Rheological heterogeneity, mechanical anisotropy and deformation of the continental lithosphere

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Abstract

This paper aims to present an overview on the influence of rheological heterogeneity and mechanical anisotropy on the deformation of continents. After briefly recapping the concept of rheological stratification of the lithosphere, we discuss two specific issues: (1) as supported by a growing body of geophysical and geological observations, crust/mantle mechanical coupling is usually efficient, especially beneath major transcurrent faults which probably crosscut the lithosphere and root within the sublithospheric mantle; and (2) in most geodynamic environments, mechanical properties of the mantle govern the tectonic behaviour of the lithosphere. Lateral rheological heterogeneity of the continental lithosphere may result from various sources, with variations in geothermal gradient being the principal one. The oldest domains of continents, the cratonic nuclei, are characterized by a relatively cold, thick, and consequently stiff lithosphere. On the other hand, rifting may also modify the thermal structure of the lithosphere. Depending on the relative stretching of the crust and upper mantle, a stiff or a weak heterogeneity may develop. Observations from rift domains suggest that rifting usually results in a larger thinning of the lithospheric mantle than of the crust, and therefore tends to generate a weak heterogeneity. Numerical models show that during continental collision, the presence of both stiff and weak rheological heterogeneities significantly influences the large-scale deformation of the continental lithosphere. They especially favour the development of lithospheric-scale strike-slip faults, which allow strain to be transferred between the heterogeneities. An heterogeneous strain partition occurs: cratons largely escape deformation, and strain tends to localize within or at the boundary of the rift basins provided compressional deformation starts before the thermal heterogeneity induced by rifting are compensated. Seismic and electrical conductivity anisotropies consistently point towards the existence of a coherent fabric in the lithospheric mantle beneath continental domains. Analysis of naturally deformed peridotites, experimental deformations and numerical simulations suggest that this fabric is developed during orogenic events and subsequently frozen in the lithospheric mantle. Because the mechanical properties of single-crystal olivine are anisotropic, i.e. dependent on the orientation of the applied forces relative to the dominant slip systems, a pervasive fabric frozen in the mantle may induce a significant mechanical anisotropy of the whole lithospheric mantle. It is suggested that this mechanical anisotropy is the source of the so-called tectonic inheritance, i.e. the systematic reactivation of ancient tectonic directions; it may especially explain preferential rift propagation and continental break-up along pre-existing orogenic belts. Thus, the deformation of continents during orogenic events results from a trade-off between tectonic forces applied at plate boundaries, plate geometry, and the intrinsic properties (rheological heterogeneity and mechanical anisotropy) of the continental plates.

Keywords: rheology; heterogeneity; anisotropy; deformation; lithosphere; continents

1. Introduction

Tectonic models frequently assume that the rheological structure of the continental lithosphere is vertically layered, laterally homogeneous, and isotropic. As a consequence, observed intracontinental deformation is assumed to depend almost exclusively on forces applied at plate boundaries and on plates geometry. Continents, however, are composed of various lithospheric domains with different ages and tectonic histories. They have been agglomerated through orogenic processes that generated a pervasive tectonic fabric within the colliding lithospheres. Even in the absence of orogenesis, continents are subjected to intraplate processes (e.g., rifting, plume-related volcanism) that may produce local thermal/rheological perturbations. Therefore, a more realistic model for the continental lithosphere is certainly mechanically heterogeneous and anisotropic. This raises the question of the influence of pre-existing rheological heterogeneities and mechanical anisotropy on the deformation of continents. Simple thermo-mechanical models give a hint that a pre-existing rheological heterogeneity may result in strain localization and lateral variation in strain regime at the scale of the heterogeneities. Similarly, results from experimental deformation and numerical modelling suggest that mantle rocks that display an olivine lattice-preferred orientation are mechanically anisotropic. If this anisotropy also exists at a large scale, it may influence the mechanical behaviour of the lithospheric mantle.

The deformation of continents is frequently characterized by relatively short-scale spatial variations in strain intensity, deformation regime, metamorphic grade and topography (vertical strain). This complexity is usually accounted for by peculiar plate boundary configurations (e.g., oblique convergence or indentation), or by changes in stress regime through time. Using natural examples and numerical models, it will be shown that complex strain fields may also result from a simple tectonic evolution affecting a heterogeneous and/or anisotropic continental plate.

2. Rheology of the lithosphere

The concepts of strength profiles and rheological stratification of the lithosphere have been introduced in geodynamics since the end of the seventies (Goetze and Evans, 1979; Brace and Kohlstedt, 1980; Kirby, 1983; Ranalli, 1986). To derive a tractable formulation of the rheology of the lithosphere, its composition is reduced to a limited number of lithological layers, each one having uniform composition and rheological parameters over its entire thickness (Fig. 1). The strength of rocks at a given depth D depends on temperature (T_D), pressure (P_D), deformation mechanism dominant at T_D and P_D, and strain rate. In a simplified approach, two main mechanisms are competing: brittle failure and dislocation creep; and it is assumed that the active mechanism is the one that requires the minimum work (Fig. 1). Vertical integration of the strength computed at different depths allows an evaluation of the total strength of a lithospheric column (England, 1983). The oversimplification of this approach was already discussed by several authors (e.g., see review in Ranalli, 1986; Paterson, 1987), and will not be further addressed in this paper. We will rather focus on two major issues concerning the deformation of the lithosphere: the control of upper mantle mechanical behaviour on the deformation of the lithosphere, and the tectonic coupling or decoupling at the crust–mantle interface.

(1) In most tectonic environments, the strength of the subcrustal mantle is the largest, with the exception of domains characterized by a very high thermal flow, where the brittle crust (whose strength is insensitive to temperature) is stiffer (Fig. 1). There is a consensus to consider that the mechanical properties of the stiffer layer determine the behaviour of the whole lithosphere (e.g., Vilotte et al., 1982; England, 1983). This may have an important consequence: upper mantle flow very likely guides the deformation of the lithosphere and the crust passively accommodates this deformation through its own behaviour, brittle or ductile depending on depth. In other words, a comparison of the relative strength of the crust and lithospheric mantle estimated from rheological profiles suggests that the fundamental mode of deformation of the lithosphere during tectonic events is determined by the upper mantle rather than by crustal tectonics. Of course, this conclusion requires a degree of coupling high enough to allow efficient stress transfer between the mantle and the lower crust.

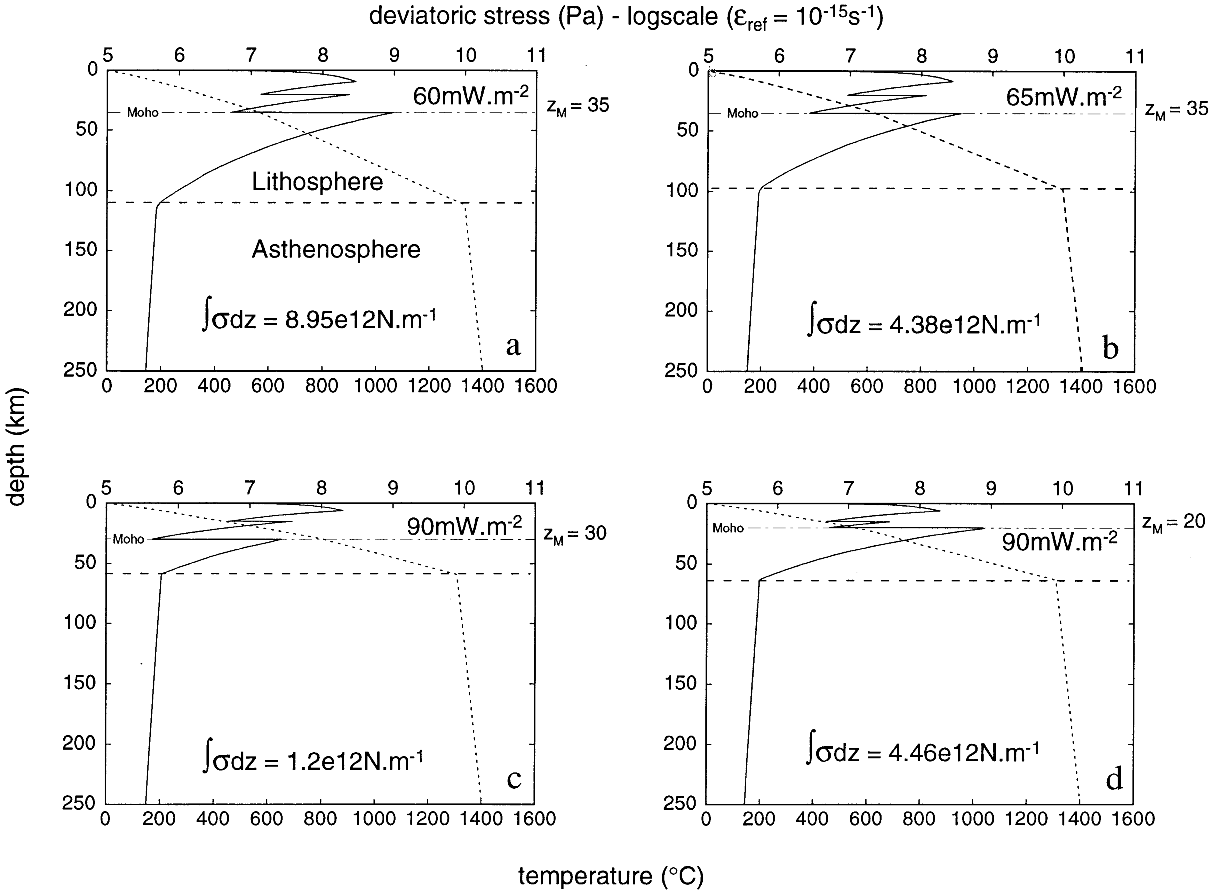


Fig. 1. Rheological profiles calculated considering that the lithosphere is composed of a quartz-dominated 'upper/middle crust layer', overlaying a plagioclase-dominated 'lower-crust' layer and an olivine-dominated 'upper-mantle' layer. Constitutive equations used to describe the deformation are the Byerlee law (Byerlee, 1978) for the brittle crust and a power law (e.g., Weertman, 1978) for ductile deformation. Input rheological parameters are from Paterson and Luan (1990) for quartz, Wilks and Carter (1990) for granulite, and Chopra and Paterson (1981) for dunite. Strength profiles are for a 'normal lithosphere' with two slightly different geotherms (a,b) and an extended lithosphere with a high surface heat flow and two different crustal thickness (c,d). It should be noticed that an increase of 5 mW m^{-2} in surface heat flow can halve the integrated lithospheric strength, and a reduction of the crustal thickness from 30 to 20 km would increase almost four times the lithosphere strength, assuming the same surface heat flow (see discussion of this assumption in the text).

(2) The description of the lithosphere as a succession of discrete layers that display contrasting rheological properties generates abrupt variations in strength at the interface between these layers (Fig. 1). These interfaces were interpreted as possible mechanical decoupling levels. Many tectonic models have included levels of decoupling, especially the lowermost crust, and it was suggested that faults frequently root into these soft levels. According to these models, coupling between the mantle and the crust should be inefficient. However, recent geophysical and geological observations tend to indicate that decoupling in the lithosphere is not as frequent as previously thought. As summarized below, a large body of data points toward continental-scale fault arrays rooted into the upper mantle rather than in the crust.

(a) Geological observations of continental-scale arrays of transcurrent faults support that these shear zones vertically crosscut the middle and lower crust where decoupling levels are expected. The Borborema Province of Brazil, for instance, provides an opportunity to observe transcurrent shear zones several hundreds of kilometres long and ten to thirty kilometres wide formed in a partially molten middle crust (Vauchez and Egydio-Silva, 1992; Vauchez et al., 1995). Although partially melted silica-rich rocks should display a low viscosity and therefore represent good candidates for a decoupling level, it is obvious from field observations that the steeply dipping mylonitic foliation is not connected with a low-angle decoupling zone. On the contrary, synmelting deformation clearly tends to localize in vertical shear zones at various scales (Vauchez et al., 1995). In addition, the chemical composition of magmas emplaced within active shear zones indicates a mantle-origin, and therefore a connection with the underlying mantle (Neves and Vauchez, 1995; Neves et al., 1996; Vauchez et al., 1997b).

(b) The fault array of Madagascar allows an observation of steeply dipping transcurrent shear zones several tens of kilometres wide and hundreds of kilometres long in the lower crust (Pili et al., 1997). In the deepest parts of the studied shear zones, mantle peridotites displaying a tectonic fabric conformable to the fabric of the crustal granulitic mylonites are exposed. No evidence of a horizontal decoupling layer has been observed. On the contrary, stable isotope studies (δ^{13}C, δ^{18}O) suggest that large volumes of CO_2-rich, mantle-derived fluids have percolated preferentially into the major shear zones, pointing to an efficient crust–mantle connection (Pili et al., 1997).

(c) Detailed studies of the gravity field over continental-scale shear zones of Madagascar and Kenya (Cardon et al., 1997; Pili et al., 1997) have shown that positive anomalies are associated to the largest shear-zones of the network, those for which a connection with the mantle was suggested from stable isotope studies. The anomalies have been satisfactorily modelled assuming a Moho uplift of ca. 10 km beneath the shear zones. This localized mantle uplift is interpreted as due to a local thinning of the crust through intense shearing, and therefore strongly support rooting of the shear zones into the upper mantle.

(d) Shear wave splitting measurements (see review in Silver, 1996) usually support that the crust and the upper mantle display similar large-scale tectonic fabrics. Shear wave splitting occurs when a radially polarized shear wave (e.g., SKS, SKKS, PKS) propagates across an anisotropic medium (e.g., the lithospheric mantle). The incident shear wave splits in two quasi-orthogonally polarized waves, and the orientation of the polarization planes is correlated to the fabric of the anisotropic layer. Especially conclusive are measurements above several major transcurrent faults which show that the orientation of the fast wave polarization plane is rotated approaching the fault, suggesting that the tectonic fabric (flow plane and direction) of the upper mantle is curved into parallelism to the shear zone. This is the case for the Great Glen fault in the Scottish Caledonides (Helffrich, 1995), the Kunlun (McNamara et al., 1994), and the Altyn Tagh (Herquel, 1997) active faults in Tibet, the North Pyrenean fault in the French Pyrenees (Barruol and Souriau, 1995; Vauchez and Barruol, 1996), the Martic line in the eastern US (Barruol et al., 1997), and the Ribeira fault array in southeastern Brazil (James and Assumpção, 1996). A similar observation was made in western North America for the San Andreas active fault (Savage and Silver, 1993). Moreover, in this area, Pn anisotropy measurements show that the fast propagation direction in the uppermost mantle beneath the fault zone is parallel to the fault direction (Hearn, 1996). This is in agreement with a preferred orientation of the *a*-axis of olivine parallel to the fault direction (i.e., the shear direction), a situation expected for strike-slip faults in the mantle. Shear wave splitting measurements in Corsica (Margheriti et al., 1996) also hint to a coherent deformation of the crust and the mantle, associated to obduction during Alpine collision; in this case, the fast shear wave is polarized in a direction normal to the trend of the belt, i.e., parallel to the direction of thrusting. The measured delay time between the fast and slow split shear waves for all these examples implies that the lithosphere is affected by the fault-related fabric over its entire thickness.

(e) Seismic profiling across major faults also provides a growing body of evidence on crust–mantle coupling. McGeary (1989) showed that the Great Glen fault is associated to a jump of the Moho which indicates that the fault penetrates the mantle. In the Alps, seismic profiles (e.g., Nicolas et al., 1990), together with gravity studies (Bayer et al., 1989), support the interpretation that the main thrusts of the belt crosscut the Moho and affect the upper mantle; an interpretation which agrees quite well with conclusions from surface geology presented by Huber and Marquer (1998). From near-vertical and wide-angle reflection surveys, Diaconescu et al. (1997) have shown that Moho offsets or even mantle reflectors are associated with many major transcurrent, extensional and thrust faults (e.g., Northern Appalachians, edge of the Colorado plateau, Urals, central Australia, Baltic shield, Superior Province of Canada). Diaconescu et al. (1997) estimate that these observations represent strong arguments against geodynamics models that favour complete decoupling at the Moho.

(f) Magnetotelluric soundings in the Pyrenees (Pous et al., 1995) have imaged a steeply dipping boundary that penetrates into the upper mantle beneath the North Pyrenean fault, suggesting that the fault crosscuts the Moho. Electrical anisotropy measurements in the Canadian shield (Sénéchal et al., 1996) and the eastern US (Wannamaker et al., 1996) show a very good agreement between the directions of conductivity anisotropy in the upper mantle and the crustal tectonic fabric, suggesting that the mantle and the crustal fabrics are similar.

There is an inconsistency between these observations that strongly support a coherent deformation of the upper mantle and the crust, and rheological models in which the lower crust cannot sustain significant stress and would act as a decoupling level. This is certainly due to the scarcity of reliable experimental rheological data for the lower crust, since many data have been obtained from Ca-poor plagioclase or in apparatus inadequate for rheological measurements at high temperature and pressure. Recent experiments point toward an activation energy for power-law creep of intermediate to calcic plagioclase-rich rocks much higher than previously thought (Seront, 1993; Mackwell et al., 1996). These data would imply that the lower crust is as stiff as the upper mantle (Seront, 1993), and therefore that a good mechanical coupling may exist at the crust–mantle transition.

Down to which depth do major faults penetrate? Whether major faults root in the sublithospheric mantle or tend to vanish in the lower lithospheric mantle due to a more homogeneous strain partition within low viscosity mantle rocks remains speculative. It should however be remembered that shear wave splitting data are, in several places, suggestive of a tectonic fabric penetrative over the entire lithosphere thickness, and therefore of a connection of lithospheric faults with the asthenosphere, which represents an efficient decoupling level in which displacement of the lithosphere relative to the lower mantle is accommodated (e.g., Tommasi et al., 1996). The time lag between the arrivals of the fast and slow split shear waves measured in active areas may even support a coherent deformation of the asthenosphere and the lithosphere. For the Kunlun fault in Tibet, for instance, McNamara et al. (1994) measured $\partial t = 2$ s and a reorientation of the fast shear wave polarization from oblique to parallel to the fault. Such delay time suggests an anisotropic layer having a fabric coherent with the crustal fault kinematics and a thickness of $\approx$200 km, much larger than the lithosphere thickness in the area (e.g., McNamara et al., 1994). We are therefore inclined to consider that major faults may crosscut the lithosphere and root into the sublithospheric mantle (Fig. 2), but this conclusion needs to be further confirmed.

In some circumstances, however, crust–mantle mechanical decoupling is likely. This is, for instance, the case in the Himalayas where interpretation of recent seismic reflection profiles (Nelson et al., 1996) supports that the main thrust faults (MCT, MBT) do not crosscut the crust–mantle interface, a model in agreement with Mattauer (1985). Indeed, Nelson et al. (1996) suggest that the middle crust is partially molten in this area and behaves as a fluid on the time-scale of deformation. Hence, observations in the Alps and the Himalayas lead to contrasting conclusions: major thrust faults may or may not affect the Moho discontinuity. Pervasive melting of the crust in the Himalayas may provide an explanation for crust–mantle decoupling in this area, since no such evidence has been reported from the Alps. It is however interesting to compare the case of the Himalayas with the Borborema shear zone system of northeastern Brazil which is not rooted in the partially molten middle crust but seems to continue down to the mantle. This may be due to a difference in deformation regime: thrust faults in the Himalayas and strike-slip faults in the Borborema Province. Besides pervasive melting in the middle-crust, decoupling was favoured in the Himalayas by continental subduction which enhanced layer-parallel forces, whereas in the Borborema Province, steeply dipping

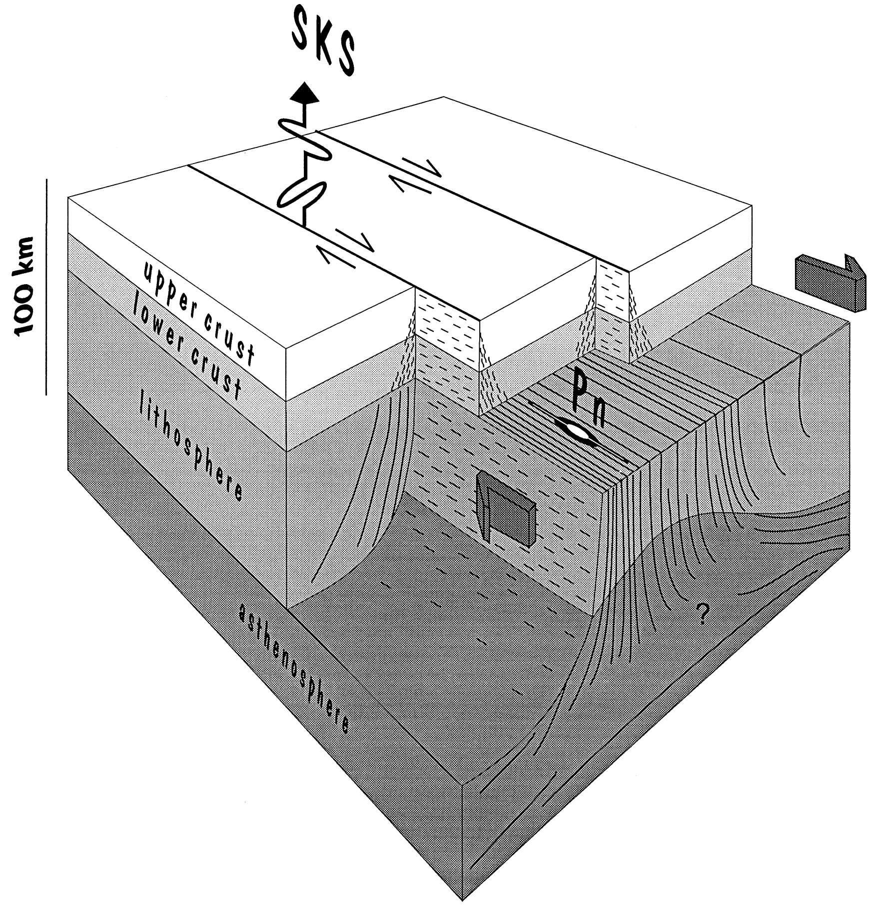


Fig. 2. Cartoon illustrating the concept of a shear zone rooted into the asthenosphere. The orientation of the fast split shear wave polarization and the fast direction for Pn are also shown. The questions of the rooting of the fault into the asthenosphere and the interaction between the fault kinematics and the deformation of the asthenosphere due to plate motion remains open. Modified from Teyssier and Tikoff (1998).

shear zones developed into an already amalgamated continent in the far field of a continental collision (Vauchez et al., 1995).

3. Lateral rheological heterogeneity and deformation of continents

Lateral rheological or strength heterogeneity may arise at various scales and from various sources. The mechanical behaviour of the continental lithosphere, however, is probably only influenced by heterogeneities which significantly modify its total (integrated) strength over areas large enough to modify the deformation field at the continental scale (Tommasi, 1995). Such heterogeneities may be generated by lateral variations either in crustal thickness or in geothermal gradient.

Because the stiffness of crustal rocks is notably lower than the stiffness of mantle rocks (perhaps with the exception of some granulites), the relative proportion of crustal and mantle material in a lithospheric section influences the total strength of the lithosphere. Assuming similar geotherms and therefore lithosphere thickness, a domain with a thick crust has a lower total strength than a domain with a normal or thin crust (e.g., Ranalli, 1986). Crustal thickness is frequently variable over a continent. Active margins above subduction zones may display an abnormally thick crust. In the Andes, for instance, a

75–80 km crustal thickness was inferred from seismic studies (Zandt et al., 1994). Thick crust also characterizes the internal domains of orogenic areas before compensation occurs. This led several authors (e.g., Ranalli, 1986; Braun and Beaumont, 1987; Dunbar and Sawyer, 1988, 1989) to suggest that subsequent deformation may preferentially localize in these domains. On the other hand, extensional deformation leading to basin development results in a significantly thinner crust and may represent stiffer domains in a continent. However, the rheological effect of crustal thickness variations cannot be considered alone since it may be balanced by the effect of others parameters, especially lithospheric mantle thickness variation, as it will be further illustrated in more detail.

Lateral variations in geotherm have a major effect on the rheology of the lithosphere due to the exponential dependence of the dominant deformation mechanisms (dislocation glide and climb) on temperature. The geotherm controls the depth of the brittle–ductile transition, and the viscosity of both crustal and mantle rocks. Hence, if the geothermal gradient is increased or decreased, weak or stiff rheological heterogeneities may be generated. Lateral variations of the continental geotherm result from either heat advection or conduction. They are often associated with variations of lithosphere thickness within a continental plate, as imaged by seismic tomography surveys. Domains displaying an abnormally thin or thick lithosphere will have a significant mechanical effect on the deformation of a continental plate, as it will be illustrated using both numerical modelling results and natural cases in the next sections.

Heat advection is an efficient process through which the lithospheric geotherm may be locally increased and the total strength of the lithosphere lowered. This may occur in many geodynamic domains where an intense magmatism allows large amounts of heat to be transferred upward. At active margins, subduction-related partial melting may generate large volumes of magma (e.g., Iwamori, 1997) which advect heat to the overlying plate. High surface heat flow has been measured in volcanic arcs (e.g., along the western Pacific or the western North American margins; see Pollack et al., 1993 for a review), and seismic tomography usually images an abnormally hot lithosphere (e.g., Van der Lee and Nolet, 1997, Alsina and Snieder, 1996 for North America, or Van der Lee et al., 1997 for South America). Since these processes are active shortly before continental collision occurs, the thermally weakened continental margin may behave as a weak domain and accommodate a large part of the deformation, at least at the beginning of the collision process. These regions will be even weaker if they display a thick crust, as the Andes do for instance.

Thermal heterogeneities may also have a composite origin. For instance, delamination, i.e., the sinking of a piece of detached lithospheric mantle into the asthenosphere, may modify the thermal/rheological structure of the continental lithosphere, due to the replacement of relatively cold lithospheric material by hot asthenospheric mantle (e.g., see Marotta et al., 1998 - this issue). This process, which may be accompanied by partial melting, would occur in domains having an abnormally thick lithosphere, as a result of continental collision for instance. It is frequently invoked to explain positive thermal anomalies and the onset, in orogenic domains, of an extensional deformation favoured by the upward rebound of the lithosphere loosing part of its mantle root.

3.1. Lithosphere extension and rifting

A positive thermal anomaly may develop in relation with extensional basins and rift formation. The analysis of the thermal and therefore rheological outcome of rifting, however, is not straightforward, since opposite trends of thermal evolution may combine. The first issue is that the lithosphere beneath rifts is thinner, whatever the precise mechanism for rifting is (e.g., Achauer and group, 1994; Gao et al., 1994; Granet et al., 1995; Slack et al., 1996).

Lithospheric thinning in continental rifts is often associated to upwelling of mantle material due to gravitational instability. Mantle plumes are thought to propagate rapidly toward the surface up to the lithosphere boundary where they are stopped. Due to adiabatic decompression and negative slope of the Clapeyron curve for peridotite, the upwelling material partially melts and large volumes of hot magma propagate into the lithosphere advecting heat toward shallower levels. Moreover, heat exchange between the plume and the lithosphere by conduction provokes an upward deflection of the isotherms and of the asthenosphere–lithosphere boundary. The geotherm is therefore steeper and the surface heat flow may reach very high values. In the Rio Grande rift, for instance, the surface heat flow may reach 120–130 mW m^{-2} (e.g., Reiter et al., 1978; Pollack et al., 1993). A recent seismic survey has shown a 7–8% reduction of P-wave velocity beneath the Rio Grande rift relative to mantle velocities beneath surrounding areas, and joint analysis of S- and P-wave delays points to temperatures in the sub-crustal mantle close to the solidus (Slack et al., 1996). The effect of a mantle plume is also well illustrated by seismic tomography studies in the French Massif Central (Granet et al., 1995) where a temperature increase of up to 200ºC in the lithospheric mantle has been evaluated (Sobolev et al., 1996). Such large temperature variations may decrease considerably the total strength of the lithosphere.

Attenuation of the lithospheric mantle cannot be considered alone since coeval extension and thinning of the crust comes to replace crustal material by stiffer mantle rocks. As the thermal anomaly starts to vanish, progressive cooling of the lithosphere may further increase the lithospheric strength. England (1983) for instance, assuming a vertically uniform stretching, has shown that after an initial decrease in the average strength due to lithosphere attenuation, conductive cooling leads to a rapid increase in lithosphere strength for strain rates of 10^{-14} s^{-1}. Since crustal and mantle thinning have opposite rheological effects, the strength of the lithosphere in extensional areas will depend on the rate of lithospheric mantle (δ) to crustal (β) extension. Geophysical surveys (e.g., Davis et al., 1993; Achauer and group, 1994; Slack et al., 1996) and petrological studies on rift-related magmatism (e.g., Thompson and Gibson, 1994) suggest that in most recent rifts, like the Rio Grande, East African and Baikal rifts, the lithosphere attenuation is larger than the observed crustal thinning. For instance, Davis et al. (1993) from a tomographic inversion of teleseismic data and Cordell et al. (1991) from gravity surveys in the Rio Grande rift, suggest a total lithospheric thinning (in the range 2–4) much larger than the maximum crustal thinning ($\beta < 2$, Prodehl and Lipman, 1989). Moreover, seismic tomography in the Kenya rift points to a lithosphere–asthenosphere boundary as shallow as 50 km, whereas the crust is still 25–30 km thick (Achauer and group, 1994). If $\delta \gg \beta$, the rheological effect of rifting should be to lower the effective yield stress of the lithosphere (Tommasi and Vauchez, 1997).

The tectonic process through which lithosphere extension occurs may also be of importance. Four main models have been suggested and they may have contrasted rheological effects. Homogeneous ($\delta = \beta$) and depth-dependent ($\delta > \beta$) pure-shear extension (see review in Quinlan, 1988) will respectively result in a stiff or weak heterogeneity directly beneath the basin. It should however be considered that homogeneous thinning does not take time into account and is therefore unlikely (e.g., Fowler, 1990). Simple-shear lithospheric extension (Wernicke, 1985) is especially interesting since (1) factors δ and β vary across the extensional domain, and (2) the lithospheric structure is asymmetric (Fig. 3). In this model, $\beta > \delta$ just beneath the basin,

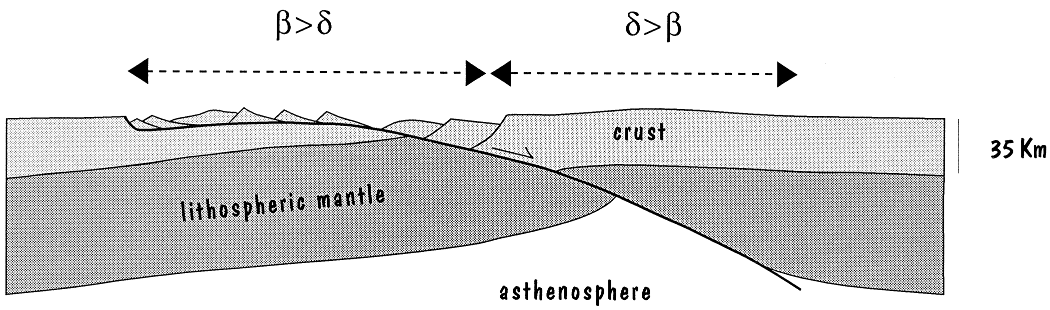


Fig. 3. Simple shear rifting of the lithosphere (Wernicke, 1985). The relative thinning of the mantle (δ) and of the crust (β) varies across the rift and may result in the development of a stiff heterogeneity below the basin, and a weak heterogeneity outside the basin, where $\delta > \beta$.

a situation that would generate a stiff heterogeneity. Outside the basin, where the detachment fault crosscuts the Moho, only the mantle is thinned ($\delta \gg \beta$), which will result in a notable decrease of the total lithospheric strength. Simple-shear extension should therefore produce a juxtaposition of a stiff domain correlated with the crustal basin and a weak domain beside the basin. The fourth model (Nicolas and Christensen, 1987) is based on a combination of seismic tomography data and field observations in rifts. It considers that rifting requires a first stage of lithospheric rupture accommodated through the introduction of a relatively narrow wedge of asthenosphere into the lithosphere. At this stage mantle thinning is considerably higher than crustal thinning ($\delta \gg \beta$) and the upwelling of an asthenospheric wedge would result in a narrow but substantially weak rheological heterogeneity. Subsequently, the rift geometry would evolve toward a classical 'passive rift' without reversing the crust/mantle attenuation ratio.

Extension of the lithosphere appears therefore as an efficient process to generate weak heterogeneities. The post-rifting evolution of these domains and the time interval between its development and reactivation are, however, crucial to its mechanical effect during subsequent deformation episodes. Morgan and Ramberg (1987) using the model of McKenzie (1978) calculated that the thermal relaxation of a palaeorift occurs in a time interval varying between 70 and 200 Ma, depending on the equilibrium thickness of the lithosphere (100 and 200 km, respectively). Moreover, for narrow rifts, significant lateral heat loss may result in a still shorter duration of the thermal anomaly. However, the model of McKenzie (1978) only simulates the thermal relaxation within the rifted zone, considering that the surrounding lithosphere is already in thermal equilibrium. If both thinned and normal lithospheres progressively cool, they may retain a rheological contrast for significantly larger time spans (Sahagian and Holland, 1993). Several other factors may counteract the strengthening effect of lithospheric cooling, like thermal blanketing due to syn-rift accumulation of sediments or post-extension subsidence and sedimentation which induce a deepening of the crust–mantle transition.

When rifting proceeds, the transition from continental to oceanic lithosphere may occur. Because the oceanic crust is thinner, oceans are much more resistant to deformation than continents at similar lithospheric thickness. However, during the initial stages of an oceanic basin ($t < 20$ Ma), the new oceanic lithosphere is extremely thin and the geotherm very steep. As a result, the lithospheric strength is significantly lower in such basins than in the surrounding continental lithosphere (Fig. 4). The newly formed oceanic lithosphere may represent a weak rheological heterogeneity and this may have important tectonic consequences. Back-arc basins along an active margin may, for instance, localize the deformation during continental collision, and impede an efficient stress transmission to the continent until they have been completely closed.

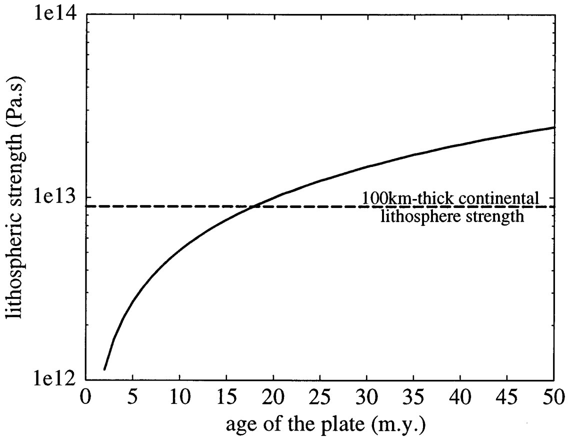
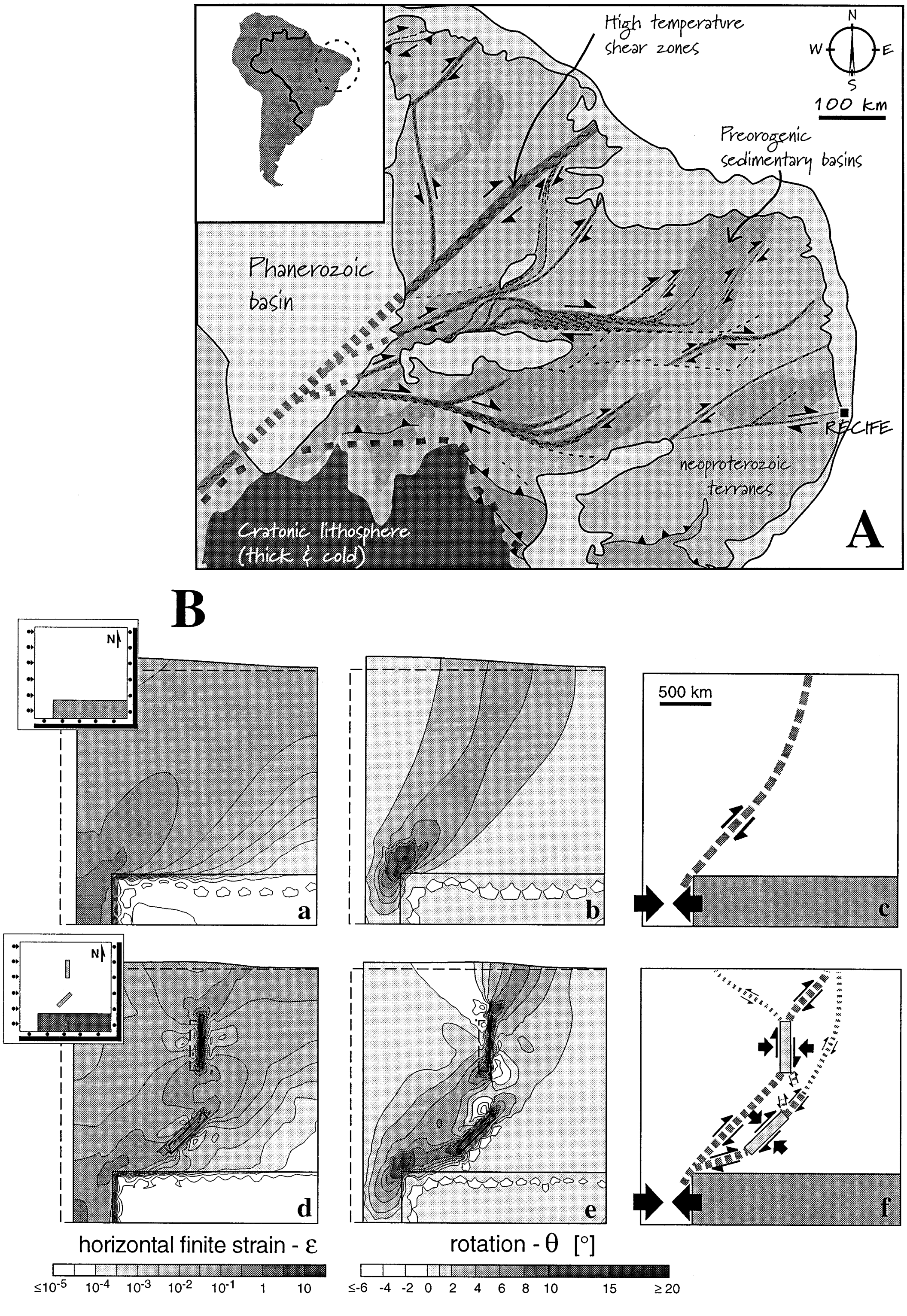


Fig. 4. Evolution of the strength of a young oceanic lithosphere with time. The strength calculated for a continental lithosphere 100 km thick (Fig. 1a) is plotted as a reference. The oceanic lithosphere remains weaker than the continental lithosphere for at least 15 Myr.

3.2. Influence of pre-existing rifts on the strain field: examples

Recent numerical models (Tommasi et al., 1995; Tommasi and Vauchez, 1997) show that thermally induced rheological heterogeneities affect strain localization, shear zone development, and the distribution of deformation regimes and vertical strain within a continental plate. The topology and boundary conditions of these models are inspired by the geological situation of the Borborema Province of northeast Brazil (Vauchez et al., 1995), where a



High temperature shear zones
N
W
E
S
100 km
Preorogenic sedimentary basins
Phanerozoic basin
RECIFE
neoproterozoic terranes
Cratonic lithosphere (thick & cold)
A
B
500 km
a
b
c
d
e
f
horizontal finite strain - ε
rotation - θ [°]
≤10-5 10-4 10-3 10-2 10-1 1 10
≤-6 -4 -2 0 2 4 6 8 10 15 ≥20

complex network of NE–SW- and E–W-trending right-lateral shear zones was formed during the Neoproterozoic (Fig. 5A). The models take into account the pre-deformational rheological structure of the lithosphere in this area, i.e., the presence of a craton (the São Francisco craton) southward, and several sedimentary basins containing felsic or bimodal volcanic layers. A 2500×2500 km quadrilateral plate (Fig. 5B), containing a stiff and one or several low-viscosity domains, simulates a 'normal' continental lithosphere surrounding a craton and containing intracontinental basins. From the first increments of deformation, both weak and stiff heterogeneities induce strain localization, due to their low effective viscosity or to stress concentration at their tips, respectively. Shear zones propagate from the heterogeneities and finally coalesce, forming a network of high-strain zones that bound almost undeformed blocks. Within this network, shear zones transfer strain between the different heterogeneities. The evolution of the system depends essentially on the geometrical distribution of heterogeneities and on their strength contrast relative to the surrounding lithosphere, i.e., on the variation in geotherm between the different domains. The resulting strain field is heterogeneous and displays rapid lateral variations in vertical and/or rotational deformation.

Convergence of two heterogeneous plates (Africa/Arabia and Europe) is also proposed to explain the origin of the Dead Sea rift (Lyakhovsky et al., 1994). In their models, the stiff Africa–Arabia and European plates are separated by a weak domain representing the young oceanic lithosphere of the Mediterranean Sea and Cungus basin (Fig. 6). The Red Sea and the site of the future North Anatolian fault are represented as damaged (less resistant) zones within the plates. The mechanical evolution of these models shows that (1) once collision between Arabia and Europe starts, strain localizes within the Cungus basin, favouring the development of a transtensional shear zone (the Dead Sea fault, Fig. 6b,c), that transfers strain from the Red Sea to the Cungus basin, and (2) the damaged zone at the location of the North Anatolian fault did not localize strain, and no escape of the Turkey microplate was observed, until the Cungus basin was completely closed (Fig. 6c). Beyond its applicability to the Dead Sea evolution, this model illustrates the role that weak oceanic basins may play during a continental collision.

The effect of pre-existing intraplate basins was documented by Armijo et al. (1996) in the Aegean region, where several rifts have formed since 15 Myr in relation with the Hellenic subduction. At ca. 5 Ma, the propagation direction of the North Anatolian fault, which accommodates the westward extrusion of Anatolia, changed from westward to southwestward (Fig. 7). Subsequently, several branches splayed off the main E–W-trending fault and propagated southwestward toward the pre-existing rifts into which they terminate. This resulted in a lithospheric-scale reactivation of the western part of the rift system (Armijo et al., 1996). We suggest that the re-orientation of the North Anatolian fault, its propagation towards pre-existing basins and finally the reactivation of these basins results from the interplay between the kinematic boundary conditions (Arabia–Eurasia collision and Anatolia extrusion, Hellenic subduction) and the rheological heterogeneity of the lithosphere in the Aegean domain due to previous lithosphere extension and rifting.

The weakness of young oceanic basins may also have favoured a change in stress regime from extensional to compressional in the northwestern margin of the China Sea, in association with an inversion of shear sense on continental-scale strike-slip faults resulting from the India–Eurasia collision. According to the model of Tapponnier et al. (1986) of the tectonic evolution of eastern Asia, oceanization of the China Sea occurred at the tip of the left-lateral Red River fault in relation with the extrusion of Sundaland (SE Asia) between 50 and 17 Ma (Fig. 8).

Fig. 5. Numerical models inspired from the situation of the Borborema Province of NE Brazil (A). A continental plate involving one high-viscosity domain (B, a–c) and one high- and two low-viscosity domains (B, d–f) is submitted to compressional deformation. Inserts show boundary conditions and topology. (c) and (f) present a tectonic interpretation of numerical models. The presence of a stiff block induces strain localization at its tip from which a shear zone originates and propagates at 45º of the compression direction. When weak domains are added (d,e), the high strain zone is partly deviated to connect with the closer weak 'basin', and a complex network of shear zones forms.

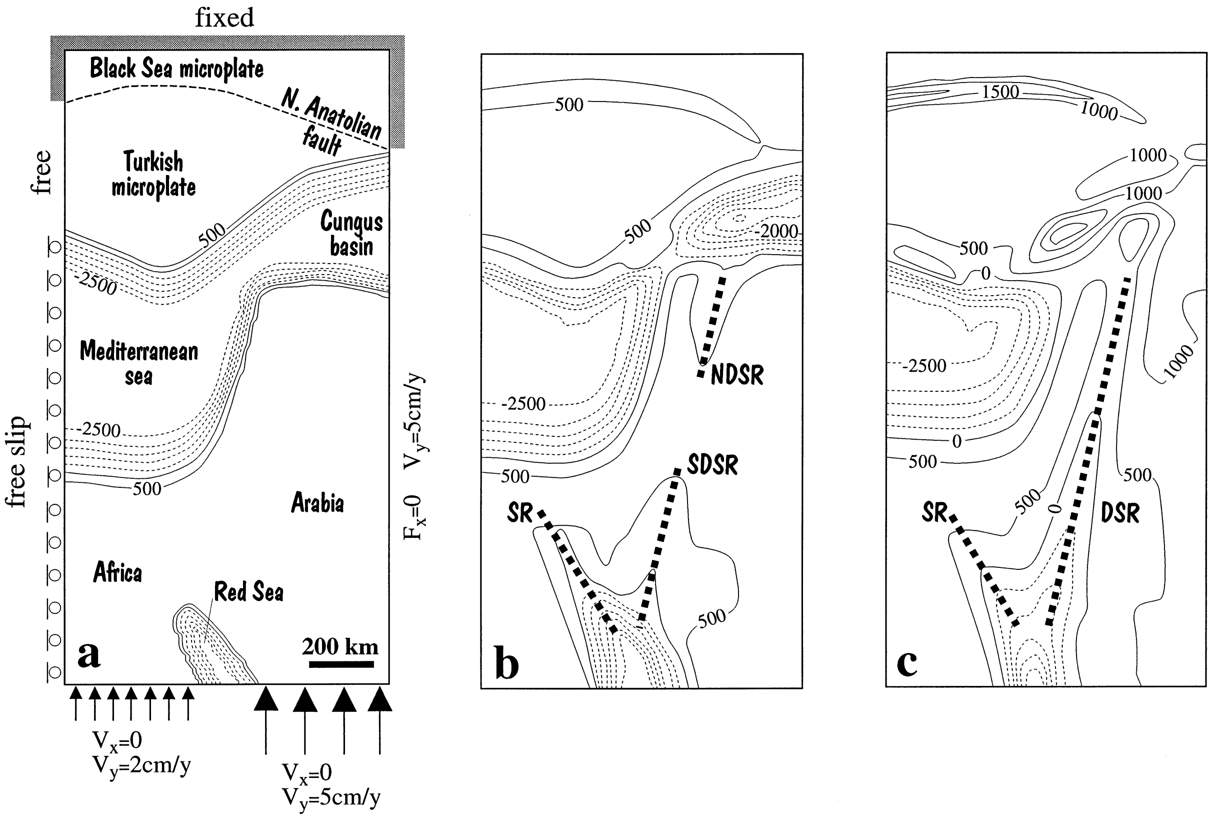


Fig. 6. Numerical models simulating the convergence of the Eurasia and Africa/Arabia plates (redrawn from Lyakhovsky et al., 1994). Three successive stages are shown illustrating the preferential deformation of the Cungus basin and the development of the Dead Sea fault zone. Most of the differential displacement between Africa and Arabia is accommodated within the Cungus basin, and the North Anatolian fault localizes strain only after the closure of the Cungus basin. Contours are topography. *DSR* = Dead Sea rift; *NDSR* and *SDSR* = northern and southern branch of the DSR; *SR* = Suez rift; boundary conditions are shown in (a).

Around 17 Myr ago, the sense of displacement along the Red River fault reversed to right-lateral to accommodate the eastward extrusion of the South China block. Currently, the state of stress in the China Sea area is compressive (Zoback, 1992). This inversion of tectonic regime could be due to the growth of the weak China Sea at the tip of the Red River fault. In the regional framework of the India–Asia collision, this rheological heterogeneity may account for an easier extrusion of the South China block than of the Sundaland block. As a matter of fact, the present-day tectonic system formed by the Red River fault and the China Sea is quite similar with the model developed for the Borborema shear zone system (Tommasi et al., 1995; Tommasi and Vauchez, 1997) in which continental-scale transcurrent shear zones terminate into weak continental basins that deform through transpression.

The active deformation of the Tyrrhenian back-arc basin may also be interpreted following a comparable scenario (J.C. Bousquet, pers. commun., 1996). Currently, the basin is characterized by a local seismicity that denotes an ongoing compression and by a high surface heat-flow. This compressional deformation of the basin is associated with left-lateral displacement along the NNW-trending Aeolian–Maltese fault system. During the Quaternary, this fault propagated from the Mediterranean basin through the southern Tyrrhenian basin to the Aeolians to accommodate a differential displacement of the western block relative to the eastern one, probably associated with the compression of the weak back-arc basin.

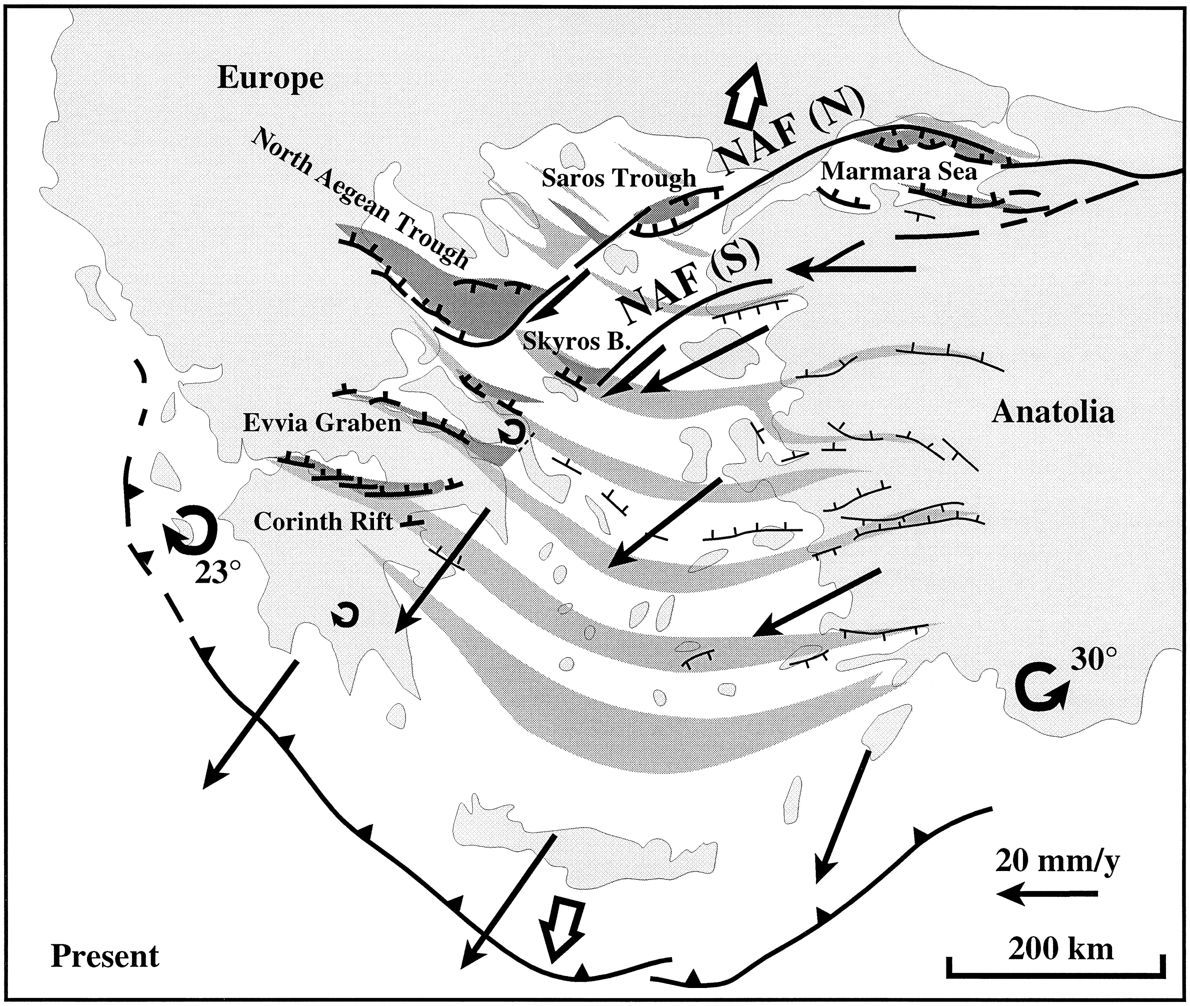


Fig. 7. Effect of pre-existing rift basins on the propagation of the North Anatolian fault (*NAF*). This figure from Armijo et al. (1996) shows the present-day kinematics in the Aegean region. The northern and southern branches of the westward propagating NAF have deviated southwestward to propagate toward the rift basins (light grey) formed between 15 and 5 Myr ago. Where they connect with pre-existing rifts, the fault segments have reactivated the rift structure (darker grey). Narrow arrows = today displacements. Curved arrows = finite rotation. Large open arrows = extension related to the Aegean subduction.

3.3. Development of stiff heterogeneities: aging of the continental lithosphere

The rheology of the continental lithosphere evolves continuously due to progressive cooling and associated thickening of the lithosphere (see review in Sclater et al., 1980) by accretion of cooled asthenospheric material. A consequence of this evolution is a strength increase with increasing tectonic age of the lithosphere. Cratons represent the best illustration of this process. Most continental plates developed by successive accretions around a cratonic nucleus. Although a long time elapsed since the assembly of these continents (~600 Myr for the African and South American plates for instance), cratonic nuclei still display lower both surface (q_s) and reduced (q_m) heat flows than adjacent terranes (e.g., in the Archaean Kaapvaal craton $q_s < 40$ mW m^{-2} and $q_m \approx 17$ mW m^{-2} Jones, 1988; Nyblade and Pollack, 1993). Thermo-barometric calibrations

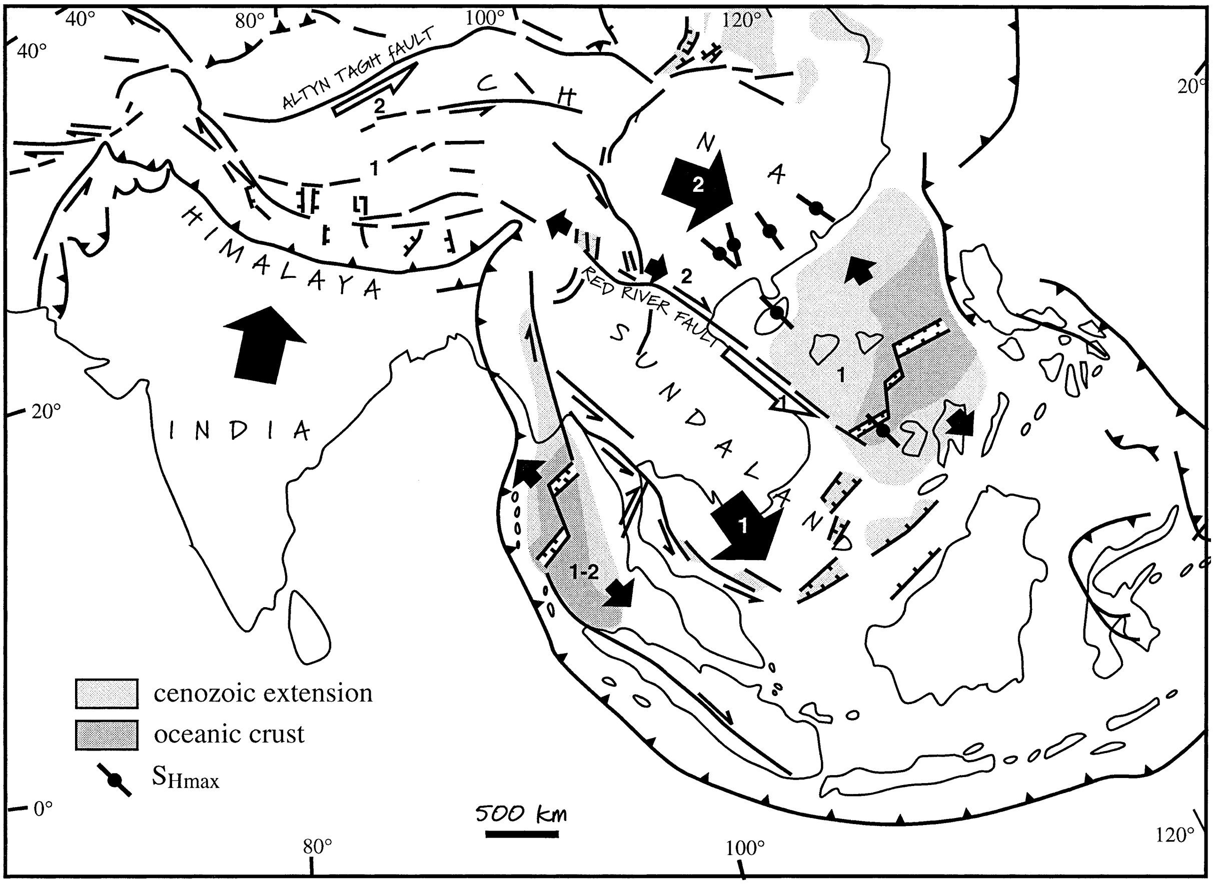


Fig. 8. Schematic map of the India–Eurasia collision system (modified from Tapponnier et al., 1986). This map shows the South China Sea basin with a domain of continental lithosphere extended during the Cenozoic, and the oceanic lithosphere. According to Tapponnier et al. (1986) this basin formed in relation with Cenozoic left-lateral shearing (open half-arrow) along the Red River fault. From 17 Ma to present, the sense of displacement along the fault reversed to right-lateral (small black half-arrow) and the northwestern margin of the China Sea is currently submitted to compression (data from Zoback, 1992). Numbers refer to period of evolution: *1* = 50–17 Ma, *2* = 17 Ma to present.

of mantle xenoliths from kimberlites also point to a low geothermal gradient and a thick lithosphere (e.g., Boyd et al., 1985). Recently, seismic tomography surveys have shown deep continental roots characterized by high seismic wave velocity, probably due to lower upper mantle temperatures, beneath cratonic domains (Grand, 1994; Polet and Anderson, 1995; Van der Lee and Nolet, 1997). This suggests that cratonic blocks have cold geotherms and, therefore, a high stiffness (Fig. 9). For such low geothermal gradients, the uppermost lithospheric mantle is able to sustain very high stresses and may be considered as almost rigid at geologically reasonable strain rates. This conclusion is in good agreement with the observation that cratons largely escaped more recent tectonic events which formed the 'mobile belts' at their boundaries.

The Precambrian shields of India, Tarim and Ankara, are good examples of stiff blocks involved in a continental collision. Observing that these blocks have not deformed significantly since the India–Asia collision, whereas younger domains were extensively deformed, Molnar and Tapponnier (1981) first suggested a dependence of the intensity and style of active tectonics on the age of the last orogenic activity in Asia, and hence on the strength contrast between the various terranes that compose the Eurasian plate. Numerical models simulating the role of the Tarim

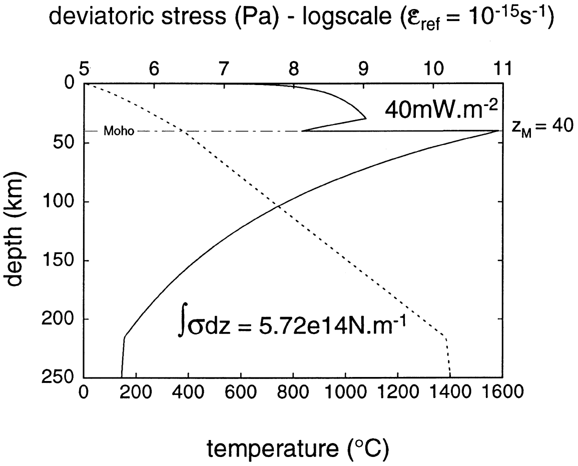


Fig. 9. Rheological profile for a cratonic area assuming a granulitic crust. Rheologic parameters and constitutive equation as in Fig. 1. The integrated strength is almost 2 orders of magnitude higher than the strength of the younger lithosphere of Fig. 1a, calculated with a normal geotherm.

shield during the India–Eurasia collision (Vilotte et al., 1984; England and Houseman, 1985) indicate that stiff inclusions within a continental plate submitted to compression induce stress concentration at their edges or tips, that favour strain localization and initiation of shear zones in the surrounding material, whereas the stiff domains tend to behave as rigid blocks and remain almost undeformed.

A similar effect was proposed for the Archaean to Palaeoproterozoic São Francisco craton of Brazil (Vauchez et al., 1994). This craton displays a thicker lithosphere (>200 km, Grand, 1994; VanDecar et al., 1995) and a lower average surface heat flow than the surrounding Pan-African belts (42 ± 5 mW m^{-2} and 55 ± 4 mW m^{-2}, respectively; Vitorello et al., 1980). At the southern termination of the craton, the Neoproterozoic Ribeira–Araçuaí belt is characterized by significant modifications of its tectono-metamorphic pattern (Fig. 10a). From north to south: (1) the structural trend is bent from N–S to NE–SW; (2) the dominant tectonic flow shifts from orogen-transverse thrusting to orogen-parallel strike slip faulting; and (3) synkinematic metamorphic conditions decrease from high to low grade. Numerical models (Fig. 10b,c) simulating the compressional deformation of a plate comprising a large stiff heterogeneity (Vauchez et al., 1994) show a partitioning of the deformation and the development of a deformation pattern similar to the one observed in the Ribeira–Araçuaí belt. The northern domain, squeezed between the heterogeneity and the converging boundary, displays a dominant pure shear deformation (Fig. 10c), characterized by normal shortening and associated thickening. At the tip of the stiff domain, a wide dextral shear zone initiates and propagates at 45º to the convergence direction (Fig. 10b). Development of continental-scale strike-slip faults due to stress concentrations at the tip of stiff domains was also suggested by Mattauer (1986) for the Chaman fault and Tommasi et al. (1995) for the Borborema shear zone system of northeast Brazil.

The Baikal rift system provides a different example of the interplay between preexisting rheological heterogeneities within a continental plate and remote tectonic forces (due to the India–Asia collision). The rift formed at the boundary between the Siberian craton and younger Mongolian belts. Palaeo- and present-day stress-field analysis (Delvaux et al., 1995; Petit et al., 1996) points to a consistent relationship between changes in the dominant tectonic regime and the geometry of the craton boundary. At the tip of the craton, in the south Baikal rift, the tectonic regime depicted by earthquake focal mechanisms changes abruptly from wrench-compressional to extensional. Nevertheless, the general agreement between average locals and regional S_{Hmax} directions suggests that the wrench-compressional tectonics in Mongolia and the extensional regime in the Baikal are both far-field effects of the India–Asia collision.

4. Mechanical anisotropy

Anisotropic strength of crystalline solids is well known and is extensively studied in material sciences, since it represents an important factor for the resistance of metals and ceramics. Structural anisotropy may result from either a grain-shape-preferred orientation, a heterogeneous distribution of the constituents in a polyphase aggregate, or a lattice-preferred orientation (LPO) in an aggregate composed by mechanically anisotropic constituents (e.g., Parnière, 1982). As most materials deforming by dislocation creep, like metals, ceramics or ice, rocks

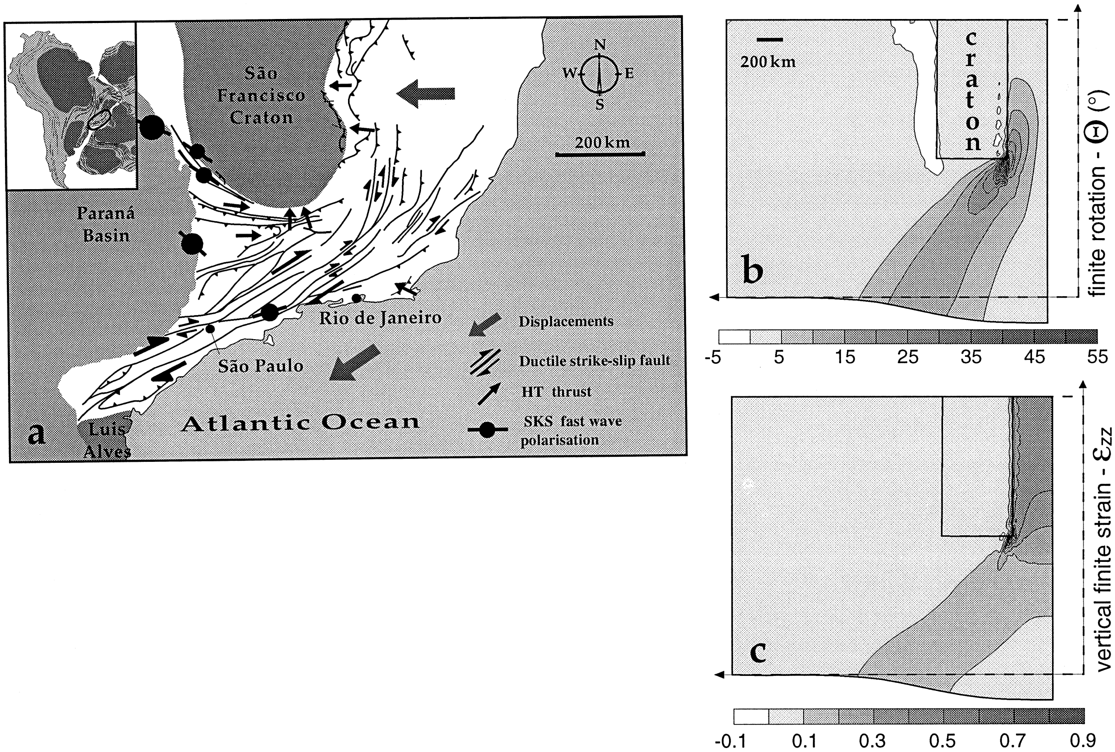


Fig. 10. Schematic structural map (a) of the Ribeira Araçuai belt and numerical model (b,c) simulating the deformation of a continental lithosphere involving a cratonic domain. The map illustrates the variation in dominant deformation regime from orogen-normal thrusting to the north to orogen-parallel transcurrent faulting to the south. Fast split shear wave polarization measured in this area is from James and Assumpção (1996). Model results show: (b) the development of a 'right-lateral shear zone' (positive finite-rotation domain) at the tip of the stiff block ('craton') and (c) a high-strain zone confined between the stiff block and the boundary of the model, where thickening dominates.

often display a mechanical anisotropy due to deformation-induced lattice-preferred orientation of its constituent crystals. A link is usually found between anisotropy and the deformational structure of rocks. In tensional experiments, for instance, flow stresses are generally lower for extension normal than parallel to the foliation plane (Heard and Raleigh, 1972; Sirieys, 1982).

Available data usually refer to the mechanical anisotropy of mica-rich crustal rocks (e.g., Borg and Handin, 1966; Gottschalk et al., 1990). However, this anisotropy is not restricted to crustal rocks. Olivine, the main constituent of the upper mantle, is also mechanically anisotropic. Microstructural analysis of deformed peridotites (Nicolas et al., 1971; Mercier and Nicolas, 1975) and experimental axial compression of olivine single crystals in different crystallographic orientations (Durham and Goetze, 1977; Bai et al., 1991) provide evidence that (1) few slip systems are available to accommodate plastic deformation, and (2) these systems display a significantly different resistance to flow. The relative strength of the different slip systems in olivine depends strongly on temperature. Under experimental conditions the (010)[100] slip system is the weakest at high temperatures ($T \geq 1200$ºC) and should therefore accommodate most of the strain (Fig. 11). A transition from this high temperature [100]-glide to low-temperature [001]-glide occurs at ca. 1000ºC (Durham and Goetze, 1977). However, at natural strain rates, this transition occurs at lower temperatures, and the activation of the low-temperature {110}[001] system in natural rocks is restricted to mylonites that accommodate the emplacement of peridotite slices into the crust. Moreover, analysis of more than 200 lattice-preferred orientations of natu-

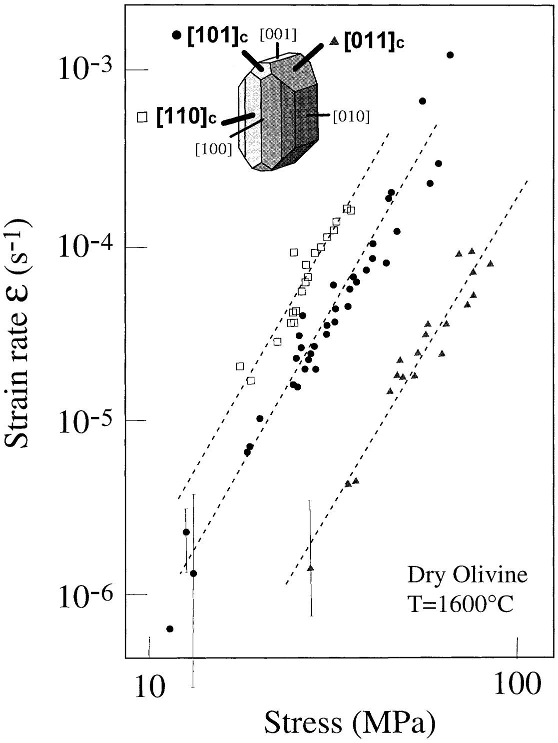


Fig. 11. Stress–strain rate plot for dry olivine single crystals compressed in three different directions ([110]*c*, [101]*c*, [011]*c*) relative to the crystallographic structure. At any strain-rate, deformation is easier for crystals shortened in the [110]*c* direction, due to the lower strength of the (010)[100] slip system. After Durham and Goetze, 1977.

rally deformed peridotites from various geodynamic environments (Ben Ismail and Mainprice, 1998 - this issue) clearly indicates that olivine deforms predominantly by glide on the (010)[100] slip system with subsidiary glide on other systems of the {0*kl*}[100] family, suggesting that, as indicated by single crystal deformation, under upper mantle conditions, the weakest slip system for olivine is (010)[100].

Upper mantle rocks develop crystallographic preferred orientations of olivine during deformation by dislocation creep; they should therefore display a significant mechanical anisotropy. Recent experimental axial compression tests of textured dunites under high temperature and pressure (P. Chopra, pers. commun., 1995; Wendt et al., 1998) confirm this hypothesis: the strength of pre-textured dunites depends on the orientation of the compression relative to the foliation. The lowest strength and the most ductile behaviour are observed for compression at 45° to the foliation, i.e. for the orientation that produces the maximum resolved shear stress for the 'weak' (010)[100] slip system. The mechanical behaviour of a model dunite (100% olivine) displaying an initial lattice-preferred orientation may also be simulated through a self-consistent viscoplastic model (Lebensohn and Tomé, 1993). In these simulations, strength ratios larger than 2 are obtained depending on the stress orientation relative to the crystallographic fabric, and the lowest strengths correspond to loading orientations resulting in a deformation accommodated by slip on the 'weak' (010)[100] or (001)[100] systems (Tommasi et al., submitted).

Mechanical anisotropy is a property of olivine single crystal and aggregates. If the lithospheric strength also is dependent on the orientation of the tectonic forces applied to the plate, this property should influence the deformation of continents. Do we have any evidence that this anisotropy is also manifested at the scale of the lithospheric mantle? Although it is not possible to answer this question with certainty, geophysical and geological observations point toward an anisotropic tectonic behaviour of the lithosphere.

4.1. Tectonic fabric of the continental lithospheric mantle

Direct observation of the continental mantle is possible in several lherzolite massifs. Although tectonic emplacement within the crust resulted in partial reworking of the initial fabric, undisturbed lithospheric mantle structures are often retained in the core of the massifs. This is, for instance, the case of the Lanzo massif (Boudier, 1978) and lherzolite bodies of the Ivrea zone in the Alps (e.g., Peselnik et al., 1977), the Ronda massif in the Betic Cordillera (Tubia and Cuevas, 1987), and the Lherz massif in the Pyrenees (Avé-Lallemant, 1967). A consistent tectonic fabric (foliation, lineation, crystallographic preferred orientation), formed under lithospheric conditions (900°C $< T <$ 1200°C), has been mapped over these massifs. This suggests that the lithospheric mantle, before it was incorporated into the crust, already displayed a fabric consistent at the scale of several tens of kilometres at least before it was incorporated into the crust.

Seismological and magnetotelluric observations allow an extension of this conclusion at a larger scale. Rock-forming minerals (especially olivine) are seismically anisotropic, their crystallographic preferred orientation therefore produces a significant seismic anisotropy of upper mantle rocks (e.g., Babuska, 1981; Nicolas and Christensen, 1987; Mainprice and Silver, 1993) and is regarded as the main source of teleseismic shear wave splitting (e.g., Silver and Chan, 1988; Vauchez and Nicolas, 1991; Vinnik et al., 1992). Short-scale spatial variations of splitting parameters on continents and a good correlation with surface geology (e.g., Silver, 1996; Vauchez and Barruol, 1996; Barruol et al., 1997) point to a dominant contribution of continental lithospheric mantle fabric to the splitting of shear waves. Most commonly, the fast split shear wave is polarized subparallel to orogenic structural trends (see review in Silver, 1996). The difference in arrival time between the fast and slow waves is usually around 1 s and may exceptionally exceed 2 s. Considering the intrinsic anisotropy of mantle xenoliths, such time lags require a layer of anisotropic mantle 100 km thick at least, with a preferred orientation of the [010] and [100] axes of olivine, respectively perpendicular and close to the strike of the orogen (Mainprice and Silver, 1993). These measurements hint at the existence of a crystallographic fabric coherent over the entire thickness of the lithospheric mantle.

Pn azimuthal anisotropy, although more sensitive to lateral heterogeneity, leads to a similar conclusion. Pn waves propagate just beneath the Moho (Fig. 2) and the fast propagation direction is parallel to the olivine [100]-axes, i.e., to the frozen flow direction (e.g., Babuska and Cara, 1992). Azimuthal variations in Pn travel time therefore reflect the structure of the uppermost lithospheric mantle only. Consistent patterns are found over large areas and display a good correlation with both shear wave splitting parameters and tectonic features at the surface (e.g., Hearn, 1996).

Electrical conductivity anisotropy evidenced by magnetotelluric soundings is interpreted as resulting from the existence of conducting graphite films along grain boundaries, with the highest conductivity parallel to the foliation trend (e.g., Sénéchal et al., 1996). It therefore records the existence of a shape-preferred orientation in the upper mantle. The depth at which electrical anisotropy occurs is related to the frequency range in which anisotropy is observed, and is therefore determinable. Although few measurements are available, they display an internally consistent anisotropy pattern, and a good correlation with shear wave splitting and surface geology. Sénéchal et al. (1996), for instance, through coupled shear wave splitting measurements and magnetotelluric soundings along a transect in the Canadian shield, retrieved similar directions of anisotropy from both datasets and, considering that the measured electrical anisotropy originates between 50 and 150 km, they concluded that the seismic anisotropy also originates from a well-defined fabric in the lithospheric upper mantle fabric. Similar preliminary results have been obtained in the Appalachians (Wannamaker et al., 1996) and the Pyrenees (Pous et al., 1995; M. Daignières, pers. commun., 1997).

The good consistency between these data strongly support that the lithospheric mantle displays a tectonic fabric consistent over its entire thickness and this fabric is laterally coherent over several tens to several hundreds of kilometres. As the lithospheric mantle displays a pervasive tectonic fabric and olivine single-crystals and aggregates are mechanically anisotropic, it may be expected that the lithosphere will also behave anisotropically when submitted to tectonic forces.

4.2. Geological examples

Mechanical anisotropy of the lithosphere underlies the classical concept of structural inheritance that was frequently invoked to explain the systematic reactivation of pre-existing tectonic structures. This concept may be found for instance in the textbook of De Sitter (1964) or in the milestone paper on tectonic (Wilson) cycles of Wilson (1966). Structural inheritance was generally related to the crustal structure. This may be true when reworking is restricted to local tectonic structures, especially fault zones. However, when the deformation is at the continent-scale and clearly affects the entire lithosphere, a crustal origin of structural inheritance is unlikely if, as suggested by rheological models, the lithospheric mantle rheology guides the mechanical behaviour of the lithosphere.

Presence of a tectonic fabric in the lithospheric mantle seems to have strongly influenced subsequent deformation of the lithosphere. In the eastern US, for instance, the 1Ga Grenville belt direction was reactivated during Neoproterozoic–early Palaeozoic extension, then during Caledonian (Ordovician), Acadian (Late Devonian–Early Carboniferous), Alleghanian (Late Carboniferous) compressive deformations and finally during early Mesozoic opening of the Atlantic Ocean. In Europe, a clear parallelism is observed in the Pyrenees between the Alpine and Hercynian structures (e.g., Vauchez and Barruol, 1996). The initial rupture between Iberia and Europe occurred parallel to the Hercynian belt in the future site of the Pyrenees. The displacement of Iberia relative to Eurasia occurred along this newly formed plate boundary (e.g., Choukroune, 1992). Finally, North–South convergence of Iberia and Europa, mainly during the Eocene, led to the formation of the Pyrenees which also parallels the older Hercynian grain. In both the Appalachians and the Pyrenees, an inherited pervasive lithospheric fabric was suggested from shear wave splitting analysis which showed a fast shear wave polarization sub-parallel to either the Grenvillian or Hercynian tectonic fabric (Vauchez and Barruol, 1996; Barruol et al., 1997).

Nevertheless, the most spectacular consequence of mechanical anisotropy of the lithospheric mantle is probably the tendency of continents to break-up along old orogenic belts (Vauchez et al., 1997a). Many examples are found in the literature: the North Atlantic rift in eastern North America (e.g., Wilson, 1966) and in Morocco (Piqué and Laville, 1996), the Rio Grande rift (Olsen et al., 1987), the Northeast China rift (e.g., Ma and Wu, 1987), the Baikal rift (e.g., Delvaux et al., 1995), the East African rift (Ring, 1994; Theunissen et al., 1996), the West African rift (e.g., Fairhead and Binks, 1991), the Cape Graben (Burke, 1976), and the Eastern Brazilian rift (Chang et al., 1992). Recent shear wave splitting surveys in the Kenya and the Baikal rifts (Gao et al., 1997) show that at the rift margin, where large-scale modification of the lithospheric mantle by upwelling asthenospheric mantle is unlikely, fast shear waves are polarized parallel to the trend of the Mozambique and Mongolian orogenic belts respectively (Fig. 12). These observations suggest that the Kenya and Baikal rifts formed parallel not only to crustal structures of the Mozambique and Mongolian belts, but also to a pre-rift fabric frozen in the upper mantle since these orogenies.

At a larger scale, the Atlantic Ocean opening was also largely guided by the tectonic fabric of the Gondwana and Laurentia continents. The initial South Atlantic rift (Fig. 13a) propagated over more than 3000 km parallel to the Hercynian Cape fold belt, then to the Malmesbury, Kaoko, Dom Feliciano, Ribeira and West Congo Neoproterozoic belts, and also to the Palaeoproterozoic Itabuna belt in the São Francisco craton. This striking parallelism is maintained even south of Africa where the relative displacement of continents was essentially transtensional. The North Atlantic initial rift propagated from the Central Atlantic along the Hercynian belt until it deviated towards the Mediterranean basin following a major Hercynian transcurrent fault which linked the Newfoundland with the south Iberian Hercynian segments (Fig. 13b). The reactivation of this fault accommodated a differential displacement between the North African and the Iberian branches of the Hercynian belt. Northward, the opening of the Bay of Biscay started when the North Atlantic rift, following the curvature of the Ibero–Armorican Hercynian segment, wrapped around Iberia and produced the rupture between Iberia and Europe along the North Pyrenean fault which reactivated Hercynian structures. Finally, as the direction of the Hercynian belt curved to almost E–W, a direction unfavourable to opening, the North Atlantic rift followed the Caledonides belt northward.

Continental-scale steep transcurrent faults (or shear zones) are preferentially reactivated during rift propagation, even when they are oblique on the general trend of the orogenic belts. This was already highlighted by Daly et al. (1989), for instance, for the upper Palaeozoic Karoo basins of South Africa: Proterozoic shear zones closely controlled the location, orientation and tectonic styles of these basins. Another example lies in the West African rift system (e.g., Fairhead and Binks, 1991), where the Benoue Trough, the Central African basins and the Cameroon volcanic line reworked the transcontinental strike-slip fault system formed by the Sanaga and Adamawa Neoproterozoic faults and their Brazilian counterparts (Vauchez et al., 1995). Finally, in the East African rift system, the southern part of the Gre-

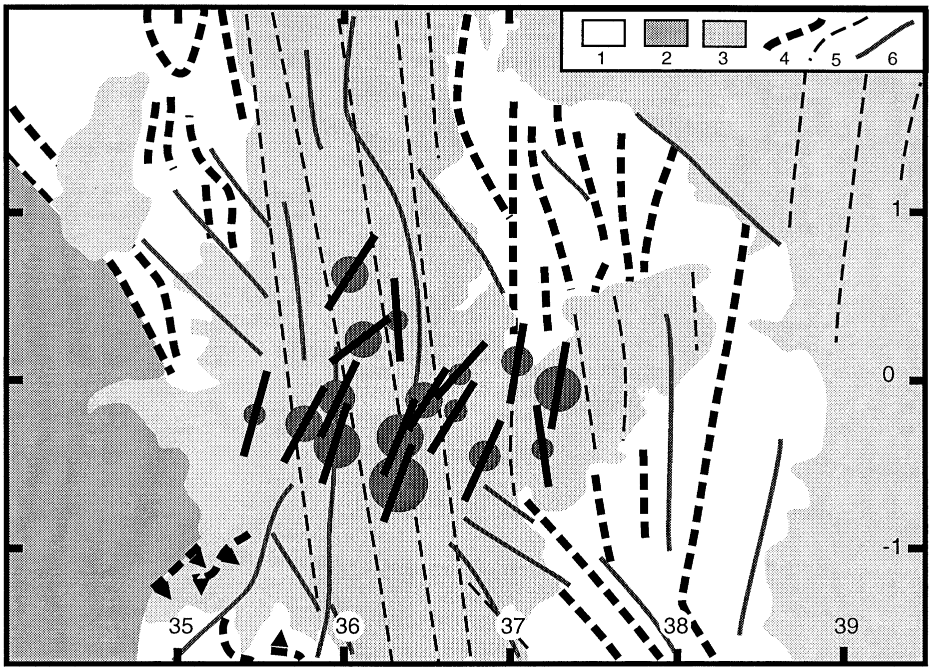


Fig. 12. Shear wave splitting measurements in the Kenya rift (Gao et al., 1997) compared to the orogenic fabric in the same area on the eastern boundary of the rift. The fast shear waves polarization direction is close to N–S and parallels the Neoproterozoic fabric of the Mozambique belt. For splitting measurements, the size of the circle is proportional to the time lag between the fast and slow wave arrivals, and the orientation of the bar is parallel to the orientation of the fast split wave polarization plane. *1* = Neoproterozoic basement; *2* = Tanzania craton; *3* = rift-related sediments; *4* and *5* = observed (4) and inferred (5) Neoproterozoic structural fabric; *6* = rift-related normal faults.

gory rift follows a Neoproterozoic lithospheric shear zone oblique on the general trend of the Mozambique belt (e.g., Cardon et al., 1997). A similar process may account for other basins oblique on the general orientation of the rift system, that otherwise follows the N–S Neoproterozoic Mozambique belt.

Mechanical anisotropy of the upper mantle may have a significant influence on intracontinental seismicity. The slow deformation of the upper mantle, which results from an interplay between the orientation of the mantle fabric and the bulk stress field, may control the orientation, faulting mechanisms and displacement velocities of active faults in the brittle crust. This may, for instance, explain the frequently observed transcurrent component of earthquake mechanisms during rifting (e.g., Doser and Yardwood, 1991). Both experiments and numerical models support that the lowest lithospheric strengths correspond to an upper mantle deformation accommodated by slip on the ‘weak’ (010)[100] or (001)[100] slip systems of olivine. Thus if, as suggested from shear wave splitting measurements (Mainprice and Silver, 1993), the continental lithosphere in collisional belts, in particular those involving continental-scale strike-slip faults, is characterized by a steep foliation and a low-angle lineation, preferential orientation of the weak (010)[100] slip system of olivine should induce reactivation of the initial fabric involving a strike-slip component.

5. Conclusions

The rheological stratification of the lithosphere was frequently illustrated by strength profiles built using (1) slicing of the lithosphere to simulate its vertical variation in composition, (2) simplified constitutive equations, and (3) experimental rheological parameters difficult to extrapolate. Strength profiles and integrated lithospheric strength are useful to

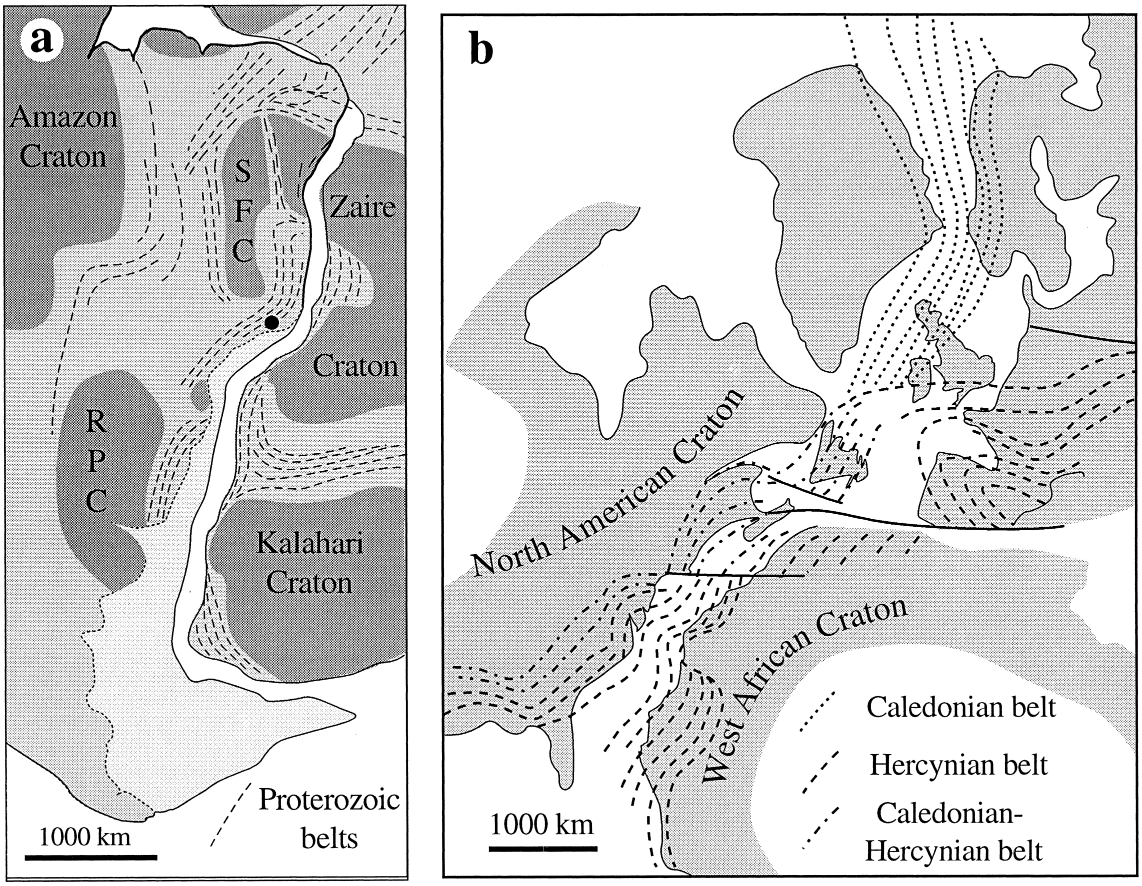


Fig. 13. Schematic maps showing the parallelism between the trend of pre-rifting orogenic belts and the initial rifting in the South (a) and North (b) Atlantic oceans. In both oceans, the initial rift followed orogenic belts or major transcurrent faults over thousands of kilometres.

compare the rheology of different geodynamic domains. Their integration into tectonic models, however, requires caution. The popular concept of decoupling levels in the middle crust and at the crust–mantle interface was proposed from this kind of approach. Although decoupling may certainly occur in some circumstances, geological and geophysical observations, together with new experimental data, clearly demonstrate coupling of the lower crust and upper mantle, especially beneath major transcurrent faults. It is therefore suggested that, in these cases, lithospheric mantle deformation guides the tectonic behaviour of the lithosphere (with the exception of domains with a high geothermal gradient), and that in most cases the crust merely accommodates the imposed deformation.

Continents grow through collisions and collage of different lithospheric blocks. After amalgamation, continents are submitted to various processes that locally modify the thermal and tectonic structure of the lithosphere; hence, they cannot be regarded as homogeneous and isotropic. The existence of stiff and weak heterogeneities significantly influence the deformation of continents in response to tectonic forces. The resulting strain field is heterogeneous: weak domains accommodate more deformation, stiff ones tend to remain undeformed, continental-scale transcurrent faults tend to develop and transfer strain between the heterogeneities. Vertical strain also is variable and this should result in complex metamorphic history, topography, and sediments distribution. Other possible causes of rheological heterogeneity still need to be investigated. Lateral variations in composition of the lithospheric mantle are obvious from geochemistry and structure studies of xenoliths. Partial melting may lead to water-depletion within the lithospheric mantle, and this may result in a local strength increase (e.g., Hirth and Kohlsted, 1995a,b).

Mechanical anisotropy of the lithospheric mantle is suggested from (1) the plasticity anisotropy of olivine single crystal and aggregates during experimental deformations and numerical simulations, (2) the observation of a strong crystallographic orientation of olivine and other mantle rock-forming minerals in mantle xenoliths and lherzolite massifs, and (3) widespread seismic anisotropy (shear wave splitting, Pn azimuthal anisotropy, P-residuals etc.) and electrical conductivity anisotropy (magnetotelluric soundings) which require a consistent tectonic fabric in the upper mantle over large areas. Mantle tectonic fabrics developed during orogenic events are probably retained over long periods provided that no subsequent process (deformation, plume upwelling etc.) deeply modifies the lithosphere. Mechanical anisotropy of the mantle is probably the main mechanism of the so-called 'tectonic inheritance' frequently invoked to explain reactivation of old structures and parallelism of successive deformations (either compressional or extensional) in continental domains. This may especially explain why continents break-up occurs parallel to ancient orogenic belts and account for many characteristics of ocean basin development.

Rheological heterogeneity and mechanical anisotropy have been addressed separately for the sake of simplicity. It is however obvious that their effects may combine during the deformation of continents. The evolution of the Baikal rift or the East African rift, for instance, probably integrates the rheological effect of the stiff Siberian or Tanzanian cratons and an anisotropy factor due to the tectonic fabric of the lithospheric mantle in the Mongolian and Mozambique belts, as suggested by shear wave splitting, results (Gao et al., 1997; Vauchez et al., in press). As a matter of fact, during a tectonic event, the deformation of the lithosphere is certainly the result of an interaction between the internal structure of the plate and the external forces applied at plate boundaries.

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