The active southwest margin of the Colorado Plateau: Uplift of mantle origin

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ABSTRACT

During Cenozoic time, the Colorado Plateau was raised about 2 km above sea level. The most-recent and best-documented uplift of the plateau (~1 km) has been concentrated at its southwest margin between 6 and 1 Ma, whereas the eastern Colorado Plateau may have been at high elevations since Eocene time. To better understand the recent tectonic activity at the southwest margin of the Colorado Plateau, we compile detailed crustal thickness and density information from seismic and gravity data for a region that includes northwest Arizona and the southern tip of Nevada. This information is used to isolate the mantle contribution to uplift. We find that there is relatively low density mantle underlying the southern margin of the plateau in northwest Arizona, which could result from about 60–80 km of thinning of the dense mantle lithosphere combined with about 100 °C of heating through a 100-km-thick mantle layer. The available estimates from earthquake-source seismology in or near the study area are compatible with this estimate of lithospheric thinning. We speculate that uplift may result from subduction-related thinning of the continental lithosphere.

INTRODUCTION

The Colorado Plateau is a major tectonic and physiographic province in the southwestern United States (Fig. 1) that has behaved as a relatively stable, coherent block during much of Phanerozoic time. A site of marine deposition during Cretaceous time, the Colorado Plateau now stands about 2 km above sea level, implying that nearly 2 km of uplift occurred during Cenozoic time. The greatest amount of uplift has apparently been along the southwestern margin of the Plateau, where elevations are often 0.5 km greater than in the center (e.g., Lucchitta, 1989). Unlike the Basin and Range Province and Rio Grande rift, which have experienced ~1 km of uplift while simultaneously undergoing horizontal extension and internal deformation, the Plateau has remained a relatively rigid block, resistant to faulting. The interpretation of an apparent rigid-block behavior for the Colorado Plateau is reinforced by paleomagnetic observations of coherent rotation of the plateau (e.g., Bryan and Gordon, 1986; Wells and Hillhouse, 1989).

Given that the Colorado Plateau is in isostatic equilibrium now (the free air gravity anomaly is nearly zero; Thompson and Zoback, 1979), and assuming that it was in the past, then some growing mass deficiency at depth must have driven its uplift. Several mechanisms have been proposed to account for the most recent phase of uplift, including thermal expansion, crustal thickening, and delamination of the lithosphere (e.g., Bird, 1979; Thompson and Zoback, 1979; McGetchin et al., 1980; Bird, 1984; Morgan and Swanberg, 1985). Because the crust is the buoyant component of the lithosphere while the mantle lithosphere acts as the dense keel (e.g., Lachenbruch and Morgan, 1990), one must first determine the density and thickness of the crust before the mantle contribution to uplift can be isolated. If it is determined that the crust is not buoyant enough to float the plateau at its current elevation, then a more dynamic mantle process is indicated. In this paper we use seismic refraction results from the Pacific to Arizona Crustal Experiment (PACE) (McCarthy and Parsons, 1994; Parsons et al., in

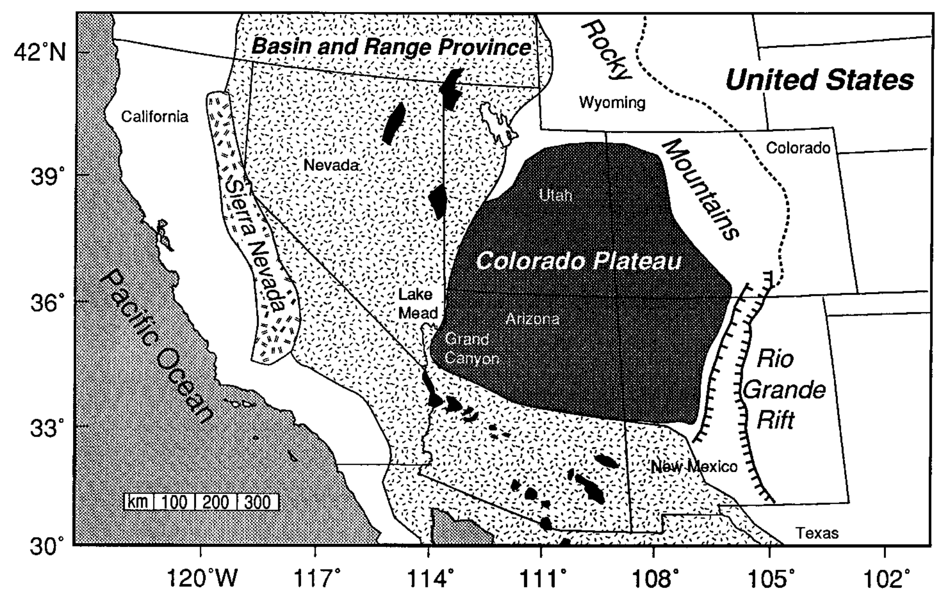

Figure 1. Selected tectonic provinces of southwestern United States. The Colorado Plateau is a roughly square block bounded by the Basin and Range and transitional zones to the west and south, the Rio Grande rift to the east, and the Rocky Mountains to the north. The black areas are locations of metamorphic core complexes.

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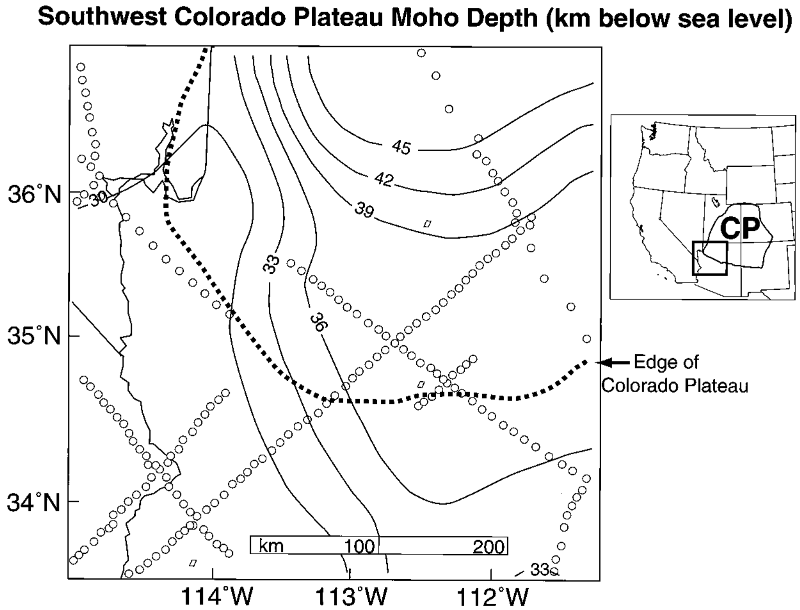

Figure 2. Smoothed Moho depths in study area in northwest Arizona. White dots depict locations of seismic refraction profiles used to create three-dimensional image of the Moho. The crust beneath metamorphic core complexes (locations shown in Fig. 1) is thin (27–30 km thick) and gradually thickens beneath margin of Colorado Plateau, reaching a maximum thickness of 48 km beneath the heart of the plateau. The Arizona-California-Nevada state lines are shown on this and subsequent figures for location purposes.

press) combined with residual gravity data (Saltus, 1991) to estimate the crustal contribution to surface elevation. We further speculate on the mantle contribution to elevation of the Colorado Plateau and/or the role of subduction in the uplift and extension of the southwestern United States.

Uplift History

Sediments deposited on the Colorado Plateau indicate that the present-day high elevation of the plateau is the result of Cenozoic tectonic events. Eocene gravels, originating from the southwest, have been deposited directly onto Paleozoic strata and indicate that the plateau stood at a lower elevation than the bordering Basin and Range Province before Eocene time (Lucchitta, 1979). Young basalt flows at the present-day western rim of the Grand Canyon and on the canyon floor are 6 and 1 m.y. old, respectively. These basalt flows are both cut by the Colorado River, requiring that the cutting of the canyon (and an implied 1 km of uplift at the southwest plateau margin) occurred during this time interval (Lucchitta, 1989). Crustal thickening suggested to have been related to horizontal subduction of the Farallon plate (between 70 and 40 Ma) (e.g., Dickinson and Snyder, 1978) and subsequent removal of the slab either by a steepening in the subduction angle or by thermal assimilation during the latest Eocene (e.g., Eaton, 1979; Zoback et al., 1981) is proposed to have elevated the eastern Colorado Plateau (e.g., Bird, 1988), which may have stayed high since that time (e.g., Gregory and Chase, 1992). The well-documented Miocene and Pliocene uplift of the Colorado Plateau occurred primarily at its southwest margin (Lucchitta, 1989).

Seismic Studies and Estimates of Crustal Thickness

In this study we use seismic refraction profiles collected by the U.S. Geological Survey as part of the Pacific to Arizona Crustal Experiment (McCarthy and Parsons, 1994; Parsons et al., in press) in conjunction with a network of older refraction profiles, teleseismic receiver function results (Ruppert, 1992), and a fan-geometry wide-angle reflection profile (Howie, 1991) to model crustal thickness from the Basin and Range Province, across the transitional margin, and into the Colorado Plateau. Collectively this regional coverage is sufficient to enable us to generate a three-dimensional image of the Moho in northwest Arizona (Fig. 2). Early seismic refraction experiments conducted on the plateau found intermediate thicknesses of 40–43 km (Roller, 1965; Warren, 1969). A reinterpretation of the Roller (1965) study by Prodehl (1979) found the same result as the initial interpretation. Hauser and Lundy (1989) combined new deep seismic reflection data recorded on the Colorado Plateau with a reinterpretation of the Roller (1965) and Warren (1969) data to suggest that the plateau is at least 50 km thick. More recently, interpretation of the PACE results across the Colorado Plateau yielded similar estimates of crustal thickness to that of Roller and Warren (Wolf and Cipar, 1993; Parsons et al., in press). Because of general agreement in the refraction results, the conflicting nature of the refraction and reflection estimates of crustal thickness, and the ambiguity of interpreting crustal thickness from vertical-incidence reflection data, we have chosen not to include the Hauser and Lundy (1989) results in this compilation.

PACE refraction results across the southern Basin and Range Province show a fairly uniform crustal thickness in the southwestern part of the study area, varying from about 27 to 30 km beneath the metamorphic core complexes along the Colorado River (McCarthy et al., 1991). Lower-crustal ductile flow combined with magmatic addition to the crust (Bird, 1991; McCarthy et al., 1991; McCarthy and Parsons, in press) have probably maintained this nearly uniform crustal thickness within the southern Basin and Range Province (Fig. 2) despite strongly varying magnitudes of crustal extension. The slope of the Moho steepens beneath the transition zone between the Basin and Range and Colorado Plateau and has a rough inverse correspondence with the shape of the smoothed topography (Fig. 3); crustal thickness in this region ranges between 30 and 36 km. The crust beneath the active margin of the plateau is thicker (38–42 km), increasing gradually to the northeast (Kohler and McCarthy, 1990; Parsons et al., in press). In southern Utah and northern Arizona the crust is thickest, reaching an estimated depth-to-Moho of 48 km beneath the Kaibab plateau where elevations are greatest (Wolf and Cipar, 1993; Parsons et al., in press).

CALCULATING THE MANTLE CONTRIBUTION TO UPLIFT

To investigate the mantle beneath the Colorado Plateau, it is first necessary to estimate the thickness and density of the crust.

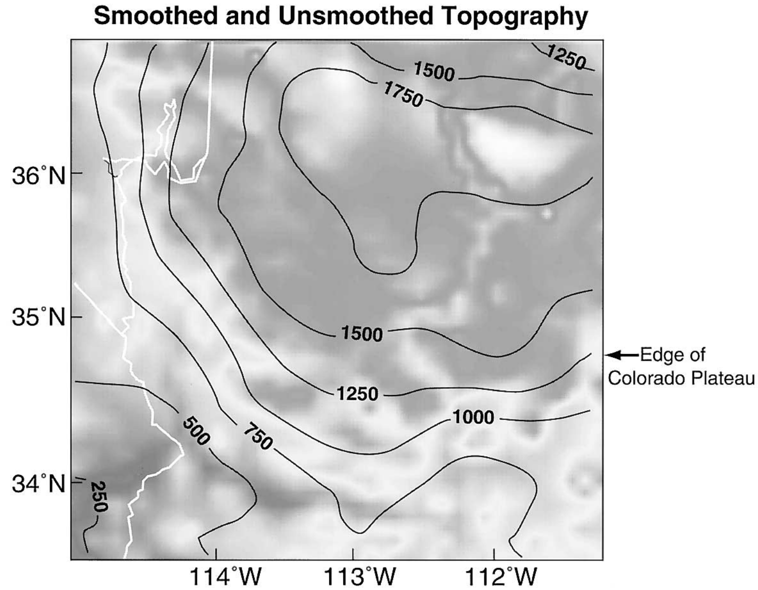

Figure 3. Topography of northwest Arizona. The gray-scale contour image is unsmoothed and shows detailed variation in elevation. Contour lines have been smoothed using a near-neighbor kernel to show general trend in elevations (in meters) that we match with isostatic modeling. Transitional edge of the Colorado Plateau corresponds to the 1250 m contour line and trends northwest across Arizona.

If these two parameters are known, then it is possible to estimate the crustal contribution to uplift and thereby isolate the mantle contribution. If the crustal buoyancy alone cannot float the lithosphere enough to account for the present elevations, then a mantle contribution is indicated. In our calculations, we use a compilation of crustal thickness incorporating the refraction studies discussed above. Subsurface crustal density cannot be measured directly and has been estimated in the following manner. Variation in the gravitational field of the basement rocks allows an estimate of the relative density of the crust down to midcrustal levels (15–20 km). Saltus (1991) compiled an isostatic residual basement gravity grid covering southern California, Nevada, Arizona, and Utah. This compilation removes the effects of sedimentary basins and isolates the gravity signature of the upper-crustal crystalline basement. The long wavelength variations that have a deep source are removed, leaving behind the short wavelength features that have their source in the upper crust. A similar map was compiled for the entire United States, and short wavelength features were found to correlate well with mapped surface features; the highs correlated in general to mafic igneous bodies and uplifted crystalline basement, whereas lows correlated to felsic intrusive bodies and depression of the crystalline basement rocks (Jachens et al., 1989). Density perturbations of the upper-crustal rocks may be approximated using the formula for the attraction of an infinite slab, where

$$\rho \approx (g/0.04L_{uc})$$

(Simpson and Jachens, 1989), where g is gravity in mgal and L_{uc} is an assumed upper-crustal thickness (15 km) (Fig. 4). We are required to make the assumption that the lower-crustal rocks are of a uniform density (~2.9 g/cm^3), because no independent means of relative density determination at depth are available. If the lower crust is of uniform density, then an incorrect density estimate for the lower-crustal layer would produce a static shift in our mantle buoyancy calculations, which is unimportant because we are primarily interested in the relative buoyancy contributions across the study area. Possibly more important are lateral density variations within the lower crust; however, seismic refraction modeling has reported relatively uniform lower-crustal velocities beneath the Colorado Plateau and transition (e.g., Prodehl, 1979; McCarthy et al., 1991; Wolf and Cipar, 1993; Parsons et al., in press), which suggests a roughly uniform composition for the lower crust across the study area. Regional lower-crustal flow that is inferred to have occurred (e.g., McCarthy et al., 1991; McCarthy and Parsons, 1994) may have homogenized the lower-crustal layer to some extent, possibly reducing errors caused by the uniform density approximation. For reference, Meissner

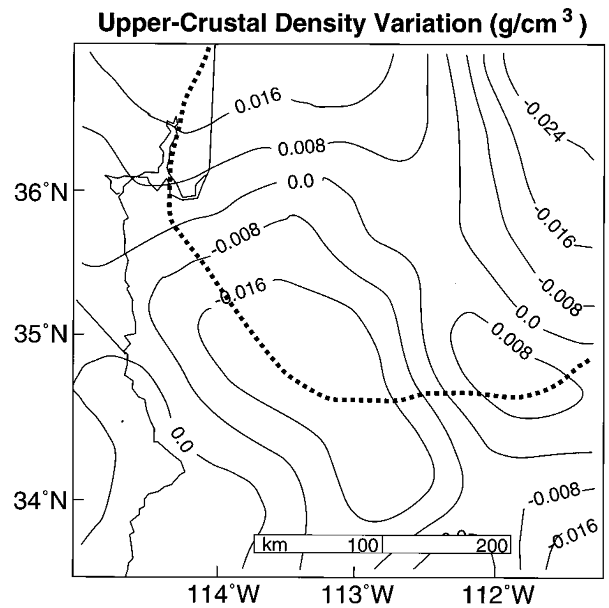

Figure 4. Relative variation in upper-crustal density across study area calculated with basin-stripped residual gravity data after Saltus (1991). Relative highs and lows are indications of high- or low-density upper crust inferred by this method. Crustal buoyancy was estimated assuming a linear relation between rock density and gravity (see text) for the upper crust, and a uniform lower-crustal density.

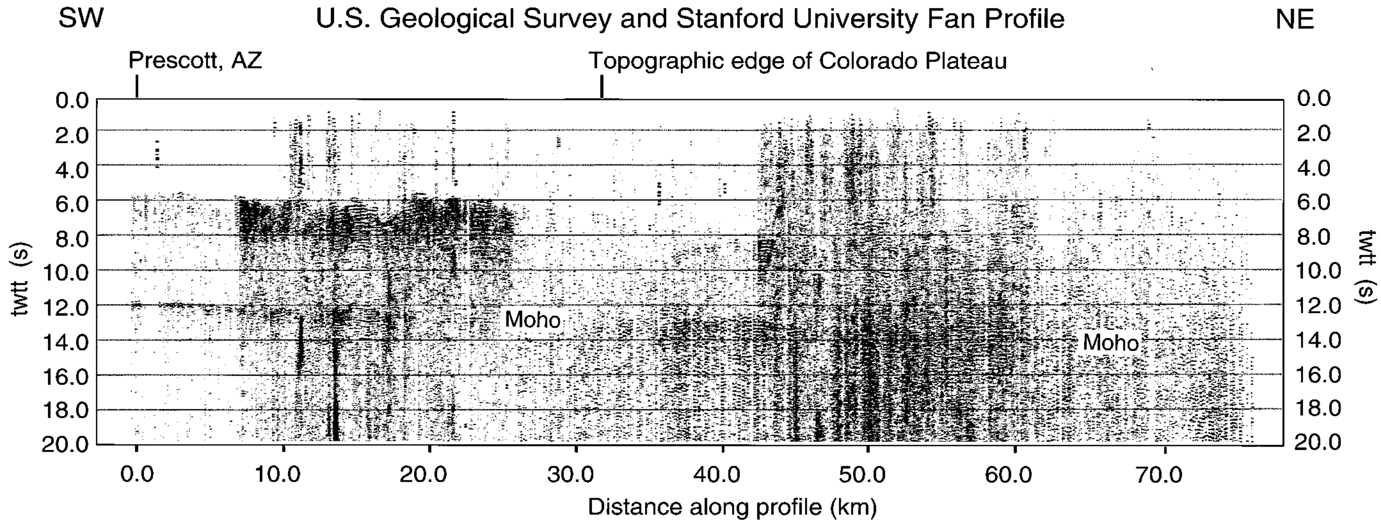

Figure 5. Wide-angle fan-geometry reflection profile recorded jointly by U.S. Geological Survey and Stanford University at southern margin of Colorado Plateau in Arizona (Howie, 1991). This profile was created by recording two shots fired broadside into the northeast-trending PACE seismic array. Bottoming points of Moho reflection from the shots located 80 km east of the recording array are thus about 40 km east of the array. The reflection from Moho (PmP) begins at 12 s two-way traveltime (twtt) (~37 km depth) at south end of profile, smoothly descending to about 14 s twtt (~42 km depth) at the north end. Moho image shows that there is no discernible step or sudden increase in crustal depth at the edge of the Colorado Plateau. Local topographic variations are not manifested as short-wavelength variations in crustal thickness but rather are compensated at shallower levels in the crust, or are supported by the strength of the crust.

(1986) gave average densities for the following lower-crustal rocks (corrected for pressure): granodiorite, ~2.72 ± 0.05 g/cm^3; diorite, ~2.85 ± 0.1 g/cm^3; diabase, ~2.95 ± 0.15 g/cm^3; and gabbro, ~3.0 ± 0.1 g/cm^3. We include an assessment of the magnitude of possible uncertainties associated with lateral density variation in the lower crust in the following section.

The Mantle Beneath the Colorado Plateau Margin

Worldwide, the Earth's crust is generally underlain by a dense mantle lithosphere layer (or lid) that acts as a keel, countering the buoying effect of the less-dense crust. To address the mantle role in uplift, we calculate the model-independent parameter of mass per unit area. The mantle mass per unit area is the required variation in mass over a given mantle volume that, combined with the buoyant crustal layer, will satisfy regional isostasy consistent with the observed topography. We calculate the mantle mass rather than a mantle density anomaly in order to combine two unknown parameters, mantle-lithosphere thickness and density contrast relative to the asthenosphere, into a single variable of mass beneath a given area. We compute this variation in mantle mass using smoothed topography (Fig. 3) for two reasons: first, the refraction data coverage is such that the depth to the Moho boundary has a lateral resolution on the order of a few tens of kilometers, and second, the short-wavelength features shown in the unsmoothed topography of Figure 3 are probably supported by the strength of the crust rather than by isostatic compensation at the Moho. For example, a high-resolution (25- to 50-m-station spacing) wide-angle reflection image of the Moho beneath the margin of the Colorado Plateau demonstrates the smoothness of that boundary as it gradually increases from 38 to 42 km in depth, despite local variation in topography (Fig. 5).

We find the relative mantle mass deficit per unit area by dividing the study area into a grid of isostatic columns. The mass in these columns is calculated using a two-layer isostatic expression with a known crustal layer and a second mantle layer that extends down to an assumed local isodensity asthenosphere at a depth of 200 km. The problem can be viewed as a group of 200-km-deep columns floating in an asthenosphere of fixed density, with each column containing the unknown boundary between the mantle lithosphere (of unknown density) and the asthenosphere. Because we treat the crust as a known quantity, and because we know the elevation of its top and bottom, thus we can calculate the mass contained within each mantle column that is required to float it to its individual height. The relative mantle mass deficit per unit area of a given column (M_n) is found by comparing the mass of a given column (I_n) to a reference column (I_{ref}), which is the column in the study area that contains the greatest amount of mantle mass (and hence has the lowest mantle buoyancy), as

$$M_n = (I_n - I_{ref}),$$

where

$$I_n = \rho_a E_n - (\rho_a - \rho_c)\, L_{cn}$$

is a modified form of the isostatic equation (e.g., Lachenbruch and Morgan, 1990). The variable L_{cn} is the thickness of the crust in column n, E_n is topographic elevation, and ρ is density with the a and c subscripts denoting asthenosphere and crust, respectively. We assume that the asthenosphere is mobile enough at 200 km depth to be at approximately the same density across the study area; thus fixing its value (we set it at 3.2 g/cm^3 after Lachenbruch and Morgan, 1990) does not affect the relative mass deficit between columns.

The isostatic equation in this form yields an image of the relative variation in the buoyant influence of the mantle. Mass variation in the mantle implies variation in its

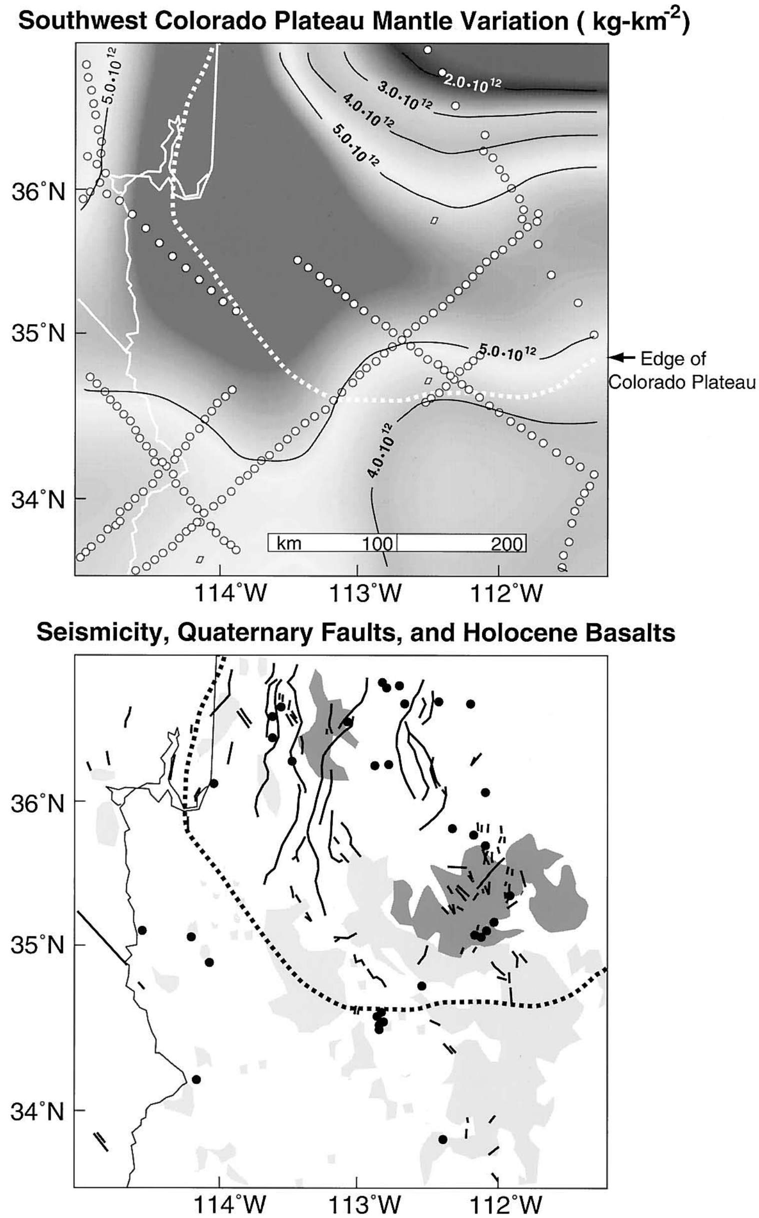

Figure 6. Representation of mantle contribution to lithospheric uplift beneath northwest Arizona. Contours of relative mantle mass deficit per unit area were made by comparing the buoyancy of isostatic columns. The crustal thicknesses and densities are treated as known quantities as determined from gravity and seismic refraction data. The asthenosphere density was held fixed at 3.20 g/cm^3. Since the mantle-lid thickness and its density relative to the asthenosphere are unknown parameters, we calculate mass variation between columns, which combines those two parameters into a single value. The numbers represent missing mass as compared with the column containing the greatest amount of mass in the study area. A zone of apparent low-density or thin mantle lid is suggested beneath the margin of the Colorado Plateau, especially beneath the Lake Mead and western Grand Canyon area. Plotted below are faults with Quaternary offsets, basaltic volcanic rocks (light gray 16–5 Ma, darker gray 5–0 Ma), and recent earthquake epicenters (black dots; M of 1.5 and greater) (after Menges and Pearthree, 1989). There is a rough correspondence between the most tectonically active areas and anomalous mantle.

density through a given volume. The primary mechanisms that can generate an apparent net loss in mantle density are (1) localized thinning of the dense mantle lithosphere and replacement by less-dense asthenosphere and (2) localized heating and thermal expansion of the mantle lithosphere and/or asthenosphere. The depth to which the observed variation in mantle mass deficit per unit area extends is unknown, but deep-seated sources are likely to have broad surface expression, whereas strong local variation in the mantle mass content is likely to have a shallower source. As such, a map of the local variation in mantle mass deficit per unit area may be viewed as a map that indicates either variations in lithospheric thickness, variations in the thermal state of the upper mantle, or a combination of the two. For reference, a 1×10^{12} kg/km^2 value translates to about 20 km of relative thinning of mantle lithosphere that is 0.05 g/cm^3 denser than the asthenosphere, or a temperature increase of about 100 °C over a 100-km-thick layer with a 3×10^{-5} °C^{-1} coefficient of thermal expansion.

Using the isostatic equation, we have computed the mantle mass deficit per unit area across the PACE transect and surrounding region in northwest Arizona (Fig. 6). The results show a relatively uniform mantle mass deficit across the Basin and Range Province at $\sim 4 \times 10^{12}$ kg/km^2 (roughly equivalent to 80 km of mantle lithosphere thinning relative to the Colorado Plateau interior). We calculate the mantle mass deficit per unit area relative to the smallest mantle contribution within the study area; thus the mantle mass deficit is shown decreasing sharply to the north and northeast, an indication that the thicker crust there has a greater isostatic contribution compared to the thinner crust of the plateau margin. The greatest relative mantle mass deficit is $\sim 5 \times 10^{12}$ kg/km^2 (approximately equivalent to 100 km thinning of mantle lithosphere) beneath the Colorado Plateau margin and is especially strong beneath southern Nevada and northern Arizona in the vicinity of Lake Mead and the western Grand Canyon (Fig. 6). An important assumption we make is that the crustal thickness has not increased at the Colorado Plateau margin since middle Miocene time. If the crust there were thickened during recent uplift, then that uplift could be explained simply by isostatic response to a thickened buoyant layer. The western Colorado Plateau is surrounded by the Basin and Range Province, which has been in a variable state of extension

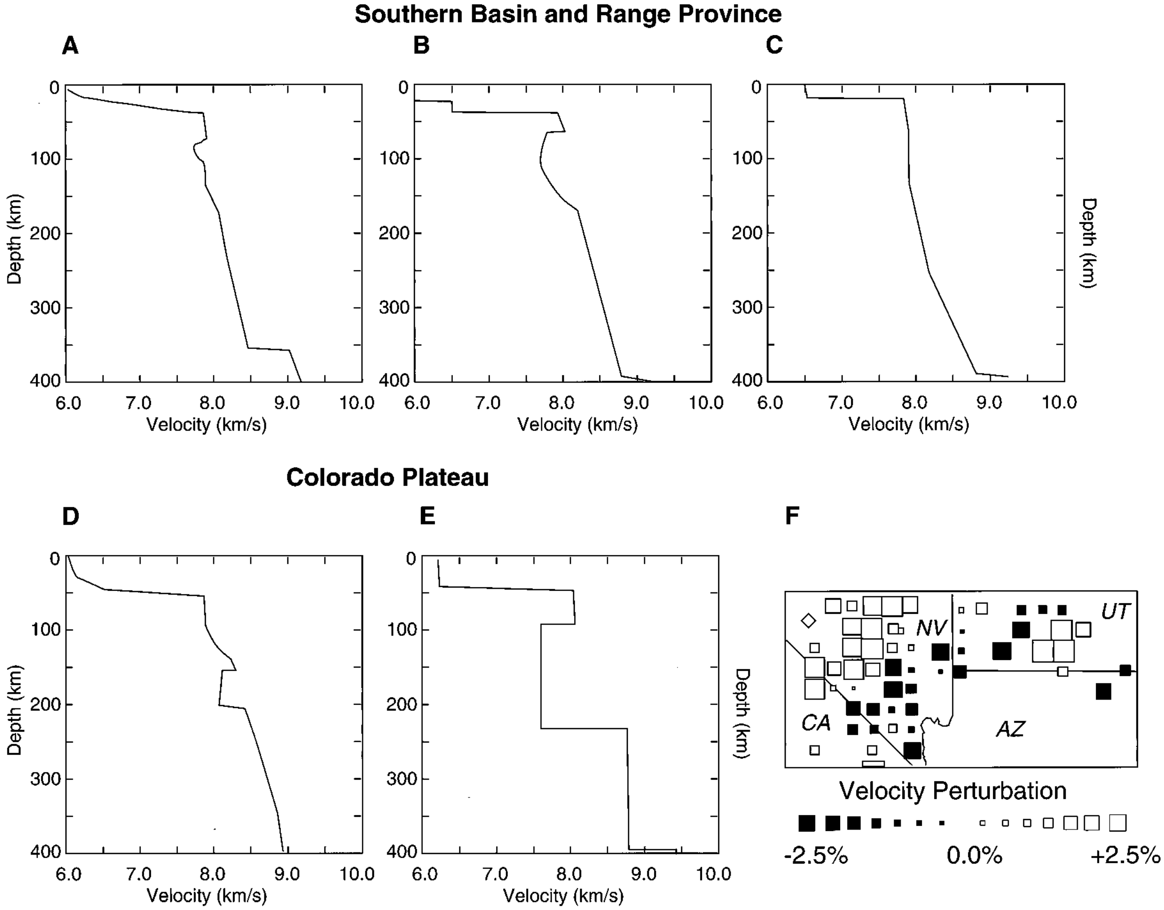

Figure 7. Selected one-dimensional velocity depth profiles of upper 400 km of southern Basin and Range Province and Colorado Plateau. A thin (20–50 km thick) high velocity lid tends to be observed above a 40- to 100-km-thick low-velocity zone in the southern Basin and Range. Mantle lid is thicker beneath Colorado Plateau at about 50–100 km thick. A P-wave delay model (window F) shows lower velocity mantle beneath southern Nevada and the Lake Mead region. References: (A) Gomberg et al. (1989), (B) Burdick and Helmberger (1978), (C) Walck (1984), (D) Wiggins and Helmberger (1973), (E) Beghoul et al. (1993), (F) Dueker and Humphreys (1990).

since at least Oligocene time; thus recent crustal thickening of the Colorado Plateau that postdates the Laramide Orogeny seems highly unlikely, nor is there any surface geologic evidence that indicates Miocene or younger compression.

We have forced all of the mass deficit to be in the mantle by assuming a known crustal buoyancy. It is useful to explore what contribution to the calculated apparent mantle mass deficit could be caused by density variations in the lower crust. If an extreme lateral variation in lower-crustal density were to exist across the study area (using Meissner's [1986] values), then complete replacement of a gabbro ($\sim$3.00 g/cm^3) lower crust with granodiorite ($\sim$2.75 g/cm^3) over a maximum 20 km thickness could produce the entire $\sim 5 \times 10^{12}$ kg/km^2 relative mass deficit. However, modeling of the seismic refraction data (Prodehl, 1979; McCarthy et al., 1991; Wolf and Cipar, 1993; Parsons et al., in press) does not show the strong lateral decrease in lower-crustal velocity that would be required with such a major compositional change ($\sim$1 km/s difference; e.g., Holbrook, 1988). In fact, the average crustal velocity is higher (suggesting an associated density increase) beneath the Colorado Plateau margin where the strongest mass deficit is calculated, than beneath the southern Basin and Range Province (McCarthy et al., 1991; McCarthy and Parsons, 1994). A mechanism for recent (<6 Ma) replacement of the southwest Colorado Plateau margin lower crust with less dense rock is difficult to conceive; magmatism at the southwest plateau margin has been primarily mafic in composition (e.g., Armstrong and Ward, 1991). Thus, although it is numerically possible to account for the entire mass variation through lower-crustal density variation, the available crustal models show no indication that this occurs.

Further insight as to whether the calculated lithospheric density variation is in the crust or mantle may be found through examination of independent seismic investigations of the mantle. Estimates of lithospheric thickness at or near the Colorado Plateau margin have been attempted using earthquake-source methods. Humphreys and Dueker (1994a) found slower relative upper-mantle seismic velocities beneath the Lake Mead area and higher relative velocities in the surrounding areas (Fig. 7F). These results were interpreted by Humphreys and Dueker (1994b) as about 50–70 km of lithospheric thinning beneath the Lake Mead–St. George (Utah) area relative to the Colorado Plateau. Scattered one-dimensional lithospheric thickness estimates are summarized by Iyer and Hitchcock (1989) as about 100 km of lithospheric thinning beneath southern Nevada relative to the eastern Colorado Plateau (Fig. 7, A–D). Lithosphere thicknesses for the Basin and Range Province and Colorado Plateau were calculated by Beghoul et al. (1993), who suggested that the mantle-lid thickness beneath the Basin and Range Province decreases about 20 km relative to the Colorado Plateau (Fig. 7E). Combining the seismic evidence for mantle-lithospheric thinning with the observed uniform crustal seismic velocities causes us to suggest that the calculated mass variation beneath our study area is primarily in the upper mantle.

DISCUSSION

An independent indication of the anomalous upper mantle beneath the Colorado Plateau margin is the recent volcanism, uplift, and extension of the region. The region where we calculate the greatest mantle contribution to isostatic uplift is located beneath the active southwest margin of the plateau where increased seismicity (e.g., Smith, 1978) and high heat flow (e.g., Lachenbruch et al., in press; Sass et al., in press) are observed. This zone of low-density mantle correlates broadly with regions of neotectonic activity, as defined by Menges and Pearthree (1989). Areas of Quaternary faulting, uplift, seismicity, and basaltic volcanism are all associated with regions of strongest mantle mass deficit (Fig. 6). A northward progression of late Cenozoic volcanic activity across the southern margin of the Colorado Plateau is well documented (e.g., McKee and Anderson, 1971; Armstrong and Ward, 1991). Moyer and Nealey (1989) summarized the trend of rhyolite eruptions across the transition from the southern Basin and Range Province onto the Colorado Plateau; these rhyolite eruptions decrease in age progressively from a maximum of 15.1 Ma at the southern edge of the plateau transition in the Casteneda Hills field to as young as 2.7 Ma at the Colorado Plateau rim in the Mount Floyd field. Nealey and Sheridan (1989) concluded that the bulk of magmatic activity at the transitional margin of the Colorado Plateau occurred from 10 to 5 Ma, and that the activity has shifted northward into the Colorado Plateau interior during the past 5 m.y. (Fig. 6). The most-recent (<5 Ma) basalts erupted on the Colorado Plateau margin are indistinguishable in composition from ocean island basalts, which implies an asthenospheric source for these basalts (e.g., Fitton et al., 1991). In addition to volcanism and uplift, the age of extension is progressively younger to the north-northwest along the Colorado Plateau transition. The onset age of low-angle detachment faulting in the southern Basin and Range Province is younger to the north, with the oldest core complexes having initiated in northern Mexico and southeast Arizona (e.g., Glazner and Bartley, 1984; Axen et al., 1993). The most-recent extensional faults (with Quaternary offset) are concentrated along the Colorado Plateau margin (Fig. 6), northeast of the now-dormant metamorphic core complexes along the Colorado River. Regardless of whether the uplift of the southwest Colorado Plateau margin is due to changes in lithospheric thickness and/or mantle density reduction through thermal expansion, it appears that processes active in the mantle are related to uplift and neotectonic activity.

Our results are an indication that the mantle beneath the southwest margin of the Colorado Plateau may be quite different from the mantle beneath the southern Basin and Range Province and the Colorado Plateau interior. Although the presence of a strong decrease in mantle mass per unit area may indicate the magnitude of mantle processes that have occurred beneath the Colorado Plateau margin, it does not by itself yield any information about the physical mechanisms that have caused the uplift there. The maximum magnitude of reduced mantle mass beneath the southwest Colorado Plateau margin as compared with the plateau interior is $\sim 5 \times 10^{12}$ kg/km^2, which translates into 100 km of lithospheric thinning or 500 °C of asthenospheric heating (assuming a linear thermal expansion relation). Clearly the magnitude of heating necessary to cause the entire reduction in mantle density is not reasonable, as extreme amounts of magmatism would be observed if the asthenosphere were to melt to the extent that a 500 °C increase in temperature predicts. Thus the most appealing explanation for reduced mass in the mantle beneath the Colorado Plateau margin is lithospheric thinning and replacement by less dense asthenosphere, perhaps in some combination with a thermal reduction in density. For example, the mantle mass difference between the southwest margin and the interior of the Colorado Plateau can be satisfied if the lithosphere beneath the margin is about 60–80 km thinner and about 100 °C warmer than that of the plateau interior. Similarly, the mantle mass difference between the plateau margin and the southern Basin and Range Province can be satisfied if the mantle lithosphere is about 20 km thinner (or about 100 °C warmer) than the southern Basin and Range lithosphere. These amounts of thinning are consistent with the observations from earthquake-source models (Fig. 7), although those models are broad averages. The origin of the apparent reduced-density upper mantle beneath the southwest Colorado Plateau is difficult to assess but might be linked to changes in tectonic style that occurred along the western North American margin during late Tertiary time.

The uplift of the southwest Colorado Plateau margin apparently occurred during the past 6 m.y. (e.g., Lucchitta, 1989) and may have resulted from the subduction that took place along the western margin of North America throughout the Mesozoic Era. During Laramide time, the rate of subduction increased (e.g., Engebretson et al., 1985), resulting in a flatter slab that extended several hundred kilometers east of the continental margin. Subduction-related magmatism shifted to the east during that time, creating a magmatic gap where the flat slab was inferred to underlie the continent (e.g., Coney and Reynolds, 1977; Dickinson and Snyder, 1978). As a possible result of low-angle subduction, basal shear may have been imparted to the continent, producing the block uplifts and monoclinal folding observed in the Rocky Mountains and eastern Colorado Plateau region (e.g., Dickinson and Snyder, 1978; Bird, 1984). Bird (1984) proposed another important consequence of low-angle subduction. He reasoned that because the Farallon slab extended farther to the east, a portion of the North American mantle lithosphere was shaved away, thereby thinning the dense mantle keel and producing uplift. Uplift was delayed until the Farallon slab was no longer in contact with the continent, since the dense oceanic lithosphere effectively replaced the shaved-off continental lithosphere. The southwest Colorado Plateau margin was probably the part of the plateau to have been underlain by the Farallon slab most recently (Bird, 1984), as evidenced by the ca. 10 Ma closing of the magmatic gap there (Armstrong and Ward, 1991) (Fig. 8). Thus the low-density upper mantle and recent (ca. 6 Ma) phase of uplift at the southwest Colorado Plateau margin may locally represent the final delayed response of lithospheric thinning

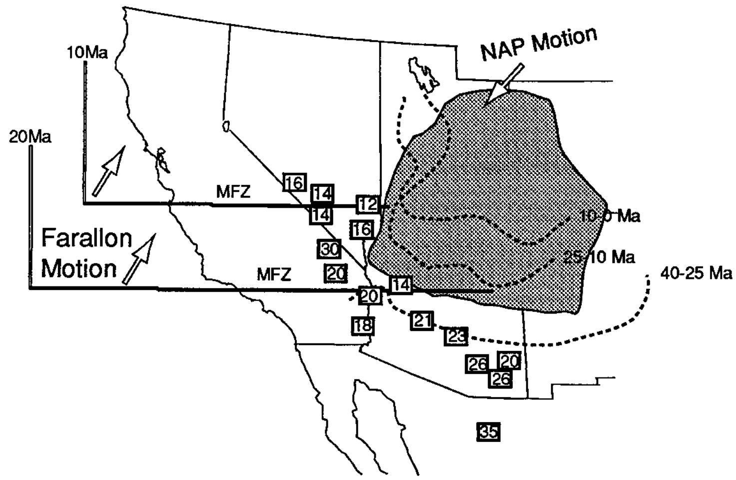

Figure 8. Locations of coherent Farallon slab (after Severinghaus and Atwater, 1991) shown with the locations and timing of onset of core complex extension and magmatism. MFZ denotes the inland extension of the Mendocino fracture zone. The dashed lines mark the local northward extent of magmatic activity with time after Armstrong and Ward (1991), and the boxed numbers are approximate time since core-complex initiation (in Ma) as summarized by Axen et al. (1993). The gray shaded area is the uplifted Colorado Plateau. The recent closing of the magmatic gap, as shown by lines of magmatic activity, may indicate when last remnants of the Farallon slab were removed from beneath Colorado Plateau lithosphere (e.g., Bird, 1984).

caused by the low-angle Laramide slab as described by Bird (1984).

CONCLUSIONS

Elevations at its southwest margin are among the highest found on the Colorado Plateau, yet models generated from seismic and gravity data indicate that the crust there lacks sufficient buoyancy and thickness (as compared with the plateau interior) to account for the elevations. Because the crust is not buoyant enough to explain the recent (<10 Ma) Colorado Plateau margin uplift, a low-density zone must exist in the mantle. Such a low-density mantle might be caused by physical thinning or thermal expansion of the dense mantle-lithosphere lid, or by a hot, low-density asthenosphere. By balancing mass in lithospheric columns, we find that the magnitude of the mass deficit in the mantle is roughly equivalent to about 60–80 km of mantle-lithospheric thinning relative to the interior of the Colorado Plateau and a net increase in mantle temperature of about 100 °C across a layer about 100 km thick. We find that about a 100 °C temperature increase beneath the plateau margin relative to the southern Basin and Range Province can explain the variation in mantle mass between those two provinces. Pliocene and younger tectonic activity (basaltic magmatism, seismicity, faulting, and uplift) has occurred in the crust above the apparent low-density upper mantle, indicating a possible causal link.

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