Tethyan oceans

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Abstract: Diachronous subsidence patterns of Tethyan margins since the Early Palaeozoic provide constraints for paleocontinental reconstructions and the opening of disappeared oceans. Palaeotethys opening can be placed from Ordovician to Silurian times and corresponds to the detachment of a ribbon-like Hun Superterrane along the Gondwana margin. Neotethys opening took place from Late Carboniferous to late Early Permian from Australia to the eastern Mediterranean area. This opening corresponds to the drifting of the Cimmerian superterrane and the final closing of Palaeotethys in Middle Triassic times. Northward subduction of Palaeotethys triggered the opening of back arc oceans along the Eurasian margin from Austria to the Pamirs. The fate of these Permo-Triassic marginal basins is quite different from areas to area. Some closed during the Eocimmerian collisional event (Karakaya, Agh-Darband), others (Meliata) stayed open and their delayed subduction induced the opening of younger back-arc oceans (Vardar, Black Sea). The subduction of the Neotethys mid-ocean ridge was certainly responsible for a major change in the Jurassic plate tectonics. The Central Atlantic ocean opened in Early Jurassic time and extended eastwards into the Alpine Tethys in an attempt to link up with the Eurasian back-arc oceans. When these marginal basins started to close the Atlantic system had to find another way, and started to open southwards and northwards, slowly replacing the Tethyan ocean by mountain belts.

There is still some confusion about what Tethys existed at what time (e.g. Şengör 1985). A consensus exists, however, regarding the presence of a mainly Palaeozoic ocean north of the Cimmerian continent(s): the Palaeotethys, a younger Late Palaeozoic–Mesozoic ocean located south of this continent – the Neotethys – and finally a Middle Jurassic ocean – the Alpine Tethys (Favre & Stampfli 1992; Stampfli & Marchant 1997), an extension of the Central Atlantic, which broke through the Pangea supercontinent. These three oceanic realms form the Tethyan domain s.l. extending from Morocco to the Far East (Şengör & Hsu 1984).

The subsidence history of these oceans to support this group’s proposed paleocontinental reconstructions is discussed here. These reconstructions have been done in the frame of the IGCP 369 project and the EUROPROBE-PANCARDI project to serve as a basis for discussion. Through the ongoing process of data collection these reconstructions have evolved and will, it is hoped, evolve further to give a larger consensus about their validity.

These reconstructions are presently displayed on the website in Lausanne (www.sst.unil.ch), focusing mainly on the western Tethyan realm and the Alpine domain. The arguments which led to the present state of these reconstruction are found in Stampfli et al. (1998 a, b, 2000) and Stampfli & Mosar (1999). Regarding the Alpine domain s.str., the reader is referred to Stampfli (1993) and Stampfli & Marchant (1997), which discuss the opening of the Piemont and Valais Oceans. The Late Variscan evolution of the western Tethyan realm is discussed in Stampfli (1996), a review paper with a large reference list about the southern Variscan domains.

Some definitions
The first geodynamically correct definition of the main Tethyan oceans, based on extensive field work in the Middle East, was given by Stöcklin (1974). He recognized a Late Palaeozoic–Triassic oceanic realm cutting through the epi-Baikalian (Pan-African–Gondwana) Platform and separating the Iranian Plate from Arabia – which he called Neotethys – and another older oceanic realm separating the Iranian epi-Baikalian (Panafrican) domain from the Variscan Turan domain to the north – which he called Palaeotethys.

Following this proposal, an investigation of the eastern Alborz Range was begun (Stampfli 1978), effectively defining it as a potential southern margin of Stöcklin’s (1974) Palaeotethys ocean. The opening of this Palaeozoic ocean was placed in Silurian time. At the same time, the ophiolites of Mashhad were...
recognized as most likely pertaining to the Palaeotethys suture [see the review of Ruttner (1993) concerning these ophiolites].

The drifting of the Irano–Afghan Block from Gondwana to Laurasia was then clearly recognized and constrained by the evolution of the microlflora of the Iranian Block from a Gondwanan affinity in Carboniferous time (Coquel et al. 1977; Chateauneuf & Stampfli 1979) to a Eurasian affinity in Late Triassic time (Corsin & Stampfli 1977). The Eocimmerian Orogeny was also defined in Iran at that time, as a result of the closing Palaeotethys and Middle Triassic collision of the Iranian Block with the Eurasian Turan Block (Stampfli 1978).

This concept was later extended further west (Turkey) and east (Tibet, Far East) by Şengör (1979, 1984); who defined the Cimmerian Block as a ribbon-like microcontinent separating Neotethys from Palaeotethys (Şengör & Hsu 1984), he also defined, at the same time, the Cimmerian deformation as non-Hercynian or post-Hercynian. Şengör’s definition of Palaeo- and Neotethys (e.g. Şengör 1989) is similar to Stöcklin’s (1974), with a major deviation which became clearer with time – Şengör viewed the opening of the Neotethys as the spreading of a back-arc ocean. This proposal implied that the Gondwana margin was an active margin in Perm-Carboniferous times and that margin would then belong to the Variscan domain s.l. [the Podataskasi zone of Şengör (1990, 1991)]. This assumption was based on an erroneous interpretation of the uplift and erosion of the Neotethys rift shoulders of northern India, Oman, Iran and Turkey as proof of Variscan deformation within the epi-Baikalian (Pan-African) domain (e.g. Oman – Michard 1982; Şengör 1990: India – Fuchs 1982; Bagati 1990: Turkey – Demirtaşlı 1984). The geometry of this Permian unconformity, its age compared to the rifting period, the geochemistry of associated basalts, the sedimentary record and the geodynamic context, imply a synrift thermal uplift and not a contemporaneous orogenic event, as clearly demonstrated in all these areas by thorough field work done in the last ten years (e.g. Mann & Hanna 1990; Pillevuit 1993; Vannay 1993; Garzanti et al. 1994, 1996; Pillevuit et al. 1997).

Similarly, Ricou (1974) and Braud (1987) never spoke of Variscan deformation or metamorphism concerning the Sanandaj–Sirjan Zone (the Iranian part of the Podataksasi Zone); they regarded the metamorphic rocks of this region as a retrogressed epi-Baikalian basement [this is also indicated on the 1: 1 000 000 map of Iran (Huber & Eftekhar-Nezhad 1978)]. Lower Permian limestones and volcanics rest on this Precambrian basement; they can be regarded as syn- to post-rift deposits of the northern margin of Neotethys. The development of greenschist facies metamorphic conditions in these areas may be related to the Permian Neotethyan rifting phase, or to younger intrusive events (Berberian & Berberian 1981) when these regions became part of the northern Neotethyan active margin.

Stöcklin’s (1974) original definition of Palaeotethys is therefore correct; it separates the Variscan domain from the epi-Baikalian (Pan-African) domain and its closure in Triassic times produced the Eocimmerian tectonic event which is always found south of the Variscan domain. The complete Triassic closure of Palaeotethys on an Iranian transect was proven later on by paleomagnetic studies (e.g. Schmidt & Soffel 1984; Lemaire 1997; Soffel & Förster 1984), confirming the conclusions reached previously based on floral and micrfloral distribution (e.g. Corsin & Stampfli 1977).

As proposed by Şengör (1979), the Cimmerian orogenesis was of collage type and never produced a large mountain belt. This can be explained by the presence of intra-oceanic arcs, back-arc marginal seas and oceanic plateaus, located between the Cimmerian and Eurasian Plates (see below), which strongly reduced the effects of crustal thickening. In many cases the Palaeotethys suture zone was used for the opening of Jurassic marginal oceans during the subduction of the Neotethys (e.g. Caspian Sea, İzmır–Ankara Suture), complicating somewhat the image one can reconstruct of the former geometry of the Palaeotethyan margins.

The reconstructions

Presented here are a set of maps, altogether with subsidence curves, for key times in the Palaeozoic and Mesozoic. The plate reconstructions were computed by the GMAP (Geographic Mapping and Palaeoreconstruction Package) program developed by Torsvik & Smethurst (1994). These maps are based mainly on a review of the following articles and books:

Most of the references concerning the palaeozoic evolution of the Alpine s.l. region have been taken from recent compilations, e.g.:

- Società-Geologica-Italiana (1979);
- IGCP project 5 – Flügel et al. (1987) and Sassi & Zanferrari (1989);
- IGCP project 276 – Baud et al. (1991b) and Carmignani & Sassi (1992);
- von Raumer & Neubauer (1993);
- IGCP project 369 (started in September 1994) – see web site www.sst.unil.ch or www.geomin.unibo.it/orgv/igcp/igcp.htm;
- PANCARDI project (a Europrobe project started in 1994) – see web site www.geofys.uu.se/eprobe/.

A Pangaea A fit is used since the Late Carboniferous (Stampfli 1996). The Pangaea B concept (Irving 1977) could be applied for the Early Carboniferous period, it would then slowly grades into a Pangaea A position towards the end of the Carboniferous. A Triassic Pangaea B, proposed by some palaeomagnetic studies (Muttoni et al. 1996; Lemaire 1997; Torq 1997), is certainly not supported by geological evidence, mainly regarding the 3000 km dextral strike-slip motion during the Triassic supposedly passing through Morocco. The feature generally used to transform the Pangaea B in A is the Tizi-n-Test Fault Zone of the High Atlas, but this has proven not to be a dextral but a sinistral Tertiary shear zone, and only of local importance (Jenny 1983). Also, the lasting marine sedimentation and calc-alkaline volcanism in southern Europe until the Late Carboniferous–Early Permian [see the review of field data in Stampfli (1996)] favour a Pangaea A model, because the Pangaea B model would imply a total closure of Palaeotethys up to the Caucasus before the Permian, but Late Carboniferous granites (De Bono 1998), or sedimentary sequences, of Greece (e.g. Phyllite–quartzite Group: Krahl et al. 1983, 1986; Krahl 1992) are not affected by Variscan metamorphism.

Prototethys

This as yet little known oceanic realm bordering Gondwana on its North African to Australian side in Late Proterozoic and/or Early Palaeozoic time will not formally be defined here. It could be characterized by the deposition of the ‘Sinian’ sedimentary cycle in many areas located in the vicinity of this ocean, as shown in Fig. 1 (Morocco, Arabia, Iran, India, China). Most of these areas are also affected by Pan-African deformation, followed by the deposition of a new cycle of sedimentation usually starting in Cambrian times. Was the Prototethys a mainly Late Proterozoic ocean or a mainly Cambrian ocean? This is still an open question. This group’s reconstructions suggest that Baltica–Siberia could have drifted away from Gondwana, opening the Prototethys in Early Palaeozoic time, a model also proposed by Torsvik & Eide (1998).

The Early Palaeozoic subsidence curves from Iran and India (Fig. 2; curves 2 and 3), possibly associated to the Prototethys thermal subsidence, could also be interpreted as resulting from the formation of a flexural foreland basin in view of the accelerating subsidence. This, together with other arguments presently in review, led this group to propose the accretion of an arc to the Prototethyan margin in Ordovician times. Followed here the idea of Şengör & Natal’in (1996) concerning the development at that time of a large intra-oceanic arc complex south of Siberia, the Kipchak Arc now forming the Kazakhstan Plate. This arc was extended to the south of Baltica and included in it were all the Alpine basement elements which comprise Ordovician granites, sometimes associated with remnants of oceanic crust or even eclogites of that age (von Raumer et al. 1993, 1998). The Rhei ocean is then viewed as the back-arc ocean located between Baltica and this Panalpine Arc. The Mauretanian ocean would open at the same time, possibly also as a back-arc basin due to the drifting away of Avalonia from the west coast of Africa. In view of its obduction onto the Baltica passive margin, the Iapetus ocean is represented here as a suprasubduction zone ocean opening at the expense of an older ocean.

Palaeotethys

The gentle docking of the ‘Pan-Alpine’ Ordovician arc was immediately followed by the southward subduction of the fast-spreading Rhei ocean. After subduction of its mid-ocean ridge, slab roll-back affected the remnant Rhei ocean and triggered the opening of Palaeotethys (Fig. 3). The ribbon-like continent being drifted away from Gondwana is therefore a composite terrane that this group term the Hun Superterrane [it contains most of the areas devastated by Attila!; see Stampfli (1996) and von Raumer et al. (1998)]. This Hun Terrane includes all the fragments accreted to Europe during the Variscan cycle and it extended eastwards to the Karakum (Turam) and Tarim areas, and possibly to the north China Block.
Fig. 1. Early Ordovician reconstruction, 490 Ma. The drifting of Baltica (including Ta, Taymir) and Siberia took place either from the south American side of Gondwana or the Indian side of Gondwana. Palaeomagnetic data do not provide enough constraints to decide on this. An intra-oceanic arc extends from Siberia to the south of Baltica [Ki, Kipchak Arc of Şengör and Natal’in (1996); and the Hun Cordillera terranes]. The opening of the associated back-arcs resulted in the seafloor spreading of the Rheic and Khanti-Mansi Oceans at the expense of Prototethys. The drifting of Avalonia (E-Av, W-Av) off the coast of western Africa probably opened the Rheic–Mauretanian Ocean at the same time. In Early Palaeozoic Gondwana is bordered by the following blocks. Hun Cordillera terranes: Early Palaeozoic active margin of the Hun composite terrane from west to east: OM, Ossa–Morena; Ch, Channel terrane; Sx, Saxo–Thuringian; Is, Istanbul; Po, Pontides; Li, Ligurian; Md, Moldanubian; MS, Moravo–Silesicum; He, Helvetic; sA, south Alpine; Pe, Penninic; AA, Austro–Alpine; Cr, Carpathian; Tn, north Tarim. Hun Gondwana terranes: blocks forming the northern margin of Palaeotethys, from west to east: Ib, Iberic; Ar, Armorica; Mo, Moesia; Ct, Cantabria; Aq, Aquitaine; Al, Alboran; Ia, intra-Alpine (Adria, Carnic, Austro–Carpathian); DH, Dinaric–Hellenic; Kr, Karakum–Turan; Pa, Pamirs; Ts, south Tarim; Qa, Qantang. The Cimmerian terrane: blocks forming the southern margin of Palaeotethys that were detached during the Late Permian opening of Neotethys, from west to east: Ap, Apulia s.str.; HT, Hellenides–western Taurides externides; Me, Menderes–Taurus; Ss, Sanandaj–Sirjan; Al, Alborz; Lt, Lut–Central Iran; Af, central Afghanistan; St, south Tibet; SM, Sibu Masu. Anamian blocks: defining the future northern and southern branch of Palaeotethys: nT, north Tibet; IC, Indochina and Borneo; sC, south China. Numbers 1–4 refer to the position of the subsidence curves of Fig. 2.
Subsidence patterns of selected areas along the Palaeotethyan margins show that the thermal subsidence of this ocean was diachronous, starting in the Early Ordovician from the east (Australia–India) to Early Silurian in the west (Turkey). From faunal (e.g. Robardet et al. 1994) and palaeomagnetic data, it appears that Avalonia (east and west, including the Brabant Terrane) drifted away from Gondwana before the Hun Terranes. In contrast, the Armorican Terrane remained close to Gondwana until the Ordovician (Perroud et al. 1984), when it accelerated its journey to collide with Laurussia in the Devonian. Therefore, Armorica s.l.
Fig. 3. Early Silurian reconstruction, 435 Ma. Baltica and Avalonia have been accreted to Laurentia–Barentsia (Ba) and the Pan-Alpine Arc accreted to Gondwana. The opening of Palaeotethys separated the Hun Superterrane from Gondwana. This opening took place in a context of diachronous back-arc spreading and slab roll-back of the Rheic Ocean. See Fig. 1 for the legend of the terranes. 1–4, Position of the subsidence curves of Fig. 2.

(Brioveria) is included in the Hun Superterrane and is placed north of North Africa based on palaeomagnetic data (see Torsvik & Eide 1998), which show a separation of Armorica from Gondwana not before Early Silurian.

Therefore, the Hun Superterrane is spread over a relatively large palaeolatitudinal area (from 60° south to the equator); large changes of facies are expected between Armorica and terranes now found in the Alps (e.g. Carnic and Austroalpine domain). For terranes located within the tropical zone (Alps, Spain, southern France) it can be shown that they present a very similar stratigraphic evolution to that of the Gondwana margin in Iran (Alborz) or Turkey (Taurus) from the Silurian to the Carboniferous; they are interpreted as representing the northern Palaeotethyan margin (Stampfli 1996). During the Carboniferous, terranes derived from the Hun Superterrane developed pelagic sedimentation followed by flysch deposits before being accreted to the Avalonia–Baltica margin. On the contrary, carbonate platform sedimentation continued along the Gondwana margin until the Namurian–Moscovian in North Africa (Lys 1986; Vachard et al. 1991), with
pelagic sedimentation persisting until Permian time in the Sicanian basin located north of Sicily (Catalano et al. 1988, 1992; Kozur 1990). East of a Palaeo-Apulian promontory [southern Greece, Turkey, Iran; see Vai (1991)], the carbonate platform lasted until Early–Middle Triassic. This shows a diachronous closure of Palaeoetothys from Moscovian to Early Triassic time from Morocco to Greece, along what was certainly a very oblique convergence zone.

The Variscan and Eocimmerian event and the marginal oceans

In Europe, the ‘Variscides collisional processes’ are generally regarded as extending from the Early Devonian to the Late Carboniferous, and the ‘Tethyan cycle’ (opening of the Alpine Tethys–Central Atlantic system) as not starting before mid-Triassic times. An apparent lack of major tectonic events during the Permian and Triassic southwestern Europe, or in the Appalachian domain, documents the welding of Gondwana with Laurasia to form the Permian Pangaea. But the Variscan domain extends over the whole Alpine area and even further in the Dinarides and Hellenides, northern Turkey and the Caucasus. It also extends in time, with deformations becoming younger, possibly grading into Eocimmerian (Triassic) deformations southwards and eastwards. As shown in Fig. 4, in the Late Carboniferous, the Palaeoetothys domain was not fully closed in southeast Europe and even persisted till the Early Triassic in the Hellenides [phyllite–quartzite Group; Krah et al. (1983, 1986), Krah (1992) and Stampfli et al. (2000)], and Middle Triassic times further East (e.g. Alborz Chain in Iran) (Fig. 5).

Stampfli et al. (1991) and Stampfli (1996) discussed this diachronous closure of the large Palaeoetothys ocean, insisting on the likely development of back-arc oceans or basins within the Permo-Triassic Eurasian margin. From the Palaeo-Apulian promontory eastwards it is quite clear that an Eocimmerian domain of deformation is found just south of a relatively undeformed Variscan domain, which is represented by a Late Carboniferous–Early Permian arc and clastic sedimentation of Verrucano type, mainly affected by extension. In contrast, the Cimmerian deformations are accompanied by Triassic flysch, mélanges and volcanics (Stampfli et al. 1995, 1998a) and even collisional-type intrusive events (Reischmann 1998), marking the closing of either Palaeoetothys or the marginal oceans.

As shown by the plate reconstructions (Fig. 4), the Permian margin of southeast Europe is of a transform type and little subduction took place along that margin at that time. However, the slab roll-back of Palaeothys rapidly induced ‘back-arc’ rifting along the whole margin after Late Permian times. East of the palaeo-Apulian promontory, this back-arc rifting graded into seafloor spreading of the Maliak–Meliata marginal ocean (Kozur 1991). The western end of the Meliata Rift (Southern Alps–Ivrea) aborted in Late Permian, whilst its eastern part (Maliak–Meliata–Dobrogea) spread [e.g. Early Triassic mid-ocean ridge basalt (MORB) pillow lava of N-Dobrogea; Niculitel formation: Cioflica et al. (1980), Seghedi et al. (1990) and Nicolae & Seghedi (1996)]. This Early Triassic seafloor spreading is accompanied by a marked thermal subsidence along the Pelagonian northern margin (Fig. 7, curve 6). At Meliata, the oceanic series are not older than mid-Triassic (Kozur 1991), but these oceanic remnants represent the accretion–obduction of the Meliata Ridge during its subduction, implying an older age for the onset of seafloor spreading.

In the Dinaro–Hellenide domain (Fig. 6), the Late Carboniferous arc (Pelagonia) collided directly with the Gondwana margin (Stampfli et al. 1995, 1998a; Vavassis et al. 1997). On the contrary, the Karakaya domain in northern Turkey (Şengör et al. 1980; Okay & Mostler 1994) is a complex Cimmerian deformation zone representing first the closure of Palaeothys between an oceanic plateau (Nilüfer Formation) and the Gondwana margin, then the closing of the Karakaya Back-arc and its northward thrusting on the Variscan Sakarya margin of Eurasia (see Okay 2000). Therefore, the Karakaya and Küre back-arc sequences (Ustaömer & Robertson 1994, 1997, 1999) are not remnants of the Palaeoetothys s.str. (see Kozur 1997b) but they represent a Mariana-type marginal ocean developing south of the Sakarya margin. During the Cimmerian collision event, the Late Permian–Early Triassic Karakaya–Küre back-arc ocean subducted southwards (Şengör et al. 1980; Tüysüz 1990; Pickett & Robertson 1996), but the former subduction of Palaeothys was to the north, under Eurasia, as clearly demonstrated in Iran (Stampfli 1978; Ruttnner 1993), Afghanistan (Boulin 1988) and in the Pontides (see Ustaömer & Robertson 1999).

Late Permian–Middle Triassic back-arc basins are also known eastwards in the Caspian (Nikishin et al. 2000), northeast Iran [e.g. the Agh–Darband sequence; Baud & Stampfli (1989) and Baud et al. (1991a)], northern
Afghanistan (Boulin 1988) and in the Pamirs (Khain 1994a, b; Leven 1995). Due to the collision of the Cimmerian blocks with the Eurasian margin (e.g. Stampfli et al. 1991; Alavi et al. 1997; Fig. 6), these back-arc basins disappeared during the Late Triassic. Along the Cimmerian orogen, development of a carbonate platform resumed in the Norian or Liassic, marking the end of this orogenic cycle. The Upper Late Triassic–Liassic deposits usually start with continental clastics derived from the Cimmerian elevations – they are found on both sides of the suture in Turkey – where they rest unconformably on older sequences [Karakaya or Sakarya basement in the Pontides – Koçyiğıt (1987) and Altnur et al. (1991): Triassic or Palaeozoic strata in the Taurus – Monod & Akay (1984)]. In Iran they are represented by the Shemshak Formation which also extends northwards on the Turan Plate (Stampfli 1978).

For the opening and oceanization of the Meliata Rift, the main geodynamic factor used is the thermal subsidence, which affected the whole Austro-Alpine domain and other internal
Fig. 5. Permian-Triassic boundary reconstruction, 248 Ma. The slab roll-back of the whole Palaeotethys induced the opening of back-arc oceans in the active Eurasian margin [Meliata, Karakaya (Ka) and Agh-Darband (Ag)] and the strong slab-pull force is opening the Neotethys Ocean. This opening is separating the Cimmerian Superterrane from Gondwana. See Fig. 1 for legend. 1-4, Position of the subsidence curves of Fig. 2.

parts of the Alps since the Permian-Triassic boundary, implying a rifting phase just before that time (Fig. 7). This subsidence induced the deposition of a more or less complete Triassic sequence, usually conformable on an Upper Permian clastic sequence, presenting a large diversity of facies and thicknesses locally approaching 3–4 km (e.g. Bernoulli 1981; Haas et al. 1995). This diversity is also linked to the counter-effect of the Eocimmerian deformation which induced local inversion or flexure of pre-existing basins, mainly in the southern margin of Meliata, and in the aborted part of the rift (e.g. subsidence curves 7 and 8 in Fig. 7). This Middle Triassic tectonic pulse is well established in the Dolomites where it is also accompanied by the emplacement of diapirs (Castellarin et al. 1996).

Without the presence of the Maliak-Meliata Ocean it would be difficult to explain the development of such a large-scale carbonate platform and marginal sequences before the opening of the Atlantic-Alpine Tethys system in the Early-Middle Jurassic. As can be seen from figure 7, this opening had no effect on all these peri-Apulian regions.
**Neotethys**

Although the geodynamic evolution of the Neotethys Ocean is now relatively well constrained, mainly through the recent findings of well-dated Wordian MORB in Oman (Pillevuit 1993; Pillevuit et al. 1997), its relationship with the East Mediterranean Basin is more debatable. The Neotethys separated the Cimmerian microcontinent(s) from Gondwana between the latest Palaeozoic and the earliest Tertiary, following a diachronous opening from east to west (Figs 2 and 5). Depending on the authors, the East Mediterranean–Ionian Sea basin is regarded as opening in the Late Palaeozoic (Vai 1994) or as late as the Cretaceous (e.g. Dercourt et al. 1985, 1993). Most people would regard this ocean as opening in Late Triassic or Early Jurassic (e.g. Robertson & Woodcock 1980; Garfunkel & Derin 1984; Şengör et al. 1984; Finetti 1985) and therefore possibly being related to the Alpine Tethys–Atlantic opening. However, as shown in Fig. 7, the Alpine Tethys opening has no tectonic or thermal effect on areas located around the Ionian Sea, although a Jurassic extensional phase is clearly recognized along the Levant Transform margin (Garfunkel 1998; Stampfli et al. 2000).

A new interpretation showing that the East Mediterranean domain corresponded to an oceanic basin since the Late Palaeozoic has been proposed by Stampfli (1989). Subsequently, new plate tectonic reconstructions considering this basin as part of the Neotethyan oceanic system have been developed (Stampfli et al. 1991, 2000; Stampfli & Pillevuit 1993) (Figs 5 and 6). This is supported by: (1) geophysical characteristics of the Ionian Sea and East Mediterranean Basin (isostatic equilibrium, seismic velocities, elastic thickness), excluding an age of the seafloor younger than Early Jurassic; (2) subsidence patterns of areas such as the Sinai margin, the Tunisian Jeffara Rift, and Sicily and Apulia s.str. (Stampfli et al. 2000) (Fig. 7), confirming a Late Permian onset of thermal subsidence for the East Mediterranean and Ionian Sea basins and the absence of younger thermal events; (3) Triassic MORB found in Cyprus in the Mamonia complex (Malpas et al. 1993), certainly derived from the East Mediterranean sea-floor, although a more exotic nature of these basalts (from the Neotethys) cannot be excluded; and (4) Upper Permian Hallstatt-type pelagic limestone, similar to those found in Oman where they sometimes rest directly on MORB (Pillevuit 1993;
Fig. 7. Circum–Apulia tectonic subsidence curves modified from Stampfli & Mosar (1999). The curves for Pelagonia are from De Bono (1998) and the timescale is from Gradstein & Ogg (1996).

Niko et al. 1996), have also been reported from the Sosio Complex (Kozur 1995) in Sicily. This Late Permian pelagic macrofauna presents affinities with both Oman and Timor, and implies a Late Permian direct deep-water connection of the East Mediterranean Basin with the Neotethys.
The Alpine Tethys, the Central Atlantic and the Vardar

Field work in the Canary Islands and in Morocco (Favre et al. 1991; Favre & Stampfli 1992; Steiner et al. 1998) has allowed the onset of seafloor spreading to be dated as Toarcian in the northern part of the central Atlantic. Similar subsidence patterns between this region and the Lombardian Basin (Figs 9 and 11) led to the proposal of a direct connection between these two areas (Fig. 8). The Lombardian Basin aborted (Bertotti et al. 1993) as it could not link up with the Meliata Ocean whose already cold oceanic lithosphere was rheologically unbreakable relative to surrounding continental areas. Therefore, the Alpine Tethys rift opened along the Meliata northern margin separating the future Austro-Carpathian domain from Europe. Thermal subsidence and spreading started in Aalenian time in the west (Fig. 9) [the Briançonnais margin of Stampfli & Marchant (1997) and Stampfli et al. (1998b)] and Bajocian time eastward [the Austro-Alpine margin of Froitzheim & Manatschal (1996) and Bill et al. (1997)]. The Alpine Tethys was linked to the Eurasian back-arc basins located further east through the Moesian-Dobrogean Transform.

A transform ocean linked up the central Atlantic and the Alpine Tethys delayed seafloor spreading and thermal subsidence in this Maghrebide Ocean is well exemplified by the subsidence curve from the Rif area (Favre & Stampfli 1992; Favre 1995) (Fig. 9). The rotation of Africa relative to Europe after the Late Triassic induced the subduction of the Meliata Ocean under the young Neo-Tethys oceanic crust. This intra-oceanic subduction gave birth to the Vardar Ocean, which had totally replaced the Meliata Ocean by the end of the Jurassic. The Vardar Ocean obducted southward onto the Pelagonian margin in the Late Jurassic (Fig. 7), then subducted northward under Moesia, finally opening the Black Sea in the Late Cretaceous, and it represents the third generation of back-arc opening in that region.

The North Atlantic and the Valais Ocean

Since the Early Cretaceous (Fig. 10), there has been no possibility for the Atlantic mid-ocean ridge to link up eastwards with another ocean. A last attempt was made through the opening of the Valais and Biscay Oceans, but by the end of the Santonian the break-up between North America and Greenland had taken place and then, in Campanian time, the Biscay Ocean aborted (Ziegler 1988a, b). Closing of the Valais Ocean took place during the opening

Fig. 8. Sinemurian reconstruction of the western Tethyan area. AA, Austro–Alpine; CA, Inner Carpathian; IzAnSi, İzmir–Ankara Ocean; MO, Moesia; Va, Vardar. ★, Magmatic arc activity. Palaeogeography modified after Ziegler 1988a. 9–21, Position of the subsidence curves of Figs 7, 9 and 11.
Fig. 9. Subsidence pattern of the Central Atlantic, Alpine Tethys system. Curve 10 (Préalpes Médianes Basin) is from Mosar et al. (1996); curves 11 (Lombard Basin) are modified from Greber et al. (1997); curve 12 (Rif) is from Favre (1995); and curves 13 have been calculated from the data of Ambroggi (1961), Adams et al. (1980) and Du Dresnay (1988). The timescale is from Gradstein & Ogg (1996).

of the Gulf of Biscay and subduction of the Alpine Tethys certainly started at the same time (Stampfli & Marchant 1997; Stampfli et al. 1998b).

Therefore, spreading of the Valais Ocean was short lived (Fig. 11). Its closure during the opening of the Biscay Ocean was only partial (there was no collision of the Briançonnais Peninsula with the Helvetic margin in the Late Cretaceous), but this event marks the onset of the inversion phase of pre-existing fault-bounded basins in many areas in Europe (Ziegler 1990).

Figures 10 and 11 show how different parts of southwest Europe or Iberia have been affected by these diverse rifting phases: the interpretation of these curves, however, requires a good knowledge of their initial position, some areas being mainly thermally influenced whereas others are more affected by faulting.
and extension of the upper crust. Generally speaking, all the areas shown on these figures have been affected by the opening of the North Atlantic and Valais Ocean system in Late Jurassic times.

Discussion – Tethys sutures in Turkey

Figure 12 shows a simplified tectonic map of the Aegean–Mediterranean area with the major sutures indicated as broad lines. The present back-stop of the East Mediterranean subduction zone (e.g. Cyprus, Crete) is regarded as the western extension of the Neotethys suture as found along the peri-Arabian ophiolitic zone (Ricou 1971) in Syria (e.g. Delaune-Mayere 1984), southern Turkey (e.g. Michard et al. 1984), Iraq and Iran (e.g. Ricou 1974; Braud 1987), and Oman (e.g. Bernoulli & Weissert 1987; Robertson & Searle 1990; Pillevuit et al. 1997). This obvious connection actually implies that the Neotethys ocean does continue into the East Mediterranean Basin, as discussed above, and provides a direct link between the pelagic Permain deposits from Sicily and their equivalent in Oman.

Actually, the Neotethys suture is composite and it contains the remnants of the southern passive margin of Neotethys (Hawasina Basin in Oman), as well as remnants of Cretaceous ophiolites (Semail Nappe in Oman) which obducted onto that margin. These peri-Arabian ophiolites correspond to the obduction of a younger intra-oceanic Cretaceous offspring of Neotethys (the Semail Ocean). Therefore, the northern margin of that Semail Ocean is the former northern active margin of Neotethys. Around the Arabic promontory (e.g. Maden Complex; Aktas & Robertson 1984) and in Cyprus (Robertson & Xenophontos 1993), this younger ocean closed during the Palaeogene. In Oman this ocean is not yet closed (Sea of Oman) and is still subducting under the Makran active margin.

As the East Mediterranean subduction zone (Aegean active margin) is certainly not older than Miocene, it does not represent the western continuation of the Neotethys active margin. In our structural scheme, a possible extension of this Late Cretaceous–Palaeogene active margin westward under the Lycian Nappes, linking up the Antalya Suture with the Axios Vardar Zone, is proposed. This connection is necessary to provide a western limit to the Greater Apulian Block (Apulia; autochthonous and para-autochthonous units of the Dinarides and Hellenides, Bey Dağları Block). This limit corresponds to the former transform that separated Neotethys from Meliata (Figs 6 and 8).

In this scheme, the İzmir–Ankara Suture represents a Jurassic back-arc ocean related to

Fig. 10 Santonian reconstruction of the western Tethyan area. 10–21, Position of the subsidence curves of Fig. 11.
Fig. 11. Circum Iberia subsidence curves, modified from Borel (1998), curves 14 and 15 are modified from Wildi et al. (1989). The timescale is from Gradstein & Ogg (1996).
the northward subduction of Neotethys since Late Triassic, this back-arc basin opened within the complex Karakaya–Palaeotethys Suture Zone. It followed the collapse of the Cimmerian Orogen and rescreened the Cimmerian Tauric–Menderes Block from Eurasia. This suture, characterized by its ‘coloured mélanges’ (Gansser 1960), can be followed up to the Iranian border. In Turkey, closure of the İzmir–Ankara Ocean postdates the Late Cretaceous ophiolite obduction found on its southern passive margin (Gutnic et al. 1979; Okay et al. 1996; Demirel & Kozlu 1998; Collins & Robertson 1998). Development of a new accretionary wedge along its northern margin is placed in the Middle Campanian and the final closure in the Eocene (e.g. Norman 1984; Koçyiğit 1991). Similar coloured mélanges are also found in Iran where they grade into Palaeogene flysch, followed by the main Eocene unconformity; they separate the Sanandaj–Sirjan Block from the central and north Iranian blocks (Stöcklin 1968, 1974, 1977, 1981). Therefore, the Cretaceous coloured mélanges in Turkey and Iran are clearly disconnected from the Neotethyan suture, and are located within the Eurasian margin to the north of the Neotethys active margin. In that sense, the İzmir–Ankara Suture is not the Neotethys suture.

The intra-Pontides (Okay et al. 1996), Meliata–Balkan Suture is of Cretaceous age and corresponds to the collision of a Vardarian arc with the Rhodope–Moesian Plate inducing the folding and northward thrusting of the Balkanide units (Georgiev et al. 1997; Tari et al. 1997). The Axios–Vardar Suture is the final Late Cretaceous–Palaeocene suture of the Vardar. The latter obducted onto the Pelagonian margin in Late Jurassic times – the front of this obduction is shown in Fig. 12. The Axios–Vardar Palaeogene Suture was somehow connected to the İzmir–Ankara Suture of similar age, although, the two domains were separated by the former Meliata–Neotethys transform.

In Fig. 12, the transported Palaeotethys suture in the Dinaro–Hellenides Zone is shown by a dashed line following the Budva–Pindos Zone under which it is hidden (De Bono 1998). The Palaeotethys Suture Zone in Turkey is possibly found in Chios and the Karaburun Peninsula (Kozur 1997a); further east it would be located in the vicinity of the İzmir–Ankara Suture Zone. Following Late Triassic or Jurassic opening of the latter, Palaeotethyan elements
could have been dispersed on both sides of the İzmir–Ankara Rift. So far, most Palaeotethyan elements have been described from the northern side of that suture.

Conclusions

Palaeocontinental reconstructions show that a large oceanic space (Palaeotethys) remained open south of the Variscides until Late Palaeozoic time. The Devonian–mid-Carboniferous collisional processes in Europe were related to accretion of the Hun Superterrane to Avalonia–Baltica. This terrane presents strong affinities with the Palaeotethyan passive margin sequences found, for example, in northern Iran (Alborz). The latter is regarded as representing the southern margin of this ocean, the Hun Superterrane the northern margin. This northern margin was separated from Gondwana in Ordovician–Silurian times. After the accretion of the Hun Superterrane to Europe, between the Late Devonian and Early Carboniferous, subduction jumped south into the Palaeotethys. Northward subduction of the Palaeotethys is responsible for Upper Carboniferous calc-alkaline intrusions and volcanism found everywhere in the Variscan Alpine domain. Closure of the Palaeotethys was achieved after the Namurian north of Africa but was diachronous going eastwards. East of a Palaeo-Apulian promontory, subduction continued into the Permian and generated the opening of the Maliak–Melia Oceans.

Concomitant Late Permian opening of the marginal Melia Ocean (within the Eurasian margin) and Neotethys (within the northern Gondwana margin) accelerated the closure of the Palaeotethys in the Dinaro–Hellenide region. Late Permian–Lower Triassic mélanges found in Greece point to a final closure of this Palaeozoic ocean at that time (Eocimmerian event). In Turkey, the collision of the Cimmerian terranes with the Eurasian margin was more complex due to the presence of large oceanic plateaux and Mariana-type back-arc basins between the two domains.

The Late Triassic–Early Jurassic intra-oceanic subduction of the Melia Ocean generated the Vardar marginal Ocean, obducted in Late Jurassic onto the Dinaro–Hellenic area. Its subsequent northeast directed subduction generated the collision of an intra-oceanic arc with the Austro–Carpathian and Balkanide areas in late Early Cretaceous times.

Figure 12 is a first attempt at showing the present-day location of the sutures of the above described oceans in the eastern Mediterranean area. It clearly shows that the Neotethys s.str. is located to the south of Turkey. The İzmir–Ankara Suture is not the Neotethys suture, it is the suture of a marginal ocean located within the Neotethyan active margin.

In a southeast transect of Europe, it can be seen how the Variscan Orogen evolved into Early and Late Cimmerian deformations. The rather clear situation found in the Appalachians cannot be extrapolated much further than western Iberia, where Laurentia and Gondwana collided in the Carboniferous–Early Permian. In the rest of Europe, this collision never happened as such. There was a collision between the Eurasian active margin and terranes derived from Gondwana. One has to wait for the anticlockwise rotation of Africa in the Cretaceous to see a collision between Europe and Africa, giving birth to the Alpine Orogen.

I would like to acknowledge helpful input from G. Borel, C. Steiner and A. De Bono for the subsidence curves for the Alpine and Hellenide regions, and Jon Mosar for the elaboration of the palaeostructures. This is a contribution to the FNS project 2000-53646.98 ‘Geodynamics of Tethyan margins’. A. H. F. Robertson and A. I. Okay reviewed this manuscript, I thank them for their constructive remarks and for sharing their knowledge of the Mediterranean area with me.

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