Strain localisation in the subcontinental mantle — a ductile alternative to the brittle mantle

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Abstract

It is now admitted that the high strength of the subcontinental uppermost mantle controls the first order strength of the lithosphere. An incipient narrow continental rift therefore requires an important weakening in the subcontinental mantle to promote lithosphere-scale strain localisation and subsequent continental break-up. Based on the classical rheological layering of the continental lithosphere, the origin of a lithospheric mantle shear/fault zone has been attributed to the existence of a brittle uppermost mantle. However, the lack of mantle earthquakes and the absence of field occurrences in the mantle fault zone led to the idea of a ductile-related weakening mechanism, instead of brittle-related, for the incipient mantle strain localisation. In order to provide evidence for this mechanism, we investigated the microstructures and lattice preferred orientations of mantle rocks in a kilometre-scale ductile strain gradient in the Ronda Peridotites (Betics cordillera, Spain). Two main features were shown: 1) grain size reduction by dynamic recrystallisation is found to be the only relevant weakening mechanism responsible for strain localisation and 2), with increasing strain, grain size reduction is coeval with both the scattering of orthopyroxene neoblasts and the decrease of the olivine fabric strength (LPO). These features allow us to propose that grain boundary sliding (GBS) partly accommodates dynamic recrystallisation and subsequent grain size reduction.

A new GBS-related experimental deformation mechanism, called dry-GBS creep, has been shown to accommodate grain size reduction during dynamic recrystallisation and to induce significant weakening at low temperatures \((T<800 \, ^\circ C)\). The present microstructural study demonstrates the occurrence of the grain size sensitive dry-GBS creep in natural continental peridotites and allows us to propose a new rheological model for the subcontinental mantle. During dynamic recrystallisation, the accommodation of grain size reduction by three competing deformation mechanisms, i.e., dislocation, diffusion and dry-GBS creeps, involves a grain size reduction controlled by the sole dislocation creep at high temperatures \((>800 \, ^\circ C)\), whereas dislocation creep and dry-GBS creep, are the accommodating mechanisms at low temperatures \((<800 \, ^\circ C)\). Consequently, weakening is very limited if the grain size reduction occurs at temperatures higher than 800 \(^\circ C\), whereas a large weakening is expected in lower temperatures. This large weakening related to GBS creep would occur at depths lower than 60 km and therefore provides an explanation for ductile strain localisation in the uppermost continental mantle, thus providing an alternative to the brittle mantle.

Keywords: Grain size; Dynamic recrystallisation; Grain boundary sliding; Ronda peridotites

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1. Introduction

For a few decades now, several examples of geological phenomena show that the lithosphere deformation is, at first order, controlled by the uppermost mantle strength with the best example certainly being the differences between wide and narrow continental rifts (Buck, 1991). The degree of the strain localisation at lithosphere-scale in a wide/narrow rift can be explained by the absence/presence of a high strength uppermost mantle, respectively (Hopper and Buck, 1993; Brun, 2001). Such narrow rifts that lead to lithosphere necking are accommodated by the development of kilometre-scale mantle shear zones/fault zones, as imaged in exhumed passive margins (Smythe et al., 1982; Brewer et al., 1983; Beach, 1986; Gibbs, 1987; Manatschal et al., 2001). The presence of a high strength uppermost mantle thus exerts a direct control on the development of mantle shear zones and subsequent lithosphere necking (Allemand and Brun, 1991; Brun and Beslier, 1996; Frederiksen and Braun 2001). However, the mechanisms responsible for strain localisation at mantle depth still remain a matter of debate.

The classical rheological layering of the continental lithosphere (Brace and Kohlstedt, 1980) predicts the existence of a brittle uppermost mantle, in agreement with the mantle earthquakes seen beneath Tibet in the early eighties (Molnar and Chen, 1983). The presence of a high strength brittle uppermost mantle thus exerts a direct control on the development of kilometre-scale mantle shear zones and subsequent lithosphere necking (Allemand and Brun, 1991; Brun and Beslier, 1996; Frederiksen and Braun 2001). However, several aspects are inconsistent with a brittle continental mantle: 1) a recent relocation in depth of deep earthquakes beneath Tibet has shown that they nucleated in the deep crust and not in the mantle (Maggi et al., 2000), which led Jackson (2002) to argue for a ductile lithospheric mantle; 2) brittle deformation has never been observed in exhumed mantle mylonites (e.g. deformed at depths of more than 30–40 km; Dijkstra et al., 2004); and 3) ductile shear zones (mylonites) are the only observed features indicating strain localisation, in both peridotite massifs (Drury et al., 1991) and xenoliths (Pike and Schwartzmann, 1977; Cabanes and Mercier, 1988; Downes, 2001; Tommasi et al., 2000). Kilometre-scale mantle structures imaged on seismic profiles across narrow continental rifts (Meier and Eisbacher, 1991) have been consistently interpreted as ductile shear zones rather than brittle fault zones (Brun et al., 1992; Goleby et al., 1988; Flack et al., 1990; Reston, 1990; Keen et al., 1991). Thus, there is a discrepancy between the mechanics of lithosphere extension, which require a high strength and localising uppermost mantle, and the geological and geophysical observations that contradict the existence of a brittle mantle. Nevertheless, a ductile-related mechanism of strain localisation might reconcile models with natural observations.

Ductile strain localisation requires a local weakening mechanism in order to reduce the material strength during deformation. Drury et al. (1991) and Dijkstra et al. (2004) have reviewed the possible weakening mechanisms acting in the lithosphere mantle: structural softening, thermal softening, melt-induced softening, geometric softening, fluids, syn-tectonic reactions softening and grain size-related softening. Among these various weakening mechanisms, syn-tectonic reaction Spinel/Plagioclase was evoked as the main weakening mechanism responsible for mantle strain localisation during lithosphere necking (Newman et al., 1999; Handy and Stüinitz, 2002; Dijkstra et al., 2004). This weakening well explains the peridotite mylonitisation during mantle unroofing from Spinel towards Plagioclase peridotite facies conditions. This Spinel/Plagioclase transition occurs at a depth of the order of 30 km for a classical lithosphere geotherm (MacDonough and Rudnick, 1998). This syn-tectonic reaction-weakening thus explains mantle exhumation during passive margin formation (Handy and Stüinitz, 2002) and hence supplies an explanation for strain localisation-type structures throughout the major part of European Plagioclase-bearing peridotite massifs as Lanzo (Boudier, 1978) or Voltri (Vissers et al., 1995), for example (Fig. 1). However, strain localisation acting at depths typical of the uppermost subcontinental mantle (30–60 kms; e.g. Peridotite spinel facies without metamorphic reaction) involves another weakening mechanism. In this case, grain size reduction by dynamic recrystallisation seems to be the best candidate as it could provide the largest amount of weakening with respect to the other weakening mechanisms (Drury et al., 1991). Furthermore, grain size reduction is the only feature that is systematically observed in exhumed mantle mylonites (Handy, 1989; Karato and Wu, 1993; Drury, 2005).

In this paper, we performed a microstructural study on the Ronda peridotite in order to infer the acting weakening mechanism at subcontinental mantle depth. The Ronda massif belongs to the internal zone of the Betics Cordillera (southern Spain) and shows a kilometre-scale strain gradient interpreted as a result of strain localisation (Balanya et al., 1993; 1997; Platt et al., 2003). The temperature conditions for the Ronda mylonite are similar to the other European peridotite massifs. Indeed, all strain localisation-type structures recorded “low” temperature conditions (around 850 °C, Fig. 1) with respect to their surrounding peridotites (around 1100 °C, Fig. 1). However, unlike the other massifs, the sheared
Ronda peridotites are characterised by grain size reduction within the spinel facies without major mineralogical changes, which are neither melting nor are a major occurrence of hydrated minerals (Obata, 1980; Van der Wal and Vissers, 1993). The Ronda example therefore provides an opportunity to study the grain size-related weakening (Drury et al., 1991) in order to understand the strain localisation process in the subcontinental mantle. This paper will first discuss the debated relationship between weakening and grain size reduction by dynamic recrystallisation in the subcontinental mantle. We will then present a microstructural study of the Ronda strain gradient that documents relevant mechanisms associated to strain localisation. Finally, a new rheological model for the continental mantle is proposed and supplies a ductile-related explanation for lithosphere-scale strain localisation.

2. Weakening and grain size reduction

2.1. Common dislocation/diffusion-related rheology

Grain size reduction by dynamic recrystallisation is classically attributed to two competing deformation mechanisms, dislocation creep ($\dot{\varepsilon}_{\text{disl}}$) and diffusion creep ($\dot{\varepsilon}_{\text{diff}}$) in the form of

$$\dot{\varepsilon}_{\text{disl}} = A_{\text{disl}} \cdot \exp \left( \frac{-Q_{\text{disl}}}{R \cdot T} \right) \cdot \tau^{n_{\text{disl}}}$$

$$\dot{\varepsilon}_{\text{diff}} = A_{\text{diff}} \cdot \exp \left( \frac{-Q_{\text{diff}}}{R \cdot T} \right) \cdot \tau^{n_{\text{diff}}} \cdot d^{-m_{\text{diff}}}$$

where, $\dot{\varepsilon}_{\text{disl}}$, $\dot{\varepsilon}_{\text{diff}}$, $\tau$, $d$, $m_{\text{diff}}$, $T$, $R$, $A_{\text{disl}}$, $A_{\text{diff}}$, $Q_{\text{disl}}$, $Q_{\text{diff}}$, $n_{\text{disl}}$ and $n_{\text{diff}}$ respectively are: the strain rates for...
dislocation creep and diffusion creep in $s^{-1}$, the stress in MPa, the grain size in $\mu m$, the grain size exponent for diffusion creep, the temperature in K, the gas constant and, also according to each deformation mechanism, the pre-exponential constant, the activation energy and the stress exponent. All values for these two flow laws are given in Table 1. The consideration of these two deformation mechanisms in the same rheological model leads to the definition of the overall strain rate $\dot{e}$ as the sum of the partial strain rates for each of the creeping mechanisms:

$$\dot{e} = \dot{e}_{\text{disl}} + \dot{e}_{\text{diff}}. \quad (3)$$

This relationship is used to construct an olivine deformation map that shows the shear stress as a function of grain size for different strain rates at constant temperatures ($850 \, ^\circ\text{C}$, Fig. 2a). In the deformation map, the dislocation creep field (GSI) represents stresses/grain sizes for which $\dot{e}_{\text{disl}} > \dot{e}_{\text{diff}}$. Therefore, the overall strain rate is the dislocation creep strain rate in the GSI field ($\dot{e} = \dot{e}_{\text{disl}}$). Similarly, $\dot{e}_{\text{disl}} < \dot{e}_{\text{diff}}$ prevailed in the GSS field and therefore $\dot{e} = \dot{e}_{\text{diff}}$. The transition line between these two fields is defined by an equal contribution of the two competing creeps to the overall strain rate $\dot{e}_{\text{disl}} = \dot{e}_{\text{diff}}$. Because dislocation creep is insensitive to grain size (Eq. (1)), strain rate isolines are horizontal in the GSI field and the slopes of the strain rate isolines in the GSS field are defined by the grain size exponent of the diffusion creep (Eq. (2)).

The dynamic recrystallisation at a constant strain rate is represented in the deformation map as a black arrow (Fig. 2a). Starting from an initial large grain size that induces dislocation creep as the dominant deformation mechanism (Fig. 2a, Kirby, 1983; Carter and Tseun, 1987), dynamic recrystallisation induces grain size reduction. Grain sizes tend towards the recrystallised grain size balance that can be defined in two different ways. First, experimental works on metals, ceramics and rocks have shown that the recrystallised grain size is related to the steady state shear stress by a power-law relation termed the piezometric relation,

$$\tau = B.D^{-p}, \quad (4)$$

where, D, $\tau$, p and B are the recrystallised grain size, the shear stress, the stress exponent and a material constant, respectively (values in Table 1). This relationship reflects the balance between the elastic energy stored in a grain and the surface energy (Twiss, 1977) and has been experimentally calibrated by Van der Wal et al. (1993) on natural olivines (VDW’s piezometer, curve (2), Fig. 2a). Second, Derby and Ashby (1987) have alternatively suggested that dynamic recrystallisation reflects a balance between grain-size reduction and grain growth (De Bresser et al., 1998, 2001). Because dislocation creep promotes grain size reduction by dynamic recrystallisation and diffusion creep can only enhance grain growth (Frost and Ashby, 1982; Handy, 1989; De Bresser et al., 1998, 2001), the recrystallised grain size should be in the neighbourhood of the boundary between dislocation creep and diffusion creep (De Bresser et al., 1998, 2001). As a result, by assuming that $\dot{e}_{\text{disl}} = \dot{e}_{\text{diff}}$ at the boundary (see above), the D-relationship for the balance hypothesis is therefore not only stress sensitive but also temperature sensitive (balance hypothesis line, curve (1), Fig. 2a) and induces significant differences in stress estimates with respect to the Van der Wal’s piezometer (curve (2), Fig. 2a). As a consequence, the amount of expected weakening during dynamic recrystallisation is drastically different depending on whether Van der Wal’s piezometer or the balance hypothesis is used to define the recrystallised grain size (Yamasaki, 2004). Following the balance hypothesis, grain size reduction is not related to a strong strength decrease (De Bresser et al., 2001; Gueydan et al., 2001) and hence cannot involve strain localisation (De Bresser et al., 2001; Ter Heege et al., 2004; Yamasaki, 2004). In contrast, if the recrystallised

Table 1
Creeping parameters for dislocation creep, diffusion creep and dry-GBS creep (Hirth and Kohlstedt, 2003) and parameters for the VDW’s piezometer (Van der Wal et al., 1993)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Meaning</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_{\text{disl}}$</td>
<td>Dislocation creep pre-exponential constant</td>
<td>1.1, $10^5$ MPa$^{-n}$.s$^{-1}$</td>
</tr>
<tr>
<td>$A_{\text{diff}}$</td>
<td>Diffusion creep pre-exponential constant</td>
<td>1.5, $10^9$ MPa$^{-n}$.s$^{-1}$</td>
</tr>
<tr>
<td>$A_{\text{GBS}}$</td>
<td>Dry-GBS creep pre-exponential constant</td>
<td>6.5, $10^3$ MPa$^{-n}$.s$^{-1}$</td>
</tr>
<tr>
<td>$Q_{\text{disl}}$</td>
<td>Activation energy for the dislocation creep</td>
<td>530 +/- 4 kJ. mol$^{-1}$</td>
</tr>
<tr>
<td>$Q_{\text{diff}}$</td>
<td>Activation energy for the diffusion creep</td>
<td>375 +/- 50 kJ. mol$^{-1}$</td>
</tr>
<tr>
<td>$Q_{\text{GBS}}$</td>
<td>Activation energy for the dry-GBS creep</td>
<td>400 kJ. mol$^{-1}$</td>
</tr>
<tr>
<td>$n_{\text{disl}}$</td>
<td>Stress exponent for the dislocation creep</td>
<td>3.5</td>
</tr>
<tr>
<td>$n_{\text{diff}}$</td>
<td>Stress exponent for the diffusion creep</td>
<td>1</td>
</tr>
<tr>
<td>$n_{\text{GBS}}$</td>
<td>Stress exponent for the dry-GBS creep</td>
<td>3.5</td>
</tr>
<tr>
<td>$m_{\text{disl}}$</td>
<td>Grain size exponent for the dislocation creep</td>
<td>3</td>
</tr>
<tr>
<td>$m_{\text{diff}}$</td>
<td>Grain size exponent for the diffusion creep</td>
<td>2</td>
</tr>
<tr>
<td>$m_{\text{GBS}}$</td>
<td>Grain size exponent for the dry-GBS creep</td>
<td>2</td>
</tr>
<tr>
<td>$p$</td>
<td>Stress exponent of D</td>
<td>1.33</td>
</tr>
<tr>
<td>$B$</td>
<td>Constant of D</td>
<td>1.5, $10^4$ $\mu$m.MPa$^{-n}$</td>
</tr>
<tr>
<td>$R$</td>
<td>Universal gas constant</td>
<td>8,314 J. mol$^{-1}$.K$^{-1}$</td>
</tr>
</tbody>
</table>

grain size is defined by the Van der Wal’s piezometer, an important stress decrease is predicted during grain size reduction that has to be achieved in the diffusion creep domain. Braun et al. (1999) consistently modelled the dynamic recrystallisation related to this rheology, i.e., based on the piezometric relationship, and they obtained strain localisation. However, the assumption of grain size reduction within the GSS diffusion creep is not supported by any published experimental results (Derby and Ashby, 1987; Drury et al., 1991).

2.2. GBS-related rheology

Recently, a new deformation mechanism was proposed to account for grain size reduction in a grain size sensitive creep (Hirth and Kohlstedt, 1995; 2003). This mechanism has been called dry-GBS creep by Hirth and Kohlstedt (2003), or GSS-power law creep by Drury (2005), and consists of grain boundary sliding (GBS) creep accommodated by dislocation and diffusion creeps (Hirth and Kohlstedt, 2003; Drury, 2005):

$$\dot{e}_{\text{GBS}} = A_{\text{GBS}} \cdot \exp \left( \frac{-Q_{\text{GBS}}}{RT} \right) \cdot \tau^{n_{\text{GBS}}} \cdot d^{-m_{\text{GBS}}},$$

where $\dot{e}_{\text{GBS}}$, $A_{\text{GBS}}$, $Q_{\text{GBS}}$, $n_{\text{GBS}}$ and $m_{\text{GBS}}$ are the strain rate, the pre-exponential constant, the activation energy, the stress exponent and the grain size exponent of the so-called dry-GBS creep mechanism, respectively. The strain rate is thus a low-power function of stress (non-Newtonian viscosity) and of grain size (GSS). The difference in the grain size exponent between this law ($m_{\text{GBS}}=2$) and the diffusion creep law ($m_{\text{diff}}=3$; Eq. (2)) marks different types of diffusion creep; namely coble creep (e.g. surface diffusion) and Nabarro–Herring creep (e.g. volume diffusion; Knipe, 1989; Wheeler, 1992), respectively. As a result, the combination of the three deformation mechanisms, i.e., dislocation creep, diffusion creep and dry-GBS creep yields the following definition of the overall strain rate:

$$\dot{\varepsilon} = \dot{\varepsilon}_{\text{disl}} + \dot{\varepsilon}_{\text{diff}} + \dot{\varepsilon}_{\text{GBS}}$$

The deformation map of Fig. 2b shows different slopes of the strain rate isolines in the dry-GBS creep and diffusion creep fields due to their different grain size exponents, and thus displays three deformation mechanism domains. Similar to the deformation map provided in Fig. 2a, far away from the transition line between each mechanism domain, the partial strain rate for the considered mechanism ($\dot{\varepsilon}_{\text{disl}}$, $\dot{\varepsilon}_{\text{diff}}$ or $\dot{\varepsilon}_{\text{GBS}}$) contributes as a whole to the overall strain rate $\dot{\varepsilon}$. In contrast, in the neighbourhood of the transition line, all partial strain rates contribute to the overall strain rate (Eq. (6)). At the exact transition line between dislocation creep and dry-GBS creep (GSI/GBS transition), dislocation and GBS strain
rates are consistently equal and much larger than the diffusion creep strain rate $\dot{e}_{\text{disl}} = \dot{e}_{\text{GBS}} > \dot{e}_{\text{diff}}$. The same reasoning for the two other transition lines yields the following relationships: $\dot{e}_{\text{disl}} = \dot{e}_{\text{diff}} > \dot{e}_{\text{GBS}}$ at the GSI/GSS transition line and $\dot{e}_{\text{diff}} = \dot{e}_{\text{GBS}} > \dot{e}_{\text{disl}}$ at the GBS/GSS transition line. At the junction of the three transition lines, each mechanism contributes to the overall strain rate in the same amount: $\dot{e}_{\text{disl}} = \dot{e}_{\text{diff}} = \dot{e}_{\text{GBS}} = 1/3 \dot{e}$.

Like the dislocation creep, the dry-GBS creep favours grain size reduction by intra-crystalline deformation and thus involves a change in the definition of the recrystallised grain size, according to the balance hypothesis of De Bresser et al. (1998; 2001). Indeed, if the recrystallised grain size well reflects the balance between grain-size reduction and grain growth (Derby and Ashby, 1987), the recrystallised grain size may be defined by the field boundary between dry-GBS and diffusion creeps at low grain sizes ($\dot{e}_{\text{diff}} = \dot{e}_{\text{GBS}}$) and by the field boundary between dislocation and diffusion creeps at large grain sizes ($\dot{e}_{\text{disl}} = \dot{e}_{\text{diff}}$). As recently discussed in Warren and Hirth (2006), a significant stress decrease during dynamic recrystallisation is predicted for low grain size while no weakenings are expected at large grain sizes, providing an explanation for weakening related to grain size reduction (Fig. 2b). Moreover, the dry-GBS creep provides an explanation for the grain size reduction accommodated by a GSS mechanism (Fig. 2a, curve 2), which plays an important role in the strain localisation process (Rutter and Brodie, 1988). Such an explanation is still lacking for the relationship based on the VDW’s piezometer that requires grain size reduction in the diffusion creep which is not supported by any experimental work.

In a recent study based mainly on olivine and orthopyroxene lattice preferred orientation (LPO) analyses from oceanic peridotites (Warren and Hirth, 2006), a similar weakening mechanism related to grain size reduction, grain boundary sliding and second phase pinning has already been proposed to explain the incipient strain localisation through oceanic peridotites. However, despite a strong prediction from its existence by Jin et al. (1998), dry-GBS creep has never been identified in continental peridotites.

3. Case study: the Ronda peridotites

3.1. Geological setting

The Ronda peridotites belong to the internal zone of the Betic Cordillera in southern Spain (Fig. 3). The cordillera is composed mainly of three stacked metamorphic nappe complexes (Tubía et al., 1992), which are (from bottom to top): the Nevada–Filabride, the Alpujárride and the Maláguide. The Ronda peridotites and its thin overlying crustal sequence (Jubrique unit; Balanya et al., 1997) belong to the upper part of the Alpujárride complex (“Los Reales” nappe, Tubía et al., 1992), which integrates three peridotite massifs: the Carraatraca massif, the Ojen massif and the Ronda massif, which is the largest one. A very similar peridotite massif is also found in the Moroccan side of the Alboran Sea in the Rif, i.e. the Beni–Bousera peridotite (Kornprobst, 1973, Fig. 1). Our study focused on the western part of the Ronda massif, which is composed of four tectonic domains (Obata, 1980; Van der Wal and Vissers, 1993) from south to north: Plagioclase (Pl)-bearing peridotite, granular peridotite, Spinel (Spl)-bearing tectonite and Garnet (Grt)/Spinel-bearing mylonite (Fig. 3; Fig. 4a).

The geodynamic history of the Ronda massif can be summarised as follows. The Ronda massif has suffered a first deformation stage marked by an almost adiabatic decompression (Frey et al., 1985) during the Jurassic Thetyan opening (Pearson et al., 1993; Van der Wal and Vissers, 1993), which unroofed the massif from high depths to a depth of approximately 70 km and produced graphitised diamonds (Davies et al., 1993). Then, during the cenozoic period, the Ronda massif underwent at least three tectonic events, as suggested by the P–T path of Obata (1980) and a consistent recent P–T path from Garrido et al. (2006; Fig. 4b): 1) a decompression event due to an extensive stage (Fig. 4b; point 1; Obata, 1980; Balanya et al., 1993; 1997) that induces the formation of the Spl-bearing tectonite and the Grt/Spl-bearing mylonite; 2) a rapid uplift of asthenosphere (Fig. 4b; point 2), which induces partial melting of the lithospheric mantle and produces the granular peridotite bounded by a recrystallisation front (Van der Wal and Bodinier, 1996; Lenoir et al., 2001); and finally 3) a compressive stage (Fig. 4b; point 3; Tubía et al., 2004) that leads to the emplacement of the peridotite domain into the crust and probably to the formation of the Pl-bearing peridotite (Van der Wal and Vissers, 1993; 1996). The peridotite massif and its country rocks are cross-cut by undeformed intrusions of acid dykes related to this compressive stage and dated at ca. 20 Myr (Priem et al., 1979). These acid dykes thus permit the dating of the end of the ductile deformation of the Ronda massif to approximately 20 Myr. The last exhumation of the peridotite lenses is attributed to a Miocene extensive stage responsible for the Alboran Sea opening (Balanya et al., 1997; Azañon and Crespo-Blanc, 2000; Booth-Rea et al., 2005). Presently, regional stresses involve a compression due to the convergence between European and African tectonic plates (Azañon and Crespo-Blanc, 2000).
3.2. Kilometre-scale ductile strain gradient

3.2.1. Controversy about the origin of the Grt/Spl mylonite

The first Cenozoic deformation stage is represented by a single extensive event, which develops the Spl-bearing tectonite and the Grt/Spl-bearing mylonite. From that view, both structures belong to a kilometre-scale strain gradient. However, Van der Wal and Vissers (1996) assumed a non-contemporaneous origin of the Spl-bearing tectonites and the Grt/Spl-bearing mylonites, as some garnets are observed around spinels in the mylonite. Indeed, this texture should indicate higher pressure for the mylonite than for the formation of the Spl-bearing tectonite (Obata, 1980) and should therefore involve a pressure increase during deformation. Nevertheless, the peculiar Grt/Spl texture in the mylonite could be explained by a variable chemical composition of the peridotite (Garrido et al., 2006). This chemical difference is assimilated in the strain-enhanced diffusion around the Al-pyroxenites (Al-pyroxenites group according to Garrido and Bodinier, 1999), which would modify the composition of the surrounding peridotite and would therefore change the position of the Grt/Spl transition to lower pressure into the P-T diagram. Furthermore, the occurrence of spinel in an eye-like garnet habitus within the mylonite and the stability of the tiny spinels confirms the strain-coeval transition from garnet to spinel stability fields. Therefore, a pressure increase is not necessary to explain the garnet occurrence in the mylonite, as shown by the P-T estimates for the pre- and syn-shearing conditions (Obata, 1980, Garrido et al., 2006). The PT path of mafic granulite lenses in the mylonite also consistently yields quasi adiabatic pressure decreases during deformation (Fig. 4, Morishita et al., 2001; Haissen et al., 2004). Consequently, we assume in this study that the Spl-bearing tectonite and the Grt/Spl-bearing mylonite were formed during a single progressive deformation. This assumption is moreover confirmed by the structural study presented below. Therefore, we will
use proto-mylonite instead of tectonite for the rest of this paper to describe the less deformed Spl-peridotites (Fig. 4a). Note that this term is only defined macroscopically and not according to the Sibson’s classification (Sibson, 1977).

### 3.2.2. The Ronda strain gradient

The north-western part of the massif shows a progressive strain increase from the Spl-bearing proto-mylonite towards the Grt/Spl-bearing mylonite. These two tectonic domains (Fig. 4a; Pictures 1–3) have approximately the same lherzolitic and harzburgitic composition with few dunites, Cr-pyroxenites and Al-pyroxenites, which mark the pyroxenite “layering” (Fig. 4a; Pictures 4-6; Van der Wal and Bodinier, 1996). The foliation strikes of the proto-mylonite and the mylonite (about N50°) are roughly parallel to the contact with the country rocks, with a mean dip angle increasing from 50° near the northern edge of the massif to almost 90° near the recrystallisation front (Fig. 4a; Van der Wal and Bodinier, 1996). Through this domain, the lineation is constant and the orientation of the pyroxenite layering, with respect to the foliation, is a good indicator of the strain intensity. Indeed, in the proto-mylonite, pyroxenite layers often form metric folds with axial planes parallel to the peridotite foliation (Fig. 4a; Picture 6), whereas boudinage is common in the mylonite. The rotation of the pyroxenites during strain involves the continuous evolution from a compressive state (folding) to an extensive state (boudinage), as suggested by the boudined limbs of the foliation.
the metric folds. This evolution, from folding to stretching of the pyroxenite layers, is consistent with a northward progressive increase of the finite strain intensity and hence, the existence of a ductile strain gradient. In the interest of being concise, a detailed study of the Ronda peridotite will be presented in a separate paper as the main objective here is to discuss the rheology of the mantle.

Temperature conditions through this strain gradient can be roughly estimated from the neoblasts which yielded a temperature close to 830–880 °C (Van der Wal and Vissers, 1993). Thus, the strain gradient developed at low temperatures, around 850 °C, as observed through other strain gradients in European continental mantle peridotites (Fig. 1). Moreover, the existence of a spatial temperature gradient during strain localisation is also inferred from the temperature conditions in the top of the peridotites, calculated in the Kinzigites (730±50 °C; Platt et al., 2003, Figs. 3–4) and in the core of the peridotites at the recrystallisation front (≥ 1200 °C; Lenoir et al., 2001). However, the real spatial evolution of the temperature from the mylonite (“cold” member) to the proto-mylonite close to the recrystallisation front (“hot” member), is very difficult to constrain. P–T paths computed in mafic granulite lenses in mylonite (Fig. 4, Morishita et al., 2001; Haissen et al., 2004) and the Grt/Spl mylonite (Obata, 1980; Garrido et al., 2006) yield a continuous cooling history during strain. This ductile strain gradient has therefore been formed according to a spatial temperature gradient between ~750 °C at the top and ~1200 °C in the core of the massif and strain localisation occurs in the coldest part of the massif, where the temperature is less than ~850 °C during deformation.

3.3. Microstructural study

3.3.1. Microstructures

A microstructural study was performed on 40 thin sections from samples taken across the strain gradient. The mineral composition is roughly constant: 65 to 75% olivine (forsterite), 15 to 23% orthopyroxene (enstatite), 2 to 7% clinopyroxene (diopside) and 3 to 6% spinel (Fig. 5). Some other phases, like amphibole and calcite, can also be observed in very small amounts. The proto-mylonite consists of centimetre-scale grains with lense-
shaped olivine clasts showing strong internal strain (undulose extinction and subgrains). In contrast, the mylonite displays grain sizes close to 140 μm near the contact with the country rocks and contains fewer porphyroclasts (Fig. 5). Thus, grain size reduction is the main microstructural feature observed with increasing strain. The mean value of the recrystallised grain size is between 250 μm and 450 μm in the proto-mylonite (grey box, Fig. 6) and is reduced to 150–220 μm in the mylonite. A scattering in the recrystallised grain size values is also observed. The recrystallised grain size values range between 800 and 100 μm in the proto-mylonite (dashed lines, Fig. 6); while in the mylonite, recrystallised grain sizes vary between 50 μm and 600 μm. Therefore, both recrystallised grain size averages and the scattering in grain sizes decrease drastically with increasing strain. Following Gueydan et al. (2005), these two combined features likely reflect an increase in shear stress and hence an increase in strain rate, marking strain localisation.

There is also another textural feature discriminating between the mylonite and the proto-mylonite. Pyroxene porphyroclasts and neoblasts of the proto-mylonite are clustered, forming centimetre-scale aggregates surrounded by olivine aggregates. In contrast, the pyroxene neoblasts in the mylonite (mainly the orthopyroxenes) are much more abundant and scattered into the olivine matrix (Fig. 5), suggesting a neighbour-switching of pyroxene-recrystallised grains with increasing strain. This feature might possibly be accompanied by a change in the intra-crystalline deformation mode, which could be detected by a change in the lattice preferred orientation (Boulier and Gueguen, 1975; Jaroslow et al., 1996; Lee et al., 2002).

3.3.2. Lattice preferred orientation (LPO)

EBSD (Electron BackScattered Diffraction) was used to measure olivine LPO (Fig. 7a). We measured the orientation of the three main olivine crystallographic axes ([100], [010] and [001]) with respect to the main strain axes. Post-experiment numerical handling was then performed in order to illustrate the fabric strength by using iso-density curves.

Fabrics are displayed in Fig. 7a. They are similar to the type A from Jung and Karato (2001) as well as to the LPO reported by Vauchez and Garrido (2001) and Tubía (1994). An easy [a]-slip system can be observed with a weak [c]-slip system, involving moderate to high strain temperatures (around 1000 °C; Tommasi et al., 2000). Of more interest, is the change in the LPO intensity along the strain gradient (Fig. 7b). Indeed, the fabric strength first increases in the proto-mylonite and then decreases in the mylonite until it reaches the edge of the massif. This specific feature is well marked on the [010]-axis. Total randomisation in the mylonite fabric strength is not observed, not even in the highest strained peridotite near the contact with the country rocks. This feature is also

![Graph showing olivine recrystallised grain size as a function of the distance to the contact of peridotites with the crustal rocks. Around 25 measurements per thin section are reported as well as the mean recrystallised grain size marked by the black squares. The mean recrystallised grain size decreases from 450–250 μm in the proto-mylonite to 220–150 μm in the mylonite (grey boxes).](Image)
observed in the Shaka oceanic peridotites (Warren and Hirth, 2006).

4. Discussion

4.1. Inferred mechanisms of strain localisation in the Ronda peridotites

Based on the above observations, the relevant weakening mechanism in the Ronda strain gradient is the dynamic reduction of grain size that is coeval with a decrease in LPO intensity and a neighbour-switching of pyroxene neoblasts, leading to their scattering within the fine-grained olivine matrix. The increase in LPO intensity in the proto-mylonite is well explained by the increase in the intra-crystalline plastic deformation. In contrast, the decrease of the LPO intensity in the mylonite might result from at least three mechanisms: 1) the increase in the amount of dynamic recrystallisation with an associated increase in disorientation; 2) the activation of different slip systems with increasing strain; or 3) the activation of grain size-sensitive deformation mechanisms that would compete with intracrystalline slips (Warren and Hirth, 2006). However, the neighbour-switching of the recrystallised grains through the mylonite tectonic domain strongly suggests that grain boundary sliding plays an important role during mylonitisation (Vissers et al., 1995; Paterson, 2001; Lee et al., 2002; Warren and Hirth, 2006) and hence, the LPO decrease is related to a grain-size-sensitive mechanism. As a result, we infer from these microstructural observations that the weakening mechanism responsible for strain localisation in the Ronda peridotites is grain size reduction accommodated by both dislocation creep and grain boundary sliding, as described in the dry-GBS creep (Hirth and Kohlstedt, 1995; 2003).

4.2. New rheological model: weakening and grain boundary sliding

To account for the main features reported in the Ronda peridotites, it is necessary to combine three deformation mechanisms for olivine, i.e., dislocation creep, diffusion creep and dry-GBS creep, in the same rheological model, as seen in the deformation map provided in Fig. 2b. Since most of the subcontinental mantle peridotites show “low temperature” conditions associated with strain localisation-type structures (Fig. 1), it is more convenient here to fix the strain rate and to discuss the temperature
conditions necessary for weakening during grain size reduction. As a result, Fig. 8 shows three olivine deformation maps using a constant strain rate ($10^{-15} \text{s}^{-1}$, $10^{-14} \text{s}^{-1}$ and $10^{-13} \text{s}^{-1}$, respectively in Fig. 8a, b and c). Temperature isolines are displayed from 650 °C to 950 °C in increments of 50 °C.

Following Derby and Ashby (1987) and De Bresser et al. (1998; 2001), dynamic recrystallisation reduces the grain size that tends towards the balance between the domains of grain size reduction (dry-GBS and dislocation creeps) and grain growth (diffusion creep). The recrystallised grain sizes are thus defined by the dislocation/diffusion boundary at high temperatures while they are defined by the dry-GBS/diffusion boundary at low temperatures (see the detailed discussion in the previous section). This modified balance hypothesis is moreover confirmed by the extrapolation of the experimentally-obtained recrystallised grain size at geological conditions ($\dot{\varepsilon} \approx 1.10^{-14} \text{s}^{-1}$; Drury, 2005). Starting from a large grain size close to 4 mm, dynamic grain size reduction at $1.10^{-15} \text{s}^{-1}$ will be successively accommodated by dislocation creep and dry-GBS creep at low temperature.

Fig. 8. Olivine deformation maps (shear stress versus grain size) at a constant strain rate, i.e. $10^{-15} \text{s}^{-1}$ (a), $10^{-14} \text{s}^{-1}$ (b) and $10^{-13} \text{s}^{-1}$ (c), with temperature isolines from 950 °C to 650 °C in increments of 50 °C. The new rheological model proposed in this paper consists of defining the overall strain rate as the sum of the partial strain rates for dislocation creep (Eq. (1)), diffusion creep (Eq. (2)) and dry-GBS creep (Eq. (4)). Due to its grain size insensitivity (GSI), the dislocation creep field displays horizontal temperature isolines. In contrast, the grain size sensitivity (GSS) of the diffusion and the dry-GBS creeps show positive slopes for temperature isolines in their respective domains and the difference in slope between the two deformation mechanisms reflects the different grain size exponents (Table 1). An initial grain size of 4 mm and temperatures between 730 and 880 °C for the Ronda strain gradient (Platt et al., 2003; Van der Wal and Vissers, 1993) were chosen to discuss dynamic recrystallisation (black arrows). The dynamic recrystallised grain size is defined by the modified “balance” hypothesis according to De Bresser et al. (1998; 2001). Each graph shows an example of grain size reduction at a low temperature (≈750 °C) and high temperature (≈870 °C) in order to represent the GBS-related weakening undergone by the recrystallised grains. As a result, even for a relatively low overall strain rate ($10^{-15} \text{s}^{-1}$), the low temperature involves a strong weakening with respect to the weakening produced at high temperatures and hence defines a range of temperature conditions, according to the overall strain rate, in which strong weakening can occur. Furthermore, this rheological model can be validated in part by comparing the predicted recrystallised grain sizes by the model with the natural recrystallised grain sizes of the Ronda strain gradient (grey rectangle on the abscise axis, cf. Fig. 6). See the text for explanation.
(Fig. 8a; thick arrow) and only by dislocation creep at high temperatures (Fig. 8a; thin arrow). A critical temperature \( T_{GBS} \) can therefore be defined; below which dry-GBS occurs as a dominant mechanism and allows the promotion of large weakenings according to the temperature. For temperatures lower than \( T_{GBS} \), the amount of weakening, e.g. stress decrease, during dynamic grain size reduction increases with decreasing temperature (Fig. 8a). This critical temperature \( T_{GBS} \) increases with an increasing strain rate (Fig. 8). \( T_{GBS} \) is 780 °C, 818 °C and 858 °C at \( 10^{-15} \text{s}^{-1} \), \( 10^{-14} \text{s}^{-1} \) and \( 10^{-13} \text{s}^{-1} \), respectively (Fig. 8a, b and c). As a consequence, the amount of weakening at a given temperature (750 °C, thick arrow) increases with an increasing strain rate (Fig. 8b and c). The weakening during the dynamic recrystallisation defined as one minus the ratio of stress after dynamic recrystallisation over initial stress at 750 °C, is around 50% at \( 10^{-15} \text{s}^{-1} \) (Fig. 8a), 65% at \( 10^{-14} \text{s}^{-1} \) (Fig. 8b) and 80% at \( 10^{-13} \text{s}^{-1} \) (Fig. 8c). This positive feedback between weakening and the local increase in strain rate is furthermore a prerequisite to accurately model strain localisation in the subcontinental mantle.

**4.3. Application to the Ronda peridotites**

In order to validate the new rheological model for the subcontinental mantle, we compared the estimates of the recrystallised grain size from the deformation maps (Fig. 8) with the average of the measured recrystallised grain size in the Ronda ductile strain gradient (the grey boxes in Fig. 6). The Ronda ductile strain gradient is formed within a spatial thermal gradient with a temperature of approximately 730 °C near the contact with the crustal rocks and around 850–900 °C in the mylonite/proto-mylonite contact (see the discussion in the previous section). Predicted recrystallised grain size \( D \) for temperatures ranging from 730 °C and 880 °C are reported in Fig. 9 as a function of the overall strain rate. Recrystallised grain sizes at the dislocation/diffusion boundary are plotted as white boxes, while recrystallised grain sizes at the dry-GBS/diffusion boundary are plotted as grey boxes. As explained earlier, dynamic recrystallisation accommodated by dislocation creep and dry-GBS creep will induce significant weakening. As a consequence, the \( D \) values in the grey boxes are most likely related to significant weakening during grain size reduction, while the \( D \) values in the white boxes are not related to weakening. The measured values of recrystallised grain sizes in Ronda are shown as dashed black boxes and black boxes for the proto-mylonite and mylonite, respectively. Based on the structural evidence for the existence of a ductile strain gradient, the proto-mylonite \( D \)-values are plotted with respect to the lower strain rate values, while the mylonite \( D \)-values are plotted with respect to the higher strain rate values. The \( D \)-values for the proto-mylonite are mostly consistent with predicted recrystallised grain sizes at the dislocation/diffusion boundary, which likely indicate high temperatures during deformation. In contrast, mylonitic \( D \)-values...
are almost entirely consistent with predicted recrystallised grain sizes at the dry-GBS/diffusion boundary, and thus, low temperatures during deformation. This comparison is fully consistent with 1) the presence of a spatial thermal gradient during deformation with mylonite being colder than proto-mylonite and 2) microstructural evidence that

**Fig. 10. Schematic drawing of lithosphere necking illustrating the ductile alternative to the mantle as the origin of a lithosphere-scale strain localisation.**

**a) Classical rheological layering: Brittle faulting**

![Diagram of classical rheological layering showing the brittle and ductile transition (BDT) at different temperatures.]

**b) Proposed model: Ductile faulting**

![Diagram showing the ductile localising rheology.]

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a) Mantle localising rheology defined after the brittle mantle inherent to the classical rheological layering (Brace and Kohlstedt, 1980). The brittle-ductile transition (BDT) is defined according to a steady-state lithosphere geotherm and standard rheological values presented in Tables 1 and 2. b) Mantle localising rheology defined after the GBS-related rheology proposed in this paper. The depth of the boundary between the no GBS- (e.g. no localising) and the GBS-related rheology (e.g. localising) is defined by temperatures close to 800 °C ($T_{\text{GBS}}$) and is denoted LMB (Localising Mantle Boundary).
show that dislocation creep is the main deformation mechanism in the proto-mylonite while the dry-GBS creep was active in the mylonite. Moreover, this comparison gives a rheological explanation for the Ronda ductile strain gradient. Due to the spatial thermal gradient that characterised the Ronda peridotites, deformation at mantle conditions leads to grain size reduction with no weakening for peridotites at temperatures higher than 800 °C and leads to grain size reduction and weakening for peridotites at lower temperatures. This difference in weakening could likely explain the structural differences between the Grt/Spl-bearing mylonites that are GBS-related structures and the Spl-bearing proto-mylonite that are dislocation-related structures. This new rheological model also predicts a decrease in the scattering of the D-value with an increasing strain rate (Fig. 9), and thus with increasing strain, as observed in the field (Fig. 6).

This good level of consistency between the rheological model prediction and the natural data exemplifies the key role played by the dry-GBS creep in the mantle rheology as well as in explaining the ductile strain localisation in the uppermost continental mantle.

4.4. Ductile alternative to the brittle mantle

The GBS-related rheology presented in this paper could possibly explain the formation of ductile shear zones in the subcontinental mantle as follows. Dry-GBS creep is activated during dynamic recrystallisation only in temperatures lower than $T_{\text{GBS}}$ which varies between 780 °C at $10^{-15}\text{ s}^{-1}$ and 858 °C at $10^{-13}\text{ s}^{-1}$ (Fig. 8). On the scale of the lithosphere, the temperature gradient required to define a temperature-dependent weakening domain by GBS-related grain size reduction is simply the lithosphere geotherm (Fig. 10, and thermal parameters in Table 2). Based on a mantle heat flux of 30 mW.m$^{-2}$, a 30 km thick crust and an overall strain rate of $5.10^{-15}\text{ s}^{-1}$, the first 35 km of the mantle are at a temperature lower than $T_{\text{GBS}}$. This critical $T_{\text{GBS}}$ temperature is approximately 800 °C for a strain rate of $5.10^{-15}\text{ s}^{-1}$. The GBS-related rheology therefore defines a ductile localising rheology in the uppermost high strength lithosphere mantle. This new rheological model moreover proposes a strain weakening mechanism often attributed as being at the origin of the continental necking (Govers and Wortel, 1993; 1995). During the unroofing of the continental mantle, as soon as the mantle rocks reach temperatures lower than $\approx$ 800 °C, weakening occurs during grain size reduction accommodated by dry-GBS creep and could lead to the onset of strain localisation. The subsequent increase in strain rate leads to an increase in weakening that again enhances the strain rate increase. This positive feed back between ductile weakening and the strain rate increase will favour strain localisation in the uppermost mantle. Mantle shear zones can therefore develop in a purely ductile uppermost mantle. Nevertheless, this process of strain localisation on the scale of the lithosphere is clearly transient and requires numerical modelling to better document interactions between cooling, grain size reduction, weakening and strain localisation.

This rheological model combining dislocation creep, diffusion creep and dry-GBS creep, thus defines a new rheological layering of the continental lithosphere: classical brittle/ductile crust overlying a purely ductile mantle with an uppermost ductile localising mantle (GSI and GBS creeping mechanisms) above a lower ductile lithosphere mantle (GSI and GSS creeping mechanisms). Thus, this new rheological layering of the lithosphere is different from the common idle of a brittle/ductile lithosphere mantle (Fig. 10a). Brittle–ductile transition in the mantle depends on the material used to model the mantle rheology (i.e. dry versus wet olivine, Karato and Wu, 1993). Based on dry Olivine dislocation creep parameters (Hirth and Kohlstedt, 2003), brittle–ductile transition in the mantle occurs at 700 °C and at a depth of 55 km, yielding a 25 km thick brittle mantle (Fig. 10a). The temperature of the brittle–ductile transition $T_{\text{BDT}}$ is therefore much lower than the critical temperature of our model $T_{\text{GBS}}$, below which ductile weakening occurs. As a consequence, the thickness of the ductile localising mantle is clearly larger than the thickness of the brittle mantle. The impact of such differences in thickness of ductile/brittle localising mantle could not be inferred through simple strength profiles and instead requires numerical simulations. Despite these differences between a classical brittle/ductile mantle and our new ductile

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Meaning</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_{\text{crust}}$</td>
<td>Density of the crust</td>
<td>2800 kg. m$^{-3}$</td>
</tr>
<tr>
<td>$\rho_{\text{mantle}}$</td>
<td>Density of the mantle</td>
<td>3300 kg. m$^{-3}$</td>
</tr>
<tr>
<td>$k_c$</td>
<td>Thermal conductivity of the crust</td>
<td>2.1 W. m$^{-1}.K^{-1}$</td>
</tr>
<tr>
<td>$k_m$</td>
<td>Thermal conductivity of the mantle</td>
<td>3.0 W. m$^{-1}.K^{-1}$</td>
</tr>
<tr>
<td>$q_m$</td>
<td>Basal heat flux</td>
<td>0.03 W. m$^{-2}$</td>
</tr>
<tr>
<td>$C$</td>
<td>Thermal capacity</td>
<td>1000 J</td>
</tr>
<tr>
<td>$r$</td>
<td>Radioactive heat production</td>
<td>30 km</td>
</tr>
<tr>
<td>$H_c$</td>
<td>Thickness of the crust</td>
<td>1.10$^{-6}$ W. m$^{-3}$</td>
</tr>
<tr>
<td>$H_r$</td>
<td>Thickness of the radioactive crust</td>
<td>8 km</td>
</tr>
<tr>
<td>$\mu$</td>
<td>Sliding coefficient</td>
<td>0.6</td>
</tr>
<tr>
<td>$g$</td>
<td>Gravitational acceleration</td>
<td>9.81 m .s$^{-2}$</td>
</tr>
<tr>
<td>$\varepsilon_0$</td>
<td>Overall strain rate</td>
<td>5.10$^{-15}$s$^{-1}$</td>
</tr>
</tbody>
</table>

localising mantle, most of the mechanical properties of the uppermost mantle are similar. Indeed, the new rheological layering of the lithosphere mantle predicts a high strength uppermost mantle that can localise strain. These two features are moreover required to accurately model the mechanics of the lithosphere extension from narrow rifting and subsequent lithosphere necking to wide rifting (see the Introduction for this discussion). Finally, this new ductile rheology of the mantle is in agreement with most geological and geophysical observations that contradict the existence of a brittle mantle.

In summary, taking into account the dry-GBS creep to model mantle rheology is crucial and permits the definition of a ductile high strength localising uppermost mantle instead of a brittle mantle. This new rheological model also permits the reconciliation between the geological/geophysical observation and the mechanics of lithosphere deformation.

5. Conclusion

In order to propose a ductile weakening mechanism for the incipient strain localisation through the uppermost continental mantle, a microstructural study was performed in a kilometre-scale ductile strain gradient in the Ronda peridotites (Betics cordillera, Spain) and shows that ductile strain localisation is coeval with:

1) grain size reduction by dynamic recrystallisation.
2) the neighbour-switching of recrystallised orthopyroxenes within fine-grained olivine matrix.
3) a decrease of the olivine fabric strength (LPO) in the more strained samples.

These results demonstrate the impact of the grain boundary sliding (GBS) during dynamic recrystallisation. Following Hirth and Kohlstedt (2003) and Drury (2005), we propose a new rheological model for the continental mantle that combines dislocation creep, diffusion creep and dry-GBS creep, a deformation mechanism that assimilates the impact of the grain boundary sliding accommodated by dislocation and diffusion creeps. Therefore, this new rheological model for the continental lithosphere mantle predicts:

1) a significant weakening during dynamic recrystallisation at low mantle temperature ($T<800 \, ^\circ C$)
2) an increase of the weakening with decreasing temperature and increasing strain rate.

Thus, this new rheological model provides an explanation for ductile strain localisation in the uppermost subcontinental mantle and hence defines a ductile alternative to the brittle mantle as the origin of the strain localisation process. Nevertheless, this rheology cannot explain large weakening at very low temperatures (500 °C) due to the infinite diffusion, thus suggesting another peculiar rheology for the low temperature uppermost mantle, perhaps strongly related to metamorphic reactions (Newman et al., 1999; Dijkstra et al., 2004).

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