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Delamination and delamination magmatism

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ABSTRACT

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Lithospheric delamination is the foundering of dense lithosphere into less dense asthenosphere. The causes for this density inversion are thermal, compositional, and due to phase changes. For delamination to occur in the specific, and probably common, case where lithospheric mantle is intrinsically less dense than underlying asthenosphere due to compositional differences, a critical amount of shortening is required for the densifying effect of cooler temperature to counterbalance the effect of composition. Crustal thickening that results from shortening may result in a crustal root that, due to phase changes, becomes denser than the underlying mantle lithosphere and should delaminate with it: most of the negative buoyancy resides at the top of the mantle and the bottom of the crust. In most cases composition is not known well enough to calculate the driving energy of delamination from densities of equilibrium mineral assemblages in a lithospheric column. Poorly known kinetics of phase changes contribute additional uncertainties. In all cases however, the effects of delamination under a region are readily recognizable: rapid uplift and stress change, and profound changes in crustal and mantle-derived magmatism (a reflection of changes in thermal and compositional structure). Characteristics of delamination magmatism are exhibited in the Southern Puna Plateau, central Andes. The consequences of delamination for theories of crustal and mantle evolution remain speculative, but could be important. Recognition of delamination-related magmas in older (including Archean) orogens may be the best way to recognize past delamination events, because the magmas are among the most indelible and least ambiguous of delamination indicators.

Introduction

Recently, lithospheric delamination has become a popular albeit controversial explanation for rapid regional uplift and extension, accompanied by lithospheric thinning and increased magmatic production (Bird, 1979; and see Kay and Kay, 1991, for a recent review). As used here, delamination involves the rapid foundering of lower lithosphere (mantle – but not excluding crust) into asthenospheric mantle. Potential energy that drives the process (e.g., Molnar and Lyon-Caen, 1988) is released as hot, low-density asthenospheric mantle replaces cold, dense lithosphere. For the purpose of the present discussion, which focuses on the consequences of the process, there is no essential difference between delamination (e.g., Bird, 1979) and convective instability of the boundary layer at the base of the lithosphere (e.g., Houseman et al., 1981; Fleitout and Froidevaux, 1982; Yuen and Fleitout, 1985; Molnar, 1990).

The physical basis for delamination has been inadequately formulated in existing models which rely on thermal expansion to generate density changes. Both compositional variability within crust and mantle, and phase changes easily generate density differences exceeding those produced by thermal expansion alone. In particular, the density of peridotite in the mantle lithosphere may be relatively low due to extraction of basalt.

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This low intrinsic density (e.g., density at a specified pressure and temperature) may stabilize lithosphere against delamination (Oxburgh and Parmentier, 1977; Jordan, 1988). In order for models to contain the essential physics of the system, which is driven by density instability, density must be treated more realistically than it is in existing models.

The observational basis for delamination is indirect, and observational opportunities are limited because delamination events are of short duration, and the resulting thermal and mechanical structures of the crust and mantle are ephemeral (e.g., Zhao and Morgan, 1987; Bird, 1991). In the recent past to present, delamination events appear to be confined to a very few places: sites are thought to exist in the Basin and Range, Tibet, and the Andes (see Dewey, 1988). However, at some times in the past (times of supercontinent accretion: Kay, 1993; and in the Archean) there are indications that delamination may have been more common than it is at present.

Often, magmatism is associated with and is a consequence of delamination. This delamination magmatism records thermal and compositional conditions in both mantle and crust, conditions that are necessary inputs for delamination models. The magmatic record is perhaps the most indelible of all evidence of delamination, being expressed and preserved in various tectonic levels in old as well as young orogens. Some delamination-related magmas are compositionally similar to magmas generated by melting of young, hot subducted oceanic crust (e.g., Defant and Drummond, 1990). The tectonic context serves to discriminate between the two types, with delamination magnetism associated with thick continental crust.

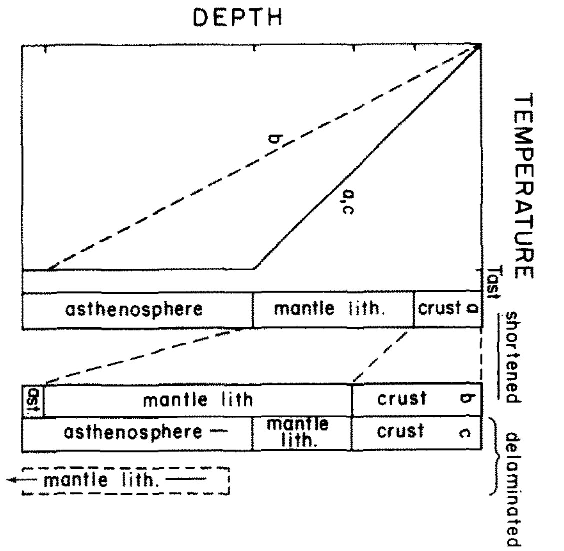

Fig. 1. Geotherms and lithospheric columns (figure modified after England and Houseman 1989) showing that shortening at constant volume cools (and therefore densifies) lithosphere, and that delamination raises the base of the thinned lithosphere to asthenospheric temperature (T_{ast}). The delaminated mantle lithosphere is shown outlined by dashed borders. Its place in section C is occupied by asthenospheric mantle.

The physical basis for delamination

The mantle and delamination models

The main rationale for invoking rapid delamination of the lower continental lithospheric mantle is a rapid ("catastrophic" - Dewey, 1988) increase in mantle temperature inferred from rapid uplift and an increase in magmatic activity. Delamination involves the abrupt change in thermal structure by advection of hotter, deeper (and geochemically distinct) asthenospheric mantle to shallow depths. An alternative to delamination, thermal thinning with no advection of heat, is unrealistic, as both conduction and radioactive heating are much too slow (e.g., Emerman and Turcotte, 1983; Yuen and Fleitout, 1985) to heat and soften large volumes of lithospheric mantle (converting them to asthenosphere) over short time intervals corresponding to observed tectonic transitions in orogenic zones. Often this advection occurs with minimal lithospheric extension. Generally, some regional extension follows, but none precedes, the delamination event.

Figure 1, after England and Houseman (1989), illustrates some of the assumptions that have become standard in delamination models (e.g., Bird and Baumgardner, 1981; Turcotte, 1989; Schmeling and Marquart, 1991). In these models, potential energy is released on replacement of lowermost mantle lithosphere by hotter and therefore less dense, underlying asthenosphere.

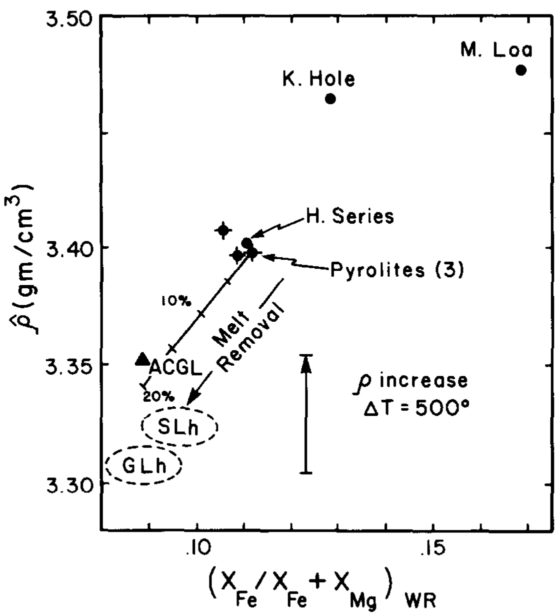

Fig. 2. Normative density (calculated at room *P* and *T* from bulk composition in the garnet lherzolite metamorphic facies) against Fe-number (molar) for mantle peridotites (figure modified after Jordan, 1979). Effect of melt removal is to decrease peridotite density: starting with a "Pyrolite" composition, removal of 10% melt is shown to be equivalent to a couple of hundred degrees of thermal expansion. *SLh* and *GLh* are spinel lherzolite and garnet lherzolite averages (samples have various Fe numbers) from Hawkesworth et al. (1990). Other points on the diagram after Jordan (1979) are as follows: *ACGL* is average continental garnet lherzolite; Kilbourne Hole (K. Hole), Mauna Loa (M. Loa), Honolulu series (H. Series) are estimated mantle compositions, and pyrolites are various mixtures (representing undepleted mantle) of Ringwood.

In the case depicted in Figure 1, the mean temperature is approximately 300°C greater for the 50 km of asthenosphere that replaces lithosphere. For a thermal expansion coefficient of 3×10^{-5} deg^{-1}, this represents an expansion of about 1% or 500 m due to delamination alone. This calculation assumes that the asthenospheric mantle and lithospheric mantle have the same composition, and therefore the same density when compared at the same *P* and *T*. The flaw in the standard delamination models is that they assume that all mantle peridotites are chemically identical. As shown in Figure 2, chemical differences among mantle peridotites result in density differences that are comparable to those resulting from temperature. Jordan (e.g., Jordan, 1979) has repeatedly emphasized how in models of lithospheric stability, where density is a critical parameter, temperature and compositional effects on density can be made to compensate. Jordan's reasoning is equally applicable to delamination.

Three processes that change mantle composition are likely to account for lithosphere–asthenosphere density contrasts. The first is removal of melt from peridotite that has crossed the solidus during advection from depth. Residual peridotite is less dense than its parental peridotite prior to removal of melt (see Fig. 2), and is likely to remain at shallow levels of the mantle (Oxburgh and Parmentier, 1977; Kay, 1980; Hawkesworth et al., 1990). On cooling, the residual peridotite would become mantle lithosphere. The second is addition of water. Addition of 1.5% water reduces peridotite density by about 1%. Both melting and water addition occur at convergent plate margins (magmatic arcs), and should exert a stabilizing influence on the arc mantle lithosphere. The third is intrusion of melts (broadly, basaltic) into the mantle. If the melts solidity to eclogite, every 10% eclogite added increases density by about 1%.

In the common cases where the combined thermal and compositional effects result in mantle lithosphere that has a lower density than underlying asthenosphere, the lithosphere will be gravitationally stable on extension. During compression however, lithosphere will become unstable, but only after a critical amount of shortening: the amount required depends on the difference in intrinsic or normative densities (e.g., at the same pressure and temperature – see Fig. 2) between asthenospheric and lithospheric mantle. This requirement for a critical amount of shortening is consistent with geological observations: as noted by Dewey (1988) compressional thickening precedes extensional collapse of many orogens. Dewey (1988) also notes that the lithospheric thinning that is compemporaneous with this extension often exceeds that expected from extension alone, and therefore the thinning has an important delamination component. The concept of a critical percentage shortening (shown on Fig. 3) offers a rational explanation for the critical

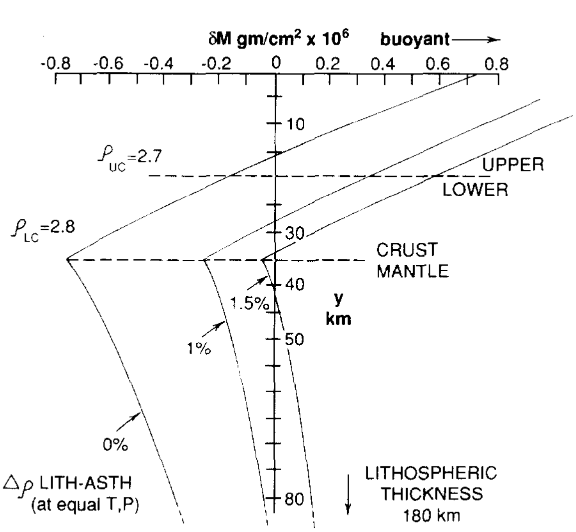

Fig. 3. Buoyancy mass per unit area of 180-km-thick mantle and (at depths less than 35 km) crustal lithosphere below a depth, as a function of that depth (as in Turcotte, 1989). Normative density (see Fig. 2) of asthenospheric and lithospheric mantle are assumed to be the same for the curve to the farthest left. Other curves are for lithosphere whose normative density is 0.033 and 0.0495 g/cm^3 (1% and 1.5%) less than asthenosphere, cases that have more buoyancy. Calculations assume a linear temperature gradient in the mantle lithosphere, from a 1300°C base to a 320°C Moho. Curves are continued into the crust assuming that a lower most crust that is 0.5 g/cm^3 less dense than the immediately underlying mantle, although for thicker crust, the lower crust may be denser than the mantle (see text).

value of the thickening factor that England and Houseman (1989) propose as a prerequisite to lithospheric instability (e.g., delamination).

An additional consequence of attaining a critical thickening factor is that lithosphere thickened in excess of this critical value will delaminate wholesale, because the lower part alone is stable (see Fig. 3). This particular behavior does not follow from the standard delamination models, that neglect compositional effects. In these models, all mantle lithosphere is unstable, and the thickness of the mantle lithosphere that delaminates is chosen arbitrarily (e.g., Houseman et al., 1981; McKenzie and O'Nions, 1983).

The lower crust and delamination models

We assume that the lower crust does not initiate delamination events, but it may well respond to them. The response depends on lower crustal density, which determines the direction of advection of lower crustal material. In all cases, delamination involves heating of the lower crust, due to rapid advection of hot asthenospheric mantle to shallow depths. Based on experimental investigations, density can be calculated for a given crustal thickness and composition (although the sluggish kinetics of mineral reactions must be considered). There are two general cases: thin and thick crust, corresponding to low and high pressure conditions in the lowermost crust. For regions with thin crust (< 50 km), like the Basin and Range, (Bird, 1988) lower crust of any composition continues to have a lower density than the mantle. In this case, lower crust is thought to flow vertically and laterally (e.g., Bird, 1991; Schmeling and Marquart, 1991), especially if it melts. Unless it is entrained by underlying delaminating mantle lithosphere (e.g., Turcotte, 1989), material of the lowermost crust remains in the crust. In contrast, for regions where compression has thickened the crust to > 50 km basaltic composition rocks in the lower crust undergo large density increases due to "eclogitic" phase transitions (see Fig. 4 and Austrheim, 1990). As pointed out by Sobolev and Babeyko (1989) crustal thickness is limited to a maximum value by these phase transitions because basaltic composition rocks are classified (by density and seismic velocity) as "crust" if they have gabbroic mineralogy, but as "mantle" if they have eclogite mineralogy. For a quartz tholeiite composition, Figure 4 shows that density changes by phase transitions (over 10%) far exceed those possible by thermal expansion. If the basaltic lower crust partially melts in response to the temperature increase resulting from delamination of the underlying mantle lithosphere, and if the melt migrates upward, the residue will be even denser (see Fig. 4).

Returning to the question of how much lithosphere delaminates, we now see that not only does the shallowest lithospheric mantle furnish the greatest negative buoyancy (Fig. 3), but that the lowermost part of thickened (> 50 km) crust, if it is basaltic, contributes negative buoyancy as well. In regions with thick crust, if lithospheric mantle delaminates, lowermost crust should also.

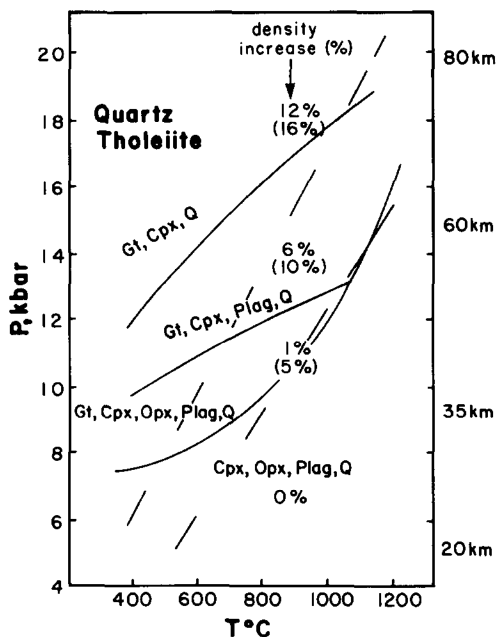

Fig. 4. Mineralogy of quartz tholeiite (basalt) composition at conditions corresponding to the lower crust in thin (20 km) to thick (80 km) crustal sections (diagram after Wood, 1987). Also shown are density increases expressed relative to lowest pressure mineral assemblage (which has a density of 3.07 g/cm^3), calculated from data listed in Wood (1987). The percentages in parentheses are the estimated densities of crystalline residues remaining after removal of 15% silicic melt (T greater than about 800°C is probably required for this) (Rapp and Watson, 1993). Note that density of the highest pressure assemblages is 10–15% higher than 2.9 to 3.1 g/cm^3, gabbroic density, and is comparable to or exceeds values for mantle peridotite (3.32 g/cm^3).

Generally, standard delamination models have neglected density changes within the mantle and the crust due to composition and phase changes. The forgoing is meant to alert the reader to the probability that these effects are important – even dominant – when compared with thermal expansion effects. If so, it can hardly be claimed that the physics of the delamination problem has been adequately portrayed.

Rheological controls on delamination

If the gravitational potential energy requirement for delamination is met, it does not immediately follow that delamination will occur. A second necessary condition is that the whole lithospheric section detaches and sinks. Thus, for completeness, a definitive statement about lithospheric rheology is required for the whole process to be plausible. A convincing rheological framework is not available without a very exact, and generally unattainable, statement about compositional heterogeneity (this is especially, but not exclusively, a problem in the crust). Even then, as emphasized in a recent review by Rutter and Brodie (1992), the application of laboratory data to the generation of large-scale structures, required for delamination, is fraught with uncertainty. For instance, according to Rutter and Brodie (1992) ".... where the entire thickness of the lithosphere must rupture, steady-state creep laws for intracrystalline plasticity of coarse-grained rocks are probably of limited applicability". Fortunately, for the delamination hypothesis, the factors influencing failure (e.g.,, fluids, stress amplification) make failure easier than otherwise predicted. An additional consideration is that if lithospheric thickening immediately precedes delamination, the lithosphere has undergone large-scale flow or failure just prior to delamination. With stress levels of similar magnitude for thickening and delamination, if the former occurs, so should the latter.

The observational basis for delamination

General observations

While the physical basis for delamination is inadequate for prediction of delamination events, the observational criteria are somewhat better for identifying where delamination has occurred. A sequence of observable structural, thermal and magmatic events have been recognized as consequent of delamination. Models have been regarded as successful if they reproduce some of these events. Observation of one or more of the events has led to claims of delamination under the Michigan Basin, Pannonian Basin, and the Apennines, as well as Tibet, the Basin and Range and the Puna.

For example, the Tibetan Plateau models of England and Houseman (1989) show that a recent period of uplift and E–W extension in this region, which is under regional compression, is accounted for by delamination (using the terminology of Bird, 1979) of underlying mantle lithosphere. Temperature increases sufficient to melt both mantle and crust are predicted by the models, thus a recent concentration of volcanism in the region is explained. Seismological observations in Tibet have defined diagnostic mantle properties in regions where delamination is suspected to have occurred recently. Seismic attenuation (Ni and Barazangi, 1984) and velocity (Brandon and Romanowicz, 1986) studies indicate that asthenospheric mantle reaches shallow depths under the Northern Tibet Plateau region, consistent with the delamination hypothesis.

As a second example, the models of Bird (1979, 1988) for the Basin and Range encompass a period of crustal thickening in the Tertiary, coincident with shallow underthrusting of an oceanic plate, followed by delamination. The uplift, crustal extension and volcanism experienced by this region in the late Tertiary are the diagnostic observable changes that constitute primary evidence that the models seek to explain. As for the Tibet example, seismic studies have identified shallow asthenospheric mantle under the Basin and Range (which has undergone extension) but not under the Colorado Plateau (which has not) – see Beghoul and Barazangi (1989).

Finally, in the Southern Puna Plateau in the Andes lithospheric thickening has been followed by rapid lithospheric thinning signaled by uplift, volcanism, and a change in stress directions. This Andean example is treated in more detail below.

In all three of the regions cited above the rapidity of the change from "pre-delamination" state inferred from various regional geological observations quoted in the cited literature to the "post-delamination" state has been rapid and substantial. Existing models treat both the rapid delamination event itself (see Bird and Baumgardner, 1983) as well as the conditions that result from the delamination and persist for several million years thereafter. From the preceding discussions of uncertainties in mantle and crustal density, it is apparent that the total amount of uplift, which is very sensitive to density, is not a very compelling constraint on the thickness of delaminated lithosphere. The rapidity of uplift is much more significant for it records how fast conditions are changing underneath an uplifted block.

A specific limitation is the non-treatment of melting in the models. While it is recognized that melting is explained (solidus temperatures are exceeded) the feedback of melting on the rheology and temperature structure (e.g., Hollister and Crawford, 1986; Rutter and Brodie, 1992) is not incorporated. The consequences of melt migration are also not identified. The probable reason for the success of models that omit magmatism is that formation of magmas in the system only serves to magnify the contrasts in both density and rheology, which control the rate and locale of the tectonics.

An independent consideration is the distinctive chemistry of the delamination-related magmas themselves, which is considered next. Of all the observable effects of delamination, the magmatic rocks are the most enduring.

Delamination magmatism: Andean examples

Investigators in the Andes have discovered two localities that exhibit criteria that indicate delamination. Magmatism in the two areas – late Tertiary volcanic rocks of the southern Puna Plateau in Argentina (Fig. 5), and Gondwana-age magmatic sequences in the Frontal Andean Cordillera in central and southern Chile – is important in defining the timing and extent of the delamination. Rapid tectonic transitions occur in both areas, identical to those identified by modelers (cited above) as accompanying delamination. The Gondwana example has been discussed in Mpodozis and Kay (1992); here we focus on the Puna example (Kay et al., 1990, and in prep.), with the aim of illustrating the magma types that accompany delamination. Complementary geophysical investigations are underway in the Puna with the aim of defining lithospheric and asthenospheric characteristics (see below). Preliminary studies of the magmas (crust and mantle-derived) show the geochemical characteristics expected of magmas created during delamination events.

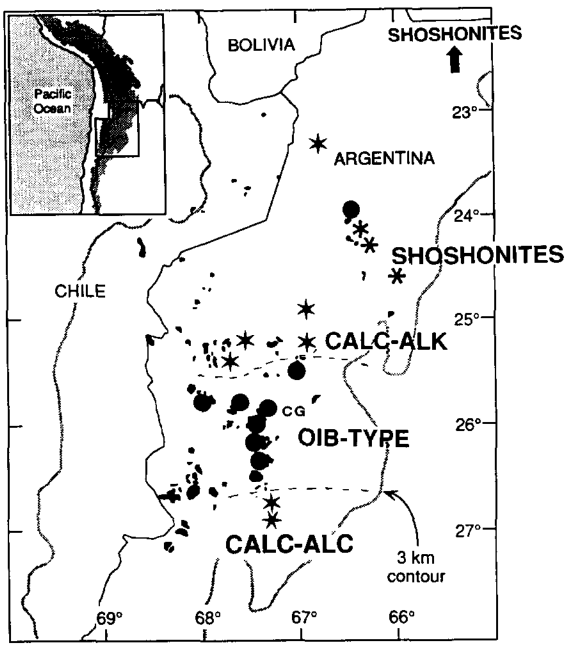

Fig. 5. Locations of mafic volcanic rocks sampled in the Puna (region above 3 km contour in map, shaded region in inset map), indicating geochemical types based on classification in Figure 7. Irregular dark regions are outcrops of mafic volcanic rocks based on mapping of TM satellite images by Eric Fielding at Cornell. No data is available for volcanic rocks from the SW Puna. *CG* = Cerro Galan (see text).

The Altiplano–Puna plateau, one of the earth's largest high standing (> 3 km) surfaces, is underlain by thickened crust – a product of crustal shortening in the Tertiary (see Isacks, 1988). The regional stress system of the southern part of the plateau, the Argentine Puna, underwent a fundamental change at about 2 to 3 m.y. ago (Allmendinger, 1986). Coincident with this change was the appearance of mafic volcanism across the region (Fig. 5). The topographic data of Isacks (1988), also discussed in Whitman et al. (1992), also shows that this region is somewhat higher than the northern Puna. We have proposed (Kay et al., 1990), drawing analogy to the models of England and Houseman (1989) for the northern Tibetan plateau, that the region of most extensive mafic volcanism outlines the area under which compressionally thickened mantle and lower crustal lithosphere delaminated. Silicic, crustally derived magmas (e.g., Cerro Galan Ignimbrite – Francis et al., 1989) erupted prior to (or coincident with) the proposed delamination (Fig. 5). Melting models constrained by the trace-element concentrations in these magmas indicate that dense mafic residues were created at the base of the crust prior to the proposed delamination, and we have suggested that these dense residues (garnet-bearing, with low modal plagioclase) have delaminated with the mantle lithosphere (Kay and Kay, 1991).

Ongoing seismic studies have outlined the geometry of the downgoing slab (Fig. 6), and have defined the lithospheric thicknesses beneath the Puna–Altiplano plateau (Isacks, 1988; Cahill, 1990; Whitman et al., 1990, 1992). The results are consistent with young mafic volcanism in the Puna behind the main volcanic arc being concentrated over a region of relatively thin lithosphere (Fig. 6). Note that the thin lithosphere is not due to backarc lithospheric extension. The volcanism is also concentrated over a prominent gap in the seismic zone. Studies of seismic wave paths suggest that the subducting slab is still present beneath this region (Whitman et al., 1990, 1992). A very reasonable explanation for this seismic gap is that the slab in this region is not generating earthquakes because it is too hot. Thus a plausible cause for the heating is that delaminated lithosphere has been replaced by hotter asthenosphere which in turn has heated the slab (Fig. 6). The distribution of volcanic rock types in the southern Puna (Fig. 5) is consistent with this interpretation (Knox et al., 1989; Kay et al., 1990, and in prep.).

The distribution, age and geochemistry of Central Andean (20–33°S) magmatic rocks, integrated with structural and other geologic data have led to a picture of changing slab geometries from the Early Miocene to Recent. The data suggest that during this time the angle of subduction has decreased in the modern "flat-slab" (28–33°S) region (Kay et al., 1987, 1991, 1992, in prep.), and has increased in the Puna north of about 24° to 25°S. (Isacks, 1988; Coira and Kay, in prep.). The intervening southern Puna region (25–27°S, Fig. 5) served as a pivotal zone where the slab neither steepened nor shallowed. Processes that shallow the subduction zone also appear to thin the lithosphere (Bird, 1988; Kay et

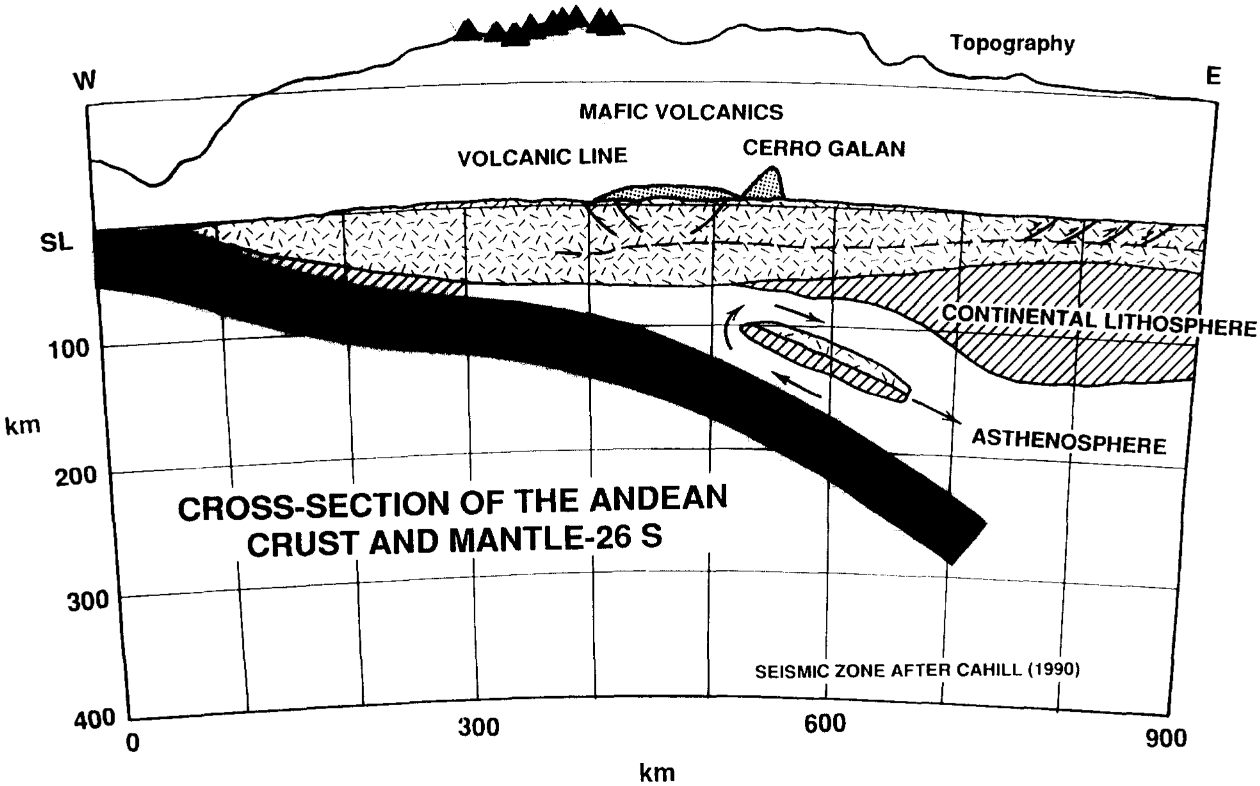

Fig. 6. Cross section cartoon at 26°S (see Fig. 5) showing delaminated block (from Kay et al., 1992, in prep.). Topography is shown at 10 × vertical exaggeration across the top. Dip of slab, lithospheric thickness and topography from Isacks (1988), Cahill (1990) and Whitman et al. (1992).

al., 1991), thus is it reasonable to suggest that the lithosphere under the southern Puna, which has not been over a very shallow seismic zone recently, should be thicker. Contradicting this expectation is the observation that the lithosphere is this region is thin. One possible explanation is

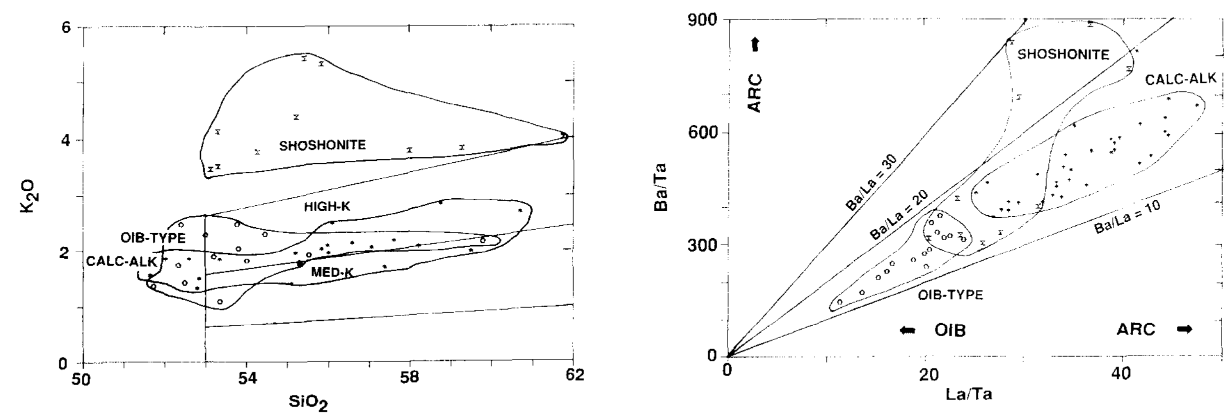

Fig. 7. Geochemical classification for centers in Figure 5 based on our analyses of young Puna mafic volcanic rocks (from Kay et al. 1992, in prep.). (a) (upper figure) K_2O va SiO_2 — Lavas of the shoshonitic series are differentiated from both calc-alkaline and Ocean island basalt (OIB)-type lavas based on this diagram. (b) (lower figure) OIB-type and calc-alkaline-like rocks are differentiated on the basis of their Ba/Ta and La/Ta ratios.

that the compression that caused Puna uplift produced a structurally thickened dense lower crust and upper mantle. Upon attaining a critical amount of shortening, this lithosphere became denser than the underlying asthenosphere, resulting in catastrophic delamination event centered in the southern Puna (Fig. 6). Our preliminary geochemical studies (Knox et al., 1989; Kay et al., 1990) and more extensive studies in progress have defined some important regional patterns in the chemical characteristics of the young Puna mafic rocks (Fig. 5). These patterns are fundamental for understanding crustal and mantle source regions for volcanic rocks in the Central Andes as well as for probing the delamination hypothesis.

The young Puna mafic rocks are generally basalts and andesites (51–58% SiO_2) that occur in small cones and isolated lava flows associated with young faults. Three general chemical subdivisions, shoshonite, Oceanic Island Basalt (OIB)-type) and calc-alkaline, have been defined based on trace-element characteristics (K and La contents; La/Ta; Ba/Ta; La/Yb ratios). Shoshonitic group lavas are recognized by their high K (Fig. 7a) and incompatible element contents and steep REE patterns (high La/Yb ratios – see Fig. 8). They occur in the east central Puna and are also found in the Altiplano (north of 23°S, see Fig. 5

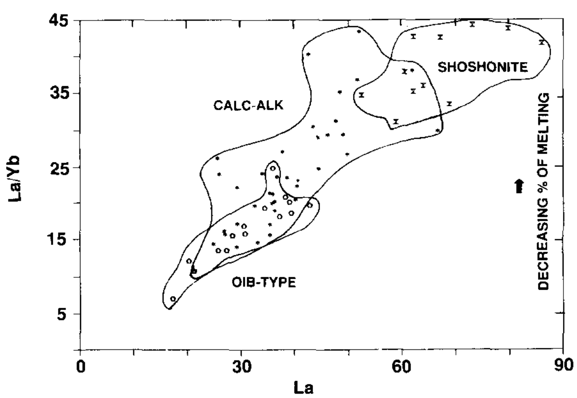

Fig. 8. La/Yb vs La (in ppm) for Puna basalts and andesites showing points and data fields for young Puna mafic rocks (from Kay et al. 1992, in preparation). OIB-type rocks have the lowest La/Yb ratios and La contents while the shoshonites have the highest. The data can be interpreted to suggest that melting percentages are higher in the source for the OIB rocks, which have the greatest volume in the field.

and summary by Soler, in prep., and LeFevre, 1979). OIB-type lavas (called Galan group by Knox et al., 1989) have OIB-like Ba/Ta and La/Ta ratios (Fig. 7b) and relatively low La/Yb ratios (Fig. 8). These lavas occur above the central part of the gap in the modern seismic zone in the southern Puna (Fig. 5). High-K calc-alkaline group lavas have more arc-like Ba/Ta and La/Ta ratios (Fig. 7b) and intermediate La/Yb ratios (Fig. 8). They occur at the edge of and above seismically active parts of the descending slab and in general, flank both sides of the OIB-type group (Fig. 5).

The chemical characteristics of these Puna lavas can be explained in a general way by considering them in the context of the regional geology and geophysics. Given the simplified assumption of a relatively homogenous peridotite source, La/Yb ratios and La contents can be used as indicators of melting percentages (e.g., Kay and Gast, 1973; Nakamura et al., 1989). In this way, the OIB-type magmas, which to a first order are the most voluminous, are also the highest percent melts (lowest La contents and La/Yb ratios). This correlates with the fact that these lavas overlie the seismic gap where the hottest mantle (most seismically attenuating, Whitman et al., 1990) occurs. Perhaps the lack of a slab signature in the OIB-like magmas can be explained by loss of the top of the subducting slab due to association with the hot overlying mantle. Degradation of the seismic zone could be associated with the influx of hot asthenosphere following an episode of lowermost crustal and mantle delamination. The loss of either large blocks or smaller ones through convective instability (e.g., Yuen and Fleitout, 1985) would then account for stress reorientation, uplift and the extrusion of OIB-type mafic magmas (Kay et al., 1990) over the last 2 to 3 m.y. in the southern Puna. Following this model, the less voluminous calc-alkaline lavas flanking the OIB-type lavas would represent intermediate percentages of melt along the cooler margins of this region where the lithosphere is thickening to the north (Whitman et al., 1992) and the slab is shallowing to the south. Finally, the least voluminous rocks, the shoshonites to the north, would represent the smallest percentage of melting.

They occur in the northeastern Puna and Altiplano where the lithosphere is thicker (Whitman et al., 1992). Thus over the whole region, the geochemistry of the mafic lavas appears to be critical in understanding the sequence of events and in defining where delamination may have occurred.

It is important to consider how events and conditions that are so easily assembled and correlated in the present Puna will look 100 m.y. in the future. Then, much of the evidence for delamination will either exist no longer (the case for seismicity and for thickness of the crust and lithosphere), or will be difficult to extract given the overprint of subsequent events, especially in the case of structure and thermal conditions. Sedimentary and igneous rocks record the tectonic conditions, and between these the igneous rocks have the highest preservation probability, as they can be recovered from various crustal levels. The distinctive sequence of high La/Yb ratio ("eclogitic" - see Kay and Kay, 1991) intermediate to silicic crustally-derived magmas followed by mafic post or orogenic or extension-related magmas has parallels in the later stages of many orogenic belts. Especially distinctive is the first appearance of OIB-type basalts, which indicate the influx of new, non-subduction-related asthenospheric mantle into the "mantle wedge" - an event that we correlate with delamination of mantle lithosphere (that has an arc-related geochemical signature). In the Puna it must be reemphasized that delamination is postulated because the total crustal extension is very limited, and is therefore insufficient to account for the observed lithospheric thinning.

Delamination and evolution of the crust–mantle system

For a long time it has been recognized that mass is being exchanged between continental crust and mantle at subduction zones (e.g., Armstrong, 1968; Von Huene and Scholl, 1991). Only more recently has delamination entered the picture as a process for creating particular types of mantle reservoirs (e.g., McKenzie and O'Nions, 1983), and then the process was envisioned as delamination of mantle only. The possibility that crystal cumulates or crystalline residues from melting sink from the uppermost mantle and lowermost crust into the asthenosphere mantle has been championed by Kay and Kay (1985, 1988, 1990, 1991) and Arndt and Goldstein (1989). The idea of relating lower crustal recycling to particular short-lived delamination events, advanced here (see also Kay et al., 1990; Kay and Kay, 1991), connects the recycling mechanism directly to currently popular theories for the development of compressional orogens - and to observable (and preservable) events in the upper crust.

At first glance, crust to mantle recycling by delamination may not seem comparable in magnitude to sediment subduction. Von Huene and Scholl (1991) have estimated that at present 0.7 km^3/yr of crustal sediment is subducted into the mantle at the 44,000 km of convergent plate margin. In contrast, the sites of delamination that may involve lower crust are few at any one time - the Andes and Tibet at present, perhaps the Alps in the future. But the masses involved in each of these delamination events is plausibly large. Globally, every million years, delamination of 20 km of lower crust under an area comparative in size to the 250 km by 100 km area outlined by the "OIB-type" mafic magmas in Figure 5 (an area outlining delamination, as argued above) would equal the Von Huene and Scholl (1991) sediment subduction estimate. Furthermore, as noted by Kay and Kay (1985, 1988) the absence in accreted terranes of mafic lower crustal sections found in oceanic arcs and plateaus, and the tendency of balanced cross sections to overestimate lower crustal as well as mantle lithospheric mass (e.g., Nelson, 1991; Kay et al., 1992) are both consistent with removal of lower crustal mass from the sections. Finally, delamination events may have been commoner at some times in the past (for instance, at times of supercontinents accretion: Kay, 1993) than they are at present.

The existence of mechanisms that recycle crust to mantle serves to emphasize that crustal growth is the net result of two-way mass transfer between mantle and crust. At present crustal additions from the mantle are largely basalt, and crustal subtractions to the mantle are the weighted sum of upper crust (subducted components) plus lower crust (delaminated, or perhaps subcrustally eroded components). In the Archean, when the net rate of crustal accumulation was greater, growth mechanisms may have been different (see Kröner, 1985). Recently, the extraction of basalt, andesite and dacite magmas directly out of ultramafic or mafic sources in the mantle has probably been the most popular crustal growth process: a recent book on the continental crust (Taylor and McLennan, 1985) does not mention delamination. Extraction of andesite from the mantle generally requires a non-peridotitic mafic source (see Kay and Kay, 1991 review). Martin (1986) and Defant and Drummond (1990) identify the mafic source as subducted oceanic crust. Examples of this process occur today, but only infrequently, in subduction zones that are extraordinarily hot, due to the youth of the subducted plate. The claim is that in the Archean, the hotter mantle temperatures and smaller plates made "hot" subduction regimes much more common, and extraction of andesite from the mantle correspondingly much more common (e.g., Tarney and Weaver, 1987). It is important to note that the andesitic to dacitic magmas associated with delamination and with hot-subduction share some geochemical features (e.g., high La/Yb ratios and Sr concentrations). This may be expected, as both originate by melting of mafic rocks at high pressure, leaving a garnet-bearing plagioclase-free "eclogitic" residue. However, we note that the tectonic contexts of these eclogite-related melts are distinct. If the country rocks are thickened crust, related to compression, now probably preserved as terranes of high metamorphic grade, then a delamination origin is indicated. Many of the Archean localities (see Taylor and McLennan, Chapter 7) fit this description. If the country rocks are oceanic lithologies, including those of oceanic island arcs and ophiolites, then an origin by shallow melting of a hot subducted slab is indicated. Few Archean localities fit this description.

The relevance of delamination for mantle evolution is direct, for crustal components are returned to the mantle. The trace element contents of the delaminated mantle and crustal masses are very different from those in other recycled crustal materials (e.g., sediment). Using analyses of mantle-derived basalts it has been easy to "rule out" the presence of recycled sediment taken alone into the mantle, but it is not so easy to "rule out" the mixture of several compositionally disparate crustal components, including sediment, each modified by complex processes occurring in subduction or collision zones. The characterization of delaminated mantle and lower crust, possibly by examination of the geochemical features of magmas derived from these transient lithospheric reservoirs, is undoubtedly important for understanding the isotopic and compositional evolution of the mantle. Equally important is the view that mantle and crust are complementary geochemical reservoirs, with andesitic crustal composition as the result of both addition of mass from the mantle, and removal of crustal mass to the mantle.

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