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Continental Delamination and the Colorado Plateau

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Continental lithosphere is in unstable mechanical equilibrium because its mantle layer is denser than the asthenosphere. If any process such as cracking, slumping, or plume erosion initially provided an elongated conduit connecting the underlying asthenosphere with the base of the continental crust, the dense lithospheric boundary layer could peel away from the crust and sink. An analytic model for sinking velocities at the critical initial time shows that instability occurs if the effective viscosities of the lower continental crust and the rising asthenosphere are no more than 10^{19} P. Analogies to subduction suggest that the mature instability would grow laterally at plate tectonic velocities; however, it would be almost aseismic. Loss of the cold mantle boundary layer would cause uplift, increased heat flow, reduced seismic velocities, and perhaps emplacement of basalt flows, mantle diatremes, and granodiorite sills. A one-dimensional thermal model of the formation of a new boundary layer predicts a half life of about 3×10^7 years for this thermal anomaly and uplift. As an example, the geologic and geophysical data from the Colorado Plateau are shown to be consistent with the hypothesis that it was uplifted by a delamination event 30 m.y. ago and perhaps a second event about 5 m.y. ago.

INTRODUCTION

The model of plate tectonics successfully explains some grand features of continental geology, such as the rifting, separation, and eventual recollision of shields in the Wilson cycle. But there are at least two continental tectonic problems which plate theory seems unable to solve: sudden uplift and igneous activity far from plate boundaries, as in the Colorado Plateau of the western United States.

Past attempts to tie the tectonics of the Rockies, the Andes, and Tibet into plate tectonics have invoked a shallow dipping slab of subducting oceanic lithosphere, which might plausibly extend inland 1000 km from a known plate boundary. This slab is regarded as either a source of magmas [e.g., *Lipman et al.,* 1971] or as the cause of induced convection in the asthenosphere above [*Toksöz and Bird,* 1977*b*]. These models are difficult to apply to the Colorado Plateau, since it was still a lowland area in the Eocene [*Eardley,* 1951] when the hypothesized shallow slab was already retreating back toward California [*Snyder et al.,* 1976]. The removal of such a dense slab could not in itself produce a net uplift but only cancel any depression that had previously been caused by the insertion of that same slab. Given these problems, it seems prudent to consider alternative models independent of classical plate tectonics.

A solution to these problems can be found if we relinquish the idea of continents as monolithic plates and consider the possibility of decoupling between crust and mantle. Some preliminary results of laboratory deformation [*Tullis and Yund,* 1977] suggest that dislocation creep within sialic crust might be appreciable at normal Moho temperatures. *Bird* [1978*a*] has shown that a horizontal 'décollement' of crust from mantle is occurring in the Zagros today and has argued [*Bird,* 1978*b*] that the Himalayan orogeny involved a vertical separation or 'delamination' of mantle lithosphere from crust. If this process can occur in a continental collision, might it not also take place within plate interiors?

This paper argues for the inherent instability of continental lithosphere, using the assumption that mantle lithosphere and asthenosphere are chemical equivalents, of different temperature and density. A simple analytic model is used to show that an initial flaw in the subcrustal lithosphere may grow at an accelerating rate by peeling away and sinking of this cold mantle below the crust. This causes a wholesale replacement of cold mantle by hot in a geologically short time, with observable geophysical and geologic consequences. The middle section of the paper gives predictions of the amplitude and time decay of these effects to allow for testing of this delamination model. Finally, the Colorado Plateau is discussed in detail to show that the predictions are consistent with published data.

SPONTANEOUS DELAMINATION OF CONTINENTAL LITHOSPHERE

Any layered assemblage of liquids is in mechanical equilibrium as along as its layer boundaries are perfectly horizontal. In this situation there are no horizontal gradients of stress to cause acceleration or shear stresses to cause deformation; hence there is no motion. The interior of a continental plate (away from orogenic areas) approximates such horizontal layering; therefore the long inactivity of continental shields does not imply that they are necessarily composed of strong rocks.

The mechanical equilibrium of the assemblage will be unstable whenever a denser layer overlies a lighter one. In this case any disturbance of the layering grows, since gravitational potential energy to drive the deformation is released by the reordering of densities. The mantle or subcrustal portion of the continental lithosphere is probably unstable by this definition because it is more dense than the seismic low-velocity zone, or asthenosphere, below it. This density difference arises primarily from thermal expansion/contraction.

In this paper it is assumed that both continental (mantle) lithosphere and asthenosphere have the same chemical composition and thermal expansion coefficient as the total oceanic lithosphere.

The oceanic lithosphere, which, according to plate tectonic theory, is formed by cooling of asthenosphere at ridges, is separated by differentiation into a basaltic crust and a depleted residue. But since both crust and residue are included in the plate, the effect of differentiation on the bulk thermal contraction and density of the plate is probably small.

Bathymetric data from *Sclater et al.* [1971] show the Pacific Ocean floor sinking from 2750-m depth toward 5900 m in 100 m.y. of cooling. If the density of asthenosphere under ridges is

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about 3.13 g/cm^3, then 6.63×10^5 g/cm^2 of mass have been lost in topography. If isostasy holds, this same mass has been added by density increases in the lithosphere. *Leeds* [1975] finds the thickness of Pacific lithosphere at 100-m.y. age to be 80 km, so the average density increase has been 0.083 g/cm^3. The average temperature drop in this thickness range should be half of the solidus or formation temperature, since a linear geotherm is well established by this time. If this average temperature drop is 650°C, we derive a (volumetric) thermal expansion coefficient of about 3.9×10^{-5}/°C. This is consistent with measured values of 3.1–4.4 $\times 10^{-5}$/°C for olivine at 800°C [*Skinner,* 1966].

Average stable continental plates have a base (defined seismologically) at about 120-km depth [*Toksöz et al.,* 1967]. The conventional interpretation of the velocity decrease here is that a small percentage of melt appears. Accordingly, the plate bottom temperature in a peridotite mantle can be estimated, after *Kay et al.* [1970], to be about 1350°C. On the other hand, the average Moho temperature at 35-km depth is probably only 550°C in platform areas with heat flows of 54 mW/m^2 [*Bird,* 1978*b*]. This means that continental mantle just below the Moho has a pressure-corrected (potential) density 0.10 g/cm^3 greater than the asthensophere below, with this anomaly declining linearly to zero at the base of the plate.

This difference in densities means that if some crack allows asthenosphere to rise up and contact the Moho, its static pressure there will be 400 bars greater than the previous lithostatic Moho pressure. Therefore given sufficiently low viscosities, the asthenosphere will expand along the Moho surface and change places with the mantle lithosphere.

This hypothetical disruption of the lithosphere would have to be longer in one horizontal dimension than the lithosphere is thick, or the lithosphere would be prevented from sinking rapidly by geometric constraints. Such long ruptures occur occasionally in geologic history when continents break up around a new spreading center. *Bird* [1978*a, b*] has argued that the subcrustal lithosphere is also broken in continental collisions by the detachment of the oceanic slabs, but other processes unconnected with differential plate movements might also cause a linear disruption. A hot rising 'plume' under a relatively moving plate interior, like the one that apparently thinned the oceanic lithosphere to 40 km under Hawaii [*Detrick and Crough,* 1978], might erode a continental plate up to the Moho. Or a linear thermal anomaly imposed on the base of the plate by convective rolls in the asthenosphere [*Richter and Parsons,* 1975] might grow by Rayleigh-Bernard instability in the lithosphere until it was completely disrupted. In this paper no attempt is made to predict all of the possible sources of linear disruptions. Instead we calculate whether an asthenospheric conduit up to the Moho can cause an instability and what observable effects can be predicted.

Stability of Disrupted Lithosphere

Assume that a broad conduit exists in the subcrustal lithosphere of a continent and is filled with mobile asthenosphere. Assume the conduit is infinitely long in the y direction, as in Figure 1. Because of symmetry we need consider only one side ($x > 0$). Also assume that the lower continental crust contains a layer of thickness S which behaves viscously with viscosity μ_c. Now, since the conduit pressure is greater than the lithostatic pressure along the Moho at $x \to \infty$ by an amount $\Delta P(0)$, the crustal material in the viscous layer will be driven to the

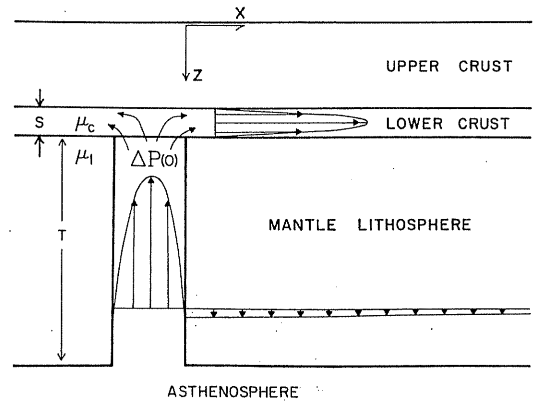

Fig. 1. Schematic diagram of a conduit through the mantle lithosphere of a continent, showing the convective flow that may lead to a delamination instability. The structure is assumed to be two-dimensional. Symbols explained in text.

right. The total horizontal flux of crustal material $F(x)$ is given by

$$F(x) = \int_{M-s}^{M} V_x(x, z)\, dz = \frac{-S^3}{12\mu_c} \frac{\partial}{\partial x} \Delta P(x) \tag{1}$$

where $z = M$ is the plane of the Moho. Because rocks are approximately incompressible, the mantle lithosphere and the upper crust must separate at a vertical velocity

$$V_z(x) = -\frac{\partial}{\partial x} F(x) = \frac{S^3}{12\mu_c} \frac{\partial^2}{\partial x^2} \Delta P(x) \tag{2}$$

The anomalous pressure in the lower crust, ΔP, is also controlled by the equation for the bending of the viscous slab of mantle lithosphere:

$$\Delta P(x) = \left(\frac{T^3\mu_l}{3}\right) \frac{\partial^4}{\partial x^4} (V_z(x)) \tag{3}$$

where T is the slab thickness and μ_l is its viscosity. Combining (2) and (3) we obtain the governing differential equation

$$\frac{\partial^6}{\partial x^6} (V_z(x)) = \theta^6 V_z(x) \tag{4}$$

where

$$\theta \equiv \left[\frac{36\mu_c}{S^3 T^3 \mu_l}\right]^{1/6} \tag{5}$$

is a constant. The boundary conditions are that $V_z \to 0$ as $x \to \infty$ and that the moment (proportional to V_z'') and the shear traction (proportional to V_z''') both vanish at the edge of the conduit ($x = 0$). They imply the unique solution

$$V_z(x) = \frac{3 \cdot \Delta P(0)}{T^3 \mu_l \theta^4} \left\{ \exp(-\theta x) + \exp\left(\frac{-\theta x}{2}\right) \cdot [\cos((3/4)^{1/2}\theta x) + (1/3)^{1/2} \sin((3/4)^{1/2}\theta x)] \right\} \tag{6}$$

for the velocity of the lithosphere. This velocity is greatest at the free end:

$$V_z(0, M) = \frac{6 \cdot \Delta P(0)}{T^3 \mu_l \theta^4} = (0.5503)\Delta P(0) S^2 T^{-1} \mu_l^{-1/3} \mu_c^{-2/3} \tag{7}$$

As discussed above, the pressure anomaly $\Delta P(0)$ will be approximately 400 bars, and the thickness of the mantle lithosphere T about 80 km. The mantle lithosphere viscosity is probably less than the 10^{24} P determined by *Walcott* [1970] for the entire lithosphere yet more than the average mantle viscosity of 10^{22} P obtained from glacial rebound studies. Here we will use an estimate of 2×10^{23} P, noting that a factor of 10 error would only affect V_z by a factor of 2. Much harder to determine are the lower crustal thickness and viscosity (S and μ_c).

In a mechanical simulation of the Zagros range, *Bird* [1978*a*] determined the ratio μ_c/S to be about 6×10^{14} dyn s/cm^3. If we estimate the thickness of the shearing layer to be 10 km, this implies a viscosity of 6×10^{20} P. Although this sounds low, it may be a maximum value for the lowermost continental crust, since the Zagrós area has a base line heat flow of only 47 mW/m^2 [*Toksöz and Bird,* 1977*a*]. In places where heat flow is greater or the crust is thicker, the Moho temperature might easily be 200°C higher. Then if the activation energy of the viscous process were perhaps 40 kcal/mol, the viscosity might drop to 7×10^{17} P! Putting these alternate estimates into (7), we find that the free end of the mantle lithosphere subsides at either 0.20 or 18 cm/yr.

As sinking proceeds, three changes occur that will modify the rate. First, asthenosphere material intrudes into the lower crustal layer. If its viscosity is much greater than that of the lower crust, the process will obviously stop. Some estimates of asthenosphere viscosity have been as low as 2×10^{19} P [*Sliboutry,* 1971] in situ, which is encouraging. Also the drop in pressure that the material experiences while rising in the conduit is likely to induce further partial melting, as it does beneath midocean spreading ridges. If a liquid basalt fraction separates at the top of the conduit [*Sleep,* 1974], then the viscosity could be very low.

A second effect is that the intruding asthenosphere will form chilled, solid boundary layers on top and bottom of the sill layer. This reduces S and competes with the third effect, which is the increase of S due to subsidence of the lithosphere. If the freezing predominates, the system is stable; if separation predominates, it is very unstable, since (7) suggests that (dS/dt) will vary roughly as S^2. If we equate the time required for the original layer thickness to be entirely blocked (without allowing for widening) to the time required for it to double in width (without allowing for freezing), we can estimate the critical viscosity. This viscosity is 1.4×10^{19} P if the frozen boundary layers thicken according to the same law as governs the growth of oceanic lithosphere at the surface. Since the growth of boundary layers follows a smaller time exponent ($t^{1/2}$) than the widening of the layer (t^1), an intrusion which passes this early critical point will accelerate until the resistance of the asthenosphere to the sinking slab becomes dominant, and this is not likely to occur until the slab droops low enough to reach the higher-viscosity 'mesosphere' inferred by *Sleep* [1979] and others.

Lateral Spread of Delamination

The farther the intrusion of asthenosphere spreads into the crust, the wider the area in which the intrusion is too thick to suffer any significant viscous pressure drop, and the wider the zone of unsupported mantle lithosphere. This drooping slab will exert downward shear force and moment on the adjacent mantle lithosphere, tending to tear it away regardless of the local crustal viscosity. Simultaneously, this growing 'paddle' will encounter growing resistance to its sinking through the mantle. The velocity of lateral spread of delamination should therefore be controlled by a balance between the slab's excess density and mesosphere viscosity, just as a similar balance in subduction zones may control the velocity of the surface plates [*Forsyth and Uyeda,* 1975].

Figure 2 is a cartoon intended to show the similarity of mantle flows between the cases of subduction (with back arc spreading) and delamination (with a favorable component of upper mantle shear). Simple thermal expansion calculations show that subcontinental lithosphere has an average excess density at least 75% of the excess density of oceanic lithosphere at 10^8 years of age. The work done in 'tearing' the lower continental crust is probably less than the interplate dissipation in subduction zones, and the mantle into which both types of slab sink is probably similar [*Okal and Anderson,* 1975]. So once the instability is well developed, it can be expected to propagate at plate tectonic rates, of the order of 5 cm/yr. It can be expected to stop whenever it encounters a continental margin or another flaw in the lithosphere like the one which initiated it.

One possible objection to this hypothesis is that delamination must be very rare, for only one midcontinent Benioff zone of seismicity is presently known (in the Hindu Kush). However, detailed studies of subducting slabs in Alaska [*Engdahl,* 1977] and Japan have shown that all internal earth-

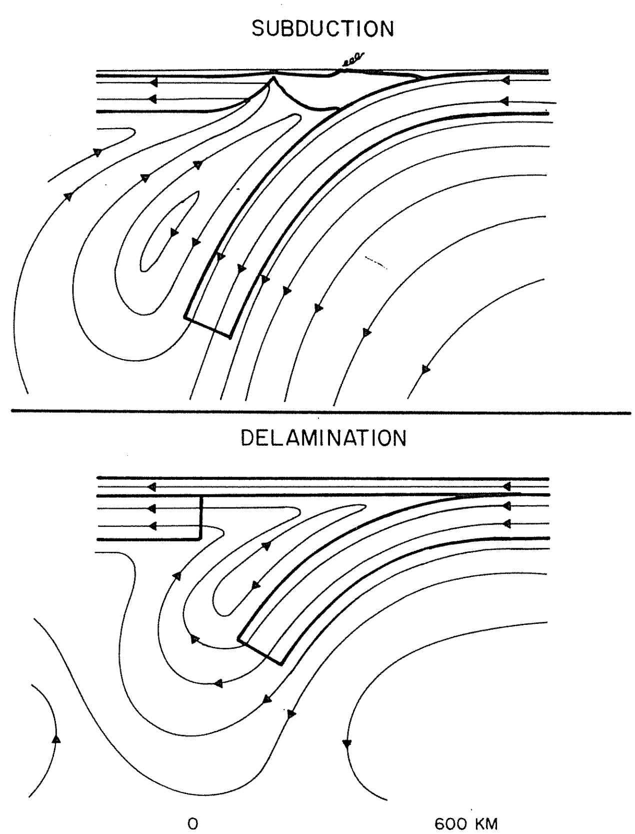

Fig. 2. Cartoon showing the similarity of upper mantle flow between subduction with back arc spreading (top) and delamination of a continent which is traveling to the left in relation to the lower mantle (below). If mesosphere viscosity supports most of the slab's weight in each case, the magnitude of the velocities will be similar.

quakes occur in the top 20–40 km of the oceanic slabs. This is the part in which temperatures are very low and precisely the part of the lithosphere (the crust) which is not deformed by delamination. Hence it could be an almost aseismic process. It should be easier to detect its aftereffects, which occur when a layer of continental crust is left sitting directly on hot asthenosphere.

CONSEQUENCES OF DELAMINATION

In this section a one-dimensional conductive model is used to predict the temperature changes during the period when a new mantle lithosphere is gradually established by cooling. This history can be used to predict quantities which depend on temperature (magma formation), on the temperature gradient (heat flow), or on the integral of temperature (elevation and gravity).

Conductive Thermal Model

It is assumed that re-formation of the lithosphere takes place primarily through conductive cooling rather than convection. (Otherwise, there are too many unknown parameters to allow for any meaningful predictions.) Thus we start from the heat equation in a solid:

$$\rho C \frac{\partial T}{\partial t} = \nabla \cdot (K \nabla T) + H \tag{8}$$

where ρ is density, C specific heat, T temperature, t time, K thermal conductivity, and H radioactive heat production. Next we subtract out the normal geothermal gradient maintained by mantle flux and heat production H and consider only the excess, anomalous temperature T' due to delamination:

$$\rho C \frac{\partial T'}{\partial t} = \nabla \cdot (K \nabla T') \tag{9}$$

(This step requires the assumption of temperature-invariant K.)

Next it is assumed that the region delaminated is several times wider than it is thick, so that in the center, horizontal gradients of temperature are negligible:

$$\rho C \frac{\partial T'}{\partial t} = \frac{\partial}{\partial z}\left(K \frac{\partial T'}{\partial z}\right) \tag{10}$$

Two natural boundaries in depth z are the earth surface and the maximum depth perturbed by delamination, at the former bottom of the lithosphere. The surface boundary is held at 0°C. The bottom boundary at 120 km is held at zero anomalous temperature ($T' = 0$), because it is likely that the asthenosphere at greater depths is in a state of local convection combined with shearing horizontal flow. The assumed initial condition is that the asthenosphere (already near an adiabatic gradient) rises to the Moho on a continuation of that adiabat to replace the fallen away lithosphere (Figure 3).

This problem has been solved numerically by expressing T' as a decaying sine series:

$$T'(z, t) = \sum_{i=1}^{N} a_i \sin\left(\frac{z\pi i}{120 \text{ km}}\right) \exp\left[-\kappa \left(\frac{\pi i}{120 \text{ km}}\right)^2 t\right] \tag{11}$$

where κ is thermal diffusivity. This satisfies the initial and boundary conditions and also (10) as long as K and κ are constants. We use $N = 120$, $K = 2.5 \times 10^5$ erg/cm s °C, and $\kappa = 0.012$ cm²/s. The effect of increasing either is obvious; less obvious is the effect of varying the crustal thickness, because it changes the initial condition as well as the length scale. Therefore two representative crusts of 30 and 40 km were separately modeled.

Results are shown in Figure 3 for various times before and after delamination. After 150 m.y. no significant thermal perturbation remains, and the cycle is completed.

Geophysical Consequences

Heat flow is easily calculated by adding the steady state flux to the derivative of the anomalous temperature:

$$K \left.\frac{\partial T}{\partial z}\right|_{z=0} = F_0 + K \sum_{i=1}^{N} a_i \left(\frac{\pi i}{120 \text{ km}}\right) \exp\left[-\kappa \left(\frac{\pi i}{120 \text{ km}}\right)^2 t\right] \tag{12}$$

Figure 4 shows the results for two different crustal thicknesses. The heat takes a few million years to diffuse through the crust, so the peak is not reached until 9–14 m.y. after delamination. These peaks are rather small (15–25 mW/m² extra flux) and would be difficult to separate from the normal background flux F_0.

The Bouguer gravity anomaly is a more useful indicator, since it is commonly measured with an accuracy of 1%. Furthermore, we can look to stable, nontectonic areas around the world to estimate the initial value of the anomaly prior to delamination (Δg_0). (According to *Woollard* [1959] the stable plate anomaly is +30 mGal over a 30-km crust and −100 mGal over 40 km of crust.) From this we must subtract the loss in gravity due to replacement of cold mantle by thermally expanded asthenosphere and add the excess attraction of the sinking and warming slab:

$$\Delta g = \Delta g_0 + \Delta g_s - 2\pi\rho\alpha G \int_0^{120 \text{ km}} T' \, dz = \Delta g_0 + \Delta g_s$$

$$- (5.25 \times 10^{-8}) \sum_{i=1 (i \text{ odd})}^{N} a_i \left(\frac{240 \text{ km}}{\pi i}\right) \exp\left[-\kappa \left(\frac{\pi i}{120 \text{ km}}\right) t\right] \tag{13}$$

where Δg is the total anomaly in milligals, Δg_s is the excess attraction of the sinking slab, α is the volumetric thermal expansion coefficient, and G is the universal gravity constant. A further effect which might increase the Bouguer anomaly is crustal uplift, which removes a thin layer of crust from the section below sea level and adds a corresponding thickness of asthenosphere. In the limit of full isostatic uplift the thermally induced part of the gravity anomaly is reduced by the factor

$$\frac{(\Delta g - \Delta g_0 - \Delta g_s)_{\text{isostatic}}}{(\Delta g - \Delta g_0 - \Delta g_s)_{\text{rigid}}} = \frac{\rho_c}{\rho_A} \tag{14}$$

where ρ_c is the density of the top of the crust and ρ_A is the density of asthenosphere. This ratio is probably about 0.85.

Figure 4 shows that transient anomalies of −110 mGal to −180 mGal can be produced in the upper 120 km and that they decay with a half life of about 30 m.y. The positive effect of the sinking slab is initially large enough to offset these, but after a while it is diminished in three ways: the slab expands from warming, it sinks farther away from the observer, and relative horizontal velocities in the upper mantle may carry it out from under the observer, so that its attraction is more sideways than down.

Many studies have shown a negative correlation between mantle temperature and seismic velocities and a positive correlation of temperature and attenuation [e.g., *Molnar and Oliver*, 1969]. Surface waves of 20- to 80-s period and vertically

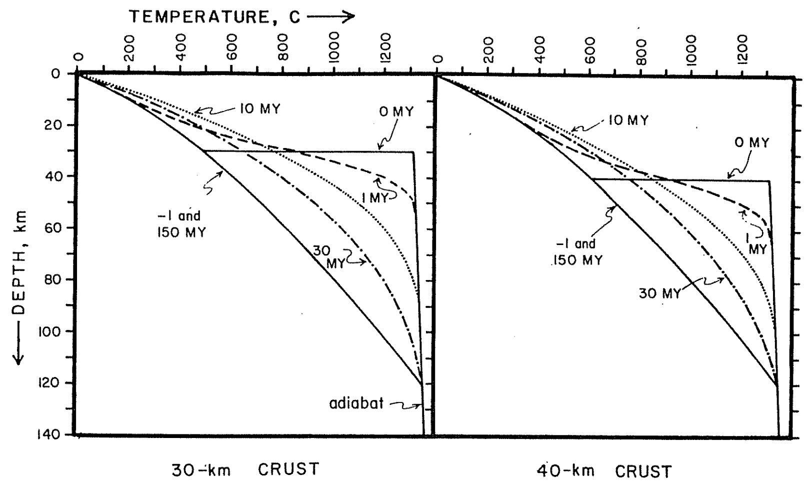

Fig. 3. Temperature profiles in a delaminated area as a function of time and crustal thickness. Delamination is represented by replacing all temperatures below the crust with an asthenospheric adiabat at time $t = 0$ m.y. Through the next 150 m.y. the thermal anomaly decays by one-dimensional conduction, with temperatures at 0 km and 120 km held fixed. The initial and final conditions are consistent with a mantle heat flux of 30 mW/m^2 and a surface flux of 54 mW/m^2.

incident S waves should give an indication of high mantle temperature by their delay and loss of high-frequency energy. This prediction must remain qualitative because of our poor understanding of the physics and geochemistry of the two-phase asthenosphere and because the rate at which these effects would decay is unknown.

Finally, it seems likely that the electrical conductivity of the upper mantle in a delaminated region should be enhanced, allowing detection by geomagnetic depth sounding or magnetotellurics. Laboratory and theoretical studies [e.g., *Chan et al.,* 1973] indicate that the presence of partial melt should raise conductivity and the intrinsic conductivity of the solid mantle minerals should increase with temperature also. Thus a delaminated region should be characterized by reduced ratios of transient vertical to horizontal magnetic fields and by low apparent resistivity at periods of thousands of seconds. Again, a quantitative prediction of this effect would be premature.

Probable Geological Consequences

The preceding effects of delamination leave no permanent record and can only be observed where it has happened in the Late Tertiary. This section considers the geological consequences that might mark an ancient event: gas-rich eruptions, basalt extrusion, silicic intrusion, and uplift. Each of these involves an all-or-nothing, nonlinear process like fault slip or a phase change. Therefore lack of one response does not rule out an ancient delamination.

Delamination is a possible solution to the problem of what initiates the eruption of gas-rich volcanics from the mantle in violent maar eruptions. The kimberlitic diatremes of South Africa (for example) give textural evidence of fluidization and transport by rapidly expanding gases rich in carbon dioxide. The ultramafic nodules included in them, the occasional presence of diamond, and the few discoveries of coesite all suggest an origin depth of 100–300 km [*Cox,* 1978]. Yet there has not to date been a satisfactory explanation of what process causes an uprise so fast that the volatile pressure becomes equal to total pressure along the way [*McGetchin et al.,* 1973]. Various convective models have been proposed to bring mantle material upward, but it is not clear why this convection should produce diatremes so rarely or why they are never erupted (to our knowledge) at midocean ridges, where convection is definitely occurring. Only the 'reversing flow' model of *Anderson and Perkins* [1975] seems consistent with the scarcity and uneven distribution of diatremes.

In the delamination model the rapid sinking of lithosphere implies an even faster ascent of asthenosphere to replace it. For example, some of the asthenosphere in the conduit of Figure 1 would rise from 120-km to 40-km depth at a minimum speed of 24 cm/yr if the conduit were 25 km wide. None of the previous models of subcontinental convection bring material so shallow so rapidly. In the process the rocks suffer a pressure drop of about 25 kbar, or three quarters of the original pressure. This means that carbon dioxide originally present in the subcontinental asthenosphere as carbonates or in a melt fraction might in the process of delamination separate out as a new liquid phase. The fluid could then segregate into pockets of considerable size because of its negligible viscosity, and buoyancy resulting from its low density would force it up by some dikelike fracturing process. A few kilometers below the surface these hot fluids would flash into gas and blast out diatremes on their way to the surface.

This interesting hypothesis cannot be called a prediction, however, because we know so little about the concentration and distribution of such volatiles. It may be that areas with such concentrations are rare in the subcontinental asthenosphere and are only found where there has been subduction, plume activity, or some other special event.

A more certain geological consequence is the eruption of basalts. The same 27-kbar pressure drop that may cause diatremes will almost certainly cause increased melting of the rising asthenosphere. If continental and oceanic mantle are equivalent as we have assumed, then the process should give rise to basalt as it does at spreading ridges. Since the uprise is so rapid, these basalts should have very high initial temperatures and should easily penetrate to the surface as low-viscosity flows.

A more difficult question is whether any contact melting of the lower crust is to be expected. The thermal calculation above shows that the Moho conditions at the time of delamination are 900°C at 8.3 kbar (30-km crust) to 965°C at 11.1 kbar (40-km crust). Afterward the temperature slowly declines by mantle cooling (Figure 3). Whether or not these conditions will produce any melting depends on the composition of the lower crust and the availability of water. Here I assume rather arbitrarily that the lower crust is dioritic and dry, so that any water to form melts must first be released by breakdown of a hydrous mineral. *Brown and Fyfe* [1970] present data for this case that show that a granodiorite partial melt would be formed in the first case only if the crust contained biotite or muscovite. In the case of a 40-km crust, phlogopite or hornblende might also break down and initiate melting. In any case the amount of melt would not be large, and the magma would be relatively cool and viscous.

Another geologic phenomenon that may be tied to delamination is regional uplift, which can be later inferred from unconformities, stratigraphic offsets, and sedimentation patterns. The excess pressure of about 400 bars that the newly risen asthenosphere exerts at the Moho will require some topographic compensation if the region delaminated is sufficiently large. To find the critical delaminated area necessary to create block uplift, consider a circular delaminated region of diameter D bounded by a vertical cylindrical fault. The requirement for motion is that

$$\Delta P(0)D > 4\tau L \qquad (15)$$

where L is crustal thickness and τ is the average shear stress required to slip the fault. For example, if τ is 1 kbar and L is 40 km, any delaminated region more than 400 km in diameter would rise between vertical faults. Regions less than this critical size would rise by a lesser amount through elastic or anelastic warping.

Here we calculate only the maximum uplift possible in the case of perfect isostasy. Then the size and time history of excess uplift are just proportional to the thermal Bouguer gravity anomaly calculated above, neglecting the effect of the slab. The initial maximum amounts are 1300 and 1050 m for 30- and 40-km crusts, respectively. But the 40-km crust would rise to a greater total height, because a 40-km thickness is usually associated with a higher steady state elevation of 1.0 km [*Woollard,* 1959]. Figure 4 shows the time decay of these transient elevations. It is almost an exponential decay after the first 10 m.y., with a time factor of approximately exp $(-t/4.5 \times 10^7$ years). This is close to the rate obtained by *Sleep and Snell* [1976] for the subsidence of the Michigan basin. They concluded that this subsidence might have been preceded by uplift caused by 'bulk replacement of the upper mantle' and thus were probably the first to suggest what is here called delamination. Unfortunately, the recent Michigan Basin Borehole Project did not conclusively test their hypothesis [*Sleep and Sloss,* 1978].

The elevation-subsidence rates obtained from these calculations peak at about 0.02 mm/yr and then fall off exponentially. It is worth noting that if surface erosion rates are faster (and they usually are), the rocks of the crust will continue to rise even during the cooling phase. The geologic record would then show a permanent offset of at least a kilometer at the fault surrounding the delaminated region.

The strongest indications of delamination in the geologic record are regional unconformities where the underlying rocks are cut by basalt dikes and diatremes about 50 m.y. older than the superjacent sediments.

Delamination of the Colorado Plateau

One promising application of this model is to the Colorado Plateau province of the western United States. It had a simple

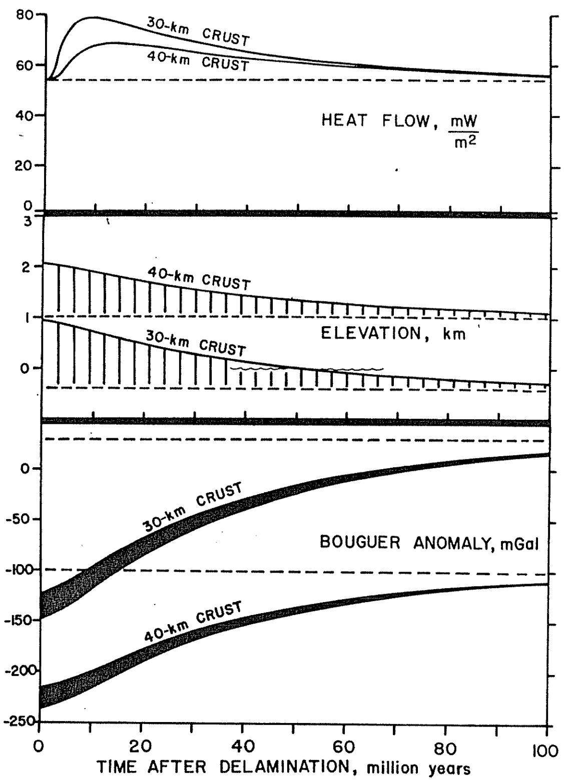

Fig. 4. Predicted heat flow, elevation, and gravity anomaly in a delaminated region as a function of cooling time, based on the one-dimensional calculation of Figure 3. Dashed horizontal lines are the assumed base line values which the solutions asymptotically approach. Uplift may fall anywhere between the extremes of none (because of crustal strength) and isostatic elevation. The gravity anomaly varies by a lesser amount depending on the amount of uplift. The effect of the detached slab, which initially offsets part of the gravity anomaly, has not been included.

geologic history until the Tertiary, when it was subjected to uplift and igneous intrusion, and no generally accepted explanation of these events is available. In this section a few data are used to constrain the free parameters in a delamination model for the region, allowing other quantities to be predicted. All the other data are clearly or arguably consistent with these predictions.

While testing this model, we will be comparing the present Colorado Plateau only with the platform areas of the eastern United States and elsewhere, not with the adjacent Rocky Mountain or Basin and Range provinces. The regions to the west were the site of pre-Tertiary continental rifting, subsidence, and subduction [*Burchfiel and Davis,* 1975], while mountains to the east were uplifted by unknown Laramide mechanisms. We may never be able to unravel these histories enough to identify the Tertiary parts of their various geophysical fields; certainly, it cannot be done in this brief paper. But the Colorado Plateau has the simplest history of any part of the West, contains no Paleozoic fold belts or intrusions, and acted as a resistant island (of intact lithosphere) in the midst of the Laramide orogeny [*Eardley,* 1951]. Therefore it is assumed here that the future plateau was no more heated, deformed, or elevated at the close of the Paleocene than the Great Plains states are today.

Previous Theories for the Plateau Uplift

The Colorado Plateau now stands at an average height of 2000 m, at least 1000 m higher than other areas of equal crustal thickness. Because of the lack of surface strains it is necessary to postulate a deep, mantle mechanism to account for this Tertiary uplift. *Kennedy* [1959] favored subcrustal convection currents to drag extra buoyant crust under the plateau but noted that there was no adjacent source area that had been depressed. *Gilluly* [1963] pointed out that thermal expansion effects in the crust are insufficient and that a phase change is a poor explanation. (We now know that mantle expansion by hydration is unlikely because temperatures are above normal.) He also favored the idea of subcrustal convection. But more recent evidence has shown that much of the isostatic support for the plateau comes from the mantle, so this theory is insufficient. Upwelling mantle flows which uplift the plateau directly, by their pressure, have been suggested many times. *Wilson* [1973] proposed a plume beneath the area; *Sbar and Sykes* [1973] generalized this to a group of plumes; and *Anderson and Perkins* [1975] proposed that reversing mantle flow caused by subduction of the East Pacific Rise might be important. A third class of theories attempts to link the uplift to plate tectonics: *Helmstaedt* [1974] argues that a slab of Farallon plate oceanic lithosphere slid below the Moho of the plateau, which rose when it was removed. *Huntoon* [1976] pointed to the regional stress change following the reorganization of plate boundaries in California. *Melosh and Ebel* [1977] suggested a shear-heating instability as a possible cause of upper mantle expansion.

None of these theories has been developed to the point where quantitative predictions of uplift or associated effects have been made and tested.

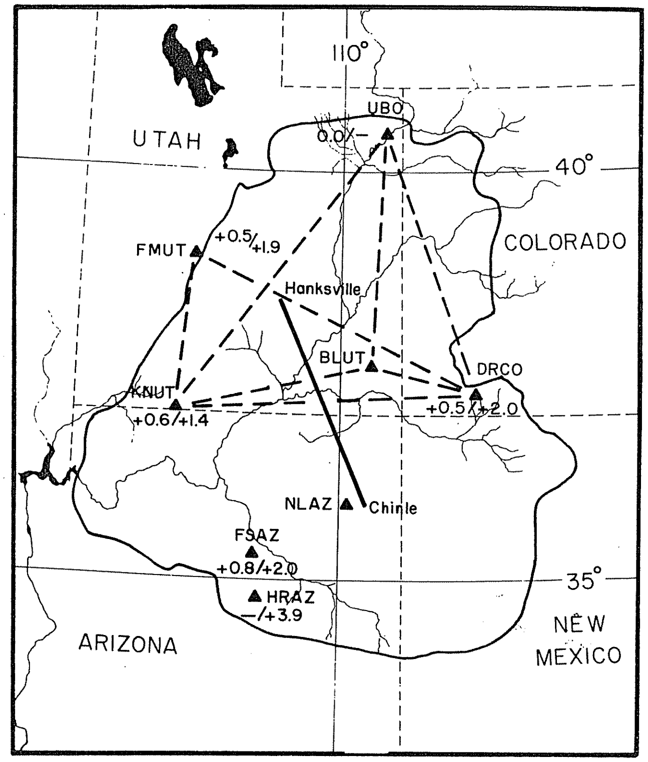

Fig. 5. Location map for seismic data on the Colorado Plateau. The plateau is outlined with a heavy curve, after *Fenneman* [1928]. Shown are the Hanksville-Chinle refraction profile of *Roller* [1965], the triangular arrays used by *Bucher and Smith* [1971] to measure Rayleigh wave phase velocities, and the stations at which *P* or *S* wave absolute residuals were determined by *Cleary and Hales* [1966] or *Hales and Roberts* [1970]. The number before the slash is the *P* wave residual in seconds (positive is late), and the number after is the *S* wave residual. A dash means no residual was determined.

Predictions of the Delamination Model

In order to construct a delamination model, one must assume an onset time, a crustal thickness, a lithosphere thickness, and the intrinsic thermal parameters of both layers. *Roller* [1965] and *Keller et al.* [1979] both find the Moho elevation to be 40 ± 3 km below sea level by different seismic methods (Figure 5), so a crustal thickness of 42 km has been adopted. Other material constants have been stated above. The only subtlety involved is in the choice of a starting time, since the plateau uplift has been variously regarded as Laramide [*Lovejoy,* 1974], late Eocene–early Miocene [*Hunt,* 1953], mid-Oligiocene [*Rowley et al.,* 1978*a*], late Miocene [*Kennedy,* 1959], or Pliocene [*Young,* 1974].

The age of 30 m.y. (mid-Oligocene) used here is based on the igneous record. *Burke and McKee* [1979] present an isochron map of the initiation of volcanism in Nevada which shows a wave of new activity advancing southwest from 42 to 28 m.y. ago at a rate of about 3 cm/yr. The left flank of this wave passes along the western boundary of the Colorado Plateau in such a way that its linear extension would pass through the Four Corners area 30 m.y. ago. Just at this time the Four Corners was intruded by the Navajo diatremes (which brought up kimberlites with mantle xenoliths) and by nearby minette plugs [*Naeser,* 1971]. This strongly suggests that the linear cause of the Nevada volcanism also passed beneath the Colorado Plateau, and it is assumed here that the cause was a regional delamination of the southwestern United States.

To complicate this simple picture, there is equally good evidence for a second igneous wave in the opposite direction. *Hamblin et al.* [1976] describe a northeastward migration of basaltic volcanism at about 3 cm/yr in the last few million years along the plateau's western boundary. Likewise, the San Francisco Peaks volcanism has migrated inward from the plateau rim, arriving 6 m.y. ago (E. W. Wolfe, personal communication, 1978). There is a second set of diatremes, at Hopi Buttes, which *Naeser* [1971] has dated at 5.5 m.y., accompanied by basalts. Their eruption was slightly preceded by regional warping, so that streams were tectonically dammed, and the Pliocene Bidihochi Formation was deposited in lakes [*Cooley and Akers,* 1961]. Thus for consistency we must also assume a second delamination which passed through northern Arizona about 5 m.y. ago and which may be still in progress!

In order to predict the consequences this would have, the thermal model discussed before was recalculated but interrupted after 25 m.y. of cooling. Then temperatures below the Moho were replaced (again) by the asthenospheric adiabat to represent the second delamination. Conduction was resumed for another 5 m.y., up to the present. The predicted heat flow, elevation, and gravity histories of this two-stage model are shown in Figure 6, with the elevation and gravity changes smoothed by estimated two-dimensional effects. The starting values 35 m.y. ago represent the assumed initial conditions based on the crustal thickness combined with *Woollard*'s [1959] compilations. Since the Colorado Plateau has been moving at 2.4 cm/yr with respect to the hot spot or average-mantle reference frame [*Minster and Jordan,* 1978], the slab from the first delamination is assumed to be far away, if not absorbed. The second (thinner) slab must still be present at some depth and will be considered below.

The same hypothesized history is shown with a time-depth plot in Figure 7, which represents the late Tertiary in the Four Corners area. The remainder of this paper will consider how well the geophysical and geologic evidence from the whole

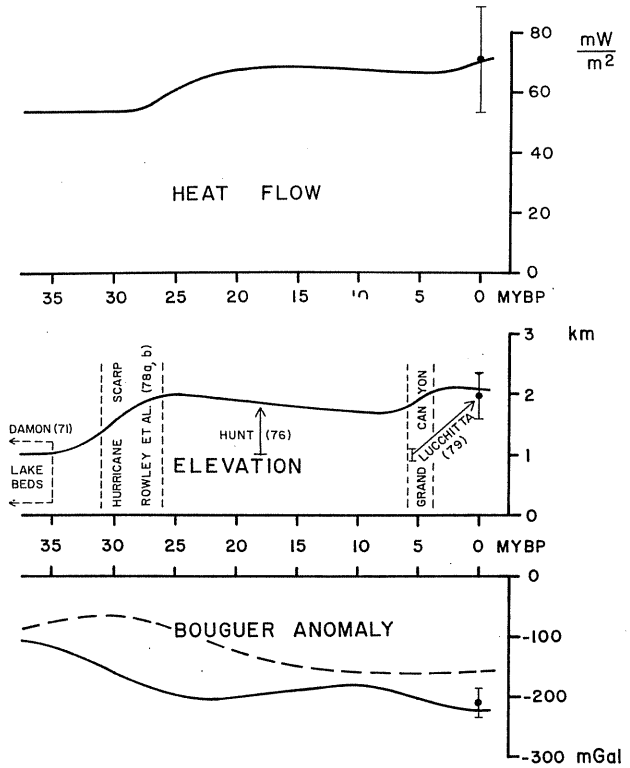

Fig. 6. Model predictions for the history of the center of the Colorado Plateau based on one delamination event 30 m.y. ago and a second one 5 m.y. ago, with propagation velocities of 3 cm/yr. Symbols at 0 m.y. ago show mean and standard deviation of present heat flow, elevation, and gravity anomaly. In the case of gravity, two possible extremes are shown: either the detached slab is instantly removed/absorbed (solid line), or else it remains a source of attraction until it sinks past 700 km at 5 cm/yr (dashed line).

Colorado Plateau match the predictions of these two figures and the preceding sections.

Geophysical Tests

The hypothesized events would have raised the present heat flow by 16 mW/m^2 (0.4 HFU; 1 HFU = 10^{-6} cal cm^{-2} s^{-1}). Added to the assumed base line flux of 54 mW/m^2 (1.3 HFU), this yields a predicted total of 70 mW/m^2 (1.7 HFU). For comparison, an average of published heat flow values has been computed within the Colorado Plateau as defined by *Fenneman* [1928], a definition which excludes the Rio Grande and San Juan volcanic fields (Figure 5). Combining the values listed by *Spicer* [1964], *Roy et al.* [1968, 1972], *Costain and Wright* [1968], *Sass et al.* [1971], and *Reiter et al.* [1975, 1979], we find that there are about 45 determinations (after holes less than 15 km apart are averaged), strongly concentrated in New Mexico and Colorado. This data set has a range of 42–133 mW/m^2 (1.0–2.95 HFU), a mean of 71 mW/m^2 (1.7 HFU), and a standard deviation of 18 mW/m^2 (0.4 HFU). Although there is some danger that this mean has been biased by magmatic heat transport, it is in excellent agreement with the model prediction. As *Reiter et al.* [1975, 1979] have stressed, the plateau heat flow is measurably above that of stable platform areas.

The prediction of present gravity is less precise, because it is hard to know how much attraction a (thin) slab of mantle lithosphere detached 5 m.y. ago would exert today. If it were completely removed by rapid heating, sinking, or upper mantle shear flow, the Bouguer anomaly should be −225 mGal (Figure 6). On the other hand, if it sank only slowly, the second delamination would not affect the Bouguer anomaly, and the present value should be −170 mGal. (Both figures include the effect of a 42-km crust uplifted against gravity forces alone.) Because the plateau has only a small positive free air anomaly [*McGinnis et al.,* 1979], it is tempting to say that the

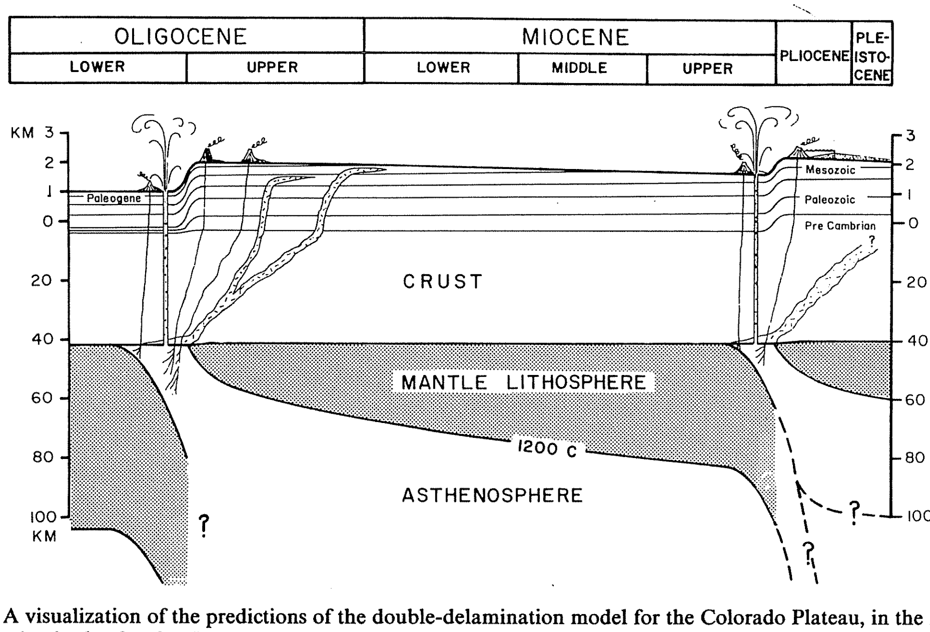

Fig. 7. A visualization of the predictions of the double-delamination model for the Colorado Plateau, in the form of a time versus depth plot for the Four Corners area. Note the scale change at sea level. Sedimentary layer thicknesses and erosion rates are only schematic. Igneous eruptions of mantle diatremes, basalt flows, and granodiorite laccoliths are only shown while in motion (and not after cooling) to maintain clarity. It is not known whether the delaminated mantle lithosphere (shaded) settles onto the mesosphere or sinks deeply into the mantle; hence the alternatives are marked by dashed lines with question marks. The lithosphere-asthenosphere boundary is always transitional; here it is represented arbitrarily by the 1200°C isotherm. Time scale by *Berggren* [1972].

situation is the former; but it is possible that crustal rigidity is creating a negative free air anomaly which masks that of the slab.

For comparison, the Bouguer map of *Woollard and Joesting* [1964] shows apparently random variation around a large negative average. The mean of 200 equally spaced points on this map is −209 mGal (within the predicted range), and the standard deviation is 23 mGal. The relative gravity maximum which follows the Colorado River through the plateau is probably caused by slight crustal arching in response to erosion.

There is considerable evidence of the predicted drop in upper mantle seismic velocities. The P_n velocity measured on the Hanksville-Chinle profile was low: 7.8 km/s [*Roller,* 1965] underlies a possible transition zone of 7.6 km/s [*Prodehl,* 1970]. Also, Rayleigh wave phase velocities determined over several large triangular arrays on the plateau (Figure 5) by *Bucher and Smith* [1971] confirmed the P_n velocity of 7.8 and established the upper mantle shear velocity as 4.25 km/s (as opposed to 4.65 km/s for normal continental shields). Travel time anomalies for body waves from distant earthquakes also confirm an anomalous slowness. *Cleary and Hales* [1966] found *P* wave delays from zero to +0.8 s, whereas a 10-km thickening of crust usually causes only +0.2 s. *S* wave delays measured by *Hales and Roberts* [1970] were proportionately larger: +1.4 to +3.9 s (Figure 5). These are absolute residuals for events where the source correction was independently determined, but the slowness also shows up in relative residuals. *Yasar and Nuttli* [1974] showed KNUT and NLAZ (Figure 5) to be the latest stations in the West for *S* waves from Japan and South America. There has not been any local study of mantle attenuation in this area, but *Der et al.* [1975] show that the Colorado Plateau has the same general high attenuation as the rest of the American West. Clearly, the upper mantle here is just as unusual (and probably as hot) as beneath the adjacent Basin and Range province.

The only geophysical data which seem inconsistent with the model are the geomagnetic depth sounding results of *Porath* [1971]. He presents models which show the Colorado Plateau almost as resistive as the Great Plains and definitely more resistive than the Basin and Range or Southern Rocky Mountain provinces adjacent. Since the data show no frequency dependence, the depth resolution is nil, but these differences are interpreted as indicating a thick lithosphere under the plateau. The model presented here has a lithosphere only 60 km thick at present (Figure 7).

There are two possible ways to excuse this embarrassing fact. First, the conductivity of the asthenosphere may have been overestimated, and the survey may only reveal differences in lower crustal conductivity. This was the interpretation that *Hyndman and Hyndman* [1968] and *Caner* [1970] preferred for the data from the Canadian Rockies. Lower conductivity of the plateau crust would be consistent with heat flow, which is lower than in surrounding regions [*Sass et al.,* 1971]. Second, it is still possible under the delamination hypothesis to have a thin asthenosphere under the plateau. This would be the case if the second mantle slab came to rest on the mesosphere, which *Sleep* [1979] locates beyond a depth of 200 km. In that case the mantle lithosphere and asthenosphere would merely have changed places, and the asthenosphere would be only about 40 km thick today (Figure 7). This suggestion rests on a single piece of evidence: *Julian* [1970] reported a step increase of *P* wave velocity with depth from 7.8 to 8.1 km/s at 100 km under the plateau. Could this perhaps be the top of the sunken mantle lithosphere?

Speculation aside, we have a problem in interpretation because seismic and geomagnetic depth sounding methods give opposite characterizations of upper mantle temperature beneath the plateau. More geomagnetic data allowing for better depth resolution are needed to solve this problem.

Geological Tests

The delamination model predicts an igneous suite of basalt plus granodiorite plus kimberlite. Above, the kimberlitic (or monchiquitic) diatremes of the Colorado Plateau and their associated basalts (or minettes) were mentioned and used to estimate the times of delamination. But something very like granodiorite is also present: the diorite laccoliths of the Henry, La Sal, and Abajo mountains. *Hunt* [1958] considered them products of crustal melting, and it is suggested here that they resulted from contact between the lower crust and hot asthenosphere in the first delamination. The upper Oligocene ages [*Snyder et al.,* 1976] of the latter two intrusions are about what we would expect (Figure 7).

The most controversial data from the Colorado Plateau have been those concerning the age of its uplift. No single date appears to satisfy all constraints, so it is common to divide the uplift into at least two phases [e.g., *Dutton,* 1970; *Huntoon,* 1976]. The delamination model predicts two such phases (Figures 6 and 7), and the evidence is consistent with the prediction. Unfortunately, all of this evidence is from the western and southwestern margins, which may not be representative.

At the close of the Laramide orogeny some 60 m.y. ago, the future plateau was above sea level but still a relative lowland. This is shown by the deposition of Eocene lake deposits on top of Laramide structures [*Eardley,* 1951]. *Damon* [1971] estimates that the average elevation may have been as much as 1200 m. This moderate elevation is what would be expected of a normal shield with slightly thick crust.

A possible indication of the first hypothesized uplift is the appearance of the Hurricane Fault scarp between 31 and 26 m.y. ago [*Rowley et al.,* 1978*a*, *b*]. This does not prove that the plateau rose; the adjacent basin to the west might have sunk, but at least there was tectonic activity associated with volcanism.

Hunt [1976] has established that 18 m.y. ago the Colorado River flowed through Peach Springs Canyon, where it cuts 1000 m below the present top of the plateau. At that time, the Gulf of California did not exist, and the river's path to the sea would have been longer. So it is not unreasonable to propose that this point on the river was several hundred meters above sea level. Furthermore, the depth of the fossil canyon has been reduced since then by erosion of the plateau, so the separation might well have been 1500 m then. By these arguments the model prediction of 1800 m (Figure 6) can be reconciled.

The second uplift that is predicted has been documented by *Lucchitta* [1979], who traced stratigraphic indicators of sea level onto the southwest corner of the plateau. He was able to show an additional 900 m uplift there, sometime in the last 5.5 m.y. A bit of circumstantial evidence that this happened in the early Pliocene is that the Colorado River shifted its course and cut the Grand Canyon prior to 3.8 m.y., when basalts began to flow into it.

Finally, we note that the predicted present elevation of 2100 m is correct. When elevations were sampled at $\frac{1}{2}°$ intervals of latitude and longitude inside *Fenneman*'s [1928] plateau boundary, the result was a mean of 1990 m and a standard deviation of 390 m, as shown in Figure 6.

CONCLUSIONS

This study demonstrates both theoretically and with a practical example that mantle convection beneath continents may extend right up into the base of the crust. The convection that removes the chilled boundary layer of mantle lithosphere from a shield is a finite amplitude instability which starts at a linear rift in the lithosphere and propagates laterally at several centimeters per year. The mantle pulls away from the crust and sinks downward in a style analogous to 'retreating subduction,' leaving in its place a layer of hot asthenosphere which has risen adiabatically to replace it. One may regard the heavy mantle lithosphere as sucking the asthenosphere upward or consider the buoyant asthenosphere to force its way between the lighter crust and heavier lithosphere; the difference is only semantic.

Although the surface effects during this process are not known, the effects after a delamination front has passed are quite predictable. The drop in pressure experienced by the rising asthenosphere causes segregations of partial melts which erupt as basalts and of carbon-dioxide-rich magmas which erupt through diatremes. The bulk replacement of mantle by hotter and less dense rock causes an increase in heat flow, a more negative Bouguer gravity anomaly, a drop in seismic velocity and Q, and isostatic uplift (if the region is large enough) by about a kilometer. The effects which are linearly related to temperature decay with a half-life of 3×10^7 years as a new boundary layer forms below the crust. After 100 m.y. the shield is re-formed, and only igneous rocks and signs of erosion mark the past events. Whether the region remains stable or goes through another convective cycle will depend on the sources of the necessary initiating flaw.

It has been shown that this cycle, repeated twice, can explain many of the events in the Colorado Plateau since the Laramide orogeny. Setting the ages of delaminations by the ages of diatremes, it was found that the model successfully predicts the heat flow, gravity anomaly, seismic velocity structure, age of basalts, presence of laccoliths, history of uplift, and present elevation. The only data which are difficult to reconcile with this model are the low mantle conductivities inferred from geomagnetic depth sounding.

This model provides a mechanism for uplift and igneous activity in continental interiors far from ridges and subduction zones. It may also contribute to the understanding of kimberlitic diatreme emplacement and thus enable us to read the ultramafic xenoliths in their proper context. If it proves successful, it may even assist in combatting the new dogma that every geologic event is a result of plate tectonics.

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