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Weakness of the lower continental crust: a condition for delamination, uplift, and escape

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Abstract

We discuss three interconnected processes that occur during continental compression and extension: delamination of the lower crust and sub-crustal lithosphere, escape tectonics (i.e., lateral crustal flow), and crustal uplift. We combine calculations of lithospheric viscosity–depth curves with geologic observations and seismic images of the deep crust to infer the mechanisms controlling these processes. The basic driving force for delamination is the negative buoyancy (in some regions) of the continental lower crust and sub-crustal lithosphere with respect to the warm, mobile asthenosphere. A phase transformation in the lower crust from mafic granulite facies to eclogite may be important for providing negative buoyancy. Where negative buoyancy exists, the onset of delamination is mainly a question of the presence of a suitable decoupling zone between the denser lithosphere and the lighter upper and middle crust. We estimate the depth to potential decoupling zones by calculating lithospheric viscosity–depth curves based on reasonable geotherms and models of lithospheric composition. Low-viscosity zones occur at three depths: (1) at the base of the felsic (upper) crust; (2) within the lower crust; and (3) several tens of kilometers below the Moho. The commonly observed absence of a high-velocity (>6.8 km/s) lower crustal layer beneath extended crust may be explained by delamination wherein decoupling occurs at the top of the lower crust. In addition to being zones of potential decoupling, crustal low-viscosity zones are avenues for lateral crustal flow, a process that is often referred to as crustal escape (e.g., eastern Tibetan Plateau). The third process addressed here, crustal uplift, is mainly found in compressional environments and can be related to mature (i.e., complete or nearly complete) delamination and/or a thick low-viscosity lower crust. Mature delamination generates crustal uplift as the sinking, dense lithosphere is replaced by the mobilized hot asthenosphere. A very different mechanism of uplift is associated with some continental high plateaus, where a high convergence rate and the lateral intrusion of cold, rigid shield crust into warm, low-viscosity orogenic crust acts like a solid piston moving into hydraulic fluid. The displacement of the low-viscosity crustal 'fluid' generates broad plateau uplifts. Modern examples are the intrusion of the Indian shield into the Tibetan Plateau and the Brazilian shield into the Andes. All of these processes, delamination, tectonic escape, and uplift are interconnected and are related to weakness in the lower crust during continental compression and extension.

Keywords: rheology; lithosphere; dynamics; delamination

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1. Introduction

The idea of lithospheric delamination was developed in the 1970s and matured in the 1980s (Bird, 1979; Kay and Mahlburg Kay, 1986, 1993; England and Houseman, 1989). The fundamental arguments centered on the observations that (1) island arcs were supposed to be the main ingredients of continental crust, but (2) island arcs were basaltic and continental crust is intermediate in bulk composition. In the following years, improved knowledge of the composition of continental crust accumulated (Mooney and Meissner, 1992; Rey, 1993; Rudnick and Fountain, 1995; Christensen and Mooney, 1995), but still, the suspicion has remained that much of the lower crustal basaltic component was lost (delaminated) or at least strongly modified. One possibility is that the crust may be modified by anatexis of the lower crust, thereby leading to differentiation into a thicker granitic–gneissic upper crust and an ultramafic residue that forms a new seismic Moho. The resulting crust is both thinner and more sialic (Meissner, 1989).

In this paper, we define *delamination* as the decoupling process of continental lower crust and upper mantle from the overlying crust. Beneath thick crust, negative buoyancy within the lower crust may be strongly enhanced by the gabbro–eclogite transition (Rey, 1993). However, a weak lower crust is also necessary for decoupling and escape tectonics to take place (Kay and Mahlburg Kay, 1993). This is in fundamental contrast to the rheology of oceanic lithosphere, which is strong and rigid throughout, and is largely incapable of decoupling (Meissner and Wever, 1986). This difference in viscosity between continental and oceanic lithosphere is crucial for delamination, and for the deformation of continents at ocean–continent collision zones. The strongest part of the oceanic lithosphere is between the depths of 30 and 50 km. In contrast, this depth range corresponds to the weakest part of thick continental lithosphere (Molnar, 1988; Meissner and Wever, 1986).

The exact process of delamination is still a matter of debate, but it is agreed that negative buoyancy and decoupling must play dominant roles. Sometimes there is direct evidence for delamination, as in the case of the Alps, where seismic reflection studies image the subducting European lower crust and uppermost mantle (Nicolas et al., 1990; Heitzmann et al., 1991). Occasionally, seismic tomography reveals pronounced low velocities in the uppermost mantle, apparently caused by mobilized asthenosphere that has replaced delaminated lithosphere. However, often only indirect evidence for mobilized asthenosphere is available. England and Houseman (1989) mention three key indicators: (1) onset of mafic volcanism, (2) a regional change of the stress system to extension, which is related to (3) regional uplift. The presence of all three indicators provides strong evidence for mature delamination. A delay of several million years may occur between the onset of delamination and the appearance of hot asthenospheric material at the crust–mantle boundary.

Significant uplift is initiated by crustal shortening and thickening, leading to stacking of nappes along thrust zones in the rigid upper crust and to ductile thickening and indentations in the middle and lower crust (Meissner, 1996). Delamination, in its mature phase, might also significantly contribute to uplift (England and Houseman, 1989). In some areas, increased crustal buoyancy due to magmatic underplating and retrograde phase changes from eclogite to gabbro might also contribute to crustal uplift. These uplifted areas store a huge amount of potential energy that may lead to tectonic collapse.

The third process discussed here is escape tectonics. This concept was developed by Tapponier et al. (1982, 1986, 1990) to describe the eastward movement of the thickened Tibetan crust perpendicular to the direction of maximum shortening. In Tibet, many extensional and strike-slip faults indicate massive transport of (at least) the upper crust to the east and southeast (Westaway, 1995). Other regions where escape tectonics may take place include the eastern Alps (escaping eastward) and Turkey (escaping westward).

2. Lithospheric strength and weakness

The strength of the lithosphere has been compared with a jelly sandwich (Meissner and Strehlau, 1982; Zuber, 1994), with the weak lower crust being squeezed between the rigid upper crust and upper mantle. This rather simplistic picture has since been refined, based on laboratory experiments and new in-situ observations of seismicity, fault zones, and deep drilling (Karato et al., 1986). As rheology plays a critical role for delamination, uplift, and escape, rheological models, including friction and creep, will have to be critically reviewed.

When trying to relate data from frictional and creep experiments to natural conditions, it is necessary to extrapolate over many orders of magnitude in time. When comparing experimentally induced microstructures with natural ones, however, good correspondence is found (Rutter and Brodie, 1991; Hirth and Tullis, 1994). Modelling maximum stress (strength) in the upper crust and ductile behaviour in the lower crust, strength and viscosity display a peak (Fig. 1). From observations of crustal seismicity, it has been shown by Sibson (1982) and Meissner and Strehlau (1982) that a peak in strength at the base of the upper crust coincides with the peak in seismicity, where the depths of both peaks are mainly a function of temperature. Byerlee's law (Byerlee, 1978) for the approximate assessment of frictional strength of the upper crust, has been confirmed by many experiments as well as by many in-situ observations. Byerlee's law is dependent on pressure (depth), but is approximately independent of temperature up to about 350°C. At greater depths, a large transition zone follows where brittle and plastic processes are mixed (Karato et al., 1986). At the KTB (Deep Continental Borehole) drill hole in Germany, some intracrystalline creep processes are found to start at about 300°C. Open cracks were also found at this depth, and some seismicity was induced by hydrofracturing (Duyster et al., 1993).

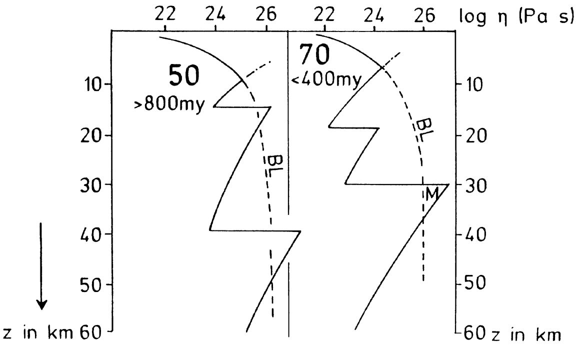

Fig. 1. Two simple viscosity–depth models (η–z) for a quartz-rich upper crust, a feldspar-dominated lower crust and an olivine-dominated mantle; logarithmic scale for η (after Eq. 4). Two heatflow provinces (50 and 70 mW/m^2) related to age provinces according to average crustal models, as often found in Proterozoic, Paleozoic, and Mesozoic stable areas. *BL* = Byerlee's Law (Byerlee, 1978); creep rate assumed to be 10^{-17} s^{-1}; *M* = mantle; lithospheric and rheologic models after Meissner and Kusznir (1987).

At naturally occurring strain rates, intracrystalline creep processes become important at about 350°C for quartz-containing rocks, at about 500°C for feldspar-dominated rocks, and at about 750°C for olivine-rich ultramafic rocks, depending on water or gas content, and on the particular mineralogy of the rocks (Strehlau and Meissner, 1987; Rutter and Brodie, 1991). These values are valid for 'normal' creep rates and are based on abundant experimental and microstructural observations of naturally deformed rocks, coupled with geothermometry, as summarised by Kirby and Kronenberg (1987). Observations of seismicity and the depth extent of fault zones seem to confirm the experimental data. The maximum depth of seismicity seems to coincide with the depth where creep processes start to dominate. Extensional seismogenic fault zones have not as yet been observed in the deeper continental crust (Meissner, 1996). For the oceanic (ultramafic) mantle, the curves of Parson and Sclater (1977) show that the deepest earthquakes coincide approximately with the 750°C isoline.

For the upper crust (and often for the uppermost mantle) we assume rock friction to be adequately described in the form:

$$\tau = \tau_0 + \mu^* \sigma \tag{1}$$

where τ = frictional shear strength, τ_0 = strength extrapolated to zero pressure, μ^* = frictional sliding coefficient, σ = confining pressure (depth).

At a certain temperature, however, temperature- and rate-dependent creep processes start to dominate when the stress associated with creep processes becomes smaller than those required to overcome friction or fracture. No more dynamic (velocity weakening) rupture can be generated (with the exception of large earthquakes, which often touch the lower crust but do not leave a permanent rupture zone). For the lower crust, and for the lower lithosphere and asthenosphere, Weertman's power law is widely held to be appropriate (Weertman, 1970; Meissner and Kusznir, 1987):

$$\dot{\epsilon} = C_n \tau^n \exp\left[-(E_c^*/RT)\right] \tag{2}$$

where C_n = a material constant; E_c^* = activation enthalpy for flow; R = gas constant; T = temperature in K; n = exponent $(1, \ldots, 4)$; τ = shear stress; $\dot{\epsilon}$ = creep rate.

This power law is only one of several ways that one might fit data points to a creep law. Other approaches are generally more complex. In the past decade, the addition of another term in Eq. 2 in the form of 'grain-size-sensitive' (GSS) creep, has been demonstrated by some researchers like Karato et al. (1986), Handy (1989) or Rutter and Brodie (1991). This additional GSS term, $d^{-\mathrm{m}}$, with d = grainsize, m = exponent, $0, \ldots, 3$, might assume a special importance if tectonic grainsize reduction occurs during high strain deformation. Other strain-weakening processes that are not grainsize-sensitive (e.g., dynamic recrystallization) can also occur at high strains, and favor localization of plastic flow into shear zones (Rutter, 1998). Such processes may weaken the crust still further after large deformation, but will not negate the general tenor of the following argument. Therefore, solving Eq. 2 for η with:

$$\eta = \frac{\tau}{\dot{\epsilon}} \tag{3}$$

one gets:

$$\ln \eta = (1/n)\left[(E_c^*/RT) - (1-n)\ln - \ln C_n\right] \tag{4}$$

In Eq. 4, the term for GSS-creep $(-m \times \ln d)$ might be added, if information on grainsize sensitivity of creep rate is available (Handy, 1989) or modifications could be made to take into account other strain dependencies.

Based on Eqs. 1–4, the idea developed that petrologic stratification of the crust and upper mantle will lead to stratification of mechanical/rheological properties. Many models were developed, starting with Sibson (1982), Meissner and Strehlau (1982), Chen and Molnar (1983), Handy (1989), Meissner (1989), Rey (1993), Singh and McKenzie (1993), Hirth and Tullis (1994).

Some parameters in Eq. 4 deserve special attention. First is the activation enthalpy, E_c^*, which can be related to rock type and deformation mechanism. A compilation of experimental data (Kirby and Kronenberg, 1987) was used to relate E_c^* approximately to the velocities of certain rock types, for example to quartz containing rocks with V_p around 5.8–6.3 km/s and $E_c^* = 130 \pm 30$ kJ/mol, to feldspathic rocks with V_p = 6.5 to 7.1 km/s and $E_c^* = 240 \pm 50$ kJ/mol, and to ultramafic rocks with V_p = 7.9 to 8.3 km/s and $E_c^* = 400 \pm 80$ kJ/mol (Meissner et al., 1991). The velocity structure of a certain area has to be known, to obtain an estimate of E_c^* values in this way.

A critical parameter for the evaluation of Eq. 4 is the temperature, T, and especially its depth dependence $T(z)$. One has to extrapolate heat flow values, q, to greater depth using conductivity equations as established by Chapman (1986) and others. Sometimes the limits of $T(z)$ can be assessed by seismicity, by seismic velocity values, or by evidence of recent xenoliths. An example of modeling Eq. 4 for two different heat flow provinces and crustal structures is given in Fig. 1. In similar models, the base of the quartz-bearing upper crust and the base of the feldspar-dominated lower crust stand out as prominent low-viscosity zones. They are the preferred candidates for detachment, delamination, and the outflow and inflow of material under appropriate stresses, for absorbing the effects of the intrusion of hard indenting material in a continent compressional collision, or for magmatic underplating and intrusions (extension).

As previously mentioned, there are several uncertainties, and modifications may be necessary for the simple models of Fig. 1. Some of these effects are shown in Fig. 2. In the brittle regime of the upper crust, calculations were performed with hydrostatic pore pressure. A higher value, as often is observed in situ, reduces the effective pressure (Oxburgh, 1972) and hence enlarges the brittle regime (Fig. 2a). The transition from the brittle to the intracrystalline plastic region has been found to be much smoother than shown in the simple models of Fig. 1 (Chester et al., 1993). Fig. 2b shows this approximation. Another modification is observed if delamination of parts of the lower crust and upper mantle has taken place, making the remaining crust sialic (Fig. 2c), and leaving the new lower crust full of seismic lamellae with very high reflectivity. The heating which follows delamination might lead to an accentuation of the differentiation process. The rising asthenosphere produces hot, mafic magmas which may underplate and weaken the remaining crust.

The largest uncertainty in the viscosity models is the creep rate. For our calculation, we have used a

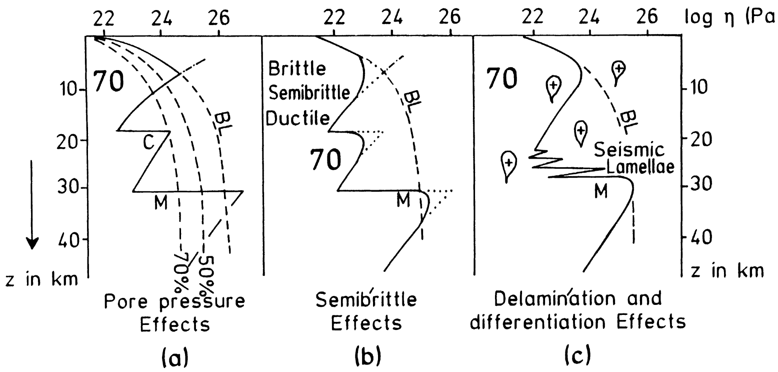

Fig. 2. Deviations from the simple models of Fig. 1 for a heatflow of 70 mW/m^2. (a) Effect of various pore pressures (% of lithospheric pressure) (Meissner and Wever, 1988). (b) Smoothing effect of semibrittle zones (Chester et al., 1993). (c) Effects of delamination (under collapse). Former mafic lower crust has disappeared and left a laminated, reflective zone with rather low average velocity (Meissner et al., 1991).

constant creep rate obtained from average plate tectonic movements. In the examples, this is 10^{-17} s^{-1}, but other constant values, such as 10^{-14} s^{-1}, do not change the shape of the curves, but only reduce the viscosity scale by about two orders of magnitude. As it is nearly impossible to predict the existence of many different crustal layers, each with limited thickness, or to calculate for possible thin near-horizontal shear zones, we can only conclude that it is possible for thin minima to exist with very low viscosity values, even several orders of magnitude lower than indicated in the figures (Figs. 1 and 2). An additional weakening effect could also be produced by the contribution of fluids. However, their existence in the lower crust is still hotly debated. Certainly the minima in our models are the preferred candidates for decoupling, delamination, and indentations. Thus, experimental, theoretical, and in-situ observations provide ample evidence that the lower crust, especially when thick and hot, is weak and ductile.

3. Delamination, uplift, and escape: examples from Tibet and other compressional orogens

All three processes, delamination, uplift, and escape, can be observed in zones of convergence, especially in continent–continent collisions. An example is the Tibetan Plateau (Fig. 3), with its average elevation of more than 4 km over an area of 5 million km^2. The collision of India with Asia, which started at about 60 Ma, continues with a rather constant convergence rate of 50 to 60 mm/yr. Even before 60 Ma, several terranes from the dispersal of Rodinia and Gondwanaland had been accreted to Asia (Dewey and Bird, 1970). By about 40 Ma, the Tethys Sea had completely disappeared, and the phase of strongest compression, uplift, and indentation began. Compressive structures and thrusts, shortening, and some thickening are observed until about 17 Ma, and E–W extension, with an accelerated uplift, started at about 14 to 11 Ma (Westaway, 1995).

There are several strong indications for delamination of the Indian lithosphere beneath the Tibetan Plateau. In the north, there was a sudden onset of mafic volcanism, a rotation of the stress system, and an acceleration of uplift at about 14 to 11 Ma. In addition, in northern Tibet, there appears to be a thin subcrustal lithosphere, as evidenced by reduced P- and S-wave velocities, strong absorption, and high Poisson ratio (Beghoul et al., 1993). These observations indicate a mature delamination process which has mobilized the asthenosphere (England and Houseman, 1988).

The Indian subcrustal lithosphere (ISCL), intruding from the south, has presently reached the middle of Tibet (McNamara et al., 1997). Data from various seismological studies indicate that there are high mantle velocities, low absorption, normal Poisson ratios, and a thick lithosphere (Beghoul et al., 1993). In southern Tibet (Fig. 4), some earthquakes occur in the uppermost mantle below an aseismic middle and lower crust. Evidence for delamination beneath

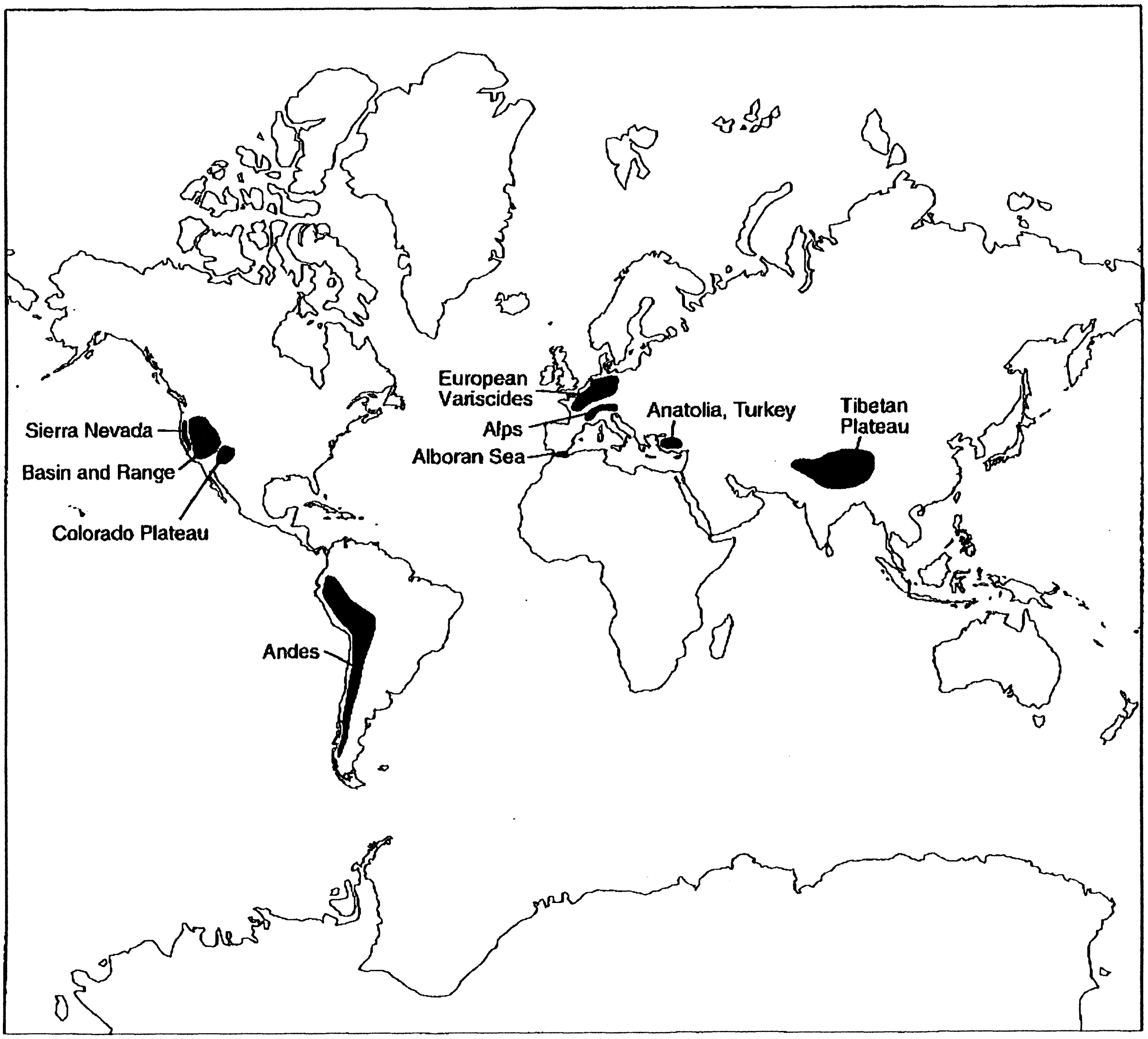

Fig. 3. Location map of regions discussed in this paper. Delamination of the lower crust and sub-crustal lithosphere, escape tectonics, and crustal uplift are widespread processes in both compressional and extensional environments.

Tibet is given by the high convergence rates which should have pushed the ISCL about 2000 km north of the Himalayas, which is considerably more than the 350 km that is inferred from geophysical data. About 1650 km of ocean and ISCL has been lost. Some material can be compressed or stacked, but certainly, the main part of the missing ISCL has delaminated by means of decoupling from a very hot and ductile middle and lower crust. This delamination has taken place beneath the Yarlung–Tsangpo Suture (YTS, the collision zone between India and Tibet), and/or farther north beneath the Bangong–Nuijang Suture (BNS). In addition, part of the ISCL may have delaminated laterally to the east, similar to the eastward escape of the upper crust (Tapponier et al., 1982, 1986). Because the middle and lower crust is extremely weak and appears to account for the accelerated uplift and escape of the Tibetan Plateau (Zhao and Morgan, 1985, 1987; Westaway, 1995), the idea of lateral escape of the ISCL below a decoupled crust is problematic. If the weak lower crust were completely decoupled, then an eastward escape

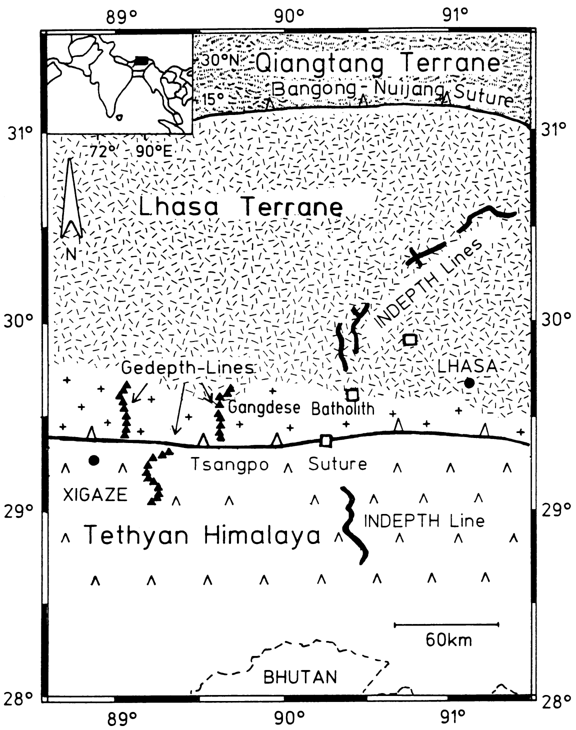

Fig. 4. Situation map of southern Tibet. Simplified geology after Tapponier et al. (1986), and Nelson et al. (1996). The two sutures are marked by ophiolites; location of INDEPTH II profiles (black lines) for obtaining structural information (Nelson et al., 1996); wide angle stations (black triangles) = GEDEPTH lines, and positions of three broad-band stations (open squares), for obtaining velocity information (Kola-Ojo and Meissner, in prep.).

of the upper crust and the upper mantle is difficult to explain. Perhaps the ductile middle and lower crust exerts a drag force on both the upper crust and the uppermost mantle. The observed S-wave anisotropy in the uppermost mantle, with its rotation of the fast S-axis to an E–W direction in the middle of Tibet, certainly supports the rotation at depth of the ISCL as well (McNamara et al., 1997). In any case, the disappearance of more than 1500 km of ISCL in northern Tibet suggests massive delamination.

In the overthickened Tibetan crust, the very ductile middle and lower crust also plays a critical role for uplift and escape. While the middle and lower crust provides a decoupling zone for delamination in the north and in the south, it apparently also plays an active role in the uplift and escape. Among the various explanations for the enormous and rather homogenous uplift of the plateau, the 'hydraulic pump' models of Zhao and Morgan (1985, 1987) and Westaway (1995) seem to provide the best explanation for the broad, flat plateau uplift, particularly for the last 14 to 11 Myr, where no compressive structures are observed on the plateau, and uplift seems to have accelerated. A rigid push from the south would have produced a topographically slanted, or at least irregular, uplift. However, the ductile 'inflow' of Indian crust from the south, with some compression below Tibet and an 'outflow' to the weak crust to the east, may explain the homogeneous uplift and the observed structures at the surface.

In order to calculate the viscosity below the Tibetan Plateau using Eq. 4, we need two relationships: the crustal or lithospheric velocity–depth structure, $V_p(z)$ (to estimate the activation energy E_c^*) and the temperature–depth relationship $T(z)$. A reliable picture of the lithospheric structure in southern Tibet (Fig. 5) has been obtained during the INDEPTH II investigations (Zhao et al., 1993; Nelson et al., 1996), also by wide-angle reflection studies (Kola-Ojo and Meissner, in prep.), and receiver functions from broadband seismograph stations (Makowski et al., 1997; Yuan et al., 1997; Fig. 5). Average P-wave velocities have been related to average activation energy E_c^*, according to Meissner et al. (1991). The temperature–depth relationship has been estimated based on the high heat-flow values in southern Tibet,

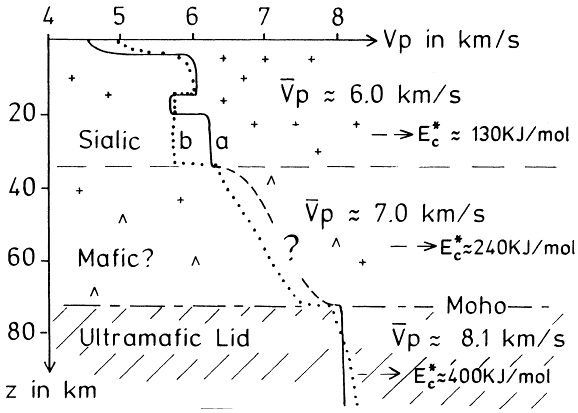

Fig. 5. Velocity models: (a) from the wide angle stations, after Kola-Ojo and Meissner (in prep.): (b) from the receiver functions (converted from S- into P-wave velocities, using a Poisson ratio of 0.25); after Yuan et al. (1997). Average V_p values related to average activation energies E_c^*, after Meissner et al. (1991).

the termination of seismicity at about 20 km, and by following conventional geotherms downward until the cold, high-velocity region of the Indian mantle is reached.

There are three principal arguments for a warm lower crust beneath the Tibetan Plateau: (1) On average, a thickened crust will assume a temperature similar to that of the ancient mantle that was located at a similar depth. A temperature of 800°C at 50 to 60 km depth is not unusual in the upper mantle. Crustal rocks at comparable depths will be much more ductile (due to lower activation energy) than mantle rocks. (2) There is considerable denudation and erosion of the uplifted Tibetan Plateau, thus deeper and warmer layers move vertically to shallower depth. (3) Delamination beneath northern and southern Tibet have transported hot asthenospheric material to the crust–mantle boundary before the cold ISCL intruded Tibet from the south. Fig. 6 is an assessment of the present temperature–depth relationship. It is based on a model for a slowly thickening crust, and considers as well some limits on temperature based on the observed seismicity in the upper crust and

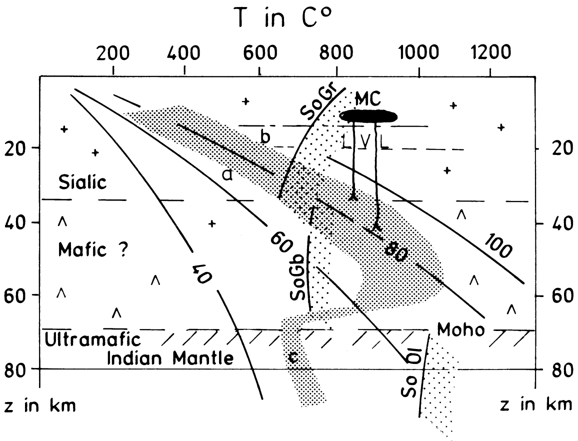

Fig. 6. Temperature model for crust and uppermost mantle below southern Tibet: (a) from seismicity (no rupture for temperatures greater than 400°C); observed seismicity limit is 20 km; (b) from reduced V_p and V_s values in LVL; (c) from sporadic earthquakes in uppermost mantle; (no rupture in mantle rocks above 700°C) and high V_p and V_s velocities (McNamara et al., 1997). M = Moho; MC = magma chambers; LVL = low-velocity channel (Nelson et al., 1996). 40–100 = geotherms for 40–100 mW/m^2 (Chapman, 1986); *SoGr*, *SoGb*, and *SoOl* are, respectively, solidi for granitic, gabbroic, and olivine-dominated (ultramafic) material.

in the uppermost (cold) Indian mantle. The largest uncertainty in this diagram is the composition of the lowermost crust. Because the velocity resolution of the available wide-angle reflection data and the receiver functions is not high, the composition of the lower crust is not well constrained. There may be a gabbro–eclogite transition, as suggested by the rather smooth receiver functions and the absence of near-vertical Moho reflections. Such material may be transported northward along with the cold ISCL and may be subjected to future delamination.

A viscosity–depth diagram with a constant creep rate was modeled (Fig. 7) based on the data shown in Figs. 5 and 6. The model of Tibet is compared with two other conventional models. The Tibetan Plateau shows much lower viscosities in the middle and lower crust than anywhere else on Earth. As mentioned above, the higher creep rates which are expected to occur within the minima of the viscosity–depth curves may have even lower apparent viscosity values than those indicated in Fig. 7. This viscosity calculation certainly supports the 'hydraulic pump' model, explaining the Tibetan uplift within the last 11–17 Myr.

The enormous uplift of the Tibetan Plateau can be compared only with the Altiplano in the Andes, although the Altiplano has a much smaller lateral scale. It is also thought to have evolved over a very long time span (at least since the Jurassic) of strong convergence in an ocean–continent col-

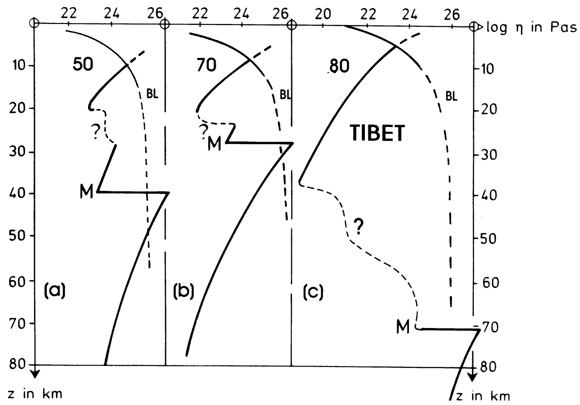

Fig. 7. Simple viscosity depth models for southern Tibet, as obtained from assessments of Figs. 5 and 6 (Eq. 4), compared with models of Fig. 1. Dashed line: uncertain composition; (a) 50 mW/m^2, (b) 70 mW/m^2, (c) 80 mW/m^2.

lision zone (Kay and Mahlburg Kay, 1986). Similar to the Tibetan orogen, an enormous mass of (oceanic) lithosphere has been guided beneath the (South American) continent. Thickening, shortening, erosion and delamination/subduction processes appear to have left the remaining thick crust even more sialic than in Tibet (Kay and Mahlburg Kay, 1993). Lateral escape seems to be absent, which might be a consequence of a narrow plate boundary and the narrow (two-dimensional) mountain belt of the Andes. Mass balance calculations, chemical analysis of volcanic rocks, and a regional absence of a subcrustal lithosphere indicate a recent massive delamination process (Kay et al., 1994). This delamination seems to be intimately connected with tectonic erosion near the trench, and with the drag of the 'flat slab' section of the subducting oceanic lithosphere, which here has removed large parts of continental lithosphere with it (Kay and Abbruzzi, 1996).

Other continent–continent collisions, such as eastern Turkey and the eastern Alps, display both considerable crustal uplift and escape tectonics. Eastern Turkey, with an average elevation of nearly 3000 m, shows a westward escape between the North- and South-Anatolian fault zones, and the eastern Alps, with an average elevation of about 2000 m, are escaping eastward (Ratschbacher et al., 1989). Both escape processes are directed towards the rheologically weakest areas in the surrounding region: Turkey moving toward the Aegean Sea, and the eastern Alps moving toward the Pannonian Basin. The initiation of delamination has been observed in the central and western Alps by detailed near-vertical seismic reflection studies by the Swiss Working Groups (Heitzmann et al., 1991), and by an ECORS–CROP investigation (Nicolas et al., 1990). Also, ancient collisions and accreted terranes, like in the European Variscides, were most probably accompanied by delamination, uplift, and escape, but because of post-collisional collapse, it is difficult to relate these processes to a specific tectonic phase. Strong thermal pulses during post-collisional times, possibly during mature stages of delamination, produced ample granitic volcanic activity in the Variscan internides, and left a surprisingly flat and thin sialic crust with a strongly laminated, and highly reflective, lower crust. Strong acoustic impedance contrasts, possibly from sill-like mafic intrusions, and a shear-induced ordering process are responsible for these continuous lamellae, which have been termed 'multigenetic' in origin (Mooney and Meissner, 1992).

4. Delamination during extension

The negative buoyancy that drives delamination during compression may also be present during extension. Delamination during the extensional collapse of an over-thickened crust is indicated in several regions. In order to maintain an isostatic balance, thick crustal roots beneath mountain belts must disappear during the process of orogenic collapse and erosion. However, the mechanism behind this disappearance is not clear. We note that in the late- or post-orogenic stage (i.e., during extensional collapse) a strong heat pulse often occurs that generates extensive granitic plutons. Lithospheric delamination seems to be a likely explanation for this heat pulse (Meissner, 1989).

Another indication of delamination in highly extended areas is the usual existence of a flat and shallow Moho with high lower-crustal reflectivity and relatively low ($\sim$6.5 km/s) seismic velocities near the base of the crust. These velocities are more typical of the continental middle crust, which suggests that the lower crust and subcrustal lithosphere were delaminated. Alternatively, anatexis that causes internal differentiation of a previous mafic lower crust into a thicker (granitic) upper crust and an ultramafic residue (which seismically correlates with shallow mantle) might also explain the presence of a thin crust without a pronounced, high-velocity lower crust. The heat for such anatexis may be provided during the mature stage of delamination of the subcrustal lithosphere.

These considerations support the suggestion that the transition from compression to extension is a critical period in the evolution of mountain belts (Rey, 1993). The change in the stress system might be due to the collapse of mountain belts together with the delamination of a crustal root and a lateral escape of crust. Mountain belts collapse under their own weight (Nelson, 1992), and lateral extrusion is powerful and dominant beneath high (compressed) plateaus (Bird, 1991). At this point, the tectonic development of collapsed mountain belts apparently

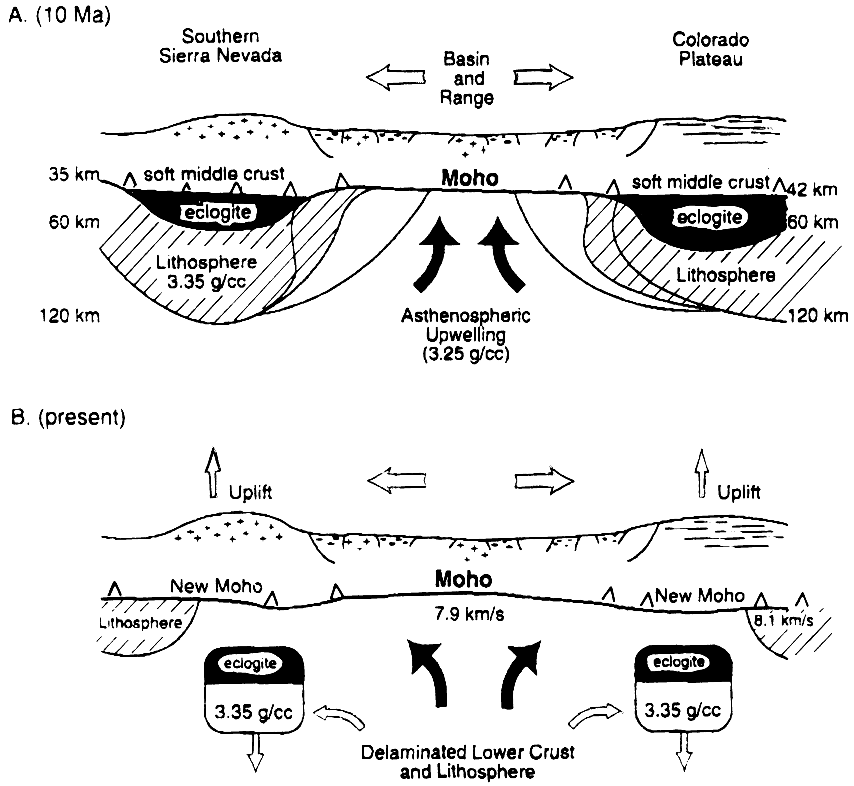

Fig. 8. Delamination of the lower crust and lithosphere during extension, as hypothesized for the Sierra Nevada, Basin and Range, and Colorado Plateau of the western United States (Fig. 3). (A) During crustal extension in the Basin and Range Province, asthenosphere upwells and thins the lithosphere. (B) A crustal low-viscosity zone is created by the heating, and delamination of the lower crust and sub-crustal lithosphere occurs on the flanks. The presence of some eclogite in the mafic lower crust is indicated by eclogite within crustal/upper mantle xenoliths found in this region.

follows two different paths: delamination is hypothesised to cause a second uplift, together with modest crustal thickening, heating and mafic volcanism (Bird, 1979; Houseman et al., 1981; England and Houseman, 1989). Others see delamination as the main cause for collapse and thinning, together with a powerful lateral extrusion under extension, producing flattening, stretching, and layering in the hot and ductile lower crust (Nelson, 1992; Rey, 1993; Meissner, 1996).

There are examples supporting both the thickening and the thinning models, and plate tectonic forces will determine the final outcome of the delamination process. If delamination occurs within a zone of strong and continuous convergence, such as in Tibet or the Altiplano, dynamic and possibly hydraulic forces keep the crust thick and buoyant, in spite of escape tectonics. If delamination occurs towards the end of an orogeny when compressive forces decay, then delamination and escape accelerate the collapse. The crust is heated, stretched, differentiated, and thinned. An example may be the end of the compressional period in the European Variscides when Pangaea was nearly completely assembled.

The final product of delamination is not always as clear as described here. The tectonic consequences of delamination may be different in adjacent areas, depending on the stage of the delamination process. One example is the Sierra Nevada in the western U.S. (Fig. 8). Here the thinning of a 65-km-thick crust to 35 km, accompanied by a nearly complete removal of a relatively cold 70 to 80 km thick mantle lid, is invoked by Wernicke et al. (1996). Also, the adjacent Basin and Range Province, with its high topography and thin, strongly laminated crust, is hypothesized to be an extensional member of a massive delamination process. Still farther east, the uplift of the Colorado Plateau may have been initiated by a delamination event at 30 Ma, followed by a second event at 5 Ma (Bird, 1979). In the Alboran Sea and the Rift–Betic Mountains, an early stage of delamination is inferred. Results from seismic tomography and gravity data indicate the existence of a delaminated lithospheric body between 70 and 150 km depth below the Alboran Sea (Seber et al., 1996). The overlying crust, at this stage of delamination, is pulled down, creating depressions in northern Morocco and southern Spain.

Both the magnitude of tectonic forces (continuous convergence vs. gradual termination) and the maturity of delamination play significant roles in the tectonic and magmatic expression of this process. During the onset of delamination, relatively little asthenospheric material is mobilized, probably not enough to stop the tendency for sinking, induced by the downgoing lithosphere (Fig. 9A,B). Only in the mature stages of delamination, when a large amount of hot asthenospheric material has been mobilized (Fig. 9C), will there be strong uplift, magmatism, and a certain change in the stress system. Alternatively, in extension, the hot asthenosphere might spread out laterally, thereby stretching, warming, and thinning the crust.

No clear examples seem to exist for delamination in a continuously extended regime. Theoretically, mafic material of a plume or some other asthenospheric upwelling could accumulate below the crust or below a thin lithosphere. If such a process is limited in space and time, the added material cools and is possibly transformed to eclogite. This process might pull the crust down, and might induce a rather vertical delamination event which could explain the origin of some basins, like the mysterious development of the Michigan basin or the North German basins. It has been suggested that such a cooling plume process is responsible for global recycling of the whole lithosphere on Venus, thus explaining the average surface age of only 500 Myr.

It is difficult to estimate the typical dimensions of lithosphere that has delaminated. Kay and Mahlburg Kay (1993) hypothesize a relatively small slice of about 120 × 30 km following the trend of shallow subduction below the central Andes (Fig. 9C). The sinking lithospheric unit of Seber et al. (1996) has

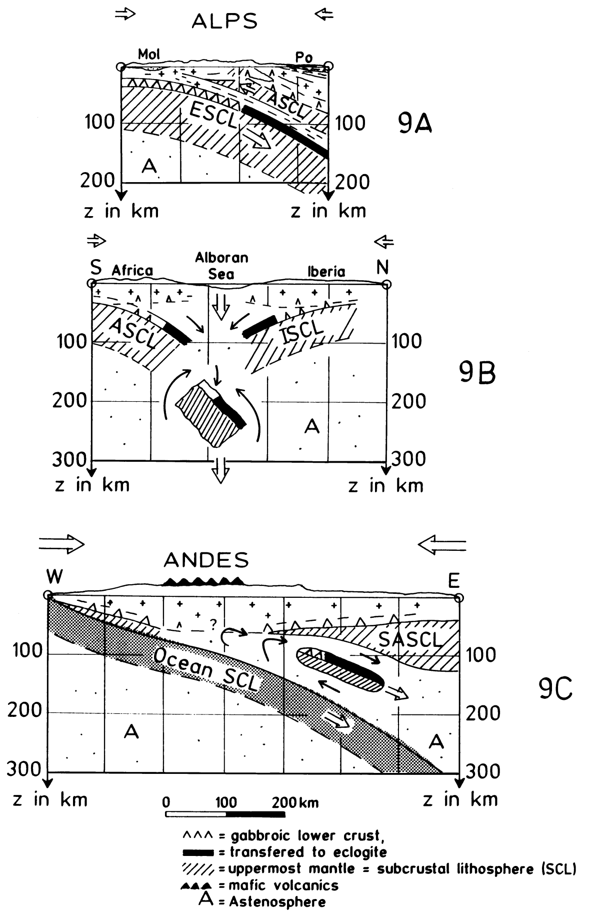

Fig. 9. Various stages of delamination under compression (where direct evidence is available, cartoons are simplified). Gabbroic lower crust supposed to be transferred to eclogite facies below about 70 km depth (black signature). (A) The Alps, after Nicolas et al. (1990) from CROP–ECORS reflection and refraction seismics (and gravity). Small convergence rate (∼1 cm/a). *Onset of delamination*. Part of the European subcrustal lithosphere (*ESCL*) = LC + UM dives beneath the Adriatic subcrustal lithosphere (*ASCL*). (B) The Alboran Sea (western Mediterranean) from various seismicity (and gravity) studies after Seber et al. (1996). Small convergence rate (1 cm/a). *Delamination in progress*. A piece of LC + UM on its way down. The African and the Iberian subcrustal lithosphere (ASCL and ISCL) + lowermost crust are bending down. (C) The Altiplano (Andes) from various geophysical studies after Kay and Mahlburg Kay (1993). Strong convergence rate (∼10 cm/a).

a dimension of at least 100 × 200 km. Beneath Tibet, even larger lithospheric slabs are missing and thought to have delaminated (Kola-Ojo and Meissner, in prep.). Other sizes mentioned so far are less well constrained, but it seems that delaminating lithosphere has dimensions of about 50–500 km in length (or width) and a thickness of 50–200 km.

5. Summary and conclusions

In this paper, we discuss three processes that occur during continental compression and extension: delamination of the lower crust and sub-crustal lithosphere, escape tectonics, and crustal uplift. Although many details concerning these processes are uncertain, geologic and geophysical data provide ample evidence for the occurrence of these processes. We combine calculations of lithospheric viscosity with geologic observations and seismic images of the deep crust to infer the mechanisms controlling these processes.

The basic driving force for delamination is the negative buoyancy of the lower continental lithosphere with respect to the warm, mobile asthenosphere. The negative buoyancy of the oceanic lithosphere as it cools and thickens is evidenced by the existence of deep ocean basins and subduction. In contrast, continental lithosphere is often older than oceanic lithosphere, and has a more complex thermal and magmatic evolution. Consequently, its density is likely to be highly variable, and in some places it may be neutrally buoyant due to the removal of basaltic melts from the mantle (Jordan, 1975, 1988). However, the lack of topography in regions of great crustal thickness (e.g., 60 km beneath southern Finland) demonstrates that dense lower crustal/lithospheric roots exist in some regions. A phase transformation in the lower crust from mafic granulite facies to eclogite facies is an example of a mechanism that would provide negative buoyancy.

Where negative buoyancy exists in continental lithosphere, the occurrence of delamination is mainly a question of the existence of a suitable zone of decoupling between the lighter portions of the crust and the denser lithosphere. Decoupling may occur in weak (low viscosity) zones within the crust or sub-crustal lithosphere. We have estimated the depth of these zones by calculating lithospheric viscosity–depth curves based on published geotherms and reasonable assumptions of lithospheric composition. Continental geotherms may be estimated on the basis of heat flow measurements, and lithospheric composition may be inferred from seismic velocity determinations and the evidence from crustal and upper mantle xenoliths. These lithospheric viscosity–depth calculations show low-viscosity zones at three depths: (1) at the base of the felsic upper crust; (2) just above the Moho within the lower crust; and (3) some tens of kilometers below the Moho. The crustal low-viscosity zones, in addition to being zones of decoupling, are avenues for lateral crustal flow (crustal escape), as hypothesized for both compressional orogens (e.g., Tibet; Tapponier et al., 1982) and extensional regimes (e.g., Basin and Range, western U.S.; Gans, 1987; Wernicke, 1992).

Crustal uplift can be related to both delamination and lithospheric strength. Mature delamination leads to crustal uplift as the dense lithosphere is replaced by lower-density asthenosphere, as in the Basin and Range Province, Colorado Plateau, and Sierra Nevada of the western U.S. A very different mechanism of uplift is associated with continental plateaus and some mountain belts. During long-active compressional orogenies, the intrusion of cold, rigid crust into crustal low-viscosity zones acts like a piston moving laterally into hydraulic fluid. The displacement of the crustal ‘fluid’ within the low-viscosity zones generates some of the most remarkable topographic features of the Earth, broad plateau uplifts. Modern examples are the intrusion of the cold crust of the Indian shield into the warm, thick crust of the Tibetan Plateau (Zhao and Morgan, 1985, 1987) and the intrusion of the Brazilian shield into the middle crust of the Andes.

Delamination of the lower crust and sub-crustal lithosphere, escape tectonics, and crustal uplift are interconnected processes that are the consequence of low viscosity within the lower crust. These processes are active during continental compression and extension, and all three processes have played a major role in determining the deep structure, composition, and evolution of the lithosphere.

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