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Controls of stable continental lithospheric thickness: the role of basal drag

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Abstract

The lithosphere is the earth's strong outer layer that remains a coherent unit for long times during plate motions. The motions lead to shearing of its hot and ductile lower part. The base of the lithosphere is placed where the shearing is large enough to effectively decouple the coherent plates from the underlying mantle. Here the depth and temperature where this occurs under stable continental areas were modeled by using the experimentally determined creep properties of olivine (which depend on temperature, pressure and fluid content) for different geotherms and different magnitudes of the basal drag. It is found that the shearing increases rapidly downward in a ca. 20–25 km thick decoupling zone where the bottom of the lithospherecan can be placed. This zone is deeper for cooler geotherms and for smaller a basal drag, while the presence of fluids makes it shallower. In dry rock this zone is at the mantle adiabat or cooler when the basal drag is >0.5 MPa. Under Archean cratons whose temperature is represented by xenolith geotherms the decoupling zone is 180–240 km deep. Higher crustal heat generation than in cratons (common in post-Archean terranes) by itself causes the decoupling zone to be a few tens of kilometers shallower, and it will be even shallower if the mantle heat flow is larger than under cratons. The base of the lithosphere need not follow an isotherm. As the lithospheric thickness depends on several factors it will change when any of these changes, which is expected to have happened many times in earth history. To persist the lithosphere should escape being sheared off laterally, regardless of its buoyancy.
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Keywords: Continental lithosphere; Lithospheric thickness; Changes of lithosphere

1. Introduction

The term lithosphere, referring to the earth's mechanically strong outer shell, was introduced about a century ago in the context of studies of isostasy (Barrell, 1914). Later this term was incorporated into plate tectonic theory to designate the earth's outer cold and strong plates that slide over the underlying asthenosphere (e.g. Isacks et al., 1968; Le Pichon, 1968). The lithosphere extends down to where the rocks are sufficiently hot and ductile to allow significant motion relative to deeper parts of the mantle. As this is not a sharply defined level it was placed in different ways, most often where some thermal condition is fulfilled, e.g. where the geotherm intersects the mantle adiabat, but other criteria were also used (Sleep, 2005; Anderson, 1995).

In the plate tectonic context the simplest approach is to view the lithosphere as comprising the layer that remains a coherent unit while participating in plate motions. Under continents the extent and history of this layer are constrained by data on mantle-derived xenoliths and xenocrysts. These reveal compositional variations in the uppermost mantle which extend to depths of 150 km to >200 km and correlate with the tectono-thermal age of the

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overlying crust (Griffin et al., 1999, 2003; Gaul et al., 2000). This indicates the longevity of the compositional domains and their long-term coherence with the crust. Their longevity is further proven by the ages of major events of melt extraction from mantle rocks and the ages of inclusions in diamonds (Richardson et al., 1985; Pearson, 1999a, 1999b; Carlson et al., 2005), and by the evolved Sr and Nd isotope compositions of mantle xenoliths from Archean terranes that indicate long periods of isolation (Menzies, 1990). Correlation with heat flow variations further shows that the uppermost mantle under Archean cratons is colder and intrinsically less dense than under younger areas (Jordan, 1988; Poudjom Djomani et al., 2001). Data on the geoid confirm the lateral extent of these differences (Doin et al., 1996). It is noteworthy that mantle xenoliths from Archean cratons equilibrated at conditions that approximate steady state conductive geotherms down to some 200 km (Finnerty and Boyd, 1987; Rudnick and Nyblade, 1999). Such a thermal regime can develop in coherently moving plates, but not in the convecting mantle where a much lower temperature gradient close to adiabatic is expected. In addition, the variations of seismic velocities under continents, though not always sufficiently well resolved, also reveal lateral differences which correlate with the stabilization age of the crust down to depths reaching ca. 250 km and can be correlated with the compositional and temperature variations of the continental lithospheric mantle, whereas at greater depths the velocities mostly tend to become rather uniform (Anderson and Polet, 1995; Van der Lee and Nolet, 1997; Simons et al., 1999; Ritsema and van Heijst, 2000; Goes et al., 2000; Röhm et al., 2000; Goes and van der Lee, 2002; Poupinet et al., 2003; Godey et al., 2004; Ritsema et al., 2004; Yoshizawa and Kennett, 2004).

Taken together these data are best interpreted as showing that the uppermost mantle beneath stable continental areas down to depths that may reach 200–250 km remained attached to the overlying crust and escaped the homogenizing effects of convection for long times, most likely since stabilization of the overlying crust. This reveals the minimum thickness of the lithosphere, but does not define its base. Sclater et al. (1980), following Parsons and McKenzie's (1978) study of oceanic lithosphere, placed the base of continental lithosphere within a thermal boundary layer, several tens of kilometers thick, in which mantle flow begins, which was often followed (e.g. Sleep, 2005). Here the purpose is to examine the thickness of stable continental lithosphere using a somewhat different approach. The basic point is that motions of plates will produce a viscous drag that shears their base, so where the shearing deformation becomes large enough the lithosphere decouples from the convecting mantle (Fig. 1; Bokelmann and Silver, 2002; Garfunkel, 2004). This approach has the advantage of allowing one to use the actual properties of the lithosphere in different places for evaluating its response to the basal drag. In what follows this concept is applied to the lithosphere of stable continental areas.

2. The model

For the present purpose, consider a vertical cross-section through the lithosphere in the direction of plate motion relative to the underlying mantle (Fig. 1). The basal drag will cause a horizontal simple shear at a rate

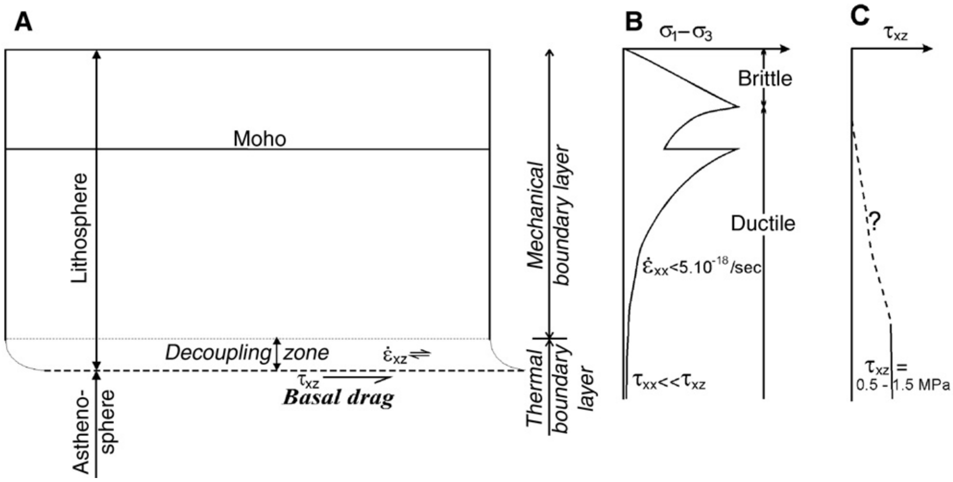

Fig. 1. General features of continental lithosphere. A. Deformation of the lithosphere due to basal drag. B. Schematic behavior stressdifference; not to scale. C. Schematic behavior of shear stress; not to scale.

of $\dot{\varepsilon}_{xz}(z)$ which increases downward because the downward temperature rise will cause the rocks to flow more readily. The velocity $v_x(z)$ of the deeper parts of the lithosphere relative to its upper surface will increase downward, and it is given by

$$v_x(z) = \int_0^z 2\dot{\varepsilon}_{xz}(z')d(z') \qquad (1)$$

($\dot{\varepsilon}_{xz}(z) = 1/2(\partial v_x/\partial z + \partial v_z/\partial x)$, but in the lithosphere $v_z(x)$ and its x derivative are effectively zero, so they do not contribute to $\dot{\varepsilon}_{xz}(z)$ and to the integral (1)). Here the base of the lithosphere is taken to be at depth l where $v_x(l)$ attains some critical value, e.g. 2 mm/y (=200 km/100 Ma), but later it will be seen that the choice of this particular value is not crucial. At such velocities material at $z>l$ is effectively no longer attached to the shallow part of the lithosphere after several hundred Ma, as it moves sideways >1000 km, which is the order of the size of many continental plates. To evaluate the integral (1) one needs to know $\dot{\varepsilon}_{xz}(z)$, which in turn depends on (a) the geotherm, (b) the rheology and (c) the state of stress in the lower part of the lithosphere. In what follows these three factors in stable continental areas are examined first, and then the implications for $v_x(z)$ are examined and discussed.

2.1. Geotherms

The temperature distribution in the lithosphere is of crucial importance in the present context because it strongly influences rock deformation. The important point is that under stable continental areas, not affected by tectonothermal events during a few hundred million years, the thermal regime is close to a conductive steady state. This allows calculating geotherms using standard methods if one knows: (a) the measurable heat flow at the surface q_0 (averaged over sufficiently large areas), (b) the amount and distribution of heat generation in the crust q_{cr}, and (c) the thermal conductivity k or the diffusivity $\kappa(=k\,\rho^{-1}\,Cp^{-1})$, which are determined experimentally. This is the basis of many studies of continental geotherms (Pollack and Chapman, 1977; Sclater et al., 1980; Morgan, 1984; Chapman, 1986; Nyblade and Pollack, 1993; Rudnick et al., 1998; Jaupart and Mareschal, 1999; Artemieva and Mooney, 2001). The value of q_0 is found to vary considerably, and often, but not always, it tends to be higher as the stabilization age of the crust decreases. Though the value of q_{cr} can only be estimated as it cannot be directly determined, it appears that its variability accounts for much of the variability of q_0. The uncertainty in q_{cr} carries over to an uncertainly in the value of $q_m(=q_0-q_{cr})$, the heat flow in the lithospheric mantle. Therefore, in practice only a range of geotherms can be estimated depending on assumptions regarding q_{cr}, but in some cases other information helps constraining the geotherms, e.g. thermo-barometry of mantle xenoliths or seismic data.

Some uncertaintly arises also because the value of k (or κ) in the mantle is not known sufficiently well. This is dominated by the properties of olivine, the main constituent of the upper mantle. Schatz and Simmons (1972) found that between 500 and 1200 °C and at low pressure k of polycrystalline Fo90 is in the range of 4–3.5 W m^{-1} °K^{-1}. Katsura (1995) studied polycrystalline Fo89 and his results show that at mantle conditions k varies from ca. 3 W m^{-1} °K^{-1} at ca. 50 km depth (T=450–500 °C) to ca. 3.5 W m^{-1} K^{-1} or slightly more at depths of 150–200 km (T=1000–1200 °C). More recent data (Gibert et al., 2003a) raise the possibility that k of olivine is several tens of percent higher than previously suggested. In the following calculations a value of k=3.5 W m^{-1} °K^{-1} will be used.

Seismic data show that continental lithosphere is often anisotropic to various degrees due to preferred lattice orientation of olivine crystals, which is also characteristic of mantle xenoliths (Silver, 1996; Fouch et al., 2000). This implies that k is also anisotropic and may very by up to 15–ca. 30% in different directions (Gibert et al., 2003b), but often will be less. Such effects were neglected, because it is not clear that they follow some systematic pattern. The resulting uncertainty is similar to the uncertainty in the absolute value of k. It is similar to or exceeds influence of other minerals (Clauser and Heunges, 1995; Gibert et al., 2003b).

Heat generation in the lithospheric mantle is neglected. Rudnick et al. (1998) favored a heat generation of 0.019 μW m^{-3} under Archean cratons, which corresponds to ca. 0.03 wt.% K_2O. This is close to the K content of primitive mantle (McDonough and Sun, 1995). However, since the Archean lithosphere underwent extensive melting (Griffin et al., 1999, 2003; Gaul et al., 2000; Carlson et al., 2005) it must have lost most of the heat generating elements, and replenishment of the entire loss is difficult to envisage. Thus even 0.01 μW m^{-3} is probably an overestimate for the regional average heat generation in Archean lithosphere, though it may apply to younger lithosphere. The resulting influence on the geotherms would be ca. 30 °C to ca. 45 °C over depth ranges of 150 km to 180 km, respectively, which is neglected.

2.2. Rheology (creep law)

Since only the lower part of the lithosphere is significantly deformed by the basal drag (see below), the rheology of only this region is considered here. This is dominated by the behavior of olivine — the main constituent of the uppermost mantle. Other minerals are less common, so they do not influence much the bulk behavior. Thus, though the composition of the lithosphere varies (Griffin et al., 1999, 2003; Carlson et al., 2005), the effects on rheology are probably small and cannot be modeled at this stage. Laboratory experiments and the textures of mantle xenoliths show that in the lithospheric mantle olivine deforms by dislocation creep, which results in a power law dependence of the strain rate on the stress and produces a preferred lattice orientation, making the deformed rocks anisotropic (Kirby and Kronenberg, 1987; Karato and Wu, 1993; Kohlstedt et al., 1995; Drury and Fitz Gerald, 1998; Hirth and Kohlstedt, 2004). When the deformation is incompressible, the strain rate tensor $\dot{\varepsilon}_{ij}$ is given by (Nye, 1957; Ranalli, 1995):

$$\dot{\varepsilon}_{ij} = A\exp\left(-\frac{E^* + PV^*}{RT}\right)\tau^{n-1}\tau_{ij} \tag{2}$$

where A is a material constant (depends on oxygen and water fugacity), E^* and V^* are the activation energy and activation volume, respectively, P is the pressure, T is the absolute temperature, R is the gas constant, τ_{ij} is the deviatoric stress tensor, and τ is the square root of its second invariant, and is given by:

$$\tau^2 = 1/2(\tau_{11}^2 + \tau_{22}^2 + \tau_{33}^2) + \tau_{12}^2 + \tau_{13}^2 + \tau_{23}^2 \tag{3}$$

This shows that any component of the strain rate tensor depends on the overall stress state, i.e. on all the components of the deviatoric stress tensor, so relation (2) will have different forms depending on the state of stress.

Let σ_i be the principal (non-deviatoric) stresses, oriented along the coordinate axes. Then in uniaxial compression $\sigma_1 > \sigma_2 = \sigma_3$, so Eqs. (2) and (3) show that

$$\dot{\varepsilon}_1 = \frac{2A\mathbf{E}}{3^{(n+1)/2}}(\sigma_1 - \sigma_3)^n \text{ and } \dot{\varepsilon}_2 = \dot{\varepsilon}_3 = -\dot{\varepsilon}/2 \tag{4}$$

where $\dot{\varepsilon}_i$ are the principal strain rates and $\mathbf{E}$ is the exponential term of Eq. (2). Similarly, when $\sigma_1 > \sigma_2 = (\sigma_1 + \sigma_3)/2 > \sigma_3$ (plane strain– pure shear in xz plane)

$$\dot{\varepsilon}_1 = \frac{A\mathbf{E}}{2^n}(\sigma_1 - \sigma_3)^n \text{ and } \varepsilon_2 = 0,\ \varepsilon_3 = -\varepsilon_1 \tag{5}$$

In plane strain $\sigma_2 = (\sigma_1 + \sigma_3)/2$, and when the principal stresses σ_1 and σ_3 are oriented at 45° to the x and z axes, then $\tau_{xz} = (\sigma_1 - \sigma_3)/2$ and $\tau_{xx} = 0$, so the deformation is simple shear in the xz plane. Then

$$\dot{\varepsilon}_{xz} = A\mathbf{E}\tau_{xz}^n = \frac{A\mathbf{E}}{2^n}(\sigma_1 - \sigma_3)^n \text{ and } \dot{\varepsilon}_{xx} = \dot{\varepsilon}_{yy} = \dot{\varepsilon}_{zz} = 0 \tag{6}$$

If σ_1 and σ_3 are in the xz plane but at an angle $\theta \neq 45°$ to the x and z axes, then $\tau_{xz} = ½(\sigma_1 - \sigma_3)\sin 2\theta \neq 0$, $\tau_{xx} = ½(\sigma_1 - \sigma_3)\cos 2\theta \neq 0$. In this case

$$\dot{\varepsilon}_{xz} = A\mathbf{E}\tau_{xz}\frac{(\sigma_1 - \sigma_3)^{n-1}}{2^{n-1}}\sin 2\theta = \frac{A\mathbf{E}}{2^n}(\sigma_1 - \sigma_3)^n \sin 2\theta \text{ and} \tag{7a}$$

$$\dot{\varepsilon}_{xx} = \dot{\varepsilon}_{xz}/\tan 2\theta \tag{7b}$$

The parameters in these equations appropriate for olivine were generally determined on the basis of uniaxial compression experiments, so the results are described in terms of Eq. (4). Chopra and Paterson (1984) obtained for dry coarse grained natural peridotite a value of 2.88×10^4 MPa^{-n} s^{-1} (to within a factor of 1.5) for the pre-exponential term, $n = 3.6 \pm 0.2$, and $E^* = 535 \pm 33$ kJ/mol, while Hirth and Kohlstedt (1996) inferred a value of 4.85×10^4 MPa^{-n} s^{-1} for the pre-exponential term in (4), $n = 3.5 \pm 0.5$, and $E^* = 535$ kJ/mol. Other studies obtained somewhat different values (Kirby and Kronenberg, 1987; Drury and Fitz Gerald, 1998). Hirth and Kohlstedt (2004), based on precise deformation studies of synthetic fine grained olivine aggregates, preferred 1.1×10^5 MPa^{-n} s^{-1} for the pre-exponential factor, $n = 3.5$, and $E^* = 530$ kJ/mol. These values will be used here. Eq. (4) shows that to apply these results to other stress systems the numeric factor in Eq. (4) must be taken into account. Other sets of values of these parameters were proposed, but they predict similar strain rates because the values of the individual parameters are correlated, as they are not measured separately but inferred from the same experiments (Hirth, 2002).

The presence of water considerably increases the deformation rate of olivine, so for "wet" conditions, i.e. in the presence of little water, the strain rates at given conditions (stress, T, P) may be 10 times or more faster than of dry rocks at the same conditions (Hirth and Kohlstedt, 1996, 2004), and thus it can significantly affect the mechanical behavior of the lithosphere (Jackson, 2002; Dixon et al., 2004; Garfunkel, 2004).

A major uncertainty regarding the creep rate concerns its pressure dependence. Estimated values of V^* gave values of 6 to 15×10^{-6} m^3/mol and even higher (Kohlstedt et al., 1995; Drury and Fitz Gerald, 1998). Hirth and Kohlstedt (2004) concluded that V^* decreases with rising pressure, so in the upper mantle it may be 6×10^{-6} m^3/mol or less. Béjina et al. (1999) found that $V^*\leq10\times10^{-6}$ m^3/mol, and preferred a value of ca. 6×10^{-6} m^3/mol. Because of the uncertainty several values will be examined.

At low stresses and when the grain size is small, the deformation is dominated by diffusion creep which results in a linear stress-strain rate dependence and does not produce a lattice preferred orientation (Karato and Wu, 1993; Kohlstedt et al., 1995; Drury and Fitz Gerald, 1998). It was suggested (e.g. Karato and Wu, 1993) that this mechanism becomes dominant at the base of the lithosphere. However, dislocation creep most probably persists to pressures of 8 GPa, equivalent to ca. 240 km depth (Li et al., 2003). Evidence for pronounced seismic anisotropy in the continental lithosphere and also at depths of 200–400 km (Silver, 1996; Gung et al., 2003; Simons and van der Hilst, 2003) also points at the operation of dislocation creep at such depths, because anisotropy is not produced by diffusion creep. Therefore dislocation creep is assumed in the subsequent modeling, though the possibility of transition to diffusion creep needs to be kept in mind.

2.3. State of stress in lower lithosphere

Eqs. (2) and (3) show, as noted above, that deformation of the lower lithosphere by dislocation creep depends on the entire stress system. This is constrained by several lines of evidence. The world stress map (Zoback, 1992) shows that in stable continental areas the maximum compressive stress is horizontal. It tends to be close to the direction of absolute plate motions because plate driving forces and plate boundary forces are the major sources of intra-plate stresses (Solomon et al., 1975; Zoback, 1992; Richardson, 1992). Intra-plate deviatoric extension occurs where rifting and uplifting take place (Zoback, 1992), but since in geologic history such events were transient and limited in space and time this is not considered here.

The magnitude of the stress difference ($\sigma_1-\sigma_3$) is limited by rock strength (Brace and Kohlstedt, 1980). In the shallow brittle levels frictional sliding on existing fractures is the limiting factor, but in the deeper parts of the lithosphere which are examined here, rocks yield by ductile creep so $\sigma_1-\sigma_3$ is determined by the long-term strain rate which is considered to be depth independent. It is often assumed that $\dot{\varepsilon}_{xx}=10^{-14}$ s^{-1} or 10^{-15} s^{-1} (shortening) which means $\dot{\varepsilon}_{xx}=0.3$ to 0.03 Ma^{-1}. Such strain rates will reduce the width of any region to less than half its original width within a few Ma to <15 Ma, and the other dimensions will have to change accordingly to preserve volume. There is no evidence for such deformation in tectonically stable areas, and it would not allow long-term preservation of the lithosphere. The long-term strain rate in stable areas is difficult to resolve, but a value of $\dot{\varepsilon}_{xx}\approx0.1$ Ga^{-1}, i.e. 3×10^{-18} s^{-1} is a likely upper bound and may well be an overestimate because even such deformation has not been documented in Precambrian shields. Zoback et al. (2002) inferred that the plate driving forces could induce at most similar strain rates, but in stable Precambrian lithosphere the strain rate should be much smaller. Eqs. (2) and (6) show that two-dimensional pure shear of an olivine-dominated rock at such a rate requires that τ_{xx} should decrease downwards from tens of MPa at mid-lithospheric depths to ≤1 MPa and ≤0.3 MPa where $T=1000$ °C and 1200 °C, respectively, depending on the geotherm (with $V^*=6\times10^{-6}$ m^3 mol^{-1} and other parameters according to Hirth and Kohlstedt, 2004), and even to lower values if water is present in significant amounts.

At the surface and at shallow depths in tectonically stable areas the horizontal shear stress (τ_{xz}) vanishes, because there is no horizontal shearing. This implies that one principal stress axis is vertical and the other two are horizontal. At greater depths, however, the viscous drag arising when the lithosphere slides over the asthenosphere will produce a horizontal shear stress τ_{xz} in the direction of plate motion (Fig. 1). It is generally considered to oppose the plate motion (Forsyth and Uyeda, 1975; Zoback, 1992). Hager and O'Connell (1981) estimated its magnitude to be about 0.5 MPa, but a range of values is expected as the drag most likely depends on plate velocity which varies by a factor of 5 or more. However, Bokelman (2002) raised the possibility that the basal drag can also act in the direction of plate motions. This will not affect the following calculations, because only the magnitude of the shearing and not its direction is considered.

As will be seen below, where $v_x(z)=2$ mm/y (see Eq. (1)) τ_{xz} is about 10 times greater than τ_{xx} that is required to produce $\dot{\varepsilon}_{xx}=3\times10^{-18}$ s^{-1}, so the deformation is close to simple horizontal shear, and according to Eqs. (7a) and (7b) the principal stress axes are inclined at an angle close to 45° to the horizontal. The change from the shallow levels where τ_{xx} is effectively horizontal occurs where τ_{xx} becomes small compared with τ_{xz}, but this occurs at depths where the shear deformation is so small that its contribution to $v_x(z)$ is negligible (see below), so it does not influence the following modeling. The very small to negligible deformation of stable continents implies that $\dot{\varepsilon}_{yy}$ is also very small, so it hardly contributes to the deformation. Thus deformation of the lower part of the lithosphere is very well approximated by simple shear in a vertical plane, taken to be the *xz* plane, which is parallel to the local direction of plate motion relative to the underlying mantle (Fig. 1). This assumes that shearing due to plate motions dominates. Small-scale irregularities in the flow of the topmost asthenosphere are possible; such complications are beyond the scope of this work which focuses on the overall effect of the plate motions which is considered to be dominant, but it will have to be considered in more advanced modeling.

Based on data from tectonically active areas (e.g. Himalaya) Jackson (2002) inferred that the crust, including its lower part, contributes more than the topmost mantle to lithospheric strength, which differs from the often accepted interpretation (shown in Fig. 1). This would not change the present analysis of areas where at these depths deformation is negligible over periods of 1–2 Ga and more, though it would become relevant at strain rates of the order of 10^{-15}/s that he considers.

3. The shearing of the base of the lithosphere

The above considerations allow calculating $v_x(z)$ that results from basal drag in various situations. For convenience Archean cratons and Proterozoic shields are considered separately.

3.1. Shearing at the base of Archean cratons

In Archean cratons $q_0 = 41 \pm 11$ mW m^{-2}, the variations reflecting mainly differences in q_{cr} (Nyblade and Pollack, 1993; Rudnick et al., 1998; Jaupart and Mareschal, 1999; Artemieva and Mooney, 2001). The value of $q_m (= q_0 - q_{cr})$ was estimated in the range from 11–15 mW m^{-2} to 20 mW m^{-2}, but part of this range appears to reflect the uncertainty in q_{cr}. Thus a range of geotherms with different temperature gradients in the mantle can be inferred. The P–T data from mantle xenoliths in kimberlites (Finnerty and Boyd, 1987; Kopylova et al., 1999; Kukkonen and Peltonen, 1999; Rudnick and Nyblade, 1999) indicate $T = 1200$–1300 °C at 200 km and temperature gradients of ca. 4.5°/km (Finland) to 5.5–6°/km (Siberia, Kaapvaal) below 100 km depth (the deepest samples display some scatter, probably due to heating by magmas that transported the xenoliths). Such gradients imply $q_m = 16$ to 18 mW m^{-2} to 22 mW m^{-2} ($k = 3.5$ W m^{-1} °K^{-1}). As

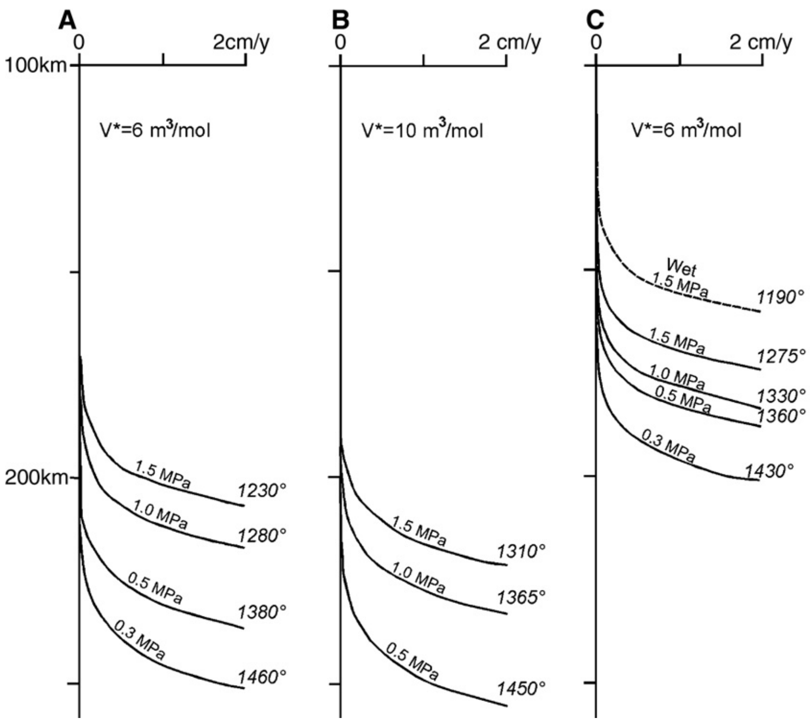

Fig. 2. Depth variation of v_x and its dependence on the basal drag. Full lines give the values for dry lithosphere; broken line — for wet lithosphere, deforming 10 times faster than dry lithosphere. Also shown are the temperatures where $v_x = 2$ cm/year. A. Variation of v_x for geotherm 1 of Fig. 3, for low activation volume. B. Variation of v_x along geotherm 1 of Fig. 3, for a high activation volume. C. Variation for a warmer geotherm (same moho temperature as geotherm 6 in Fig. 3, but with a mantle temperature gradient of 6°/km. Note that when the basal drag is 0.3 MPa the temperature where $v_x = 2$ cm/y is above the mantle adiabat shown in Fig. 3.

noted, these xenolith geotherms have the features of conductive geotherms, and therefore are often considered to represent the thermal steady state of Archean cratons.

Fig. 2 shows how $v_x(z)$ depends on τ_{xz} in an Archean craton with a representative xenolith geotherm (q_0=41 mW m^{-2}, T(200 km)=1200 °C, temperature gradient of 5°/km below 100 km; with k=3.5 W m^{-1} °K^{-1} then q_m=17.5 mW m^{-2}; geotherm 1 in Fig. 3), using the parameters listed by Hirth and Kohlstedt (2004). It is seen that $v_x(z)$ is negligible over a considerable depth interval, which represents the undeformed mechanical boundary layer, as defined by Parsons and McKenzie (1978) and Sclater et al., 1980). Here the term is used to emphasize negligible deformation at the conditions envisaged, though at higher stress levels (and resulting strain rates) this level may appear weak (cf. Jackson, 2002).

Below a certain depth, which depends on τ_{xz} and the geotherm, $v_x(z)$ increases appreciably and reaches a value of 2 mm/year over a depth interval of 20–25 km. The shearing in this interval allows decoupling of the lithosphere from the underlying mantle and thus this zone defines the base of the lithosphere. When $v_x(z)>$ 1 mm/year its value increases downward very rapidly, so choosing a threshold greater than 2 mm/year will hardly change the thickness of the decoupling zone. This decoupling zone becomes deeper and hotter as τ_{xz} decreases. Also, increasing V^* makes the decoupling zone deeper, other things being equal. Thus, when $V^*=10.10^{-6}$m^3/mol the decoupling zone for a given τ_{zx} is ca. 20 km deeper and thus is ca. 100 °C hotter than when $V^*=6.10^{-6}$ m^3/mol (Fig. 2). Within the decoupling zone the magnitude of τ_{xx} that produces $\dot{\varepsilon}_{xx}=3\times10^{-18}$ is 5–10 times smaller than τ_{xz}, which shows that simple shear indeed approximates well the total strain Such a behavior of $v_x(z)$ is found also in all other models, regardless of the geotherms.

In the mechanical boundary layer the strain is negligible, so heat is necessarily transported by conduction, as is assumed in the calculations. This assumption breaks down, however, within the decoupling zone. Where $v_x(z)$=2 mm/year the shearing strain rate is about 10^{-15} s^{-1} and it increases to ca. 4.4×10^{-14} s^{-1} and ca. 3.9×10^{-14} s^{-1} where $v_x(z)$=2 cm/year when $V^*=6.10^{-6}$ m^3/mol and 10×10^{-6} m^3/mol, respectively. As the strain rate increases the role of heat conduction in the decoupling zone decreases in favor of heat advection, so the temperature gradient decreases. This may become important already where $v_x(z)<2$ mm/year, so then the strain rate will increase more slowly than in the model. Thus the decoupling zone is a part of the thermal boundary layer on top of the convecting mantle. Within this layer the temperature gradient decreases downward until it reaches

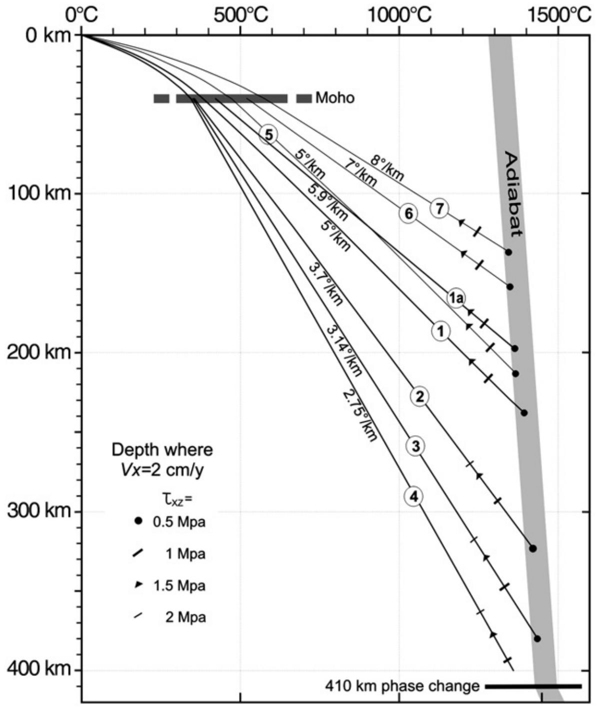

Fig. 3. The depth where v_x=2/cm/y along different geotherms for different values of the basal drag τ_{xz}. See text for discussion of the goetherms.

the adiabatic gradient, but the exact relations are beyond the scope of this work. Still, the sensitivity of strain rate to temperature assures the fast downward increase of the shearing at least in the upper part of this layer, so the simple model allows evaluating the thickness of the decoupling zone probably to within 5–10 km.

The variability of q_0 of Archean cratons records different geotherms, because it implies different moho temperature. A difference of 10 mW/m^2 in q_0 due entirely to a difference in q_{cr} implies a change of the moho temperature by some 50 °C (when conductive equilibrium is attained). This translates into a similar temperature difference at greater depth when q_m is the same, which in turn will change the depth of the decoupling zone by 10–15 km, other things being equal. Of course q_m can also vary, e.g. under the Kaapvaal craton the temperature gradient is somewhat higher than along geotherm 1, while q_0=45 mWm^{-3} (Rudnick and Nyblade, 1999), so the moho temperature is also higher (geotherm 1a, Fig. 3). For a given τ_{xz}, and the same V*, the decoupling zone is then some 30 km shallower than along geotherm 1 (Fig. 3).

The position of the decoupling zone relative to the mantle adiabat varies depending on the geotherm: it becomes deeper and hotter as τ_{xz} decreases, all other things being equal. Figure 3 shows adiabats with a potential temperature of 1280 °C–1320 °C, which is compatible with the temperature at the 410 km phase change (Akaogi et al., 1989; Katsura et al., 2004). Since the decoupling zone cannot be hotter than the adiabat, this requires that $\tau_{xz} \geq 0.5$ MPa when V*=6.10^{-6} m^3/mol (Fig. 3). When $\tau_{xz} \geq 1.5$ MPa the decoupling zone will be ≥150° colder than the adiabat, which is perhaps not be realistic. If a larger value of V* is appropriate, τ_{xz} should be even higher for the decoupling zone to be colder than the adiabat. Thus, if temperatures in the decoupling zones are not much colder than the adiabat, then τ_{xz} is expected to be similar to the stresses estimated by Hager and O'Connell (1981) and Doin et al. (1996). Hotter adiabats with potential temperatures of 1400–1450 °C were also considered in view of xenolith thermobarometry (Rudnick and Nyblade, 1999) or interpretations of the conditions of MORB formation (Green and Falloon, 1998). However, such adiabats do not seem to agree with the properties of the 410 km phase change as determined in the laboratory.

The foregoing results apply to dry olivine, but the presence of water can have an important effect (Garfunkel, 2004), as laboratory experiments demonstrate that it can considerably increase the strain rate at a given stress level (Karato and Wu, 1993; Kohlstedt et al., 1995; Hirth and Kohlstedt, 1996; Drury and Fitz Gerald, 1998; Hirth and Kohlstedt, 2004). Since the effect depends on the water content it cannot be represented by a unique set of parameters. To illustrate the significance of this effect models in which the strain rate is taken as 10 times faster than in dry rock are also shown in Fig. 2. It is found that this does not change much the thickness of the decoupling zone, but at any given value of τ_{xz} it will cause it to be 15–20 km shallower and thus 75 °C–100 °C cooler than in dry rock. Higher water content will lead to even thinner lithosphere.

These models show that a dry Archean lithosphere with a xenolith-type geotherm is expected to be 180–230 km thick, and its base is close, to or somewhat cooler, than the adiabat when τ_{xz} is 0.5–1.5 MPa and V*=6.10^{-6} m^3 mol^{-1}. This implies that the xenoliths of deepest origin were derived from near the base of the lithosphere. Such a picture is also compatible with the seismic data cited above which show that the high-velocity keels beneath the Archean cratons extend down to 180 up to ca. 250 km, while the deeper levels are uniform and can interpreted as having been homogenized by convection. For example, under the Kaapvaal craton the decoupling zone is expected to be some 170 km deep when τ_{xz}=1 MPa (Fig. 3), which is about the depth of the high velocity keel inferred by for this craton. However, the results of James et al. (2001) reveal a much deeper change in velocitiy anomalies, so though the interpretation of these results may be problematic (Priestley and McKenzie, 2002), other geotherms should also be examined.

Several authors (Jaupart and Mareschal, 1999; Artemieva and Mooney, 2001) suggested that beneath some cratons q_m=15–11 mW/m^2. This corresponds to gradients of 4.3–3.14 °C/km if k=3.5 W m^{-1} °C^{-1} or to 3.75–2.75 °C/km if k=4 W m^{-1} °C^{-1} (the value adopted by Artemieva and Mooney, 2001) i.e. considerably less than in xenolith geotherms, which gives rise to "cool geotherms" (geotherms 2, 3, 4 in Fig. 3, though geotherm 4 may be too cold to be realistic). With such geotherms the moho temperature will also be lower than along xenolith geotherms. In such cases for any given τ_{xz} the decoupling zone is deeper than along xenolith geotherms, other things being equal. The temperature of the decoupling zone increases as it becomes deeper, by ca. 50 °C over a depth interval of 150 km, which expresses the effect of pressure on creep that opposes the effect of temperature. In dry lithosphere with V*=6.10^{-6} m^3/mol the decoupling zone can be at 300–350 km or even deeper when $\tau_{xz} \geq 0.5$ MPa (otherwise it would be hotter than the adiabat). As in the other models, smaller V* or the presence of water will make the decoupling zone shallower.

Low values of q_m can explain very low values of q_0 that are found in some places, though the influence of groundwater flow should also be considered (Kukkonen, 1988). Such low values of q_m are not very well constrained, however, as they are calculated as the difference between q_0 and q_{cr}, which involves some uncertainty. Moreover, the regional and global tomographic studies cited above do not reveal high-velocity keels under cratons that were suggested to extend to 300–350 km. In particular, seismic data regarding the West African and Siberian cratons (Ritsema and van Heijst, 2000; Priestley and Debayle, 2003), two cratons that are >300 km thick according to Artemieva and Mooney (2001), show that their high velocity roots extend only to ≤250 km. It is also noteworthy that some global tomographic studies show thicker keels than regional studies (Artemieva and Mooney, 2002), the latter having a better depth resolution (Priestley and McKenzie, 2001). These problems show that the existence of very thick lithospheric keel with cool geotherms that differ considerably from those in adjacent areas needs further verification, though it cannot be excluded and should be kept in mind.

3.2. The lithosphere of Proterozoic shields

The Archean–Proterozoic transition was a period when the course of the earth's evolution was modified. One of the results was a change in the average composition of the continental crust and the lithosphere (Taylor and McLennan, 1985; Griffin et al., 1999, 2003; Gaul et al., 2000). However, when particular areas are considered individually they display considerable variability. The difference from Archean cratons is expressed in the often slower seismic velocities in the uppermost mantle beneath Proterozoic and younger terrains (e.g. Röhm et al., 2000 Goes et al., 2000; Goes and van der Lee, 2002; Godey et al., 2004; Poupinet et al., 2003; Ritsema et al., 2004), but considerable local variability of the seismic velocities exists (e.g. under the Baltic shield: Bruneton et al., 2004). These differences from Archean cratons are expected to be expressed also in the thickness of the lithosphere.

The important thing regarding the thickness of the lithopsheric is that on a global scale Proterozoic (and younger) terranes tend to have a higher heat flow than Archean terranes, though in some particular areas the values overlap. The global average for Proterozoic terranes is $q_0 = 55 \pm 17$ mW m^{-2}, and is even higher in younger terranes (Nyblade and Pollack, 1993; Pollack et al., 1993), i.e. distinctly higher than the average in Archean cratons. The considerable variability of q_0 results mainly from variations in the heat generation in the crust, which quite often is higher than in Archean crust (Morgan, 1984, 1985; Jaupart and Mareschal, 1999; Artemieva and Mooney, 2001). However, as noted above, the value of q_m ($=q_0 - q_{cr}$) is not tightly constrained, reflecting the uncertainty in q_{cr}. Some authors (e.g. Morgan, 1985; Jaupart and Mareschal, 1999) concluded that q_m does not vary much between Archean and younger regions, whereas others thought that it is higher in the younger regions, tending to be broadly correlated with q_0 (Pollack and Chapman, 1977; Artemieva and Mooney, 2001).

In both cases, however, higher values of q_{cr} will cause the moho temperatures to be higher when a thermal steady state is approached. By itself an increase of 15 mW m^{-2} in q_{cr} will increase the moho temperature by 70–80 °C. As the moho temperatures serve as anchors for the downward continuation of the geotherms (e.g. Morgan, 1985), the mantle temperatures will necessarily also be higher than where q_{cr} is low. If q_m is similar to that under Archean cratons, the geotherms in the lithospheric mantle will be parallel to those in the latter. This is represented by geotherm 5 in Fig. 3 that differs from geotherms 1 only in having a 100 °C higher moho temperature. This causes the decoupling zone at the base of a dry lithosphere to be ca. 20 km shallower for any given τ_{xz}, signifying a somewhat thinner lithosphere. The value of q_m can be the same because the temperature drop is smaller across the thinner lithosphere, so the temperature gradient remains the same. The influence of the various parameters on the thickness will be the same as in Archean cratons.

If q_m is higher than under Archean cratons, the lithospheric mantle will be even hotter, mainly because of the higher slope of the geotherm (which is added to the moho temperature), so the decoupling zone will also be shallower. For example, when the temperature gradients in the mantle are 7° and 8 °C/km (i.e. $q_m = 24.5$–28 mWm^{-2} if $k = 3.5$ Wm^{-1}°C^{-1}), then $v_x(z)$ will reach 2 mm/y at depths of ca. 140 km and 125 km, respectively when $\tau_{xz} = 1$ MPa and $V^* = 6.10^{-6}$ m^3/mol (geotherms 6, 7 Fig. 3). The lithosphere will be even thinner if it contains some water and/or if τ_{xz} is higher (Fig. 3 and 4). Fig. 3 shows that with the creep parameters used here the thermal gradients in the mantle should be that high or even higher for the lithosphere to be only 100–120 km thick, unless τ_{xz} is considerably higher than envisaged here or the lithosphere is strongly hydrated; in such cases the transition zone will ≥150 °C cooler than the adiabat. Conversely, lower values of q_m imply a thicker lithosphere.

It is noteworthy that in some Proterozoic terranes q_0 is very low, e.g. in the Meso-Proterozoic Grenville province in Canada where on average $q_0 = 41$ mW m^{-2}, the same as in the adjacent Archean craton (Jaupart and

Mareschal, 1999). In Canada Godey et al. (2004) did not identify a velocity or temperature differences between the two regions down to a depth of 200 km, which supports the notion that here they are underlain by lithosphere with essentially the same geotherms and thickness. However, further south in the USA Van der Lee and Nolet (1997) found a lower Vs under the Grenville province than under the nearby Archean basement, which correlates with differences in heat flow (Nyblade, 1999). This suggests compositional and temperature variations in the lithopsheric mantle beneath this province, and stresses the importance of the lateral variability of the lithosphere of Proterozoic terranes.

Thus, though when Proterozoic terranes are considered as a group their lithosphere is expected to be on average thinner than under Archean cratons, such a generalization cannot be applied when dealing with individual areas. Where q_{cr} is large enough it will cause the lithosphere to be thinner by a few tens of kilometers than where it is low even if q_{m} is the same, but this cannot be translated into a simple age-lithospheric thickness relation. The difference between the seismic velocities in the uppermost mantle beneath Archean cratons and younger shields can give additional insights, though their interpretation is not simple because they express the combined influence of both compositional and temperature differences. Godey et al. (2004) inferred an average temperature difference of some 150 °C between Archean and younger lithosphere under North America. The above considerations show that about a half of this can be accounted for by differences in the moho temperature resulting from variations in q_{cr} alone, while the rest may express a higher average temperature gradient. The required difference is no more than 1–1.5 °C/km on average, which means that q_{m} may be on average up to 3.5–6 mWm^{-2} higher under the post-Archean terranes. The question remains, however, as to what extent q_{m} correlates with the tectonothermal age of the lithosphere and the crust. If such a relation exists, it should somehow be maintained despite plate motions amounting to a few thousand kilometers relative to the underlying convection system on time scales of 100 Ma. On the other hand, the influence of q_{cr} will be maintained regardless of the plate motions, so if its effects on lithopsheric thickness influence the flow this could allow q_{m} to remain higher under thinner lithosphere (Ballard and Pollack, 1987), but this is outside the scope of the present study.

4. Discussion

The foregoing considerations show that the base of the continental lithosphere is defined by a decoupling zone some 20–25 km thick that separates it from the underlying convecting mantle. The depth and temperature of this zone depend primarily on the geotherm, the magnitude of the basal drag, and the presence of water. For any given geotherm the decoupling zone will be deeper and hotter when τ_{xz} is smaller, while for a given τ_{xz} (along different geotherms) this zone will be hotter the deeper it is (Fig. 3). As the geotherms differ from place to place, and τ_{xz} is expected to vary because it should depend on plate velocities which are quite variable, the base of the lithosphere need not follow any particular isotherm and its temperature may vary (by some 100 °C?), but it must be colder than the adiabat. With the creep parameters used here, and assuming an adiabat with a potential temperature of ca. 1300 °C this requires that $\tau_{xz} \geq 0.5$ MPa.

Regarding the conditions at the base of the continental lithosphere, two possibilities can be envisaged. One is that the xenolith geotherms represent the steady state geotherms in Archean cratons. In this case dry Archean lithosphere is expected to be 180–250 km thick, which is compatible with seismic information. In post-Archean terranes where q_{cr} is higher the geotherms should be hotter, which will cause the lithosphere to be thinner (by up to a few tens of kilometers, depending on the value of q_{cr}) for any given τ_{xz}. The difference can be enhanced by the presence of water, by higher values of τ_{xz}, and when q_{m} is higher. It can be envisaged that in extreme cases the combined effect of all these factors could cause the lithosphere to be only 100–120 km thick.

The other possibility is that the low estimates of q_{m} (14–11 mW m^{-2}) in Archean cratons define the thermal steady state in the lithospheric mantle, so its temperature is considerably lower than that recorded by the xenoliths. In such "cool" models the lithosphere is 300 km to >350 km thick if its base is not much colder than the adiabat. If q_{m} is similar in Archean and post-Archean terranes the lithosphere of the latter will be somewhat thinner because of the influence of the often higher values of q_{cr}. Higher values of q_{m} will also cause the lithosphere to be thinner. However, it can be thin (100–120 km) only if q_{m} is much higher than in the cool keels of Archean cratons. In this case q_{m} and the lithospheric thickness will vary much more than in the previous models (as was envisaged by Pollack and Chapman, 1977; Artemieva and Mooney, 2001).

The "cool" models raise several problems. The low values of q_{m} are uncertain, as noted above. In some cases the presence of the inferred thick lithosphere is not supported by seismic tomography. It is also not clear how the P–T arrays recorded by xenoliths from Archean cratons formed. Is it a coincidence that they resemble conductive geotherms that are compatible with the surface heat flow? How were such P–T arrays produced over depth intervals of >100 km? If q_m does not vary much, then the lithosphere of post-Archean terranes is expected to be quite thick, >200 km, but would such a thick lithosphere be easily affected by rifting and basaltic intraplate igneous activity which are widespread in post-Archean terranes? If, as seems likely, such activity tends to occur in thinner lithosphere (ca. 150 km or less?), then q_m and temperatures should be significantly higher under post-Archean terranes than under Archean cratons. The implied lateral differences in q_m, i.e. in the heat delivery from the convecting mantle, may result from lateral variations of lithospheric thickness (Ballard and Pollack, 1987), but can they be maintained while the plates move a few thousand kilometers? These problems suggest that models of the first type should be tentatively preferred, but the "cool" models cannot be rejected and should be further studied.

The relations outlined above raise the possibility of long-term changes of plate thickness. The important point is that the main controlling factors – the basal drag, presence of water, and perhaps also q_m – most likely change with time. The basal drag can change when plates move over different parts of the convecting system (rising, descending or horizontal flow), when the direction and velocity of plate motions change, or when the mantle flow changes (e.g. in response to reorganization of the mid-oceanic ridges and subduction zones). When τ_{xz} increases, previously undeformed portions of the lithosphere will be sheared and displaced sideways. Hotter material that comes instead can heat and weaken the remaining base of the lithosphere, enhancing its deformation erosion of its basal part. When τ_{xz} decreases, top parts of the underlying mantle will begin to move with the overlying lithosphere and thus become a part of it (and be cooled). Rising plumes can add water to the lithosphere and weaken it, causing it to deform more readily. The widespread evidence for metasomatism that includes formation of hydrous minerals records the occurrence of events of water addition (Menzies and Hawkeswort, 1987). Niu (2005) showed that there is a great potential for addition of water by dehydration of subducting slabs, and this may have a significant, even dominant, role in thinning the lithosphere. The two effects – weakening by water addition and changes in the basal drag– can of course operate simultaneously.

In the case of Archean cratons an increase in τ_{xz} and/or softening of the lowermost lithosphere will move peripheral parts of their compositionally distinct keels sideways, so this material will no longer be beneath the cratons and most likely will eventually mix with the convecting mantle so that it will be permanently removed from the cratons. Thus buoyancy of lithospheric keels is not a sufficient guarantee for the survival of ancient lithosphere because it does not protect it from being swept away sideways.

An example of thinning of the lithosphere is provided by the Archean North China (Sino-Korean) craton (NCC) which was still ≥200 km thick in the Ordovician, but was considerably thinned in the Late Mesozoic and later times, based on data on xenoliths in volcanics of these ages (Menzies et al., 1993; Fan et al., 2000; Griffin et al., 1998; Xu, 2001, Zheng et al., 2001; Gao et al., 2002). The nature of Cretaceous basic magmas that invaded the NCC indicates that its lower part was already modified when at that time (Zhang et al., 2003, 2004). Here only a few points relating these processes to the geologic history of the NCC and the surrounding regions can be mentioned.

The late Paleozoic and Triassic setting of the region was dominated by the assembly of China and nearby regions along ca. W–E trending zones of plate convergence and collision. Later, in the Mesozoic, the situation changed drastically as the eastern margin of the area came under the influence of the Pacific subduction system (e.g. Faure and Natalin, 1992). In this setting a new stage of igneous activity and deformation began in east China (Ma and Wu, 1987; Liu, 1987; Tian et al., 1992; Menzies et al., 1993; Ratschbacher et al., 2000; Ren et al., 2002; Zhang et al., 2004; Wu et al., 2005). In particular, the now modified part of the NCC was invaded in many places by basic magmas and was also deformed in the Cretaceous, while in the Cenozoic a new pulse of igneous activity took place and a prominent system of extensional basins formed over the craton. Such events are exceptional when compared with the behavior of other Archean cratons worldwide.

The processes that modified the NCC seem to have occurred under a much larger area all along eastern China where Cretaceous and Cenozoic igneous activity and intraplate deformation by rifting and faulting took place (Ma and Wu, 1987; Liu, 1987; Tian et al., 1992; Ren et al., 2002). This area differs from the more western part of China in its gravity anomalies, topography, lower seismic velocities in the uppermost mantle, thinner crust, and higher heat flow (Liu, 1987; Tian et al., 1992; Griffin et al., 1999; Hearn et al., 2004). All these features define a distinct belt that cuts across the pre-Jurassic grain of the country but is generally parallel to the Pacific subduction system and extends to a distance of ca. 1500–2000 km west of it.

These relations point at a process related to this subduction system. Griffin et al. (1998) advocated thermal erosion and chemical modification resulting from convective overturn related to the subduction as the mechanism of thinning of the lithosphere of the NCC, while Xu (2001) invoked thermo-mechanical (and chemical) erosion of the base of the lithosphere. Niu (2005) proposed that the lithosphere of the eastern China belt was hydrated by water derived from subducted slabs that extended far west of the Pacific, and later this softened part of the lithosphere was later removed by flow in the mantle above the slabs. In other models of subduction zones (Schmidt and Poli, 1998; Stern, 2002) the descending slabs are dehydrated mostly within a few hundred kilometers of the subduction zone where the slabs are not deeper than ca. 200 km, but Niu (2005) pointed at the possibility that water can be released also at later stages of slab descent when they reached considerable distances west of the subduction zones. This is an attractive argument, but it requires further evaluation, as it is also not clear whether the slabs extended under the entire width of the east China belt already during the Mesozoic stages of modification of the lithosphere. However, even if hydration and metasomatism were of prime importance, this had to be accompanied/followed by removal of the lower lithosphere of the NCC, which requires flow in the underlying asthenosphere.

The descent of the slabs that were subducted along the periphery of the Pacific Ocean is expected to induce a strong circulation under the continent, especially if the slabs retreated (rolled back) (Garfunkel et al., 1986; Ribe, 1989; Kincaid and Sachs, 1997). Such a sub-lithospheric flow is also required to replenish the mantle wedge with hot material and thus sustain magma generation along the subduction zone despite the continuing cooling from below of the mantle wedge by the cold descending slabs. Such a flow system could form only when the Pacific subduction system began to dominate the evolution of easternmost Asia, but not in the late Paleozoic and Early Mesozoic times when N–S plate convergence shaped this area, so a change in the convection regime is implied. The establishment of this flow system could increase the drag at the base of the lithosphere of the region, as discussed above, and therefore it is here assumed to have been an important agent for thinning the NCC, especially during the Mesozoic stage of its modification (though mantle plumes could also contribute). This would be much enhanced if addition of water weakened the base of the lithosphere, as envisaged by Niu (2005).

The widespread normal faulting and rifting and the thin crust of the modified region show, however, that the process included not only removal of the lower part of the lithosphere but also rejuvenation of major older faults as the Tan Lu Fault (Ratschbacher et al., 2000) as well as some widespread stretching. This could have resulted from the drag at its base once it was weakened. However, this (and the young igneous activity) seems to point at a more complex process than envisaged here, which requires further study.

The Indian shield may be another place where lithosphere was modified. It experienced a phase of very rapid northward drift in Cretaceous and Early Cenozoic times (Scotese et al., 1988), so in this period the basal drag could have been considerably greater than in earlier and later times. This is expected to thin the lithosphere, as discussed above. The thinning could have been variable because different parts of the shield have different ages, so the lithosphere probably varied laterally. Negi et al. (1986) and Pandely and Agrawal (1999) concluded that this actually took place, mainly on the basis of heat flow data, but other researchers dispute this conclusion (Gupta, 1993). More information about heat flow and generation and precise tomographic data are needed to test this possibility. The Late Proterozoic Arabian shield may have also been modified. It is a region of Neogene–Quaternary uplifting and volcanism that accompanied the opening of the Red Sea, but extend >500 west of it (Choubert and Faure-Mauret, 1975), which suggests addition of new hot material beneath the shield. This is supported by temperatures recorded by mantle xenoliths that significantly exceed the steady state geotherm inferred from the heat flow (Stein et al., 1993), which point at heating, at least locally, at depths of some 80–100 km. Moreover, velocities of surface waves traveling beneath the western part of the shield are notably slow (Hadiouche and Zuren, 1992), and the attenuation of seismic waves traveling beneath the shield is greater than under other Precambrian shields (Mitchell et al., 1997), which further supports the interpretation that the thermal regime beneath the shield changed by emplacement of hot material, most likely related to the breakup of the African–Arabian continent, but extending ca. 500 km or more inland of the site of continental splitting along the Red Sea. More data about heat flow and the deep structure of the shield are needed to test these interpretations.

These examples illustrate how continental lithosphere can be modified in different tectonic settings. More data on the uppermost mantle are needed to test the operation and relative importance of different processes that modify the lithosphere.

Another question raised by the above considerations is how the thick lithosphere of Archean cratons was stabilized in the first place, whereas later in earth history only thinner continental lithosphere formed. In the

Archean the crustal heat generation was about double the present one, while the mantle temperature is thought to have been higher than today, so geotherms in the material that now forms the lithosphere of cratons should have been warmer than today. Thus the above considerations show that if the basal drag was the same as today, the lithospheres should have been thinner than today. Does this mean that Archean convection was less vigorous than today, or had somewhat different characteristics? Clearly, these questions need special attention.

In conclusion, the relations presented above identified the main parameters that control the thickness of continental lithosphere and provide a framework for evaluating these parameters. These relations also indicate how changes in their values can modify the lithospheric thickness, and that such changes most likely occurred many times in earth history. The relations presented above can be improved when better knowledge of the creep parameters, thermal conductivity and anisotropy, state of stress and history of metasomatism in the continental lithospheric mantle becomes available. In this work the continental lithosphere was treated as a distinct entity, whose interaction with the convecting mantle was specified in terms of parameters such as the basal drag, q_m, and the supply of fluids (water). This allows modeling the influence of these factors while taking into account the properties of the lithosphere. However, full understanding of these factors requires models of mantle convection that can resolve the details of the flow and the interaction between the convecting mantle and the continental lithosphere.

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