# Rheology and deep tectonics

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**Abstract**
The distribution of the rheological properties of the lithosphere in space, and their variations in time, have a profound effect on the resulting tectonic deformation. A classical way of estimating these properties makes use of rheological profiles (strength envelopes). Although rheological profiles are based on assumptions and approximations which limit their resolving power, they are an efficient first-order tool for the study of lithosphere rheology, and their application clarifies the dynamics of tectonic processes. Two examples of the interaction of rheology and tectonics are discussed, namely, the post-orogenic relaxation of Moho topography (which is an additional factor to be considered in tectonic inversion), and the strength control on the level of necking in extension (which may lead to apparent local isostasy at passive continental margins and in sedimentary basins).

**Key words** *rheology of the lithosphere – strength envelopes – relaxation of Moho – level of necking*

## 1. Introduction

The rheological properties of the lithosphere are the filter function that transforms tectonic, gravitational, and thermal forces into observable deformation. Assuming that higher-order time derivatives of stress and strain are negligible, the *rheological equation* of polycrystalline aggregates has the general form

$$R\,(\sigma, \dot{\sigma}, \varepsilon, \dot{\varepsilon}, \{M_i\}, \{S_i\}) = 0 \qquad (1.1)$$

where $\sigma$ and $\varepsilon$ are the stress and strain tensors, an overdot denotes total time derivative, and $\{M_i\}$, $\{S_i\}$, $i = 1, 2, \ldots n$ are material parameters and state variables, depending on material, temperature, pressure, stress, and chemical environment.

Explicit forms of eq. (1.1) (usually, those for Coulomb-Navier shear failure in the brittle regime, and for dislocation-recovery power-law creep in the ductile regime) are used to construct *rheological profiles* (*strength envelopes*) which give a first-order representation of the variations of the rheological properties of the lithosphere in time and space. Strength envelopes are useful to estimate important parameters such as the depth of soft layers within the lithosphere and the lateral variations of total lithospheric strength (see Ranalli, 1995, 1996; Cloetingh and Burov, 1996 for examples of applications).

After a brief review of governing equations and parameters, this paper offers a discussion of the assumptions and limitations involved in the estimation of strength envelopes, and some preliminary suggestions as to how they may be overcome. Then, some important consequences of the rheological properties of the lithosphere (relaxation of Moho topography, level of necking in extension) are illustrated, as examples of the importance of rheology in geodynamic processes possibly affecting the Mediterranean region.

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## 2. Lithosphere rheology: strength envelopes

The long-term rheological behaviour of the lithosphere is strongly dependent upon temperature. For $T < T^*$, where $T^*$ depends on the material under consideration, the behaviour is brittle. At higher temperature, the behaviour is ductile. Brittle behaviour, in a static sense, is adequately described by the Coulomb-Navier shear failure criterion (Jaeger and Cook, 1979; Ranalli, 1995), also referred in the geological literature as Byerlee's law (Byerlee, 1967). Since the upper lithosphere is criss-crossed by pre-existing faults of random orientation with respect to the present stress field, the critical stress for sliding depends on material parameters, orientation of the fault, and orientation of the stress field. Explicit expressions for the general case are available (Ranalli and Yin, 1990; Yin and Ranalli, 1992). The corresponding form of eq. (1.1) is particularly simple where the pre-existing faults are cohesionless and ideally oriented with respect to the stress field (Sibson, 1974). Then one can write the failure criterion as

$$\sigma_1 - \sigma_3 = \alpha(f, \delta, \mu)\, \rho g z\, (1 - \lambda) \qquad (2.1)$$

where $\sigma_1$ and $\sigma_3$ are maximum and minimum compressive stress, $\rho$ is the average density of overlying rock at depth $z$, $g$ is the acceleration of gravity, $\lambda$ the pore fluid factor (ratio of pore fluid pressure to lithostatic pressure), and $\alpha$ a dimensionless parameter depending on type of faulting $f$, stress ratio $\delta = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$ where $\sigma_2$ is the intermediate principal stress, and friction coefficient $\mu$. The latter is usually in the range $0.5 < \mu < 0.8$, as determined both by experiment (Byerlee, 1967) and by inversion of tectonic data (Yin and Ranalli, 1995). For $\mu = 0.75$, $\delta = 0.5$, the parameter $\alpha$ equals 3.0, 1.2, and 0.75 in the case of thrust, strike-slip, and normal faulting, respectively (Ranalli, 1995).

In the ductile field, plastic deformation of silicate polycrystals usually takes place by dislocation-recovery (power-law) creep, except in very fine-grained ($\lesssim 10\ \mu$m as an order of magnitude) materials such as can be found in mylonite zones. Equation (1.1) then takes the form

$$\sigma_1 - \sigma_3 = \left(\frac{\dot{\varepsilon}}{A}\right)^{1/n} \exp\left(\frac{E}{nRT}\right) \qquad (2.2)$$

where $\sigma_1 - \sigma_3$ is the stress difference required to maintain the steady-state strain rate $\dot{\varepsilon}$, $R$ is the gas constant, $T$ is absolute temperature, and $A$, $n$, and $E$ are material parameters. (The parameter $A$ is actually slightly temperature-dependent, and the activation enthalpy has been taken equal to the activation energy $E$, thus neglecting pressure effects which are of minor importance in the lithosphere; *cf.* Poirier, 1985; Ranalli, 1995 for discussion).

While the frictional criterion (eq. (2.1)) is temperature-independent and practically the same for all common rocks, the expression for power-law creep (eq. (2.2)) is highly temperature-dependent, and the critical stress depends on the material under consideration. The creep parameters for various rocks and minerals are listed in table I. Parameters representative of the ductile behaviour of upper crust, lower crust (when different than the upper crust), and upper mantle are given in table II.

Strength envelopes are estimated by comparing brittle and ductile critical stress differences as functions of depth. They show that the rheological structure of the lithosphere is a strong function of depth, with soft layers (usually but not always the lower crust) separating harder (brittle or ductile) layers. (For examples of strength envelopes in various areas of the Mediterranean region, see Viti *et al.*, 1997). They also show that the total strength of the lithosphere (integral of the critical stress difference with respect to depth) is a function of position and is of the same order ($\simeq 10^{12} - 10^{13}$ N m$^{-1}$) as tectonic forces (Ranalli, 1991, 1995).

The input parameters for the estimation of strength envelopes are structure and composition of the lithosphere, tectonic regime, strain rate, temperature distribution with depth, and material coefficients for both brittle and ductile regimes. All of these are subject to uncertainties which are difficult to quantify (see Ranalli, 1995, 1996 for detailed discussion). Compilations of creep parameters for various rocks (*cf. e.g.*, Kirby and Kronenberg, 1987) show varia-

**Table I.** Creep parameters of lithospheric rocks and minerals (compiled from various sources; Ranalli, 1995).

| Material | $A\,(\mathrm{MPa}^{-n}\,\mathrm{s}^{-1})$ | $n$ | $E\,(\mathrm{kJ\ mol}^{-1})$ |
|---|---|---|---|
| Quartz | $1.0 \times 10^{-3}$ | 2.0 | 167 |
| Plagioclase ($An_{75}$) | $3.3 \times 10^{-4}$ | 3.2 | 238 |
| Orthopyroxene | $3.2 \times 10^{-1}$ | 2.4 | 293 |
| Clinopyroxene | 15.7 | 2.6 | 335 |
| Granite | $1.8 \times 10^{-9}$ | 3.2 | 123 |
| Granite (wet) | $2.0 \times 10^{-4}$ | 1.9 | 137 |
| Quartzite | $6.7 \times 10^{-6}$ | 2.4 | 156 |
| Quartzite (wet) | $3.2 \times 10^{-4}$ | 2.3 | 154 |
| Quartz diorite | $1.3 \times 10^{-3}$ | 2.4 | 219 |
| Diabase | $2.0 \times 10^{-4}$ | 3.4 | 260 |
| Anorthosite | $3.2 \times 10^{-4}$ | 3.2 | 238 |
| Felsic granulite | $8.0 \times 10^{-3}$ | 3.1 | 243 |
| Mafic granulite | $1.4 \times 10^{4}$ | 4.2 | 445 |
| Peridotite (dry) | $2.5 \times 10^{4}$ | 3.5 | 532 |
| Peridotite (wet) | $2.0 \times 10^{3}$ | 4.0 | 471 |

**Table II.** Typical creep parameters of lithospheric layers.

| Layer | $A\,(\mathrm{MPa}^{-n}\,\mathrm{s}^{-1})$ | $n$ | $E\,(\mathrm{kJ\ mol}^{-1})$ |
|---|---|---|---|
| Upper crust | $10^{-6}$-$10^{-3}$ | 2.0-3.0 | 150-230 |
| Lower crust | $10^{-3}$-$10^{-2.5}$ | 3.0-3.2 | 230-270 |
| Upper mantle | $10^{3}$-$10^{4.5}$ | 3.0-4.0 | 470-535 |

Upper crust = felsic to intermediate; lower crust = intermediate to basic (excluding mafic granulite); upper mantle = peridotitic (wet or dry).

tions, not untypically of ± 25%, for similar materials. Uncertainties in geotherms can be as much as ± 150°C at the Moho (Fadaie and Ranalli, 1990; Lamontagne and Ranalli, 1996). For these reasons, rheological profiles – while useful in providing a first-order estimate of the rheology of the lithosphere – cannot be used to resolve the detailed rheological structure, except in very well-constrained areas.

## 3. Towards «second generation» strength envelopes

The limitations mentioned in the previous section deal mainly with uncertainties in input parameters. More basically, the estimation of rheological profiles involves assumptions about the underlying physical processes which drastically simplify them and consequently limit the applicability of the inferred profiles. These methodological assumptions may be classified into two groups, the first having to do with the actual rheology, and the second with the conditions of deformation (homogeneous strain and constant strain rate).

The assumption of Coulomb-Navier frictional rheology overlying power-law ductile rheology is an oversimplification in at least two respects (in addition to the obvious consideration that the brittle/ductile transition in nature is transitional rather than abrupt). The validity of Byerlee's law is experimentally confirmed only up to pressures corresponding to mid-crustal depths. At higher pressures, the critical stress becomes much less dependent on lithostatic pressure (in fact, possibly near-independent; see Ord and Hobbs, 1989; Shimada, 1993). The form of the rheological equation for this «high-pressure failure» is unknown, but its occurrence has the effect of reducing the extremely high stress differences (of the order of 1000 MPa) required for failure in the middle crust and upper mantle in regions of low geothermal gradient and low pore fluid pressure (see fig. 1). The actual conditions for the transition are uncertain, but it occurs at depths where the temperature is less than that commonly associated with the brittle/ductile transition (300 ± 50°C for quartz-rich rocks, 450° ± 100°C for feldspar-rich rocks, and 650° ± 100°C for olivine-rich rocks; *cf.* Scholz, 1990; Ranalli, 1995 for discussion).

The second oversimplification in the rheological behaviour applies to the ductile field. It is commonly assumed that the rheology is uniform below the brittle/ductile transition, the only variations (for a given material) being related to temperature. Studies of tectonically exposed upper mantle rocks, however, have shown that the ductile field can be subdivided into two subfields, a low-temperature one where ductile flow tends to concentrate along shear zones, and a high-temperature one where flow is pervasive (Drury *et al.*, 1991; Vissers *et al.*, 1991). This is also shown in fig. 1. It is interesting to note that the transition temperature from shear to bulk ductility (≃ 950 ± 50°C; Drury *et al.*, 1991) is similar to that inferred by Ranalli (1994) as determining the effective thickness of the lithosphere in flexure.

These complications in rheology are possibly related to the distribution of seismicity with depth, which usually shows a linear increase followed by an exponential decrease; *cf. e.g.*, Lamontagne and Ranalli (1996). In any event, the lower limit of seismicity should be controlled by dynamic factors – *i.e.*, the velocity weakening/velocity strengthening transition; *cf.* Scholz (1990) – and its identification

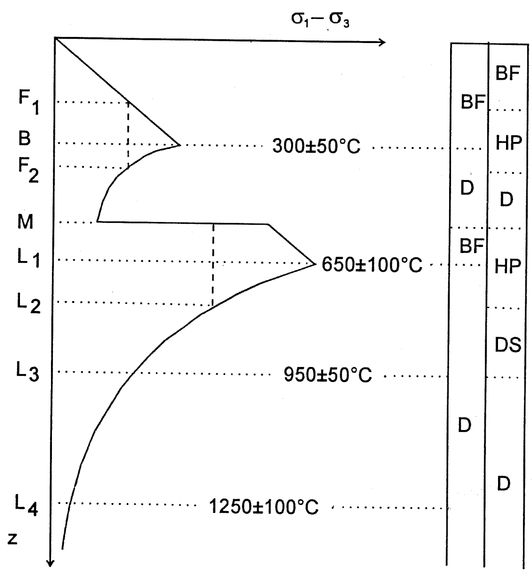


**Fig. 1.** Type rheological profiles, critical temperatures (when known), and corresponding variations of rheology with depth, assuming uniform crust and low (continental anorogenic) geothermal gradient. Standard strength envelope (full line), with critical depths $B$ (crustal brittle/ductile transition) and $L_1$ (mantle brittle/ductile transition) leads to the rheology shown in the left column (*BF*... brittle frictional; *D*... ductile). Changes due to high-pressure failure (dash lines), with critical depths $F_1$, $F_2$ (upper and lower limits of high-pressure failure in the crust), $L_2$ (high-pressure failure/shear zone ductility transition in the mantle), and $L_3$ (shear zone ductibility/bulk ductility transition in the mantle), lead to the rheology shown in the right column (*HP*... high-pressure failure; *DS*... shear zone ductility). The depths $M$ and $L_4$ correspond to Moho and lithosphere/asthenosphere boundary, respectively.

with the brittle/ductile transition is to some extent coincidental, simply implying that the transition depths for the two processes are similar.

A second methodological assumption, concerning the conditions of deformation, is that of homogeneous strain (that is, in one-dimensional rheological profiles, $d\varepsilon/dz = 0$, or equivalently $d\dot{\varepsilon}/dz = 0$). This restricts application of profiles in their present form to compressional or extensional pure shear. The lithosphere, however, is equally likely to deform in simple shear or in a more complex manner (*cf. e.g.*, Buck, 1991; Rutter and Brodie, 1992). The implications for strength envelopes are illustrated in fig. 2 for the case of extension. The «apparent» strain rate $\dot{\varepsilon} = (\Delta L/L)/t$, where $t$ is time, is not relevant to the physical process. The actual strain rate is the shear across the shear zone, $\dot{\varepsilon}_s = v_s/d$, where $v_s$ is relative velocity and $d$ the thickness of the shear zone. Similarly, the total strength of the lithosphere is given by the integral of the critical stress difference along the length of the shear zone.

Another assumption also concerns the conditions of deformation, since it postulates a constant strain rate (*i.e.*, $d\dot{\varepsilon}/dt = 0$). This imposes restrictions on the kinematic boundary conditions of the process (*cf.* McKenzie, 1978; England, 1983). A constant strain rate $\dot{\varepsilon}(t) = \dot{\varepsilon}_0$ implies that the width of the structure (for instance, in pure shear extension) varies as $L = L_0 \exp(\dot{\varepsilon}_0 t)$, where $L_0$ is the initial width. Since the velocity is $v = dL/dt = L_0\, \dot{\varepsilon}_0 \exp(\dot{\varepsilon}_0 t)$, it follows that the velocity increases with time, $dv/dt > 0$. (Conversely, the condition $v = v_0$ results in $d\dot{\varepsilon}/dt < 0$). In other words, constant strain rate deformation, as assumed in rheological profiles, must of necessity be transient since it implies an exponentially increasing velocity of convergence or divergence.

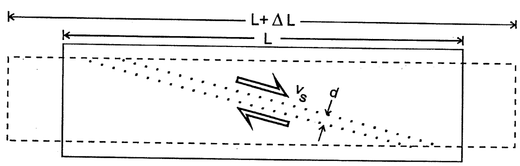


**Fig. 2.** Extension of the lithosphere in simple shear. Increase in width $\Delta L$ is accomplished by relative motion, with relative velocity $v_s$, across the low-angle shear zone of thickness $d$.

The above considerations do not deny the very real usefulness of rheological profiles in estimating the first-order rheological properties of the lithosphere. It is important, however, not to overlook their limitations, and to devise ways to take into account more realistic rheologies and conditions of deformation.

## 4. Examples: relaxation of Moho topography; level of necking

A partial list of geodynamic processes where rheology plays a central role, together with a few subjectively chosen key references, may include: fault reactivation (Yin and Ranalli, 1992); gravitational spreading of thickened crust (Molnar and Lyon-Caen, 1988); lateral extension of indented lithosphere (Tapponnier and Molnar, 1976); delamination of the crust (Bird, 1979); post-orogenic extension in collisional belts (Mareschal, 1994; Platt and England, 1994); slab breakoff at the end of subduction (Davies and von Blanckenburg, 1995); evolution of critically tapered accretionary and orogenic wedges (Davis *et al.*, 1983); formation of doubly vergent compressional orogens (Willett *et al.*, 1993); and origin and evolution of sedimentary basins (Cloetingh, 1995). Some of these processes are briefly discussed in Ranalli (1995).

The Mediterranean area is a prime illustration of the interactions of rheology and deep tectonics. The papers in this volume review the evidence and propose models for its interpretation. In this section, two processes are chosen as examples. The answers they provide – and the questions they raise – are very relevant to Mediterranean tectonics.

The first example deals with the possibility of *post-thickening relaxation of Moho topography* (see also Ranalli, 1992, 1996). Since the lower crust is a rheologically soft layer under all but the lowest geothermal gradients, and isostatic equilibrium does not imply hydrostatic equilibrium, the possibility exists that the lower crust can be extruded (*i.e.*, flow outwards) from beneath regions of high topography (*cf. e.g.*, Bird, 1991). Lower crustal flow results in relaxation of Moho topography. Consequently, together with erosion and isostatic rebound, it can affect both the depth of the post-orogenic Moho and the exhumation history of metamorphic terranes in orogenic belts as inferred from pressure-temperature-time paths (for an application of *PTt*-paths to the study of Alpine tectonics, see Spalla *et al.*, 1996).

The problem can be analyzed simply by considering the Moho as the interface between two semi-infinite fluids (consideration of the finite thickness of the layers does not introduce substantial modifications), representing the lower crust and upper mantle, with densities and viscosities $\rho_c$, $\eta_c$, $\rho_m$, $\eta_m$, respectively. The initial shape of the interface (*i.e.*, at the end of orogenic crustal thickening) is sinusoidal, with amplitude $z_0'$ and wavelength $\lambda_0$. If the two fluids are linear, incompressible, and in plane strain, the relaxation of the interface at time $t$ is given by (Ramberg, 1968)

$$z_0(t) = z_0' \exp(kt) \tag{4:1}$$

where

$$k = \frac{\lambda_0 g (\rho_c - \rho_m)}{4\pi(\eta_c + \eta_m)} \tag{4.2}$$

(Note that $k < 0$ since $\rho_c < \rho_m$). The relaxation rate, therefore, depends strongly on temperature, since viscosity depends exponentially on $T$ (see eq. (2.2)). Applying this argument to the Earth, it follows that the relaxation of Moho topography after an orogenic event depends on the subsequent thermal history of the lithosphere.

Equations (4.1) and (4.2) hold for linear (Newtonian) materials with stress exponent $n = 1$ (see eq. (2.2). However, linear and power-law ($n > 1$) viscosities are similar in the lithosphere and upper mantle for realistic values of strain rate and grain size (*cf.* Ranalli, 1995 for a discussions of creep mechanisms and effective viscosities). Therefore, use of nonlinear viscosities in the above solution gives a first-order idea of the Moho relaxation time. Results are shown in fig. 3, where the relative relaxation is given as a function of time for different Moho temperatures. Parameter values are $\lambda_0 = 1000$ km, $\rho_m - \rho_c = 500$ kg m$^{-3}$. The first parameter, *i.e.*, the wavelength of Moho deflec-

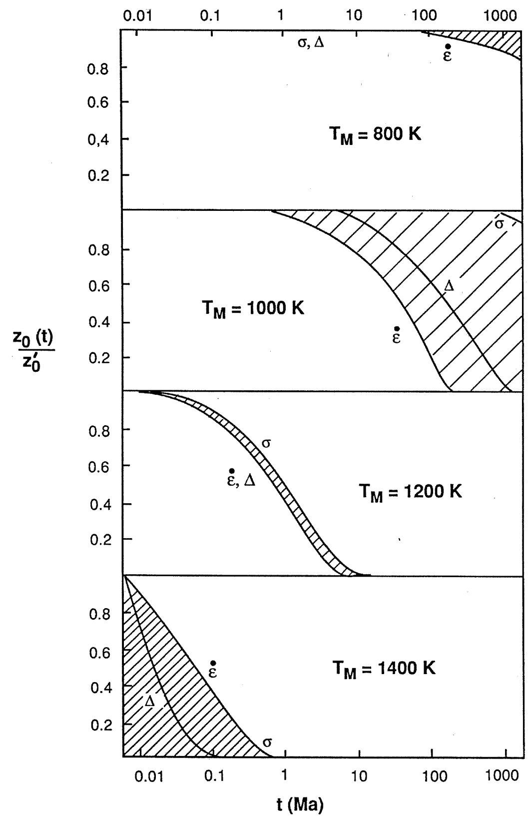


**Fig. 3.** Relaxation of Moho topography ($z_0(t)/z_0'$) as a function of time for various Moho temperatures $T_M$ (in degrees Kelvin). Curves based on the assumption of constant stress, constant strain rate, and constant viscous dissipation are denoted by $\sigma$, $\dot{\varepsilon}$, and $\Delta$, respectively (values given in the text). Shaded areas represent the range of values of relaxation covered by these three different assumptions. For instance, for a period of 100 Ma, significant relaxation (say, $z_0/z_0' \lesssim 0.3$) does not occur for $T_M = 800$ K and occurs only for constant strain rate conditions for $T_M = 1000$ K, while virtually complete relaxation occurs under any conditions for $T_M \geq 1200$ K.

tion resulting from orogenic compression, is an order of magnitude estimate, accurate within a factor of two in most cases. The viscosities $\eta_c$ and $\eta_m$ are estimated, at the appropriate temperature, using creep parameters for feldspar-rich lower crust and peridotitic upper mantle (see table II), under conditions of constant stress ($\sigma$ = 10 MPa), constant strain rate ($\dot{\varepsilon} = 10^{-15}$ s$^{-1}$), or constant viscous dissipation ($\Delta = \sigma\dot{\varepsilon} = 10^{-8}$ W m$^{-3}$). Further details are given in Ranalli (1996). As both wavelength and viscosities are estimates reflecting idealized conditions, results are to be taken only as an indication of the order of magnitude of Moho relaxation.

The above estimates show that Moho relaxation does not occur if the Moho temperature remains low ($T_M \lesssim 500$°C) after orogenic thickening. As Moho temperature increases, significant relaxation begins to take place, in times of the order of 100 Ma for $T_M \simeq 800$°C, and less than 10 Ma for $T_M \gtrsim 1000$°C. In other words, a moderate heating episode lasting 100 Ma, or an intense heating episode even if lasting only a few million years, lead to significant flattening of the Moho. This flattening is an additional factor to be considered, in addition with changes in tectonic stresses and detachment of lithospheric root, in models of post-orogenic extension.

The second example of rheological control on deep tectonics deals with the *necking of lithosphere in extension*. Thinning of the lithosphere under tensional forces tends to concentrate in zones of finite width, leading to formation of sedimentary basins, rifts, and passive continental margins (for a review of extensional processes in the lithosphere, see Cloetingh, 1995). This concentration of deformation results in necking of the lithosphere (see fig. 4a-c). The level of necking can be defined as the level which would remain horizontal during extension in the absence of isostatic restoring forces. The lithosphere, however, responds to isostatic forces by flexure, since it has finite strength. Consequently, if the level of necking is below that required for isostatic equilibrium, the lithosphere shows upward flexure; if the level is above that required for isostatic equilibrium, the flexure is downwards (Braun and Beaumont, 1989).

Although it is generally assumed that the level of necking is controlled by the distribution of strength with depth, the details are not clear. Necking in some cases coincides with the level of maximum strength, but in others appears to be the combined effect of two strong layers. Estimates of its depth in various extensional basins of the Mediterranean area vary from 4-6 to 25-35 km (Cloetingh *et al.*, 1995).

An interesting aspect of the problem is that, although the rheology of the lithosphere is such that it should respond to vertical loads by flexure, sometimes local isostasy accounts for the evidence (*cf. e.g.*, Cabal and Fernandez, 1995). In these cases, therefore, the necking level must be that resulting in no net vertical restoring force. An estimate of this equilibrium level can be obtained from a mass balance of lithospheric columns before and after thinning (see fig. 4a-c). Expressing it in terms of the thickness $h_a$ of asthenospheric material replacing colder lithospheric upper mantle, it is

$$h_a = \left(\frac{\beta - 1}{\beta}\right) \frac{h_c(\rho_c - \rho_0) + h_m(\rho_m - \rho_0)}{\rho_a - \rho_0} \tag{4.3}$$

where $\beta$ is the stretching factor (ratio of final to initial width of stretched lithosphere), $h_c$ and $h_m$ the initial thicknesses of crust and lithospheric upper mantle, respectively, and $\rho_c$, $\rho_m$, $\rho_a$ and $\rho_0$ the densities of crust, lithospheric upper mantle, asthenosphere, and basin infill (water or sediments). Equation (4.3) is valid for homogeneous pure shear (the ratio of crustal to mantle thickness does not vary during extension). If this condition does not hold, the equilibrium necking level depends on the relative stretching factors of crust and mantle, but it can still be estimated if these factors are known.

The value of $h_a$ in eq. (4.3) depends on densities and initial thicknesses of crust and lithospheric upper mantle. For reasonable values of these parameters (total lithospheric thickness 100 km, one-third of which consisting of continental crust; standard densities used in isostatic

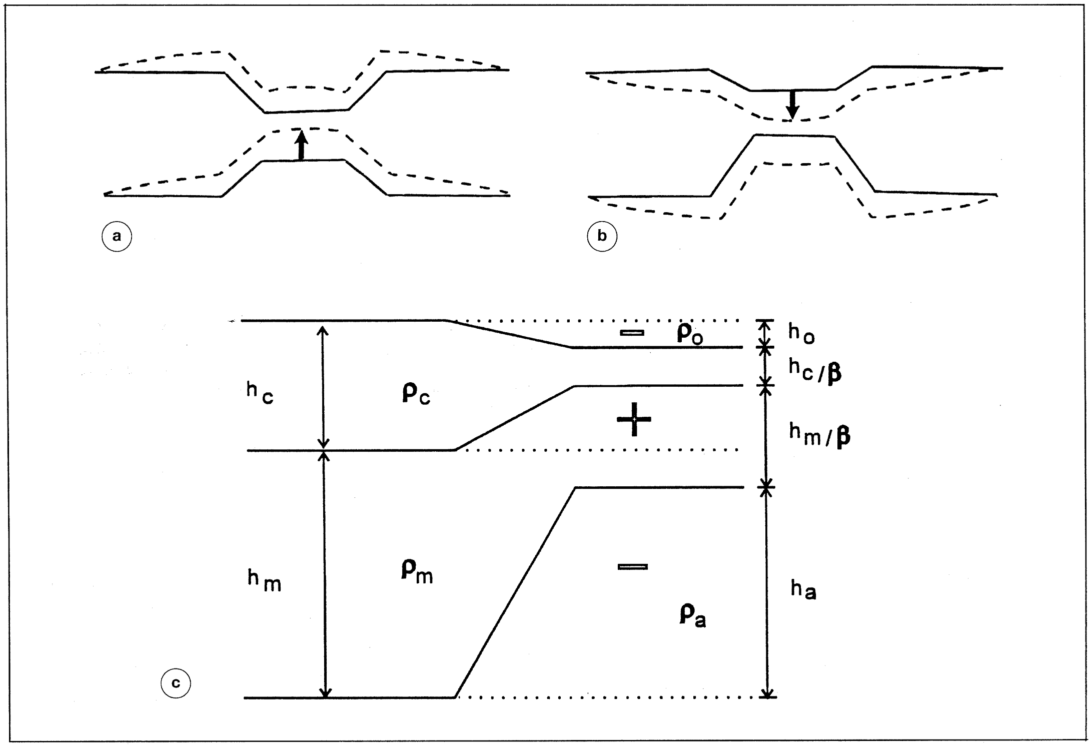


**Fig. 4a-c.** a) Upward; b) downward flexure of necked lithosphere; c) estimation of equilibrium level of necking (resulting in no flexure) for stretching factor $\beta$. Large plus and minus signs denote positive and negative mass anomalies.

calculations), one has $h_a$ (km) $\simeq$ 90 $(\beta - 1)/\beta$ and 100 $(\beta - 1)/\beta$ in the case of sediments and water infill, respectively. In order to associate necking with a definite depth $z_n$, linear interpolation between the downward displacement of the upper boundary of the lithosphere and the upward displacement of the lower boundary can be used, to obtain

$$z_n = \frac{h_c + h_m}{h_a + h_0} h_0 \tag{4.4}$$

that is, $z_n \simeq 4$ or 10 km for water or sediments infill, respectively. The critical level of necking is independent of the stretching factor $\beta$, since both $h_a$ and $h_0$ vary as $(\beta - 1)/\beta$. Levels of necking in this depth range are not uncommon (*cf.* Cloetingh *et al.*, 1995). Interestingly, for above-average temperature gradients (which should apply to extensional basins), an upper- to mid-crustal strong layer is found at similar depths (*cf.* Ranalli, 1995). Consequently, extension of continental areas with high heat flow (*e.g.*, $q \gtrsim 80$ mW m$^{-2}$), where the rheological profile of the lithosphere shows only one strong layer in the upper-middle part of the crust, results in levels of necking compatible with local isostasy. Since the level of necking has important consequences for margin uplift and basin stratigraphy, this process gives an example of how rheology controls not only tectonic, but also supracrustal phenomena such as sedimentation and erosion.

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