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# Seismicity, structure and strength of the continental lithosphere

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## Abstract

Geological–geophysical transects of active and ancient plate margins as well as structural studies of exhumed fault rocks confirm the general validity of laboratory-based strength–depth profiles for the long-term ($10^6$–$10^7$ a) rheology of the continental lithosphere. The deep structure of orogens and rifted margins requires that a relatively strong, olivine-rich upper mantle underlies a weaker granitic to dioritic crust. The continental crust itself comprises one or more viscously deforming layers that are overlain by a stronger brittle layer. This view refutes recent speculation that because the thickness of the seismogenic zone within the crust varies with the effective elastic thickness of the lithosphere, the crust must be stronger than the upper mantle. We argue that seismicity is an ambiguous indicator of strength and propose that earthquakes are more reasonably interpreted as a manifestation of transient mechanical instability within shear zones. Shear zones are often long-lived zones of weakness in which viscous mylonitic creep is punctuated by ephemeral high-stress events involving fracture, frictional melting, and a temporary, local loss of cohesion. Seismicity may therefore be used to locate current zones of episodic decoupling between and beneath crustal blocks in active interplate fault systems.


*Keywords:* seismicity; rheology; shear zones

## 1. The debate

Recently, some geophysicists have reinterpreted long-period teleseismic and gravity data from Alpine–Himalayan and central Asian mountain forelands to propose that at least in these locations, the lower continental crust is seismic and elastic and therefore stronger than the underlying, aseismic upper mantle [1–3]. The methods [4], assumptions, and logic underlying their proposal are straightforward: the close correlation of the effective elastic thickness of the lithosphere, $T_e$, with the thickness of the seismogenic zone, $T_s$, in these regions is attributed to stresses induced by lithospheric flexure that are periodically relieved by faulting [5]. Earthquakes are therefore assumed to occur in strong, elastic layers of the lithosphere, whereas weaker layers are presumed to yield and to undergo aseismic creep.

The proposal that the lower continental crustal is stronger than the upper mantle [3] contradicts a widely accepted lithospheric strength model—nicknamed variously the jelly sandwich or pine tree model—in which

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a weak, viscously deforming lower crustal layer (the jelly) is sandwiched between an overlying brittle upper crust and an underlying stronger, sometimes even brittle, upper mantle (the bread slices). This model of rheological stratification with a strong upper mantle [6] is based on the extrapolation to natural conditions of experimental constitutive equations for stable frictional sliding [7] and steady-state viscous creep [8]. The creep laws are derived in the laboratory at high temperatures to compensate for experimental strain rates that are unavoidably high for rock mechanicists to outlive their experiments. Originally, this model of rheological stratification was applied to the crust [9] but was eventually extended to the whole continental lithosphere [10].

From its inception, most experimentalists and structural geologists agreed that the jelly sandwich model is simplistic and should be applied to nature with caution. In part, this skepticism belies the realization that extrapolating laboratory rheological data to natural strain rates rests on many questionable or even unrealistic assumptions:

(1) uniform, constant strain rate [9] or stress [11];
(2) steady-state deformation of monomineralic aggregates (usually quartzite and dunite) with a strain-invariant microstructure;
(3) steady-state geotherm;
(4) constant kinematic configuration, usually simple or pure shear parallel to subhorizontal lithological layering; and
(5) size-independent strength [12].

Clearly, extrapolation of laboratory rheologies for barrel-shaped, mm- and cm-sized samples based on such assumptions yields only semiquantitative predictions of lithospheric strength at best [13]. In particular, assumptions (1) and (5) lead to consistent overestimates of lithospheric strength [12].

Despite the evident drawbacks of the jelly sandwich model, we will show that it explains the overall structure of orogens and rifted margins remarkably well. We then highlight some of the discrepancies between notions of lithospheric strength based on structural observations, experimental and theoretical rock mechanics, and geophysical models. Based on the distribution of fault rocks in exhumed continental crust, we argue that seismicity probably reflects transient, mechanical instabilities within shear zones and is therefore a poor indicator of lithospheric strength. Finally, we present an integrated geological–geophysical model for seismicity and crustal strength and discuss some implications of seismicity for mechanical coupling in the lithosphere.

## 2. Long-term rheology revealed by lithospheric structure

Geological structure on any scale reflects the rheology of rocks and their constituent phases, as well as their thermal and kinematic history. In areas where the kinematic history is well constrained, structure can be used to draw inferences about long-term lithospheric strength. Integrated geological and geophysical studies in several areas, particularly the central Alps (Fig. 1), the northern Tibetan plateau (Fig. 2), the Rhine Graben (Fig. 3), and the Galicia continental margin of offshore Spain (Fig. 4), yield insight into the relative strength of the continental crust and lithospheric mantle. We define upper and lower crust in lithological rather than seismological or rheological terms because lithology is more readily related to reflective seismic properties and is one important determinate of long-term rheology.

The central Alpine orogen is bounded by thrusts rooting at, or just above above, a mafic lower continental crust (Fig. 1). The lower crust of the upper Apulian plate is detached from its underlying mantle and forms a slender, northward-tapering wedge between the down-going European lithosphere and partly exhumed nappe edifice of the Alpine orogen accreted to the upper plate [14]. The nucleation of detachments (Fig. 1) at the base of the granitic crust indicates that this part of the crust was relatively weak. Furthermore, the existence of a lower crustal wedge derived from the upper plate suggests that the Moho was a first-order mechanical discontinuity. This discontinuity was probably preconditioned by preorogenic rifting which accentuated strength contrasts between the crust and the mantle in the Apulian passive continental margin [15]. We note that most current seismicity in this part of the Alps is located above most intracrustal detachments in the

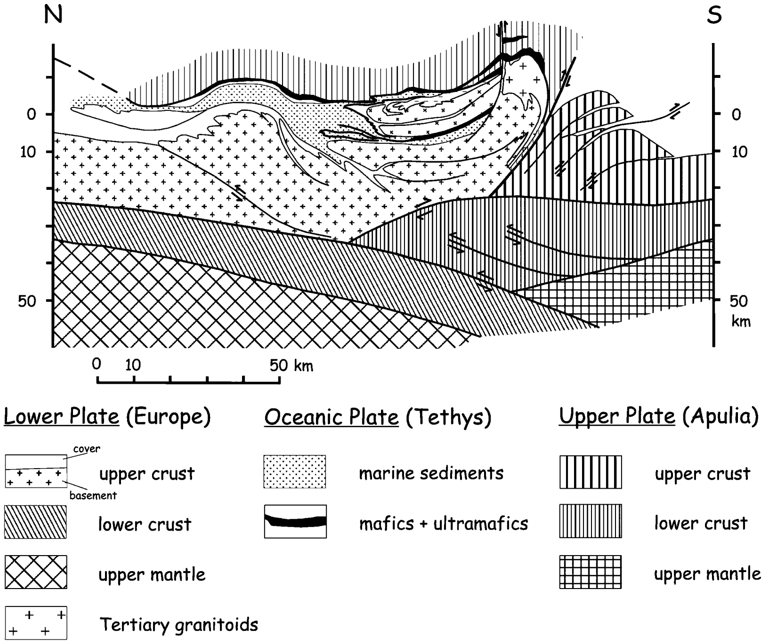



Fig. 1. EGT (NFP20E) transect of the central Alps after Pfiffner et al. [74].

uppermost 10 km of the crust [16] although infrequent, low-magnitude ($M_L \leq 4$) seismicity has been noted in the lower crust of the down-going European plate [17].

The Altyn Tagh–Kunlun transpressional fault system (Fig. 2) coincides with the northernmost of three lithospheric subduction zones underlying the Tibet plateau. The locations of these subduction

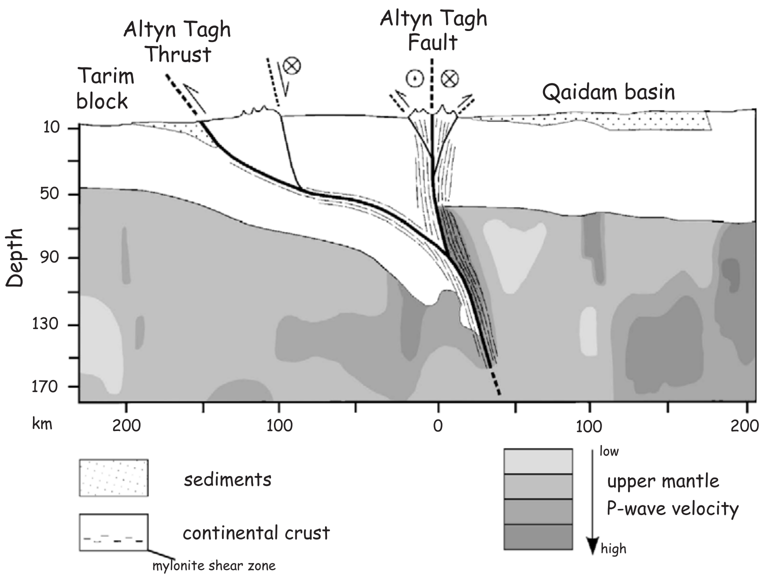



Fig. 2. Lithospheric structure along a transect of the Altyn Tagh fault system, northern Tibet, after Wittlinger et al. [18].

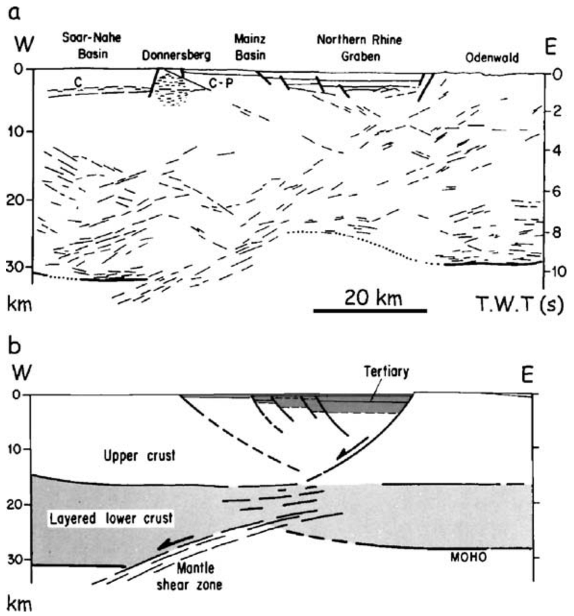



Fig. 3. Structure of the northern Rhine Graben: (a) line drawing of the DEKORP-ECORS deep seismic profile after Wenzel et al. [75]; (b) tectonic interpretation after Brun et al. [76].

zones are inferred from successive offsets of the Moho imaged by seismic tomography and receiver functions [18]. These fault systems accommodated oblique stepwise growth of the Tibet plateau and eastward lateral escape of the thickened Tibetan crust during Tertiary to Recent north–south shortening [19]. In Fig. 2, oblique convergence is accommodated by the Altyn Tagh strike-slip and thrust faults, both of which root in the upper mantle. The lithospheric mantle is displaced, indicating that it was strong compared to the shortened crust above.

In addition to the Alpine and Tibetan examples, there are several other well-documented cases where subducted lithospheric slabs extend downward from thickened crust and thrust offsets of the continental Moho (circum-Mediterranean Tertiary orogenic belts: e.g., [20]; Red Bank thrust in central Australia: [21,22]). In areas of active convergence, seismicity occurs both in the crust and the underlying mantle (Aegean, Carpathians: [20]). All these examples of so-called thick-skin tectonics indicate that the upper mantle is strong compared to the overlying crust.

The Rhine Graben imaged by deep seismic DEKORP-ECORS profiles (Fig. 3) comprises conjugate, steeply dipping, Tertiary normal faults that conjoin at the top of the reflective lower crust at about 15 km depth. There, the lower crust with layering likely inherited from late- to post-Variscan extension [23] is pervasively thinned and becomes less reflective. The Moho is displaced vertically some 4 to 6 km along a west-dipping, normal-sense shear zone that is offset from the graben axis. In the southern part of the graben, deeper earthquakes are located above the thinned lower crust, within the graben, and partly within the unthinned lower crust [24,25]. The vertical variation of deformational style—brittle faulting in the upper crust, distributed thinning of the lower crust, and localized deformation in the uppermost mantle—indicates that the lower crust decoupled the overlying crust from a stronger upper mantle.

A mechanical discontinuity at the Moho can also be inferred from deep-seismic reflection profiles of the Galicia rifted margin (Fig. 4). There, the seaward dip of brittle, normal faults, and the landward dip of tilted crustal blocks indicate a top-to-the-ocean shear sense along a master fault, corresponding to a seaward increase in thinning. Mantle rocks exhumed in the distal part of the margin display the opposite shear sense within a landward-dipping mantle shear zone [26]. Decoupling in the lower crust and opposite shear senses in the lower crust and lithospheric mantle are

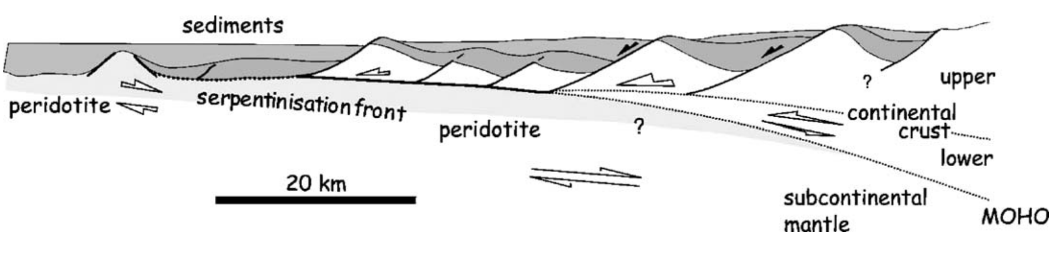
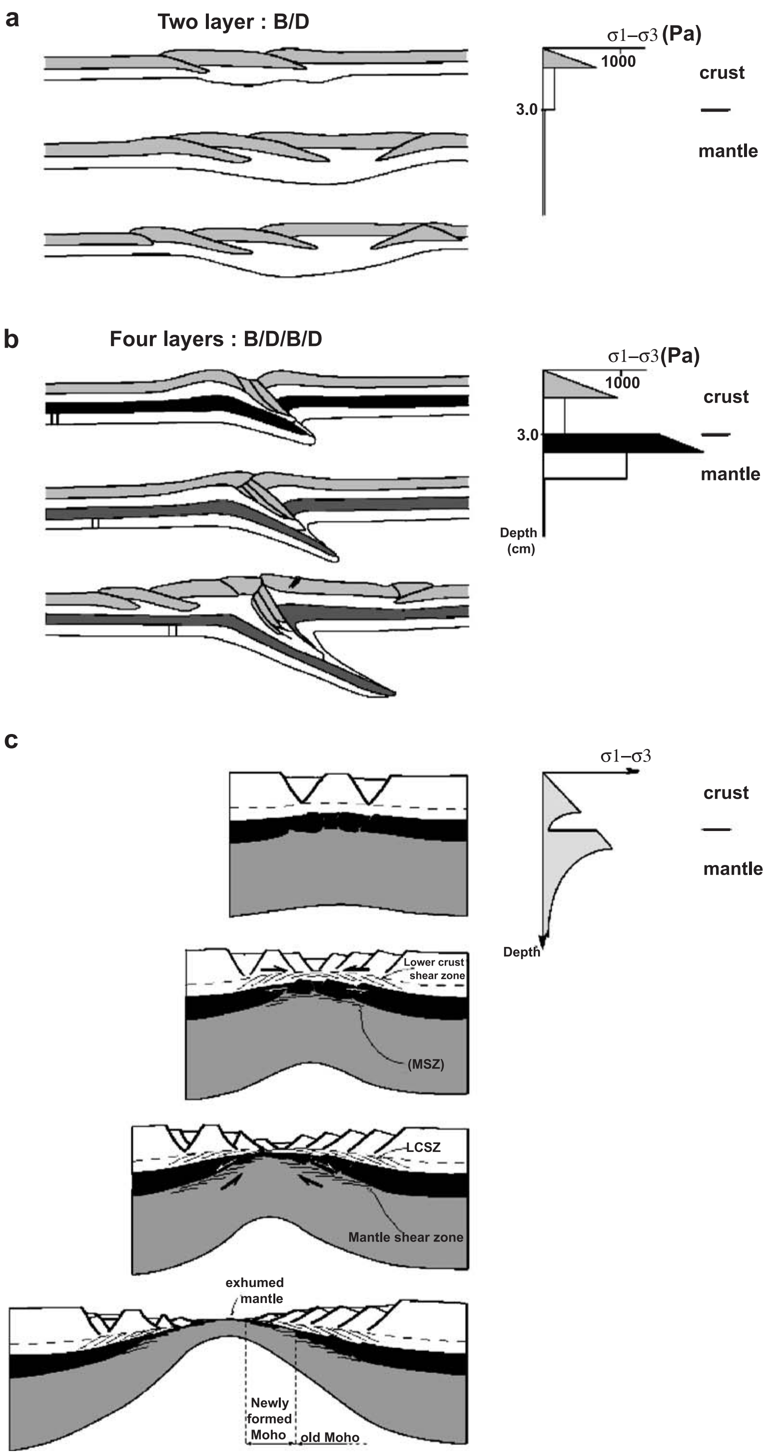



Fig. 4. Cross section of the eastern North Atlantic rifted margin off the coast of Galicia, Spain, after Boillot et al. [77].



a
Two layer : B/D
σ1–σ3 (Pa)
1000
3.0
crust
mantle
b
Four layers : B/D/B/D
σ1–σ3 (Pa)
1000
3.0
crust
mantle
Depth (cm)
c
σ1–σ3
crust
mantle
Depth
Lower crust shear zone
(MSZ)
LCSZ
Mantle shear zone
exhumed mantle
Newly formed Moho
old Moho

both diagnostic of a strong layer in the uppermost lithospheric mantle [27].

Although it is beyond the scope of this paper to review tectonic modeling, such models provide tests for a large variety of simplified strength profiles. In shortening models, only configurations with a stronger upper mantle layer develop realistic orogenic structures (e.g., compare Fig. 5a and b, see also [28,29]). Early on, this layer is offset by a thrust that eventually cuts across the lithosphere and facilitates crustal subduction (Fig. 5b). In extensional models (Fig. 5c), this strong mantle layer develops necking instabilities that lead to extreme crustal thinning and mantle exhumation, as observed in distal parts of rifted continental margins [27]. Although the uppermost mantle may weaken locally where affected by syntectonic hydration reactions and grain size reduction [15,30], the lithospheric mantle as a whole remains stronger than the overlying crust during rifting. Regarded generally, these results corroborate the idea that a strong lithospheric mantle localizes deformation at the lithospheric scale in both compression and extension [31].

To conclude this section, a wealth of integrated geological–geophysical studies reveal that on time scales of orogenesis and rifting ($10^6$–$10^7$ a), the lithospheric mantle is more competent than either the asthenosphere or the overlying continental crust. There are three first-order mechanical discontinuities in both shortened and extended continental lithosphere: (1) the crust–mantle boundary, which corresponds to the transition from quartz-, mica-, and/or feldspar-rich rocks above to olivine-rich rocks below; (2) the interface between dry, mafic (feldspar- and amphibole-rich) lower crust rocks below and hydrous, granitic rocks above; and (3) the sediment-basement contact. As discussed below, these discontinuities are long-lived, and with progressive strain, become the sites of pronounced weakness.

## 3. Lessons from structural studies in the field and the lab

The structural style of folds, boudins, and cleavages in exhumed rocks is a semiquantitative indicator of their competence contrast, and if one assumes uniform or nearly uniform strain rate, is also diagnostic of their relative strength [32,33]. Fig. 6 summarizes the competence contrast of lithospheric rocks and minerals at geological strain rates and at temperatures estimated from geothermometry and mineral equilibria in coexisting, syntectonic minerals [34]. It reveals two striking features: (1) olivine and pyroxene aggregates in the upper mantle are much stronger than feldspar, quartz, or mica aggregates in the crust, and (2) mineral aggregates with a small syntectonic grain size, generally $<10$ μm for naturally hydrous quartz [35] and olivine [30], are weaker than coarse-grained aggregates at temperatures and strain rates near the transition from viscous creep to frictional sliding.

Experimental and theoretical work on the rheology of polycrystalline aggregates has shown that the amount of weakening in the presence of a weak mineral phase depends on the interconnectivity of this phase and on its strength contrast with the other stronger phases [36]. The interconnection of even a small amount of weak phase can induce a significant drop in aggregate bulk strength. For example, the creep strength of a granitic rock containing 90% feldspar and only 10% interconnected quartz at a strength contrast of 5:1 is only 20% above that of a pure quartzite [36]. However, the magnitude of long-term weakening associated with the development of shear zones may be more modest than for a weak aggregate in a polyphase rock because shear zones do not always anastomoze completely on the lithospheric scale, in which case, stress is very heterogeneous in the deforming system (rock plus shear zones). Bulk rock strength therefore never drops far towards the uniform-stress lower bound and the strength of the

Fig. 5. Analogue brittle-ductile models of lithospheric scale deformation. B = brittle sand layer with Coulomb frictional sliding rheology; D = ductile silicon layer with Newtonian rheology. Shortening models in (a) and (b) after Davy and Cobbold [78]. (a) Two-layer shortening model in which thrust faults do not crosscut the whole lithosphere, leading to rather homogeneous thickening above the Moho; (b) four-layer model where the presence of a strong uppermost mantle localizes deformation, allowing subduction of the underthrusted mantle; and (c) progressive stretching of a four-layer brittle-ductile continental lithosphere leading to mantle exhumation at the passive margin, after Brun and Beslier [27]. The strong mantle layer undergoes necking at the rift axis. The bulk deformation is pure shear, with the lower crust sheared top-to-the rift axis and the ductile part of the mantle sheared top-away from this axis.

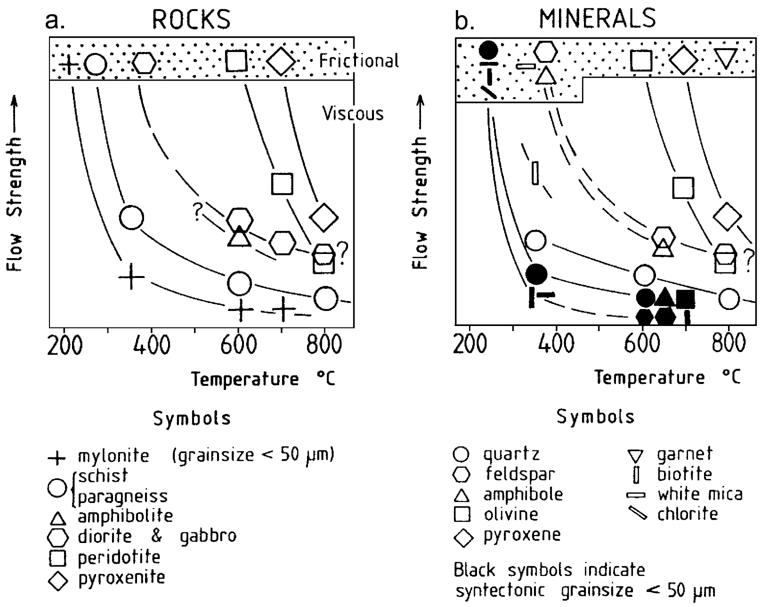



Fig. 6. Competence contrast of naturally deformed (a) minerals and (b) rocks as a function of temperatures in various paleodepths of exposed crustal cross sections (modified from Fig. 9 in [34]).

system remains within 15–20% of the bulk strength of homogeneously deforming rock [37].

Regardless of the view of weakening adopted, it is clear that shear zones accommodate most of the strain in the lithosphere [38]. More importantly, shear zones are long-lived, with life spans approaching those of the plate margin structures in which they occur. Mylonitic shear zones can develop rather quickly ($\leq 10^3$–$10^4$ years; [15]), but they remain active for about $10^3$–$10^7$ years, a range estimated from measured shear strains of between 10 and 100 for lower crustal and upper mantle shear zones [30,39] and shear strain rates for dislocation creep bracketed at $10^{-11}$–$10^{-14}$ $s^{-1}$. Cataclastic shear zones in shallower levels of the crust probably have similar life spans because although they have higher displacement and strain rates, they are generally narrower for a given length than mylonitic shear zones and therefore accommodate greater shear strain.

Although shear zones are weaker than their surrounding rocks over long time spans, they are inherently unstable structures and cannot be assumed to have supported a constant load throughout their history. This is inferred from the preservation of one or more generations of ultracataclasites, including pseudotachylites [40], in cataclastic and mylonitic shear zones. Pseudotachylites are particularly interesting in this respect because they represent ephemeral melt generated by intense frictional heating [41] at seismic displacement and strain rates. These rates vary from average values of 1–10 m/s to peak values of 10–100 m/s, $10^{-4}$–$10^{1}$ $s^{-1}$ near the tips of a propagating rupture [42,43]. Mutual overprinting relations between mylonites and pseudotachylites within some shear zone segments indicate that, locally, periods of creep are punctuated by short episodes of rapid, transient slip and rupture [44–47]. This behaviour has been documented in an exposed cross section of rifted continental crust, the Ivrea Zone, where mylonitic shear zones thinned the crust and the upper mantle in Early Mesozoic time [48]. Pseudotachylites formed during strain localization in retrograde mylonitic shear zones that were active during rifting, both in granitic, intermediate crustal rocks [34] and in lherzolitic upper mantle rocks [49,50]. Differential stress was high in these shear zones at rupture (>200 MPa) but relaxed rapidly to a much lower level of $\leq 10$ MPa [30]. In both the crustal and mantle rocks, initial fracturing was accompanied by the ingress of metamorphic fluids, leading to the syntectonic growth of hydrous minerals (mica, epidote, amphibole) at the expense of stronger reactant minerals (feldspar, garnet, pyroxene).

The observations above are consistent with the idea that the brittle-viscous transition is a zone of long-term weakness punctuated locally and temporally by ephemeral, high-stress events, particularly in the presence of fluids or melts [51]. Episodic loading and rapid unloading of shear zones is associated with seismicity, a characteristic of the brittle-viscous transition within broad depth intervals of the crust ("schizosphere"; [52]) and the upper mantle. Stable frictional sliding and steady-state creep—both basic assumptions of the jelly sandwich model—are therefore not appropriate descriptions of the short-term rheology of the lithosphere.

## 4. What do seismicity and gravimetry tell us?

The jelly sandwich model was attractive in the first place because experimentally derived strength–depth envelopes for quartz and olivine at natural creep strain rates [53] fit well with three other independent geophysical observations: (1) the downward increase of differential stress in the uppermost crust from in situ stress depth measurements [9], (2) the depth-dependent variation of earthquake frequency and magnitude, and (3) the positive correlation of postseismic stress drop with heat flow in different crustal provinces [6,54]. Implicit in using these observations to validate laboratory steady-state strength curves is the notion that strong rocks fail more frequently and at higher applied differential stresses than weak rocks. Indeed, this is precisely the logic recently used [3] to infer the high relative strength of the lower continental crust based on the occurrence of earthquakes within the crust and the absence thereof in the upper mantle.

Yet, is seismicity really a diagnostic criterion for the relative strength of rocks on the lithospheric scale? We don't think so, for the following reasons: Earthquakes are elastic waves emanated from a stressed surface when the strain energy stored along and around that surface is released suddenly. Although it is true that larger earthquakes indicate more strain energy released at rupture than for smaller earthquakes, it does not follow that adjacent aseismic layers have less strain energy stored and are therefore weaker. In other words, the lack of seismicity is ambiguous because it tells us that either no stress has accrued at all or the stress accrued is less than that required to fracture the rock (i.e., the failure or rupture strength). This in turn is either because the rock is weak (preventing stress from accumulating) or because the rock is strong (preventing it from fracturing).

As previously pointed out [3], gravity analyses yield only an estimate of the effective elastic thickness or rigidity of the lithosphere, not the actual depth to which elastic strength occurs. In fact, rocks always have elastic strength, even at great depth and very high temperatures. It is well known that the elasticity and yield strength of olivine in the mantle are greater than those of quartz, feldspar, or micas at comparable pressure and temperature [53]. Furthermore, the long-term rheology of the lithosphere is viscoelastic rather than elastoideal plastic so that creep at constant temperature relaxes stresses, causing the effective elastic thickness of the lithosphere to decrease during the initial stages of loading ([5], p. 283). Therefore, it is impossible, on the basis of gravity data alone, to determine whether the elastic strength of the lithosphere resides in the crust or, as we believe, derives primarily from the strength of the lithospheric mantle.

A correlation of effective elastic thickness, $T_e$, with the thickness of the seismogenic zone, $T_s$, is not a universal feature of the continental lithosphere. In a recent data compilation ([5], Fig. 6.41), it was shown that although $T_e$ and $T_s$ have similar depth ranges in rifted and young orogenic continental lithosphere ($<5$ km, 20–25 km), the depth-frequency peaks for $T_e$ and $T_s$ do not always match. Also, stable cratons show no peak in $T_e$ to 100-km depth, whereas seismicity clusters in the uppermost 25 km (see [5], p. 284 for brief speculation on why).

Arguing that seismically active layers above the calculated lower depth limit for the effective elastic thickness of the lithosphere have greater long-term strength [2] presumes that rocks below that depth limit have negligible elasticity and creep strength. Comparing the thicknesses of the elastic and seismically active layers in the lithosphere provides no mechanistic explanation for why seismicity occurs, especially within zones of long-term weakness.

## 5. Other mechanisms of seismicity and lithospheric strength

The jelly sandwich model of a rheologically stratified lithosphere is inadequate to explain seismicity and other short-term geophysical features, a conclusion already reached some time ago by the rock mechanics community [55]. Other rheological models proposed in recent years are compatible with the field observations above and explain seismicity in terms of mechanical instability rather than rock strength:

(1) In the shallow crust, time- and displacement-dependent changes in the frictional coefficient [56,57] and/or other properties of cataclastic fault rock (e.g., the development of a cohesive, impermeable gouge; [58]) can lead to frictional instabilities along a fault surface, provided that the fault rock is stiffer than its host rock or that externally imposed perturbations in sliding velocity are sufficiently large [59]. The resulting reduction in shear strength with sliding velocity (velocity weakening) leads to accelerating slip along fault surfaces [59].
(2) In deeper parts of the crust, viscous mylonitic creep is usually velocity strengthening, preventing deformation from attaining seismic rates during localization. However, because creep is thermally activated, runaway weakening may occur during near-adiabatic shearing below a critical temperature which is material-dependent (e.g., 200–300 °C in quartz; [45,60]).

As in (1), seismic instabilities develop if the loading agent (the host rock) is compliant enough, allowing the strain energy stored therein to be released abruptly along the fault. Likewise, syntectonic reaction of nonhydrostatically stressed minerals can lead to the spontaneous nucleation of weak, fine-grained products, especially if the reaction is exothermal, preventing it from "freezing". The reaction rate is enhanced for metastable reactants deformed outside of their pressure–temperature stability field [61,62]. The associated stress drop in the shear zone is larger for reactions with a large change in molar volume, which together with the exothermal condition, restricts this type of seismic instability to prograde solid–solid reactions at great depth in the mantle, e.g., in subducting lithospheric slabs. Endothermal reactions, especially those involving a fluid phase, can also induce rapid weakening during initial fracturing and spontaneous phase nucleation, but beyond this, grain growth tends to stabilize creep [30].

Clearly then, earthquake magnitude does not depend primarily on the absolute or peak strength of a rock layer but on the difference in rates of stress drop for the fault rock and its host rock at the time of rupture (Fig. 7). Initial localization is aseismic if the rate of stress drop in the instability region is low compared to the stiffness of the surrounding rock (inset a, Fig. 7). However, deformation can become unstable if the system is perturbed, for example, if the effective pressure or temperature decreases, or if strain rate increases. Rupture initiates where stresses exceed the fracture strength, preferentially at the tips of the shear zone. If the ensuing stress drop along the deforming segment is faster than in the adjacent locked or more slowly deforming rock, the elastic strain energy stored in the compliant host rock is released seismically along the fault, and earthquake magnitude is proportional to the area delimited by the divergent stress–decay curves (inset b, Fig. 7). Rupture terminates when stresses at the shear zone tips decay to below the fracture strength.

Earthquake frequency depends primarily on the (re)loading rate of the deforming system (fault plus host rock), which in turn depends on several interrelated, intrinsic and extrinsic factors including the rate of sliding (or strain rate), temperature, effective pressure, and of course, rock composition, microstructure, and fault-zone thickness [63]. Episodic fault slip is possible at conditions favouring instability close to the transition from stable to unstable behaviour [55], and oscillatory creep related to these instabilities at depth is expected on either side of the rupture surface at distances comparable to the depth of the seismogenic zone [64]. Transient behaviour is manifested in the rock fabric as mutually overprinting cataclasites and mylonites within zones of high shear strain (Fig. 6), as described above.

Conceivably, creep instabilities leading to rupture and even seismicity occur in mylonitic shear zones both above and below the effective lower depth limit for the elastic lithosphere calculated from gravity and topography. This neither confirms nor refutes the

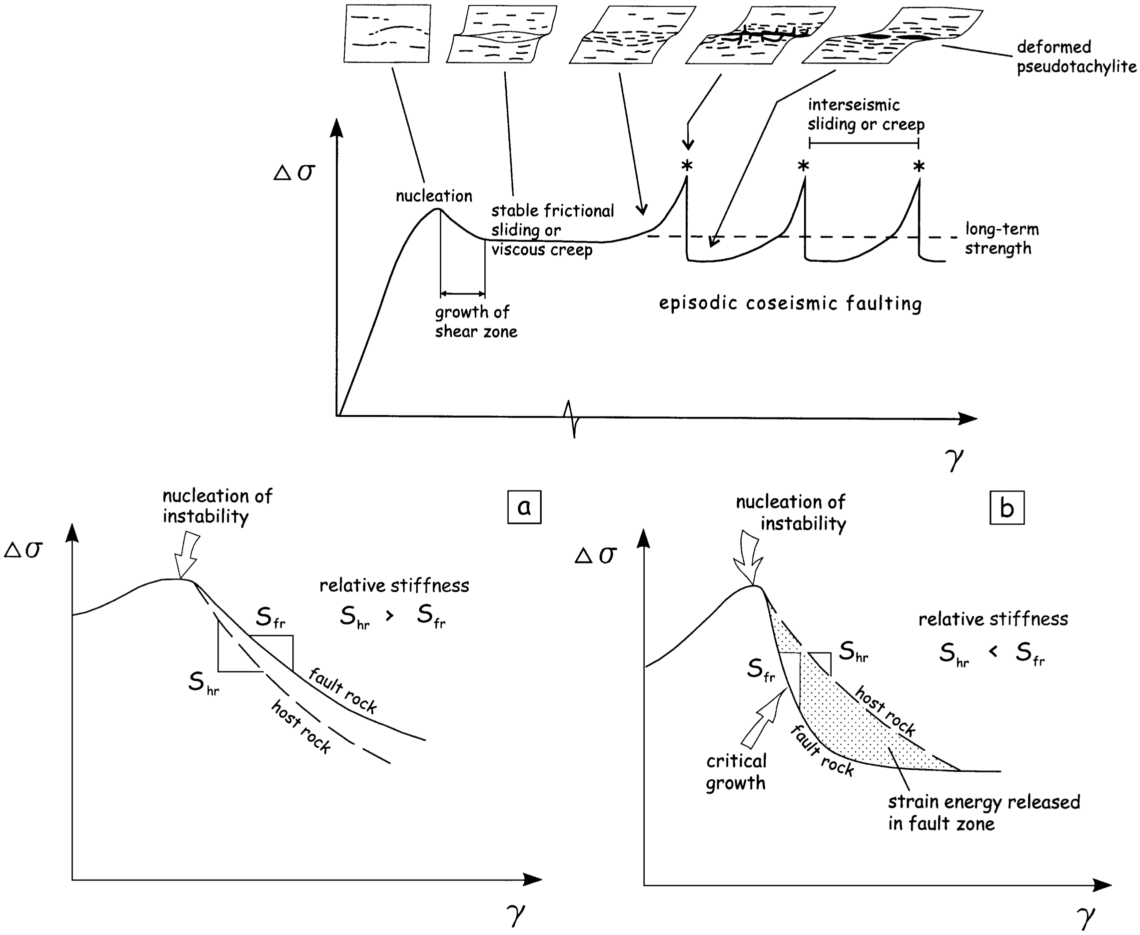



Fig. 7. Stress–strain evolution for a volume of rock undergoing deformation to frictional sliding or creep at a constant slip or strain rate. Insets above depict the corresponding structural evolution of fault rocks. Initial localization just before or upon attainment of peak strength leads to strength drop as shear zone grows. In this case, the host rock is stiffer than the fault rock, and the fault zone stabilizes (inset a). Unstable behaviour is associated with fluctuating strength and seismicity (asterisks) if the host rock is more compliant than the fault rock (inset b). Pseudotachylite forms at asperities in the fault zone and is overprinted during subsequent deformation. Strain energy released in the fault zone is proportional to the hatched area in (b) between the stress–decay curves for the host rock and the fault rock (see text).

notion that the current effective elastic thickness of the lithosphere, $T_e$, tracks with or slightly exceeds the thickness of the seismically active layer, $T_s$, in rifts and active convergent margins [1,5]. Indeed, the notion that flexed lithospheric plates have an elastic core which is able to support bending stresses on long geological time scales does not preclude local yielding within that core so long as the overall rate of weakening due to strain localization is balanced by the rate of loading due to convergence and tectonic accretion.

We note that proponents of the weak upper mantle hypothesis rejected the idea of seismicity induced by fault instability as "improbable and unnecessarily complicated" ([1], p. 496; [3], p. 7), presumably because seismicity was considered only in the context of the effective elasticity of lithospheric plates as gleaned from gravimetry. If one accepts the fault instability hypothesis, however, then the earthquakes relocated at the base of the foreland crust in the Alpine–Himalayan and central Asian mountains [2,3] are not indicative of greater lower crustal strength but of active thrusting along shear zones that are weaker than the adjacent un- or less deformed lower crust and upper mantle.

## 6. A model for seismicity and episodic crustal detachment

Fig. 8 summarizes the proposed evolution of crustal strength, rock structure, and rheology for one earthquake cycle in a generic strike-slip plate boundary fault. The model is based on the side-driven fault-slip model of Tse and Rice [59] but incorporates qualitative aspects of experimental and natural rock deformation cited in previous work and as discussed above. Although the Tse–Rice model applies only to a single strike-slip fault rather than a fault array, its basic characteristics have been extended to large thrust faults in subduction zones [65,66]. The main difference is that thrusts and normal faults are expected to show pronounced structural and rheological asymmetries between their hanging- and footwalls due to heat advected from the warm to the cold fault block.

Lithospheric strength during an earthquake cycle is both time- and displacement-dependent (Fig. 8): overall strength increases with progressive deformation to peak values at $t_1$ before coseismic stress release in the upper crust from $t_1$ to $t_2$ results in short-term loading of the underlying crust and upper mantle at $t_2$. The upper crust is subsequently reloaded as the overall strength of the lithosphere recovers and acquires its long-term profile at $t_3$.

The minimum in the $t_3$ long-term strength profile at about 10-km depth in Fig. 8a reflects syntectonic grain-size reduction at very high differential stresses prior to and during coseismic slip. Small syntectonic grain sizes favour the activity of grain size-sensitive creep mechanisms and induce a drop in solid-state

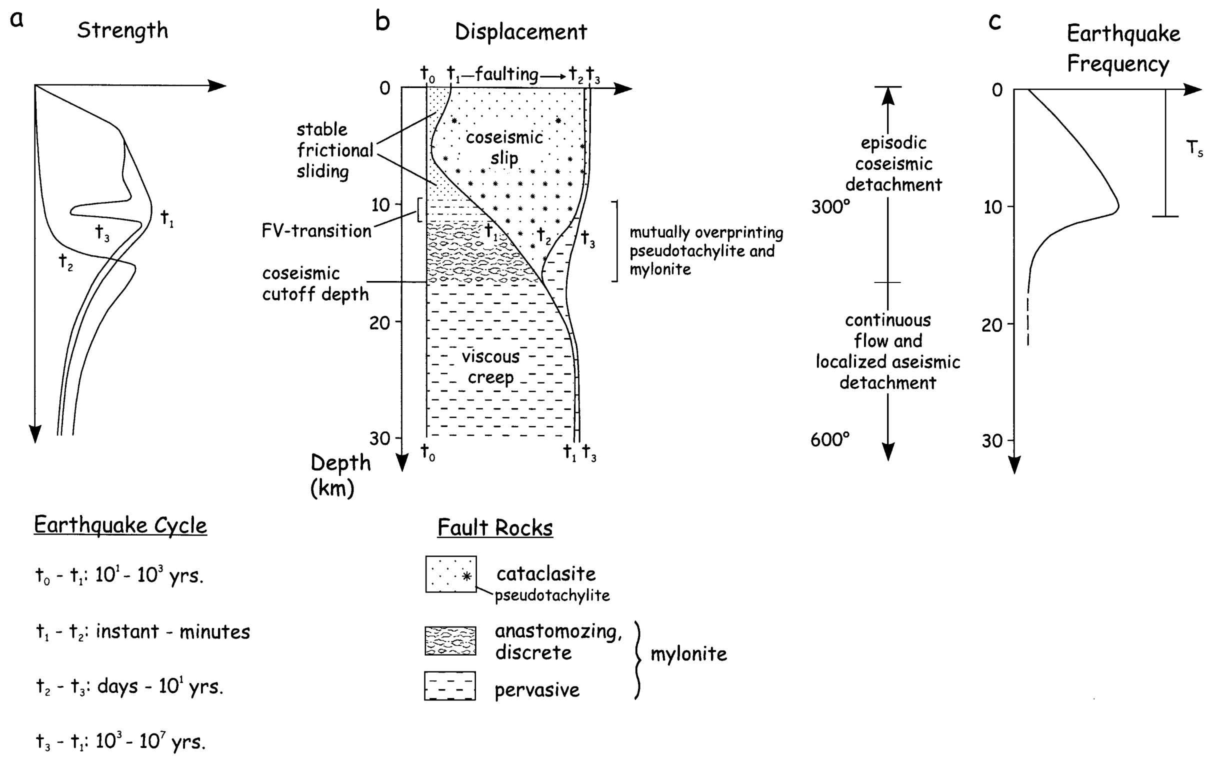



Fig. 8. Schematic plots of (a) strength and (b) displacement versus depth for an idealized strike-slip fault in continental crust during one earthquake cycle (b adapted from Fig. 7 in [58]). Each line in (b) represents the locus of points in the fault at a given time (one line each for times $t_o$ to $t_3$), but the time interval between any two neighboring lines is not identical. The crust can be subdivided into an upper layer of episodic, coseismic detachment (decoupling) and a lower layer of largely aseismic, viscous creep. Fault-rock nomenclature from [79]. The long-term strength of the crust increases slowly during the interseismic period beginning at $t_3$ and ending at the next major slip event (see text). (c) Generic earthquake frequency versus depth diagram, $T_s$ is thickness of the seismogenic zone.

viscosity [51]. It is important to note that the cutoff depth for coseismic slip at ca. 16 km exceeds the depth interval for the high-strain transition from frictional, cataclastic flow to viscous mylonitic creep (marked "FV-transition" in Fig. 8b) as well as the thickness of the seismogenic zone, $T_s$ (Fig. 8c). The base of $T_s$ is defined as the depth below which earthquake frequencies and magnitudes decrease rapidly [54]. Obviously, the extent to which coseismic slip propagates downwards into rocks that usually undergo aseismic creep depends on the slip magnitude at a given geothermal gradient and overall strain rate [59]. Downward coseismic rupture stops where velocity-strengthening processes (viscous creep) prevent stresses at the tips of shear zones from attaining the fracture strength of the surrounding rock. Fracturing and even seismicity can be triggered below the coseismic cutoff depth by seismic events above, especially if rocks are locally stressed to near their fracture strength, but seismicity below this depth is limited in magnitude and frequency.

The main point of Fig. 8 is that the cutoff depth for coseismic slip separates a lower lithospheric level in which aseismic viscous creep ensures intracrustal strain continuity from an upper level in which attachment is punctuated by episodic, coseismic slip and detachment. Intracrustal detachment can therefore be coseismic and episodic, as well as aseismic and long-term, as at the FV-transition or at first-order lithological contacts like the crust–mantle boundary.

The debate on lithospheric strength addressed in this paper therefore has interesting implications for the related issue of whether the strength of the continental crust is governed by crustal blocks and intervening subvertical shear zones [3] or by viscous drag of the underlying mantle. These scenarios correspond to "side-driven" [59] and "bottom-driven" [67] fault-slip models. In deeply eroded, ancient fault zones, uniformly oriented stretching lineations on alternately steep and predominantly subhorizontal foliations have been cited as possible evidence for basal attachment in bottom-driven systems [68]. Indeed, it has been argued that basal attachment is necessary to explain strain partitioning along oblique-convergent plate boundaries [69]. Molnar [69] also proposed the activity of subvertical shear zones in the strong upper mantle to account for a deficit in heat flow beneath the San Andreas Fault Zone. The possibility that seismicity is associated with domains of long-term weakness (rather than strength) may make seismicity a useful tool for identifying active parts of long-lived shear zones between and beneath crustal blocks in major fault zones. This is supported by recent high-resolution seismological [70] and theoretical studies [71,72] indicating that seismicity is concentrated at the interface between actively slipping and effectively locked segments of faults.

## 7. Conclusions

Even jelly can be food for thought as long as it is not consumed in excess! To a first approximation and for long geological times ($\geq 10^6$ a), the jelly sandwich model of steady-state rheological stratification is consistent with the location of weak zones within orogens and rifted continental margins as inferred from structural studies and from the geometry of reflection seismic images (Figs. 1–5). It also corroborates the competence contrast of naturally deformed rocks and minerals (Fig. 6). On all length scales, olivine-rich rocks of the upper mantle are stronger than quartz-, feldspar-, and mica-rich rocks of the crust, except in shear zones where the syntectonic grain size is sufficiently small to allow viscous grain-boundary sliding. However, the jelly sandwich model fails to explain why long-lived zones of weakness sometimes contain pseudotachylites, rocks that are indicative of ephemeral high-stress events including seismicity (Fig. 7).

Unfortunately, gravity studies of orogenic forelands provide us only with estimates of effective elastic thickness (or rigidity) on a time scale for which the current topography can be argued to have existed, at most, $10^6$–$10^7$ a [73]. Comparing these estimates with depth–frequency curves for current seismicity tells us little, if anything, about the strength of the continental lithosphere over longer periods of time and does not explain seismicity satisfactorily.

We argue that seismicity is induced by frictional and viscous instabilities arising from strain-dependent changes in the rheology of the fault rocks with respect to that of the host rock. Transient instabilities potentially nucleate in all levels of the lithosphere but especially in segments of active shear zones that are weaker than their surrounding rocks (Fig. 8). Steady state is therefore not a valid concept when applied to the short-term rheology of the lithosphere. It follows that seismicity is not a diagnostic criterion for rock strength.

Rather, earthquakes may be used to locate active weak zones of transient detachment and attachment within the continental lithosphere. When combined with studies of fault rock distribution and geometry in exhumed fault zones, this may be useful for constraining the direction and magnitude of forces acting on blocks within interplate fault systems.

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