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Notes

Overview of tectonic settings related to the rifting and opening of Mesozoic ocean basins in the Eastern Tethys: Oman, Himalayas and Eastern Mediterranean regions

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Abstract: A combination of geophysical studies and deep-sea drilling have in the past suggested that orthogonally rifted margins fall into two end-members: volcanic-rifted margins (e.g. eastern Greenland) and non-volcanic rifted margins (e.g. Iberia–Newfoundland conjugate). This paper explores the rifted margins of the Eastern Tethys stretching from the Eastern Mediterranean, through Oman to the Himalayas. Rifting in these area was typically pulsed, extending over more than 50 Ma. The timing of final continental breakup ranged from Late Permian in the east, in Oman and the Himalayas, to latest Triassic–earliest Jurassic in many parts of the Eastern Mediterranean (e.g. Antalya in SW Turkey; Pindos in Greece). Rifting in the Himalayas and Oman gave rise to a proximal to a distal ramp geometry with scattered seamounts (continental fragments and atolls) located adjacent to the rifted margin. The Eastern Mediterranean was palaeogeographically varied, and was characterized by a number of mainly elongate continental fragments (tens to several hundreds of kilometres long by tens of kilometres wide). These microcontinents subdivided the Eastern Tethys in the Eastern Mediterranean region into several small ocean basins, which rifted at more or less the same time in latest Triassic–earliest Jurassic time.

All of the rifted margins of the Eastern Tethys are associated with rift-related volcanic rocks. However, with the exception of the Permian Panjal Traps in the Himalayas, the volumes of magma and corresponding thermal doming were less than for the ideal Volcanic-rifted margin (i.e. eastern Greenland). None of the Eastern Tethyan rifted margins show evidence of features characteristic of Non-volcanic rifted margins (e.g. sea-floor serpentinite exhumation), in contrast to the Iberia–Newfoundland conjugate or the Alps. Most of the Eastern Tethyan rifted margins appear to correspond to an ‘intermediate’ type, characterized by pulsed rifting, limited rift volcanism and a narrow continent–ocean transition zone. Such ‘intermediate-type’ rifted margins may remain to be explored in the modern oceans by deep-sea drilling.

There is little evidence to support previous suggestions that the Eastern Tethyan rifts can be considered as back-arc basins above either northward- or southward-dipping subduction zones.

Here it is suggested that the Eastern Tethys documents a fundamentally different type of rifting from either the ‘Volcanic-related’ or ‘Non-volcanic’ intracontinental rifts known from the Alps or the North Atlantic region.

The dominant controls of rifting are seen as the traction of rising asthenosphere on the base of the lithosphere, related deviatoric tensional stresses, inherited and thermally induced weaknesses in the crust, and slab-pull. Specifically, in the Eastern Tethyan region continental breakup was probably triggered by a combination of long-term asthenosphere flow, slab-pull related to subduction beneath Eurasia and melt-induced crustal weakening associated with pulsed rifting or plume effects. Final continental breakup corresponds to a major (‘Cimmerian’) convergent phase along the opposing Eurasia margin, which further supports the role of plate boundary forces in Eastern Tethyan rifting.

The early Mesozoic oceanic basins opened, probably associated with northwestward propagation of a spreading centre through the already weakened periphery of Gondwana, adjacent to less deformable Palaeotethyan oceanic crust.

After a lengthy period of passive margin subsidence, locally punctuated by crustal extension and related volcanism, or plume effects, the rifted margins were finally tectonically emplaced during mid-Mesozoic, late Mesozoic or early Cenozoic time in different areas.

Our knowledge of the processes of continental rifting to form ocean basins is still far from complete. Only a handful of rifted margins have yet been studied by academic drilling and, as yet, there is not one complete study of a conjugate margin.

Marine geophysical studies of rifted margins leave many unanswered questions in the absence of direct sampling by drilling. In this respect the record of former rifted margins emplaced on land in orogenic belts around the world provides a

valuable source of information. This is clearly shown by the Alps, which can be correlated with the rifted margin of the Central North Atlantic (e.g. Manatshal & Bernoulli 1999; Manatshal *et al.* 2007). A difficulty, however, is that rifted margins in orogenic belts have experienced collisional processes: they are typically deformed, metamorphosed and only sparsely preserved. Possibly for this reason the evidence of rifted margins in orogenic belts has received less attention in recent discussions of rift processes than they deserve. Here, we will take a 'virtual fieldtrip' around rifted margins preserved in the Eastern Tethyan region to facilitate comparisons with our knowledge of rifted margins in the Alps and the modern oceans. The discussion draws on the writer's field experience of these areas, but it should be noted that it is not the intention to present new basic data here.

The approach adopted is to focus on the *preserved stratigraphic record* of the continent–ocean transition zone of the rifted margins of the Eastern Tethyan region. This can be achieved through an integration of several lines of evidence: first, stratigraphic data from surface exposure and wells, mainly based on dating using fossils; secondly, sedimentological evidence, for example to shed light on proximal–distal sedimentary relations; thirdly, geochemical evidence that can help determine the tectonic settings of eruption of rift-related igneous rocks; and, fourthly, structural evidence that allows deformed units to be restored to their original relative positions along rifted margins. Additional information (e.g. from

palaeomagnetism, seismic reflection/refraction and subsidence curves) can be integrated where possible.

The purpose of this contribution is, therefore, to provide an overview of rift settings and processes related to the opening of part of the Mesozoic Tethyan Ocean, generally known as the Eastern Tethys. The evidence comes from a wide region encompassing Oman, the Himalayas and the Eastern Mediterranean region (Fig. 1).

The Tethys refers to ocean basins located between Africa and Eurasia that have now more or less entirely closed and been replaced by mountain chains (e.g. the Alps and the Himalayas). The Tethys is divisible into the Western Tethys and the Eastern Tethys. The Western Tethys is mainly located in the Western Mediterranean region; it opened during Jurassic time in concert with the opening of the Central Atlantic and closed during the latest Cretaceous–Early Cenozoic, influenced by the convergence of the African and Eurasian plates, creating the Alps and adjacent mountains (Bernoulli & Jenkyns 1974; Ogg *et al.* 1983; Manatschal *et al.* 2007). By contrast, the Eastern Tethys encompassed a much larger oceanic realm stretching from the Eastern Alps, through the Eastern Mediterranean, the Middle East, to Central Asia and beyond. It shows a long and complex history related to the opening and partial closure of a super ocean to the east, Panthalassa (e.g. Smith & Briden 1977; Scotese 1991). The Eastern Tethys is subdivided into an older 'Palaeotethys', essentially pre-Late Jurassic in age, that is

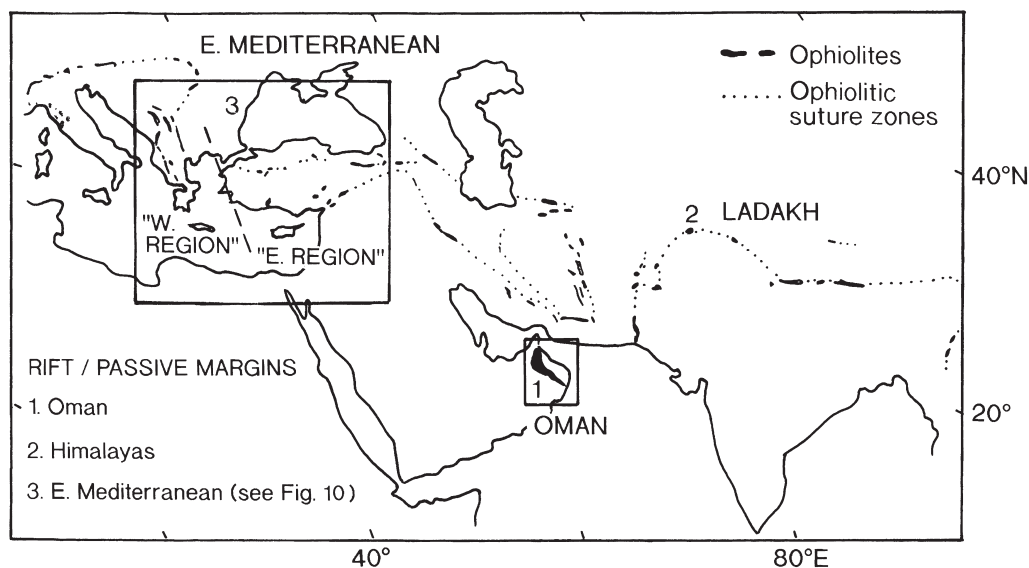


Fig. 1. Outline map of the Eastern Tethyan region showing the main areas discussed in this paper.

widely believed to be rooted near the southern margin of Eurasian (e.g. Pontides of Turkey; Caucasus of Black Sea–Caspian region), and a younger ‘Neotethys’ that is generally seen as being located further south (e.g. Taurides of Turkey; Hellenides of Greece: e.g. Robertson & Dixon 1984; Şengör 1984; Dercourt *et al.* 1986).

There is an on-going largely semantic discussion concerning the use of such terms as ‘Palaeotethys’ and ‘Neotethys’. Some authors have used these terms in different ways over the years, and, unhelpfully, have linked them to their own favoured tectonic models, thus ensuring that the terms are model dependent and inherently untestable. Here, we are concerned only with the younger Mesozoic–Early Cenozoic ‘Neotethys’, and we will refer to different parts of this system, informally, in our discussion of rift settings (e.g. South Tethys; North Tethys, etc.).

Volcanic v. Non-volcanic rifted margins

In recent years rifted continental margins have been divided into ‘Non-volcanic’ and ‘Volcanic’ rifted types, based largely on the study of the Central and North Atlantic regions by the Ocean Drilling Program (Whitmarsh *et al.* 2001). The tectonic architecture of one conjugate Non-volcanic rifted margin setting has been explored, utilizing two legs of drilling on the Iberia margin and, so far, one leg of drilling on the Newfoundland margin (Tucholke *et al.* 2004, 2007). Additional evidence has come from the drilling of a relatively small number of other settings, notably the Equatorial African transform margin (Mascle *et al.* 1987), and from back-arc basins including the Western Mediterranean Tyrrhenian Sea (Kastens *et al.* 1988) and the SW Pacific Woodlark Basin (Taylor *et al.* 1999a, b; Goodliffe & Taylor 2007). Several other rifted margins were studied mainly using geophysical techniques and have been assigned, tentatively either to Non-volcanic or Volcanic-rifted margin types. Non-volcanic rifted margins include the Great Australia Bight–Wilkes Land (Antarctic) conjugate margin (Direen *et al.* 2007), whereas Volcanic-rifted margins include the Falklands–Malvinas and the Brazilian margin segment of the South Atlantic.

It should be noted that the rift geometry and settings of many rifted margins beneath the world ocean have not been studied by academic drilling and remain largely unknown. Geophysical data alone (e.g. gravity, magnetic and seismic reflection studies) cannot distinguish contrasting margin settings with confidence. For example, in the absence of wide-angle seismic refraction data it is impossible to determine whether inferred seaward-dipping

seismic reflectors may simply represent rift-related volcanics or instead true marginal oceanic crust. In some areas (e.g. eastern Brazil margin) little faulting is imaged seismically in the upper crust, which might give the false impression that little extension has taken place when in reality much extension was accommodated by exhumation of lower crust–lithospheric mantle. Such strain localization has implications for the interpretation of ancient rifted margins in orogenic belts where only the stratigraphically higher levels of the rifted margins are typically preserved. The theoretical subsidence curves of Volcanic v. Non-volcanic margins differ markedly (Skogseid & Eldholm 1988), but this is of limited value in practice as few if any volcanic rifted margins have been deeply drilled and accurately dated.

At present, the available information suggests that contrasting Non-volcanic-rifted and Volcanic-rifted settings may characterize different segments of rifted margins, as inferred around Australia (Direen *et al.* 2007). However, to what extent are the Non-volcanic v. Volcanic margin types in reality end members of a spectrum of margins that exhibit more or less evidence of volcanism and fault-controlled extension (i.e. ‘intermediate’ rift types)? Some rifts might show a combination of rift volcanism and seaward-dipping reflectors, as suggested for the Baltimore Trough and Carolina Trough (White & McKenzie 1989), although deep-sea drilling is needed to confirm this. The evidence presented in this paper suggests that such ‘intermediate-type’ rifts characterize most of the rifted margins of the Eastern Tethys.

Recognizing rift settings in orogenic belts

Before proceeding it is useful to consider what are the key features of both of the end-member rifted margin types, as currently understood, that might be preserved in a mountain belt. The essential features in the stratigraphic record that could allow different tectonic settings to be recognized (e.g. rift; accretionary prism; foreland basin) have been identified and termed tectonic facies (Robertson 1994).

In mountain belts generally the proximal (inboard) parts of rifted continental margins are formed on relatively thick continental crust and these are commonly well preserved. By contrast, the distal (outboard) parts of rifted margins are characterized by markedly thinned continental crust and/or marginal oceanic crust, which rarely survive collision and emplacement processes. Where preserved, such distal units typically take the form of imbricate thrust sheets or chaotic units known as *mélange* that underlie large emplaced

ophiolites (e.g. Robertson 2002, 2004). Typically, we are only able to observe the more proximal, stratigraphically higher parts of the former continent–ocean transition. The Alps, where deeper crustal levels are exposed (e.g. intrusive igneous rocks), is an exception.

For an ideal Volcanic-rifted margin the basket of features that could allow this setting to be identified in the field include thick proximal–distal units ranging from subaerial flood basalts (massive flows) and subordinate pyroclastic sediments (e.g. air-fall tuffs), to subaqueous pillow lavas, hyaloclastites and volcanoclastic turbidites (Planke *et al.* 2000; Campbell 2001; Rainbird & Ernst 2001). In some cases such features (e.g. dyke swarms and flood basalts) cover wide areas and could be preserved subaerially after breakup and subsidence (e.g. dyke swarms of the eastern Greenland margin: Karson & Brooks 1999). However, in other areas (e.g. Brazil margin) there is less evidence of a volcanic-rifted margin onshore where mainly basement rocks are exposed. In orogenic belts evidence of magmatism (e.g. sheet flows) could be exposed in foreland areas that underwent less deformation than the main suture zone or escaped later collisional deformation altogether. Domed uplift of at least several kilometres is expected resulting from thermal effects (probably related to magmatic underplating: White & McKenzie 1989) and this would result in an important regional unconformity.

On a subsided Volcanic-rifted margin, an ideal (imaginary) deep hole would pass downwards through deep-sea sediments into thick sequences of subaqueous basalts, then subaerially erupted flood basalts, at least hundreds of metres thick, showing a plume-type geochemical signature. Beneath would be a rifted basement and probably some synrift terrigenous sediments.

By contrast, an ideal Non-volcanic rifted margin would be characterized by crustal thinning, normal faulting and related footwall uplift affecting a wide area (Ziegler & Cloetingh 2004). Lithosphere extension is generally seen as being accommodated by faulting in the upper crust, although there are exceptions (e.g. eastern Brazil margin: Karner *pers. comm.* 2005). Lithosphere stretching may lead to rift magmatism (White & McKenzie 1989; Kendall *et al.* 2005). Rifting is commonly pulsed, lasting over tens of millions of years. Accompanying flexural uplift can lead to deep erosion of a rift shoulder. The structural style varies greatly depending on whether rifting is symmetrical or asymmetrical. Passive margin subsidence ensues after sea-floor spreading begins (Watts 1982; Karner *pers. comm.* 2005). As extension intensifies prior to continental breakup deep crust and upper mantle are exhumed and covered directly by

mid-ocean ridge (MOR)-type basalts, based on evidence from ODP Site 1277, Leg 120 (Tuckolke *et al.* 2007; Robertson *in press*) and the Alps (Manatschal *et al.* 2007).

Thus, an ideal deep-sea core through a Non-volcanic rift basin directly on the locus of final continental breakup would be expected to pass through a deep-sea terrigenous passive margin sequence, into MOR-type basalt (tens to several hundreds of metres) and then into exhumed peridotite (mantle lithosphere).

In principle, it should be relatively simple to differentiate between a distal Volcanic-rifted v. a Non-volcanic rifted margin setting, where preserved in an orogenic belt. However, uncertainties include the following: (1) the ‘ideal’ sections have never been completely drilled and are thus questionable; (2) ‘intermediate’ types of rift may exist, as noted above; (3) the key lower parts of the succession may not be preserved or exposed; and (4) the rift may be asymmetrical with key information missing from the conjugate margin where this is not preserved. These problems can be minimized if the entire rift setting is considered including proximal rift units, which have a higher preservation potential. Also, other evidence can help, notably the geochemistry of the rift-related extrusive rocks, which may help discriminate between, for example, plume- and non-plume-influenced settings.

Nature of the database

With the reader in mind who is not a field geologist it is worth spending a moment to outline the sources of information available for syntheses of rift settings in orogenic belts. Our knowledge of the geology of each of the areas discussed (Himalayas, Oman and Eastern Mediterranean), like many others, stems from more than 100 years of research. Each area needed to be geologically mapped, a process that in areas such as the Himalayas remains incomplete, and the units geologically dated mainly using fossils and radiometric techniques. Each unit has to be characterized using a range of, for example, petrological, sedimentological, geochemical, geophysical and other techniques. Structural geology is critical to allow units to be restored to their predeformational state as far as possible.

As information increases it becomes possible to produce regional, to global, palaeogeographic maps and plate-tectonic models. Commonly there are inherent assumptions in such plate-tectonic syntheses; for example, as to whether most ophiolites (i.e. emplaced oceanic lithosphere) have formed at divergent spreading centres or at convergent margins (above subduction trenches). The role of rift settings in several alternative plate-tectonic

models of the Eastern Tethys is outlined towards the end of this paper.

Briefly, some of the main contributions that have provided the regional database for this study of rift settings are as follows. For Oman, following early work, it was the search for hydrocarbons that resulted in the first comprehensive mapping and regional study (Glennie *et al.* 1973), followed by research triggered by international interest in the Oman ophiolite (Coleman 1981; Lippard *et al.* 1986; Nicolas 1989). Mapping of the country as a whole was completed by a national geological survey under contract to the government of Oman (Le Metour *et al.* 1995). At the same time, academics from a number of countries, including Britain, France, Japan, Switzerland and the USA, provided much additional, more specialist information, and Omani nationals increasingly contributed. For the Himalayas, much of the relevant information came from numerous expeditions over many decades especially by Italian, French and British groups, with Indian, Pakistani and other Earth scientists playing an increasingly important role. In the Eastern Mediterranean the contributions are even more diverse with the nationals of each of the many countries in the region providing much information, supplemented by long-standing research traditions, especially in France, Germany and the UK.

The present summary of rift settings draws on all of the above contributions. Nevertheless, the writer is one of the very few people who has first-hand field experience (over 35 years) of all of the rift settings discussed here (Oman, the Himalayas and the Eastern Mediterranean), and this is the first attempted synthesis of the rift settings from these regions as a whole based on individual fieldwork.

Rifted margins of the Eastern Tethys

In slightly more detail, the Eastern Tethyan region extends from the Eastern Alps, through former Yugoslavia, the Balkans, Greece, Turkey, former southern USSR, the Middle East, Afghanistan, to the Himalayas and beyond (Fig. 1). The northern margin of the Eastern Tethys corresponds to the southern margin of Eurasia (Fig. 2). The southern margin of the Eastern Tethys is effectively the northern margin of the supercontinent, Gondwana and satellite microcontinents. During the Mesozoic, Gondwana episodically fragmented and continental units of various scales and locations drifted away, many towards Eurasia (Şengör *et al.* 1984; Scotese 1991). These fragments range from India (on a continent scale), to Madagascar (Hanken 1994), the Iranian and Tauride microcontinents (Şengör 1984), to smaller fragments (e.g.

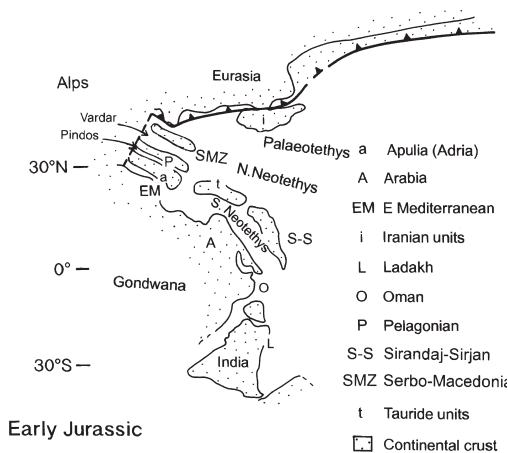


Fig. 2. Simplified palaeogeography of the Eastern Tethyan region. Continental fragments rifted from Gondwana and drifted towards Eurasia. Continental breakup took place during the Late Permian in Oman and the Himalayas, but not until the latest Triassic–earliest Jurassic in the Eastern Mediterranean. See the text for explanation.

Seychelles: Plummer & Belle 1995). Many other continental fragments probably remain to be discovered beneath the world ocean and in orogenic belts (Smith 1999). Continent-scale–microcontinent-sized fragments within the oceans were surrounded by rifted margins of various scales and types. Some of these margin units are now preserved, albeit deformed, in mountain belts.

Exceptionally, a few of the rifted margins of the Eastern Tethys remain as undeformed deep-water successions. The most notable example is the North African–Levant margin of the Central and Eastern Mediterranean, that remains to the south of the Alpine–Mediterranean orogenic front (e.g. Garfunkel 1998, 2004, 2006). Of such undeformed rifted margins, only the NE margin of Australia (Baillie *et al.* 1994) has been explored so far by deep-sea drilling (von Rad *et al.* 1989). Evidence of exotic units in Indonesia suggests that continental fragments were rifted from the Australian continent and drifted northwards, followed by accretion (Pigram & Pannabeau 1984). The possible driving mechanisms of the rifting of such continental fragments are discussed towards the end of this paper.

Most of the presently available useful information concerning rift settings comes from a few areas, notably: (1) the Oman region; (2) the Himalayas; and (3) the Eastern Mediterranean region. Each of these regions shows evidence of rifting that was essentially orthogonal to the adjacent continent. By contrast, several other areas appear to have been characterized by

transform-rifted (or oblique-rifted) settings (e.g. western Pakistan; SE Oman; East Africa), but these will not be considered further here.

Below, for each of the regions (Oman, the Himalayas and the Eastern Mediterranean) the settings of rifting will be summarized, alternative interpretations will be outlined and a restored margin geometry will be proposed. We will also consider if each of these cases corresponds to either a Volcanic-rifted margin or a Non-volcanic rifted margin end member.

Oman rifted margin

Rift-passive margin evolution is excellently exposed in the Oman Mountains, and has probably been studied there in more detail than any other South Tethyan region (Fig. 3). This area exposes the rifted margin of the Arabian continent that was deformed and tectonically emplaced in latest Cretaceous time (Fig. 4). After this emplacement passive margin conditions resumed on the Arabian margin, which has only more recently begun to show the signs of continental collision with the Eurasian margin in the north (i.e. adjacent to the Zagros; Fig. 3). The rift and passive margin units in Oman have thus never experienced 'full' continental collision, in contrast to the Himalayas or the Eastern Mediterranean regions discussed later in the paper. The emplacement of the rift/passive margin units onto the Arabian continental margin (Fig. 4) is widely interpreted as being driven by the collision of the Arabian continental margin with a subduction trench (Glennie *et al.* 1973;

Lippard *et al.* 1986; Robertson 1987). In this model, Tethyan oceanic crust was consumed in a NE-dipping intra-oceanic subduction zone until the trench collided with the margin; the higher levels of the passive margin were then detached and emplaced continentwards. Parts of the leading edge of the passive margin (in the south) were subducted and underwent high pressure–low temperature (HP–LT) blueschist metamorphism, followed by rapid exhumation, by latest Cretaceous time (El-Shazly *et al.* 1990; Searle & Cox 1999; Searle *et al.* 2004).

Prior to the Upper Cretaceous deformation, the Arabian continental margin in Oman can be restored as an orthogonal rift segment, approximately 800 km long. This was bounded by a major transform margin in the south (Masirah margin) and by a smaller transform (or obliquely rifted segment) in the north (Dibba zone) (Cooper 1990; Robertson & Searle 1990). The conjugate margin of the Oman rift is assumed lie within Iran to the north. It has been suggested that a continental fragment was rifted from Oman, the Sanandaj-Sirjan zone (Fig. 2), and that Neotethyan oceanic crust separated this from the Lut continental block to the north (Glennie 2000). However, insufficient work has been carried out in these areas to allow detailed correlation.

Evidence of initial rifting comes from the undeformed continental margin, rather than the over-riding, tectonically transported units. The *in situ* continental margin exhibits a variably complete Late Precambrian–Early Palaeozoic prerift sedimentary sequence that was unconformably overlain by a Late Permian transgressive succession, beginning with thin clastics and passing directly into much thicker shallow-marine carbonates (Glennie *et al.* 1973; Rabu *et al.* 1990). The unconformity was exploited as a Late Cretaceous emplacement related thrust in some areas, whereas in others a spectacular angular unconformity is exposed (Mann & Hanna 1990). The underlying Pre-Permian sequence is deformed, cleaved and folded and was subjected to regional low-grade metamorphism in contrast to the overlying Upper Permian and younger sequence. The deformation and low-grade metamorphism are widely explained in terms of a 'Hercynian' orogenic event affecting the South Tethyan margin, including Oman, during Late Palaeozoic time (Mann & Hanna 1990; Şengör 1990). On the other hand, Stampfli *et al.* (2001) suggest that the absence of preserved sequences between the Late Precambrian–Cambrian? basement and the Late Permian transgression (at least in the Jebel Akhdar; Fig. 3) resulted from rift-related uplift and deep erosion (i.e. rift-shoulder uplift). Rift-related uplift and erosion has indeed taken place, as subsidence curves imply an exhumation

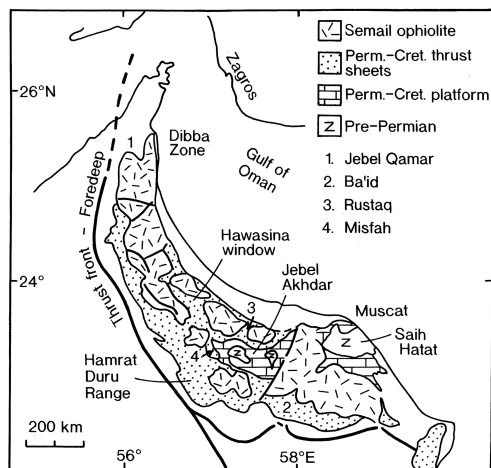


Fig. 3. Sketch map of Oman showing the location of the emplaced rifted margin sediments and volcanics located beneath and to the west of the overriding Semail ophiolite.

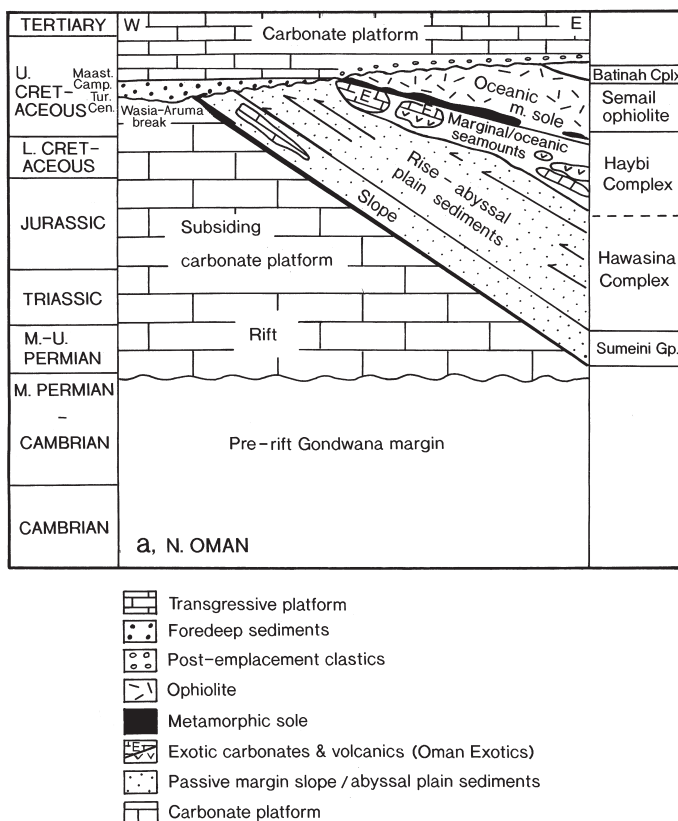


Fig. 4. Simplified tectonostratigraphy showing the relation of the rifted margins sediments and volcanics (Hawasina Complex) beneath the Semail ophiolite, also the relative positions of the Haybi Complex and the Oman Exotics. Some of the Hawasina Complex was also 'bulldozed' ahead of the Semail ophiolite (in the Hamrat-ad-Duru). From Robertson (2004); based on Glennie *et al.* (1973).

of up to 1 km (Stampfli *et al.* 2001). However, a purely extensional setting is unlikely to explain the pervasive deformation and low-grade metamorphism of the pre-rift successions in the central Oman Mountains (i.e. Jebel Akhdar; Saih Hatat), and it is likely that the area experienced compressional deformation prior to rifting and continental breakup during the Permian.

The overlying Permian rift-related sequence passes upwards into thick platform carbonates of Triassic–Late Cretaceous age, several kilometres thick, forming a regional-scale carbonate shelf sequence that is well exposed in the Jebel Akhdar, Saih Hatat and the Musandam Peninsula (Glennie *et al.* 1973; Rabu *et al.* 1990) (Fig. 5a). A foredeep-type sequence (Muti Formation) of Late Cretaceous age at the top the carbonate platform succession records flexural subsidence of the platform related to overthrusting of the margin and ophiolitic units from the Tethyan Ocean (Robertson 1987).

The relatively *in situ* carbonate platform (Jebel Akhdar) is overthrust by a stack of thrust sheets (Fig. 4). In general, higher thrust sheets originated progressively further out into the Tethyan Ocean. However, out-of-sequence thrusting has played a role in some areas so that an accurate reconstruction cannot always be guaranteed simply by unstacking the thrust sheets (Robertson & Searle 1990).

The lowest major thrust sheet is the thickest, most intact, and preserves the Upper Permian and Mesozoic proximal slope successions related to the carbonate platform (Sumeini Group; Fig. 4). This unit was thrust only a relatively short distance onto the *in situ* carbonate platform (several kilometres), and is traditionally described as being 'parautochthonous' (Glennie *et al.* 1973). Successions begin with Upper Permian shelf-to-slope redeposited carbonates that document a rifted carbonate platform edge setting (Watts & Garrison 1986). Above this, the succession records relatively

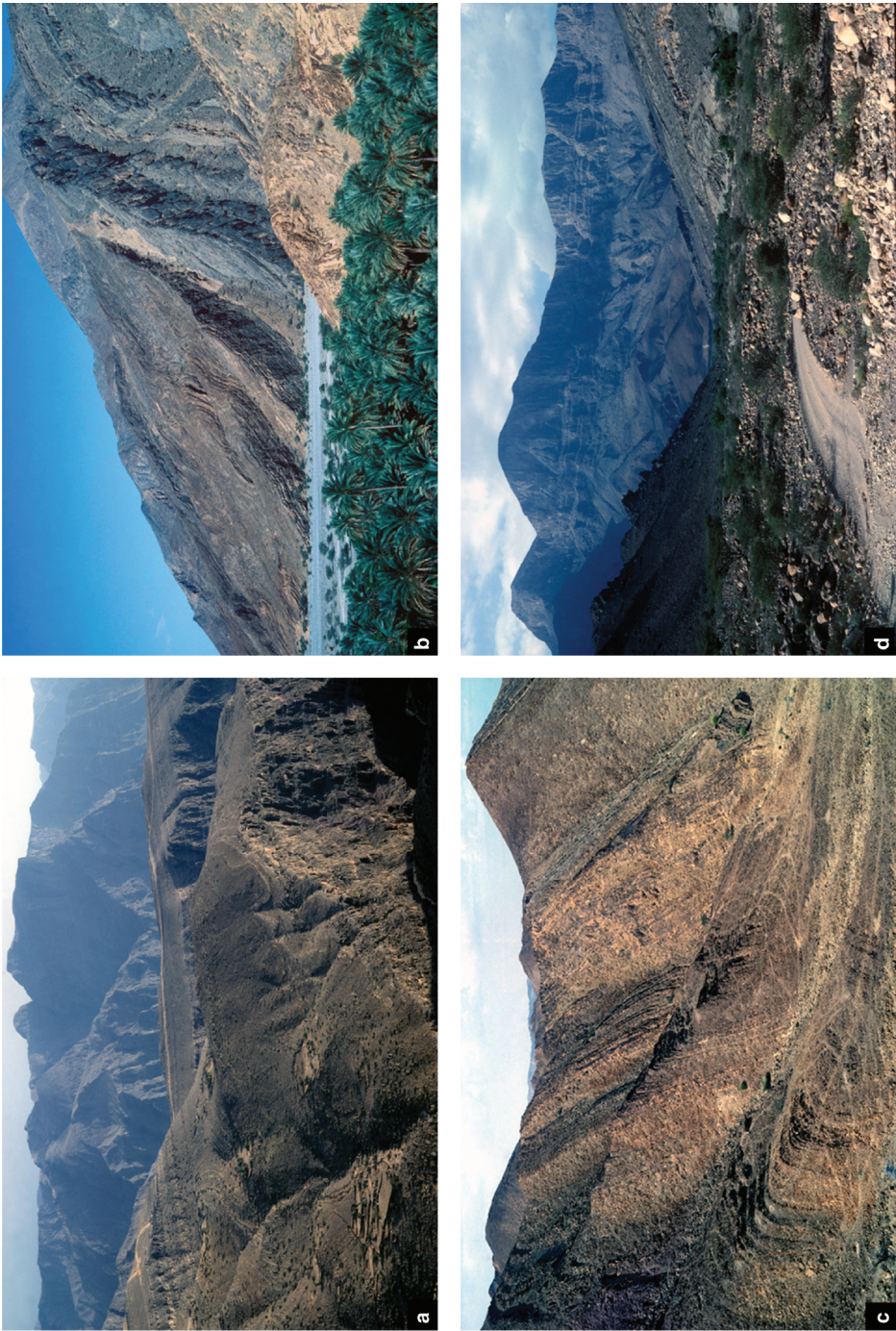


Fig. 5. Field photographs of the Oman rifted margin. (a) Mesozoic carbonate platform bordering the Musandam Peninsula (north of the Dibba Zone), Northern Oman Mountains; (b) proximal passive-margin slope sequence in structurally low thrust sheet (Upper Jurassic–mid-Cretaceous interval); Wadi Mi'adin; Central Oman Mountains; (c) distal slope/basin-plain sequence in structurally higher thrust sheets (Upper Jurassic–Lower Cretaceous); Hamrat ad Duru; west of the Central Oman Mountains; (d) Triassic Oman Exotic structurally overlying the Jebel Akhdar carbonate platform succession. Within this exotic (Jebel Misfah) Triassic alkaline lavas and sills are covered by thick Late Triassic shallow-water carbonates. See the text for discussion.

proximal slope conditions (Fig. 5b) throughout the Mesozoic until Late Cretaceous time (Watts & Garrison 1986) when syntectonic sedimentation appeared (Muti Formation: Robertson 1987). Successions characteristic of the passive margin slope include detached blocks, slump sheets and debris flows, mainly composed of limestones derived from the adjacent carbonate platform. There are also spectacular slope channels infilled with mainly shallow-water carbonates deposited by mass-flow processes.

The slope sequences are typically overlain by the Hawasina Complex, made up of a thick pile (several kilometres) of relatively thin thrust sheets, individually tens to several hundreds of metres thick (Figs 4 & 5c). In general, the stratigraphically thicker thrust sheets exhibit relatively proximal slope facies, whereas the stratigraphically thinner thrust sheets preserve more distal deep-water successions (e.g. Hamrat Duru Range, central Oman Mountains; Fig. 3). However, out-of-sequence thrusting complicates the restored geometry in some areas. The most proximal successions are exposed structurally above the Jebel Akhdar carbonate platform, or the Sumeini Group (e.g. Wadi Guweyza, central Mountains). These sediments are similar to the Sumeini Group slope sequences. The slope and basinal units represent preserved parts of a once contiguous depositional system (Bernoulli & Weissert 1987; Cooper 1990; Blechschmidt *et al.* 2004). In general, the oldest facies preserved in the Hawasina Complex are Early Triassic redeposited limestones, passing upwards into shales and then into thin radiolarian cherts of Middle Triassic age. These are overlain, in turn, by Late Triassic redeposited carbonates and quartzose sandstone turbidites (Zulla and Al Ayn formations). The Early Jurassic is dominated, after a probable hiatus, by shales, radiolarian cherts and subordinate redeposited carbonates and sandstones. The Middle Jurassic is mainly limestone turbidites with abundant redeposited ooids derived from the Jebel Akhdar carbonate shelf (Guwayza Formation). Above this, the Upper Jurassic–Lower Cretaceous interval is characterized by radiolarian chert (Sid'r Formation). Early Cretaceous (Hauterivian–Barremian) time saw a return to the accumulation of redeposited carbonates (Nayid Formation).

The successions within the well-ordered thrust sheets of the Hawasina Complex (as exposed in the Hamrat Duru Range; Fig. 3) are no older than Early Triassic. However, an isolated short intact succession of pillow lavas that is depositionally underlain by Upper Permian radiolarian cherts has been discovered in the Hawasina window (De Wever *et al.* 1990). These Permian volcanics are geochemically transitional between alkali

basalts and mid-ocean ridge basalt (MORB) (Maury *et al.* 2001) and, on this basis, could have erupted in either transitional oceanic or seamount settings. A plume-related setting was suggested mainly on the basis of 'enriched' chemistry (Lapierre *et al.* 2004).

The Hawasina Complex of mainly deep-water sediments is structurally overlain by a mélange known as the Haybi Complex (Searle & Malpas 1980) (Fig. 4). This includes up-to-mountain-sized blocks, the Oman Exotics (Glennie *et al.* 1973) that are critical to the interpretation of the Oman rifted margin. There are two main types of exotic. The first type of exotic includes Permian shallow-marine carbonates, and is represented by the Jebel Qamar exotics in the Northern Oman Mountains (Dibba Zone), and the second is the Ba'id Exotic in the southern Mountains (south of Saih Hatat; Fig. 3). These exotics include thin Permian shallow-water limestones overlain by Triassic limestones. Jebel Qamar is unique, as Late Permian carbonates there are stratigraphically underlain by Lower Palaeozoic siliciclastic sediments (Hudson 1960; Robertson *et al.* 1990; Pillevuit 1993). These sediments are reportedly overlain locally by a thin unit of shallow-marine sediments of Early Carboniferous age, which are associated with alkaline basalts (Niko *et al.* 1996). This is significant as it suggests that extension may have been active as early as the Carboniferous as a precursor to Permian rifting, at least in this area. The Late Permian Jebel Qamar shallow-water carbonates are overlain without a break by Late Triassic neritic carbonates. Further south, the huge Ba'id Exotic (Fig. 3) begins with Upper Permian platform carbonates and passes upwards into Middle–Late Triassic pelagic limestones, including slowly deposited (i.e. condensed) facies (Hallstatt facies) and pink ammonite-rich facies (Ammonitico Rosso). In addition, in the Central Mountains (north of Jebel Akhdar; Fig. 3) the small Rustaq Exotic includes a short succession of Upper Permian pelagic carbonates, underlain by basic volcanics of transitional to MOR type (Pillevuit 1993).

The second and more numerous type of exotic is dominated by large masses of Upper Triassic shallow-water limestone (Fig. 5d). These carbonates are underlain by basic volcanic rocks and overlain, in some areas, by a thin pelagic carbonate succession. The type example of this exotic is Jebel Misfah, which is located along the NW margin of the Jebel Akhdar carbonate platform (Fig. 3), and includes Jebel Kawr, making up a vast exotic mass that is tens of kilometres long. Voluminous lavas and sills beneath the Upper Triassic neritic carbonate successions are geochemically of within-plate type (Lippard *et al.* 1986). Overlying shallow-water carbonates, several thousands of

metres thick, of Late Triassic Carnian–Norian age are covered by ammonite-bearing pelagic carbonate (Ammonitico Rosso) of Early–Middle Jurassic age and then by additional pelagic carbonates of Late Jurassic–Early Cretaceous age (Glennie *et al.* 1973; Pillevuit *et al.* 1997).

The sediments of the deeper-water Hawasina Complex (Fig. 4) locally include reworked limestones of both Late Permian and Late Triassic ages in different areas. For example, the Late Triassic Misfah Exotic is in structural contact with Late Triassic radiolarites and redeposited limestones, including numerous up-to-house-sized blocks of shallow-water carbonate (e.g. along the south margin of the Jebel Akhdar, near Al Hamra). Elsewhere, isolated detached blocks of Late Triassic neritic, or pelagic, carbonate are locally present within the mid part of the stack of thrust sheets forming the Hawasina Complex (Cooper 1990).

The Oman Exotics are in places structurally overlain by the metamorphic sole of the Oman ophiolite, which ranges upwards from greenschist- to amphibolite-facies conditions (Glennie *et al.* 1973; Searle *et al.* 1980; Lippard *et al.* 1986) (Fig. 4). The low-grade metamorphic rocks include alkaline volcanics, meta-cherts and meta-carbonates. The carbonates include recrystallized equivalents of the Oman Exotics in some areas (e.g. Hawasina window), which is notable as it helps to confirm that these exotics formed in a relatively distal position along the margin or within the South Tethyan Ocean.

The various units exposed in different structural units in Oman document rifting related to opening the South Tethys. Several different interpretations have been proposed for the timing and setting of this rifting. The main debatable issues are the palaeogeographical setting of the two different types of Oman exotics and the nature of the crustal basement on which the deep-sea Hawasina Complex formed. A complicating factor is that some of the exotics are located in a high structural position above the distal margin units of the Hawasina Complex (Fig. 4), whereas others, including the huge Misfah Exotics, are located near the base of the thrust stack close to the autochthonous Jebel Akhdar carbonate platform (Fig. 3). There are two main possible explanations of this tectonic setting. First, the Triassic exotics might have formed in originally different palaeogeographic settings on the rifted margin; for example, with the exotics like Jebel Misfah located in a proximal position near the parent platform but other Triassic exotics (e.g. those of the low-grade metamorphic sole) formed more distally. Secondly, the Triassic exotics might all have formed in a relatively distal (oceanward) setting but then have reached their present position in response to structural

complexities during emplacement. Plausibly, the bulldozing of the thinner, more incompetent, deep-sea sediments of the Hawasina Complex took place ahead of the advancing Semail ophiolite nappe leaving the thick relatively competent lavas and carbonates of the Triassic exotics (e.g. Jebel Misfah) in direct contact with the *in situ* Jebel Akhdar carbonate platform beneath. In reality, both factors are likely. The Triassic exotics formed in a range of palaeogeographical setting probably in a relatively outboard (oceanic setting), but they sometimes also experienced complex thrusting which in some cases left exotics in an anomalously low structural position (e.g. Jebel Misfah).

Based on mapping of the entire Oman Mountains, Glennie *et al.* (1973) initially proposed two alternative reconstructions of the rifted margin. In the first model, the ocean began to open during Permian time: the Permian exotics formed part of the rifted margin edge (i.e. contiguous with a larger parent carbonate platform), whereas the Triassic exotics developed within the ocean basin at a later stage (i.e. as seamounts). In a second model the ocean opened during the Triassic, following earlier rifting. In this case, the Permian exotics relate to synrift faulting, whereas the Triassic exotics formed at or near a Triassic spreading ridge.

Both of these models, either of Permian or Triassic continental breakup, were developed by later workers. Members of the Open University (UK) research group favoured Triassic continental breakup, with the Permian exotics recording rifting (i.e. as synrift facies) and the Triassic exotics being seen as igneous seamounts (i.e. post-rift facies) capped by carbonate build-ups within the Tethyan Ocean (Searle & Graham 1982; Lippard *et al.* 1986). Following the discovery of Late Permian radiolarites associated with basaltic volcanics of alkaline–transitional MOR type, others favoured Permian continental breakup following pulses of rifting since Late Carboniferous time (Blendiger *et al.* 1990; Stampfli *et al.* 2001). In this interpretation it was generally accepted that the Triassic exotics (e.g. Jebel Misfah) originated as oceanic seamounts (atolls). However, the Late Permian exotics could be seen, either as part of the rifted margin of a parent Jebel Akhdar carbonate platform (Blendiger 1988; Pillevuit 1993; Pillevuit *et al.* 1997) or as being formed within the Permian ocean after continental breakup (Bernoulli & Weissert 1987; Béchennec *et al.* 1988).

Glennie (2000) has also suggested a two-stage rifting model. A narrow ocean basin opened in the Late Permian ('Neo-Tethys 1') bordered by Permian platform units on both margins. The Qamar and Ba'id exotics represent parts of the eastern margin of this small ocean basin ('Kawr Ridge'). The rift axis then 'jumped' oceanward

during the Triassic, and the NE margin of the initial rift remained as part of the subsequent SW margin of a new ocean basin of Triassic age ('Neo-Tethys') that opened to the NE. In this interpretation the Upper Triassic exotics (e.g. Jebel Misfah) would relate to Triassic rifting of a more easterly, younger basin, rather than as oceanic seamounts as in most other interpretations. However, in the writer's view it is doubtful if the localized Permian exotics really represent the easterly margin of a Permian rifted ocean basin, rather than merely fragments of a single westerly Permian rifted margin; these were left stranded when a Permian ocean, initially narrow, later widened with the genesis of Triassic oceanic crust.

The question of whether or not the Late Permian exotics with shallow-water carbonates originated along the rifted margin of Arabia or within a Permian ocean is central to an understanding of the rift setting in Oman. Here, it is important to note that (unusually) the Hawasina Complex in the area of the Ba'id Exotic includes Permian deep-water sediments (Blendiger 1988), with carbonate turbidites (reef-derived), radiolarites and minor pelagic limestones (with cephalopods) all being present. These deep-water successions have been restored as parts of a south-facing carbonate slope/basin (Blendiger 1988). The most proximal-type slope facies are located in a structurally high position in the local stack of thrust sheets, whereas the more distal facies are located near to the base of the thrust sheet. This suggests that the relatively proximal facies restore to a position within the rift basin, away from the parent carbonate platform (Jebel Akhdar). However, Blendiger (1988) believed that the proximal facies instead originated in a setting close to the Jebel Akhdar (effectively as part of it) and that they have reached their present structural position in response to out-of-sequence thrusting. In this interpretation, the distal facies were first thrust continentwards (towards the SW) over more proximal facies. The thrust sheets were then re-thrust, placing the originally proximal units at the highest level of the thrust stack. These Hawasina thrust sheets structurally overlie the large Ba'id Exotic just to the north, which, as noted above, includes Permian neritic carbonate and Triassic pelagic carbonates. The Ba'id platform unit was assumed to have formed part of the Jebel Akhdar platform during Permian–Mesozoic (Blendiger 1988; Pillevuit 1993). However, structural evidence was not presented to support this interpretation.

In the author's view the Ba'id Exotic is unlikely to represent part of the original Jebel Akhdar carbonate platform for several reasons: First, the Ba'id Exotic is structurally underlain by lithologies including volcanics and radiolarites, typical of the

distal Haybi Complex in other areas, rather than the proximal slope facies (i.e. the Sumeini Group). Secondly, if the Ba'id Exotic was contiguous with the parent Jebel Akhdar platform it is unclear why a Jurassic–Cretaceous neritic succession was not deposited there as well; structural removal is a possibility, but why? It is suggested here that the Ba'id Exotic represents a small, rifted platform unit that was positioned outboard of the main carbonate platform (Jebel Akhdar) and surrounded by oceanic crust of Upper Permian age. The Permian and younger facies described by Blendiger (1988) would have accumulated on the oceanward side of this exotic platform unit, far from the parent Jebel Akhdar carbonate platform.

A similar problem (and possible solution) applies to the Permian Jebel Qamar Exotics in the Northern Oman Mountains (Dibba Zone). A high structural position, and thus a relatively outboard palaeogeographical setting, was inferred by Robertson *et al.* (1990) (Fig. 6), a setting also favoured by Glennie (2000). However, Pillevuit (1993) re-interpreted these exotics as part of the autochthonous Jebel Akhdar carbonate platform. In this interpretation the exotic material has somehow to be relocated from within the autochthon to a structurally high position above the Hawasina-type rocks in this area during Late Cretaceous emplacement. However, no supporting structural evidence for this was provided. If the Permian exotics formed part of the proximal rifted margin of the Jebel Akhdar carbonate platform how were they exhumed from deep within the regionally thick (many kilometres) Permian–Late Cretaceous shallow-water carbonate succession? The Permian exotics lack any known post-Triassic sequences, which would need to have been structurally removed without trace in this interpretation. The author considers it more likely that the Jebel Qamar Exotics formed in a relatively outboard (oceanward) setting, where shallow-water carbonates were never deposited as the exotics had by then subsided into deep water.

In summary, a reconstruction of the rifted Oman margin is shown in Figure 7. The Ba'id and Qamar exotics are seen as rifted continental fragments (detached fault blocks) within Upper Permian oceanic crust (of unknown width). The Rustaq Exotic is seen as representing a small volcanic high within Upper Permian oceanic crust. The Triassic exotics are reconstructed as seamounts (atolls), mainly located in an oceanic setting as in most previous interpretations. In this reconstruction, the Permian Ba'id Exotic subsided and was covered by pelagic carbonate during the Triassic.

The Oman Mountains as a whole are seen as the product of near-orthogonal rifting. However, the Northern Oman Mountains is an exception, as transform or at least oblique rifting has been widely

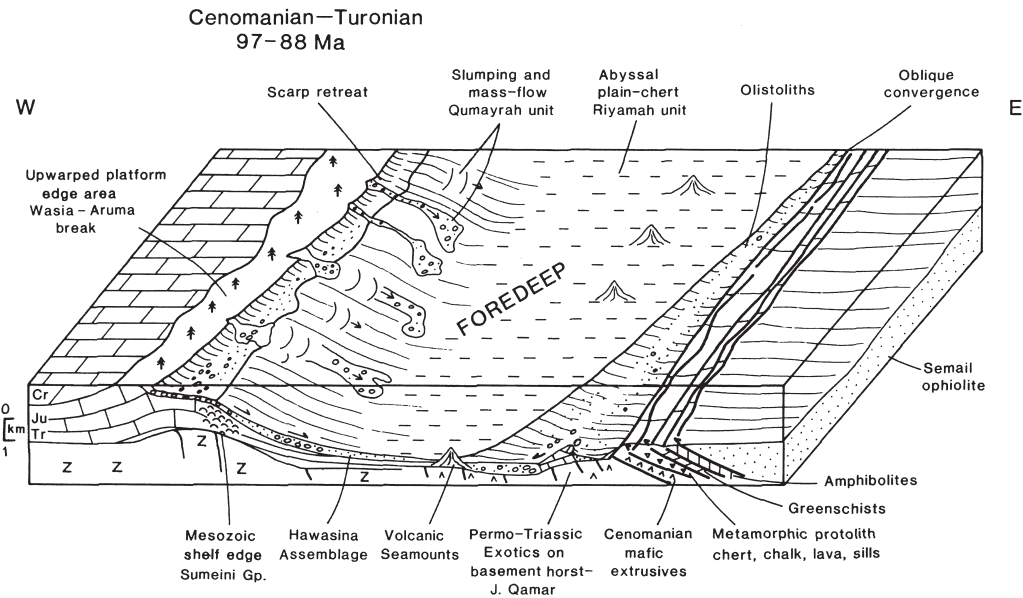


Fig. 6. Relation of the platform margin to the slope successions, based on evidence from the Diba Zone, Northern Oman Mountains (from Robertson *et al.* 1990).

inferred for this area (for references see Robertson & Searle 1990). The unique setting of the Jebel Qamar Exotics in the Northern Mountains (Diba Zone; Fig. 3) with a Lower Palaeozoic prerift basement and both Permian and Triassic shallow-water

carbonates may relate to a setting of transform, or oblique rifting. This rift setting could have triggered the detachment of a fragment of prerift continental basement, which was then ‘parked’ in a relative out-board setting on the margin as the rift basin opened

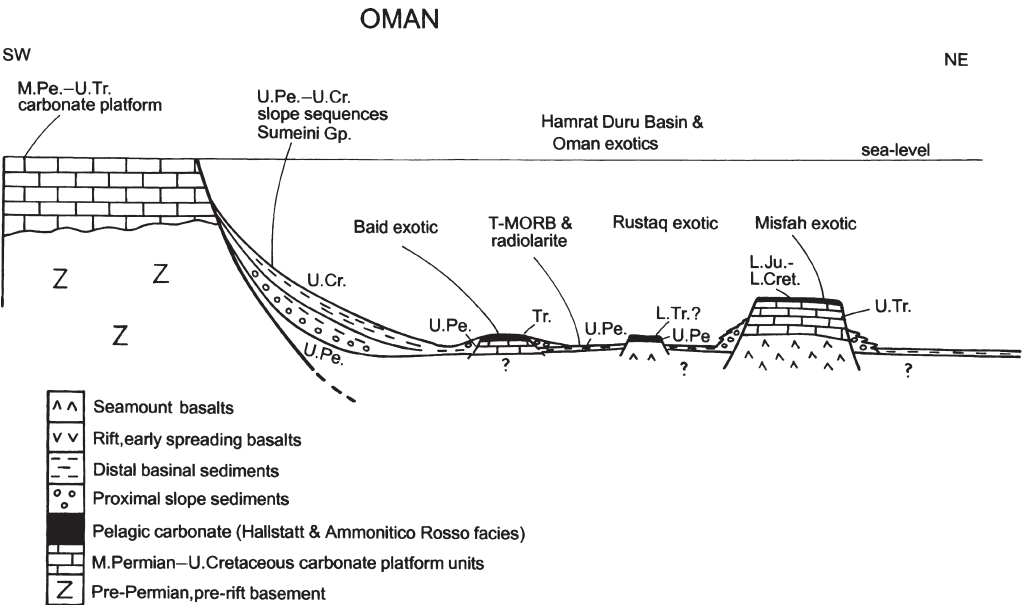


Fig. 7. Restored relations of the Oman passive margin sediments to the Triassic Exotics located on oceanic crust outboard of the Oman rifted margin (from Robertson & Searle 1990).

in the Permian. This could be comparable with the setting of the marginal high of continental crust outboard of the deep-marine Ivory Basin of the Equatorial African margin (Mascle *et al.* 1987). Also, the Hawasina Complex in the Dibba Zone, unlike most other parts of the Oman rifted margin, remained magmatically active during Triassic time possibly reflecting volcanism along a 'leaky transform'. Finally, if transform related, this part of the margin may have remained buoyant and undergoing unusually slow thermal subsidence, which would have allowed platform up-building during the Triassic to keep pace with subsidence of the Jebel Qamar exotic carbonate platform (Robertson *et al.* 1990).

The Oman rifted margin can now be compared with the Volcanic-rifted v. Non-volcanic-rifted margin types. The Oman rifted margin is dissimilar to the ideal Volcanic-rifted margin as voluminous rift magmatism is absent, despite the possible plume-like signature recorded in the Permian volcanics (Lapierre *et al.* 2004). The conjugate margin, inferred to be located in Iran, is deformed and not well documented but does not seem to be of Volcanic-rifted type either (Glennie 2000), although more evidence is needed. However, there is little field evidence to support a correlation with a Non-volcanic rifted margin either. The crust beneath the Hawasina Complex is assumed to be oceanic rather than continental, as otherwise a restored width of several hundreds of kilometres of continental crust needs to have been subducted in latest Cretaceous time, which is unlikely (Robertson & Seale 1990). Fragments of Permian oceanic, or transitional, crust are locally preserved (i.e. in Wadi Hawasina and also as the Rustaq Exotic). The nature of the crust beneath this fragment of earliest formed MOR-basalt is unknown, so it cannot be determined whether or not any (simple shear-type) exhumation affected the Oman margin, as for the Iberia–Newfoundland conjugate. However, there is no known evidence, for example, of contemporaneous (Permian) serpentinite-derived clastics from exhumed asthenosphere within the Permo-Triassic deep-sea sediments of the Hawasina Complex. This contrasts with the Alps, Iberia and Newfoundland rifted margins, where serpentinite-derived clastics were shed from exhumed low-angle fault scarps (e.g. Tucholke *et al.* 2004, 2007). Also, there is no evidence of exhumed deep continental crust in Oman as anticipated if the margin underwent intense stretching and exhumation as along the Iberian rifted margin. If lower continental crust had been exhumed it is likely (but not inevitable) that some of this material would have been incorporated into the thrust stack when the ocean was closing during the latest Cretaceous, but, as yet, this has not been discovered. In summary,

Oman appears to differ somewhat from either of the ideal rift end members.

Himalayan rifted margin

Rifted margin units of Permian and Mesozoic age are well exposed in the Himalayas. Proximal (continentward) settings are well exposed, for example in the Salt Ranges of Pakistan and in the Kashmir further north (Gaetani & Garzanti 1991). More distal settings, located near the rifted platform edge are well exposed in the Zaskar Mountains (Figs 8 & 9a). Continental slope and oceanic units are exposed in a variably deformed and metamorphosed state within the Indus suture zone to the north. This suture zone was formed in response to closure of the South Tethyan Ocean that bordered the Indian continent to the south. Similar rift-related and other Tethyan units are exposed along strike for hundreds of kilometres within the Indus–Tsangpo suture zone, where they are, however, generally less well known.

It is generally agreed that, during the Permian and Triassic, India remained south of the equator, still attached to Gondwana (Fig. 2), and that it finally detached and drifted rapidly northwards during the Late Cretaceous (from Cenomanian time, *c.* 95 Ma, onwards). The progressive closure of the South Tethyan Ocean culminated in diachronous collision of India with Eurasia (e.g. Scotese 1991). Thus, unlike Oman, the effects of this collisional deformation have to be unravelled before the rift setting can be reconstructed.

Continental rifting during the Permian, to the south of the Indus–Tsangpo suture zone (e.g. in Kashmir and parts of the Zaskar Mountains), was marked by the deposition of transgressive clastic sediments above a regional low-angle unconformity, and also by the extensive outpouring of tholeiitic flood basalts, known as the Panjal Traps, during Early Permian time. These volcanics reach hundreds of metres in thickness and thin northwards towards the axial rift zone. Rifting apparently began as early as Early Carboniferous time (Vannay & Spring 1993). Stratigraphic relations suggest rift-related exhumation of more than 2 km (Gaetani & Garzanti 1991). The rift volcanics and sediments are overlain transgressively by a mixed terrigenous–carbonate succession of Late Permian age. Late Triassic–Middle Jurassic time was marked by a predominant accumulation of shallow-marine platform and proximal slope carbonates. Middle Jurassic time was characterized by a regional discordance associated with condensed sedimentation, faulting and the disintegration of the platform edge. The Cretaceous was then associated with deep-water deposition of shales and pelagic carbonates that were accompanied, during the Early

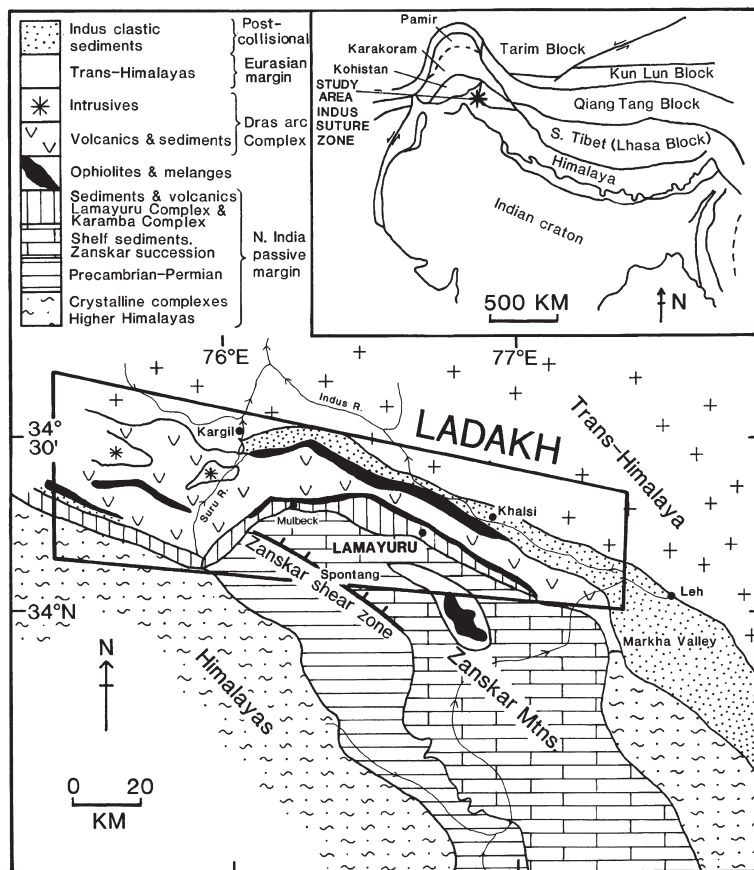


Fig. 8. Outline geological map showing the location of the rifted continental margin sediments exposed in the Zaskar Mountains and the Indus–Tsangpo suture zone in Ladakh, northern India (from Robertson & Degnan 1993).

Cretaceous (late Aptian), by further alkaline volcanism along parts of the rifted margin (Garzanti 1993). The emplacement of oceanic units (e.g. Spontang ophiolite) was followed by collision with Eurasia during Paleocene–Eocene time (*c.* 50 Ma). The timing of emplacement of the ophiolites and other Tethyan oceanic rocks, including the rift-related units discussed here, are reported to have taken place either during the Late Cretaceous as in Oman (Searle *et al.* 1988) or (less probably) not until the Early Cenozoic (Gaetani & Garzanti 1991). This still contentious aspect is outside the scope of this paper.

The development of the deep-water rift and passive margin is best documented in Ladakh (northern India) (Searle *et al.* 1988). The more proximal successions are represented by the regional Zaskar platform, a mixed carbonate–siliciclastic succession that contrasts with the more carbonate-dominated platform succession, as discussed earlier from Oman

(Jebel Akhdar). The Zaskar platform succession is highly deformed by collision-related folding and thrusting (Fig. 9b). Recently, the stack of thrust sheets has been restored (Steck 2003) to reveal a transition over approximately 250 km from an area in the south (Chamba Basin) where basement is directly overlain by rift-related Early Permian volcanics and volcanoclastic sediments (Panjal Traps). This has been interpreted as an area of flexural uplift (rift shoulder), together with an area slightly further north (Tandi syncline), where basement rocks are overlain by a relatively thin and poorly dated Permo-Triassic and possibly younger succession (Steck 2003). Further north, Palaeozoic rocks are overlain by a thin Upper Palaeozoic (Carboniferous–Permian) succession, overlain in turn by a thick shelf succession (Surchu–Baralacha; Kharnag–Tso Morari), which then passed northwards into a slope succession (Nyimaling–Markha), as exposed within the Indus suture zone.

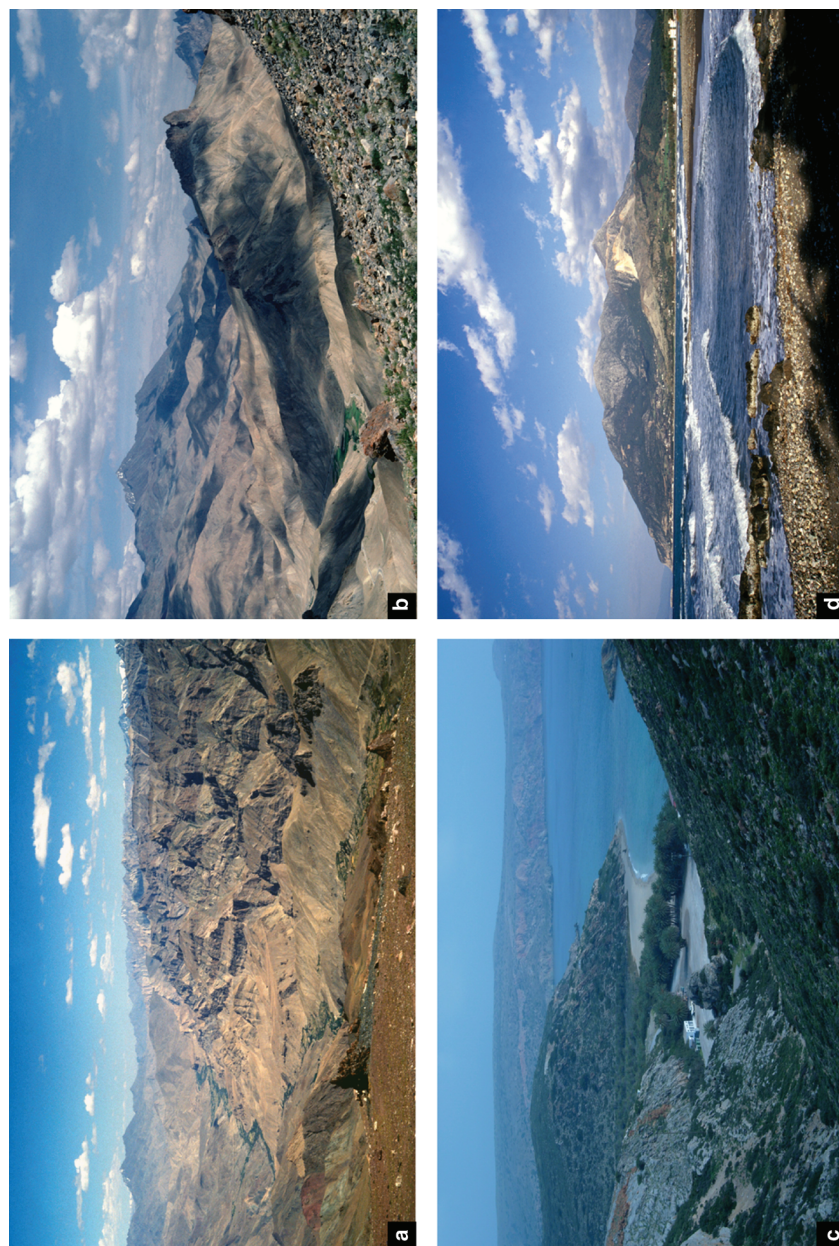


Fig. 9. Field photographs of the Himalayan and Eastern Mediterranean rifted margins. (a) Mesozoic Zanskar platform, backthrust over the Mesozoic deep-water passive margin sequence (Lamayuru-Karamba unit) during the mid-Cenozoic; Sapi-La area; western Ladakh; (b) from left to right: far distance-Zanskar shelf succession; middle distance, right-Mesozoic deep-water passive-margin slope succession (Lamayuru Complex); the pinnacles are former carbonate-filled channels, now vertical; also the high hill, left of centre (dark), represents an accreted oceanic arc unit (Dras arc complex); Namika-La area, east of Mulbeck; (c) Triassic of Phyllite-Quartzite unit, southern Greece; foreground, Triassic shallow-water limestones and sandstones; beyond the bay, Triassic deltaic sandstones; in the distance, downfaulted (overlying) regional-scale thrust sheet of Mesozoic carbonate platform rocks (Tripolitza unit); the assemblage is restored as a Triassic rift and post-rift capping carbonate platform; Vai area, eastern Crete; (d) northern Syria. Jurassic-Cretaceous carbonate platform succession (Jebel Agraia) with overthrust deep-water passive margin slope and ophiolitic material above (wooded on right); taken from near Raas-Al Bassit looking north. See the text for discussion.

The distal deep-water successions of the Indus suture zone are represented by deformed greenschist-metamorphosed facies known as the Lamayuru Complex (Bassoullet *et al.* 1981; Robertson & Degnan 1993; Sinha & Upadhyah 1994) (Fig. 9b, c). The base of the stratigraphic succession of the Lamayuru Complex is tectonic so that the nature of the substratum remains unknown. The oldest preserved units, only exposed locally (e.g. in the Markha Valley, Eastern Ladakh), are Late Permian redeposited carbonates (turbidites) that accumulated in a base-of-slope setting (Stutz 1988). Triassic time was marked by the accumulation of deep-water terrigenous sediments, mainly as deep-water turbidites that contain quartz, mica and lithic grains, ultimately derived from the Precambrian Indian craton to the south, and redeposited carbonates were supplied from the Zaskar shelf. The Early Jurassic sequence was dominated by relatively thin-bedded calciturbidites, rich in ooids derived from the Zaskar carbonate platform. During the early Middle Jurassic (Aalenian) the carbonate platform edge disintegrated in response to a regional pulse of extension, giving rise to exotic blocks of platform carbonate in proximal slope areas ('Mulbeck exotics'). Further east, numerous submarine channels carried large volumes of shallow-water carbonate detritus from the eroding platform into the basin. The Cretaceous record of the Lamayuru Complex is limited by strong deformation. However, it includes sandstones and pelagic limestones (Bassoullet 1978). The stratigraphically lower part of the Lamayuru Complex is intruded by swarms of diabase sills of chemically alkaline composition (Searle 1983; Robertson 2000).

The Lamayuru Complex in the west, in the vicinity of the Suru River (Fig. 8) is in thrust contact with a generally more distal unit of deep-water sedimentary and volcanic rocks known as the Karamba Formation or Karamba Complex (Sutre 1991; Robertson & Sharp 1998). The preserved record of this key unit begins with the deposition of siliciclastic turbidites, subordinate redeposited limestones and basic volcanics of Middle–Late Triassic age. Early–Middle Jurassic time was characterized by the deposition of intercalated radiolarian sediments, siliciclastic turbidites and redeposited shelf-derived calciturbidites. Radiolarites, dated as Middle Jurassic in age (Danelian & Robertson 1997), are exposed as sedimentary intercalations with alkaline basaltic lavas and volcanoclastic sediments, together making up a succession several hundreds of metres thick. The Karamba Formation is interpreted as representing the stratigraphically higher and relatively distal part of the North Indian passive margin (Fig. 10). In this interpretation the Lamayuru Complex

formed the more proximal part of the margin and passed distally (oceanwards) into the Karamba Formation, at least in the west (Suru River–Shegol area), where exposure is sufficient to determine primary relationships. Counterparts of the Karamba Formation are likely to have existed along strike to the east and west, although these are fragmentary, or absent, possibly because they were removed at some stage of the multiphase emplacement history that affected this margin. Basin-plain sediments similar to the Hawasina Complex in Oman (Hamrat Duru Group) are not preserved in Ladakh, again probably owing to tectonic removal during emplacement.

The Karamba Formation alkaline volcanics are definitely interbedded with Indian margin-derived sediments including quartzose terrigenous turbidites and shales (Danelian & Robertson 1997; Robertson & Sharp 1998). This establishes that the Karamba Formation formed an essential part of the North Indian rifted margin and rules out an earlier suggestion that this unit represents an oceanic seamount that was accreted and obducted onto the North Indian margin during closure of Tethys (Sinha & Misra 1994). If it were an oceanic seamount, coarse terrigenous turbidites would be absent, as for the Triassic exotics in Oman, whereas the volcanics are clearly interbedded with terrigenous sediments. The Karamba Formation is also important as it is indicative of alkaline magmatism in a deep-water rifted margin setting during the Middle Jurassic; i.e. long after the inferred continental breakup of this part of the North Indian margin during Late Permian time.

Evidence of more distal, oceanic units is preserved as *mélange* that is mainly exposed above the Zaskar shelf succession, beneath the Spontang ophiolite (Reuber *et al.* 1992; Corfield *et al.* 1997), and is also known locally in the Indus suture zone further north (Bassoullet 1978; Robertson 1998a). Recent mapping has confirmed the existence of a discrete thrust sheet, termed the Photang thrust sheet, beneath the Spontang ophiolite (Corfield *et al.* 1997). In one succession (south of Photaskar), Upper Permian chemically evolved alkaline basalts, several hundreds of metres thick, are interbedded and overlain by red silty carbonates. Pillow lavas of possibly Late Triassic age occur above this, and then Cretaceous pelagic limestones. A second succession begins with Upper Permian limestone capped by a metalliferous crust, and is then overlain by radiolarian mudstones, tuffs and varicoloured limestones. These sediments contain ammonites and conodonts of probable Late Triassic age. Additional (undated) lavas occur above this. Elsewhere in the Photaskar thrust sheet, limestones and marls of probable Cretaceous age overlie Triassic massive lavas and pillow lavas. Another smaller

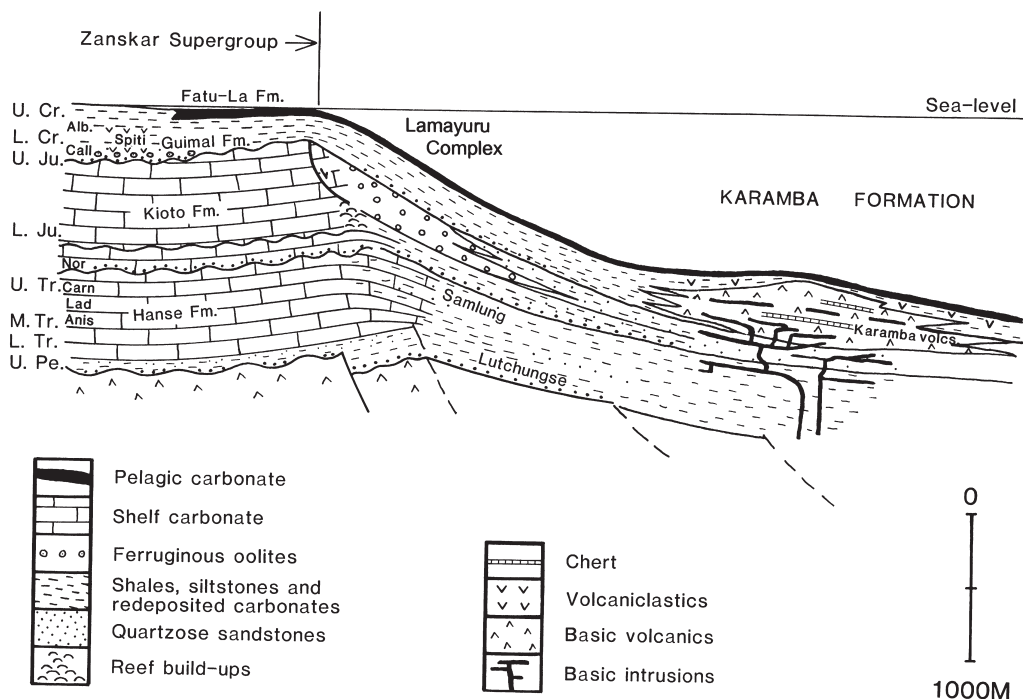


Fig. 10. Restored relations of the passive margin sediments and igneous rocks on the Zanskar shelf and the slope successions of the Lamayuru Complex, Himalayas (from Robertson & Degnan 1993).

'exotic' includes pillow lavas, overlain by Late Triassic pink pelagic limestone (Hallstatt facies). Additional smaller exotics comprise Permian bioclastic limestone, locally overlain by Upper Cretaceous pelagic limestone (Reuber *et al.* 1992). Elsewhere, the mélangé is associated with debris flows made up of limestones of different ages, shed from volcanic edifices that were capped by carbonate rocks of Permian and Triassic age.

In addition, a small but instructive 'exotic' occurs within the Indus suture zone near the Lamayuru monastery (Bassoullet *et al.* 1981; Robertson 1998a). The intact succession within this 'exotic' records Upper Permian shallow-marine carbonate deposition followed by tilting, fissuring and subsidence. The rotated fault block was then covered by ammonite-bearing pink pelagic limestone and volcanoclastic sediment during Early Jurassic time. Fe/Mn crusts accumulated, followed by the eruption of alkaline basalts, probably during Late Triassic time. Much of the history of rifting is therefore encoded within this single exotic block.

Numerous exotics are also exposed within the Indus suture zone in Eastern Ladakh (e.g. Markha Valley), although these have undergone greenschist facies metamorphism and are more recrystallized (Fuchs 1986). Kilometre-sized exotics there

comprise thick piles of within-plate-type basaltic lavas, overlain by shallow-water limestones of Late Triassic age, containing the large bivalves *Megalodonta* (author's unpublished data).

The North Indian margin is interpreted to have undergone a series of extensional rift events from Early Carboniferous time (i.e. pulsed rifting), culminating in continental breakup probably during Late Permian time, with thermal subsidence from Late Permian time onwards (Gaetani & Garzanti 1991). The Early Permian outpouring of tholeiitic continental flood basalts (Panjal Traps) was associated with thermal uplift, perhaps related to magmatic underplating (Stampfli *et al.* 2001). The exotics include Permian shallow-water carbonates (Lamayuru and Photaskar exotics) that are interpreted as remnants of several relatively small carbonate platforms that were positioned within Upper Permian oceanic crust. However, it is unclear whether the basement was composed of either oceanic or continental crust. In addition, Triassic seamounts formed within the South Tethys (e.g. Markha Valley Exotics), much as in Oman.

After the inferred Permian continental breakup and onset of thermal subsidence, the North Indian continental margin experienced several subsequent

pulses of renewed extension and related alkaline volcanism. A strong pulse (or pulses) of extension during Middle Jurassic time resulted in a regional discordance, collapse of the platform edge and the alkaline magmatism within the deep-sea slope succession (Karamba Formation). A pulse of extension is also recorded on the Zaskar shelf in the form of Early Cretaceous alkaline volcanism. These rift pulses can be related to the attempted break-way

of India from Gondwana, which was finally achieved during Late Cretaceous time, followed by passive margin subsidence.

In summary, the inferred setting of the North Indian rifted margin is shown in Figure 11a, b. The repeated pulses of extension and alkaline volcanism for more than 100 Ma differs strongly from the ideal rifted margin, which simply subsides after spreading begins.

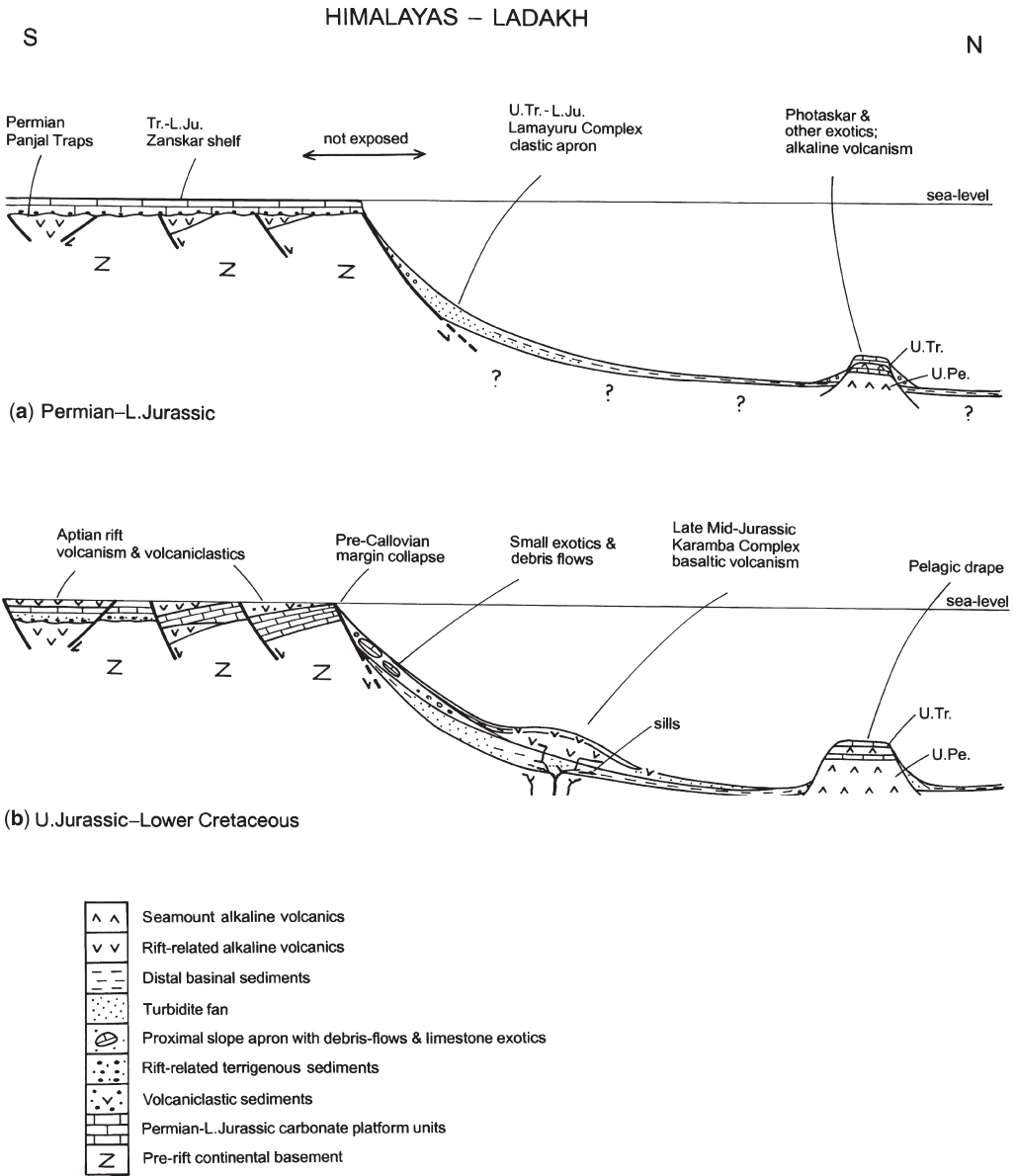


Fig. 11. Summary of the inferred tectonic setting of the north-Indian passive margin. (a) During Permian–Early Jurassic and (b) during Late Jurassic–Early Cretaceous. See the text for discussion.

The North Indian margin can also be compared with the ideal division into Volcanic-rifted and Non-volcanic-rifted margin types. On this basis, the North Indian margin is quite similar to the Volcanic-rifted type. The extensive Lower Permian magmatism (Panjal Traps) of the proximal rifted margin can be related to the effects of a mantle plume, resulting in regional-scale alkaline magmatism (Panjal Traps), thermal uplift and deep erosion. However, the volcanism is focused well south of the axial rift, unlike, for example, the east Greenland margin, which suggests that the main role of plume magmatism may have been to thermally weaken the crust rather than control the locus of final breakup. Unfortunately, the conjugate margin, assumed to lie somewhere further north in Tibet, is not available for study and thus whether or not the rifted margin exhibited a asymmetrical geometry is unknown. On the other hand, there is no field evidence of the features that characterize Non-volcanic rifted margins; for example, exhumed deep crust or mantle, or low-angle detachment faulting, although as in Oman

this could reflect the uncertainties of tectonic emplacement.

Eastern Mediterranean rifted margins

The Eastern Mediterranean region exhibits considerably greater diversity of rift-related settings than either Oman or the Himalayas. There is thus considerable potential if the rift settings can be disentangled from the complicated emplacement history. One of the main features of the Eastern Mediterranean region is that several sutures are present, associated with rift-related and ophiolitic rocks. This opens the possibility that several different types of rifted margin could be present so that individual cases need to be considered separately.

Role of microcontinents and multiple small ocean basins. In the Eastern Mediterranean region rift-related units are exposed in several generally subparallel belts (or zones; Fig. 12). In many current interpretations each of these belts represents the suture of a Mesozoic small ocean basin (e.g.

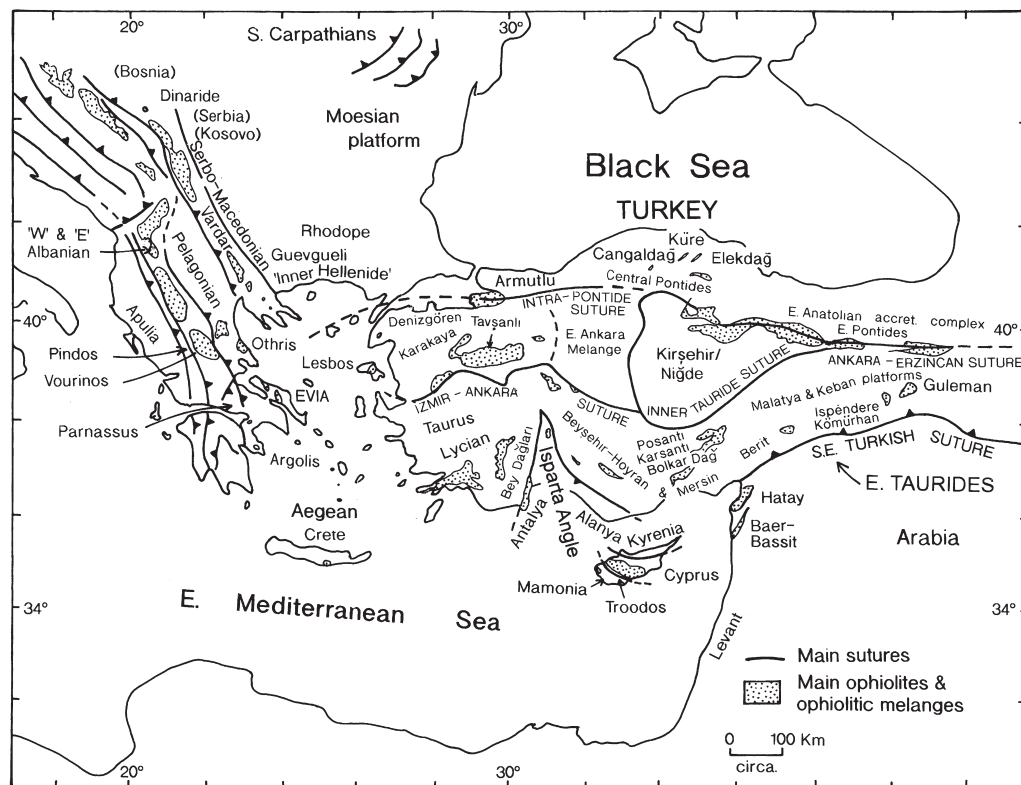


Fig. 12. Outline tectonic map of the Eastern Mediterranean region showing the locations of the main ophiolitic sutures in relation to the rift/passive margin units discussed here.

Robertson & Dixon 1984; Şengör 1984; Smith 1993, 2006; Channel & Kozur 1997; Golonka & Ford 2000; Stampfli *et al.* 2001; Stampfli & Borel 2002; Robertson *et al.* 2004). Many workers, therefore, reconstruct the Eastern Mediterranean region as a number of microcontinents, with individual rifted margins of different scale and local setting, separated by small oceanic basins (Figs 2 & 12). However, there is no agreement as to how such microcontinents should be reconstructed in space and time, or on the rift processes and settings involved. In addition, an earlier and by no means extinct view is that only one main Mesozoic ocean basin existed in the region (until the Late Cretaceous), and that the subparallel zones of rift-related, ophiolitic and other rocks resulted from long-distance overthrusting during the Late Cretaceous–Early Cenozoic closure of Tethys (Aubouin *et al.* 1970; Bernoulli & Laubscher 1972; Ricou *et al.* 1974; Dercourt *et al.* 1986; Stampfli *et al.* 1991; Ricou 1996). Terrane dispersal should also be considered as a means of duplicating continental fragments within a single oceanic basin and this is considered further below.

In one view Permo-Triassic spreading took place in a single Tethyan ocean between Eurasia and Gondwana, while rift basins developed along the southern margin (e.g. Ionian rift basin; Pindos rift basin). The spreading axis then effectively jumped southwards initiating spreading in the Eastern Mediterranean, but not until the Early Cretaceous (Dercourt *et al.* 1986). In most of the more recent models, however, spreading is assumed to have started much earlier along the northern margin of Gondwana (i.e. Permian or Triassic; see below).

In general, throughout many parts of the Eastern Mediterranean region there is evidence of rifting of Gondwana-related units during Late Palaeozoic time, followed by sea-floor spreading during the Mesozoic. This was followed by progressive closure of Tethys in this region as Africa and Eurasia converged during Late Mesozoic–Early Cenozoic time (Livermore & Smith 1984). By contrast, the opposing, northerly continental margin was characterized by the accretion of continental fragments at least from Late Palaeozoic time onwards, most of which ultimately originated from Gondwana and drifted northwards until they collided with Eurasia (Dercourt *et al.* 1986, 2000; Stampfli & Borel 2002).

In the writer's opinion, several large continental fragments and additional smaller ones must have existed in the Eastern Mediterranean region. In the south, Gondwana was bordered by a large microcontinent known as Adria or Apulia. The timing of separation of this fragment from Gondwana is still debated (Late Permian, Triassic or

Jurassic–Cretaceous), but by the late Mesozoic an ocean basin was clearly present in the east (Ionian Sea); this basin continues to be subducted northwards until today, proving that oceanic crust exists in this region (e.g. Dercourt *et al.* 2000). Further north, a large elongate continental fragment throughout former Yugoslavia, Albania and Greece is known as the Pelagonian microcontinent (i.e. the Korabi unit in Albania and the Drina–Ivanjica unit in former Yugoslavia). Further east in the Turkish region the Tauride microcontinent (one or several units) extended from the Aegean Sea to Iran. The Pelagonian and Tauride microcontinental units cannot be traced across the Aegean Sea, and so it is not known whether they were linked in any way. The Tauride microcontinent was itself probably split into several smaller rifted fragments (e.g. eastern and western Tauride). Also, outlying continental units (e.g. Niğde-Kırşehir massif, Turkey; Fig. 12) can be considered either as smaller microcontinents or as promontories of larger continental units. Additional microcontinents existed further east, in the region of Iran (e.g. Lut Block), although these will not be discussed here.

In general, three main Mesozoic basins are recognized, each with their own rifted margin that will be discussed below. The first is the 'Ionian Sea', which separated Gondwana from the Adria (Apulia) microcontinent. The second is the Pindos Ocean in the Balkan area, which separated Adria from the Pelagonian microcontinent to the north. This can be broadly correlated with the 'Southern Neotethys' further east, which separated Gondwana from the Tauride microcontinent. The third is the Vardar Ocean in the Balkan region that can be broadly correlated with the 'Northern Neotethys' further east and which separated the Tauride microcontinent from Eurasia. Several other oceanic elements constituted marginal basins within the Eurasian margin and these will not be discussed here (e.g. Guevgueli, northern Greece; Küre, northern Turkey; see Robertson *et al.* 1996).

Below, each well-documented rifted margin segment associated with each of these three oceanic basins will be discussed individually and then compared to provide a regional picture. It is worth noting that useful insights into rift settings and processes are possible even if agreement as to the regional palaeogeography remains elusive.

Rifted margins of North Africa and the Levant. The most southerly located passive margin in the Eastern Mediterranean region is the rifted margin of North Africa–Levant (Fig. 12). This extends from the Western Mediterranean region (adjacent to Morocco, Algeria and Tunisia), eastwards through the Ionian Sea into the Eastern Mediterranean, to the south of the Sea of Crete (off

LEVANT MARGIN

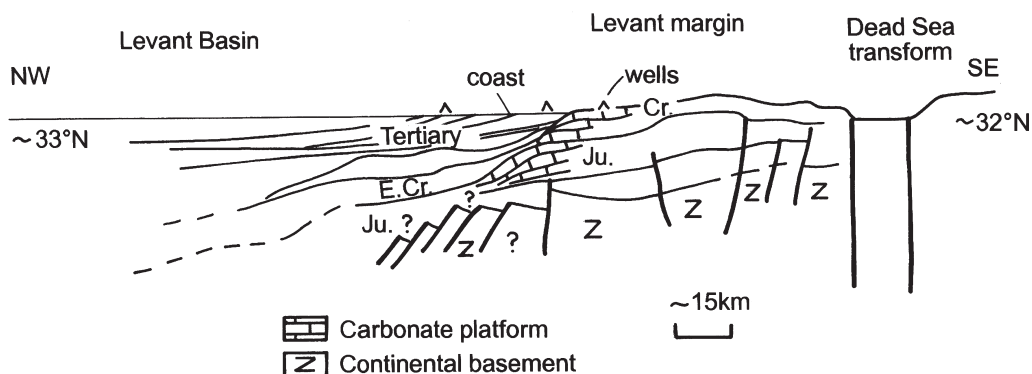


Fig. 13. Morphology and structure of the rifted margin of the Southern Neotethys in the Levant Basin. From Garfunkel (1998).

the coast of Libya) and then into the easternmost Mediterranean (south of Egypt). The margin then takes a nearly right-angle bend to follow the Levant margin, off Israel, Syria and southern Turkey, before bending eastwards around the Arabian margin (northern Syria and Turkey) and then extending southeastwards (through Iran) to link with the Oman margin that was discussed earlier. Limited on-land exposure data, combined with scarce available industry well and seismic data (e.g. Makris *et al.* 1983; Ben-Avraham *et al.* 2002), chart a history of several rifting events affecting the North African margin during Late Palaeozoic–Mesozoic time (Guiraud *et al.* 2001). Important rift events appear to have occurred during the Late Carboniferous and the Triassic, but it is not possible to determine from the evidence of rifting from the North African margin alone (on land) when exactly spreading began in the adjacent South Tethyan oceanic basin: estimates range from Late Permian, to Middle–Late Triassic, to Late Jurassic–Early Cretaceous (see Briand 2000). However, it can be inferred that a limited amount of sea-floor spreading took place in the latest Triassic–earliest Jurassic from evidence from several areas (i.e. Cyprus, Levant; SW Turkey; see below). It has also been argued that extensive Late Jurassic–Early Cretaceous volcanism throughout the region is rift related (Laws & Wilson 1997).

In the east, the Levant margin is orientated north–south with respect to the main North African margin, and has often been referred to as a transform-rifted margin (e.g. Robertson & Dixon 1984; Stampfli *et al.* 2001). However, Garfunkel (1998, 2004) argued that the available evidence is more consistent with near-orthogonal rifting along the Levant margin segment, implying the North

African margin further west (off Egypt) is instead of transform-rifted type. Based on seismic and well evidence, the deep structure of the Levant margin (Fig. 13) does appear to contrast markedly with, for example, the Ivory coast transform-rifted-type margin (Masclé *et al.* 1987). However, it should be noted that transform-rifted margins have so far been little investigated and may show a range of architectures that remain largely unexplored. There is also the evidence of a pulse of alkaline volcanism of Late Jurassic–Early Cretaceous age as documented in wells and from surface exposures (Garfunkel 1998, 2004); some consider this to be rift related (e.g. Laws & Wilson 1997), others plume related (Garfunkel 1998).

In the west, Early Permian–Early Triassic deep-water pelagic facies in western Sicily and eastern Crete include a cosmopolitan warm-water radiolarian fauna, which points to the existence of a deep-water connection between the Eastern Mediterranean and, for example, Oman by this time (Kozur & Krahel 1984; Catalano *et al.* 1991; Kozur 1993). Triassic rift-related rocks in Crete and in the Peloponnese are exemplified by the Phyllite–Quartzite unit, as discussed further below (Fig. 9c). We may, therefore, assume that a significant amount of rifting and subsidence took place along the North African margin prior to the Early Permian, probably beginning in the Late Carboniferous based on the presence of deep-water terrigenous turbidites of this age (e.g. in western Sicily). However, there is no evidence of preserved oceanic crust or of related passive margin sequences in these areas (i.e. in Sicily and Crete) that could indicate the formation of oceanic crust during Late Permian time, in contrast to suggestions made by Stampfli *et al.* (2001, 2003). There are

also different interpretations of whether the Eastern Mediterranean oceanic basin was connected to the Western Tethys through the Ionian Sea (through Sicily) to the Alpine Tethys (Piedmont–Valais Ocean). Many workers inferred such an oceanic connection, albeit narrow (Dercourt *et al.* 1986; Robertson & Grasso 1995; Ricou 1996), whereas others interpreted Adria as a promontory of Gondwana with no marine connection (Robertson & Dixon 1984; Stampfli *et al.* 2001). Much depends on the age and structure of the crust beneath the Ionian Sea. Recent geophysical studies (Catalano *et al.* 2001) suggest that this crust is likely to be of Late Jurassic–Early Cretaceous (rather than, for example, Triassic) age, and this favours a correlation with the Alpine Tethys and the Central North Atlantic.

According to the model of Stampfli *et al.* (2001), spreading began in the Eastern Mediterranean during Late Permian time leading to the opening of a 'Southern Neotethys'. Any younger volcanism, notably Early Jurassic alkaline volcanics (Asher Formation) as cored in a deep well (Atlit-1, near Haifa), represent a reactivation of the passive margin. On the other hand, based on a detailed study of the available seismic, well and surface exposure data, Garfunkel (1998) argued that major rifting dates from the Triassic and that the Early Jurassic volcanics (Asher Formation) associated with thick clastic sediments reflect continental breakup during Early Jurassic time to form a small ocean basin directly to the NW. Subsidence curves for the Palmyra rift to the NE appear to support rifting during Permian time, but there is little evidence of continental breakup of the main Levant margin prior to Early Jurassic time. This conclusion is supported by a recent reinterpretation of seismic and well data from the Levant Sea (south of Cyprus) that suggests that this basin only experienced rifting, and that spreading took place further north in the vicinity of Cyprus during the Early Jurassic (Gardosh & Druckman 2006).

In summary, the North African margin documents pulsed rifting during Palaeozoic–Early Mesozoic time, followed in the east (Levant) by continental breakup in the latest Triassic–earliest Jurassic to form a small ocean basin in the Eastern Mediterranean (Herodotus Basin area). However, breakup was probably delayed until the Late Jurassic in the west (Ionian Sea).

We now go on to consider the evidence of tectonically emplaced units, especially in northern Syria and western Cyprus to evaluate further the timing and setting of rifting of the South Tethyan Ocean bordering Gondwana.

Rifted Arabian margin: northern Syria. From northern Syria through Iran to Oman, the northern

margin of Arabia experienced the emplacement of continental margin and oceanic (ophiolitic) units during latest Cretaceous (Maastrichtian) time (Ricou *et al.* 1974). The driving mechanism of the emplacement is considered to have been trench-margin collision, as noted previously for Oman (Yilmaz 1993). The emplaced rifted margin units are well documented in the Baer–Bassit area of northern Syria, where they take the form of dismembered thrust sheets and mélange beneath over-riding ophiolites (Delaloye & Wagner 1984; Al-Riyami *et al.* 2002b) (Fig. 14). The proximal slope of the passive margin (i.e. Arabian carbonate margin) was concealed by thrusting, in contrast to Oman, and is buried somewhere to the north. More distal margin units are exposed, however, above the autochthonous carbonate platform (Fig. 9d). The succession in dismembered thrust sheets and blocks begins with Upper Triassic alkaline volcanics, interbedded with deep-water radiolarian cherts and pelagic limestones, passing upwards into a mainly siliceous sedimentary sequence; there is then an incoming of calcareous facies in the Early Cretaceous including quartzose sandstone turbidites and pelagic carbonates (Delaune-Mayère 1984; Al-Riyami & Robertson 2002). The distal margin was characterized by one, or several, large volcanic build-ups of alkaline basalt (Tamima unit), interbedded with radiolarite and capped by condensed pelagic limestone of Late Cretaceous (Cenomanian) age. This exotic unit is interpreted as one, or several, seamounts located not far outboard of the passive margin (Al-Riyami & Robertson 2002) (Fig. 15).

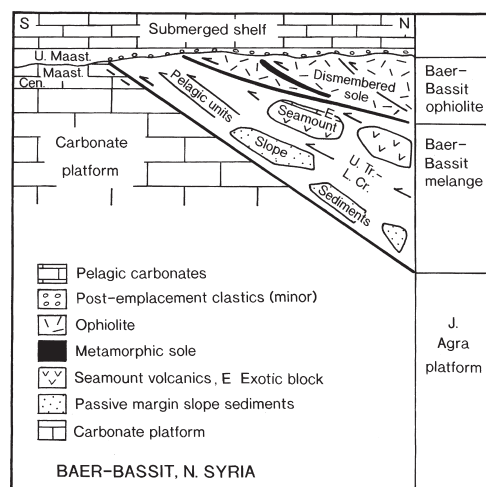


Fig. 14. Simplified tectono-stratigraphic relations of rifted margin and ophiolitic units emplaced from the South Tethyan Ocean onto the Arabian margin in northern Syria (Baer–Bassit region).

N SYRIA

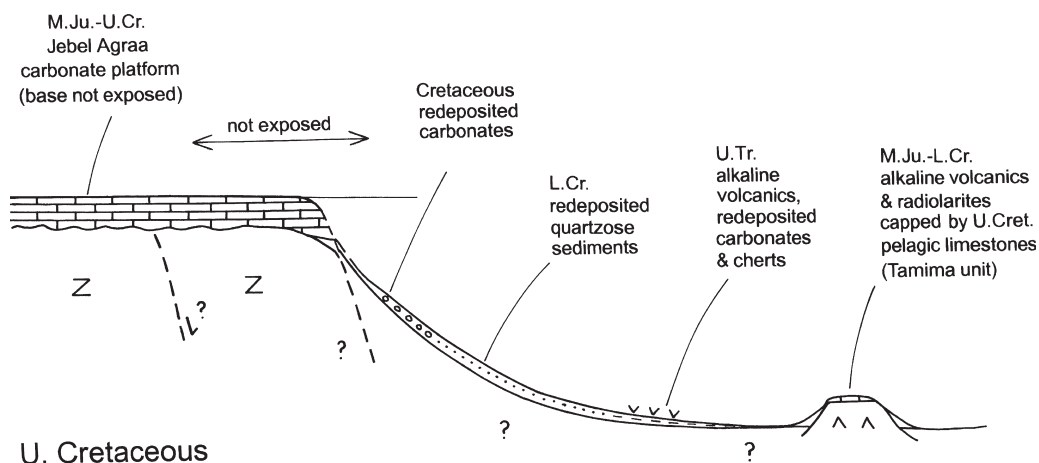


Fig. 15. Restored section across the southern margin of the Southern Neotethys in northern Syria (Baer–Bassit region). Note the age and relative locations of the passive margin and exotic units. See the text for explanation.

The Baer–Bassit area is restored as part of the rifted margin of the Arabian continental margin that stretched all the way to Oman. The Triassic sediments and volcanics there record continental breakup in Late Triassic time (or possibly slightly later). There is no evidence of pre-existing Permian deep-water facies. Seamounts formed adjacent to the margin in Late Jurassic–Early Cretaceous time, as indicated by the age of interbedded radiolarites (Al-Riyami *et al.* 2002a). Some MOR-type basalt is present in the metamorphic sole of the Upper Cretaceous Baer–Bassit ophiolite (Al-Riyami *et al.* 2002b), but this remains undated.

Rifted South Tauride margin: SW Turkey and western Cyprus. During Late Mesozoic–Early Cenozoic time, the oceanic crust of a South Tethyan ocean ('South Neotethys') was mainly subducted northwards (e.g. Dercourt *et al.* 1986, 2000; Yilmaz 1993; Robertson 1998b; Robertson *et al.* 2004). In response, the northerly conjugate margin of the South Tethys was mainly destroyed, possibly as a result of 'subduction erosion'. Exceptionally, in one well-studied area, the Isparta Angle area of SW Turkey (Fig. 12), the rifted margin stepped northwards on a regional scale creating an oceanic embayment. Today, the Isparta Angle has an inverted V shape. However, this is a relatively recent feature (mid-Cenozoic) resulting from crustal rotations during continental collision (Kissel & Poisson 1986). The Isparta Angle restores effectively as a gulf of the South Tethys that divided the Tauride microcontinent into western and eastern

segments. By contrast, Dilek & Rowland (1993) restored the two limbs of the Isparta Angle as the conjugate margins of a small Cretaceous oceanic basin within the Isparta Angle. However, these two limbs are more likely to represent different parts of the south-facing margin of the south Tethys Ocean ('Neotethys'). The setting of a regional-scale northward oceanic embayment eventually favoured the preservation of rifted margin units originally several hundreds of kilometres long, known as the Antalya Complex ('Antalya Nappes'). Deformation and emplacement of the Antalya Complex in latest Cretaceous–Early Cenozoic time was dominated by strike-slip/transpression (Woodcock & Robertson 1982). This setting of oblique convergence in turn favoured the dissection of the former rifted margin into small elongate 'microterranes', in which long vertical successions are preserved separated by steep faults. This contrasts with the mainly low-angle thrust sheets as seen elsewhere (e.g. in Oman and the Himalayas).

The emplaced rifted margin units are exposed around the periphery of the Isparta Angle (Waldron 1984). However, the exposures are most extensive and complete in the classic SW area, SW of Antalya city (Fig. 16) (see Robertson *et al.* 2003 for a recent review of the research in this key area). Proximal successions of the rifted margin, generally lacking volcanic rocks, are preserved in thrust sheets directly structurally above the *in situ* Tauride carbonate platform (Bey Dağları autochthon; Figs 17, column 1 & 18a, b). These units include detached blocks of Late

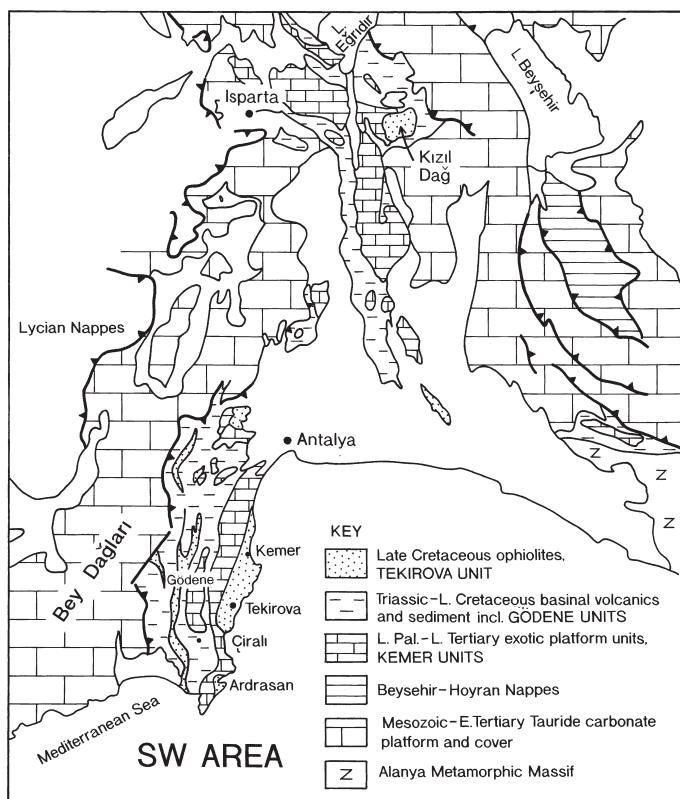


Fig. 16. Outline geological map of the Isparta Angle showing the Antalya Complex and other tectonic units, notably the Mesozoic rift-related basinal volcanics and sediments (from Robertson 1993).

Triassic highly fossiliferous reef limestone in more proximal slope settings. Lower-slope deposits include thick-bedded terrigenous turbidites and redeposited limestone (e.g. channelized conglomerates) (Fig. 17, columns 2 and 3). Distal-type successions (Fig. 17, column 4) are dominated by Late Triassic sandstone turbidites, pelagic limestones, calciturbidites and radiolarian sediments, and include pelagic carbonates. In most areas no basement to these thrust sheets is preserved.

Further east and structurally higher (Gödene unit; Figs 17, column 5 & 18b), thick successions of Upper Triassic (Carnian) pillow lavas, massive lavas and lava breccias are interbedded with minor amounts of deep-water hemipelagic limestone (Halobia Limestone) and radiolarite; detached blocks of shallow-marine coralline limestone are also locally exposed. The lavas and sediments show evidence of syndepositional normal faulting (Fig. 18c). The extrusive sequence, in places, is depositionally overlain by contrasting sediments, exposed on topographical 'highs' and in 'lows'. The 'highs' are depositionally overlain by pink

pelagic carbonate, rich in manganese oxide and ammonites of latest Triassic age (Ammonitico Rosso). The 'lows' locally begin with terrigenous turbidites and quartzose conglomerates, in places containing well-rounded clasts. This establishes that terrigenous rocks were exposed locally within the axial rift zone. Above this come Late Triassic hemipelagic Halobia limestones, with carbonate derived from adjacent carbonate platforms, and above this deep-water mainly siliceous and radiolarian sediments of Jurassic–Early Cretaceous age (Robertson & Woodcock 1981) (Fig. 17, column 5). This unit (Gödene Zone) is interpreted as volcanic seamounts within the continent–ocean transition zone.

Eastwards again, the volcanic–sedimentary successions of the Gödene zone are overthrust by, or in high-angle contact with, thick units of Mesozoic platform carbonates (Kemer units or 'Upper Antalya Nappe'); these sediments directly overlie a basement of Lower Palaeozoic–Middle Triassic prerift–synrift sedimentary rocks (Figs 17, column 6 & 18d). In places, marginal facies that

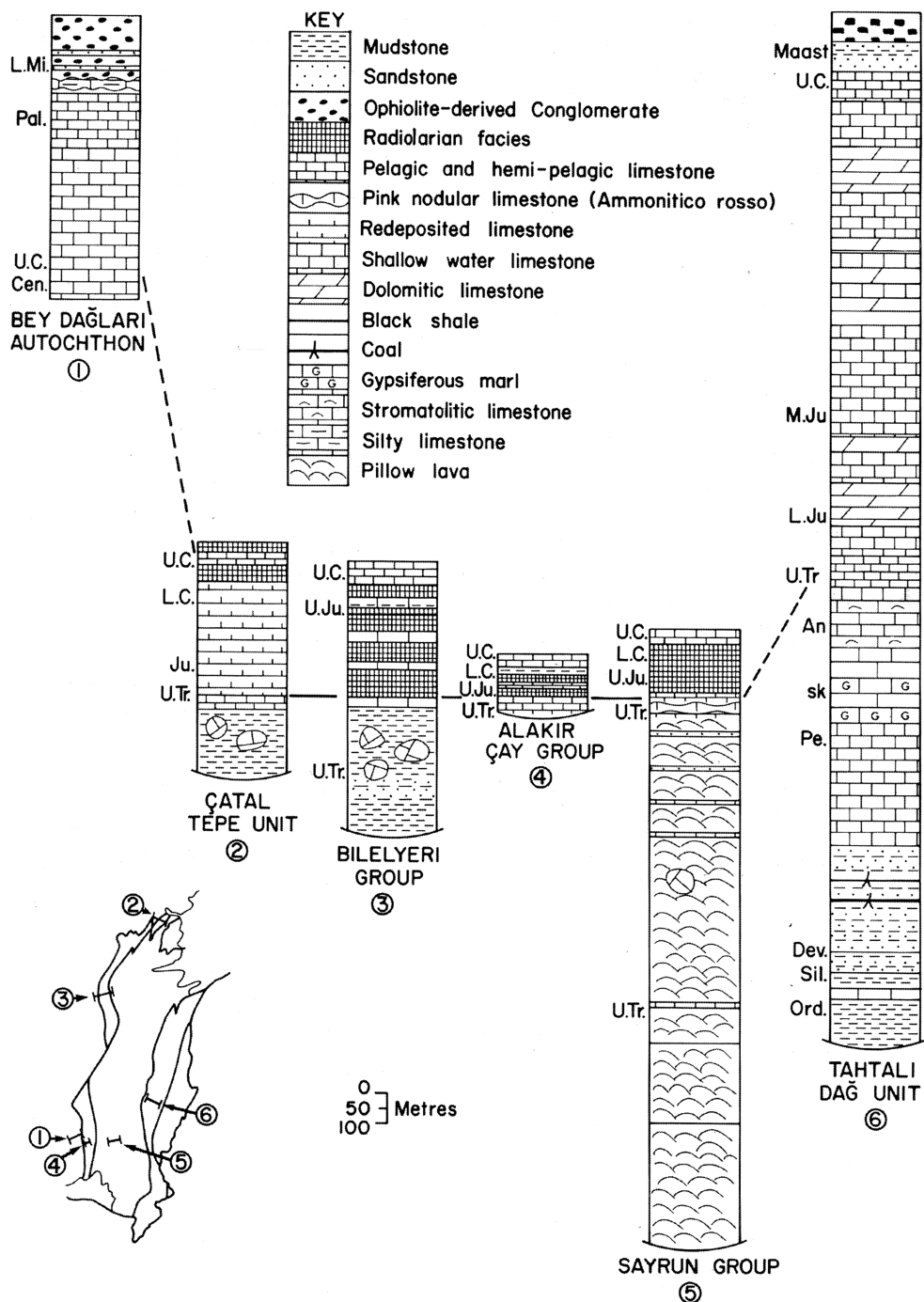


Fig. 17. The main sedimentary and volcanic successions exposed on the southwestern part of the Isparta Angle (SW segment). Note the parent carbonate platform (column 1), the rifted margin slope units (columns 2–4), the Mesozoic volcanic-related units (Gödene Zone) (column 5) and the Mesozoic carbonate platform floored by pririft continental basement (column 6; Kemer Zone). The Upper Cretaceous Tekirova ophiolite is exposed further east along the coast (see Fig. 16 for an outline tectonic map).

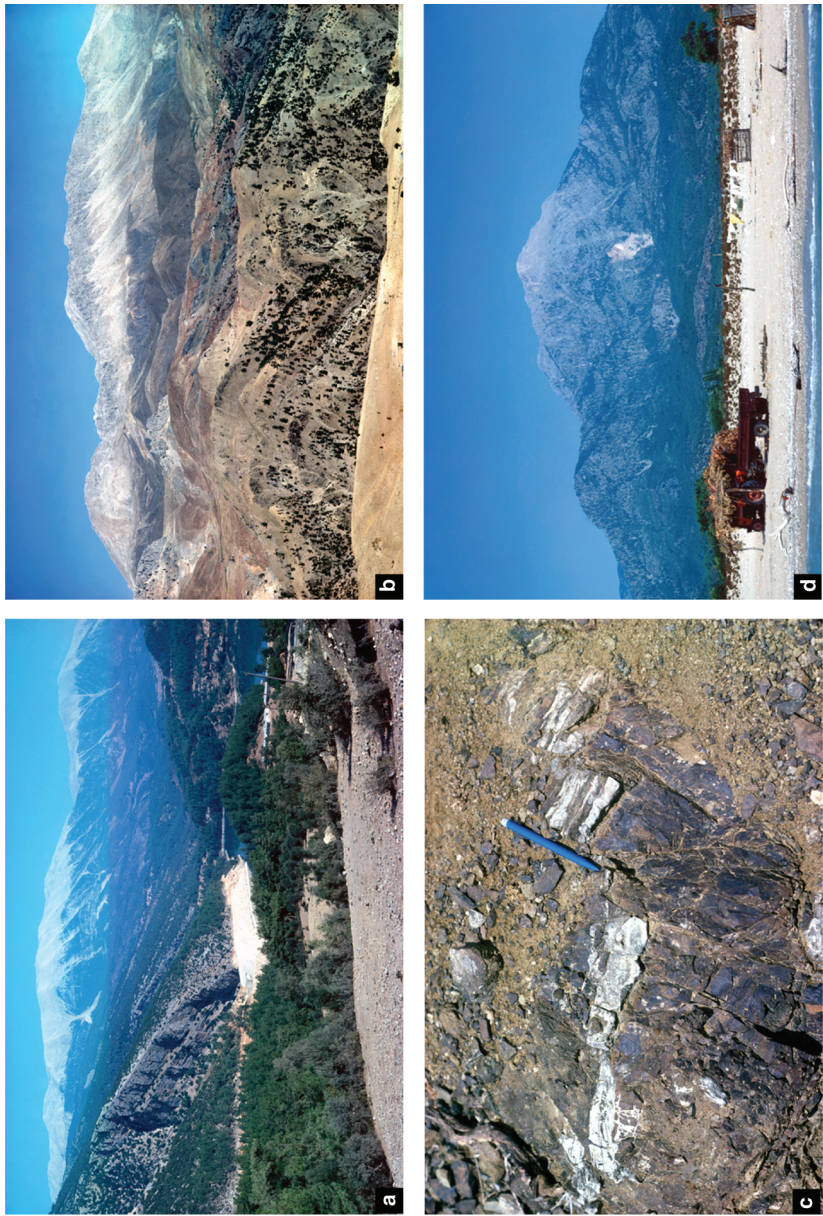


Fig. 18. Field photographs of rifted margin units in the Antalya Complex (Isparta Angle region, SW Turkey). (a) Mesozoic carbonate platform (Bey Dağları) dipping eastwards beneath emplaced Triassic–Cretaceous deep-water sedimentary units (wooded; Kumluca zone); (b) view from the autochthonous Mesozoic carbonate platform (Bey Dağları) across folded and imbricated Triassic–Cretaceous deep-water sedimentary units (Kumluca zone), then across Late Triassic deep-water transitional basalts (Gödene Zone), with Mesozoic–Triassic–Cretaceous shallow-water carbonates above (Çalbalı Dağı; Kemer zone). The contact between the lavas and the limestones here is tectonic; the limestones are restored as exotics constructed on ‘highs’ within the continent–ocean transition zone. Ocean crust (Tekirova Zone) lies over the skyline to the east, beyond which is Antalya Bay; (c) small normal faults within Upper Triassic (Carnian) subaqueous massive lava flows; erupted within the continent–ocean transition zone. The interbedded carbonate (peri-platform) ooze was shed from adjacent exotic carbonate platforms; Çalbalı Dağı; Kemer zone; (d) unbroken Palaeozoic–Upper Cretaceous sedimentary succession (prerift; synrift and post-rift); a rifted fragment of continental crust is capped by shallow-water carbonates (Tahtalı Dağı; view west from Antalya Bay). See the text for discussion.

were derived from the Triassic shallow-water carbonate platforms overlies, or interfinger with, volcanic-sedimentary successions (Robertson & Woodcock 1982).

Further east again, the Kemer units are in moderate- to high-angle fault contact with Upper Cretaceous ophiolitic rocks of the Tekirova unit, as exposed along the Mediterranean coast (Juteau 1975) (Fig. 16). The emplacement of these ophiolitic rocks onto the margin has been explained by dominantly strike-slip processes during Late Cretaceous–Early Cenozoic time (Woodcock & Robertson 1982).

Taken together, the successions restore as a transition from a proximal to a distal rifted margin from west to east in the SW Antalya area (in present-day co-ordinates; Fig. 19). Initial rift volcanism of Early–Middle Triassic age was mainly tuffaceous and was accompanied by evidence of subsidence and extensional faulting. This was followed by continental breakup and was accompanied by extensive outpouring of volcanic rocks (up to hundreds of metres thick) during Late Triassic (Carnian) time. These volcanic rocks are of alkaline–transitional MOR basalt type and were erupted at the same time as the accumulation of peri-platform ooze (Halobia Limestone) and ribbon radiolarites in deep water (Robertson & Waldron 1990). The presence of locally overlying lenses of turbiditic sandstones and redeposited terrigenous conglomerates (e.g. mass-flow-type conglomerates) is important in establishing that rift-related igneous crust formed along the continental margin rather than out in the ocean (Fig. 19). An origin as emplaced open-ocean igneous seamounts, for example, similar to the Triassic exotics in Oman (which lack interbedded terrigenous sediment) can be ruled out.

The Kemer units are interpreted as one, or probably several, rifted continental fragments that were overlain by Late Triassic shallow-water carbonate build-ups. Their setting was similar to the Jebel Qamar Exotics in northern Oman, as discussed earlier. However, the Antalya off-margin platforms continued to be the sites of shallow-water deposition and were not covered by pelagic carbonates until Late Cretaceous time. Several factors may contribute to this contrasting subsidence history. First, the original rifting was probably oblique, which could have favoured transform-type rifting of continental fragments – as, for example, along the Dibba Zone of northern Oman and the modern Gulf of Guinea transform margin (Mascle *et al.* 1987). Secondly, it is possible that only a small amount of oceanic crust formed during Late Triassic–Early Jurassic time within the adjacent South Tethys and thus tectonic subsidence was limited. Thirdly, the marginal platforms (i.e.

Kemer units) were located within a semi-isolated re-entrant of the South Tethyan oceanic basin, comparable to the setting, for example of the modern Caribbean (Waldron 1984; Robertson 1993), which may have favoured the persistence of shallow-water carbonate accumulation on marginal highs. The net effect was that tectonic and thermal subsidence was sufficiently limited to allow carbonate platform up-building to be maintained from Late Triassic to mid-Cretaceous time. This was followed by regional submergence and a transition to pelagic carbonate deposition during the Late Cretaceous.

The off-margin Kemer units and adjacent ophiolitic units were strongly deformed during latest Cretaceous–Early Cenozoic time by a combination of thrusting–strike-slip–transpressional processes. However, platform deposition (mainly hemipelagic) continued on the parent Bey Dağları carbonate platform into Early Cenozoic time. After regional deformation and emplacement ended (Middle Eocene) the oceanic basin in the Isparta Angle remained partially open to a surviving remnant of the South Tethyan oceanic basin that is, today, the Eastern Mediterranean Sea.

In summary, the importance of the Antalya Complex (SW area in particular) for the understanding of rift settings is that the continent–ocean transition is unusually intact and well exposed there, especially volcanics and sediments formed during the transition to spreading oceanic crust of Late Triassic age.

Additional important evidence of the northern, rifted passive margin of a South Tethyan Ocean ('South Neotethys') comes from the Mesozoic rocks of west Cyprus known as the Mamonia Complex. The Mesozoic allochthonous rocks occur in the west of the island and also in the south (i.e. as the Moni Mélangé; Robertson 1977). It might, therefore, be expected that these rocks relate to the southern margin of the South Tethys. However, this seems unlikely for several reasons. First, the leading edge of the southern margin, represented by the Eratosthenes Seamount, did not begin to collide with southern Cyprus until the Plio-Quaternary; during the late Mesozoic this unit was still located well to south, as shown by the results of deep-sea drilling (Leg 160; Robertson 1998b). Secondly, the units exposed in the Mamonia Complex (Mamonia nappes of Lapierre 1972) can be lithologically correlated with the SW area of the Antalya Complex, discussed above (Robertson & Woodcock 1982). Thirdly, the Troodos ophiolite, Cyprus and probably the adjacent Mamonia Complex underwent 90° counter-clockwise rotation during latest Cretaceous–Early Cenozoic time (Moores & Vine 1971; Clube & Robertson 1986; Morris 1996). Thus, the Mamonia rocks restore to an originally northerly position. Taken together,

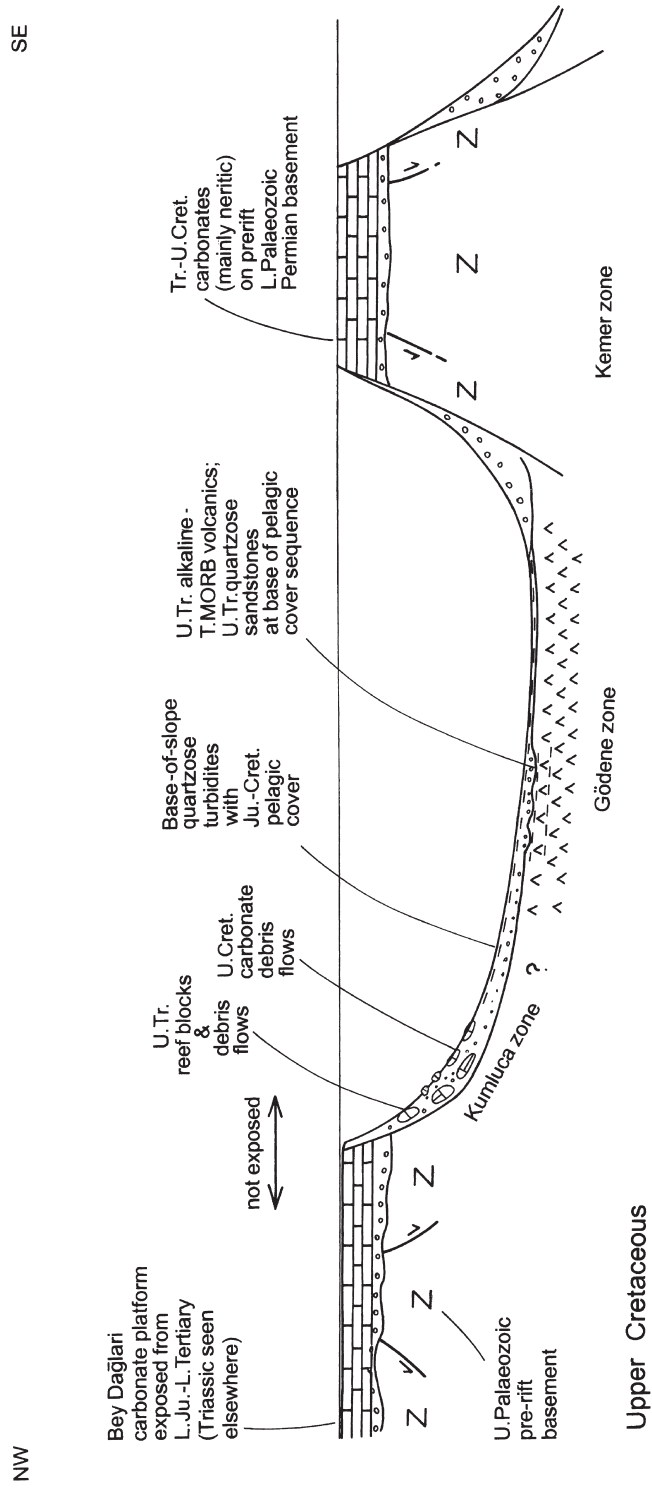


Fig. 19. Restored relations of the passive margin sediments and igneous rocks on the shelf and the slope of the SW Isparta Angle (i.e. SW segment of the Antalya Complex). Carbonate platform units there are underlain by pre-rift crust ('Upper Antalya Nappe') restored in more outboard settings.

LATE TRIASSIC

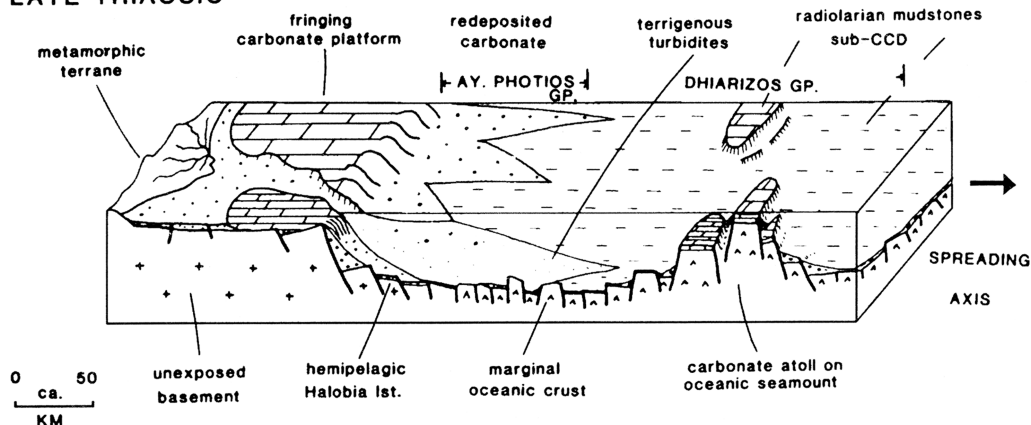


Fig. 20. Reconstruction of the Upper Triassic tectonic setting of the Mamonia Complex, western Cyprus. Note that the parent carbonate platform is not exposed in Cyprus (from Robertson 1990).

these factors favour a north margin-related origin of the Mamonia Complex, together with the Antalya Complex in the Taurides, southern Turkey.

In west Cyprus only the more distal (oceanward) rifted margin successions of an original continent–ocean transition are exposed. Dismembered sedimentary successions, termed the Ayios Photios Group, restore as Upper Triassic deep-water terrigenous turbidites, passing into Jurassic mixed calcareous–siliciclastic and siliceous sequences that accumulated in a deep-water base-of-slope setting (Fig. 20). Early Cretaceous time saw deposition of the thick-bedded quartzose sandstones (Akamas Sandstone) that were probably redeposited from a coastal, aeolian setting by gravity-flow processes. Consequently, the Mamonia sedimentary successions (Ayios Photios Group) restores as the slope/lower slope of a Triassic rifted margin (Robertson & Woodcock 1979) (Fig. 20).

An originally more distal (oceanward) unit, known as the Dhiarizos Group, comprises strongly deformed Late Triassic basalts of MOR type (Malpas *et al.* 1992), interbedded with radiolarites and locally overlain by ammonite-bearing pelagic limestones. Local successions continue upwards into relatively thin-bedded terrigenous and calcareous turbidites, shales and pelagic limestones (Mavrokolympus unit; Swarbrick & Robertson 1980). Consequently, as for the Antaya Complex (Gödene Zone), these igneous rocks were formed adjacent to the rifted margin and not within an open ocean far from a continent. In addition, there are also scattered blocks of alkaline volcanics associated with Upper Triassic reefal carbonate build-ups (Lapierre 1972). These lavas are interpreted as seamounts in an oceanic setting (albeit

marginal) comparable to that of the Triassic Oman Exotics (Figs 21 & 22). In summary, the chief importance of the Mamonia igneous rocks is that they provide evidence of Triassic MOR-type basalt with the South Tethys ('South Neotethys'). This is inconsistent with some tectonic models that suggest that no spreading occurred in this area until the Cretaceous (Dercourt *et al.* 1986; Dilek & Rowland 1993).

Taking the evidence from both the southerly and the northerly rifted margin units, discussed above, the South Tethys in the Eastern Mediterranean region can be restored as an east–west-trending Mesozoic small oceanic basin. Pulsed rifting on a regional scale (Late Carboniferous–Late Permian–Early/Middle Triassic) was followed by continental breakup during latest Triassic–earliest Jurassic time, associated with extensive alkaline–transitional MOR-type volcanism in different areas.

These South Tethyan settings differ from both the ideal Volcanic- and Non-volcanic rifted margin types. There is little evidence of regional-scale voluminous plume-type volcanism with the implied thermal uplift and deep erosion. However, the chemistry of the rift volcanics is similar to some plume-type volcanics (Dixon & Robertson 1999), while perhaps not uniquely fingerprinting a plume origin. In addition, there is little in common with the Non-volcanic type margins, as rift-related volcanics (hundreds of metres thick) are present (e.g. Antalya Complex) and there is no exposed evidence of simple shear-type detachment faulting or deep-crust exhumation. An 'intermediate' type of rift setting is again implied with some, but not profuse, rift volcanism and some extension-related faulting.



Fig. 21. Field photographs of rifted margin units in Cyprus, Greece, Albania and Bosnia. (a) Cyprus. Exotic blocks of Upper Triassic reefal limestone associated with Triassic alkaline lava (view west); interpreted as build-ups on small seamounts during initial sea-floor spreading; Petra tou Romiou (Aphrodite's rock); SW Cyprus; (b) Greece. Several imbricate thrust sheets of Triassic–Early Cenozoic deep-water sedimentary rocks that originated as the rifted passive margin of Adria (Apulia) adjacent to the Pindos Ocean; each subparallel ridge is an imbricate thrust slice; near Kato Klitoria; NW Peloponnese, Greece; (c) Albania. View westward over proximal carbonate slope facies (Krasta unit), with the autochthonous platform (Albanian Alps) behind; more distal units including ophiolite in the foreground (right); (d) Bosnia. Upper Jurassic–Lower Cretaceous deep-water siliciclastic sandstone turbidites and shales deposited along the rifted northern margin of the Dinaride continent, adjacent to the Dinaric Ocean, a westward continuation of the Pindos–Mirdita Ocean in Greece and Albania.

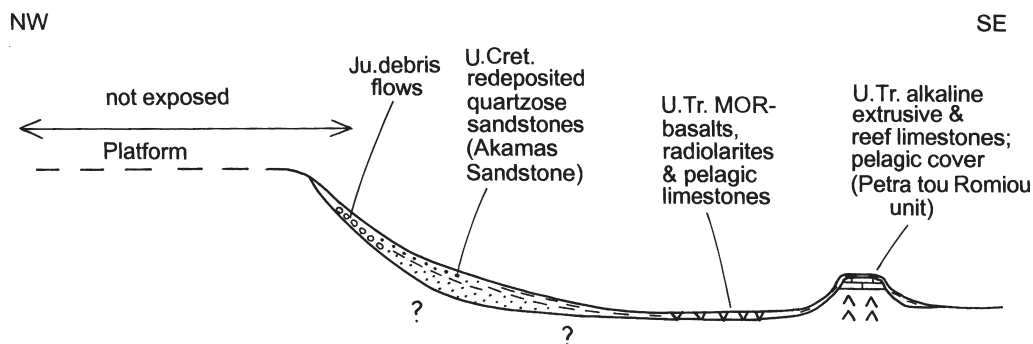


Fig. 22. Restored section across the southern margin of the Southern Neotethys in western Cyprus. Note the age and relative locations of the passive margin and exotic units. See the text for explanation.

Rifted North Tauride margins. Further north, the northern margin of the Tauride microcontinent was bordered by the North Tethyan Ocean ('North Neotethys'), which was located between the Tauride microcontinent and Eurasia (Figs 2 & 12). The Tauride 'microcontinent' is traditionally divided into an unmetamorphosed unit in the south, known as the Taurides *sensu strictu* and a metamorphosed unit further north (affected by late Mesozoic–early Cenozoic convergent tectonic), termed the Anatolides (Şengör & Yılmaz 1981). Regionally, these two units formed the Mesozoic Anatolide–Tauride platform. In most interpretations, the ophiolitic suture to the north of the Anatolide–Tauride platform, known as the Izmir–Ankara–Erzincan suture zone, is seen as accommodating the main oceanic separation between North Africa and Eurasia during Mesozoic–Early Cenozoic time (Şengör & Yılmaz 1981; Robertson & Dixon 1984; Dercourt *et al.* 1986, 1990; Göncüoğlu *et al.* 1996–1997, 2003; Okay *et al.* 2001; Robertson *et al.* 2004). Alternatively, this suture was recently interpreted by Stampfli & Borel (2002) as a back-arc basin located above a northward-dipping subduction zone. These authors further suggested that a major Late Palaeozoic–Early Mesozoic 'Palaeotethyan' suture separated the Anatolide and Tauride units (i.e. that these were of quite different palaeogeographic origin). However, no such suture has been identified regionally, and the stratigraphy of the Anatolide and Tauride units can be correlated at least from the Early Triassic onwards (Şengör & Yılmaz 1981; Demirtaşlı *et al.* 1984; Göncüoğlu *et al.* 1996–1997, 2003; Okay *et al.* 2001; Robertson *et al.* 2004). Alternative models are discussed further below, but meanwhile it is assumed that the long-established interpretation of the northern margin of the Anatolides–Taurides as a rifted passive margin is essentially correct.

There is well-documented evidence of Triassic rifting and continental breakup along the southern side of this northerly 'Neotethyan' oceanic basin; i.e. along the northern margin of the Tauride–Anatolide microcontinent (Monod 1977; Demirtaşlı *et al.* 1984; Özgül 1984; Robertson & Pickett 2000; Göncüoğlu *et al.* 2003). This margin is preserved in several emplaced units over more than 600 km from east to west (Fig. 12). Relatively proximal units are exposed within the Tauride microcontinent (i.e. within the Tauride–Anatolide platform, e.g. Geyik Dağ), whereas more distal units are preserved within overriding thrust sheets. The more proximal units (relatively autochthonous) document subsidence and both terrigenous and calcareous sedimentation related to rifting during the Triassic; this culminated in the establishment of a subsiding passive margin during Late Triassic–Early Jurassic time.

During the Late Cretaceous–Early Cenozoic, the continental margin and ophiolites were thrust southwards onto the Tauride microcontinent as a large pile of thrust sheets ('Bozkir nappes'; Şengör & Yılmaz 1981). As in Oman, the more distal parts of the rifted margin were accreted and thrust southwards onto the Tauride microcontinent, where they are exposed, generally beneath overriding ophiolites. However, the more proximal margin units (platform/slope) remained attached to the downgoing plate. As a result, the leading edge of the Tauride margin was subducted and underwent LT–HP (low temperature–high pressure) metamorphism in mid Late Cretaceous time, as well documented in NW Turkey (Okay *et al.* 1998, 2001). Continentward parts of the subducted platform (i.e. the Anatolide carbonate platform) experienced lower grade metamorphism, and areas even further south (i.e. the Tauride carbonate platform) escaped metamorphism altogether.

The northern margin of the Tauride–Anatolide microcontinent (southern margin of the North Tethys) is possibly the most complete example of an emplaced continental margin anywhere in the Tethyan region, as different shallow to deep crustal level units are all exposed along strike in different areas of the suture zone (Izmir–Ankara–Erzincan suture zone; Fig. 12), as summarized below.

The rifted passive margin subdivides into two contrasting segments. In the west (to the west of the longitude of the Isparta Angle; Fig. 12) continental slope/abyssal units were thrust southwards onto the continental margin (Menderes Massif) as a large pile of thrust sheets, initially in latest Cretaceous time (de Graciansky 1972; Poisson 1977, 1984; Collins & Robertson 1998, 1999; Danelian *et al.* 2006). These far-travelled thrust sheets, known as the Lycian Nappes (Fig. 23), restore as proximal, to more distal, slope/abyssal plain successions. As in Oman, the higher units in the thrust stack were derived from a more oceanward setting. The more proximal settings comprise fragments of prerift continental basement, overlain by shallow-water carbonates, whereas the more distal

units (Köyceğiz unit) begin with Triassic MOR-type lavas, overlain by mainly redeposited carbonates of Jurassic–Late Cretaceous age, representing a deep-water continental slope succession. As in many other areas the proximal edge of the rifted platform was concealed by overthrust units.

The Lycian thrust sheets restore as a margin that rifted in the Triassic, with the genesis of oceanic crust during latest Triassic–earliest Jurassic time. An oceanward-dipping sedimentary ramp developed above a basement that is mainly not preserved. Higher in the pile of thrust sheets more distal units, including MOR-type basalts, radiolarites and pelagic carbonates are interpreted as accreted abyssal plain material, and are exposed within a *mélange* (Lycian *Mélange*) that is located beneath an overriding regional-scale ophiolite (Lycian ophiolite) at the highest levels of the thrust stack. The *mélange* includes large blocks of Late Triassic shallow-water carbonate, overlain by pelagic sediments (Domuz Dağ unit) (Poisson 1977, 1984). These exotics were interpreted either as rifted fault blocks floored by thinned continental crust (Poisson 1977) or as oceanic seamounts underlain by volcanic rocks (Collins & Robertson 1997, 1998). The actual setting may be indeterminate as no basement to the succession is exposed.

The palaeogeography of the rifted margin differed considerably further east (east of the longitude of the Isparta Angle; Fig. 12). In this area, another series of thrust sheets encompassing the same age range was emplaced southwards onto the Tauride carbonate platform during latest Cretaceous time (Beyşehir–Hoyran–Hadım nappes; Monod 1977; Özgül 1984; Andrew & Robertson 2002). These units restore as a more complex rift palaeogeography than the Lycian nappes (Fig. 24). The Tauride carbonate platform in this area (Geyik Dağ) was bordered by a Triassic rift basin; this infilled with Triassic silicic tuffs, terrigenous and carbonate sediments, mostly turbidites and debris flows (Huğlu unit). The rift basin was then covered with mainly siliceous and pelagic carbonate sediments of Jurassic–Late Cretaceous age. The siliceous tuffs may reflect melting (anatexis) of underlying siliceous continental basement rocks that are not exposed. The rift was bordered, oceanwards, by a carbonate platform (Boyali Tepe unit), on which shallow-water carbonates accumulated in Late Triassic time. This platform later subsided and was overlain by pelagic Ammonitico Rosso, then by radiolarites and pelagic carbonates during the Jurassic–Early Cretaceous (Fig. 25).

Finally, an additional unit is exposed to the SE on the southern margin of the Tauride carbonate platform (locally termed the Bolkar Dağ), known as the Mersin *Mélange*. In essence, this is interpreted as a highly dismembered equivalent of the

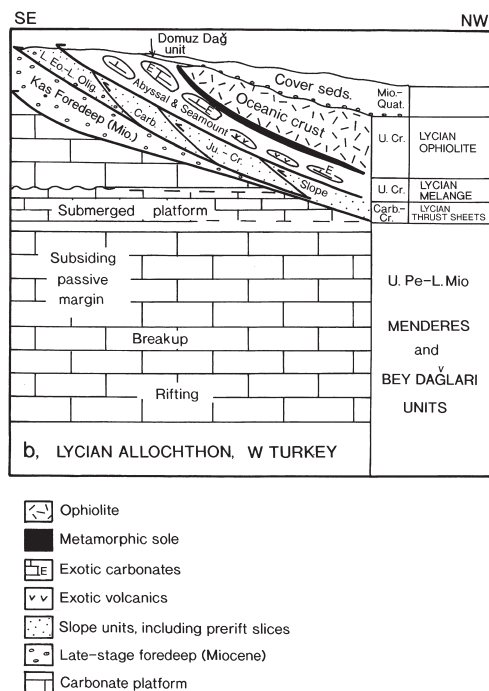


Fig. 23. Simplified tectonostratigraphic relations of rifted margin and ophiolitic units of the Lycian Allochthon (Lycian nappes), NW Turkey, inferred to have been emplaced from the Mesozoic North Tethyan Ocean onto the Tauride microcontinent (Menderes Massif).

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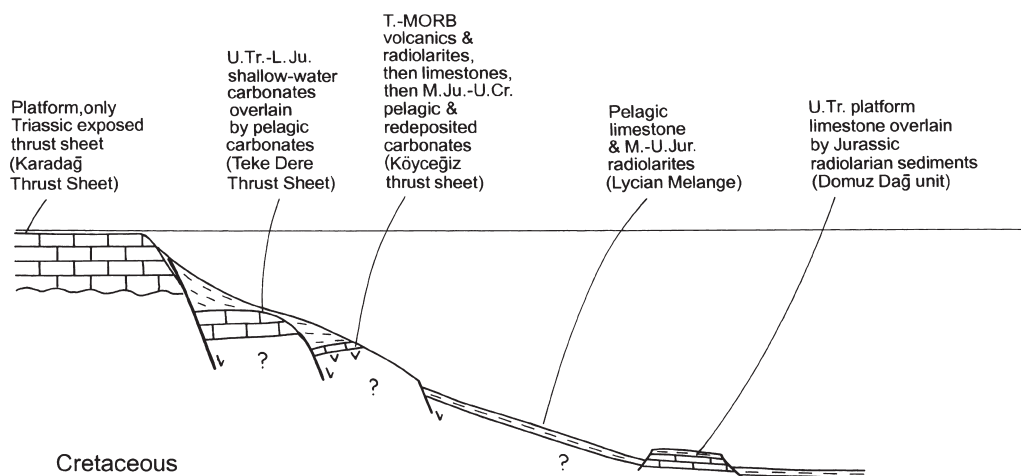


Fig. 24. Restored relations of the passive margin sediments and igneous rocks on the shelf and the slope successions represented by the Lycian nappes, which were rooted within NW Turkey. The rift and passive margin-related successions are exposed in different thrust sheets over a wide region and thus this reconstruction is tentative. From Robertson (2004).

Beyşehir–Hoyran–Hadım nappes. It includes Permian shallow-water limestones, Upper Triassic neritic carbonates, Triassic intermediate-silicic volcanics, deep-water terrigenous and calcareous sediments, and younger pelagic sediments and radiolarites (Parlak & Robertson 2004). This unit therefore provides additional evidence of rifting and continental breakup along the northern margin of the Tauride microcontinent.

Summarizing, the entire northern margin of the Tauride microcontinent (southern margin of North

Tethyan oceanic basin) represented by the Tauride and Anatolide units, the Lycian nappes, the Beyşehir–Hoyran–Hadım nappes and the Mersin Mélange shows evidence of rifting and continental breakup during Late Triassic–Early Jurassic time. The margin geometry was relatively simple in the west, with a single oceanward-dipping ramp bordered by subsiding rift blocks (Fig. 24). Further east, the Tauride microcontinent was bordered by a rift basin and a subsiding marginal platform, beyond which Mesozoic oceanic crust

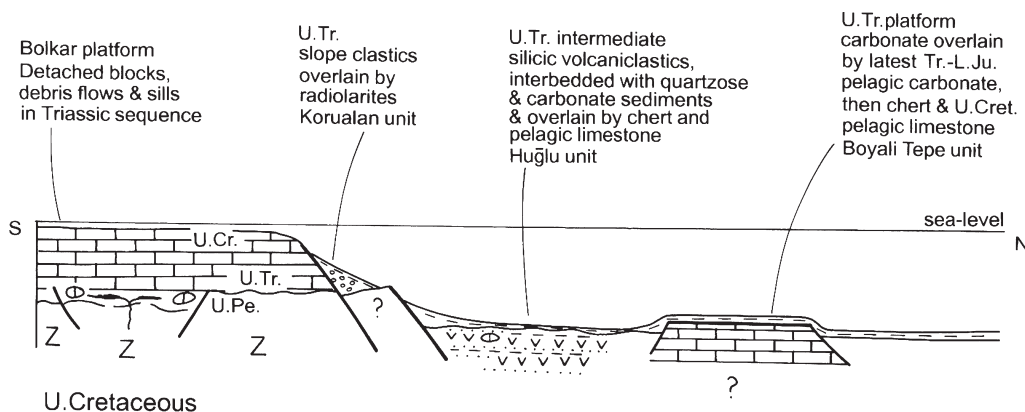


Fig. 25. Restored relations of the passive margin sediments and igneous rocks on the shelf and the slope successions, represented by the Beyşehir–Hoyran–Hadım nappes, central Turkey, which were rooted within the Inner Tauride Ocean. These allochthonous units underwent complex thrusting during latest Cretaceous and Middle–Late Eocene time and thus cannot be restored simply by simply unstacking the thrust pile. See the text for discussion.

formed from latest Triassic–earliest Jurassic time onwards (Fig. 25).

The extensive record of rift-related volcanism, but with an absence of exhumed continental basement, suggests that the south margin of the North Tethyan Ocean does not correspond to either of the ideal Volcanic-rifted or Non-volcanic rifted margin types, but is again more of an ‘intermediate’ type.

South Tethyan margins in the Balkan region.

Moving westwards back into the Balkan region, we have already considered the margin of the most southerly oceanic basin (Ionian and Herodotus basins) that separated North Africa from the Adria (Apulia) microcontinent (Fig. 2). This ocean basin is nowadays still subducting northwards along the Aegean arc, extending from western Greece to south of Crete. As noted earlier in the paper, the timing of initial rifting of this oceanic basin is controversial, as its margins remains deeply buried beneath the Sea of Crete and the Ionian Sea. However, breakup may have taken place during the Late Jurassic–Early Cretaceous (Catalano *et al.* 2001) (Fig. 2). The southerly rifted margin of the Ionian oceanic basin is represented by onshore and offshore areas of North Africa. By contrast, the northerly margin is largely beneath the sea or was removed by ‘subduction erosion’ related to Cenozoic northward subduction. However, parts of the northerly margin are exposed within a stack of thrust sheets in Crete and the Peloponnese, as mentioned earlier in the paper.

Northerly rifted margin of Adria. Further north, the next oceanic basin, the Pindos Ocean, was bordered on its southern side by the rifted margin of Adria (Apulia). Its conjugate margin is represented by the southerly margin of the Pelagonian microcontinent (Robertson *et al.* 1991; Smith & Rassios 2003; Smith 2006). An alternative (but not favoured) interpretation is discussed later in the paper. The southerly margin of the Pindos oceanic basin is well exposed in the Pindos Mountains, stretching from Albania to southern Greece (Dercourt 1964; Aubouin *et al.* 1970; Fleury 1980). To the west, the *in situ* ‘foreland’ is represented by a Mesozoic carbonate platform (Gavrovo–Tripolitza platform) that overlies Triassic synrift sediments (Phyllite–Quartzite unit) and an inferred prerift continental basement (only rarely exposed). The Adria (Apulia) foreland was overridden by a well-organized stack of thrust sheets that developed along the rifted margin to the east during Early Cenozoic time when the Pindos ocean basin finally closed (Degnan & Robertson 1998, 2006).

The Mesozoic volcanic–sedimentary rift succession (i.e. Phyllite–Quartzite unit) and the

overlying carbonate platform (Gavrovo–Tripolitza Nappe) are separated by a major Cenozoic detachment. The underlying Triassic rift succession was detached and subducted together with the downgoing plate where it experienced HP–LT metamorphism before being exhumed (Seidel 1978; Seidel *et al.* 1982; Zulauf *et al.* 2002). By contrast, the more oceanward part of the rifted margin (Gavrovo–Tripolitza nappe) and the former slope and abyssal plain units (Pindos–Olonos nappe) were retained within the overriding plate at a high structural level and remained unmetamorphosed.

The underlying Phyllite–Quartzite unit records pulsed rifting during Late Palaeozoic–Early Mesozoic Triassic time, with extensive extrusion of intermediate-composition–silicic volcanics, and deposition of mixed tuffaceous–carbonate–terrigenous sediments. Evidence from Crete indicates this volcanism was mainly of Early Triassic age (Krahl *et al.* 1983; Krahl & Kaufmann 2004), contrasting with a previously assigned Late Permian age (Seidel 1978).

Facies evidence suggests that the latest stage of rifting during the Late Triassic was followed by passive margin subsidence and the construction of a north-facing carbonate platform (several kilometres thick) that persisted from Late Triassic to Early Cenozoic time. The proximal slope of the carbonate platform was mainly overthrust and concealed, as in many other areas, but locally appears beneath the overriding thrust sheets in the western Peloponnese, Greece. By contrast, the base-of-slope to abyssal-plain sediments (equivalent to the Hawasina Complex in Oman) are well exposed and form a well-ordered imbricate stack of thrust sheets (Pindos–Olonos nappes) (Fig. 21b). Unlike most of the other Tethyan rifted margins, overriding ophiolites are absent above these distal margin/oceanic sediments, and largely for this reason the deep-sea sedimentary successions are unusually intact stratigraphically (Neumann 2003). The Pindos–Olonos thrust sheets are stacked in a regular piggy-back fashion, which allows the original proximal–distal settings to be restored with confidence.

The Pindos–Olonos succession, as documented in the NW Peloponnese (Degnan & Robertson 1998, 2006), begins with locally preserved tholeiites that are positionally overlain by deep-water pelagic limestones, containing the thin-shelled bivalve, *Halobia*. The succession continues upwards through a range of deep-water redeposited terrigenous and carbonate sediments, siliceous radiolarian sediments and pelagic carbonates. It then culminates in thick Early Cenozoic terrigenous turbidites (Pindos flysch). In southern Greece these terrigenous sediments were mainly derived from Adria (Apulia) to the west (Piper 2006), whereas counterparts in northern Greece (north of the Gulf of

Corinth) include ophiolitic detritus (Faupl *et al.* 2002) that was derived from the overriding plate (Pelagonian microplate) as the Pindos ocean closed.

A thin unit of *mélange* is entrained beneath the basal thrust of the Pindos–Olonos thrust sheets (e.g. Degnan & Robertson 1994; Pe-Piper & Piper 2004). This includes numerous blocks of intermediate-silicic tholeiitic extrusives and related pelagic sediments, locally dated as Triassic in age, together with terrigenous sediment derived from the foreland (Degnan & Robertson 1998). The *mélange* is interpreted as a remnant of the original continent–ocean transition of the Pindos Ocean that was accreted and then emplaced onto the foreland during Early Cenozoic closure of the Pindos Ocean (Degnan & Robertson 2006).

The thrust front directly above the foreland in the west includes several large (up to kilometre-sized) blocks of basaltic volcanics, overlain by Triassic ammonite-bearing pelagic limestones (e.g. Glafkos and Meghdovas units). These ‘exotics’ are interpreted as igneous extrusions capped with stratigraphically condensed pelagic sediments that originated along the rifted margin; these then remained as relatively elevated blocks during the passive margin subsidence phase. Clastic sediments by-passed these discontinuous marginal ‘highs’ to accumulate in deeper-water slope–abyssal plain settings (Fig. 26b), and were much later detached and incorporated into the base of the thrust stack when the Pindos ocean basin finally closed during the Early Cenozoic (Robertson *et al.* 1991).

In an area north of the Gulf Corinth the rifted margin of Adria (Apulia) was bordered by a rifted fragment some tens of kilometres across, known as the Parnassus unit (Celet 1960) (Fig. 12). This unit was, in turn, bordered to the west by several thrust sheets, termed the Vardoussia unit (Ardaens 1978); these can be restored as the westerly, rifted margin of a discrete Parnassus microcontinent (Degnan & Robertson 1998) (Fig. 26a). The Vardoussia succession begins with Middle Triassic, mixed carbonate–clastic sediments, tuffs and volcanics, and passes upwards into redeposited slope carbonates, derived from the Parnassos carbonate platform to the east (Ardaens 1978; Pe-Piper & Piper 2004).

The deep-water succession bordering Adria (Pindos–Olonos nappes) can be traced northwards through northern Greece and Albania, as far as the Scutari–Peć lineament (Fig. 12). This is interpreted as an important continental-margin transform fault, which segmented the Adria margin into palaeogeographically contrasting areas (e.g. Dercourt *et al.* 2000; Robertson & Shallo 2000). Beyond this, similar deep-water Mesozoic–Early Cenozoic sediments are exposed in Budva Zone of former

Yugoslavia, but there is no evidence that this basin advanced beyond the rift phase (i.e. it represents an aulacogen). The extension of the Pindos Ocean is instead located northwards of the Dinaride continent (related to Apulia), where it includes a deformed passive margin succession, known as the Bosnian Flysch (Fig. 21d), which is in structural contact with the Dinaride ophiolite belt (e.g. Robertson & Karamata 1994; Karamata 2006). Discussion of the possible linkage, westwards through Bosnia, Croatia, Hungary and Austria, to the Western Tethys rifted margin, although fruitful, is outside the present scope.

In summary, evidence from the emplaced northern margin of Adria shows that, following Late Palaeozoic pulsed rifting, the Pindos Ocean began to open during latest Triassic–Early Jurassic time, associated with the development of a proximal carbonate platform, passing eastwards into a distal deep-water succession. Part of the rifted continent–ocean transition is preserved as accreted blocks within *mélange*. Rift-related fault blocks adjacent to the ocean subsided and were covered by pelagic carbonate. The rifted margin was bordered by one much larger rifted fragment (Parnassos), with its own small passive margin (Vardoussia).

Southerly rifted margin of the Pelagonian microcontinent. Here, it is assumed that the conjugate margin of Adria (Apulia) is represented by the southerly rifted margin of the Pelagonian microcontinent. Alternative models are discussed in the section following this overview of Balkan rifted margins.

The former southerly margin of the Pelagonian microcontinent is now represented by thrust sheets and *mélange*; these are exposed structurally above the Pelagonian Zone to the east. These marginal units are most intact in the Othris Mountains of central Greece (Smith *et al.* 1995), but they are also exposed, typically, as *mélange* and dismembered thrust sheets over a wide area (Fig. 12), notably in the Pindos Mountains (Aydella *Mélange*: Jones & Robertson 1991, 1994) (Figs 27 & 28) on the large island of Evia (Ferrière *et al.* 1988; Robertson 1991; Danelian & Robertson 2001), and in the Argolis Peninsula (Baumgartner 1985; Clift & Robertson 1990) further south. In the interpretation favoured here these early Mesozoic rift-related units were emplaced northeastwards from the Pindos Ocean. The driving force was the collision of a subduction trench with the Pelagonian microcontinent during Middle–Late Jurassic time (Robertson *et al.* 1991; Smith & Rassios 2003; Rassios & Moores 2006; Smith 2006).

Along the former southern margin of the Pelagonian microcontinent, as exposed in the

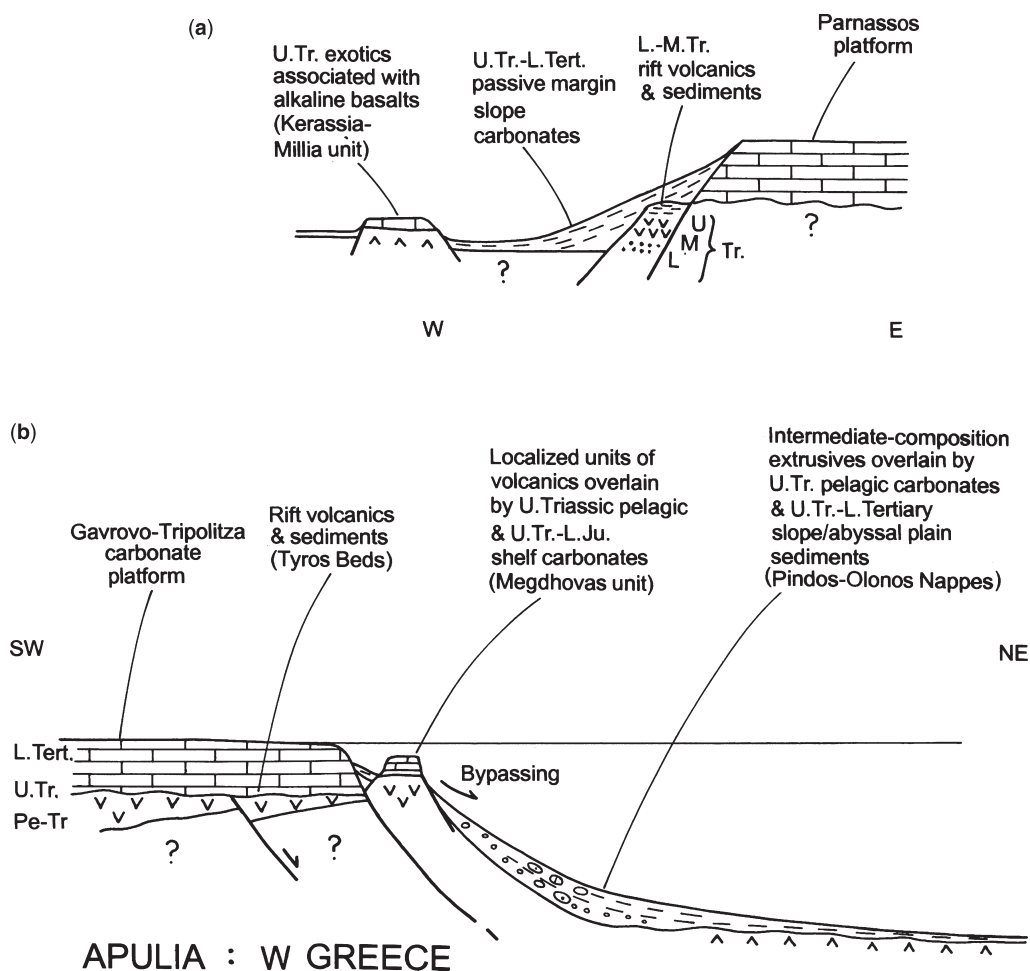


Fig. 26. Restored relations of the passive margin sediments and igneous rocks on the shelf and the continental slope successions related to the SW margin of the Mesozoic Pindos Ocean exposed in western Greece. (a) Central western Greece (north of the Gulf of Corinth) where the rifted margin is inferred to have been bordered by a small microcontinental unit, the Parnassos platform (see Fig. 12), and its passive margin (Vardoussia unit). (b) Rifted margin of Adria (Apulia) facing eastwards into the Pindos ocean, as exposed within the Peloponnese, NW Greece. (a) and (b) are effectively conjugate margins. See the text for discussion.

regional Pelagonian nappe, Late Palaeozoic metamorphic basement rocks are depositionally overlain by Upper Permian non-marine clastic sediments (Mountrakis 1986). Above, the Triassic succession includes bimodal volcanics and mixed carbonate-clastic successions, then thick Jurassic platform carbonates, that are well exposed on the island of Evia. The Pelagonian platform cannot be considered as a single homogeneous entity on a regional scale, but was palaeogeographically varied. Several intraplateform basins can be reconstructed. These were flooded by intermediate-composition volcanics and

volcaniclastic sediments, in turn overlain by redeposited platform-derived carbonates, as seen in the Argolis Peninsula (Clift & Robertson 1990; Clift 1995) and probably also in northern Greece (Sharp & Robertson 2006).

The proximal-distal deep-water rifted margin can be restored from overriding mélangé and thrust sheets (Fig. 29). Successions are most intact, stratigraphically, in the Othris area where both relatively proximal deep-water slope to base-of-slope successions and more distal rift-related alkaline volcanics (Agrilia lavas) are present (Smith *et al.*

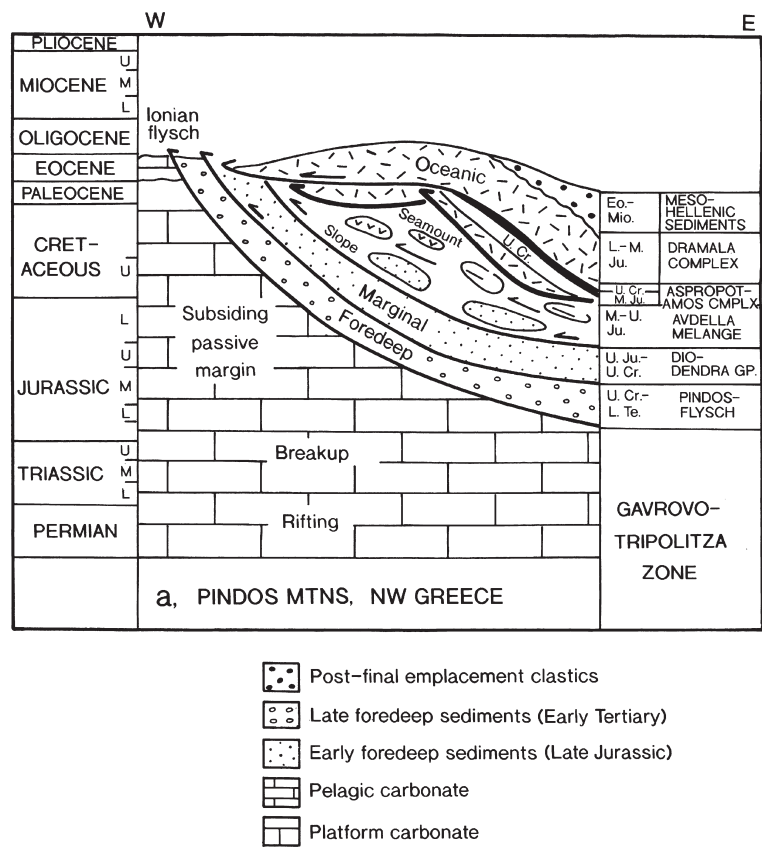


Fig. 27. Simplified tectonostratigraphic relations of rifted margin, oceanic and ophiolitic units of the Pindos allochthon in the Pindos Mountains, NW Greece, inferred to have been emplaced from the Pindos Ocean onto the Pelagonian microcontinent. The initial emplacement occurred in Late Jurassic time, followed by rethrusting towards the SW during Early Tertiary time. From Robertson (2004).

1975; Price 1976). A wide range of Triassic rift-related rocks are also well exposed further NW in the Avdella Mélange of the Pindos Mountains (Jones & Robertson 1991, 1994). Lithologies include terrigenous turbidites and platform-derived carbonates, interspersed with alkaline to transitional type composition volcanics and radiolarian cherts. Shallow-water carbonates are interpreted as small platforms that developed on subsiding marginal fault blocks. The Jurassic was characterized by passive margin subsidence with the deposition of slope carbonates and pelagic sediments (e.g. in the Othris area; Price 1976).

The passive margin phase was followed by subduction and tectonic accretion, culminating in the emplacement of the rifted margin and associated ophiolitic units onto the Pelagonian microcontinent during Middle–Late Jurassic time (Fig. 28). After this emplacement shallow-water carbonate platform deposition resumed on the Pelagonian

microcontinent during the Late Jurassic (locally) and the Cretaceous (e.g. on Evia). During the Early Cenozoic continental collision finally sutured the Pindos Ocean, and the Pelagonian microcontinent was mainly detached from its prerift basement and internally imbricated to form several huge thrust sheets that were emplaced to the SW.

Summarizing, the inferred northerly conjugate margin of the Pindos Ocean (southern margin of the Pelagonian microcontinent) records rifting during Triassic time associated with widespread volcanism, commonly of alkaline (transitional) type; this was followed by passive margin subsidence during the Early Jurassic. The NE part of the Pindos Ocean (bordering the Pelagonian microcontinent) was deformed during Middle–Late Jurassic time, associated with the emplacement of rift-margin units and ophiolites onto the Pelagonian microcontinent to the NE (present co-ordinates). By contrast, the conjugate Adria (Apulia) margin to the

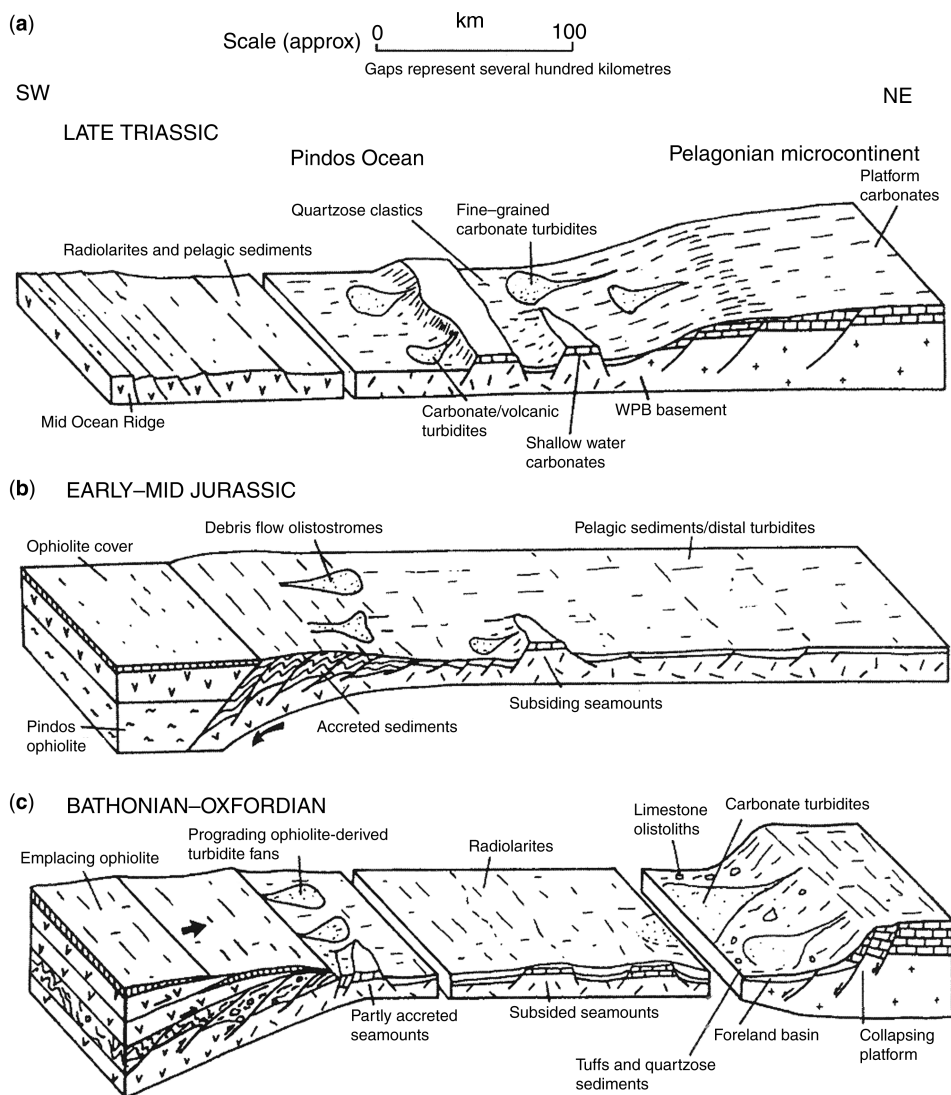


Fig. 28. Block diagrams showing the restored relations of the passive margin sediments and igneous rocks on the shelf and the slope successions of the represented by the Avdella Melange, Pindos Mountains, NW Greece. Note the role of accretionary processes. From Jones & Robertson (1994).

south escaped this deformation and remained as a subsiding passive margin until final closure of the Pindos Ocean in Early Cenozoic time.

The conjugate rifted margins of the Pindos Ocean do not show evidence of voluminous plume-type regional magmatism as in the Volcanic-rifted margin type, or the simple shear-controlled basement exhumation as seen in Non-volcanic margins. On the other hand, Triassic rift volcanism (up to hundreds of metres thick) is regionally extensive and arguably compatible with a plume origin

(Dixon & Robertson 1999). The rifted Pindos Basin again appears to document an 'intermediate' rift setting compared to the two end members.

North Tethyan margins in the Balkan region. To the north of the Pelagonian microcontinent the third Mesozoic oceanic basin is known as the Vardar Ocean (Figs 2 & 12). The Vardar Ocean extended westwards from northern Greece, through former Yugoslavia and Hungary to the Eastern Alps. Indeed, it was the Late Cretaceous closure of this

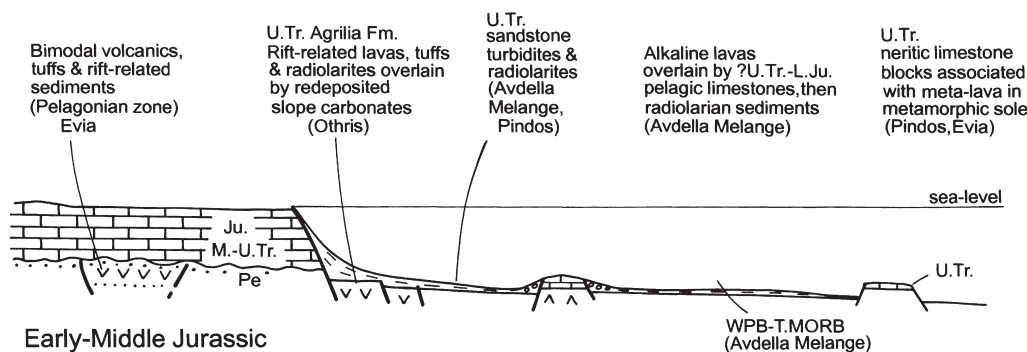


Fig. 29. Restored relations of the rifted margin of the Pelagonian microcontinent, passing into the Pindos Ocean in central and northern Greece (Othris, Evia and Pindos area). See the text for discussion.

ocean basin that triggered the initial compression of the northern margin of the Jurassic Ligurian–Piedmont Ocean, as seen in the Alps (see Manatschal *et al.* 2007). The Vardar Ocean is the most difficult of all of the Eastern Tethyan ocean basins to interpret, mainly because of only limited study, regional metamorphism of some units and limited exposure. Here it is assumed that the Vardar Ocean was bordered to the south by the northern rifted margin of the Pelagonian microcontinent and by its conjugate, the southern margin of the Serbo-Macedonian continent. However, alternative interpretations are mentioned in the following section.

The Vardar Zone is a composite unit and is characterized by several subunits, namely the Almopias unit in the west, the Paikon unit in the centre and the Peonais unit in the east (Mercier 1968, 1979). Each of these three units is internally complex. Successions exposed in each unit are shown in Figures 30 & 31. The northern margin of the Pelagonian microcontinent (Pelagonian Zone) includes Triassic–Jurassic metamorphosed shallow-water carbonates associated with alkaline meta-basalts that are assumed to have erupted during Triassic rifting (Sharp & Robertson 2006). The Almopias Zone was thrust SW over the Pelagonian Zone during the final Early Cenozoic closure of the Vardar Ocean. No rifted-margin slope sediments are known either in the Pelagonian Zone or the Almopias unit. The Almopias unit includes MOR-type basalts (Sharp & Robertson 1994, 2006) in depositional contact with Triassic radiolarites (Stais & Ferrière 1991), and is interpreted as a fragment of Triassic Vardar ('Almopias') oceanic crust. The Paikon unit is dominated by a Jurassic volcanic arc that relates to partial closure of the Vardar Ocean (Mercier 1986; Brown & Robertson 2003, 2004). The Peonais unit is characterized by

a Jurassic ophiolite (Guevgueli ophiolite) that is interpreted as a marginal basin above a northward-dipping subduction zone (Bébién *et al.* 1987). The Peonais unit includes Jurassic terrigenous turbidites (Svoula Flysch: Kauffmann *et al.* 1976), interpreted as part of the deep-water slope sequence of the Serbo-Macedonian continental unit to the north. The southern margin of the Serbo-Macedonian Zone (rifted northern margin of the Vardar Ocean) exposes poorly dated Permo-Triassic alluvial conglomerates, overlain by Triassic bimodal MOR-type tholeiites, in turn passing into a Late Triassic carbonate platform succession (Stais & Ferrière 1991).

In summary, the Vardar Ocean is well exposed in northern Greece and is marked by a poorly preserved Triassic rifted margin on its southern side (Pelagonian microcontinent) and a better preserved Triassic rifted margin on its northern side (Serbo-Macedonian continent). This includes both proximal non-marine, shallow-marine and deeper-marine successions. 'Permo-Triassic' rifting was followed by passive margin subsidence by Late Triassic time as the Vardar Ocean opened. The record is too sparse to allow useful comparison with either a Volcanic-type or a Non-volcanic-type rifted margin. However, it is notable the rift volcanics on the northern (Serbo-Macedonian) margin are unusually voluminous and could show a plume influence based on the published geochemistry (Dimitriadis & Asvesta 1993). This concentration of preserved rift volcanism on the northern margin is also suggestive of the rift being asymmetrical.

The Serbo-Macedonian Zone to the north has in the past been regarded either as part of the regional south margin of Eurasia or as a discrete microcontinent. Recently, Himmercus & Reichmann (2006) have identified what they consider to be a

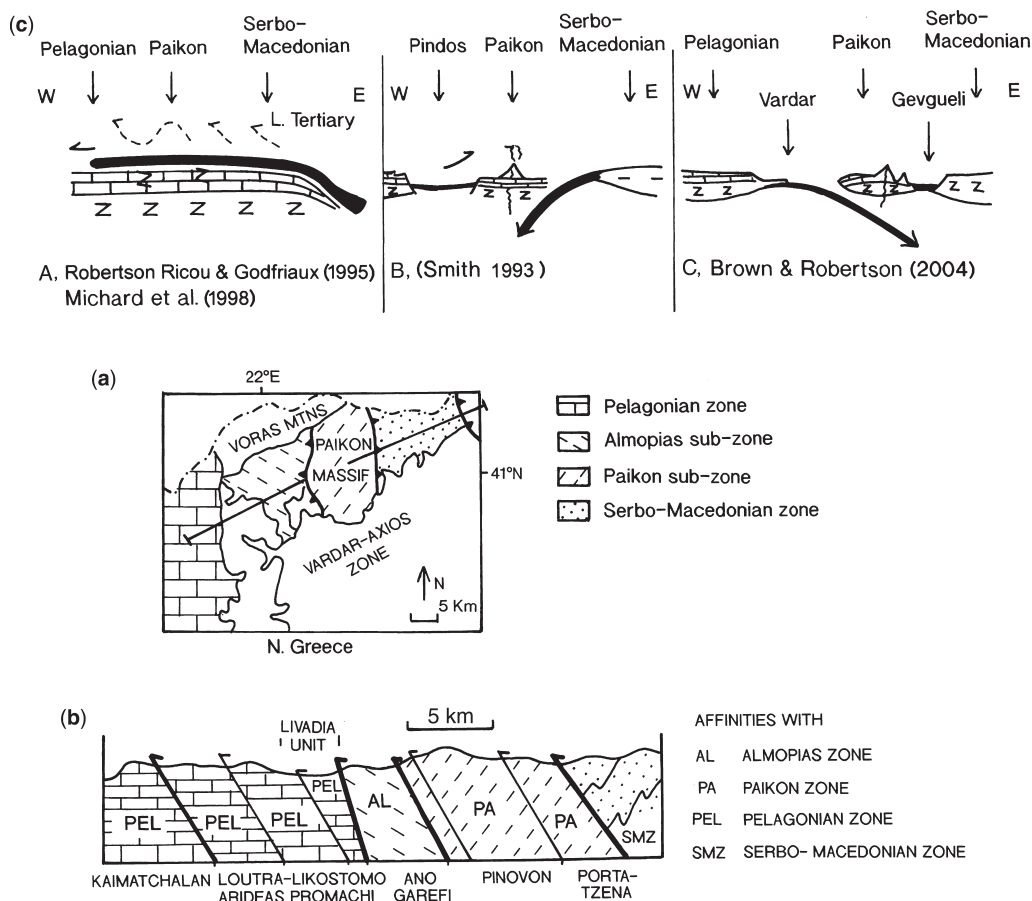


Fig. 30. Regional setting of the Vardar Zone, northern Greece between the Pelagonian microcontinent and the Serbo-Macedonian continent to the north. (a) Outline geological map; (b) simplified cross-section. Triassic margin units are exposed in the Pelagonian Zone and the Almopias Zone. (c) Alternative tectonic models for the Vardar Zone during Middle–Late Jurassic time: (A) thrusting from an ocean far to the north beyond the Serbo-Macedonian continent; (B) subduction to SW below the Pelagonian continent; and (C) subduction to NW below the Serbo-Macedonian continent. (D) is supported by structural and geochemical evidence. See the text for data sources and explanation.

‘mid-Alpine’ suture within the high-grade metamorphic rocks of the Serbo-Macedonian Zone. This suture may possibly be an equivalent of the Izmir–Ankara–Erzincan suture in Turkey, which, as noted in a previous section, is widely interpreted as the main suture of the Northern Tethys Ocean. Further north again, the true Eurasian continental margin is represented by the northern part of the Serbo-Macedonian Zone, the Rhodope massif and the Strandja massif. The southern margin of Eurasia in this region appears to have experienced a long history of active margin processes (Karamata & Janković 2000; Karamata & Vujnović 2000; Karamata 2006) comparable with the Eurasian

margin of Turkey, further east, as exposed in the Pontides and the Caucasus (e.g. Adamia *et al.* 1995; Nikishin *et al.* 2001).

Alternative Tethyan reconstructions

In the preceding sections, the Mesozoic tectonic setting of the Eastern Mediterranean has been interpreted in terms of several microcontinents separating three main interlinked small oceanic basins bordered by mainly Triassic rifted margins. However, it is fair to note that there are several quite different interpretations in the literature. In the past there have been various attempts to

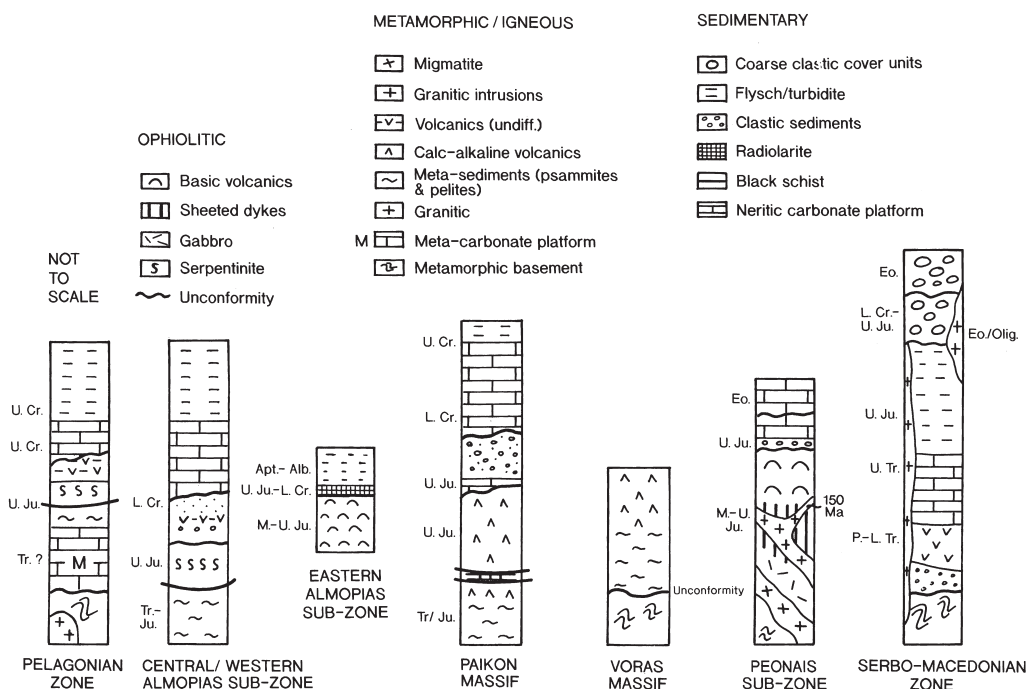


Fig. 31. Summary of successions within and adjacent to the Vardar zone, northern Greece. 1–2, rifted margin of the Pelagonian continent; 3, Upper Triassic oceanic crust; 4–5, continental margin-type arc constructed on continental crust of the Serbo-Macedonian zone; 6, back-arc marginal basin, rifted within Serbo-Macedonian continental crust; and 7, Serbo-Macedonian continent. See Figure 30a for the location of units and the text for discussion.

simplify an admittedly complex palaeogeography. Several of the currently debated tectonic reconstructions are shown in Figures 32 & 33.

In one school of thought (Fig. 32a) the Mesozoic ophiolites and related rifted margin units were thrust hundreds of kilometres generally southwards from a single, wide Mesozoic Tethys located adjacent to Eurasia (Dercourt *et al.* 1986, 2000; Stampfli *et al.* 1991). For Turkey and Cyprus, this implies thrusting of all the ophiolites (e.g. Troodos ophiolite) and the rifted margin units (e.g. Mamonia Complex, western Cyprus; Antalya Complex, SW Turkey) over the Tauride–Anatolide platform from a single Mesozoic Tethys to the north during latest Cretaceous–early Cenozoic time (Ricou *et al.* 1974). However, this can be excluded as these allochthonous unit would have had to have been thrust through a stratigraphical succession, which, at least locally (e.g. Bey Dağları carbonate platform; Isparta Angle apex; Fig. 12), remained intact during the time in question (Late Mesozoic–Early Cenozoic). In the Balkan region the situation is less clear-cut as no such intact ‘blocking’ successions are preserved. It has been argued for this area that all of the Mesozoic ophiolites and the rifted margin units originated in the northerly Vardar

Ocean or even further north (i.e. an ‘ultra-internal origin’) (Michard *et al.* 1998), and were then thrust over the Pelagonian Zone into the Pindos Zone, which is considered as a continental rift (Aubouin *et al.* 1970; Şengör 1984; Dercourt *et al.* 1986, 2000; Stampfli *et al.* 1991). However, the available structural evidence from the Jurassic ophiolites, especially the Vourinos ophiolite, is mainly indicative of emplacement from a Pindos ocean, towards the Pelagonian Zone to the NE (Rassios *et al.* 1994; Rassios & Moores 2006) (Fig. 34). In addition, structural evidence from several of the rifted margin units, especially Othris, supports emplacement from a Pindos ocean located to the SW (Smith *et al.* 1979; Robertson *et al.* 1991) (Fig. 35).

Another option is that some of the sutures can be considered as exotic terranes, in which different parts of the same Tethyan oceanic basin and its margins were interleaved by strike-slip. Specifically in the Balkan region, it has been suggested that the Pindos and Vardar sutured units formed in a single Mesozoic ocean basin that was later dissected and displaced to form two subparallel exotic terranes (Smith & Spray 1984). However, in the author’s view the tectonostratigraphy of the Vardar Zone

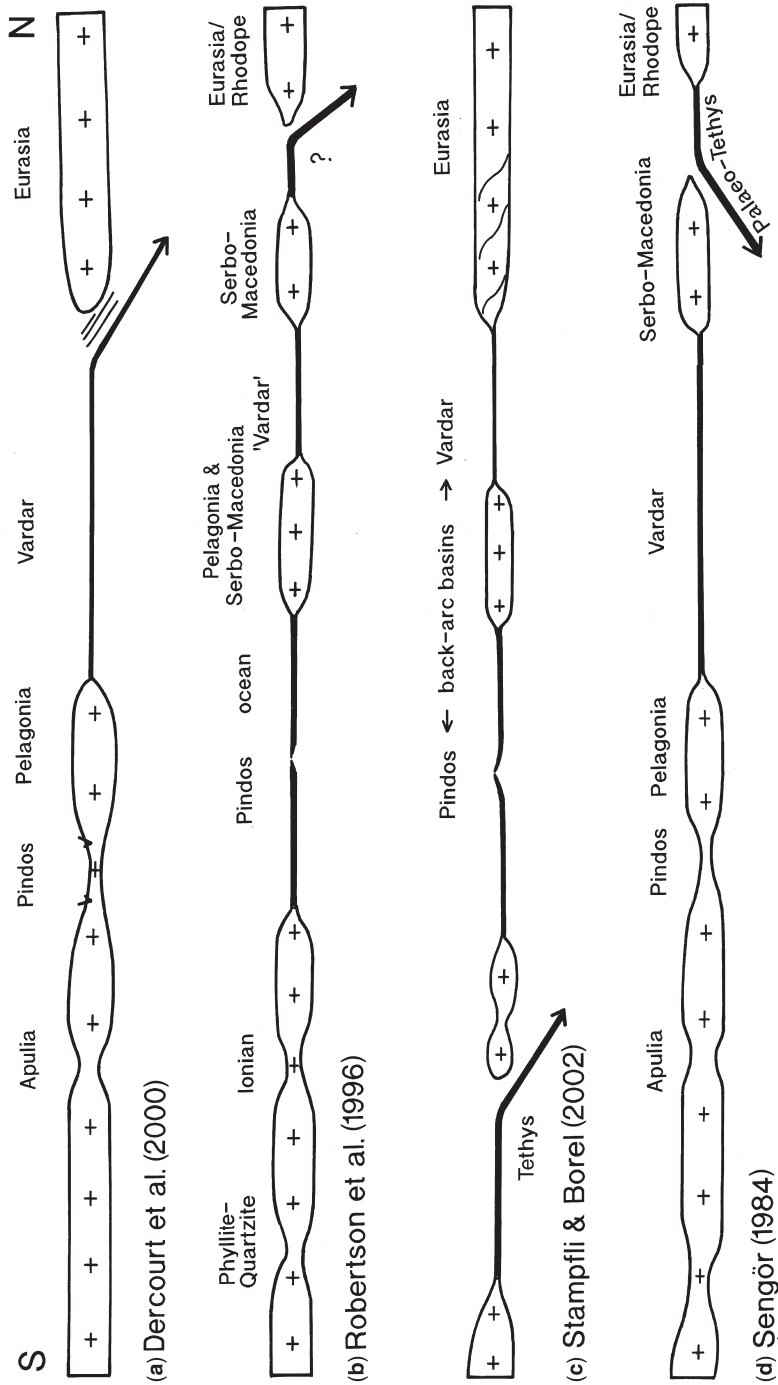
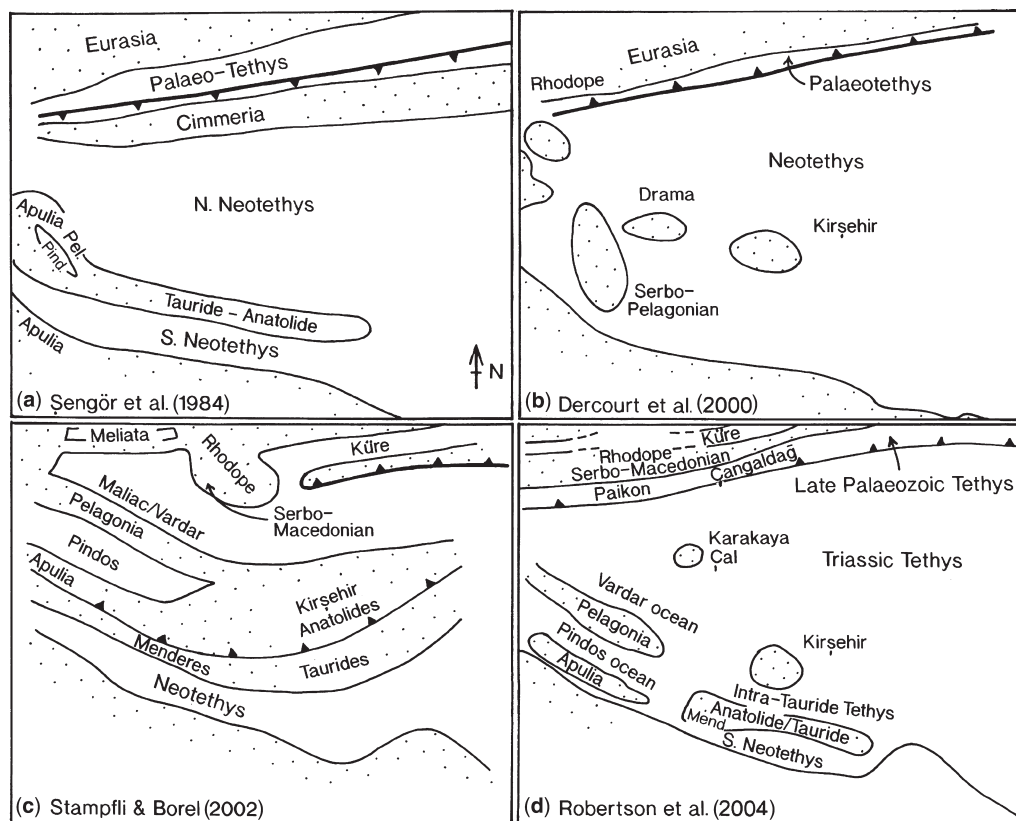


Fig. 32. Cartoons of alternative tectonic reconstruction of Tethys in the Balkan region. (a) Assumes only one Mesozoic Tethyan ocean existed and that the ophiolites and some of the ophiolitic units were thrust hundreds of kilometres generally southwards to their present positions (Dercourt *et al.* 2000). (b) Assumes the presence of marginal rifts (i.e. Phyllite-Quartzite unit in Crete and south Peloponnese), but with the Pindos and Vardar zones as small Triassic oceanic basins bordered by rifted margins (Robertson *et al.* 1996; and this paper). (c) Assumes a north-dipping subduction zone now located in the south Aegean region (i.e. Crete and south Peloponnese), with the Pindos and Vardar zones representing back-arc basins. (d) Postulates southward subduction and marginal basin formation. See the text for discussion.



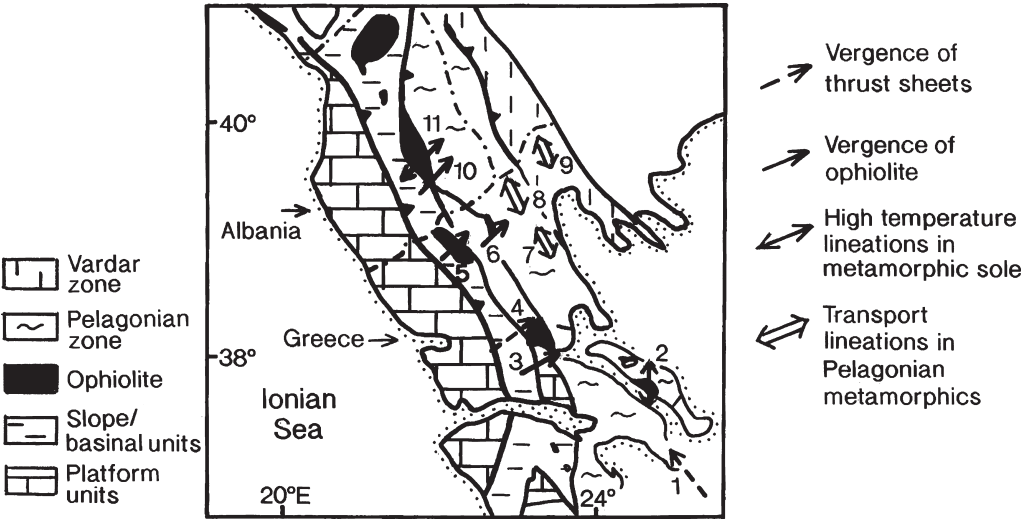
MID-LATE TRIASSIC

Fig. 33. Sketch palaeogeographical maps of the Eastern Mediterranean region for the key Late Triassic period (arranged in chronological order): (a) southward subduction model (Şengör 1984); (b) northward subduction with continental rifts bordering Gondwana (Dercourt *et al.* 1986, 2000); (c) northward subduction and marginal basin formation (Stampfli & Borel 2002); (d) northward subduction with continental fragments rifted from Gondwana (Robertson *et al.* 2004; this paper). The main differences between (b) and (d) is the size and location of continental fragments and the timing of spreading. See the text for discussion.

and the Pindos Zone are sufficiently different to make this unlikely. Also, the tectonic contacts are low-angle thrusts and the required major high-angle strike-slip faults have not been identified. For the ‘Turkish region’ application of the terrane hypothesis would imply that the Northern and Southern Tethyan units were once part of a single ocean that was tectonically duplicated during latest Cretaceous–Early Cenozoic time. This, again, seems unlikely; the tectonostratigraphy of rifted margin units north and south of the Tauride–Anatolide platform differ in important respects, and tectonic contacts are, again, low-angle thrusts (except locally as in the Antalya Complex, SW Turkey). In this tectonic interpretation the Tauride–Anatolide platform needs to be part of the north Gondwana margin from which it was transported laterally, ending up outboard (i.e. northward) of

oceanic units. However, in the east the Arabian margin did not suture until the Miocene, and further west in the easternmost Mediterranean the North African (north Gondwana) margin remains in a syn- to pre-collisional setting. Removal of a regional-scale terrane from North Africa and interleaving with oceanic units is difficult to envision. Finally, the near orthogonal (north–south) Africa–Eurasia convergence pathway (Livermore & Smith 1984; Smith 2006) does not provide any obvious driving mechanism for long-distance east–west terrane displacement.

In summary, the author does not believe that the multiple sutures can be explained, either by very long-distance southward thrusting of oceanic units or by terrane displacement and interleaving of units from a single Mesozoic ocean basin. However, this does not exclude an important but more



Pre-later Cretaceous–Early Cretaceous transgression

Fig. 34. Summary of structural evidence of emplacement directions of allochthonous units in Greece and Albania. This supports emplacement from a Pindos oceanic basin to the SW (in present co-ordinates) during Late Jurassic time. See the text for data sources and discussion.

local role for strike-slip during rifting and later emplacement, as exemplified by the Antalya Complex. It is therefore, assumed that the Eastern Mediterranean region, unlike Oman or the

Himalayas, was characterized by several contemporaneous small oceanic basins interspersed with microcontinents during Mesozoic–Early Cenozoic time.

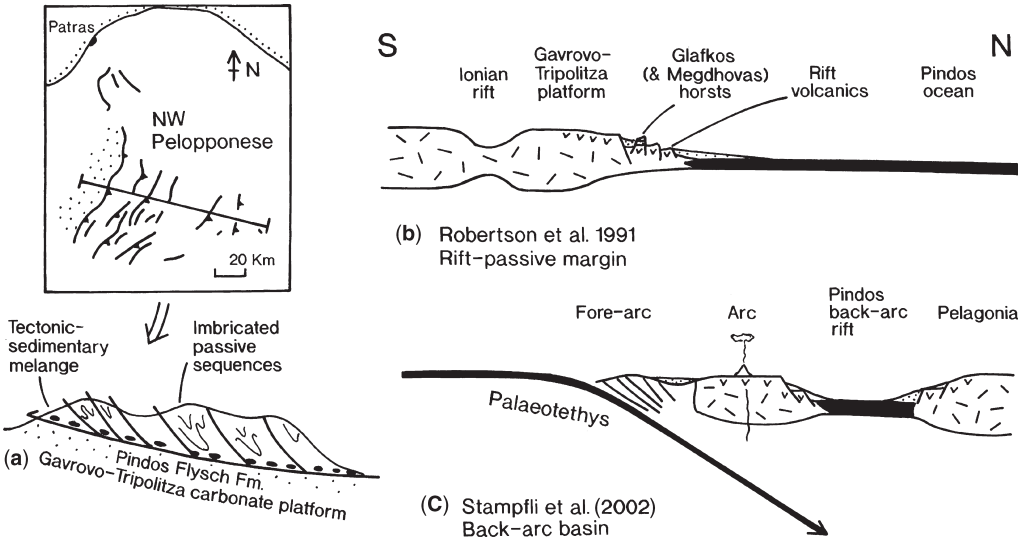


Fig. 35. Comparison of two tectonic models for the setting of the Triassic rifted-margin units in the NW Peloponnese, Greece. (a) Outline structural map and section showing thrust slices of Triassic–Early Cenozoic deep-water sediments of the Adria (Apulia) rifted passive margin (Pindos–Olonos nappes). (b) Interpretation as a rifted margin of a small Pindos oceanic basin (favoured here). (c) As a back-arc-basin above a north-dipping subduction zone. See the text for discussion.

Why alternative back-arc settings are unlikely?

In the foregoing discussion it has been assumed that several continental fragments rifted from Gondwana, without any local association with subduction (e.g. Robertson & Dixon 1984; Dercourt *et al.* 1986, 2000; Robertson *et al.* 1991; Smith 1993, 2006). However, several authors have suggested that some, or even all, of the Eastern Tethyan rifts represent back-arc basins (Pe-Piper 1982; Şengör 1984; Stampfli *et al.* 2001) that opened related to subduction beneath a continental borderland. A back-arc setting is quite attractive in principal as it provides a viable mechanism to open peripheral rift basins. Such rifts might be comparable with, for example, the Woodlark back-arc basin, SW Pacific, which developed within a long-lived active margin setting (see Goodliffe & Taylor 2007) or the Tyrrhenian Sea, Western Mediterranean Sea, which opened in a regional convergent setting (Kastens *et al.* 1988). Even if valid, this would not preclude valuable insights on rift processes, as shown by studies of the Woodlark rift basin, which is located in a regional back-arc setting (Goodliffe & Taylor 2007).

There is widespread evidence of rifting along the southern margin of Eurasia during Late Palaeozoic–Early Mesozoic time related to generally northward subduction (see Robertson 2002, 2004, 2006*b*). This includes the Jurassic Guevgueli marginal basin, NE Greece (Bébian *et al.* 1987) and the Triassic–Jurassic Küre Basin, northern Turkey (Ustaömer & Robertson 1994, 1977). These back-arc basins are associated with coeval arc volcanism (e.g. Paikon arc, northern Greece). However, it should be noted that these units are unrelated to the Triassic rift settings discussed in this paper.

Several different back-arc-type settings have been proposed for the Eastern Tethyan region, as follows.

Rifting above a southward-dipping subduction zone

The subduction zone dipped southwards from a pre-existing ocean to the north (Palaeotethys) beneath the northern margin of Gondwana as a result a continental fragment ('Cimmeria') rifted and drifted into Tethys, opening one or several back-arc basins (Figs 32d & 33a). This model has been applied to the Turkish area (Şengör & Yılmaz 1981) and Oman (Şengör 1984), but it can also be applied to the Balkans and the Himalayas. However, several points oppose this interpretation. First, the requisite magmatic arcs of the appropriate age (e.g. Triassic) appear to be absent. They should

have drifted away from Arabia as marginal basins opened, but are yet to be identified in accretionary terranes further north (e.g. Triassic Karakaya mélange terrane, northern Turkey). Secondly, in many areas (e.g. Himalayas, Oman, Baer-Bassit, Cyprus, Pindos) the Triassic rift-related basalts lack a subduction influence as expected for a subduction-related setting (Dixon & Robertson 1993; Robertson 2002). On the other hand, Triassic volcanics of the Balkans (e.g. southern Greece, Albania, Serbia, Eastern Alps, Dolomites) commonly do show a subduction-related chemistry, which requires an explanation. Thirdly, the deep-sea sedimentary covers of the more easterly rift settings mentioned above (Himalayas, Oman, Baer-Bassit, Cyprus, Pindos) are mainly terrigenous (Robertson 1981, 1986). This contrasts strongly with marginal basin settings such as the Woodlark rift basin (e.g. Taylor *et al.* 1999*a, b*; Robertson *et al.* 2001) and the Tyrrhenian Sea (e.g. Marsili Basin: Kastens *et al.* 1988). Fourthly, multiple marginal basins of similar age (Triassic) would be required to explain the multistranded nature of the Early Mesozoic small ocean basins, as discussed earlier in the paper. The absence of related Early Mesozoic magmatic arcs is thus even more problematic. Finally, the subduction-related rift model requires additional elements, notably accretionary wedges and fore-arc basins of the appropriate ages (infilled with arc-derived volcanoclastic sediments), but these do not appear to be present in the requisite locations.

Rifting above a northward-dipping subduction zone

In this model a subduction zone dipped northwards beneath the Eurasian margin, continuously from Late Palaeozoic to Early Cenozoic time (Stampfli *et al.* 2001; Stampfli & Borel 2002; Stampfli Kozur 2006) (Figs 32c & 33c). Northward-dipping subduction is indeed widely inferred (Bernoulli & Laubscher 1972; Robertson & Dixon 1984; Dercourt *et al.* 1986, 2000). For example, there is evidence of northward subduction beneath the Eurasian margin (e.g. Pontides, northern Turkey), in the form of Late Palaeozoic granites (Ustaömer *et al.* 2005; Okay *et al.* 2006), and a Triassic volcanic arc (Çangaldağ: Ustaömer & Robertson 1997). In most of these earlier interpretations the subduction zone was located close to the Eurasian margin, whereas the continental terranes further south (e.g. Pelagonian Zone, Serbo-Macedonian Zone, Anatolide platform) were seen as fragments rifted from Gondwana, as discussed previously. However, Stampfli *et al.* (2001, 2003) position the Permo-Triassic subduction zone south of these

continental units. In this interpretation the continental fragments formed part of the southern margin of Eurasia by Late Palaeozoic–Early Mesozoic time (they probably rifted from Gondwana at a much earlier stage, but this is not discussed here).

The northward-dipping subduction model (Stampfli *et al.* 2001, 2003; Stampfli & Borel 2002) has two key testable implications. The first is that a Late Palaeozoic–Early Mesozoic northward-dipping subduction zone was located in the southern Aegean region (Fig. 32c), and between the Anatolides and Taurides in southern Turkey. The second is that northward subduction was compensated for by the opening of a wide, southerly ‘Neotethys’ during Late Permian–Triassic time (Fig. 33c).

These implications have been tested by detailed fieldwork in the Turkish and Cyprus areas, and have lead to the conclusion that, first, there is no preserved evidence of a north-dipping subduction zone within or adjacent to the Tauride–Anatolide units in southern Turkey (Robertson *et al.* 2004). Secondly, unlike Oman, there is no evidence of Permian sea-floor spreading in the Eastern Mediterranean, based on a combination of marine and land-based studies (discussed earlier in this paper). For the critical westerly areas (i.e. Crete, southern Peloponnese, Evia and western Sicily) the main conclusion is that rifting rather than subduction/accretion characterized the Late Palaeozoic–Triassic history of the south Aegean region (e.g. Crete and Peloponnese; Fig. 35b), where there is no evidence of early Mesozoic ophiolites, metamorphism or contemporaneous arc volcanism (Robertson 2006b). In addition, there is no evidence of sea-floor spreading in the south–Aegean during Permian time (see earlier discussion). Other objections to northward-dipping subduction in a southerly location (south Aegean) are similar to the southward-dipping subduction model, and include the need for the formation of multiple marginal basins (e.g. Pindos and Vardar), yet with an absence of coeval magmatic arcs, fore-arc basins and accretionary prisms. Notably, the northern margin of the Vardar Zone records a rifted margin without a subduction influence or arc-derived volcanoclastic sediments. Also, in Greece, calc-alkaline plutonic rocks within the Pelagonian micro-continental unit are Late Palaeozoic in age, much older than the Triassic rifting (Mountrakis 1986; Kotopouli *et al.* 2000).

Finally, an even more complicated scenario, one involving multiple marginal basins, has been suggested based mostly on igneous geochemical evidence. In this, a Late Palaeozoic–Early Mesozoic subduction zone dipped northwards (Pe-Piper 1982), as in Stampfli *et al.*’s (2001) interpretation, and an additional Triassic subduction dipped southwards from an inferred Pindos back-arc basin

(Pe-Piper & Piper 2004). This model was introduced to explain the igneous geochemical evidence from the Triassic rift basalts, but runs into the same difficulties as with the other subduction-related model, especially the absence of evolved magmatic arcs and accretionary prisms of the appropriate age and location.

The geological evidence therefore does not support formation of the Triassic rift basins on the Eastern Tethys as marginal basins above either northward- or southward-dipping subduction zones. However, an outstanding problem is the tectonic setting of some of the Triassic rift volcanics in the Balkan region (eastern Crete, Peloponnese, Albania, Serbia, Eastern Alps, Dolomites). Geochemical plots show considerable variation in chemical composition within and between the suture zones in Greece, with no simple pattern emerging (e.g. MOR-type basalts in one suture v. subduction influenced in another) (Fig. 36). Many of the basalts in both the Pindos and the Vardar zones show a clearly expressed subduction geochemical signature (e.g. Pe-Piper & Piper 2004). An alternative suggestion to Triassic active subduction is that the subduction influence was inherited from earlier subduction in the region (Robertson & Dixon 1984). A possible explanation

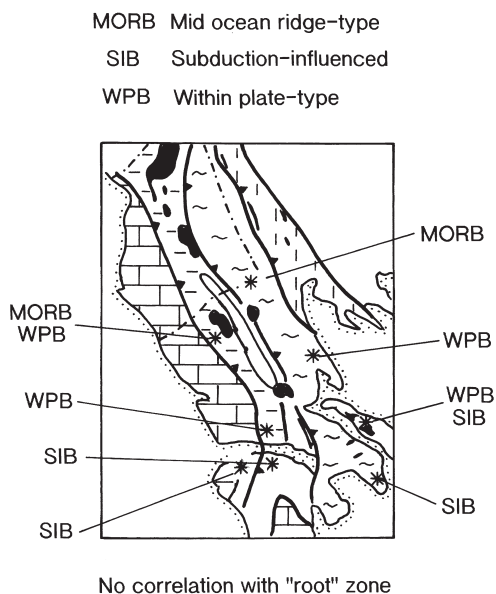


Fig. 36. Summary of the inferred tectonic settings of Triassic rift basalts analysed at Edinburgh University, School of GeoSciences. Samples were analysed over a number of years by S. A. M. Brown, P. D. Clift, P. J. Degnan, J. E. Dixon, G. Jones, A. H. F. Robertson and I. R. Sharp. An additional large dataset was published by Pe-Piper & Piper (2004).

is that southward subduction beneath the north margin of Gondwana was active during the Late Palaeozoic Hercynian 'orogeny' and that the subduction signature of Triassic basalts relates to melting of subduction-contaminated lower crust and upper mantle. These volcanics could represent small melt fractions that were strongly influenced by the composition of the underlying crust and mantle lithosphere (Dixon & Robertson 1993). The south Aegean and Balkan region was characterized by calc-alkaline magmatism (e.g. extensive granite extrusion and andesitic volcanism) during Late Palaeozoic time, related to the Hercynian orogeny (Mountrakis 1986; Kotopouli *et al.* 2000). This pre-existing magmatism was possibly the source of the subduction signature in some of the Triassic rift-related volcanic rocks rather than contemporaneous Triassic subduction.

During the Upper Palaeozoic (Early Carboniferous) subduction was perhaps both northwards and southwards, i.e. beneath both Gondwana and Laurasia (Eurasia). This culminated in complete closure of the 'Hercynian' ocean in the west (Dercourt *et al.* 1986, 2000; Ziegler 1990; Okay *et al.* 2006), whereas Tethys (Palaeotethys) remained partly open further east. Subsequent subduction was northward-directed beneath Eurasia during the Triassic and the late Mesozoic–Early Cenozoic. A relict south-facing Hercynian slab could thus have remained along the north Gondwana margin and, if so, was reactivated during Triassic rifting, resulting in subduction-influenced volcanism as in some, but not all, areas of the Balkans (e.g. eastern Crete but not western Crete; southern Pindos but not northern Pindos, etc.) (Fig. 36).

In summary, the Eastern Tethyan rifts cannot be considered as subduction-related back-arc marginal basins; where present, a subduction influence is seen as being relict from previous, probably 'Hercynian', subduction.

Comparisons between the Eastern Tethyan and Alpine–North Atlantic rift settings

Taken together, each of the rifted margins summarized here, Oman, the Himalayas and the Eastern Mediterranean, show many features in common, as follows.

- Rifting did not take place in the midst of a continent like the Atlantic or the Western Tethys (Alps), but was located along the northern margin of the pre-existing supercontinent, Gondwana, facing outwards into an older ocean (Palaeotethys).
- Rifting was pulsed prior to onset of sea-floor spreading. In Oman, the margin experienced pulses of rifting dating from the Carboniferous,

with initial spreading in the Late Permian. In the East Mediterranean (e.g. Crete) rift events along the North African–Levant margin occurred during Late Carboniferous–Permian and the Early–Middle Triassic, preceeding breakup and spreading during latest Triassic–earliest Jurassic time. Rifting and crustal extension in the Eastern Mediterranean had advanced sufficiently to form a deep-water basin with open-marine circulation during Late Carboniferous–Permian time, but there is no evidence of oceanic crust of this age, unlike in Oman. In the Himalayas, pulses of extension and related alkaline magmatism continued to affect the subsiding passive margin after continental breakup. Renewed volcanism, possibly rift-related, is also present in the Levant.

- Rifting was associated with widespread volcanism. In the Himalayas, Early Permian volcanism (Panjal Traps) affected a very wide area of the rifted margin. However, the thickness of volcanics decreases northwards towards the rift axis. Late Permian–Early Triassic rift-related volcanism also affected the south Aegean (e.g. Crete). Triassic rift-related volcanism affected large areas (e.g. Pindos rift in Greece; Antalya rift in SW Turkey). Taken together, the volcanics show a transition from intermediate-silicic composition lavas in proximal settings, to transitional MORB basalts in more oceanic settings. The volcanics appear to have extended right up to the edge of the rifted margin, as documented in SW Turkey (Isparta Angle), and thus the continent–ocean transition zone appears to have been quite narrow (kilometres rather than tens of kilometres).
- There is little evidence of the involvement of large-scale regional mantle plumes in the breakup of continents in the Eastern Mediterranean region. Large-scale domal uplifts and associated regional-scale deep erosion are not observed. Unconformities relate more to foot-wall uplift during pulsed rifting (e.g. Late Triassic of Crete). However, small-scale plume-type influences are consistent with the geochemical evidence. Also, the thick and regionally extensive Permian Panjal traps and associated regional uplift in the Himalayas are suggestive of a regional-scale plume influence.
- The rifted margins were palaeogeographically varied, ranging from oceanward-deepening ramps of deep-sea sediments above a volcanic basement (e.g. Lamayuru/Karamba Formation, Himalayas; Lycian nappes, NW Turkey) to more complex settings characterized by large marginal rift basins and off-margin carbonate platforms (e.g. Beyşehir–Hoyran nappes, central Turkey). In addition, several individual rifted margins

were bordered by rifted fragments up to tens of kilometres in size (e.g. Vardoussia, Greece; Kemer units in the Isparta Angle, SW Turkey).

- Along the margins of the ocean basins marginal highs originated as small platform units, rifted from the larger, parent platforms. Such fault blocks were capped by cephalopod-bearing pelagic limestones, and include manganese crusts of hydrothermal origin (Rustaq Exotic, Oman; Gödene Zone, Antalya Complex, SW Turkey). The rift blocks were isolated within marginal oceanic crust as spreading began. Such blocks include synrift/prerift facies in some areas (e.g. Jebel Qamar exotics, Oman; Kemer units; Antalya). The fault blocks were covered by shallow-water carbonates, then by pelagic carbonates. This facies change was triggered by tectonic subsidence, perhaps accentuated by eustatic sea-level rise. Exceptionally, in SW Turkey (Antalya), shallow-water carbonate deposition kept pace with tectonic subsidence or sea-level change for more than 100 Ma. This setting of reduced subsidence

probably relates to a setting of transform rifting or oblique rifting, and the local palaeogeographic setting of this area as an embayment of the Southern Tethys.

- True volcanic seamounts formed further out in the Tethyan Ocean and were overlain by thick, rapidly subsiding shallow-marine carbonate platforms, passing upwards into deeper water pelagic carbonates (e.g. Oman Triassic Exotics).

Western Tethys–North Atlantic

Following the (incomplete) drilling of the Iberia–Newfoundland conjugate transect, the rift/breakup history of the Central Northern Atlantic can now be correlated with the Western Tethyan region (Manatchal *et al.* 2007; Tucholke *et al.* 2007). Together, these regions provide the type example of a Non-volcanic rifted margin (Fig. 37a). This margin type has several key characteristics. First, mantle lithosphere was exhumed along the opposing rifted margins (i.e. Iberia and Newfoundland in the Central Atlantic; Eurasian and African

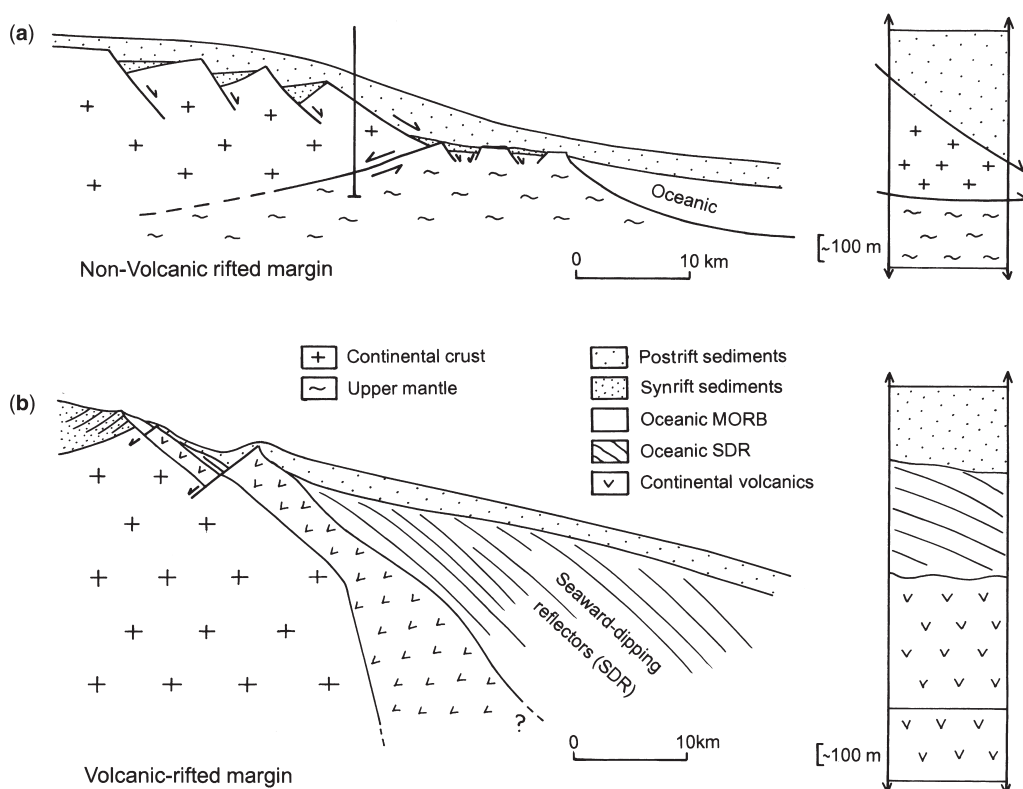


Fig. 37. Comparison of ideal (a) 'Non-volcanic' versus (b) 'Volcanic'-rifted margins, based mainly on ODP drilling in the north Atlantic. Much greater thicknesses of volcanic rocks are to be expected on volcanic-rifted margins, but without evidence of exhumed lower continental crust and mantle. See the text for data sources and discussion.

margins in the Alps). Secondly, the exhumed mantle is directly overlain by MOR-type basaltic volcanics (Tucholke *et al.* 2004), as discovered at ODP Site 1277 (Robertson 2007). Thirdly, the conjugate margins were, in turn, overlain by deep-water sediments, as the facies vary markedly under the influence of local palaeogeography. For example, the Newfoundland rifted margin was strongly influenced by the input of terrigenous clastic sediments that were locally derived from adjacent continental basement to the west and north (e.g. Avalon platform), whereas the sediments on the Iberia margin are more calcareous and pelagic owing to the trapping of clastic sediments by the rifted Galicia rifted basin to the east.

The evolution of the Western Tethys is now quite well understood. The central North Atlantic and the Piedmont–Ligurian Ocean in the Alpine region opened in concert. The Western Tethyan rifts, as exposed in the Alps, are characterized by the exhumation of lower crust–upper mantle and covering by MOR-type basalts (see Manatschal *et al.* 2007). This rift basin was clearly asymmetrical. The Eurasian margin of the Piedmont–Ligurian Ocean was relatively abrupt (upper plate) and was dominated by terrigenous sediments, whereas the conjugate Adria (Apulia) margin (lower plate) exhibits a greater abundance of pelagic sediments (e.g. radiolarites, pelagic carbonates) reflecting isolation of the rifted margin from the ocean by a wide rifted borderland (e.g. in the Dolomites, northern Italy).

On an oceanic scale, the Western Tethys can be seen as a gulf of the central North Atlantic. However, one uncertainty is whether only a single ocean basin existed in this region (Piedmont–Ligurian Ocean), as is likely, or if two subparallel oceanic basins co-existed (Stampfli & Marchant 1997). Despite the similar breakup history, the evolution of the Central Atlantic and Western Tethys eventually diverged. The North Atlantic continued to open, whereas the northerly Eurasian margin of the Western Tethys began to be deformed in latest Cretaceous time. This was related to closure of the most westerly of the Eastern Tethys ocean basins (Vardar Ocean), and was followed by complete closure of the Western Tethys during Early Cenozoic time.

Contrasting with the Non-volcanic rifted margin, Volcanic-rifted margins are mainly known from two legs of drilling of the Eastern Greenland margin (e.g. Larsen & Saunders 1998; Fitton *et al.* 2000) (Fig. 37b). The conjugate volcanic rifted margin (e.g. Vøring plateau) remains less well understood. Key features of volcanic-rifted margins are: first, seaward-dipping seismic reflectors originated as lava flows that were erupted as flood basalts. Secondly, there is a narrow continent–ocean transition zone (several kilometres), with MOR-type

oceanic crust reaching in close to the rifted margin. Thirdly, the rifted margin may be characterized by swarms of diabase dykes, as exposed in eastern Greenland. Fourthly, the volcanism is associated with regional thermal uplift, which leads to the development of a regional domed unconformity in the stratigraphic record.

From this summary it is clear that the Eastern Tethyan rifts differ markedly from either of the ideal ‘Non-volcanic’ or ‘Volcanic’ end members. In no case is there evidence of the exhumation of basement lower crust or upper mantle as in Non-volcanic margins (although preservation is always questionable). Also, rift volcanism is more extensive than on known Non-volcanic margins (although deep rift basins there remain largely unexplored). On the other hand, with the exception of the Himalayas, there is little signs of regional-scale plume-type magmatism and thermal doming. In general, volcanism is sparse along the Tethyan suture zone extending through Asia, as noted by Şengör *et al.* (1991).

The Eastern Mediterranean rifts in particular appear to record a different setting to either the ideal ‘Non-volcanic-rifted’ or the ‘Volcanic-rifted’ types.

Pure shear versus simple shear models

The evidence from drilling in the Iberia–Newfoundland conjugate shows that detachment faulting and exhumation play an important role in rifting, comparable to the Basin-and-Range of Western USA. However, drilling and geological observations of ancient margins cannot distinguish between these two alternative mechanisms and other techniques, mainly geophysical, must be used. Future drilling of the S reflector off Iberia, a possible simple shear detachment fault, is clearly a high priority for the IODP (Integrated Ocean Drilling Program). Some authors have argued that inherited crustal structure plays a key role in controlling the mechanism of rifting; it is suggested that pure shear dominates when rifting is at a high angle to the inherited tectonic fabric (i.e. zones of crustal weakness) but that simple shear results when the rifting is subparallel to the inherited grain (Ziegler & Cloetingh 2004). However, this assumes that the mechanism of rifting, by pure or simple shear, can be determined unambiguously, which is not the case in even the well-documented Iberia–Newfoundland conjugate. Recent modelling studies suggest that depth-uniform lithosphere stretching can replicate the structure and subsidence histories of intracontinental rifts but that depth-dependent stretching plays an important role during final continental breakup, with the lower

crust and mantle lithosphere being extended much more than the upper crust (see Kusznir & Healey 2006; Kusznir & Karner 2007). This situation is well documented, for example, for the east Brazil margin (Karner pers. comm. 2005).

A related approach has been to use the stratigraphic record of rifted margins to calculate subsidence profiles and thus attempt to determine the timing and processes of rifting (e.g. Brunet & Cloetingh 2003). The calculated shapes of the subsidence curves differ for pure shear and simple shear rifted margins during the postrift phase, thereby allowing a theoretical distinction using the stratigraphic record (Watts 1982; Ziegler *et al.* 2003; Karner *et al.* 2007). The subsidence curves mainly utilize shallow-water successions deposited on the more proximal rifted margin segments that remained relatively undeformed. In contrast, the deep-water units formed within the continent–ocean transition zone are generally allochthonous, deformed or metamorphosed, limiting their value for subsidence analysis.

Utilizing, this approach Wooler *et al.* (1995) calculated subsidence curves for a number of rifted margins of both the Western Tethys (e.g. Apennines) and the Eastern Tethys (from Greece to Oman). They concluded that there was little evidence that rifting and continental breakup were related to the activity of large-scale mantle plumes ('active rifting'), especially since major domal unconformities appeared to be absent from the areas studied. This situation contrasts with, for example, the Jurassic of the North Sea, a prime candidate for rifting associated with a mantle plume (Ziegler 1990; Underhill & Partington 1993). Wooler *et al.* (1995) emphasized the important role of rifting during Triassic time. The calculated subsidence curves suggested that extension occurred even in some marginal platforms where rift-related faults are not obvious in the field (e.g. Triassic of the Dolomites). However, such rift faults may be obscured, for example by slope facies; also many (or even most) rift faults were later inverted and developed into thrusts, again obscuring pre-existing rift normal faults. Wooler *et al.* (1995) argued that a dominantly pure shear rift model could explain the subsidence curves of the Tethyan rift units they studied.

Stampfli *et al.* (2001) carried out a comparable study, utilizing subsidence curves calculated from evidence of rift units in the Himalayas, Oman, Zagros and the Eastern Mediterranean regions. These authors argued that major rift-related subsidence began throughout the Eastern Tethyan region during Carboniferous time, with further widespread rifting during the Permian. They proposed that sea-floor spreading began during Late Permian time throughout the South Tethyan

region, including the Eastern Mediterranean (based on the presence of inflections in the calculated subsidence curves). They also argued that the subsidence curves obtained by them could be modeled using a simple shear rift model (e.g. for Oman).

Subsidence curves provide valuable insights on rifting, especially the timing of initial rifting and the role of extension-controlled subsidence, including areas where rift faults are largely absent, not exposed or converted into emplacement thrusts. However, they have a number of limitations. (1) Subsidence curves may not provide a good guide to the timing of final continental breakup, especially where a breakup unconformity is either subdued (e.g. Newfoundland margin) or apparently absent (e.g. South American margin). (2) Subsidence curves are influenced by whether a depth-dependent or a depth-independent stretching model is adopted, which cannot be assumed *a priori* for any given setting (see Kusznir & Karner 2007). (3) Subsidence curves can only be properly modelled where the geometry of the complete conjugate margin pair is known, showing whether rifting was symmetrical or asymmetrical. Even in the modern ocean (e.g. Iberia–Newfoundland) this information remains incomplete. For the Alps, asymmetrical rifting is well established from the stratigraphy (see Manatschal *et al.* 2007). However, in the ancient Eastern Tethyan examples discussed here, only one margin is readily available for study (e.g. Oman).

In summary, subsidence curves alone do not permit a convincing distinction between either pure shear or simple shear models, and additional constraints are needed, for example, from geophysical evidence.

Rift driving mechanisms

Based on a combination of the global data base and modelling, the main controls of rifting have recently been considered as a combination of plate-boundary forces, frictional forces acting on the base of the lithosphere (as asthenosphere convects) and the tensional stress (deviatoric tension) that develops above upwelling asthenosphere (Ziegler & Cloetingh 2004). Modelling studies suggest that continental breakup and the onset of sea-floor spreading may relate to an upwelling, divergent flow field within asthenosphere and continental lithosphere, implying that upwelling flow may be localized beneath rifts and zones of continental breakup (see Kusznir & Healy 2006; Kusznir & Karner 2007). The relation, if any, of such upwelling flow to 'mantle plumes' is currently controversial. The main role of plumes may be to facilitate crustal weakening rather than as a primary rift/

breakup driving mechanism (Ziegler & Cloetingh 2004).

For the Eastern Tethyan rifts, more specifically, several driving mechanism can be briefly considered as follows.

- *Asthenosphere convection* exerting forces on Gondwana lithosphere leading to rifting. This is likely to be a key long-term process, but the forces exerted may be insufficient to cause breakup and sea-floor spreading (Smith 1999). The convection cells are assumed to lie beneath the African continent, rather than specifically at the margins of Gondwana.
- *Plume-related magmatism* (i.e. 'active rifting': White & McKenzie 1995). The South Tethyan margins were all volcanically active and, thus, could have been influenced by the activity of mantle plumes, especially for the Early Permian Panjal Traps, northern India. There is little evidence for a string of plumes rising along the Gondwana margin, which would be expected if this were the dominant rift process. On the other hand, strain weakening related to rift volcanism, whether or not plume related, is likely to have played a role in final continental breakup (Hopper & Buck 1996). Once rifts crust becomes magmatically active strain weakening may allow continental breakup to proceed (Hopper & Buck 1996). A model of melt-assisted rifting has recently gained support from seismic anisotropy studies using shear-wave splitting in the Northern Ethiopia rift (Kendall *et al.* 2005).
- *Slab-pull*. This could relate to subduction beneath the Eurasian margin, or elsewhere. There is, indeed, evidence from the Eurasian margin (e.g. Pontides, Caucasus) of northward subduction, HP–LT metamorphism and accretion (e.g. in the central Pontides, northern Turkey) during the Middle–Late Triassic when the Anatolian microcontinent was rifting from North Africa and drifting towards Eurasia (Okay 2000; Robertson *et al.* 2004). However, it seems unlikely that slab-pull, as a plate boundary force, could by itself initiate continental rifting (Smith 1999; see below). Also, such slab-pull would not propagate across an intervening spreading ridge, if present.
- *Ridge push*. This is not considered to be a significant plate boundary force in rifting (Ziegler & Cloetingh 2004).

Several other factors may have influenced the rift setting. The first is that the crust around Gondwana was already thinned and weakened by pulses of rifting and related magmatism, potentially dating from Carboniferous time, or even the time of pan-African continental accretion. Theoretical arguments suggest that crust within the continent–

ocean transition zone is relatively weak compared to full-thickness continental crust or oceanic crust (Vink *et al.* 1984; Steckler & ten Brink 1986; Lavie & Steckler 1997). Secondly, a spreading axis may have propagated around the Gondwana margin. A spreading centre appears to have formed adjacent to Oman in the Late Permian and then migrated westwards along the continental borderland of Gondwana. The propagating spreading tip reached the Eastern Mediterranean region by Late Triassic–Early Jurassic time, opening the South Tethyan oceanic basins. However, at present, the timing of spreading initiation in the large area between Oman and the Eastern Mediterranean (i.e. Zagros suture) remains poorly constrained and thus it is not known if spreading (i.e. continental unzipping) was progressive or episodic.

A migrating spreading centre can be compared with the westward-stepping migration of the Woodlark spreading centre during Late Miocene–Quaternary time (see Goodliffe & Taylor 2007). In this case, the spreading tip migrated through a broad rift zone, within crust that was previously deformed and thermally immature (Taylor *et al.* 1999a, b; Goodliffe & Taylor 2007). The spreading tip migrated westwards, episodically over several million years, a timescale that might not be currently detectable if applicable to continental breakup in the Eastern Tethyan region. Also, the spreading tip was unstable and jumped tens of kilometres or more northwards or southwards as it migrated westwards, thus, in effect, isolating continental fragments offset by transform faults.

The opening of the South Tethyan ocean basins bordering Gondwana can also be compared with the opening of the Arctic Ocean. In this region, an elongate continental fragment, the Lomonosov Ridge, more than 1500 km long by less than 150 km wide (comparable to the scale to the Tauride or Pelagonian microcontinents), rifted from Siberia during the Paleocene, opening the Arctic Ocean along the ultra-slow-spreading Gakkel Ridge (Lawver *et al.* 1990; Kristoffersen 1990). This rifting took place as a spreading centre migrated from the North Atlantic, potentially exploiting a pre-existing zone of crustal weakness in northern Eurasia. Similarly, the Eastern Tethyan rifts are seen as exploiting pre-existing weakened continental crust, following earlier deformation and coeval rift volcanism.

Working model of Eastern Tethyan rifting

The inferred timing of rifting is summarized in Figure 38. The Eastern Tethyan rifting differed fundamentally from the intracontinental breakup of the

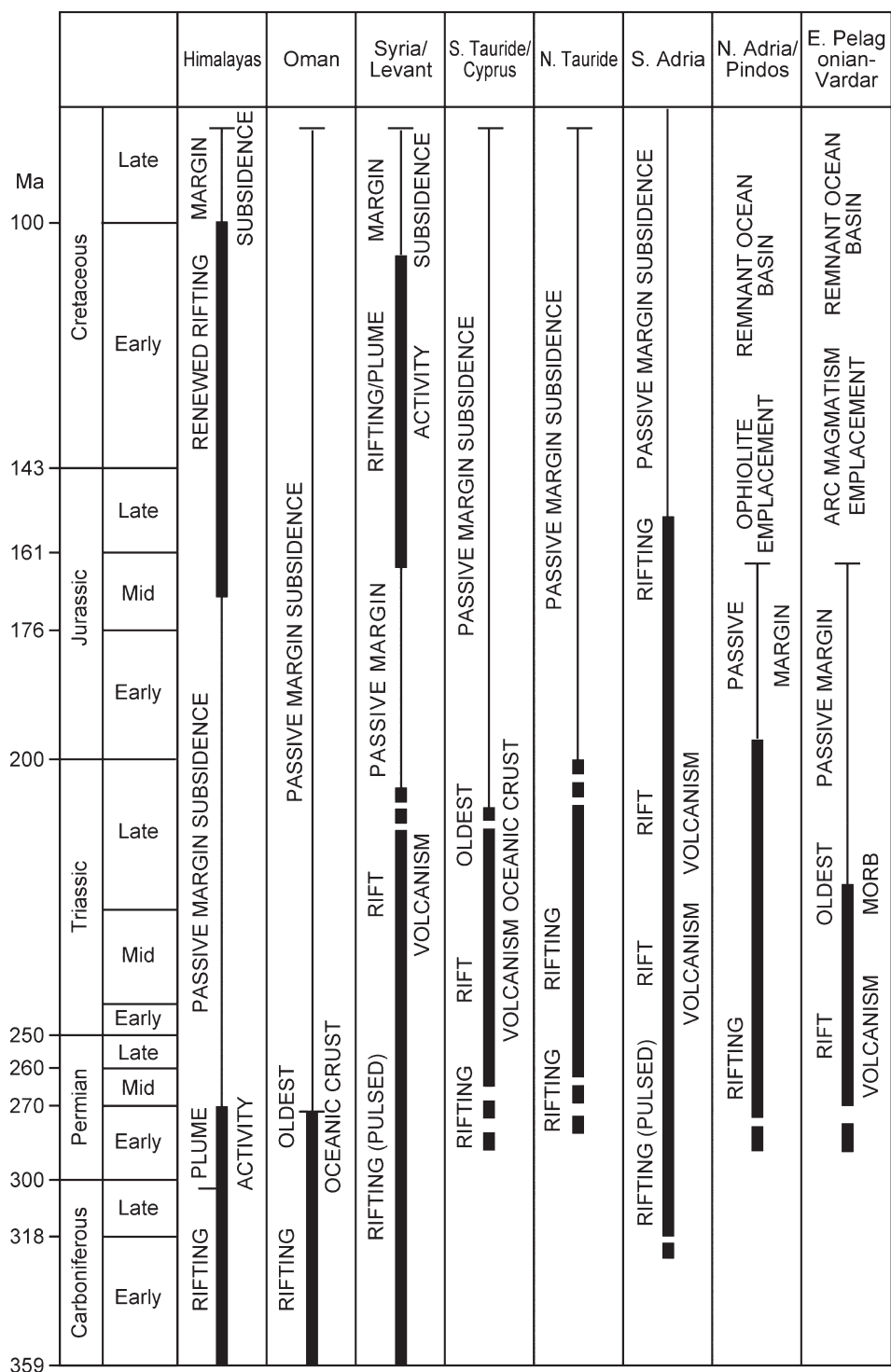


Fig. 38. Summary of the timing of rifting, continental breakup and sea-floor spreading inferred for areas of the Eastern Tethys discussed in this paper. Timescale: Gradstein *et al.* (2004).

North Atlantic (e.g. Newfoundland–Iberia conjugate) or Western Tethys (i.e. Alps). There was a strong tendency for the pre-existing Gondwana margin to fragment into microcontinents, which contrasts with the North Atlantic–Western Tethyan (Alpine) region where rifting took place amidst a pre-existing continent. The Eastern Tethyan breakup was characterized by the detachment of elongate crustal fragments from a continental margin adjacent to a pre-existing wide ocean (Palaeotethys; Fig. 39a, b).

The main forces promoting rifting can be considered as, first, the regional-scale mantle convection beneath Gondwana and, secondly, slab-pull beneath Eurasia, or elsewhere. These forces individually were capable of initiating rifting but probably not continental breakup. Breakup occurred, not randomly in time, but instead concentrated in the Late Permian in the east (Oman and Himalaya) and in the latest Triassic–earliest

Jurassic in the Eastern Mediterranean area. When breakup finally occurred it was quite drastic, initiating several subparallel small ocean basins at more or less the same time in the Eastern Mediterranean region.

Pulsed rifting affected the Northern Gondwana margin in the Eastern Mediterranean. Extension there was sufficient to promote adiabatic melting and extensive rift volcanism by the Early Triassic, probably facilitating final breakup by the process of strain weakening.

The breakup in the Eastern Mediterranean region corresponded to a time of major tectonic activity along the opposing Eurasian margin, where subduction, accretion and arc volcanism were taking place. If a spreading ridge had been subducted during this time, this could have accentuated slab pull on the southern margin and provided the extra force to trigger final continental breakup.

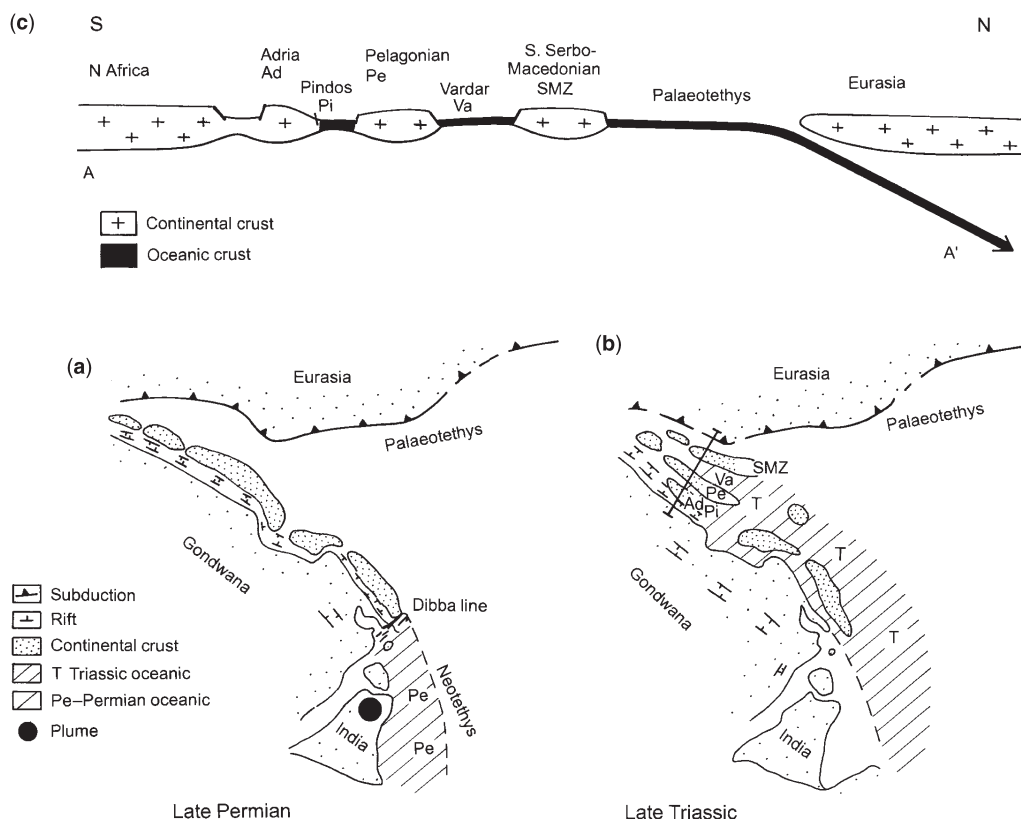


Fig. 39. Summary of the tectonic settings of rifting in the Eastern Tethys from the Eastern Mediterranean to the Himalayas. (a) Spreading adjacent to Gondwana in the Late Permian; (b) the spreading axis propagates into the Eastern Mediterranean region during the Late Triassic–Early Jurassic, isolating several microcontinents within oceanic crust (Neotethys); and (c) setting of microcontinents rifted from Gondwana in the Balkan region (Greece; Albania) during Late Triassic–Early Jurassic time. See the text for explanation.

Where present, plumes facilitated continental breakup, particularly by further weakening the crust. For the Himalayas, breakup (probably plume-assisted) took place in the Late Permian (Fig. 39a). Late Permian breakup also took place in Oman, but without evidence of a major local plume effect, although the chemistry of the volcanics is compatible with a plume influence.

By the latest Triassic–Earliest Jurassic passive margin subsidence was underway in the Eastern Mediterranean region (Fig. 39b). The resulting assemblage of continental fragments and small ocean basins is shown in Figure 39a–c. Final breakup in the Western Mediterranean took place later, coeval with the Jurassic opening of the central North Atlantic.

In the Himalayas, the rifted passive margin was later reactivated several times related to breakaway and the northward drift of India towards Eurasia. The Eastern Mediterranean rifted margins were also magmatically reactivated in some areas long after spreading had begun (i.e. in the Levant).

The Eastern Tethyan rift first underwent deformation related to convergent-margin tectonics in the Middle–Late Jurassic in the Balkan region and the Late Cretaceous in the ‘Turkish’ region, Oman and the Himalayas.

Conclusions

The rifted margins of the Eastern Tethys (Neotethys) are distributed throughout a number of suture zones, from the Eastern Alps to the Himalayas and beyond. These units provide a useful source of information on the anatomy of rifts and the processes of rifting and continental breakup. The uppermost levels of the slope–abyssal plain settings of rifted margins were typically detached during oceanic basin closure and emplaced onto continental margin as stacks of thrust sheets, overlain by ophiolites. Proximal rift/upper slope areas were commonly obscured by overriding thrust sheets, but are occasionally exposed, as in Oman. The leading edge of the rifted margin was typically underthrust or subducted (e.g. in NW Turkey and Saih Hatat, southern Oman Mountains).

All of the Eastern Tethyan margins were associated with rift-related volcanism and are thus dissimilar to ideal Non-volcanic rifted margins, as presently known (e.g. Iberia–Newfoundland transect; Western Tethys in the Alps). However, with the possible exception of the Himalayas, they are also dissimilar to Volcanic-rifted margins (e.g. Eastern Greenland), especially as voluminous plume-related magmatism and corresponding

regional updoming are generally absent. The Eastern Mediterranean rifted margins, therefore, exemplify a diversity of rift-related settings, and show that the current classification into ideal ‘Volcanic-rifted’ and ‘Non-volcanic rifted’ end members is oversimplified.

Rifting was typically pulsed, over up to 100 Ma (e.g. North African–Levant; Himalayas). The Himalayan margin was reactivated after spreading began, marked by faulting and extension-related alkaline magmatism, focused within the continent–ocean transition zone. The Eastern Mediterranean Levant margin was also magmatically reactivated, with extensive alkaline volcanism long after spreading began in the area. The production of extension-related melts as in the Eastern Mediterranean Triassic rifts is likely to have facilitated final continental breakup (controlled by strain weakening).

The extension of a previously weakened continental crust bordering less deformable oceanic crust acted to favour the detachment of elongate crustal slivers, hundreds of kilometres long by tens of kilometres wide. These fragments drifted northwards towards Eurasia as microcontinents, opening up small ocean basins in their wake. The major control on this long-lived northward drift was northward subduction beneath Eurasia.

On a smaller scale, rifting and continental breakup resulted in marginal fault-blocks (including prerift/synrift stratigraphy) being detached and isolated within marginal oceanic crust. These ‘highs’ were capped by shallow-water or pelagic facies controlled by the variable effects of subsidence versus eustatic sea-level change. Volcanic seamounts located further out in the oceans were capped by subsiding carbonate platforms (atolls).

Final continental breakup appears to have been triggered by a combination of long-term asthenosphere convection and plate boundary forces, especially slab-pull related to subduction beneath Eurasia. Where present, mantle plumes facilitated continental breakup.

Rifting processes and products are thus diverse in space and time, and much additional basic information is still needed from the modern oceans and orogenic belts to further test and constrain alternative tectonic models.

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