ELSEVIER

Tectonophysics 252 (1995) 253–267

TECTONOPHYSICS

Geological remarks on the relationships between extension and convergent geodynamic settings

Carlo Doglioni

Centro di Geodinamica, Università della Basilicata, 85100 Potenza Italy

Received 18 January 1994; accepted 3 November 1994

Abstract

In convergent geodynamic settings, extensional and compressional tectonics occur contemporaneously or in different times and places, but they are deeply linked to each other in a number of different ways. The main types of extensional regimes here considered are for west-dipping subduction zones: (1) back-arc extension (e.g., Tyrrhenian Sea); (2) uplift and related extension due to asthenospheric wedging (e.g., Apennines); (3) extension at the subduction hinge (e.g., Marianas); (4) extension due to increasing length of arcs (e.g., Apennines). For east-, northeast- or north-northeast-dipping subduction zones: (5) extension induced by collapse of orogens (e.g., Alps, Himalayas); (6) extension induced by differential lithosphere–asthenosphere coupling in the hanging wall of a subduction zone (e.g., Aegean rift); (7) rifting induced by orogenic roots anchored into the mantle (e.g., Atlantic and Tethys). Other types may be considered: (8) inversion of plate motion, from convergent to divergent relative plate vectors due to external velocity fields (e.g., Basin and Range); (9) transtension may occur along oblique ramps of thrust arcs due to differential forward propagation of thrusts planes (e.g., southern arm of the Barbados accretionary prism?); transtension and transpression also occur contemporaneously along an undulate strike-slip setting, or extension can accompany foreland dipping duplexes; (10) apparent extension due to down-section cut of thrust planes along subduction zones. Each type of extension generates different shape, depth and relationships between hanging wall and footwall of the associated normal faults and shear zones.

1. Introduction

Much debate has arisen over the last few years concerning the extensional tectonics observed in many orogenic belts (e.g., Coward et al., 1987; Jolivet, 1993). The extension processes described in the Alps, Mediterranean in general, Cordillera and Himalayas have been taken as a mechanism for the uplift of metamorphic core complexes (Lister et al., 1984; Platt, 1986, see a discussion in Michard et al., 1993), and have also been proposed as a mechanism for the collapse of several orogens (e.g., the Himalayas, Dewey, 1988; the Pannonian Basin, the Basin and Range, or the Apennines, Carmignani and Kligfield, 1990). The aim of this paper is to compare and discuss the main geologic framework between extensional and compressional tectonics in different geodynamic contexts. In particular it addresses the different origins of the tensional environments occurring contemporaneously or after compression in different geologic settings.

Recently, it has been suggested that the average westward plates motion (Le Pichon, 1968) is controlled by lateral heterogeneities in the upper mantle

SSDI 0040-1951(95)00087-9

(Ricard et al., 1991; O; Nelson and Temple, 1972; Uyeda and Kanamori, 1979; Doglioni, 1993a, b; Moser et al., 1993; Smith, 1993). A few general consequences of this relative motion are here proposed in the differentiation of extension types. Particular attention will be given to the differences between extension related to W-dipping versus E- or NE-dipping subduction zones. It will be shown that the different types of extension in convergent settings have different origins and different relationships with the compression. Each type of extension is characterized by different depths and shapes of the related decollement planes. A few cartoons are presented to illustrate the main geodynamic differences between the several types of extensional environments.

2. Back-arc extension

The opening of back-arc basins is peculiar to W-dipping subduction. They open very quickly at average rates of 3–7 cm/yr (Mariana Arc, Hussong and Uyeda, 1981; Tyrrhenian Sea, Moussat et al., 1985; Kastens et al., 1988) up to rates of more than 10 cm/yr (Bismark Sea, Taylor, 1979), and they generate new oceanic crust. They have a semicircular shape. The present back-arc basins such as the Caribbean Sea, New Scotia Basin, Parece Vela Basin, Banda Sea, Japan Sea, Bering Sea, Tyrrhenian Sea and the Pannonian Basin show a recent evolution and they all are associated with 'W'-dipping subduction. Accepting that the lithosphere is moving westward in relation to the asthenosphere (Le Pichon, 1968; Ricard et al., 1991), W-dipping subduction generates an obstacle to the related eastward–northeastward mantle counterflow. In fact all these subduction zones have an arcuate shape, suggesting they are obstacles to an opposite flow. Back-arc opening in this view is the consequence of the loss of lithosphere due to subduction opposing the relative mantle flow (Doglioni, 1991). In fact, subduction zones following the mantle flow, dipping to the E, NE or NNE, show very different characters and not typical back-arc basins (see later type 6). Extension in the back-arc may be considered as type 1 (Fig. 1). The western margin of the back-arc is more linear with respect to the eastern one which is an arc migrating and verging 'eastward'. Normal faults and ductile shear zones associated with back-arc extension flat out at the base of the lithosphere.

While the back-arc extension grows eastward, the back-arc basin also shows thermal subsidence (e.g., in the Tyrrhenian Sea, Kastens et al., 1988; Pannonian and Alboran basins, Royden et al., 1983; Morley, 1993). The extension in the back-arc basin regularly exhibits a discontinuous eastward propagation, generating a large-scale boudinage, as evidenced by the undulations of the crustal thickness. This type 1 extension occurs while compression is active to the east in the accretionary wedge. In contrast with linear rift zones (e.g., Atlantic passive margins), back-arc extension regularly shows eastward polarity of propagation, semicircular or arcuate geometry and shorter life (10–50 Ma). With the appearance of oceanic crust in the back-arc basin the classic switch from rifting to drifting stage may occur, with the transition being diachronous and propagating toward the northern, eastern and southern terminations of the arc. Higher tectonic and thermal subsidence rates characterize the back-arc basins with respect to classic linear continental and oceanic rift zones (Kobayashi, 1984; Morley, 1993), as demonstrated by Kastens et al. (1988) in the Tyrrhenian Sea where 700 m/Ma of subsidence have been calculated for the Marsili Basin.

West-dipping subduction zones may be distinguished according to whether there is active convergence between the two plates (e.g., the western Pacific subduction zones, Aleutians, Japan, Marianas, Fiji, Nankai), or if there simply is the roll-back of the subduction hinge without convergence among the two plates (e.g., Barbados, Apennines, Carpathians and Banda arcs, due to slab pull or eastward mantle push). The first type of west-dipping subduction may be considered 'active' in which two reference points between the subducting plate and the western margin of the back-arc basin are converging. The second type may be considered as 'passive' in which there is no convergence between the two reference points of the subducting plate and the western part of the back-arc basin. In fact, once west-dipping subduction is activated, the 'eastward' mantle flow provides the push for the eastward roll-back of the subduction zone, either 'active' or 'passive'. In the first type two converging plates are involved (active), whereas in the second type the west-dipping subduction develops within one single plate (passive).

3. Uplift and related extension due to asthenospheric wedging

While extension in the back-arc setting is accompanied by subsidence, the extension in the ridge above the subduction hinge is coeval to uplift (e.g., Apennines). Rates of uplift in the Apennines are in the order of 0.5–1 mm/yr (e.g., Cosentino and Gliozzi, 1988). The kinematics of W-dipping subduction seem to predict mantle wedging at the top of the subduction hinge (Fig. 1), with omission of the lithospheric mantle; this should induce uplift and extension of the overlying crust (type 2 of extension, Fig. 1). This crust is composed mainly of shallow rocks offscraped from the top of the subducting lithosphere, and metamorphic rock relicts of the former Alpine-type E-dipping related subduction. In the Apennines, the sedimentary cover extends to depths of at least 15–20 km, with a new 'Moho' indicated at 25–30 km. Then it is necessary to question what has happened to the crystalline crust that usually is 20–30 km thick. During W-dipping subduction we might kinematically interpret that this crust is mainly lost by subduction due to different behaviour of the decollement planes in the W-dipping subduction zones with respect to the E- and NE-dipping ones (Doglioni, 1992). The superficial

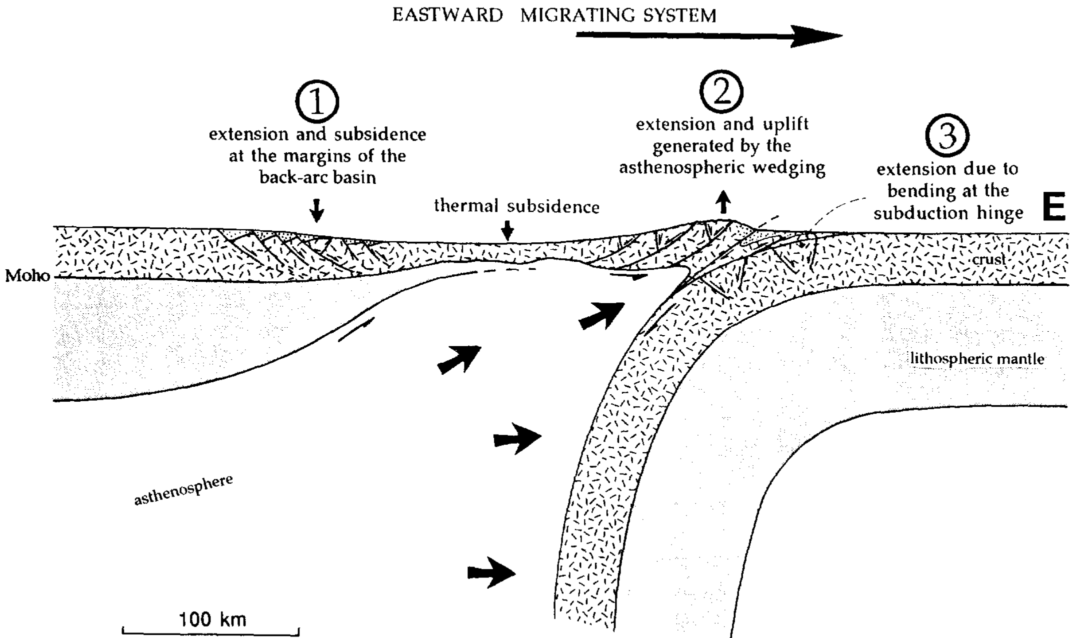

Fig. 1. Three different types of extensional environments may be observed in a section of a W-dipping subduction system. Type 1 is the extension generated by horizontal stretching during the back-arc opening; in a map view the western margin is generally linear while the eastern one is arcuate. In contrast with type 1 which occurs with general subsidence, type 2 is an extension coeval with uplift and may be interpreted as due to the bending due to the upward push generated by the asthenospheric wedging at the subduction hinge. Type 3 is the foreland extension generated by bending of the subducting lithosphere. All the three types of extension are 'eastward' migrating. Each extension type has different decollement planes: type 1 has the basal decollement at stretched lithosphere–asthenosphere boundary; type 2 has the decollement between the upper dragged layers of the crust and the asthenosphere; type 3 has normal faults terminating in the neutral crustal zone of folding where flexural slip may form (lower crust?).

material that is accreted in the frontal parts and is not subducted, is delaminated and uplifted by the mantle wedging, which is like a mouse running underneath a carpet and generating an uplifting wave during its passage. This uplift is short-lived (1–3 Ma?) and upward bounded (1–3 km) due to the limited amplitude of the wave and its eastward migration. The vertical component of this 'eastward'-migrating wave generated by the mantle wedge should then be responsible for the main relief associated to W-dipping subduction, and for the eastward migration of the drainage divide, accretionary wedge and foredeep basin in all these types of geodynamic environments (e.g., Apennines, Carpathians, Japan). This could explain why the amount of extension is much lower in comparison to the amount of shortening in the Apennines (Bally et al., 1986): the shortening is entirely related to the amount of subduction, while the extension is due to the later uplift generated by the underlying following mantle wedge. This uplift and extension may also be enhanced by the thickening of the underlying accretionary wedge. Like type 1, type 2 extension occurs while compression is active in the deeper accretionary wedge located to the east. The normal faults and shear zones flat out at the base of the anomalous light crust, above the asthenospheric wedging, with an interbedded thin or absent lithospheric mantle. The asthenospheric wedge in the hanging wall of the W-directed subduction zone can generate HT-LP metamorphic conditions affecting the overlying residual crust.

4. Extension at the subduction hinge

It is well known that folds have an external area of extension and a compression in the core underneath a neutral area (e.g., Twiss and Moores, 1992). This simple geometric constraint is valid at any scale and can explain why in several subduction hinges extensional tectonics occur (type 3 of extension, Fig. 1), in paradox with the convergent geodynamic context (Karig and Sharman, 1975; Bradley and Kidd, 1991). In fact extension is observed in the Marianas trench and in the Apulian foreland of the Apenninic subduction zone (Tropeano et al., 1994). Type 3 extension occurs while compression is active to the west in the accretionary wedge and it is confined in the upper layers of the lithosphere or crust, above the neutral line of the lithospheric bending. Subduction hinge extension may develop in a context of generalized subsidence due to the roll-back of the subduction hinge, or in a context of uplift in case of presence of an outer rise or bulge. In both cases the extension migrates eastward and lives the time of active bending of the lithosphere.

5. Extension due to increasing length of an arc

We expect extension along the strike of an arc during its evolution for simple geometric reasons (type 4 extension). The growth of an arc should generate horizontal stretching parallel to the arc trend due to the increasing length of any pre-arching marker line (Fig. 2). Extension has to be a natural accommodation during the development of an arc (Ghisetti and Vezzani, 1979; Knott and Turco, 1991; Oldow et al., 1993) generated by a W-dipping subduction (Doglioni, 1991). Type 4 extension generates grabens trending perpendicularly to the arc, with a comb distribution (Fig. 3). This extension occurs while compression is active in the accretionary wedge, where counterclockwise rotations occur in the northern arm of the arc, and clockwise rotations in its southern part (e.g., the Apenninic arc, Sagnotti and Meloni, 1993) in order to accommodate the arc migration. Grabens perpendicular to the Apennines trend occur both in the foreland and in the belt. Normal faults and shear zones are detached at the base of the lithosphere.

6. Extension induced by morphologic instability

Type 5 of extension may be considered the one generated by the relaxation of orogens due to gravitational collapse (England and Houseman, 1985; Dewey, 1988; Ratschbacher et al., 1989; Malavieille, 1993). In contrast with type 2 which is associated to W-dipping subduction, this extension is typical of orogens due to E- or NE- to NNE-dipping subductions (e.g., Himalayas, Andes, Alps). When an orogenic wedge thickens above the critical taper of the wedge, the morphologic gradient is able to induce lateral spreading and collapse of the orogen. This type of extension is quite well established and recognized in the western and central Alps, Himalayas and Andes (Davis et al., 1983; Burg et al., 1984; Burchfield and Royden, 1985; Dewey, 1988; Selverstone, 1988; Ratschbacher et al., 1989; Pêcher, 1991). Decollement planes are probably flattening out in the middle–upper crust; since the orogen is active, contractional tectonics are expected at depth. Type 5 extension (Fig. 4) is located in the structurally and morphologically uplifted core of the orogens that show a double vergence and an overthickened crust (e.g., Cordillera, Himalayas). This is in contrast with type 2 extension which occurs above a thinner crust

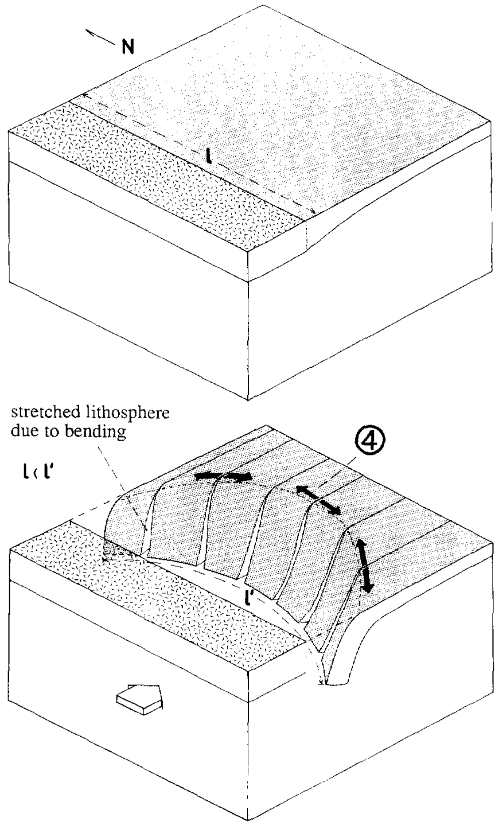

Fig. 2. Type 4 extension may be considered that generated by the lengthening of the arc during the arc growth in a W-dipping subduction setting. This extension strikes parallel to the arc and terminates in depth in the asthenosphere (modified after Doglioni, 1991).

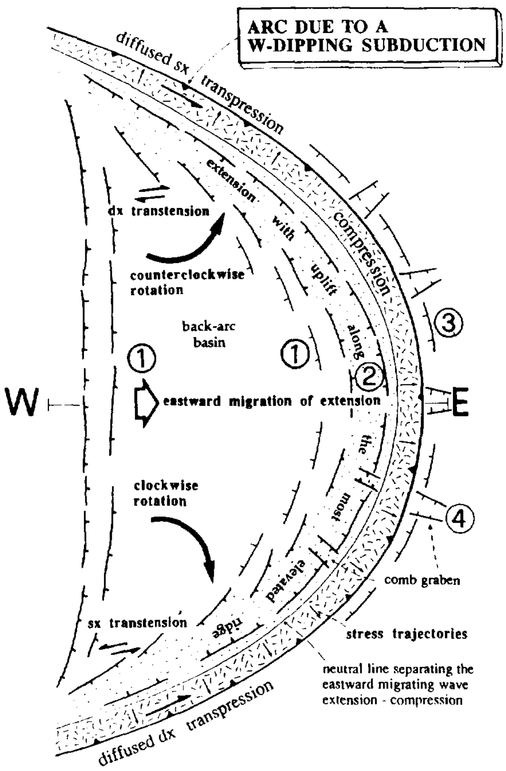

Fig. 3. Map view of a model of W-dipping subduction and location of the different extensional environments described in the former figures (modified after Doglioni, 1991).

and above orogens verging only to the east (e.g., Apennines). Type 5 extension may occur while compression is active in the whole accretionary wedge. When a margin of the orogen is weak or unconstrained, lateral extrusion has been proposed (Ratschbacher et al., 1991a, b).

Seismic images of active accretionary wedges (Bally, 1983; Von Huene, 1986) very often show normal faults in the upper or slope parts of the wedge, suggesting morphologic instability. Type 5 extension appears to be frequent for structurally and morphologically high orogens due to E- or NE-dipping subduction and should be distinguished from extension due to back-arc opening (type 1) or asthenospheric wedging at the hinge of W-dipping subduction (type 2) which have a different geodynamic context, low structural and morphologic eleva-

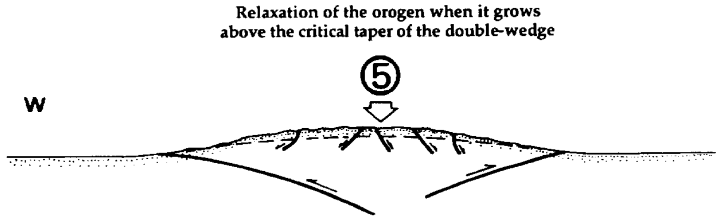

Fig. 4. Type 5 extension is the one generated by the relaxation of orogens due to gravitational collapse (e.g., Dewey, 1988). This extension is typical of overthickened crusts and is found associated to orogens due to E-, NE- or NNE-dipping subduction zones (e.g., Himalayas, Andes, Alps).

tion, and decollement planes and shear zones related to the extension reaching the base of the lithosphere (compare Figs. 1 and 4).

Extension and coeval compression also occur along passive continental margins (e.g., Namibia, Gulf of Mexico, Australia; Gerrard and Smith, 1982; Wu et al., 1990; Weimer and Buffler, 1992) where huge slides of sedimentary cover fall toward deeper areas, with extension in the source area and compression in the accumulation zone. Decollement planes for this type of event are usually shallow (2–4 km), at evaporitic or shale layers. However, this extension is not related to active convergent margins, but it may be later involved in subduction zones and related thrust belts, and it may be useful to keep in mind its peculiarities in order to make its distinction possible.

7. Extension induced by differential lithosphere–asthenosphere coupling in the hanging wall of a subduction zone

One of the most debated types of rifts are the so-called 'back-arc basins' related to E-, NE- and NNE-dipping subduction (e.g., the Aegean Sea or the Indonesian back-arc). In this paper they are considered to have a different origin from back-arc basins due to W-dipping subduction because of the following points: (a) they have lower rates of opening with respect to back-arcs related to W-dipping subduction; (b) they are characterized by thick continental crust (20–25 km), while back-arcs subducting in the opposite direction probably experienced fast generation of new oceanic crust; (c) they may be inactive during contemporaneous subduction, while this does not occur for a W-dipping subduction; (d) they have opposite polarity of opening toward WSW or SSW; (e) they often are inverted in a compressional regime; the system is composed by three plates (A–B–C in Fig. 5), in contrast with back-arc basins due to W-dipping subduction where the system may develop with two converging plates or within one single plate (Fig. 1).

The Aegean Sea is generally considered a back-arc basin due to NE-dipping subduction (Le Pichon and Angelier, 1979). However, the Aegean Sea is characterized by a relatively thick crust (20–25 km) according to Makris (1978), in spite of long-standing subduction, probably active at least since Cretaceous times (Meulenkamp et al., 1988). Extension and associated magmatism are SSW-migrating (Jackson and McKenzie, 1988), and they developed particularly only since the Oligocene, while subduction began much earlier. 'Normal' back-arc basins are associated with W-dipping subduction zones and they open very fast (10–20 Ma) and are always contemporaneous with the subduction. Moreover they are characterized by oceanization and eastward migration of extension and related magmatism, features directly surrounded by a frontal accretionary wedge. Instead the accretionary wedge of the Hellenic subduction zone is the southeastern prolongation of the Dinaric thrust belt, where no back-arc extension occurs.

No relative motion occurs between the central and eastern Mediterranean, since both sides belong to the same African plate. However, the plate is subducting both below Greece and Cyprus, but at different velocities, or more exactly the hanging-wall plate is overriding at different velocities. The Cyprus subduction in fact is slower as indicated by the minor shortening of the Quaternary sediments and by the lower seismicity with respect to the Hellenic subduction. If no relative motion between the central and eastern Mediterranean occurs, the different convergent rates at the two subduction zones have to be related to differential velocities between the hanging-wall plates, enabling a faster motion of Greece southwestward in relation to Turkey responsible for the extension in between. In other words, Turkey is moving away from Greece toward the northeast, and the two parts do not converge. Type 6 extension (Fig. 5) may or may not be coeval with compression elsewhere, and the related normal faults and shear zones should flat out in the decollement planes at the base of the lithosphere.

Accepting the 'east-northeastward' mantle flow indicated by the hot spot reference frame (Ricard et al., 1991), the origin of back-arc basins due to E- or NE-dipping subduction should not be due to lithospheric disappearance as found in W-dipping subduction zones, but rather to differential drag of the lithosphere in the hanging wall of the subduction due to different viscosity contrasts controlled by lateral heterogeneities (composition, heat flow, fluids, etc.) that control the amount of decoupling at the interface between lithosphere and asthenosphere.

8. Rifting induced by orogenic roots anchored into the mantle

The opening of the Atlantic (Tankard and Balkwill, 1989; Ziegler, 1993) and Tethys (Dercourt et al., 1986; Ziegler, 1988) oceans strictly followed the pre-existing lithospheric thickening of the Variscan and Caledonian orogens (Bernoulli and Lemoine, 1980, McClay et al., 1986; Dewey, 1988; Fig. 6). The Iapetus Ocean, in turn, was closed by the Caledonian orogen. The Late Eocene, Early Oligocene extension in front to the western Alps and producing the Rhône and Rhine grabens occurred immediately after the biggest collisional and lithospheric thickening event in the Alps (Laubscher, 1983). The Red Sea and the East Africa rift zones followed or were conditioned by the inherited pattern of pan-African pre-Cambrian orogens (McConnell, 1972; Smith and Mosley, 1993; Ring, 1994). All these observations suggest that the physical anisotropies in the lithosphere, particularly the lithospheric thickening due to continental collision, are controlling the location of rift zones. Glazner and Bartley (1985) documented that after 10–20 Ma an overthickened lithosphere becomes weaker with respect to the cratonic lithosphere and orogens can become good sites for rifting development. Kusznir and Park (1987) and Gaudemer et al. (1988) have shown that extension over-

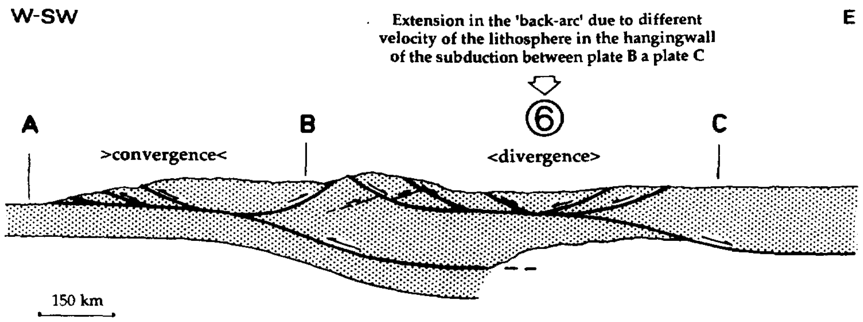

Fig. 5. Type 6 extension may be considered the so-called 'back-arc' basin forming in the hanging wall of E- or NE-dipping subductions. In contrast with back-arc basins related to W-dipping subduction zones where the back-arc is formed beacause of the loss of lithosphere caused by subduction, in this type 6 the extension forms because the hanging-wall lithosphere is split in two distinct diverging plates, like normal rifting. In fact crust and lithosphere in this type of 'back-arc' often remain thick and continental in origin and are not regularly contemporaneous with subduction. The system is composed by three plates (A–B–C), in contrast with back-arc basins due to W-dipping subduction where the system may develop with two converging plates or within one single plate. The classic example is the Aegean sea, formed since the Miocene on top of a NE-dipping subduction zone active since at least the Cretaceous.

printing large orogens is much more widespread (e.g., 1000 km) with respect to intraplate extension that is usually confined to narrow areas usually no wider than 100 km (e.g., the East African rift system), and they also demonstrated the thermal control of wide orogenic wedges on their collapse. These data indicate that the lithosphere is actively controlling rift zones, more than the underlying convecting deep mantle. Sabadini et al. (1992) indicated that the thickening of the lithosphere controls the deviatoric stress and its trajectories: a thick continental lithosphere is able to induce extension in its western margin and compression to the east due to the drag produced by the deep lithospheric roots into a relative asthenospheric flow. Consequently the overthickening induced by collisional events can control later extension, this fact predicting a continuous opening and closing of the oceans, the so-called 'Wilson cycle', extension–compression and then again extension (Vink, 1979). Immediately after a compressive event (i.e. the Late Carboniferous Variscan closure) a new rifting may start (i.e. the Atlantic and Tethys riftings) if the lithospheric overthickening is joined by the relative westward drift of the lithosphere with respect to the asthenospheric mantle, and by the lateral asthenospheric anisotropies inducing velocity variations in the overlying lithospheric plates. Continental lithosphere is weaker than oceanic lithosphere by about a factor of 3 (Vink et al., 1984): this is a very important factor in controlling the location of rift zones which are in fact more diffused in continental rather than oceanic settings. Moreover, Kusznir and Park (1987) have indicated that thicker continental crust is relatively weaker with respect to normal thickness crust because its strength is a function of the geothermal gradient, composition and thickness. For example a Devonian collapse-basin has been described by Norton (1986) and Seranne et al. (1989) on the thick and thermally perturbed orogenic Caledonian crust.

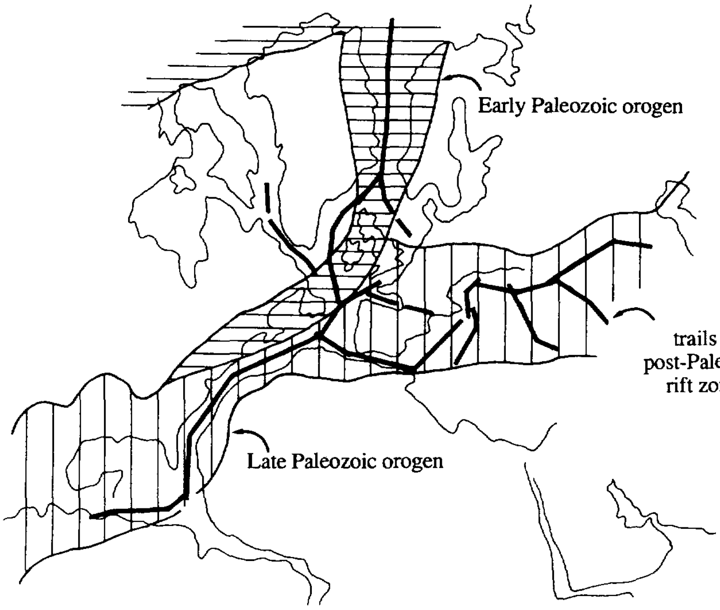

Fig. 6. This map shows the distribution of the Paleozoic collisional belts, namely Variscan (or Hercynian) and Caledonian, and the trace of the later Atlantic or Tethyan rifting which strictly followed the pre-existing lithospheric thickening due to the orogens (modified after Bernoulli and Lemoine, 1980). This observation shows that the lithospheric thickness directly controls the location of rift zones.

Type 7 (Fig. 7) of extension related to convergent settings appears due to the lithospheric roots of orogens generated by E-, NE- or NNE-dipping subduction which provide a 200–300-km-thick continental lithosphere and an obstacle to the relative east- or northeast-ward 'asthenospheric flow'. Extension might result from the greater drag and anchoring of the continental lithospheric roots within the asthenospheric mantle, the western lithospheric side remaining more to the west due to the minor drag, generating rift processes. We have seen so far type 1–6 extensional tectonics, these types are coeval and related to convergent settings. Type 7 extension occurs after the compression event. The normal faults accompanying the rift (Buck, 1991) may emerge in any part of the inherited orogen, as a function of their deep trajectory as proposed in Fig. 7. Other trajectories of the normal faults and shear zones may be transferred and displaced with respect to the emergence across decollement layers in the lower crust (Reston, 1990). The same notion should be valid for the thick lithosphere of a craton (i.e. Asia, Africa and America). Type 5 extension, which is caused by gravitational collapse, forms during and after the orogenic thickening, probably enhancing the later post-orogenic type 7 extension.

Longitudinal (or normal to the mantle flow) thickness variations of the lithosphere can also induce plate rotation (triangle effect, Doglioni, 1993a). The thick continental lithosphere compared to lateral thinner oceanic lithosphere is also capable of generating velocity variations between continental and oceanic realms. The biggest stress accumulations occur at the strongest lithospheric anisotropies (i.e. the continental margins) where thickness and compositional gradients are maximal. Indeed, we observe based on geological evidence that in the past subduction zones were generated along continental margins and that they were controlled by geographical distribution: an E-dipping subduction zone developed only in relation to a thin, possibly oceanic lithosphere located to the west with respect to a thicker continental lithosphere (e.g., the Andean subduction has been active for several hundred million years because to the west of it thinner oceanic lithosphere (i.e., the Nazca plate) was maintained. In the opposite case, we observe that W-dipping subduction begins only in relation to thin lithosphere positioned to the east with respect to a thicker continental lithosphere (i.e., all the W-dipping subduction zones of the western Pacific margin were located to the east of the thick Asiatic continental lithosphere). All this indicates that rift zones and later or earlier subduction zones are strictly controlled by the nature of the lithosphere and by interference of these zones with the associated asthenospheric flow.

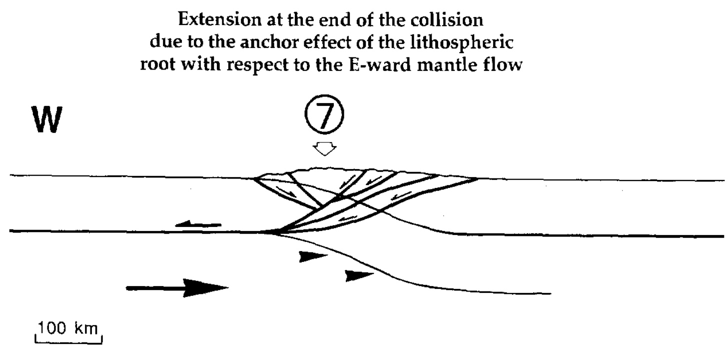

Fig. 7. Type 7 extension related to convergent settings may be considered a classic rift forming in correspondence of earlier orogens, e.g., the Atlantic opening on the Caledonian and Hercynian sutures due to the obstacle generated by the root of the thrust belt. The trace of future possible normal faults is only schematized, showing that the superficial emergence of those faults may be located in any part of the inherited orogen.

9. Inversion of plate motion, from convergent to divergent relative plate vectors

Plate velocity relative to the mantle is controlled by the degree of decoupling occurring across the asthenosphere. The highest decoupling occurs at the base of the Pacific plate. Assuming the 'westward' drift of the lithosphere, variations in the degree of decoupling determine convergent or divergent plate margins as a function that the plate to the east is moving faster or slower toward the west in relation to the plate to the west (Doglioni, 1993a). The Basin and Range extension in western North America (e.g., Wernicke, 1981; Coney and Harms, 1984; Davis and Lister, 1988) occurred when the East Pacific oceanic ridge subducted underneath the Cordillera (Atwater, 1970): the high velocity of the Pacific plate toward the west-northwest could not be compensated by the slower westward motion of the North American plate. Type 8 extension (Fig. 8) is clearly different in origin from the earlier ones, being mainly controlled by the inversion of relative plate vectors when the intermediate plate is subducted (e.g., Farallon). This should have inverted the system, from subduction to extension, being the vector of the Pacific plate greater toward the west-northwest with respect to North America. In fact the extension overprinted the North American Cordillera (Jones et al., 1993, and references therein) where the intermediate plate was lost in subduction, whereas it is still active where there still is a slower plate to the east of the Pacific Rise (e.g., Cascadia, Andes). Therefore, type 8 extension is post-convergence, and decollement planes and shear zones cross-cut the entire lithosphere and flat at its base (Wernicke, 1985). The Basin and Range is considered a classic area for the uplift of the metamorphic core complex, but it seems to have a very different geodynamic origin with respect to the for-

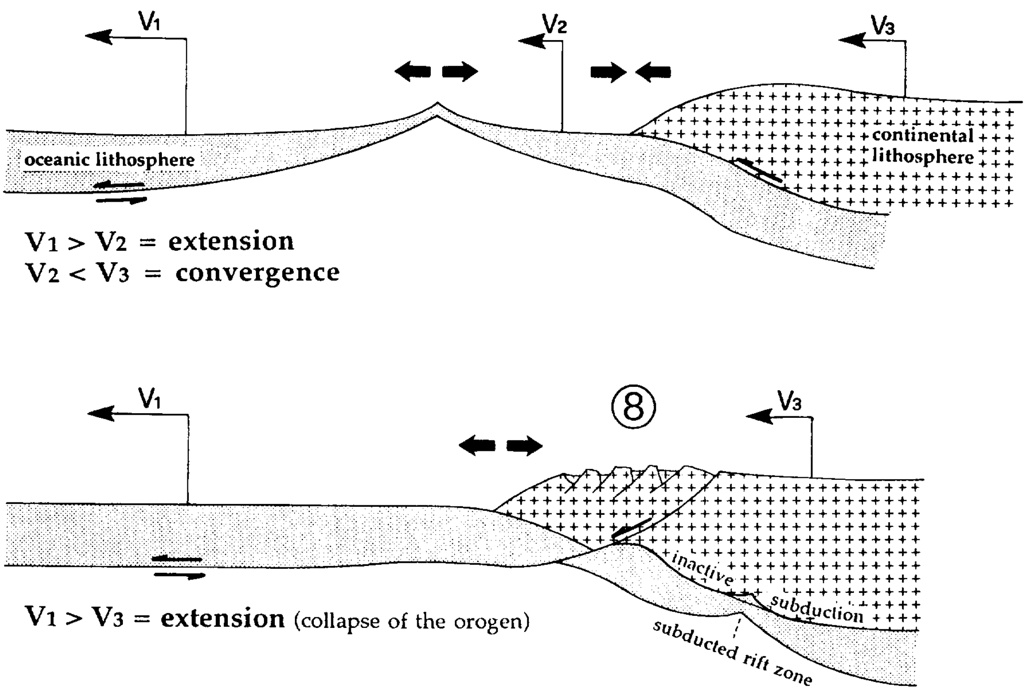

Fig. 8. Type 8 extension may be considered the one in which an oceanic ridge is subducted and the hanging-wall plate becomes slower with respect to the footwall. The orogen that formed during the earlier active subduction phase is then cross-cut by normal faults due to the inversion in relative plate vectors. Classic example could be the Basin and Range extension, a classic metamorphic core complex.

mer types of extension. For a discussion on the rheological differences of the Basin and Range with other extensional settings see for instance Braun and Beaumont (1989). However, the Cordillera can also have experienced other extensional types (e.g., 5).

10. Other types of relationships between extension and convergent geodynamic settings

Arcs along thrust belts may propagate toward the foreland more than adjacent thrusts, e.g., due to a larger presence of evaporites in the decollement plane, or to a larger roll-back of the subduction hinge, etc. The arc can experience transtension along its margins due to the slower forward propagation of the internal thrusts (Fig. 9). The decollement of the extension should correspond to the decollement of the main structure, e.g., the thrust plane or even the base of the lithosphere.

It is well known that along undulate transpressive belts transtensional pull-aparts may form, being co-existing convergent and divergent tectonic settings (Reading, 1980; Royden, 1985; Zoback, 1991).

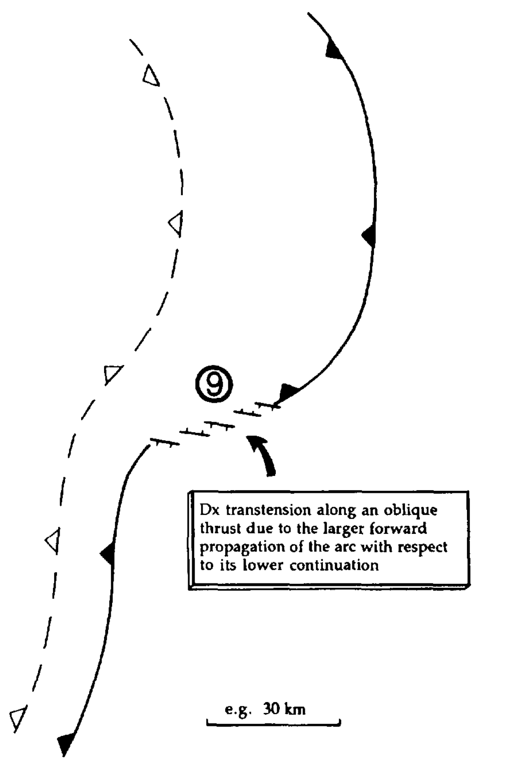

Fig. 9. Another type (9) of extension in convergent geodynamic setting could be the transtension generated along oblique ramps of thrust planes that largely propagate toward the foreland with respect to their along-strike continuation. En-échelon pattern of grabens or pull-aparts are expected. The basal decollement of this extension should coincide with the decollement of the thrust plane. Examples could be observed in the western margin of the Ionian arc (offshore eastern Sicily) or the southern arm of the Barbados accretionary wedge.

11. Apparent extension due to down-section cut of thrust planes along subduction zones

During subduction, accretion of the lower plate should conceptually be determined by slices having a lower decollement cross-cutting down-section the upper layers of the inclined subducting crust. The E-dipping subduction provides a way to bring down a reference point to higher P and T conditions, in a prograding metamorphic evolution. During collision, the thrust planes of the upper plate propagate into the lower plate, passing underneath to the reference point which is then uplifted by the new thrust, bringing again the reference point to higher structural and morphological conditions of lower P and T in a retrograding metamorphic path. During subduction, the lithosphere in the footwall is inclined (20–30°). Therefore, any horizontal foreward propagating shear plane is cutting down-section the lower plate. This thrust plane presents 'normal fault' displacement, being the hanging wall stratigraphically lowered in relation to the footwall (Fig. 10). The thrust plane may likely be later folded and uplifted by younger deeper thrusts. The resulting antiformal stack at the surface may show normal fault displacement, even if no extension occurred on it. However, not direct field evidences can prove such kinematics.

Apart from classic flattening in the ductile field (e.g., along fold limbs), apparent normal faulting in convergent settings may also be seen in foreland dipping duplexes (Boyer and Elliott, 1982) or synthetic and antithetic Riedel planes in the footwall of thrust planes (e.g., Ratschbacher et al., 1989; Schmid

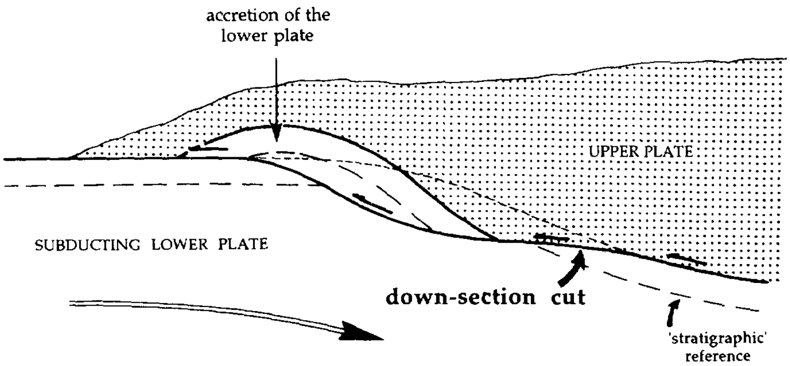

Fig. 10. During subduction, accretion of the lower plate should conceptually be determined by slices having a lower decollement cross-cutting down-section the upper layers of the inclined subducting crust. With the forward migration of the system, such a slice may later be included in an antiformal stack, duplexes or other geometries. The original down-section cut of the basal thrust may then be tilted and to appear as a 'normal fault' displacement, without any real extension.

and Haas, 1989), when those features are not pre-existing inherited passive margin real normal faults.

12. Concluding remarks

Previous work has shown that there are different types of geologic environments where extension is generated in relation to convergent systems (Table 1). This paper tries to propose that the general statement of 'metamorphic core complex extension' widely used in the literature should be distinguished on the basis of the associated geodynamic setting. The uplift of deep crustal rocks at the surface is mainly considered as related to deep thrust planes that are associated to thrust belts that form with E-, NE- or NNE-dipping subduction zones; later exten-

Table 1
Main extensional types related to convergent geodynamic settings. Other differences like shape, time evolution and rheological parameters vary in each type of setting

Geodynamic setting planes	Subduction polarity	syn- or post-conv.	Generalized subsidence or upliftf	Basal decollement	Extension migration	Number of plates	Opening rates
1 Back-arc basin extension	westward	syn	subsidence 700 m/Ma	lithosphere–asthenosphere	eastward	1 or 2	3–10 cm/yr
2 Asthenospheric wedging-related ext.	westward	syn	uplift 500 m/Ma	crust–asthenosphere	eastward	1 or 2	
3 Subduction hinge extension	westward	syn	subsidence or uplift	upper lithosphere	eastward	1 or 2	
4 Increasing arc length extension	westward	syn	subsidence	lithosphere–asthenosphere	eastward	1 or 2	
5 Morphologic gradient-related ext.	eastward or northeastward	syn	uplift	middle–upper crust		1 or 2	
6 Hanging-wall gradient velocities	eastward or northeastward	syn	subsidence	lithosphere–asthenosphere	southwestward	3	0.1–5 cm/yr
7 Lithospheric roots-related extension	eastward or northeastward	post	subsidence 100 m/Ma	lithosphere–asthenosphere		1 to 2	0.1–1 cm/yr
8 Inversion of velocity gradients	eastward or northeastward	post	subsidence	lithosphere–asthenosphere		3 to 2	1–10 cm/yr
9 Transfer zones-related extension	both west and east	syn	subsidence or uplift	basal thrust plane		1 or 2	

sion of any of the former types may affect such orogens, in particular type 1 (back-arc extension, e.g., the Tyrrhenian Sea overprinting the Alpine orogen), or type 5 (extension induced by collapse of orogens, e.g., Alps, Himalayas), or type 6 (extension induced by differential lithosphere–asthenosphere coupling in the hanging wall of a subduction, e.g., Aegean rift), or type 7 (rifting induced by orogenic roots anchored into the mantle, e.g., Atlantic and Tethys), or type 8 (inversion of plates motion, from convergent to divergent relative plate vectors, e.g., Basin and Range). Type 9 could be considered the extension resulting from greater foreland propagation of thrust arcs with respect to adjacent more internal thrusts. Several other extensional environments may develop internally to thrust belts for different tectonic reasons. Different shape, location and structural depth of the decollement planes characterize each extension type. Each type should also have different rheological parameters, having different geothermal gradients and different sections of crust or lithospheric mantle involved. The variations in lithosphere rheology are a fundamental factor in controlling different strength envelopes (Ranalli, 1991). Each type of extension should consequently have different depth and shape of necking of the lithosphere (Cloetingh, 1992). Different types of extension may overlap earlier different features, e.g., the Pannonian back-arc extension superimposed the northern Dinarides and the eastern Alps, or type 1 extension might have interacted with type 5, related to the collapse or extrusion of the earlier orogens (Ratschbacher et al., 1991a). Type 5, gravitational collapse, may enhance type 7, due to lithospheric roots-related extension. An interplay between the different types of extensional settings may be expected. Different types of extension can overprint the entire variety of pre-existing tectonic fabrics. It can also be argued that the discussions on the modes of extension (i.e. pure shear vs. simple shear) have been based on original different types of extensional environments. Magmatic signatures could also contribute to characterize these different extensional settings.

Acknowledgements

The paper strongly benefited from the critical reading by L. Ratschbacher, C. Morley, S. Cloetingh and two anonymous referees. Discussions with F. Horvath, F. Mongelli, P. Pieri, G. Ranalli and M. Tropeano were very helpful. The research was supported by CNR and ASI.

References

Atwater, T., 1970. Implications of plate tectonics for the Cenozoic evolution of western North America. Geol. Soc. Am. Bull., 81: 3513–3536.

Bally, A.W., 1983. Seismic Expression of Structural Styles. A Picture and Work Atlas. Am. Assoc. Pet. Geol., Stud. Geol., 15, 3 Vols.

Bally, A.W., Burbi, L., Cooper, C. and Ghelardoni, R., 1986. Balanced sections and seismic reflection profiles across the central Apennines. Mem. Soc. Geol. Ital., 35: 257–310.

Bernoulli, D. and Lemoine, M., 1980. Birth and early evolution of the Tethys: the overall situation. Mem. Bur. Rech. Géol. Min., 115: 168–179.

Bostrom, R.C., 1971. Westward displacement of the lithosphere. Nature, 234: 356-358

Boyer, S.E. and Elliott, D., 1982. Thrust systems. Am. Assoc. Pet. Geol. Bull., 66: 1196–1230.

Bradley, D.C. and Kidd, W.S.F., 1991. Flexural extension of the upper continental crust in collisional foredeeps. Geol. Soc. Am. Bull., 103: 1416–1438.

Braun, J. and Beaumont, C., 1989. Contrasting styles of lithospheric extension: implications for differences between the Basin and Range Province and rifted continental margins. In: A.J. Tankard and H.R. Balkwill (Editors), Extensional Tectonics and Stratigraphy of the North Atlantic Margins. Am. Assoc. Pet. Geol. Mem., 46: 53–79.

Buck, W.R., 1991. Modes of continental lithospheric extension. J. Geophys. Res., 96: 20,161–20,178.

Burchfield, B.C. and Royden, L.H., 1985. North–south extension under the convergent Himalayan regime. Geology, 13: 679–682.

Burg, J.P., Brunel, M., Gapais, D., Chen, G.M. and Liu, G.H., 1984. Deformation of leucogranites of the crystalline Main Central Sheet in southern Tibet (China). J. Struct. Geol., 6: 535–542.

Cadek, O. and Ricard, Y., 1992. Toroidal/poloidal energy partitioning and global lithospheric rotation during Cenozoic time. Earth Planet. Sci. Lett., 109: 621–632.

Carmignani, L. and Kligfield, R., 1990. Crustal extension in the Northern Apennines: the transition from compression to extension in the Alpi Apuane core complex. Tectonics, 9: 1275–1303.

Cloetingh, S., 1992. Lithospheric dynamics and the tectonics of sedimentary basins. Proc. K. Ned. Akad. Wetensch., 95: 349–369.

Coney, P.J. and Harms, T.A., 1984. Cordilleran metamorphic core complexes: Cenozoic extensional relics of Mesozoic compression. Geology, 12: 550–554.

Cosentino, D. and Gliozzi, E., 1988. Considerazioni sulle velocità di sollevamento di depositi eutirreniani dell'Italia meridionale e della Sicilia. Mem. Soc. Geol. Ital., 41: 653–665.

Coward, M.P., Dewey, J.F. and Hancock, P.L. (Editors), 1987. Continental Extensional Tectonics. Geol. Soc. London Spec. Publ., 28.

Davis, D., Suppe, J. and Dahlen, F.A., 1983. Mechanics of fold-and-thrust belts and accretionary wedges. J. Geophys. Res., 88: 1153–1172.

Davis, G.A. and Lister, G.A., 1988. Detachment faulting in continental extension; perspectives from the southwestern U.S. Cordillera. In: S.P. Clark et al. (Editors), Processes in Continental Lithospheric Deformation. Geol. Soc. Am., Spec. Pap., 218: 133–159.

Dercourt, J. et al., 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. Tectonophysics, 123: 241–315.

Dewey, J.F., 1988. Extensional collapse of orogens. Tectonics, 7(6): 1123–1140.

Doglioni, C., 1991. A proposal of kinematic modelling for W-dipping subductions — possible applications to the Tyrrhenian–Apennines system. Terra Nova, 3: 423–434.

Doglioni, C., 1992. Main differences between thrust belts. Terra Nova, 4: 152–164.

Doglioni, C., 1993a. Geological evidence for a global tectonic polarity. J. Geol. Soc. London, 150: 991–1002.

Doglioni, C., 1993b. Some remarks on the origin of foredeeps. Tectonophysics, 228: 1–20.

England, P.C. and Houseman, G.A., 1985. The influence of lithospheric strength heterogeneities on the tectonics of Tibet and surrounding regions. Nature, 315: 297–301.

Gaudemer, Y., Jaupart, C., Tapponnier, P., 1988. Thermal control on post-orogenic extension in collision belts. Earth Planet. Sci. Lett., 89: 48–62.

Gerrard, I. and Smith, G.C., 1982. Post-Paleozoic succession and structure of the southwestern African continental margin. In: J.S. Watkins and C.L. Drake (Editors), Studies in Continental Margin Geology. Am. Assoc. Pet. Geol. Mem., 34: 49–74.

Ghisetti, F. and Vezzani, L., 1979. The geodynamic evolution of the crustal structures of Calabria and Sicily. Proc. I.G.U. Comm. Geomorph. Surv. Mapping, 15 Plen. Meeting, Modena–Catania, pp. 335–347.

Glazner, A.F. and Bartley, J.M., 1985. Evolution of lithospheric strength after thrusting. Geology, 13: 42–45.

Hussong, D.M. and Uyeda, S., 1981. Tectonic processes and the history of the Mariana arc: a synthesis of the results of the Deep Sea Drilling Project Leg 60. Init. Rep. Deep Sea Drill. Proj., 60: 909–930.

Jackson, J. and McKenzie, D., 1988. Rates of active deformation in the Aegean Sea and surrounding areas. Basin Res., 1(3): 121–128.

Jolivet, L., 1993. Extension of thickened continental crust, from brittle to ductile deformation: examples from Alpine Corsica and Aegean Sea. Ann. Geofis., 36(2): 139–153.

Jones, C.H., Wernicke, B.P., Farmer, G.L., Walker, J.D., Coleman, D.S., McKenna, L.W. and Perry, F.V., 1993. Variations across and along a major continental rift: an interdisciplinary study of the Basin and Range Province, western USA. Tectonophysics, 213: 57–96.

Karig, D.E. and Sharman, III G.F., 1975. Subduction and accretion in trenches. Geol. Soc. Am. Bull., 86: 377–389.

Kastens, K., Mascle, J. and 19 others, 1988. ODP Leg 107 in the Tyrrhenian Sea: insigths into passive margin and back-arc basin evolution. Geol. Soc. Am. Bull., 100: 1140–1156.

Knott, S.D. and Turco, E., 1991. Late Cenozoic kinematics of the Calabrian arc. Tectonics, 10: 1164–1172.

Kobayashi, K., 1984. Subsidence of the Shikoku back-arc basin. Tectonophysics, 102: 105–117.

Kusznir, N.J. and Park, R.G., 1987. The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness. In: M.P. Coward, J.F. Dewey and P.L. Hancock (Editors), Continental Extensional Tectonics. Geol. Soc. London Spec. Publ., 28: 35–52.

Laubscher, H.P., 1983. The Late Alpine (Periadriatic) intrusions and the Insubric Line. Mem. Soc. Geol. Ital., 26: 21–30.

Le Pichon, X., 1968. Sea-floor spreading and continental drift. J. Geophys. Res., 73(12): 3661–3697.

Le Pichon, X. and Angelier, J., 1979. The Hellenic arc and trench system: a key to the neotectonic evolution of the eastern Mediterranean area. Tectonophysics, 60: 1–42.

Lister, G.S., Banga, G. and Feenstra, A., 1984. Metamorphic core complexes of Cordilleran type in the Cyclades, Aegean Sea, Greece. Geology, 12: 221–225.

Makris, J., 1978. The crust and upper mantle of the Aegean region from deep seismic soundings. Tectonophysics, 46: 269–284.

Malavieille, J., 1993. Late orogenic extension in mountain belts: insights from the Basin and Range and the Late Paleozoic Variscan belt. Tectonics, 12(5): 1115–1130.

McClay, K.R., Norton, M.G., Coney, P. and Davies, G.H., 1986. Collapse of the Caledonian Orogen and the Old Red Sandstone. Nature, 323: 147–149.

McConnell, R.B., 1972. Geological development of the rift system of Eastern Africa. Geol. Soc. Am. Bull., 83: 2549–2572.

Meulenkamp, J.E., Wortel, M.J.R., Van Wamel, W.A., Spakman, W. and Hoogerduyn-Strating, E., 1988. On the Hellenic subduction zone and the geodynamic evolution of Crete since the late Middle Miocene. Tectonophysics, 146: 203–216.

Michard, A., Chopin, C. and Henry, C., 1993. Compression versus extension in the exhumation of the Dora-Maira coesite-bearing unit, Western Alps, Italy. Tectonophysics, 221(2): 173–194.

Morley, C.K., 1993. Discussion of origins of hinterland basins to the Rif–Betic Cordillera and Carpathians. Tectonophysics, 226: 359–376.

Moser, J., Matyska, C., Yuen, D.A., Malevsky, A.V. and Harder, H., 1993. Mantle rheology, convection and rotational dynamics. Phys. Earth Planet. Inter., 79: 367–381.

Moussat, E., Rehault, J.P., Fabbri, A. and Mascle, G., 1985. Evolution géologique de la Mer Tyrrhenienne. C.R. Acad. Sci. Paris, 301(7): 491–496.

Nelson, T.H. and Temple, P.G., 1972. Mainstream mantle convection: a geologic analysis of plate motion. Am. Assoc. Pet. Geol. Bull., 56: 226–246.

Norton, M.G., 1986. Late Caledonide extension in western Norway: a response to extreme crustal thickening. Tectonics, 5(2): 195–204.

Oldow, J.S., D'Argenio, B., Ferranti, L., Pappone, G., Marsella, E. and Sacchi, M., 1993. Large-scale longitudinal extension in the southern Apennines contractional belt, Italy. Geology, 21: 1123–1126.

Pêcher, A., 1991. The contact between the higher Himalaya crystallines and the Tibetan sedimentary series: Miocene large-scale dextral shearing. Tectonics, 10(3): 587–598.

Platt, J.P., 1986. Dynamics of orogenic wedges and the uplift of high pressure metamorphic rocks. Geol. Soc. Am. Bull., 97: 1037–1053.

Ranalli, G., 1991. Regional variations in lithosphere rheology from heat flow observations. In: V. Cermak and L. Rybach (Editors), Exploration of the Deep Continental Crust. Springer-Verlag, Berlin, pp. 1–22.

Ratschbacher, L., Frisch, W., Neubauer, F., Schmid, S.M. and Neugebauer, J., 1989. Extension in compressional orogenic belts: The eastern Alps. Geology, 17: 404–407.

Ratschbacher, L., Frisch, W. and Linzer, H.G., 1991a. Lateral extrusion in the Eastern Alps, 2. Structural analysis. Tectonics, 10(2): 257–271.

Ratschbacher, L., Merle, O., Davy, P. and Cobbold, P., 1991b. Lateral extrusion in the Eastern Alps, 1. Boundary conditions and experiments scaled for gravity. Tectonics, 10(2): 245–256.

Reading, H.G., 1980. Characteristics and recognition of strike-slip fault systems. In: P.F. Ballance and H.G. Reading (Editors), Sedimentation in Oblique-Slip Mobile Zones. Int. Assoc. Sedimentol., Spec. Publ., 4: 7–26.

Reston, T.J., 1990. The lower crust and the extension of the continental lithosphere: kinematic analysis of BIRPS deep seismic data. Tectonics, 9(5): 1235–1248.

Ricard, Y., Doglioni, C. and Sabadini, R., 1991. Differential rotation between lithosphere and mantle: a consequence of lateral mantle viscosity variations. J. Geophys. Res., 96(B5): 8407–8415.

Ring, U., 1994. The influence of pre-existing structure on the evolution of the Cenozoic Malawi rift (East African rift system). Tectonics, 13(2): 313–326.

Royden, L.H., 1985. The Vienna Basin: a thin-skinned pull-apart basin. In: K.T. Biddle and N. Christie-Blick (Editors), Strike-Slip Deformation, Basin Formation, and Sedimentation. SEPM, Spec. Publ., 37: 319–338.

Royden, L.H., Horvath, F., Nagymarosy, A. and Stegena, L., 1983. Evolution of the Pannonian Basin system, 2. Subsidence and thermal history. Tectonics, 2: 91–137.

Sabadini, R., Giunchi, C., Gasperini, P. and Boschi, E., 1992. Plate motion and dragging of the upper mantle: lateral variations of lithospheric thickness and their implications for intraplate deformation. Ann. Geophys., 10(Suppl. I): C83.

Sagnotti, L. and Meloni, A., 1993. Pleistocene rotations and strain in southern Italy: the example of the Sant'Arcangelo basin. Ann. Geofis., 36: 2, 83–95.

Schmid, S.M. and Haas, R., 1989. Transition from near-surface thrusting to intrabasement decollement, schlinig thrust, eastern Alps. Tectonics, 8(4): 697–718.

Selverstone, J., 1988. Evidence for east–west crustal extension in the Eastern Alps: implications for the unroofing history of the Tauern Window. Tectonics, 7: 87–105.

Seranne, M., Chauvet, A., Seguret, M. and Brunel, M., 1989. Tectonics of the Devonian collapse-basin of western Norway. Bull. Soc. Géol. Fr., (8), V. 3: 489–499.

Smith, A.D., 1993. The continental mantle as a source for hot spot volcanism. Terra Nova, 5: 452–460.

Smith, M. and Mosley, P., 1993. Crustal heterogeneity and basement influence on the development of the Kenya rift, East Africa. Tectonics, 12(2): 591–606.

Tankard, A.J. and Balkwill, H.R., 1989. Extensional tectonics and stratigraphy of the North Atlantic margins. Am. Assoc. Pet. Geol. Mem., 46: 1–641.

Tropeano, M., Marino, M. and Pieri, P., 1994. Evidenze di tettonica distensiva plio-pleistocenica al margine orientale della Fossa Bradanica: l'horst di Zagarella. Il Quaternario, 7(2): 597–606.

Taylor, B., 1979. Bismark Sea: evolution of a back-arc basin. Geology, 7: 171–174.

Twiss, R.J. and Moores, E.M., 1992. Structural Geology. Freeman, New York, NY, 532 pp.

Uyeda, S. and Kanamori, H., 1979. Back-arc opening and the mode of subduction. J. Geophys. Res., 84(B3): 1049–1061.

Vink, G.E., 1979. Continental rifting and the implication for plate tectonic reconstructions. J. Geophys. Res., 87(B13): 10,677–10,688.

Vink, G.E., Morgan, W.J. and Zhao, W.L., 1984. Preferential rifting of continents: a source of displaced terranes. J. Geophys. Res., 89(B12): 10,072–10,076.

Von Huene, R., 1986. Seismic images of modern convergent margin tectonic structure. Am. Assoc. Pet. Geol. Stud. Geol., 26: 1–60.

Weimer, P. and Buffler, R.T., 1992. Structural geology and evolution of the Mississippi fan fold belt, deep Gulf of Mexico. Am. Assoc. Pet. Geol. Bull., 76(2): 225–251.

Wernicke, B., 1981. Low-angle normal faults in the Basin and Range Province — nappe tectonics in an extending orogen. Nature, 291: 645–648.

Wernicke, B., 1985. Uniform-sense normal simple shear of the continental lithosphere. Can. J. Earth Sci., 22: 108–125.

Wu, S., Bally, A.W. and Cramez, C., 1990. Allochthonous salt, structure and stratigraphy of the north-eastern Gulf of Mexico, Part II. Structure. Mar. Pet. Geol., 7: 4, 334–370.

Ziegler, P.A., 1988. Evolution of the Arctic–North Atlantic and Western Tethys. Am. Assoc. Pet. Geol. Mem., 43: 1–206.

Ziegler, P.A. (Editor), 1993. Geodynamics of rifting, Vol. II. Case history studies on rifts: north and south America and Africa. Tectonophysics, 213: 1–284.

Zoback, M.D., 1991. State of stress and crustal deformation along weak transform faults. Philos. Trans. R. Soc. London, Ser. A, 337: 141–150.