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Mechanical controls on collision-related compressional intraplate deformation

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Abstract

Intraplate compressional features, such as inverted extensional basins, upthrust basement blocks and whole lithospheric folds, play an important role in the structural framework of many cratons. Although compressional intraplate deformation can occur in a number of dynamic settings, stresses related to collisional plate coupling appear to be responsible for the development of the most important compressional intraplate structures. These can occur at distances of up to ±1600 km from a collision front, both in the fore-arc (foreland) and back-arc (hinterland) positions with respect to the subduction system controlling the evolution of the corresponding orogen. Back-arc compression associated with island arcs and Andean-type orogens occurs during periods of increased convergence rates between the subducting and overriding plates. For the build-up of intraplate compressional stresses in fore-arc and foreland domains, four collision-related scenarios are envisaged: (1) during the initiation of a subduction zone along a passive margin or within an oceanic basin; (2) during subduction impediment caused by the arrival of more buoyant crust, such as an oceanic plateau or a microcontinent at a subduction zone; (3) during the initial collision of an orogenic wedge with a passive margin, depending on the lithospheric and crustal configuration of the latter, the presence or absence of a thick passive margin sedimentary prism, and convergence rates and directions; (4) during post-collisional over-thickening and uplift of an orogenic wedge. The build-up of collision-related compressional intraplate stresses is indicative for mechanical coupling between an orogenic wedge and its fore- and/or hinterland. Crustal-scale intraplate deformation reflects mechanical coupling at crustal levels whereas lithosphere-scale deformation indicates mechanical coupling at the level of the mantle-lithosphere, probably in response to collisional lithospheric over-thickening of the orogen, slab detachment and the development of a mantle back-stop. The intensity of collisional coupling between an orogen and its fore- and hinterland is temporally and spatially variable. This can be a function of oblique collision. However, the build-up of high pore fluid pressures in subducted sediments may also account for mechanical decoupling of an orogen and its fore- and/or hinterland. Processes governing mechanical coupling/decoupling of orogens and fore- and hinterlands are still poorly understood and require further research. Localization of collision-related compressional intraplate deformations is controlled by spatial and temporal strength variations of the lithosphere in which the thermal regime, the crustal thickness, the pattern of pre-existing crustal and mantle discontinuities, as well as sedimentary loads and their thermal blanketing effect play an important role. The

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stratigraphic record of collision-related intraplate compressional deformation can contribute to dating of orogenic activity affecting the respective plate margin.

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1. Introduction

Intraplate compressional structures range from uplift of broad arches and accelerated basin subsidence, involving whole lithospheric buckling, to crustal-scale folding and, by reactivation of pre-existing crustal discontinuities, to upthrusting of basement blocks and inversion of tensional hanging-wall basins. Intraplate compressional features play an important role in the structural framework of many cratons. Such structures can develop (1) in forelands of orogens in response to collisional mechanical coupling of the orogenic wedge with the foreland plate, (2) in back-arc areas during periods of increasing convergence rates between the colliding plates, (3) in oceanic basins where they may be considered as precursors to the development of new subduction zones, (4) along major wrench systems, (5) in multi-directional rift systems in response to stress re-orientation, (6) on passive margins, partly in the projection of intra-oceanic transform faults (Ziegler et al., 1995).

The World Stress Map demonstrates that horizontal compressional stresses can be transmitted over great distances through continental and oceanic lithosphere (Zoback, 1992). Although a number of dynamic processes contribute towards the build-up of intraplate compressional stresses, forces related to collisional plate coupling appear to be responsible for the most important compressional intraplate deformations. These can occur in hinterland as well as in foreland domains of orogenic wedges and at distances of up to ±1600 km from contemporaneous thrust fronts (Ziegler et al., 1995).

The stratigraphic record preserved in sedimentary basins permits to closely date the age of extensional as well as compressional intraplate deformation. Therefore, intraplate deformation plays a pre-eminent role in monitoring the timing of stress changes within continental cratons. As such, the stratigraphic record of collision-related compressional intraplate deformation can contribute towards the dating of orogenic activity affecting a respective plate margin.

The spatial and temporal development of compressional intraplate deformations is controlled by the interaction of fluctuating intraplate stresses and spatial and temporal strength changes of the lithosphere. The strength configuration of the lithosphere primarily depends on its thermo-mechanical structure that can change due to deformation of the lithosphere and its transient thermal equilibration. In many intraplate settings sedimentary basins are characterized by relatively steady state strengths prior to the onset of compressional intraplate deformation (van Wees and Stephenson, 1995; Ziegler et al., 1995). As the onset of compressional basin reactivation reflects a relative increase or a reorientation of the intraplate stress field, intraplate deformations are particularly sensitive recorders of the timing of stress fluctuations. On the other hand, given a relatively constant stress field, spatial variations in the onset of compressional intraplate deformation are controlled by the spatial strength distribution within the lithosphere. However, as the strength of the lithosphere increases during basin inversion, locking of earlier inverted basins can control the progressive propagation of far-field compressional deformations into the interior of continental cratons (Ziegler, 1987; van Wees et al., 1992; Ziegler et al., 1995).

In this paper we summarize the occurrence of collision-related intraplate compressional structures, explore the evolution of the continental mantle-lithosphere, analyze the effects of different modes of rifting and sediment loading on the rheological structure of the lithosphere, and discuss mechanical controls on the development of collision-related compressional intraplate structures in terms of orogenic processes affecting the respective plate margins. We realize that this paper assumes a considerable knowledge of global geology and that it would have been desirable to illustrate all examples discussed. However, this would have expanded this already long paper beyond acceptable limits. Therefore, we have endeavoured to provide the reader with a comprehensive reference list.

2. Occurrence of collision-related intraplate compressional deformations

Compressional and transpressional intraplate structures can occur at distances of up to 1600 km from a collisional margin, as indicated, for instance, by the Paleocene deformation of the northern Alpine foreland (Ziegler, 1990) and the latest Carboniferous–Early Permian development of the Ancestral Rocky Mountains in the foreland of the Appalachian–Ouachita–Marathon orogen (Ross and Ross, 1985, 1986; Kluth, 1986; Stevenson and Baars, 1986; Oldow et al., 1989; Ziegler, 1989). Stress-induced whole lithospheric buckling, leading to the uplift of arches with a wavelength of 500–750 km (Nikishin et al., 1993; Burov et al., 1993), can affect even broader areas of continental cratons, as shown by the end-Silurian development of the Transcontinental and Labrador arches of North America (Ziegler et al., 1995). Similarly, Plio–Pleistocene accelerated subsidence of the North Sea Basin and contemporaneous uplift of the Fennoscandian Shield can be related to deflections of the lithosphere in response to the build-up of the present-day compressional stress field of northwestern Europe (Cloetingh and Kooi, 1992; van Wees and Cloetingh, 1996). This stress field reflects a combination of forces related to collisional coupling of the Alpine orogen with its foreland and Atlantic ridge-push forces; the latter appear to play a dominant role only along the Atlantic margins (Müller et al., 1992; Gölke and Coblentz, 1996).

Collision-related compressional intraplate structures can occur in back-arc domains of island arcs, in hinterlands (back-arc areas) of continent–ocean collision zones (Andean-type orogens), as well as in the foreland (fore-arc) and hinterland (back-arc) of continent–continent (Himalayan-type) collisional belts (Fig. 1B,C; Ziegler, 1989; Ziegler et al., 1995).

Inversion of back-arc rifts, such as the Eo–Oligocene extensional basins of the Sunda, South and East China shelves, that were compressionally deformed during Mio–Pliocene times (Sage and Letouzey, 1990; Letouzey et al., 1991; Matthews and Bransden, 1995; Sibuet and Hsu, 1997), is thought to result from an acceleration of convergence rates between the colliding plates, their increased mechanical coupling and the transmission of compressional stresses into the back-arc domain of the overriding plate (Uyeda and McCabe, 1983; Cloetingh et al., 1983, 1989; Jolivet et al., 1989; Ziegler, 1993). This model may also apply to the Devonian–Early Carboniferous back-arc rifts and small oceanic basins of the Rhenohercynian Basin that were closed at the onset of the Variscan orogeny (Ziegler, 1990). On the other hand, Palaeogene development of the basement involving the Rocky Mountains thrust belt in the Cordilleran hinterland is probably related to collisional over-thickening of the Laramide orogen at high strain rates, resulting in stress propagation into the American craton (Egan and Urquhart, 1993; Ziegler et al., 1995). Upthrusted basement blocks, as well as inverted tensional basins, occur also in the hinterland of the Andes in northern Argentina, Peru and Columbia (Uliana et al., 1995; Tankard et al., 1995; de Urreiztieta et al., 1996). A conceptual model for such deformations is given in Fig. 1B.

In contrast, the latest Cretaceous and Cenozoic compressional intraplate structures of western and central Europe developed in a foreland (fore-arc) setting with respect to the Alpine and Carpathian subduction systems (Ziegler, 1988, 1990). Similarly, the latest Cretaceous and Neogene compressional structures of the northern parts of the Arabian Peninsula, including inversion of the Palmyra Trough and the Syrian arc system as a whole, developed in the foreland domain of the South Anatolian–Tauride orogen (Yilmaz, 1993; Salel and Séguret, 1994; Litak et al., 1997). Also the Permian to Early Jurassic inversion structures of the Timan–Pechora area occur in a foreland setting with respect to the Polar Ural–Pay–Khoy subduction system (Nikishin et al., 1996). A further example of compressional foreland deformation is provided by the Late Carboniferous–Early Permian structures on the Sahara Platform that developed during the terminal phases of the Hercynian orogeny; contemporaneous hinterland compressional intraplate deformation controlled the development of the Ancestral Rocky Mountains of the U.S.A. (Ziegler, 1989; Ziegler et al., 1995).

The following focuses on the deformation of passive margins and continental platforms, located in foreland settings, that were subjected to intraplate compression during the subduction of oceanic domains separating them from arc–trench systems, and/or during their incorporation into continent–

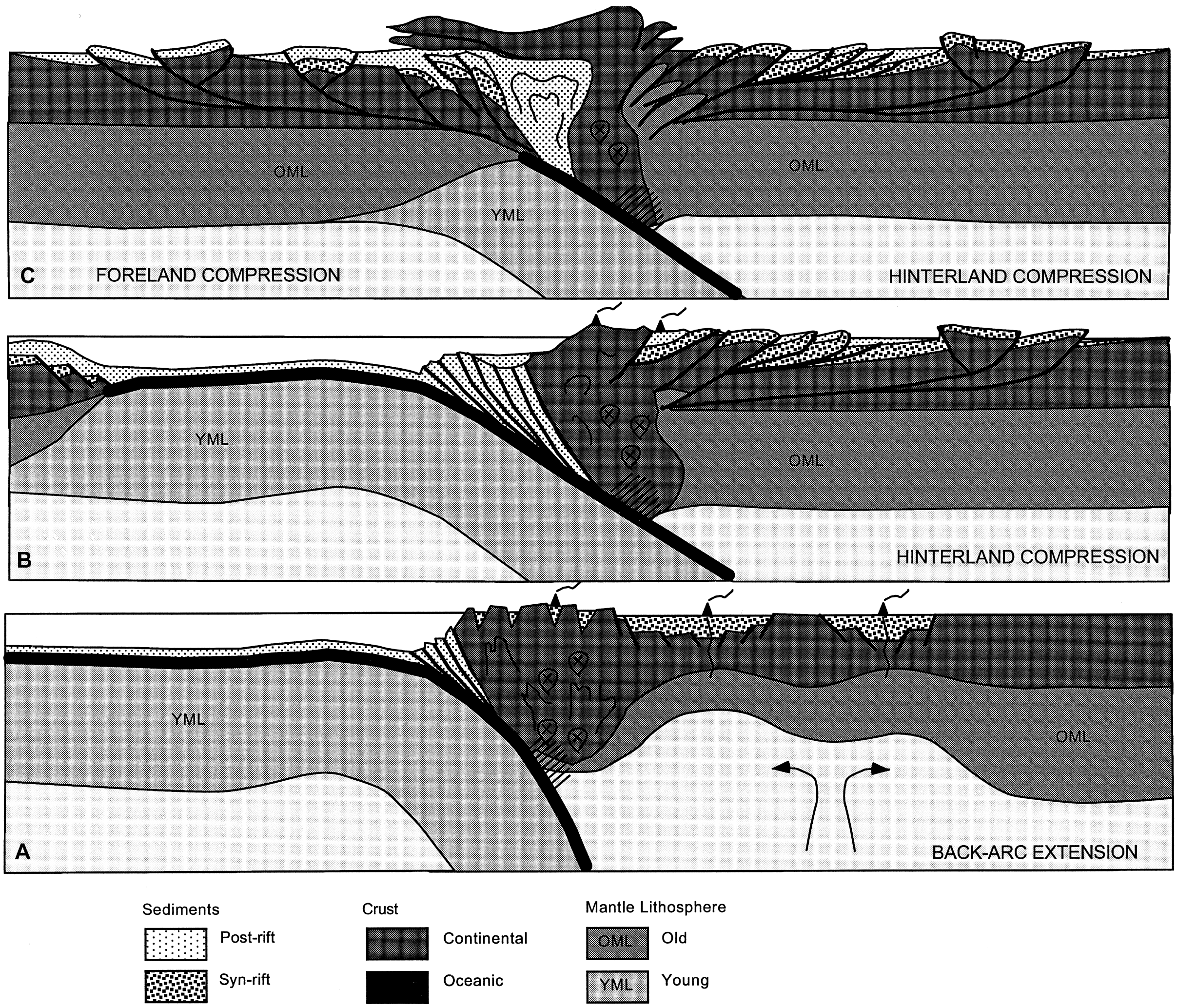

Fig. 1. Conceptual model illustrating intraplate (A) back-arc extension, (B) back-arc (hinterland) compression and, (C) collisional foreland and hinterland compression (not to scale).

continent collisional orogens. In this context, it is essential to consider the rheology of oceanic and continental lithosphere in terms of their yield strength and capability to transmit horizontal compressional stresses.

3. Oceanic and continental lithosphere

In contrast to continental lithosphere, the lithosphere of the present oceanic basins is young and ranges in age from Middle Jurassic to Recent (Cande et al., 1989). Its crustal parts consist of MORB basalts, sheeted dykes and gabbros; lateral thickness and compositional variations depend on spreading rates, the presence or absence of ridge-centred plumes (N- and E-MORB) and intraplate hotspot activity. The mantle part of oceanic lithosphere is composed of asthenospheric material depleted in basaltic components, dominated by harzburgite and dunite, that grades at depth into lherzolite, representing solidified, less depleted asthenospheric mantle (Bott, 1982; Gass, 1989; Wilson, 1989). In view of its rather uniform composition, oceanic lithosphere has a relatively simple rheological structure that basically depends on its thermal regime, controlling its age-dependent thickness. As mature oceanic lithosphere is considerably stronger than all types of continental lithosphere (Burov and Diament, 1995; Bertotti et al., 1997), stresses exceeding the strength of continental lithosphere can be transmitted through it.

Continental lithosphere is generally much older and has a more complex, depth-dependent rheological structure in which the thickness, mineralogical composition, fluid content and thermal regime of the crust, as well as the thickness of the mantle-lithosphere play a dominant role. The thickness of old continental mantle-lithosphere is mainly a function of ancient depletion events (magma extraction), increasing its buoyancy, and to a lesser degree of the potential temperature of the asthenosphere. Studies of mantle xenoliths indicate that the ancient continental mantle-lithosphere of South Africa is much more depleted (prevalence of harzburgite and dunite over lherzolite) and thus has a lower density than e.g. the Palaeozoic lithosphere of Europe (M. Wilson, pers. commun., 1997). Stabilized continental lithosphere consists of the mechanically strong upper crustal layer that is separated by the mechanically weaker lower crust from the strong upper part of the mantle-lithosphere. Pre-existing crustal discontinuities can significantly weaken the crust (see Fig. 3; Kusznir and Park, 1987; Buck, 1991; Stephenson and Cloetingh, 1991; Cloetingh and Banda, 1992; Ziegler et al., 1995; Cloetingh and Burov, 1996).

Published synthetic strength envelopes for continental mantle-lithosphere are based on the assumption that it is exclusively composed of solidified asthenospheric material. However, old continental mantle-lithosphere is probably heterogeneous and comprises depleted harzburgite, less depleted lherzolite and layers of 'frozen-in' partial melts (Wilson, 1989). Moreover, depending on its evolution, the continental mantle-lithosphere may contain, at least in its upper parts, significant amounts of continental crustal material, the rheological property of which differs from that of peridodites and pyroxenites. During orogenic processes continental crustal material can be subducted to depths of 100 to 150 km and more, and, due to eclogitization at depths between 50 and 75 km, depending on temperatures and the availability of hydrous fluids, is transferred across the geophysically defined Moho (V_p break over from $\leq$7.8 to 8.0–8.2 km/s) into the mantle (Goffé and Chopin, 1986; Austrheim, 1991; Andersen et al., 1991; Eide and Torsvik, 1996; Austrheim et al., 1997; Bousquet et al., 1997; Marchant and Stampfli, 1997). Apparently, some of this subducted crustal material can be retained in the mantle-lithosphere over long time spans and can give rise to discrete reflection bundles that are probably related to density and velocity contrasts between eclogites and pristine mantle material (Austrheim and Mørk, 1988). Examples of mantle reflectors discussed below support the concept that the continental mantle-lithosphere may indeed contain crustal material.

In this respect it ought to be kept in mind that the density of felsic eclogites ($\rho = 3.06$–3.08) is lower than that of the average mantle ($\rho = 3.35$), the density of intermediate eclogites ($\rho = 3.31$–3.37) is similar to that of the mantle, whereas mafic eclogites ($\rho = 3.53$–3.56) are denser than the mantle (Le Pichon et al., 1997). Therefore, subduction slabs consisting of eclogitized oceanic lithosphere are prone to detach from the lithosphere and to sink into the mantle, whereas subducted continental crustal material can be retained in the mantle. On the other hand, during post-orogenic re-equilibration of the lithosphere, subducted and eclogitized material that is retained in the mantle-lithosphere will be heated over a time span of some 20–30 million years and will be transformed, depending on the availability of water, at depths shallower than 65–75 km into less dense granulite ($\rho = 2.75$–3.43) (Le Pichon et al., 1997). This probably accounts for the retention of subducted continental material within the mantle-lithosphere over very large time spans.

For instance, on a deep reflection-seismic line crossing the western parts of the North Carpathian foredeep, south-dipping lower crustal reflectors of the down-flexed European foreland crust terminate at a depth of $\pm$12 s TWT (35–40 km) against a sub-horizontal reflection band, interpreted as the Moho (Tomek, 1993; Tomek and Hall, 1993); this interpretation is supported by refraction-seismic data (Tomek, 1988). This suggests, that during the Neogene development of the Carpathian foreland basin, European continental crustal material was subducted, entered the eclogite stability field and thus was transferred across a newly formed Moho into the mantle-lithosphere. A second example comes from the Hebrides Shelf (Scotland) where a sub-horizontal upper mantle reflector is observed (W-reflector) that is located about 6 s TWT beneath the present Moho. The W-reflector is interpreted as a fossil Moho that had marked the boundary between crustal and ultramafic mantle material at the end of the Caledonian orogeny; if this interpretation applies, the interval between the old and present Moho is likely to contain significant amounts of continental crustal rocks in eclogite- and granulite-facies (Snyder and Flack, 1990; Morgan et al., 1994). Alternatively, the W-reflector may correspond to the base of a major magmatic crustal underplate (M. Wilson, pers. commun., 1997), in which case it would not be relevant to our argument.

In this respect it should be noted that the geophysically defined continental Moho is not always a uniform, sharp discontinuity, but rather a complex and variable transition zone. Generally it ranges in thickness between less than 1 km and 5 km, but can expand to some 10 km. The normal-incidence reflection seismically defined Moho is characterized by very variable reflectivity and reflection patterns and does not always coincide with the Moho defined by wide-angle reflections, perhaps due to crustal anisotropy or to the gradational nature of the crust–mantle boundary (Bott, 1982; Jones et al., 1996; Hammer and Clowes, 1997; Sapin and Hirn, 1997). As the Moho can sharply truncate the orogenic fabric of the crust (e.g. Ziegler, 1996; Abramovitz et al., 1997), it is questioned whether the geophysical and petrological crust–mantle boundaries always coincide.

The concept that continental mantle-lithosphere can indeed contain crustal material is compatible with the occurrence of upper mantle reflection bundles, the dip of which generally conforms with the orogenic fabric of the overlying crust (e.g. Gulf of Bothnia: Ansorge et al., 1992; southern Baltic Sea: BABEL Working Group, 1993; Abramovitz et al., 1997; Skagerrak Telemark tongue: Lie and Husebye, 1994; southern North Sea: Kearey and Rabae, 1996; MONA LISA Working Group, 1997; Hebrides Shelf Flannen thrust: Morgan et al., 1994; Carpathians: Tomek, 1993; central Urals: Berzin et al., 1996; Abitibi province, Canada: Clowes et al., 1996). The amplitude of these dipping reflection bundles indicates that they originate from interfaces characterized by significant acoustic impedance contrasts. As such, they are generally interpreted as being associated with compositional changes, related to relict subduction zones along which oceanic and/or continental crustal material was inserted into the mantle (Lie and Husebye, 1994; Morgan et al., 1994; Clowes et al., 1996; Price et al., 1996). The age of such relic subduction zones ranges from Neoproterozoic (Baltic Shield) to Cenozoic (Carpathians).

Mass balance studies on arc–trench systems and orogenic belts suggest that significant amounts of crustal and even sedimentary material can be subducted to great depths (von Huene and Scholl, 1991, 1993; Ziegler et al., 1996; Marchant and Stampfli, 1997). This material is partly recycled into the sublithospheric mantle by means of slab detachment (Andersen et al., 1991; Armstrong, 1991; Davies and von Blanckenburg, 1996) and partly brought back to higher crustal levels during phases of collisional over-thickening and back-folding of orogenic wedges (Schmid et al., 1996) and possibly also by buoyancy forces (Chemenda et al., 1996). Moreover, during the post-orogenic re-equilibration of the lithosphere with the asthenosphere, eclogitic material (V_p = 8.0–8.4, ρ = 3.06–3.56) contained in orogenically thickened crustal root zones is heated and can be transformed above depths of 65–75 km into granulites (V_p = 7.0–7.8, ρ = 2.75–3.43) and amphibolites (V_p = 6.6–7.2, ρ = 2.83–3.06); as these phase transformations are associated with a density and velocity decrease, they entail crustal uplift and a re-equilibration of the Moho (Le Pichon et al., 1997). Nevertheless, some of the subducted crustal material is apparently retained within the continental mantle-lithosphere, as suggested by the occurrence of mantle reflectors. Compositional changes associated with ancient subduction zones presumably form discontinuities within the continental mantle-lithosphere.

Furthermore, the refraction seismically defined velocity layering of continental mantle-lithosphere (Ansorge et al., 1992; Thybo and Perchuc, 1997), as well as the composition of mantle xenoliths (peridodites, eclogites and granulites), indicate that it is characterized by considerable petrological heterogeneities and stratifications, including volatileenriched zones (Gurney et al., 1991; Saunders et al., 1992; Menzies and Bodinier, 1993). Moreover, the thermal boundary layer of old continental mantle-lithosphere is metasomatically enriched with volatiles due to very-low-fraction melts leaking up from the asthenosphere through time (McKenzie, 1989; Saunders et al., 1992; Smith, 1993; Thompson and Gibson, 1994).

From the above we conclude, that the composition of old continental mantle-lithosphere, although dominated by a peridodite lithology, differs from that of relatively young oceanic mantle-lithosphere and contains, at least locally, crustal material in eclogite and/or granulite facies. Only in areas where during a rifting event the old mantle-lithosphere was completely removed and replaced by upwelling asthenosphere is it likely that the new mantle-lithosphere, that formed during the post-rift stage, consists exclusively of solidified asthenospheric material.

In view of this, we suspect that the rheological properties of newly accreted and old subcontinental lithospheric mantle are not necessarily the same. However, at this time, we are unable to quantify this statement. Nevertheless, we assume that mantle discontinuities related to ancient subduction zones, containing granulites and rheologically weak eclogites (Austrheim et al., 1997), and granulitic lower crustal roots that developed in response to plume-induced magmatic underplating (Touret, 1995; J.L.R. Touret, pers. commun., 1997), present potential weakness zones that are prone to tensional, as well as compressional reactivation. As such, they may play an important role in localizing extensional and compressional intraplate deformation. Moreover, boundaries between old and young, newly accreted subcontinental mantle-lithosphere probably present additional weakness zones, as suggested, for instance, by the Early Cretaceous nucleation of subduction zones along the South Alpine and Austroalpine lower plate margins (Favre and Stampfli, 1992; Stampfli and Marchant, 1997).

4. Rifting of continental lithosphere

During rifting, culminating in the opening of new Atlantic-type oceanic basins, the continental lithosphere is stretched and the subcrustal mantle thermally attenuated, particularly in the presence of a mantle plume. Upon termination of rifting activity, or after crustal separation has been achieved and the respective passive margins have moved away from the seafloor spreading axis, the thermally destabilized continental lithosphere re-equilibrates with the asthenosphere (McKenzie, 1978; Steckler and Watts, 1982; Wilson, 1993a; Ziegler, 1996). During this process new mantle-lithosphere, consisting of solidified asthenospheric material, is accreted to the attenuated old continental mantle-lithosphere. This aspect is emphasized in Fig. 2 as rheological properties of old and new mantle-lithosphere may differ.

Rifts superimposed on ancient suture zones often display the geometry of simple-shear lithospheric extension (e.g. central Atlantic, Gulf of Suez, Favre and Stampfli, 1992). Pre-existing crustal and mantle–lithospheric discontinuities, as well as the mechanical anisotropy of the lithospheric mantle, probably determine their location and polarity (Piqué and Laville, 1995; Vauches et al., 1997). On the other hand, rifts cross-cutting the orogenic fabric of the crust (e.g. Labrador Sea, North Sea rift) are more prone to display a pure-shear geometry (Ziegler, 1996). Under conditions of pure-shear lithospheric extension, conjugate passive margins are likely to display at the crustal separation stage a more or less symmetrical lithospheric configuration. Therefore, their post-rift evolution is similar with variations in their rheological structure depending largely on the thickness of the passive margin sedimentary wedge (starved versus overfilled, e.g. SE Greenland and Labrador margins; Chalmers and Laursen, 1995). However, under conditions of simple-shear extension, the lithospheric configuration of conjugate margins can differ considerably at the end of the rifting stage (Fig. 2). At lower plate margins, the crust can be highly extended whereas the continental mantle-lithosphere may be little attenuated; in their distal parts, the old mantle-lithosphere can be denuded and in sheared contact with rotated upper crustal fault blocks and syn-rift sediments. In contrast, at upper plate margins, the crust may be less extended whereas the continental mantle-lithosphere can be strongly attenuated with asthenospheric material ascending close to the base of the crust (Wernicke, 1985; Wernicke and Tilke, 1989; Boillot et al., 1989; Lister et al., 1991; Favre and Stampfli, 1992; Piqué and Lav-

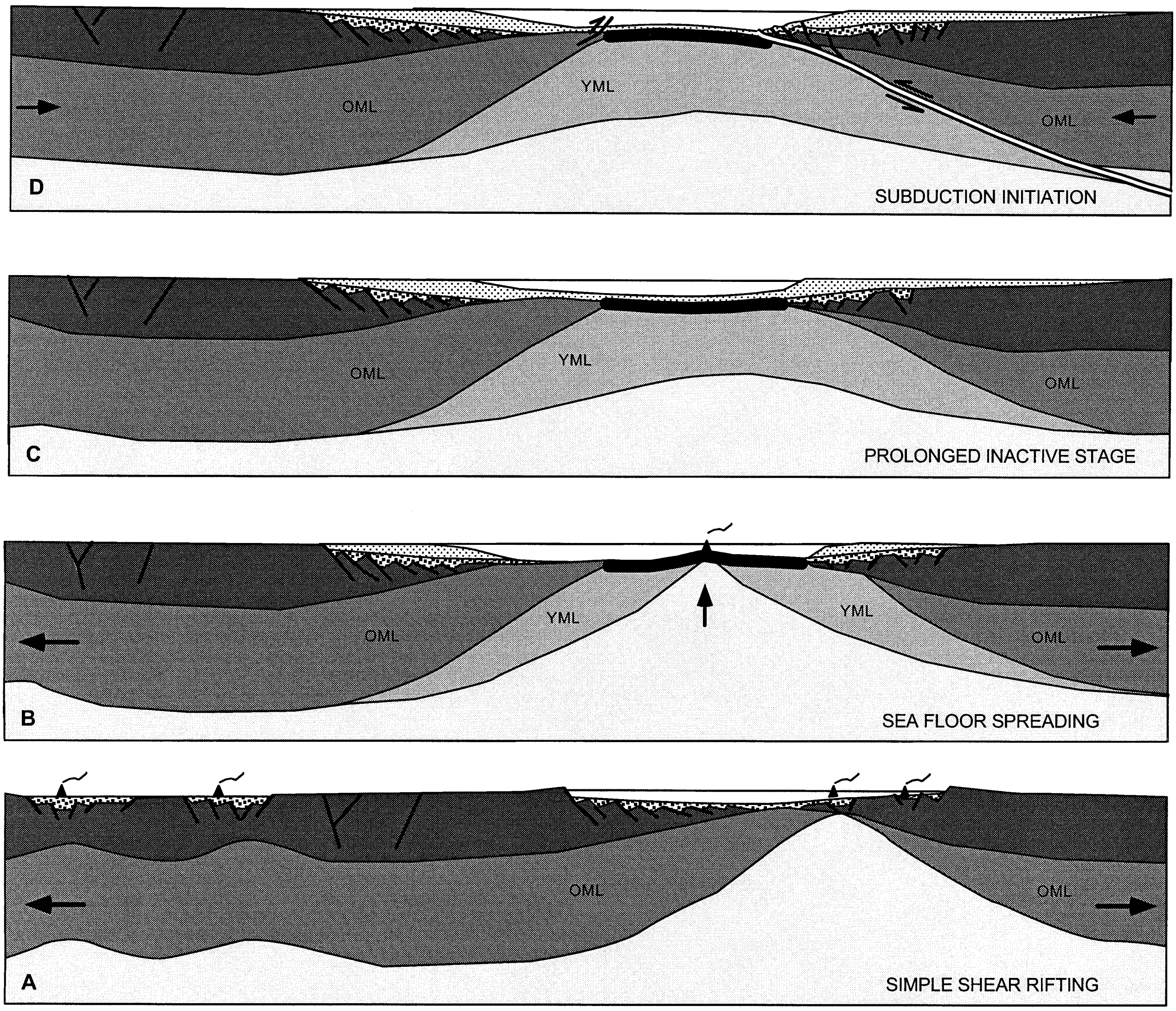

Fig. 2. Conceptual model illustrating (A) simple-shear rifting leading to (B) opening of an oceanic basin, followed by (C) post-seafloor spreading evolution of mantle-lithosphere, and (D) inception of a new subduction zone at the upper plate continental margin, employing the boundary between old and new mantle-lithosphere, and contemporaneous minor intraplate deformations along the conjugate lower plate margin (not to scale, legend same as Fig. 1).

ille, 1995; Brun and Beslier, 1996; Froitzheim and Manatschal, 1996).

These differences in lithospheric configuration of simple-shear conjugate margins at the end of the rifting stage evidently have repercussions on their post-rift subsidence history and on their rheological structure, even after full thermal relaxation of the lithosphere. In order to quantify these effects, we have applied a recently developed 1D two-layered lithospheric stretching model, incorporating the effects of heat production by the crust and its sedimentary thermal blanketing, in an effort to analyze the thermo-mechanical evolution of the lithosphere and to predict its palaeo-rheology (van Wees et al., 1996; Bertotti et al., 1997; van Wees and Beekman, 1998).

For modelling purposes we chose a time frame of 100 Ma. Of this, the first 10 Ma (between 100 Ma and 90 Ma in Fig. 3) correspond to the rifting stage, culminating in separation of the conjugate upper and lower plate margins, and the following 90 Ma to

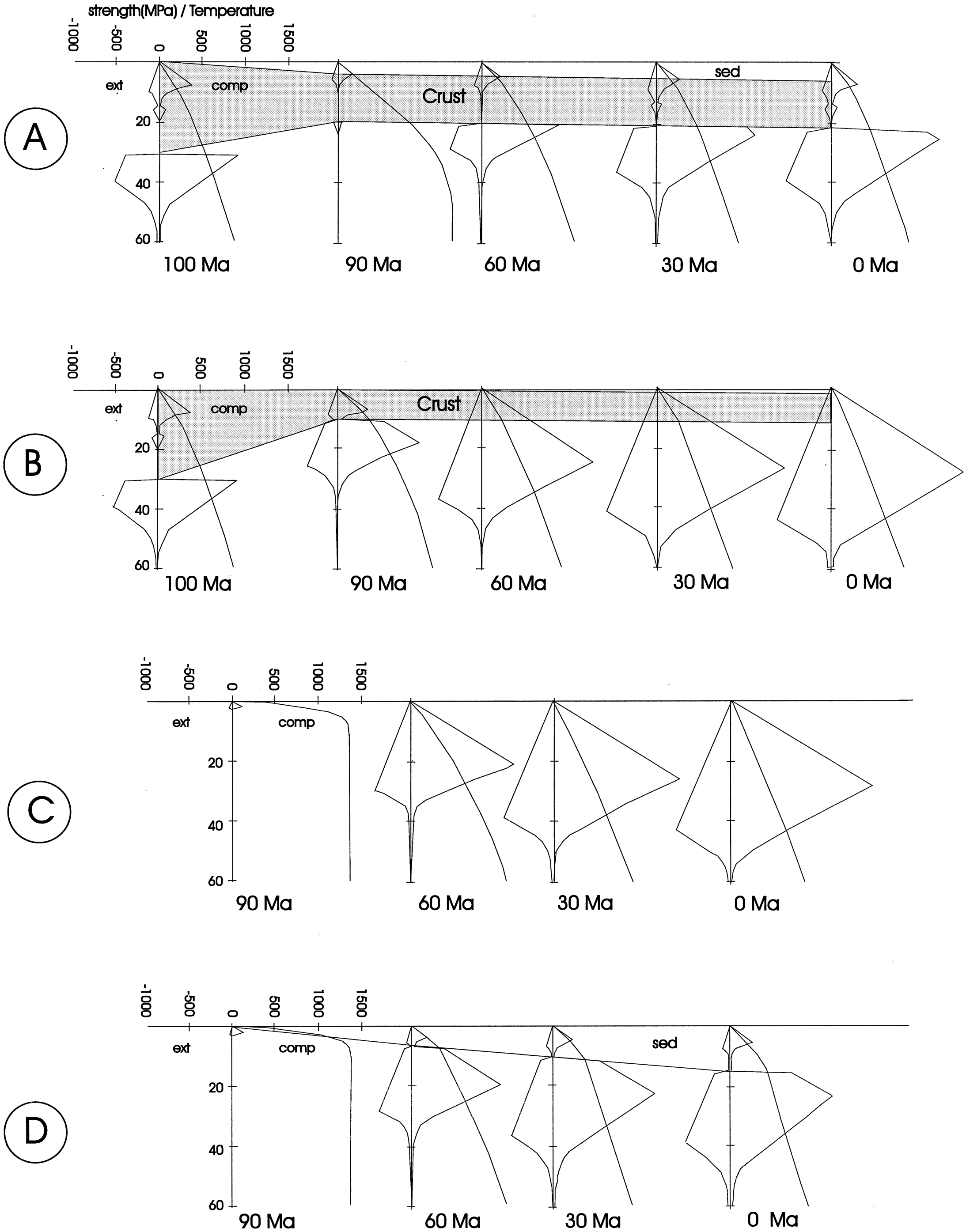

Fig. 3. Depth-dependent rheological models for the evolution of lower plate and upper plate passive margins and oceanic lithosphere. Modelling parameters are listed in Tables 1 and 2. (A) Upper plate passive margin ($\delta = 2$, $\beta = 10$), characterized complete sediment fill of accommodation space at end of post-rift phase ($\rho_s = 2100$, 7.084 km). (B) Lower plate passive margin ($\delta = 3$, $\beta = 1.1$), marked by sediment starvation at end of post-rift phase (1 km sediments). (C) Oceanic lithosphere with a thin sedimentary cover of 1 km. (D) Oceanic lithosphere with a gradually increasing sedimentary cover reaching a maximum of 15 km.

Table 1
Modelling parameters

Symbol	Name	Value
a	initial lithospheric thickness	100 km
c	initial crustal thickness	30 km
T_a	asthenospheric temperature	1333ºC
k	thermal diffusivity	1×10^{-6} m^2 s^{-1}
ρ_c	surface crustal density	2800 kg m^{-3}
ρ_m	surface mantle density	3300 kg m^{-3}
ρ_w	water density	1030 kg m^{-3}
ρ_s	density sediments	2100–2650 kg m^{-3}
α	thermal expansion factor	3.2×10^{-5}ºC
δ	crustal stretching factor	
β	subcrustal stretching factor	

the seafloor spreading stage during which oceanic lithosphere is accreted to the diverging plates. For modelling purposes we assumed at the end of the rifting stage for the upper plate margin a crustal thickness of 15 km ($\delta = 2$) and a remaining mantle-lithosphere thickness of 7 km ($\beta = 10$), and for the lower plate margin a crustal thickness of 10 km ($\delta = 3$) and a mantle-lithosphere thickness of 63.6 km ($\beta = 1.1$) (Fig. 3A,B; Tables 1 and 2). Results show that through time the evolution of strength envelopes for lower and upper plate passive margins differs strongly. In principle, during rifting increased heating of the lithosphere causes its weakening; this effect is most pronounced at the moment of crustal separation. However, upper and lower plate margins show a very different evolution, both during the rifting and post-rift stage.

At the moment of crustal separation, upper plate margins are very weak due to strong attenuation of the mantle-lithosphere and the ascent of the asthenospheric material close to the base of the crust. During the post-rift evolution of such a margin, having a crustal thickness of 15 km, the strength of the lithosphere increases gradually as new mantle is accreted to its base and cools during the re-equilibration of the lithosphere with the asthenosphere (Fig. 3A and Fig. 4A). In contrast, the evolution of a sediment-starved lower plate margin with a crustal thickness of 10 km is characterized by a syn-rift strength increase due to extensional unroofing of the little attenuated mantle-lithosphere; the strength of such a margin increases dramatically during the post-rift stage due to its progressive cooling (Fig. 3B and Fig. 4B). At the time frame of 0 Ma, a sediment-starved lower plate margin is considerably stronger than the conjugate upper plate margin for which a sedimentary cover of about 7 km was assumed. The strength evolution of an upper plate margin is initially controlled by the youthfulness of its mantle-lithosphere, its thicker crust and later by the thermal blanketing effect of sediments infilling the available accommodation space. On the other hand, the strength of oceanic lithosphere, that is covered by thin sediments only, increases dramatically during its 90 million years of evolution and ultimately exceeds the strength of both margins, even if these are sediment starved (Fig. 3C and Fig. 4). However, the strength of 90 Ma oceanic lithosphere that has been progressively covered by very thick sediments is significantly reduced (Fig. 3D) to the point that it approaches the strength of a sediment-filled lower plate margin (Fig. 4).

To test the effects of sediment infill and thermal blanketing on the strength evolution of upper and lower plate passive margins, we ran a wide range of models, assuming that sediments completely fill the tectonically created accommodation space (sediment overfilled), adopting different sediment densities and corresponding sediment thickness variations

Table 2
Rheological and thermal parameters of crust and lithospheric materials

Layer	Rheology	Conductivity W m^{-10} C^{+1}	Heat production W m^{-3}
Sediments (models A, B, C)	quartzite	2	0.2×10^{-6}
Sediments (model D)	quartzite	1.4	1.8×10^{-6}
Upper crust	quartzite	2.6	1.88×10^{-6}
Lower crust	diorite	2.6	0.5×10^{-6}
Mantle	olivine	3.1	0

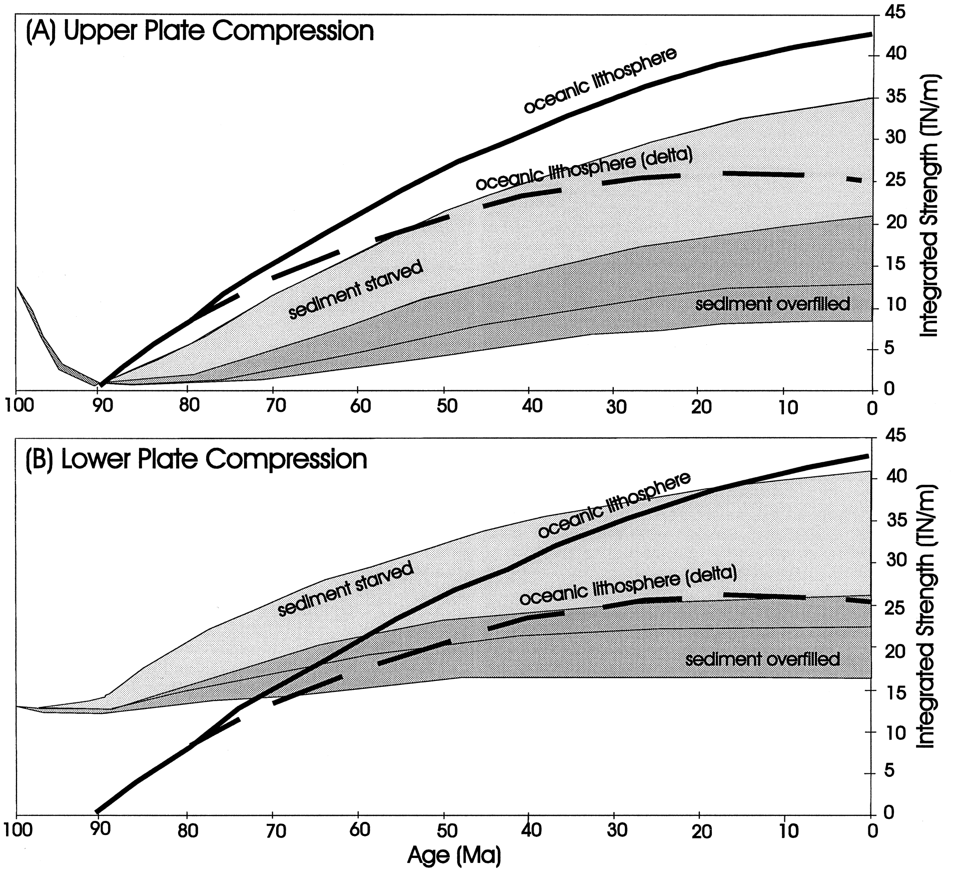

Fig. 4. Integrated compressional strength evolution of sediment-starved and sediment-filled (A) upper plate, and (B) lower plate passive margins, compared to the integrated strength evolution of oceanic lithosphere with thin and thick sediment cover as in Fig. 3. Shaded areas demonstrate strong sensitivity of integrated strength to sediment infilling, ranging from sediment starvation (highest strength values) to complete fill of accommodation space (dark shading, lowest strength values). Curves in dark shaded area correspond to different sediment densities and related range in sediment thickness (see Table 3).

(Table 3; Fig. 4). Results show that the integrated strength of a margin is markedly reduced by a thick syn- and post-rift sedimentary prism. However, despite the strong sediment fill effect on the integrated strength, earlier identified first-order differences between upper and lower plate margins remain very clear.

Compared to oceanic lithosphere, both with a 1-km sedimentary cover (Fig. 3C and Fig. 4) and a 15-km-thick cover (Fig. 3D), a sediment-overfilled

Table 3
Sediment-overfilled scenario (up to water surface), adopted in Fig. 4

		Tectonic air-loaded subsidence (m)	Accumulated sediment fill (m)		
			$\rho_s = 2100$	$\rho_s = 2500$	$\rho_s = 2650$
Upper plate	syn-rift	1474	4396	7065	9148
	post-rift	2375	7084	11384	14739
Lower plate	syn-rift	2599	7752	7752	16130
	post-rift	3332	9939	15972	20679

upper plate margin (applying $\rho_s = 2100$) is characterized by lower integrated strength values throughout its evolution. However, for a lower plate margin conditions are dramatically different. Up to 20–70 million years after rifting, the lower plate integrated strength values are significantly higher than those for oceanic lithosphere, both with a thin and a thick sedimentary cover.

From our strength calculations it is evident that at any stage the upper plate margin is weaker than oceanic lithosphere and the conjugate lower plate margin. This suggests that the upper plate margin is the most likely candidate for compressional reactivation and the initiation of a subduction zone. For realistic sediment density and infill values ($\rho_s <= 2500$), also the upper plate margin tends to grow stronger in time, indicating that after a prolonged post-rift stage (>70 million years) the observed localization of deformation and subduction along such a margin, instead of on the adjacent continent, should be facilitated by weakening mechanisms that are not incorporated in our standard rheological assumptions for the lithosphere, such as pre-existing crustal and mantle discontinuities and the boundaries between old and newly accreted lithospheric mantle.

Our modelling shows that the compressional yield strength of passive margins can vary considerably, depending on their lithospheric configuration, sedimentary cover and age. Lower plate margins, at which much of the old continental mantle-lithosphere is preserved, are considerably stronger than upper plate margins at which asthenospheric material is accreted to the strongly attenuated old mantle-lithosphere. Although mature oceanic lithosphere is characterized by a high compressional yield strength, it can be significantly weakened in areas where thick passive margin sedimentary prisms or deep-sea fans prograde onto it (e.g. Gulf of Mexico: Worrall and Snelson, 1989; Niger Delta: Doust and Omatsola, 1989; Bengal fan: Curray and Moore, 1971).

5. Scenarios for fore-arc and foreland compressional deformation

We envisage four collision-related scenarios for the build-up of horizontal intraplate compressional stresses in fore-arc and foreland domains, namely: (1) during the initiation of subduction zones along passive margins or within oceanic basins; (2) during impediment of subduction processes caused by the arrival of a more buoyant crustal element, such as an oceanic plateau, transform ridge, arc or a micro-continent at a subduction zone; (3) during the initial collision of an orogenic wedge with a passive margin; (4) during post-collisional over-thickening of an orogenic wedge.

5.1. Initiation of subduction zones

As we deal with a finite globe and interlocking lithospheric plates, generation of new oceanic lithosphere is compensated elsewhere by subduction of a commensurate amount of lithosphere, and vice versa. Nucleation of new subduction zones in oceanic basins or at continental margins is probably related to the build-up of far-field regional compressional stress fields resulting from plate interaction during phases of plate-boundary re-organization. Plate-moving mechanisms, controlling the interaction of lithospheric plates, and their relative importance, have been reviewed by Bott (1993), Pavoni (1993), Wilson (1993b), Ziegler (1993) and Ziegler et al. (1995). The present stress regime of the globe, as well as current plate movements, can be adequately explained in terms of plate-boundary forces and collisional resistance and do not require a significant contribution from sub-lithospheric mantle flow (Zoback, 1992). However, Phanerozoic motions and interaction of lithospheric plates suggest that shear-traction exerted by the convecting asthenosphere on the base of the lithosphere can play an important role as a plate-moving mechanism, particularly when operating in constructive interference with plate-boundary forces (Ziegler, 1993; Ziegler et al., 1995). In view of this, we envisage a greater contribution of mantle dynamics to plate kinematics than suggested by the Orowan–Elsasser model (see Bott, 1982).

An example of plate interaction controlling the development of a new subduction zone is the Late Jurassic–Early Cretaceous closure of the Vardar and Dinaric oceanic basins that is related to space constraints developing in the Mediterranean domain due to opening of the central Atlantic and the onset of a sinistral translation between Laurasia and Africa–

Arabia. Similarly, Late Cretaceous and Palaeogene propagation of Alpine subduction systems into the oceanic basins of the western Mediterranean area is related to the collisional interaction of the Africa–Arabian and Eurasian plates during the progressive opening of the South Atlantic–Indian Ocean (Smith and Spray, 1984; Robertson and Dixon, 1984; Scotese et al., 1988; Ziegler, 1988, 1990, 1993; Ricou, 1996).

As thermally stabilized oceanic lithosphere, covered by thin sediments, is very strong, major stresses are required to imbricate it (Fig. 3C). Ridge-push forces alone are not large enough to imbricate such oceanic lithosphere and cause the initiation of its gravity-driven, passive sinking into the asthenosphere (Cloetingh et al., 1983, 1989; Mueller and Phillips, 1991; Ziegler, 1993). As spreading axes are the rheologically weakest part in actively opening oceans, they are prone to be overpowered by compressional stresses. However, even the lithosphere of extinct spreading ridges becomes very strong with time (e.g. Indian Ocean, Labrador Sea; see Cande et al., 1989). Only along passive margins where major sedimentary complexes prograde onto oceanic lithosphere, attaining thicknesses in the order of 10 to 15 km, is the oceanic lithosphere significantly weakened by deep burial and thermal blanketing (Fig. 4; Erikson, 1993); this renders it prone to deformation upon build-up of a major regional compressional stress field during phases of plate-boundary re-organization (Ziegler et al., 1995). Nevertheless, the adjacent, deeply buried distal parts of continental margins can be rheologically even weaker and are thus even more prone to fail (Fig. 4). Moreover, pre-existing crustal and lithospheric discontinuities, such as fault systems of sheared margins and intra-oceanic transform faults, are also prone to reactivation upon build-up of compressional stresses (e.g. compressional deformation of central Indian Ocean: Chamot-Rooke et al., 1993; Beekman et al., 1996; Puysegur ridge: Collot et al., 1995).

As the weakest part of the lithosphere is the locus of compressional strain concentration, the lithosphere of young oceanic basins will be preferentially imbricated in the area of spreading axes or along transform faults (e.g. earliest Late Cretaceous activation of the peri-Arabian and inner-Tauride intra-oceanic subduction zones and subsequent ophiolite obduction: Michard et al., 1994; Polat et al., 1996; Hacker and Gnos, 1997). Under such conditions, it is unlikely that the flanking passive margins will be deformed. This is exemplified by the evolution of the Oman passive margin that remained stable during the end-Albian development of the intra-oceanic Semail arc–trench system but was subsequently destabilized during their collision, preceding its incorporation into a flexural foreland basin (Robertson, 1987).

However, in oceanic basins in which seafloor spreading has ceased some time ago and the orientation of transform faults is such that they cannot be compressionally reactivated under the prevailing stress regime, compressional deformations will be preferentially located along and within the passive margins prior to the development of a new subduction zone along one of them (Fig. 2D). Once a subduction zone has developed, possibly employing pre-existing weakness zones, such as the flexural boundary between continental and oceanic crust beneath a thick passive margin sedimentary prism (Erikson, 1993), deep-reaching shear zones, or perhaps the boundary between old and new mantle-lithosphere, it will be the locus of stress release and no further intraplate deformations will occur along the conjugate margin. A possible example of this is provided by the Eocene compressional structures occurring at the continent–ocean transition along the Armorican lower plate margin of the Bay of Biscay. These developed during the westward propagation of the Pyrenean subduction zone along the Iberian upper plate margin and its advance into the North Atlantic Ocean (Grimaud et al., 1982; Masson and Parson, 1982; Masson et al., 1994). During the Oligocene and Miocene phases of the Pyrenean orogeny no further deformations occurred along the Armorican margin.

5.2. Subduction impediment

Subduction processes can be impeded if buoyant material, such as a spreading ridge, seamount, transform ridge, oceanic plateau or microcontinent, collides with a mature arc–trench system (Mueller and Phillips, 1991; Cloos, 1993). If such a subduction obstacle is relatively small (e.g. seamount), either the entire edifice is subducted or its crustal part is sheared off and incorporated into the accretionary wedge (Lallemand et al., 1989; von Huene et al., 1997). However, larger obstacles, such as aseismic ridges, can provide sufficient subduction resistance to deform the arc–trench geometry, causing the build-up of compressional stresses in the subducting plate and potentially its imbrication (e.g. central Japan Nankai Trough, Bonin and the Zenisu ridges; Lallemand et al., 1989, 1992). A microcontinent, depending on its dimension, lithospheric configuration and rheological structure (sediment and crustal thickness), may be subducted or resist subduction to the degree that compressional stresses build up within the subducting plate until either the upper parts of the microcontinent are sheared off and subduction of its lower parts commences, or a new subduction zone develops along its distal, non-collisional margin (Fig. 5B,C). Prior to the subduction of a microcontinent, or subduction progradation to its distal margin, compressional stresses building up in the lower plate may be large enough to cause deformation of a passive margin and its shelves that is separated from the colliding microcontinent by a relatively old, and thus rheologically strong oceanic basin (Fig. 5C). However, once subduction of the microcontinent has commenced, or a new subduction zone has been activated along its distal margin, compressional stresses within the lower plate relax due to decoupling of the subducting and overriding plates as a consequence of sediment subduction (von Huene and Scholl, 1993; Genser et al., 1996).

The Eratosthenes Seamount is a modern example of a microcontinent that collided with an arc–trench system (Fig. 6). Its crustal configuration and magnetic and gravity signatures indicate its continental nature (Makris et al., 1983, 1994; Makris and Wang, 1994). The Eratosthenes Seamount collided during the Late Miocene to Early Pliocene with the

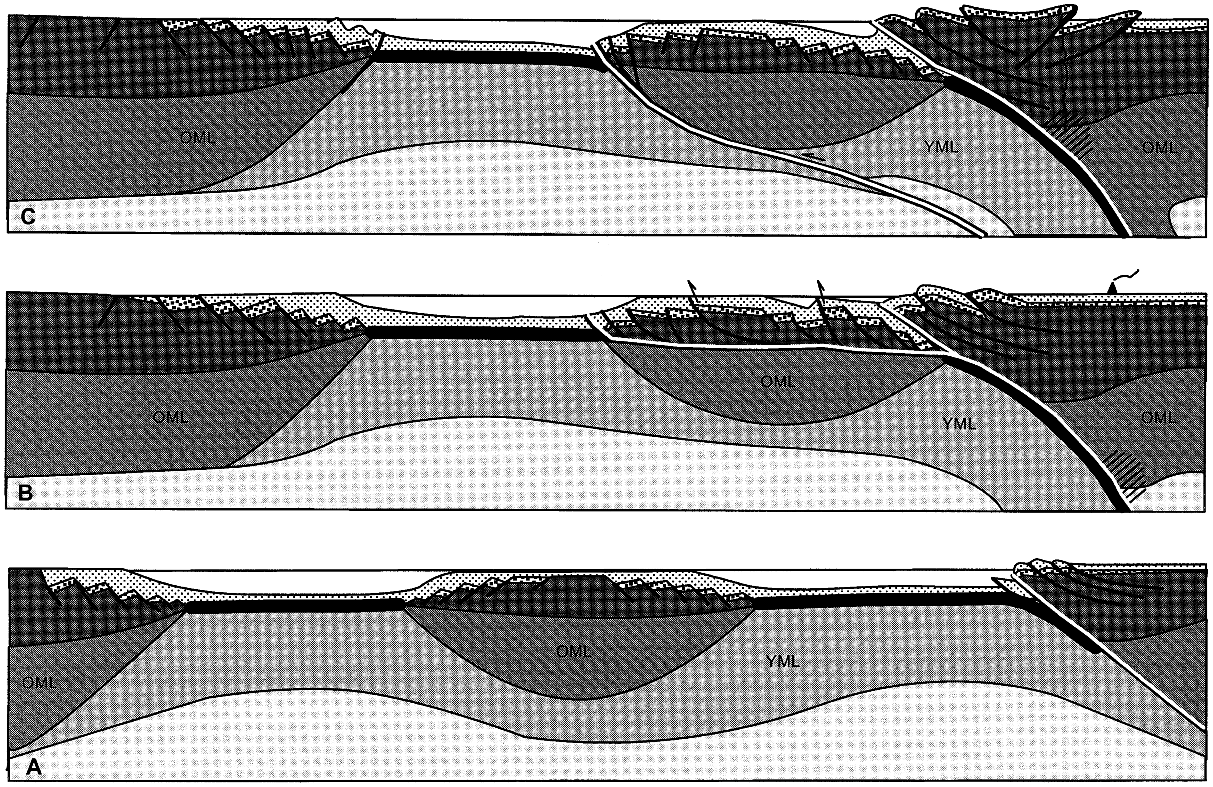

Fig. 5. Conceptual model illustrating (A) subduction of oceanic lithosphere separating a microcontinent from an arc–trench system, (B) collision of a microcontinent with an arc–trench system and detachment and deformation of its crustal parts, (C) collision of a microcontinent and subduction progradation to its distal margin, accompanied by intraplate deformation of a passive margin separated from the microcontinent by an oceanic basin (not to scale, legend same as Fig. 1).

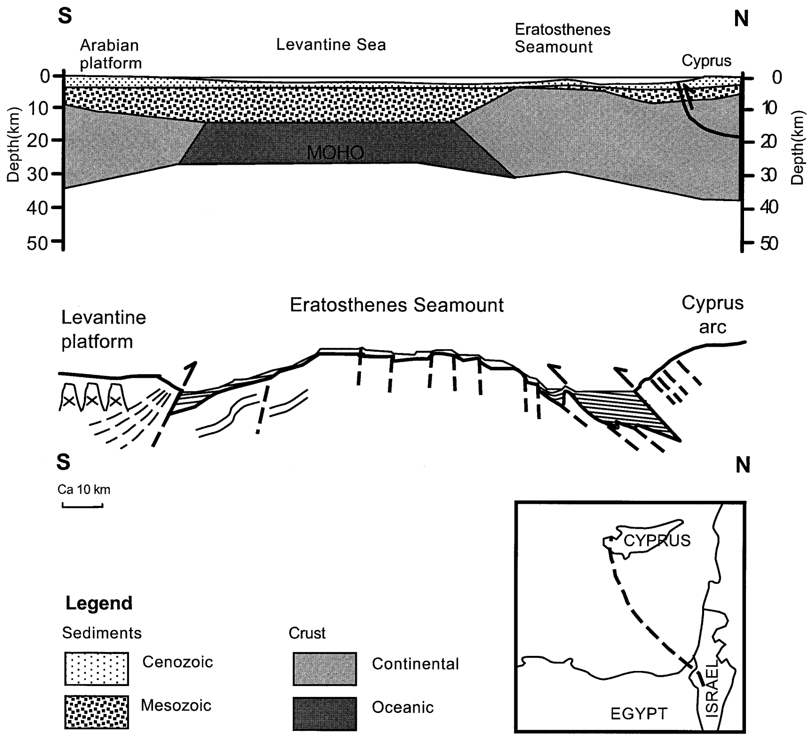

Fig. 6. Crustal structure of the southeastern Mediterranean Basin (after Makris et al., 1983) and shallow structure of the Eratosthenes Seamount and Cyprus arc (after Robertson et al., 1993).

Cyprus arc (Robertson, 1994; Robertson and Shipboard Scientific Party, 1996) and is now in a state of either incipient subduction and partial detachment from the lithosphere or subduction progradation to its distal, non-collisional margin. Post-collisional deformation of this seamount includes the development of small-scale crestal flexural extensional faults and of thrust structures along its northern and southern flanks (Sage and Letouzey, 1990; Robertson et al., 1993). Although the oceanic Herodotus plain, located to the south of the Eratosthenes Seamount, and the area of the Nile cone show no evidence of deformation (Sage and Letouzey, 1990; Harms and Wray, 1990), the southeastern Levant margin experienced a Pliocene phase of, presumably stress-induced, subsidence acceleration (Tibor et al., 1992). The ultimate fate of the Eratosthenes Seamount is uncertain.

An example of a microcontinent involved in the early phases of the Alpine collisional history is the Briançonnais terrane that was limited to the south by the Jurassic–Early Cretaceous Piemont–South Penninic Ocean and to the north by the Early Cretaceous oceanic Valais Trough (Stampfli, 1993; Florineth and Froitzheim, 1994). During the Late Cretaceous–Palaeogene phases of the Alpine orogeny, the South Penninic Ocean closed and the Austroalpine orogenic wedge began to collide with the Briançonnais terrane. During the Paleocene, the Helvetic Shelf, forming the northern margin of the Valais Trough, was affected by broad lithospheric deformation, regional uplift and erosion, and localized, fault-controlled deformation. This was contemporaneous with basin inversion in the Channel and North Sea area (Ziegler, 1990; Ziegler et al., 1995). This suggests that, during the subduction of the South Penninic Ocean and ultimately the collision of the Austroalpine orogenic wedge with the Briançonnais terrane, the latter initially resisted subduction to the degree that horizontal compressional stresses were transmitted through the lithosphere of the Valais Trough into the Helvetic Shelf, causing its deformation. Although incorporation of the Briançonnais into the accretionary wedge did not commence before Middle to Late Eocene times (Stampfli et al., 1998), the Briançonnais terrane and the Valais Ocean were subducted during the Late Eocene and Oligocene and later parts of them brought back to the surface in conjunction with Neogene back-thrusting of the Penninic and Austroalpine nappes and imbrication of the northern and southern foreland basement (Schmid et al., 1996; Stampfli and Marchant, 1997).

In northern Libya, mild Santonian and strong Paleocene inversion of the north Cyrenaican passive margin, resulting in uplift of the Jebel Akhdar anticlinorium (Duronio et al., 1991; El Hawat and Shelmani, 1993; Anketell, 1996), probably can be related to the collision of the Hellenides–Rhodope orogenic wedge with the passive margin of the composite Pelagonian–Apulian–Taurus Platform, which, at that time, was still separated from the African margin by an oceanic basin (Ziegler, 1988; Ricou, 1996). Paleocene compressional intraplate deformations are also evident in northern Egypt, such as inversion of the Cretaceous Abu Gharadig rift in the Western Desert (Hantar, 1990) and of Triassic–Early Jurassic half-grabens in the northern Sinai (Mustafa and Khalil, 1990; Aal and Lelek, 1994). During the Eocene–Oligocene deformation of the Apulian–Taurus Platform and the Early Miocene development of the modern Hellenic arc–trench system, only minor compressional deformations occurred along the Cyrenaican margin. However, during the Early Pliocene, it was affected by a last phase of basin inversion (El Hawat and Shelmani, 1993) that probably coincides with the collision of the Hellenic arc with attenuated African continental crust and early deformation phases of the Mediterranean Ridge thrust belt. The latter now encroaches on the Cyrenaican promontory (Sestini, 1984; Lallemant et al., 1994; Robertson and Grasso, 1994).

5.3. Collision of orogenic wedges with passive margins

During the initial collision of an orogenic wedge with a continent, major compressional stresses can be transmitted into the continent as a consequence of subduction resistance, giving rise to large-scale intraplate deformations (Fig. 6). However, once continental lithosphere starts to be subducted and major nappes are emplaced on the passive margin, a flexural foreland basin develops. At this stage, depending on whether or not the orogenic wedge and the foreland plate are mechanically coupled, intraplate compressional deformation can continue or will cease. We suspect that the crustal configuration of a passive margin, as well as the presence or absence of a thick sedimentary prism, play an important role in terms of its mechanical coupling with the orogenic wedge, particularly during the initial collision stage and also during the subsequent phases of nappe emplacement and foreland basin development (see below).

An example of major intraplate deformation occurring on a passive margin during its initial collision with an orogenic wedge is provided by the Late Cretaceous and Paleocene evolution of the foreland of the eastern Alps and Carpathians. The stratigraphic and structural record of this part of the European Tethys Shelf indicates that compressional stresses began to be exerted on it during the late Turonian, intensified during the Senonian and reached a peak during the Paleocene. Resulting deformation includes inversion of the Polish Trough and upthrusting of the array of basement blocks forming the Bohemian Massif; the latter are detached from the lithosphere at a mid-crustal level (Wagner et al., 1997) and, as such, resemble the Rocky Mountains. Senonian and Paleocene inversion structures occur also along the northern margin of the Bohemian Massif (Altmark–Brandenburg and Subhercynian basins) and the Rhenish Shield (Lower Saxony Basin), along the Fennoscandian border zone and in the Central North Sea. Subsurface data show that intense deformation of the European Shelf extends deeply beneath the nappes of the northern Carpathians and the eastern Alps. Considering that the European foreland crust extends some 150 km beneath the eastern Alps (Tari, 1996), the most distal Paleocene intraplate compressional structures occur about 1500 km to the northwest of the Paleocene East Alpine–Carpathian orogenic front (Ziegler, 1990; Ziegler et al., 1995).

At the transition from the Jurassic to the Cretaceous, the intra-oceanic Meliata arc–trench system collided with the southern margin of the Austroalpine block. Cretaceous deformation of the Austroalpine terrane involved step-wise detachment of its upper crust and sedimentary cover, their stacking into major nappes, and subduction of its lower crust and mantle lithosphere. This was followed by the subduction of the Penninic Ocean (Willingshofer et al., 1997). Late Cretaceous transmission of compressional stresses through the lithosphere of the Penninic Ocean, in which seafloor spreading had ceased in the Oxfordian, may be responsible for the Senonian deformation of the European passive margin. Although flysch sedimentation continued along the European margin into the Middle Eocene, the intensity of its Paleocene deformation suggests strong mechanical coupling between the Austroalpine orogenic wedge and its foreland during their initial collision, both at crustal and mantle lithospheric levels. During the Eocene, the flexural foreland basin of the Alps and northern Carpathians developed, indicating that, by this time, the distal parts of the European Shelf were overridden by the advancing orogenic wedge and that subduction of European lithosphere had commenced. During the Oligocene to Middle Miocene, the nappe systems of the eastern Alps and northern Carpathians were emplaced on the European foreland crust (Roeder and Bachmann, 1996; Zimmer and Wessely, 1996; Bessereau et al., 1996). The lack of Eocene and younger compressional foreland deformation suggests, that during nappe emplacement the orogenic wedge and the foreland were mechanically decoupled. However, the present-day stress regime of Europe and Plio–Pleistocene upwarping of the Bohemian Massif, possibly involving lithospheric buckling, reflect renewed strong coupling between the Alpine–Carpathian orogenic wedge and its foreland (Ziegler, 1990, 1994).

A further example of disruption of a passive margin, prior to its incorporation into a flexural foreland basin, is the northern shelf of the Rhenohercynian back-arc basin out of which the external parts of the Variscan fold belt developed. This oceanic basin closed during the Visean due to its southward subduction beneath the advancing Variscan orogenic wedge. During the Early Carboniferous, the Rhenohercynian Shelf was occupied by a broad carbonate platform. At the transition from the Visean to the Namurian, this platform was tectonically destabilized, as evident by the development of a regional unconformity associated with crustal warping and block faulting. During the Namurian, the Rhenohercynian Shelf subsided rapidly in conjunction with its incorporation into the flexural Variscan foreland basin. Compressional deformation of the Variscan foreland resumed only during the Westphalian late phases of the Variscan orogeny, at the end of which nappe systems had overridden the foreland crust by as much as 200 km (Ziegler, 1990; Oncken et al., 1998).

Also this example illustrates that the Variscan orogenic wedge and the Rhenohercynian passive margin were mechanically coupled during their initial collisional, but were decoupled during the main phase of nappe emplacement and foreland basin development; their renewed mechanical coupling is indicated during the late orogenic phases, probably in response to over-thickening of the orogenic wedge.

5.4. Post-collisional over-thickening of orogenic wedges

During the post-collisional evolution of an orogen, over-thickening of the orogenic wedge can be accompanied by the development of major intraplate compressional structures. This applies to the hinterland of Andean-type orogens (Fig. 1B) as well as to the fore- and hinterland of Himalayan-type orogens (Fig. 1C and Fig. 7C).

During the post-collisional evolution of an orogen, development of compressional intraplate deformations depends largely on the interplay between the integrated strength of the orogenic wedge and orogeny-related stress sources (convergence rates, slab pull, topography). At low convergence rates, the over-thickened lithosphere of an orogen has a low integrated strength due to the great thickness of the crust and its heat production (Cloetingh and Banda, 1992; Genser et al., 1996; Okaya et al., 1996). Moreover, density increase of underthrust cratonic crustal material due to eclogitization at depths of 45–50 km exerts a negative load on the orogen and impedes its uplift (Henry et al., 1997). However, at high convergence rates, underthrusting of cool foreland lithosphere can cause cooling and considerable strengthening of the orogenic wedge (Bousquet et al., 1997). Under such conditions, eclogitization of the underthrust low-density foreland crust occurs only at depths of 70–75 km, accounting for an orogen to attain elevations of some 5 km, as seen in the Himalayas (Henry et al., 1997). Body forces inherent to the thickened crust and topographic elevation of an orogen are counteracted by slab pull forces; only

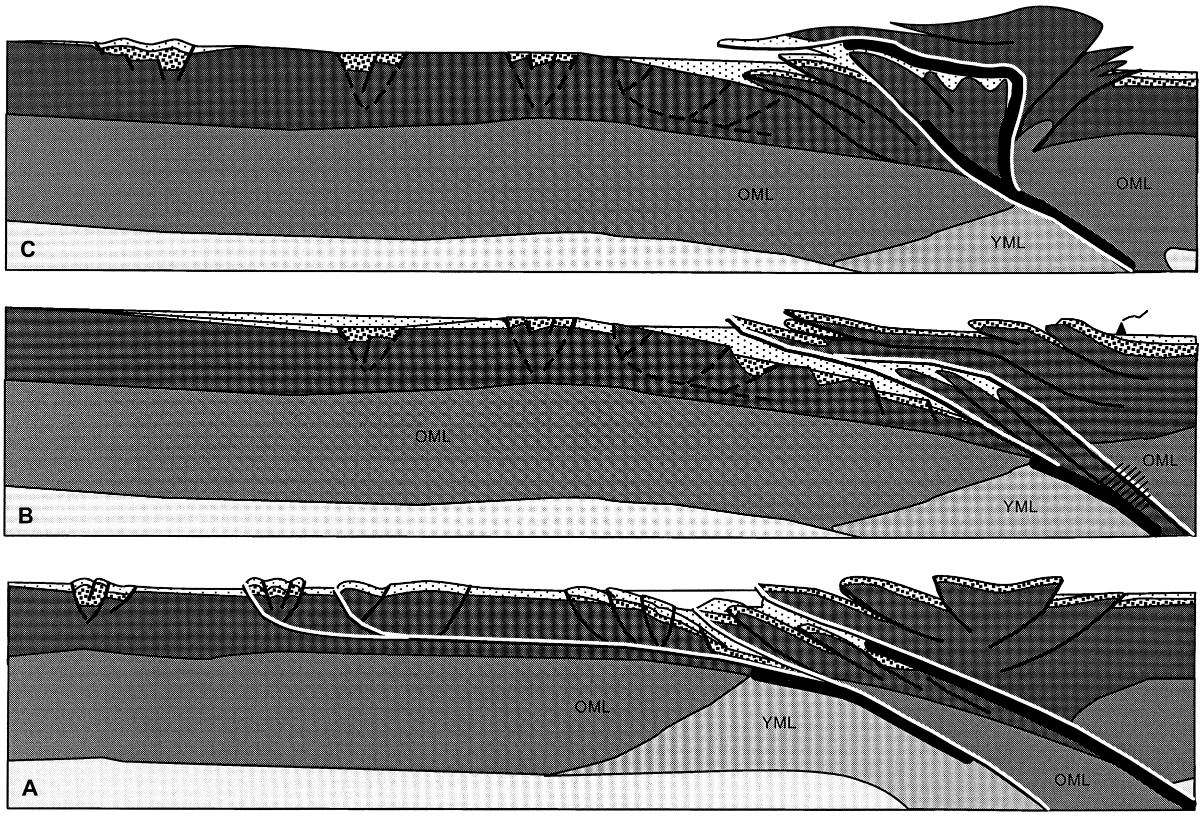

Fig. 7. Conceptual model illustrating (A) initial collision of an orogenic wedge with a passive margin, inducing foreland intraplate compressional deformations, followed by (B) subduction of foreland lithosphere, accompanied by emplacement of nappes on foreland, foredeep basin development and lithospheric buckling, or (C) subduction of foreland lithosphere, accompanied by imbrication of the foreland crust during nappe emplacement, basin inversion in the distal foreland, back-thrusting of the orogenic wedge above a mantle back-stop and imbrication of the hinterland crust (not to scale, legend same as Fig. 1).

after slab detachment and ensuing further uplift of the orogen, are compressional body force stresses exerted onto its fore- and hinterlands (Fleitout and Froidevaux, 1982; Bott, 1990, 1993). Such body forces can be maintained and possibly even enhanced during the post-orogenic evolution of an orogen due to heating of its crustal root zone, ensuing conversion of eclogites to lower-density granulites above depths of 65–75 km, and resulting isostatic uplift and collapse of the orogenic wedge (Le Pichon et al., 1997; Bousquet et al., 1997). Furthermore, in continent–continent collisional belts, progressive involvement of little attenuated crust, its partial detachment from the lithosphere and incorporation into the orogenic wedge, results in progressive thickening of the latter. In this process, development of a mantle back-stop at the leading edge of the overriding plate, beneath which mantle and crustal material of the lower plate is subducted, and above which the orogenic wedge is back-folded, plays an important role (Beaumont et al., 1994; Roure et al., 1996). Development of a mantle back-stop, and resulting strain partitioning in oblique collision zones (Ziegler and Roure, 1996), provides for strong coupling of the colliding continents at crustal and subcrustal levels. Consequently, the potential of a post-collisionally overthickened orogenic wedge to exert major compressional stresses on its fore- and hinterland, inducing far-field intraplate deformations, is largely a function of convergence rates, its topographic evolution, slab detachment, and the development of a mantle back-stop. Significantly, we know of no examples of post-orogenic foreland compressional deformations that could be exclusively related to body forces inherent to orogenic wedges.

The concepts developed above can be applied to the Oligocene and later evolution of the western and central Alps that are characterized by a system of external crystalline massifs, involving European and Apulian basement. In the Central Alps, thrust propagation into the European fore-arc crust commenced during the Late Oligocene, whereas thrust propagation into the Apulian back-arc crust began during the Early Miocene. It is noteworthy that imbrication of the European foreland crust was preceded by the Early Oligocene detachment of the Cretaceous–Palaeogene subduction slab and was accompanied by high convergence rates of some 5 mm/year and major back-folding of the orogenic wedge above the Adriatic mantle back-stop (Schmid et al., 1996; Ziegler et al., 1996). Back-thrusting and development of a mantle back-stop is also evident in the western Alps (Ziegler and Roure, 1996). In the distal Alpine foreland, Oligocene and younger inversion of tensional Mesozoic basins in the Western Approaches–Celtic Sea and Channel area, located up to 1200 km to the northwest of the contemporaneous Alpine deformation front, can be related to horizontal compressional stresses projecting from the Alpine collision zone into its European foreland (Ziegler, 1987, 1990).

Similarly, the Permo–Carboniferous deformation of the Sahara Platform, located in a foreland position with respect to the Variscan and Appalachian–Mauretanides orogens, reflects increasing mechanical coupling between these orogenic wedges and their African foreland. This deformation includes the inversion of the early Palaeozoic Saoura–Ougarta rift, reactivation of Pan-African fracture systems associated with the El Biod axis and upwarping of the broad Reguibat arch. Nearly contemporaneous development of the Ancestral Rocky Mountains, located in a hinterland position with respect to the Appalachian–Ouachita–Marathon orogen, also reflects increasing mechanical coupling between this orogenic wedge and the American craton, presumably due to over-thickening of the orogen at high strain rates (Ziegler, 1988, 1989; Ziegler et al., 1995).

Imbrication of the Precambrian autochthonous crystalline basement of the Scandinavian Caledonides (Gee et al., 1985) reflects also increasing collisional coupling between the orogenic wedge and its foreland during the late phases of the Caledonian orogeny. Moreover, development of the regional pre-Middle Devonian unconformity on the East European craton (Milanovsky, 1987) reflects the combined effects of a eustatic low-stand in sea-level (Johnson et al., 1985; Alekseev et al., 1996) and stress-induced broad lithospheric deflections that must be related to collisional coupling of the Caledonides and their foreland (Nikishin et al., 1996). Similarly, development of the regional base-Devonian unconformity on the Laurentian craton is attributed to lithospheric buckling in response to collisional coupling of the Pearya–Caledonides–Appalachian system of orogens with their foreland (Ziegler, 1989; Ziegler et al., 1995).

6. Mechanisms controlling mechanical coupling of orogenic wedges and forelands

The timing and intensity of intraplate compressional deformations indicate that mechanical coupling between an evolving orogen and its foreland can vary considerably. For instance, for the East Alpine–North Carpathian orogen, intense Senonian/Paleocene foreland deformations indicate strong mechanical coupling between the foreland and the orogenic wedge during their initial collision stage; however, during the Eocene–Early Miocene emplacement of nappes on the foreland, mechanical coupling between the orogen and its foreland was at a low level. In contrast, post-collisional increasing coupling between the west and central Alpine orogen and its foreland is reflected by Oligocene and younger progressive imbrication of the external massifs and by basin inversions in the distal foreland.

Processes governing mechanical coupling/decoupling of an orogenic wedge with its foreland are still poorly understood and require further research. However, we suspect that, during the initial collision of an orogenic wedge with a passive margin, the degree of their mechanical coupling is probably controlled, apart from convergence rates and directions (orthogonal, oblique), by the crustal configuration of the continental margin (buoyancy, cf. Cloos, 1993; rheological structure, cf. Cloetingh and Banda, 1992)

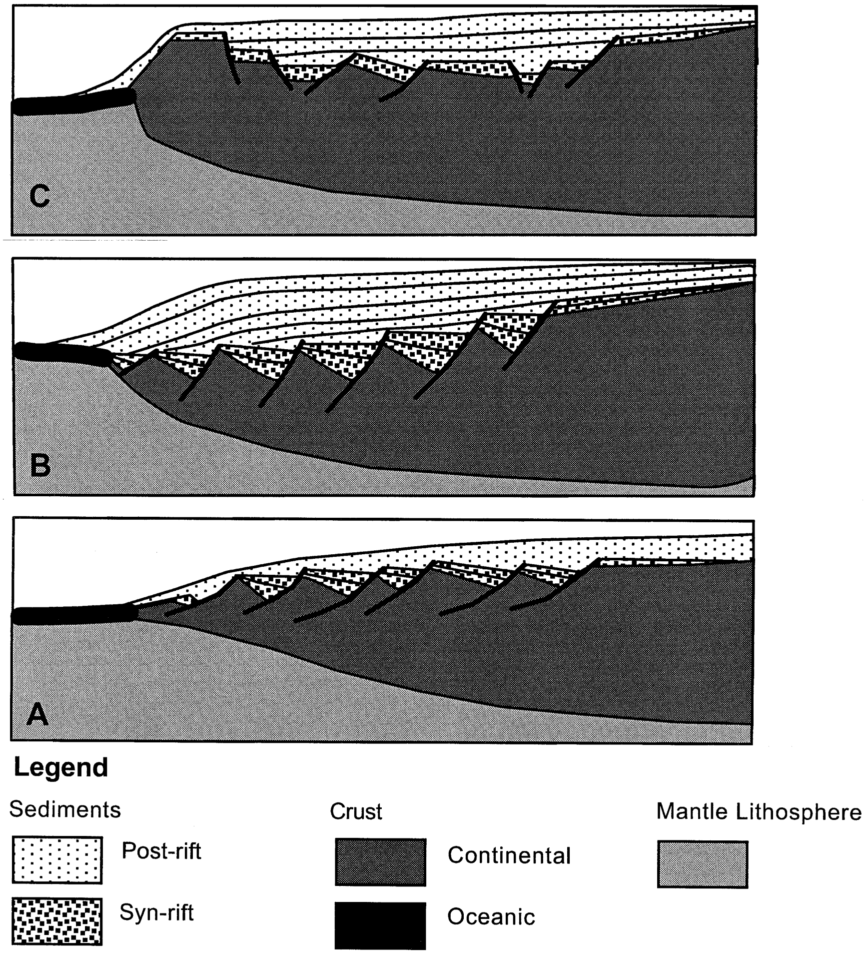

Fig. 8. Variations in configuration of passive margins having a bearing on their subduction resistance (not to scale). (A) Sediment-starved margin (e.g. Goban Spur). (B) Sediment over-filled margin (e.g. Campos Basin, Brazil; Gulf of Mexico). (C) Margin characterized by an outer basement high (e.g. East Newfoundland Basin, Mid-Norway Basin).

and whether it is characterized by a thick sedimentary prism or is sediment starved (Fig. 8).

Continental margins characterized by an abrupt continent–ocean transition, typical for sheared margins (e.g. Ivory Coast–Ghana: Basile et al., 1993) and upper plate margins (e.g. Grand Banks, Newfoundland: Keen et al., 1987; western margin of Rockall–Hatton Bank: White, 1988; Bassi et al., 1993), as well as lower plate margins characterized by a major marginal high (e.g. Vøring Plateau, Mid-Norway Shelf: Planke et al., 1991), may present major subduction obstacles, particularly if sediment starved. In contrast, continental margins characterized by a drawn out, gradual thinning of the continental crust, typical for lower plate margins, whether sediment starved (e.g. Mazagan margin, Morocco: Winterer and Hinz, 1984; Goban Spur, Armorican margin: White, 1988) or sediment over-filled (e.g. Campos and Santos basins, Brazil: Chang et al., 1992), may more readily be subducted. In the presence of a thick passive margin sedimentary prism, development of thin-skinned thrust sheets and sediment subduction will allow for mechanical decoupling of the subducting foreland plate and the orogenic wedge due to the build-up of high pore fluid pressures in the subducted sediments (von Huene and Lee, 1982; von Huene and Scholl, 1991, 1993). In the North Japan arc–trench system, this mechanism seems to play an important role in the distribution of earthquake hypocentres, indicating that the subducting and overriding plates are decoupled to depths of 23 km whereas they are strongly coupled at depths between 23 and 90 km (Huang et al., 1997).

Mechanical coupling at crustal levels is required for compressional crustal detachment in a foreland, as seen in the Rocky Mountains and the Bohemian Massif. Coupling at the level of the mantle-lithosphere is required for whole lithospheric buckling, as evident for instance during the end-Silurian deformation of the North American craton (Ziegler et al., 1995; Cloetingh and Burov, 1996). Lithospheric over-thickening of an orogenic wedge at high strain rates, as well as its topographic elevation, play an important role in its mechanical coupling with the foreland, both at crustal and mantle–lithospheric levels. In the case of continent–continent collisional belts, slab-detachment and development of a mantle back-stop (Roure et al., 1996) significantly contribute towards mechanical coupling of an orogen and its fore- and hinterland. Oblique collisions can account for lateral variations in the intensity of mechanical coupling between an orogenic wedge and its foreland (Robertson, 1994), as reflected by the Paleocene deformation of the European passive margin in the Alpine–Carpathian domain.

7. Conclusions

Collision-related compressional intraplate structures can be associated with Andean- and Himalayan-type orogens and can occur in back-arc as well as in fore-arc domains. Such deformation, ranging from crustal- to lithosphere-scales, can involve reactivation of pre-existing crustal discontinuities, resulting in the inversion of tensional hanging-wall basins and upthrusting of basement blocks, as well as whole lithospheric buckling, including uplift of broad arches and accelerated subsidence of post-rift basins.

Back-arc compression associated with island arcs and Andean-type orogens occurs during periods of increased convergence rates between the subducting and overriding plates and as a consequence of over-thickening of the orogenic wedge. In fore-arc and foreland domains, horizontal intraplate compressional stresses can develop:

(1) prior to the initiation of a new intra-oceanic subduction zone or one associated with a continental margin, possibly employing the boundary between old and younger mantle lithosphere;

(2) during the collision of an oceanic plateau or a microcontinent with an arc–trench system as a consequence of subduction resistance;

(3) during the initial collision of an orogenic wedge with a passive margin;

(4) during post-collisional over-thickening of a Himalaya-type orogenic wedge at high strain rates.

Continent–continent collisional orogens can be associated with foreland and hinterland compressional intraplate deformation.

The build-up of collision-related compressional intraplate stresses is indicative for mechanical coupling between an orogenic wedge and its forelands. Crustal-scale deformation reflects mechanical coupling between a foreland and an orogenic wedge at crustal levels. Pre-existing crustal discontinuities significantly weaken the strength of the crust and play a crucial role in localizing crustal-scale compressional intraplate structures. During their development, the upper crust can be detached from the lithosphere at the level of the rheologically weak lower crust. The depth of potential crustal detachment levels depends on the thickness of the sedimentary overburden and the thickness, composition, thermal state and water saturation of the crust (Lankreijer et al., 1997). Lithosphere-scale compressional deformation indicates mechanical coupling between a foreland and an orogenic wedge at the level of the mantle-lithosphere. Spatial strength variations within the mantle-lithosphere probably play an important role in localizing lithosphere-scale deformations. Such strength variations can be related to lateral changes in crustal thickness, thermal anomalies and possibly also to ancient subduction zones, presenting discontinuities within the continental mantle-lithosphere. Moreover, boundaries between old and newly accreted continental mantle-lithosphere may provide additional discontinuities.

The intensity of collisional coupling between an orogen and its fore- and/or hinterland is temporally and spatially variable. The build-up of high fluid pressures in subducted sediments may account for mechanical decoupling of an orogenic wedge and its fore- and/or hinterland. Nevertheless, the stratigraphic record of collision-related intraplate compressional deformations can contribute to dating of orogenic activity affecting the respective plate margin. However, in view of the different scenarios developed above, their interpretation in terms of collisional events requires supporting stratigraphic and structural evidence.

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