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Continental rifting parallel to ancient collisional belts: an effect of the mechanical anisotropy of the lithospheric mantle

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Abstract

Analysis of major rift systems suggests that the preexisting structure of the lithosphere is a key parameter in the rifting process. Rift propagation is not random, but tends to follow the trend of the orogenic fabric of the plates, systematically reactivating ancient lithospheric structures. Continental rifts often display a clear component of strike–slip deformation, in particular in the early rifting stage. Moreover, although the close temporal and spatial association between flood basalt eruption and continental breakup suggests that mantle plumes play an important role in the rifting process, there is a paradox between the pinpoint thermal and stress perturbation generated by an upwelling mantle plume and the planar geometry of rifts. These observations suggest that the deformation of the lithosphere, especially during rifting, is controlled by its preexisting structure. On the other hand, (1) the plasticity anisotropy of olivine single crystal and aggregates, (2) the strong crystallographic orientation of olivine observed in mantle xenoliths and lherzolite massifs, and (3) seismic anisotropy data, which require a tectonic fabric in the upper mantle coherent over large areas, suggest that preservation within the lithospheric mantle of a lattice preferred orientation (LPO) of olivine crystals may induce a large-scale mechanical anisotropy of the lithospheric mantle. We use a polycrystal plasticity model to investigate the effect of a preexisting mantle fabric on the continental breakup process. We assess the deformation of an anisotropic continental lithosphere in response to an axi-symmetric tensional stress field produced by an upwelling mantle plume by calculating the deformation of textured olivine polycrystals representative of the lithospheric mantle at different positions above a plume head. Model results show that a LPO-induced mechanical anisotropy of the lithospheric mantle may result in directional softening, leading to heterogeneous deformation. During continental rifting, this mechanical anisotropy may induce strain localisation in domains where extensional stress is oblique (30–55°) to the preexisting mantle fabric. This directional softening associated with olivine LPO frozen in the lithospheric mantle may also guide the propagation of the initial instability, that will follow the preexisting structural trend. The preexisting mantle fabric also controls the deformation regime, imposing a strong strike–slip shear component. A LPO-induced mechanical anisotropy may therefore explain the systematic reactivation of ancient collisional belts during rifting (structural inheritance), the plume–rift paradox, and the onset of transtension within continental rifts. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: upper mantle; deformation; anisotropy; rifting; olivine; petrofabrics; plate tectonics; reactivation; transform faults; mantle plumes; rheology; Atlantic Ocean

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1. Introduction

Continental rifting is a complex process that results from the application of a tensional stress field to a long-lived and, hence, already structured lithosphere. In the past years, discussion has centred on the nature and causes of this tensional stress field. Proposed models range from purely passive rifting, in which extension and lithospheric thinning result from tensional stresses arising from the force distribution at plate boundaries [1], to active rifting, in which tension develops in response to an ascending mantle plume [2]. The close temporal and spatial association between flood basalt eruption and continental breakup [3,4] suggests that mantle plumes play a fundamental role in the development of major oceanic basins. There is, however, a paradox between the pinpoint thermal and stress perturbation generated by an upwelling mantle plume and the planar geometry of rifts. In addition, extensional stress associated with the lithospheric uplift should be maximum above the plume apex [5]. Locations of initial plume impacts are usually poorly constrained (e.g. Greenland or the Central Atlantic [3]). Yet, continental breakup is generally offset from either the centre of the flood basalt province (e.g. Ethiopia and Parana [4]) or the seismic low velocity anomalies interpreted as recording the initial conduits of the Parana and Decan plumes [6–8]. These observations suggest the existence of additional extensional forces and/or an influence of the structure of the lithosphere.

Evidence for an active role of the structure of the lithosphere in the rifting process also comes from the observation that rift propagation is not random, but tends to follow the trend of the pre-existing orogenic fabric of the plates, systematically reactivating ancient lithospheric structures (e.g. the Atlantic and West Indian Oceans, and the Rio Grande, Baikal, East African and Northeast China rifts). The evolution of the Atlantic Ocean, for instance, is marked by three episodes of rapid propagation of the spreading ridge following the emplacement of large basaltic provinces. This led Hill [3] to propose that the rise of the Fernando de Noronha, Parana and Iceland plumes resulted in reinforcement of a previously sluggish tensional regime allowing the start of new spreading ridges. It is however striking that, in each case, rifting was not limited to the region above the plume head, but initiated more or less simultaneously along segments more than two thousands of km long [9], which follow almost exactly the trend of Hercynian, Pan-African and Caledonian belts (Fig. 1).

Yet, although this so-called structural inheritance has been noted for more than three decades [10,11], the physical basis for this phenomenon is still poorly understood. Large-scale intraplate rheological heterogeneities may lead to strain localisation and control the rift location. Crustal thickening within orogenic domains, for instance, may halve the total strength of the continental lithosphere [12]. If rifting starts shortly after an orogeny, thickened crust domains may act as lithospheric weaknesses that favour the localisation of the rupture process. However, such a process fails to explain the parallelism between rifts and orogenic belts when rifting postdates orogeny by several hundreds of million years (e.g. the South Atlantic), since this time lag allows the excess crustal thickness to be removed by erosion, isostatic compensation or gravitational collapse. Cratonic nuclei, on the other hand, represent stiff rheological heterogeneities within a continental plate; they may induce stress concentration and, hence, strain localisation at its boundaries [13,14]. Deflection of hot plume material by the cratonic root [15,16] may also weaken the pericratonic domains, leading to strain localisation. Yet, rheological heterogeneities fail to explain why rifts formed far away from any cratonic nuclei follow the trend of ancient orogenic belts (e.g. the South Atlantic rift and the East African rift north of the Tanzanian craton). They also cannot explain why continental rifts often display a component of strike–slip deformation, in particular during the early rifting stage (e.g. the East African [17], North Atlantic [18] and Baikal [19] rifts).

An alternative (and complementary) model to explain the structural inheritance was proposed by Vauchez et al. [20] based on the analysis of shear wave splitting data in the Appalachians and the Pyrenees. In this model, the source of the structural inheritance lies in a mechanical anisotropy

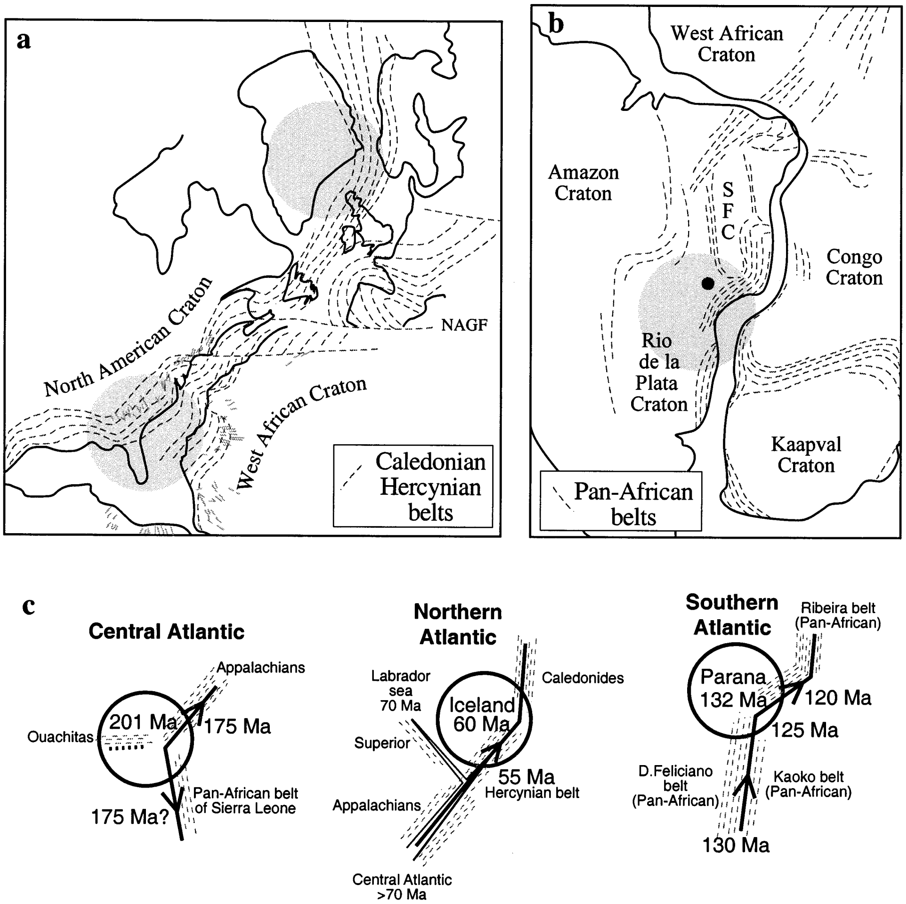

Fig. 1. Schematic map of (a) circum North Atlantic Caledonian–Hercynian belts and (b) circum South Atlantic Pan-African belts. Grey circles indicate likely locations (inferred from flood basalts distribution) of the major plumes associated with the Atlantic rifting [4]. Light grey lines in (a) mark the Central Atlantic giant radiating dike swarm [35]. Dark grey circle in (b) shows the location of the slow seismic anomaly interpreted by VanDecar et al. [6] and Schimmel et al. [7] as the remnant conduit of the Parana plume. NAGF: Newfoundland–Azores–Gilbraltar fault zone. (c) Schematic outline of the plume–preexisting structures (dashed grey lines) interaction in the Atlantic. Black arrows indicate the progression of oceanisation; ages for flood basalt events and initiation of oceanisation at different latitudes from [4].

of the lithospheric mantle due to the preservation, within the uppermost mantle, of a lattice preferred orientation (LPO) of olivine crystals formed during the main tectonic episode that shaped the plate. Olivine is the most abundant (60–70%) and the most deformable mineral in the upper mantle; it should therefore control the upper mantle rheology. Olivine displays a strong mechanical anisotropy. A crystal oriented to deform through ‘easy’ (010)[100] slip displays strain rates one order of magnitude higher than a crystal that deforms through ‘hard’ (010)[001] slip [21]. Analysis of upper mantle rocks brought to the surface by tectonic and magmatic events suggests that deformation by dislocation creep leads to the development of well-defined olivine LPO within the uppermost 200 km of the mantle [22]. Seismic anisotropy data, like SKS waves splitting [23] and Pn azimuthal anisotropy [24], suggest that these LPO are coherent over large distances ($>$ 50 km). If the preservation of well-developed olivine LPO within the lithospheric mantle also generates a mechanical anisotropy at these larger scales, this may result in a directional softening that will control strain localisation in the uppermost mantle and, hence, continental breakup.

2. Modeling the rifting of a structured lithosphere

Polycrystal plasticity models relate the aggregate strength or deformation to the slip systems' activity and, hence, to the orientation of the crystals. They allow therefore an investigation of the mechanical anisotropy induced by a preexisting LPO. These models are based on two assumptions: (i) the crystals that constitute the aggregate deform homogeneously by intracrystalline slip on selected crystallographic planes and (ii) the polycrystal (aggregate) behaviour may be calculated as an average of the grain responses. The lower and upper bounds are represented by the stress equilibrium [25] and *Taylor* [26] models that impose either homogeneous stress or strain within the aggregate. Intermediate solutions are obtained using tangent viscoplastic self-consistent (VPSC) models [27,28] or the more recent 'affine' [29] and variational [30] estimates. These models allow both microscopic stress and strain rate to differ from the corresponding macroscopic quantities. In the anisotropic VPSC model [28] used in the present calculations, strain compatibility and stress equilibrium are ensured at the aggregate scale, i.e. the volume averaged grain stresses and strain rates (s, $\dot{\varepsilon}$) are equal to the polycrystal stress and strain rate ($\bar{\Sigma}$, $\bar{D}$). Grain stresses and strain rates are related to the macroscopic quantities through:

$$\dot{\varepsilon}_{ij} - D_{ij} = -\tilde{M}_{ijkl}(s_{kl} - \bar{\Sigma}_{kl}) \quad (1)$$

Table 1
Slip systems data

Slip system	Critical resolved shear stress τ_0^a	Stress exponent n
(010)[100]	1	3.5
(001)[100]	1	3.5
(010)[001]	2	3.5
(100)[001]	3	3.5
{011}[100]	4	3.5
{031}[100]	4	3.5
{110}[001]	6	3.5

[a]Normalised relative to τ_0(010)[100].

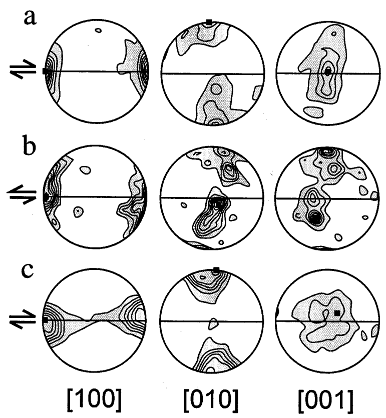

Fig. 2. Olivine LPO used in the models: (a) dunitic xenolith from Victoria, Australia (2197 measurements); (b) spinel–harzburgite from the Oman ophiolite (300 grains); (c) LPO calculated using the VPSC approach for a dextral simple shear ($\gamma = 2$, 1000 olivine grains, random initial LPO). Equal area projections, lower hemisphere. Contours at one multiple of a uniform distribution interval. Maximum density represented by a full square. Full line marks the shear plane, shear direction is horizontal.

where $\tilde{M}$ is the interaction tensor that depends on the rheological properties of the aggregate and on the grain shape.

For a given set of slip systems and an initial LPO, the VPSC model calculates the aggregate yield strength (or strain rate), the activity of the slip systems and the LPO evolution in response to an imposed macroscopic deformation (or stress) history. Under upper mantle conditions, olivine deforms essentially by slip on {0kl}[100] and {hk0}[001] systems, mainly (010)[100], (001)[100] and (010)[001] [31]. Critical resolved shear stresses and stress exponents for these slip systems (Table 1) were derived from power-law constitutive relations obtained for single crystals submitted to axial compression in various orientations under high temperature (1300–1500°C), high pressure (300 MPa) conditions [21]. Comparison of olivine LPO predicted by various polycrystal plasticity models with those measured in naturally and experimentally deformed peridotites suggests that the VPSC model with these input parameters offers a good approximation to the peridotites' deformation under upper mantle conditions [32].

To simulate the preexisting fabric of the lithospheric mantle we used: a random LPO, olivine LPO measured in two naturally deformed peridotites (a dunitic xenolith from Victoria, Australia and a harzburgite from the Oman ophiolite), and a LPO calculated using the VPSC approach for a dextral simple shear deformation (1000 olivine grains) [32]. These LPO patterns (Fig. 2) correspond to those most commonly observed in naturally deformed peridotites. In all models, the olivine LPO is oriented as if developed in a lithospheric scale vertical wrench or transpressional shear zone [33]. In this case, the shear plane is vertical and both the [100] and [010] maximum concentrations are horizontal, [100] being parallel to the flow direction (*X*) and [010] normal to the shear plane. This assumption is based on SKS splitting data on continental regions [23,34]: the fast shear wave polarisation plane, which contains the [100] axes maximum concentration, is usually parallel to the trend of orogenic belts, and the large delay times ($\geq$1 s) suggest that the almost vertically incident SKS waves propagate close to the [001] axes maximum concentration, i.e. within the shear plane, but at a large angle to the shear direction.

In order to estimate the effect of a preexisting mantle fabric on the rifting process, we investigate the deformation of a pre-structured mantle in response to an axi-symmetric tensional stress field, similar to the one expected to develop above an upwelling mantle plume (Fig. 3). Numerical models of the elastic deformation of the lithosphere above a mantle upwelling [5] show that both radial and tangential horizontal stresses are extensional and that, except for the region above the plume apex, tangential stresses are more extensional than radial ones. Such a stress state is expected to lead to radially oriented extensional fractures, like the radially fractured domes observed on Venus by the Magellan mission [5] or the giant radiating dike swarm that predates the central Atlantic opening [35]. The use of an axi-symmetric stress field has the advantage of not favouring any particular direction of extension; it allows us therefore to isolate the effect of a LPO-induced mechanical anisotropy on the deformation. Moreover, the results of such models may

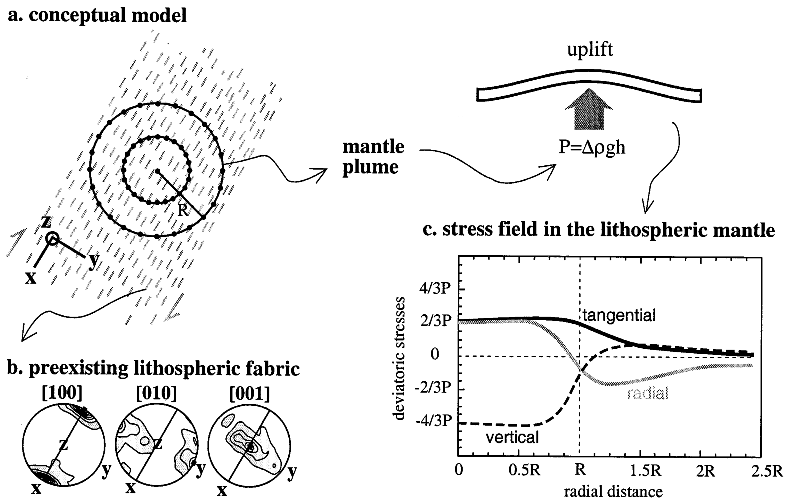

Fig. 3. (a) Outline of the conceptual model. Strain rates are calculated at points at fixed distances to the plume apex: just above it, at a distance of half of the plume head radius, and above the plume head periphery (black circles). (b) In each model series, all points display the same preexisting structure (dashed grey lines) and the olivine LPO is oriented as if developed in a lithospheric scale strike–slip fault; this preexisting structure defines the reference frame of the models: *X* is the shear direction, *Y* is the normal to the shear plane, and *Z* is vertical. (c) Deviatoric stresses induced in an elastic lithosphere by buoyancy forces associated with the upwelling, less dense plume material.

also be interpreted in terms of a two-dimensional tensional stress field (due to far-field, plate tectonics-related forces) applied at various orientations relative to the initial mantle fabric.

The ideal approach to study this problem would be a three-dimensional multi-scale deformation model in which an olivine polycrystal is associated with each grid point of the geodynamic model. However, such a three-dimensional model, in which the upper mantle viscosity is a tensorial entity that depends exponentially on both temperature and pressure, is beyond our present computational possibilities. Thus, we approach the effect of a preexisting mantle fabric on the continental breakup process by imposing an extensional stress field and calculating the instantaneous deformation (no LPO evolution) at different positions above the plume head: (i) above the apex, (ii) on a circumference of half of the plume head radius and (iii) on a circumference above the plume head periphery (Fig. 3). The same preexisting LPO is associated with all grid points. Three model series were run: one for each olivine LPO displayed in Fig. 2. The deformation of an isotropic aggregate submitted to the same stress field was also calculated; it is used as a reference. Lithospheric stresses induced by an upwelling plume were calculated using ADELI, a three-dimensional Lagrangian finite element code [36]. In this model (Fig. 3c), both the crust and the lithospheric mantle display an elastic behaviour and the upwelling plume is represented as an upward pressure imposed at the base of the lithosphere over a region of radius R.

3. Model results

Our models suggest that textured olivine polycrystals display a strongly anisotropic behaviour. The calculated deformation is anisotropic even for the region directly above the apex of the plume, which is characterised by an isotropic horizontal stress field (tangential and radial stresses are equal). Under these conditions, extensional strain rates are higher parallel to the dominant orientation of [100] axes, i.e. to the main glide direction for olivine (Fig. 4). However, strain rates displayed by textured polycrystals are significantly lower than those of an isotropic polycrystal submitted to the same stress field.

The effect of the preexisting texture is stronger at points located above the external half of the plume head, which are submitted to an anisotropic stress field (tangential stresses are higher than radial ones, Fig. 3c). Fig. 5a displays the variation in strain rate intensity (Von Mises equivalent strain rate $D_{eq} = (2/3 D_{ij} D_{ij})^{1/3}$ normalised by the strain rate of an isotropic aggregate submitted to the same stress field) as a function of the obliquity between the radial tension direction and the preexisting flow fabric for points located above the plume head periphery for the three models. Domains in which the tensional stress is

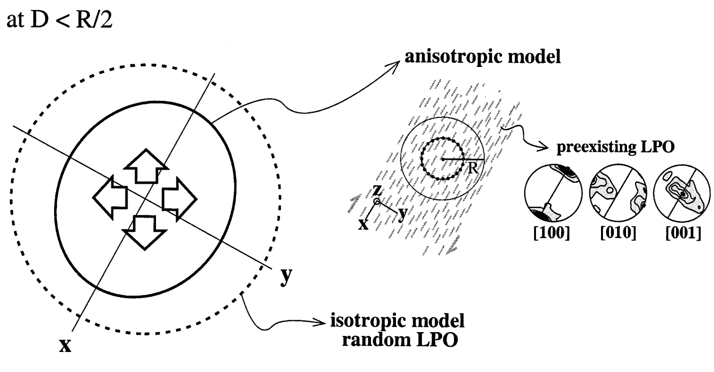

Fig. 4. Horizontal strain rate envelopes for points above the plume head apex in an isotropic model (dashed line) and in a model with a strong preexisting olivine LPO (full line). Insert displays the orientation of the LPO, which defines the reference frame in the anisotropic model.

oblique to the preexisting fabric deform up to six times faster than those domains in which the tensional stress is normal or parallel to the preexisting fabric. The former domains also deform up to three times faster than an isotropic lithosphere.

Qualitatively, all models with a non-random olivine LPO display similar results. The magnitude of anisotropy depends nevertheless on the intensity of the LPO, which may be characterised by the integral of the orientation distribution function, the *J*-index [37]. This dependence is not linear (Fig. 5a). A clear anisotropy is observed in the model with a weak, but well developed initial LPO (Australian xenolith, $J=5$), but this anisotropy is not doubled when initial LPO twice as strong are used, like the Oman peridotite LPO ($J=10.48$) or the modeled LPO ($J=9.29$). The higher anisotropy of the latter, in spite of the lower *J*-index, is due to its simpler LPO pattern characterised by single maxima distributions of both [100] and [010] axes.

Analysis of the relative magnitude of the different components of the strain rate tensor (in the *XYZ* structural frame) shows that the strain regime also varies as a function of the orientation of the tensional stress relative to the preexisting olivine LPO in the lithospheric mantle (Fig. 5b). Domains in which the tension is oblique to the initial LPO, i.e. those that display the higher strain rates, deform by transtension, i.e. a combination of extension normal to the preexisting structural trend (Dyy), vertical shortening (Dzz) and simple shear parallel to the preexisting structure (Dxy). Coaxial deformation is only observed in regions where the tensional stresses are either parallel or normal to the preexisting fabric. In that case, however, strain rates are significantly lower than when the tensional stresses are oblique to the preexisting fabric.

Finally, comparison of strain rate intensities displayed by a textured olivine polycrystal located

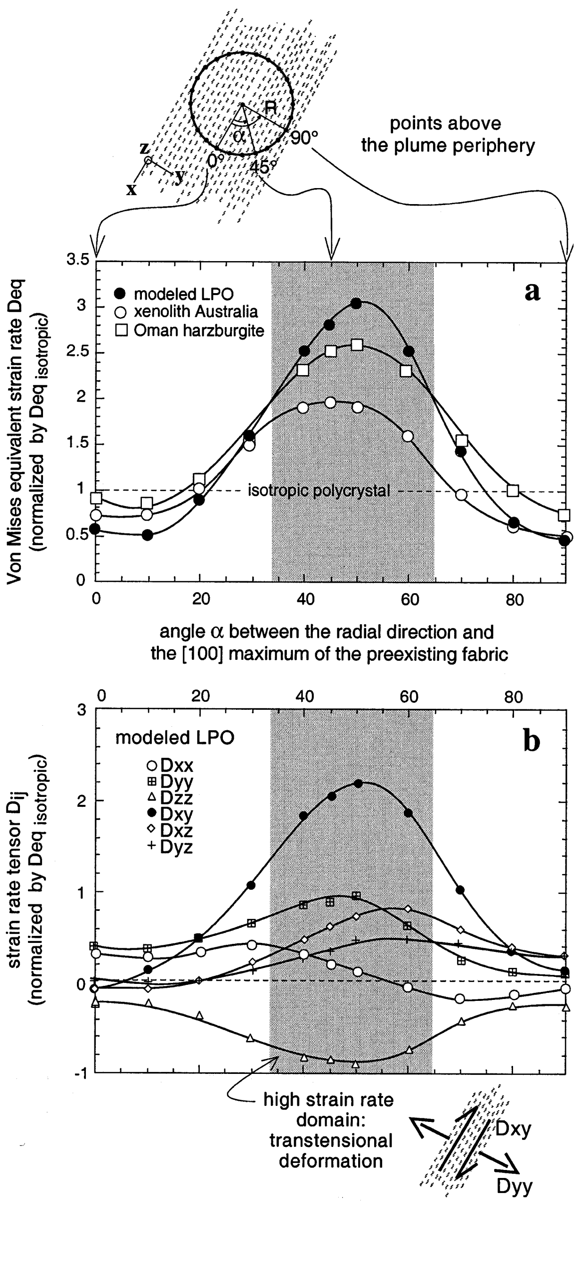

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Fig. 5. (a) Strain rate (Von Mises equivalent strain rate, normalised relative to the isotropic behaviour) as a function of the orientation of the radial tensional stress relative to the [100] axis maximum of the preexisting LPO for points above the plume head periphery. LPO for the three models are presented in Fig. 2. (b) Normal and shear components of the strain rate tensor (normalised by the Von Mises equivalent strain rate displayed by an isotropic polycrystal) for the model in which the initial LPO is the model aggregate (Fig. 2c). The reference frame is defined relative to the preexisting mantle fabric: *X* is parallel to the [100] axis maximum, i.e. parallel to the preexisting structural trend, *Y* is normal to the preexisting foliation and *Z* is vertical. Positive normal strain rates denote extension and negative ones, shortening. Grey region marks orientations that may trigger strain localisation.

above the plume periphery and those displayed by the same polycrystal above the apex reveals that the domains above the plume periphery in which extensional stresses are oblique to the preexisting fabric also deform faster than points directly above the plume conduit. These peripheral domains therefore display the fastest strain rates of the whole system.

4. Discussion

When extrapolated to the scale of lithospheric plates, these model results offer an explanation for the transition from a radial symmetry stress field to a planar deformation, the offset of rifts relative to the plume apex, the reactivation of ancient structures and the onset of transtension within continental rifts. The models show that a mantle plume rising beneath a structured continental plate induces an anisotropic deformation within the lithospheric mantle. Regions where the extensional stresses generated by the upwelling plume are oblique (30–55°) to the preexisting mantle fabric may experience significantly higher strain rates than surrounding domains (Fig. 5). This may result in strain localisation and development of local instabilities that are further amplified due to the non-linear rheology of upper mantle rocks under lithospheric conditions [38]. The directional softening associated with the olivine LPO frozen in the lithospheric mantle may also guide the propagation of the initial instability that will follow the preexisting structural trend. In addition, the initial mantle fabric may control the style of deformation, imposing a shear component parallel to the preexisting structures (transtensional deformation).

Although the models presented above only consider plume-related stresses, their predictions remain valid for an extensional stress field due to plate boundary forces. Except when the maximum extensional stress is parallel to the preexisting structural trend, deformation will be characterised by a reactivation of the ancient fabric. A far-field extensional stress oblique to the preexisting structural trend will add to the plume-related stresses and enhance the transtensional character of the deformation, whereas a far-field extensional stress normal to the structural trend will enforce the coaxial component of the deformation (extension).

The analysis of the rift systems that gave rise to the Atlantic Ocean (Fig. 1c) shows that, in all three examples, rifting is offset from the supposed location of the plume head apex (which is nevertheless poorly constrained for both the central Atlantic and Iceland plumes) and that the extensional deformation localised where the preexisting orogenic fabric is oblique to the plume-induced radial and tangential tensional stresses. Moreover, at least in the North Atlantic, there is clear evidence for an early transtensional deformation [18]. In contrast to the Central Atlantic, in the South and North Atlantic, initial rifting did not start at the plume periphery, but it progressed northwards from the Southwest Indian and Central Atlantic ridges, respectively. This suggests that extensional stresses associated with these preexisting oceanic spreading centres contributed to the continental breakup. Yet, in both cases, continental breakup occurred almost simultaneously over thousands of km shortly after the flood basalt emplacement and followed the preexisting lithospheric structure of the plate, evidenced by SKS splitting data in SE Brazil [39,40], eastern North America [41] and western Europe [42]. This suggests that the interaction of plume-related stresses with the preexisting structure of the lithosphere resulted in development of strain instabilities in domains where tensional stresses were oblique to the inherited mantle fabric. These instabilities worked as 'attractors' leading to an acceleration of the ridge development. For instance, the South Atlantic ridge propagated from the Agulhas basin up to the Parana flood basalt province in less than 5 Myr (Fig. 6); the first magnetic anomalies observed in this domain ranging from M11 to M4 [43,44].

Rift initiation offset from the centre of the flood basalt province as well as reactivation of the preexisting lithospheric structures in domains in which the tensional stress is oblique to those structures are also observed in the Southwest Indian ridge and in the Afars-East African rift system. Moreover, both rift systems underwent a sig-

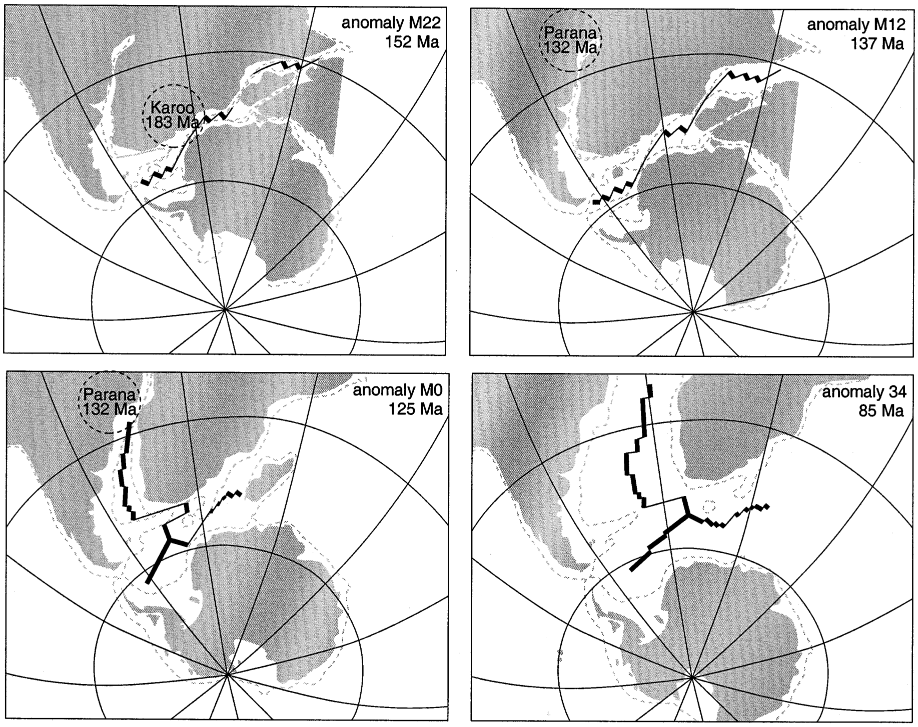

Fig. 6. Schematic reconstruction of the southern Gondwana dispersal (modified from [44]). Spreading centres and transform faults are shown as heavy and light black lines, respectively. Dashed lines mark the continental platforms. Ages for the Karoo and Parana flood basalt events from [4].

nificant component of motion parallel to the rift trend in their early stages. In the Southwest Indian Ocean, for instance, this early transtensional deformation is clearly evidenced by the length of active transform segments relative to the ridge segments at M22 time (Fig. 6). The Red Sea opening also involved a large amount of strike–slip deformation, as did the western branch of the East African rift [11], and, although the present day extension in the eastern branch is roughly normal to its trend, there is evidence for dextral transtension during the Miocene [45].

Some continental rifts did develop oblique or even at a high angle to the preexisting structural trends like the Rhine Graben in Western Europe or the Tucano-Jatoba Basin in Brazil. It is interesting to note, however, that none of these rifts evolved up to continental breakup and initiation of an oceanic basin. The Rhine Graben is a particularly interesting example since it was probably formed in response to far-field tectonic forces (regional seismic tomography studies do not image low velocity anomalies that might be assigned to a mantle plume beneath the graben [46]). The north–south orientation of the graben suggests that the maximum tensional stress was applied parallel to the Hercynian structural trend in this region (evidenced by surface geology and shear wave splitting data [47]). For such an orientation of the tensional stress relative to the preexisting mantle fabric, our models predict that reactivation is extremely difficult (this orientation corresponds to the lowest strain rates in Fig. 5b) and that extension will take place parallel to the preexisting structural trend. This should result in formation of a rift normal to the collisional belt trend, like the Rhine Graben.

Reactivation of a preexisting upper mantle fabric is not restricted to rifting episodes. Directional weakening resulting from frozen-up olivine LPO within the lithospheric mantle may also explain the long-lived nature of some major transform faults, which are systematically reactivated during both collisional and extensional episodes. An example of such a major transform is the Newfoundland–Azores–Gibraltar Fault zone (Fig. 1a). During the Hercynian orogeny, this fault acted as a dextral strike–slip boundary [48]. Subsequently, during the early Central Atlantic rifting, it acted as a major transform, which accommodated the differential motion between Africa and Europe. Indeed, the initial opening of the Central Atlantic ocean, in the mid- to late Jurassic, took place almost simultaneously from Florida to the Azores–Gibraltar transform (the first Central Atlantic magnetic anomaly, M25, is identified along this entire segment [43]), but further northward propagation of the Central Atlantic leading to separation between Eurasia and North America did not occur before late Cretaceous.

Analysis of these observations in the light of our model results points to a major role of the preexisting lithospheric structure in plate tectonics. On the other hand, recent studies [49,50] have shown that a 'self-lubricating' or strain-softening rheology is necessary to produce realistic plate tectonics in 3D convection models. We suggest that the preservation within the lithospheric mantle of a LPO of olivine crystals formed during the major tectonic episodes that shaped the plates (i.e. the preservation of a structural memory at the lithosphere scale) leads to a directional strain softening that may explain the perennial nature of plate boundaries and their systematic reactivation. This LPO-induced strain softening is particularly effective at strike–slip (transform) boundaries, which are precisely the most difficult to generate in a self-consistent way in convection models.

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