea level is entirely caused by thermal contraction.

**Subduction Zone Rollback Models**

At many island arcs, the subducting slab retreats away from the arc in the hotspot frame of reference [Molnar and Atwater, 1978, Chase, 1978, Dewey, 1980]. This peeling or rolling back of a slab generates and maintains back arc extension. At continent-ocean subduction zones, rollback is probably the principal means by which potential energy stored in thickened continental lithosphere is released.

Rollback is a natural consequence of subducting old oceanic lithosphere which is colder, and therefore denser, than the mantle through which it sinks [Le Pichon and Angelier, 1981]. Within the subducting slab, the vertical negative buoyancy force $F$ can be resolved into two components: one component at right angles to the dipping slab $R$ and a second component along the slab $P$ (Figure 6a). If the slab is in static equilibrium, $P$ is balanced by viscous forces on the two surfaces of the slab and by resistive forces at the slab tip. The only way in which $R$ can be supported is if the pressure within the asthenospheric mantle beneath the dipping slab is...
Figure 5. Potential energy calculations for different columns of lithosphere. (a) cartoon illustrating two end-members for which potential energy difference is calculated. Reference column of continental lithosphere with 30 km thick crust is in isostatic equilibrium with standard mid-ocean ridge column. (b) Calculated potential energy difference ($10^{-3}$ Nm$^{-1}$) as a function of thickening for two end-members shown in Figure 5a. Upper curve is labeled "convective removal": lithosphere which has been thickened by factor $f$ but lower part of lithosphere has been removed to maintain lithospheric thickness at 125 km. Lower curve is labeled "uniform thickening": lithosphere allowed to thicken uniformly. If convective removal of thickened lithospheric mantle occurs at $f = 1.8$, then the path taken is shown by solid arrowheads. A value of +0.5 is maximum level of potential energy difference likely to be sustainable on Earth. (c) and (d) Potential energy calculations for southern Africa compared with surrounding ocean basins. Crustal thickness taken from Qiu [1995].

Greater than the pressure above the slab. This pressure difference is superimposed upon the hydrostatic pressure gradient and so will tend to drive mantle material from high to low pressure thus facilitating rollback (Figures 6b and 6c). The buoyancy forces generated by slabs greatly exceed other convective forces within the mantle, and so motion will largely be controlled by the density excess of the slab. The component of velocity that is normal to the slab is determined by the ease with which material can flow around the slab. Flow can easily occur in the strike direction (i.e., around the edges of the slab), but the increase of viscosity with depth means that flow around the tip of the slab is more difficult. Hence the normal component of velocity is greatest at subduction zones that have a short strike length (< 1000 km).

When subduction zone rollback is initiated at a continent-ocean margin, it is unlikely that the extensional deviatoric stresses within thickened continental lithosphere can be transmitted through rigid oceanic plate to the mid-ocean ridge. As a result, continental lithosphere will collapse by rapid extension into the space provided.

and obtained 160 mm/yr for the opening of Lau basin with convergence rates of 240 mm/yr across the Tonga Trench.

Once rollback commences, it should continue until the system completely runs out of dense oceanic lithosphere. Thus there are two principal ways by which rollback ceases. First, it ceases when the retreating subduction zone collides with mid-oceanic ridge. Thus it is likely that the Scotia Arc will continue to roll back until it meets the mid-Atlantic ridge. Second, when the arc collides with continental lithosphere (i.e., dense oceanic lithosphere is completely consumed) rollback ceases. This more interesting case is of especial importance within continent-ocean-continent collisions where the original passive continental margins form irregular boundaries (e.g., during the closure of the Tethyan Ocean). When an arc collides with an irregular continental boundary, the subducting slab will probably split into two parts with each segment continuing to roll back.

As well as generating rapid back arc extension within a zone of overall convergence, rollback is also an excellent means for generating rapid rotations about vertical axes. The classic examples come from southeast Asia where palaeomagnetic declination anomalies are associated with the opening of many marginal seas (see Jarrard and Sasajima [1980] for summary). Palaeomagnetic studies from many of these island arcs clearly demonstrate the role of rapid block rotations about vertical axes in back arc basins. The largest rotations occur in the Japan, Ryukyu, and Marianas basins where rotations of up to 80° have often been observed and where major differences in the amount of rotation experienced by different segments of the same arc occur. For example, palaeomagnetic data from southwest Japan is plentiful and clearly demonstrates clockwise rotations of 38°–47°. Most of this rotation occurred between 16 and 14 Ma [e.g., Otofuji et al., 1991]. In contrast, northeast Japan rotated counterclockwise by ~ 32°. These estimates yield an angular velocity of 20°/Ma. It is interesting to note that the physiography of the Japan Sea is a mirror image of that of the western Mediterranean (Figure 7).

### Subduction Rollback in the Alboran Sea

We contend that the Neogene structure and evolution of the Betic-Rif mountain belts are best explained by the westward rollback of a short east dipping subduction zone [Prizot de Lamotte et al. 1991, Royden, 1993]. Since late Oligocene times, the African plate has
converged with Iberia by about 300 km [Dewey et al., 1989].

An important argument in favor of a subduction rollback model in the western Mediterranean is that the boundary separating extension from shortening follows the oroclinal bend around the Straits of Gibraltar. The Flysch nappes and External Zone “olistostromes” west of Gibraltar may be analogous to the imbricated trench sediments which are associated with arcs. If so, their location and strike can be used to pinpoint the final position of the inferred north-south striking subduction zone. Continental lithosphere at the Straits of Gibraltar has probably prevented rollback from proceeding any farther. In contrast, Royden [1983] suggests that rollback may have continued as far west as the Horseshoe Seamounts.

As described above, volcanism is relatively young when compared with deformation of Internal Zones. It is coeval with and younger than extension and is diffusely located within the Alboran Sea, onshore North Africa, and in southeast Spain. Such a pattern is difficult to attribute to either a north or south dipping subduction zone, but it is compatible with an east dipping subduction zone which rapidly rolled back to the west. The range of composition of volcanic rocks from calc-alkaline to alkaline and basaltic is typical of small back arc basins where both arc and extensional volcanism occur. In the Alboran Sea, no calc-alkaline volcanic rocks have been reported between the Straits of Gibraltar and ~ 4° west (Figure 2). Assuming that the final surface position of the Alboran subduction zone was somewhere between the Straits of Gibraltar and the westerly limit of deformation (the “olistostromes”) and that the calc-alkaline volcanic edifices in the Alboran Sea represent the dismembered remnants of a small arc, then the final arc-trench gap was of the order of 200 km. This distance is similar to the arc-trench gap between the Aeolian Islands and the Calabrian trench. The lack of a clearly defined volcanic arc in the Alboran Sea may be attributed to one of two causes: (1) late Miocene to Recent strike-slip faulting postdating the cessation of subduction and thus dismembering the arc, or (2) an originally narrow subduction zone that moved rapidly, preventing a large, stationary arc edifice from forming.

In summary, the geology of the Betic Rif orocline and formation of the Alboran Sea can be accounted for by a north-south striking subduction zone which originally dipped east. Rocks that now make up the basement of the Alboran Sea and the Internal Zones of the Betic and Rif mountains originally lay to the east of the subduction zone and formed a collisional wedge. As the subduction zone rolled back, thickened crust behind the subduction zone extended rapidly to fill the space generated by the retreating subduction zone. Rollback of the subduction zone and outward displacement of the extending Internal Zones-Alboran Sea impinged on the passive margins of both the Iberian and African plates, causing the onset of oblique thrusting and rotations in the External Zones of the Betics and Rif mountains in the early Miocene.

Shortening in the External Zones and extension in the Alboran Sea terminated by the late Tortonian (Figures 3 and 8), suggesting that the subducting slab had rolled back as far as the Straits of Gibraltar by this time and that the Alboran Sea basin had achieved its present-day dimensions. As the narrow subduction zone peeled back, rotational deformation on the margins was taken up by oblique-slip thrust faulting in the External Zones. Large rotations are probably also associated with major extensional faults in the Internal Zones and Alboran Sea, similar to those observed in the Aegean. The only published palaeomagnetic data for the Internal Zones come from the Cabo de Gata volcanic province where rotations are associated with

Figure 7. Bathymetry of (a) western Mediterranean, (b) South China Sea, and (c) Japan Sea. Note similar physiographies.
late Miocene strike-slip deformation in the Carboneras fault zone [Calvo et al., 1994]. Pervasive west directed stretching lineations measured in metamorphic rocks of the western and central Betics [e.g., García-Dueñas et al., 1988] are consistent with this part of the orogen having undergone the largest amount of westward translation behind the retreating slab.

On the basis of the timing of onset and termination of both shortening in the External Zones and extension in the Internal Zones and Alboran Sea, we estimate that rollback occurred in the Alboran Sea between 23 and 10 Ma. Depending upon where the postulated subduction zone started east of Iberia, it travelled between 500 km (i.e., the length of the Betic orogen) and 900 km, yielding a rollback rate of 42–75 mm/yr with associated rotation rates of between 6°–18°/Ma.

Evolution of the Western Mediterranean Sea

By elaborating upon the ideas of Alvarez et al. [1974], Rehault et al. [1985], Dewey et al. [1989], Royden [1993], and Tricart et al. [1994], we propose that the now extinct subduction zone beneath the Alboran Sea was originally contiguous with the Calabrian Arc, currently located just north of Sicily. This original arc split into two shorter arcs, probably in the middle Miocene when the central segment of an original northward dipping subducting slab collided with the North African continental margin to form the Kabylies.

The logical step is to link the Tyrhenian subduction system to the subduction zone responsible for formation of the Betic-Rif orocline. In the Oligocene, prior to the opening of the western Mediterranean, the rocks now making up the metamorphic belts in Corsica, Sardinia, Calabria, Sicily, the Kabylies of North Africa, and the Rif and Betic Internal Zones must have been grouped behind a north dipping subduction zone as shown schematically in Figure 9a. The similarity in the geology of these now dispersed metamorphic belts has been noted by many previous authors (compare the AlKaPeCa terrane of Bouillin et al. [1986]). Once the subduction zone began to roll back, extension commenced and Corsica and Sardina rotated counterclockwise, reaching their current position by 19±1 Ma [Mon-
Figure 9.
There is general agreement that the three Balearic and north Tyrrhenian basins initiated in a back arc setting behind the north dipping subduction zone in the early Miocene [Cohen, 1980, Rehault et al., 1985, Malinverno and Ryan, 1986, Kastens et al., 1988, Dewey et al., 1989]. As a consequence of this extension, Corsica and Sardinia rotated counterclockwise, colliding with continental Adria, and initiated the formation of the Apennines [Alvarez et al., 1974]. Continued south and southeast directed rollback of the subduction zone led to the formation of the southern Tyrrhenian Sea and the Calabrian Arc (see Dewey et al., [1989] for further details). As shown in Figure 8, the oldest documented extension in the western Mediterranean area occurs onshore in southern France where the narrow Oligocene graben of Camargue and Provence have an east-west strike [Rehault et al., 1985]. The extensional graben in western Sardinia which extends from Cagliari to Cagliari now has a north-south strike, but during the late Oligocene-early Miocene, this basin was parallel to the Gulf of Lions graben [Rehault et al., 1985]. By early Miocene times (23–20 Ma), extension started in the Valencia Trough, Balearic basin, Alpero-Provencal basin, and Alboran domain, suggesting that the same mechanism is responsible for the formation of these basins. The onset of rifting in the Valencia Trough gave rise to clockwise rotation of the Balearic Islands [Parés et al., 1992]. The dispersion of the other units ultimately led to the formation of new oceanic crust in the Alpero-Provencal basin (21–18 Ma [Rehault et al., 1985]).

We suggest that later collision of the Kabylies continental fragments with North Africa forced this subduction zone to divide into two segments: a short segment oriented north-south which subsequently tore around to form the present-day Alboran Sea and a longer segment which continued to evolve until the present day, forming the Tyrrhenian Sea and the Calabrian Arc (Figure 9b). By 18 Ma, the rotation of Corsica and Sardinia was complete, shortening had commenced in the northern Apennines, and the collision of the Kabylies with Africa had commenced [Cohen, 1980, Wildi, 1983, Tricart et al., 1994]. Shortening was also underway in the External Zones of the Betics and Rif. Hence the subduction zone must have been approximately located as sketched in Figure 10b. Continued convergence led to the evolution of two increasingly looped branches of the original subduction zone in the Alboran and Tyrrhenian seas, respectively. By Tortonian times, the Alboran Sea had evolved to its current size and shape. At the Calabrian Arc, rollback has continued throughout the late Miocene until the present day (Figure 9b) [Anderson and Jackson, 1987]. The current position and asymmetry of the flysch basins from Sicily around to the Rif in the western Mediterranean supports this subduction zone geometry, provided that the Flysch nappes represent the deformed deposits of a trench-forearc basin.

**Volcanism**

Figure 10b summarizes the age and distribution of Neogene volcanic rocks in the western Mediterranean. A similar pattern to that already described for the Betic-Rif mountain chain is evident. Volcanism accompanied and postdated the extension episode, the earliest volcanism tending to be calc-alkaline followed by more alkaline (potassium-enriched) volcanism (compare Figures 8 and 10). The earliest documented Neogene volcanism in the western Mediterranean area is calc-alkaline and occurs on both the western side of Sardinia (24–13 Ma), on the margins of the Ligurian Basin (34–20 Ma) [Bello, 1981, Rehault et al., 1985], and in the Valencia Trough. These volcanic rocks are related to the earliest phases of back arc evolution which began as the Gulf of Lions, the Ligurian, and the Valencia Trough basins opened, and as Sardinia and Corsica began their counterclockwise rotation. In the Valencia Trough, considerable quantities of volcanic rock have been encountered in wells and can be inferred from seismic data [Maillard et al., 1992].
Figure 10.
The majority of the volcanic activity was associated with the opening of the basin in the early Miocene and, where sampled, these volcanics are dominantly calc-alkaline in composition. A second phase of alkaline volcanism occurred in the latest Miocene and Plio-Quaternary both in the Valencia Trough (8–1 Ma) and on Sardinia (4–2 Ma). On the northern margin of Africa in the Kabylie and Tell ranges of Algeria and Tunisia, volcanic rocks are mainly calc-alkaline in composition and span the middle Miocene (i.e., both during and after collision of the Kabilies with North Africa [Bellon, 1981]. A small amount of late Miocene alkaline volcanism also occurs.

In the Calabrian Arc-Tyrrhenian Sea-Apennines system, volcanism is notably younger but again clearly linked to the back arc evolution of the Tyrrhenian Sea. The earliest volcanism occurs on the eastern edge of Corsica associated with the oldest northern end of the Tyrrhenian Sea. Subsequently, volcanism in the northern Apennines became younger from east to west and is spatially associated with the eastward migration of extension [Serri et al., 1993]. Apennine volcanic rocks are heterogeneous, but they all generally show an enrichment in potassium. The Aeolian Islands and associated seamounts comprise the volcanic arc associated with the Calabrian subduction zone [Ellam et al., 1989]. This arc has been active for the last 1.3 Ma, and the volcanic rocks exhibit a range of compositions from tholeiitic to calc-alkaline and shoshonitic. A modest K-enriched trend with time can be observed [Beccaluva et al., 1981]. The Tyrrhenian Sea has two basalt-floored sub-basins, the Vaslov and Marsili basins (Figure 10b). Both formed very rapidly between the late Miocene and the Recent. Basaltic crust formation was widespread during the Pliocene in the Vaslov basin, and in the Marsili basin, basaltic crust formation began at ~2 Ma [Kastens et al., 1998, Kastens and Muscle, 1990]. Hence there is clear evidence that the Tyrrhenian basin is younger toward the southeast, as predicted by subduction rollback models. Two tholeiitic basaltic seamounts occur in the southern Tyrrhenian Sea in the area presumed to be underlain by oceanic crust. The composition of basalts both drilled on the seafloor and collected from the seamounts in the Tyrrhenian Sea have a range of chemistries from alkali to calc-alkaline affinities.

The range of composition of Neogene volcanic rocks throughout the western Mediterranean is typical of small back arc basins where arc and extensional volcanism overlap in a diffuse manner and where volcanic arcs sensu stricto have a limited spatial distribution. Striking similarities exist at the eastern and western extremities of the western Mediterranean Sea. The Aeolian volcanic arc is located ~300 km northwest of the current position of the Calabrian subduction zone.

Rotations

Considerable palaeomagnetic data have been collected around the western Mediterranean, and rotations about vertical axes on a variety of scales have been documented. A generalized summary of the main trends and sense of rotations is shown in Figure 10a. Early work in the region concentrated on using palaeomagnetic data to construct polar wandering curves for rigid microplates within the Tethyan suture realm (e.g., the Adria microplate), but it rapidly became apparent that superimposed on the movement of microplates there were local small-scale relative rotations of fault blocks within the deformed regions [e.g., Channell, 1986]. At the largest scale, the onset of the south-southeastern migration of the subduction zone in late Oligocene and early Miocene led to the opening of the Ligurian basin and counterclockwise rotation of Corsica and Sardinia [Montigny et al., 1981]. The asymmetric opening of the Valencia Trough presumably generated clockwise rotation of the Balearic Islands. A recent synthesis of palaeomagnetic data from the Balearic Islands and adjacent areas onshore Iberia by Parés et al. [1992] identify three rotational components. The youngest rotations (20°) on Mallorca and Menorca occur in upper Miocene rocks and are attributed to rotations on listric faults that occurred in the late Miocene and Pliocene After subtracting the effects of late Miocene-Pliocene rotations, the magnetic directions in the Balearic Islands have been rotated clockwise by variable amounts when compared to the Spanish margin. Parés et al [1992] attribute these rotations to differential rotation of thrust sheets in the early Miocene, including a compo-

Figure 10. (a) Generalized summary of sense and magnitude of Neogene-recent rotations about vertical axes from palaeomagnetic data. Timing of rotation given in italics beneath amount of rotation. Dotted line shows original positions of Sardinia, Corsica, and Balearic Islands. Except for Corsica, palaeomagnetic analyses record variable rotations associated with shortening structures. Note that in many cases, data from numerous thrust sheets are summarized at one point. Data are compiled from references discussed in text. Ornament for mountain belts is given in Figure 1. (b) Age and distribution of volcanic rocks in western Mediterranean: calc-alkaline volcanic rocks are distinguished from alkaline volcanic rocks (compiled from Bellon [1981], Hernandez and Bellon [1985], Hernandez et al. [1987], Rehault et al. [1985], Maillard and Mauffret [1993], Ellam et al. [1989], Beccaluva et al. [1981], Serri et al. [1993], and Kastens et al. [1988]).
The Betic and Rif External zones and the Apennines and Sicilian thrust belts all have variable magnitude rotations associated with thrusting [Platman et al., 1993, Allerton et al., 1993, Jackson, 1990, Oldow et al., 1990, Scheepers et al., 1993, Allerton, 1994]. Sense of rotation is reversed on the opposing sides of both the Gibraltar and Calabrian Arcs. In both Sicily and in the Betics, a decrease in the amount of rotation is observed toward the distal parts of the thrust belts [Allerton et al., 1993, Oldow et al., 1990]. We note in passing that detailed palaeomagnetic data from the central Apennines may support the existence in some locations of small Miocene-Recent clockwise rotations [Mattei et al., 1995]. Rotations associated with thrusting indicate oblique convergence and so the edges of the retreating subduction zone are not defined by large-scale wrench faults. A combination of transcurrent motion caused by rollback and shortening caused by continuing plate convergence is clearly an efficient mechanism for generating oblique convergence, especially at the edges of an arcuate subduction zone.

Over the last 10 years, a similar pattern of rotations has been documented in other parts of the Mediterranean Sea. In the Aegean, Kissel and Lay [1988] and Speranza et al. [1995] have measured rotations associated with both thrust and normal faulting. A testable prediction of the rollback model arises from the fact that the collision of the Kabylies with the African margin has been documented in other parts of the Mediterranean has evolved in a fundamentally different way from the Aegean Sea at the other end.

Further work is needed to understand how a subduction zone starts to roll back. It is also unclear why some subduction zones should remain static over long periods of geological time, forming relatively simple linear mountain chains, and others such as those of the Tethyan belt and the western Pacific are very mobile, become arcuate, and lead to the formation of complex orogens (e.g., Figure 7). Short, highly mobile subduction zones that have generated arcuate mountain systems may be difficult to detect in the geological record because they may not form well-developed volcanic arcs. Once the back arc basin closes in the continuing cycle of convergence, there may be little evidence of the former existence of such a subduction zone.

Appendix

The gravitational potential energy calculations shown in Figure 5 were carried out using the approach described by England and Houseman [1988]. We briefly describe a corrected version of their formulation which we have used to calculate Figure 5. All symbols and their values are identical to those used by England and Houseman [1988].

Crust and lithospheric mantle are assumed to have uniform compositions, and the density in each case is given by

\[ \rho = \rho_0 (1 - \alpha T) \]

where \( T \) is the temperature in degrees Celsius and \( \alpha \) is the coefficient of thermal expansion. The variable \( \rho_0 \) is the density of either crust or mantle at 0°C. We assume that the geotherm is linear through a lithosphere of original thickness \( a \) and that the temperature beneath the lithosphere is \( T_1 \).

Let us consider the case of thickened continental lithosphere in isostatic equilibrium with a standard column of continental lithosphere (Figure 5). If the crust and lithosphere are thickened by factors \( f \) and \( \gamma \), respectively, then isostatic balance between the thickened column and the reference column means that \( \Delta e \), the


\[ \Delta \rho_a = (\rho_m - \rho_c) (f_e - f_c) + \frac{\alpha T_1}{2a} \left[ (\rho_m - \rho_c) f_e (f_e - 1) \left( \frac{f_e}{\gamma} \right) + a^2 \rho_m (\gamma - 1) + \rho_m a f_e \left( \frac{f_e}{\gamma} - 1 \right) \right] \]  

(2)

\[ \Delta \Psi, \text{ the difference in gravitational potential energy between the thickened reference lithospheres, can be determined by integrating over the stress distribution as a function of depth:} \]

\[ \Delta \Psi = \int_{\Delta \sigma}^{\infty} \sigma_t dz - \int_{0}^{\infty} \sigma_\tau dz \]  

(3)

Since \( \sigma = \rho g z \), the density profiles, which are determined using Equation (1), have to be integrated.

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