Mantle plume influence on the Neogene uplift and extension of the U.S. western Cordillera?

Tom Parsons

U.S. Geological Survey, MS 999, 345 Middlefield Rd., Menlo Park, CA 94025

George A. Thompson, Norman H. Sleep

Department of Geophysics, Stanford University, Stanford, CA, 94305

ABSTRACT

Despite its highly extended and thinned crust, much of the western Cordillera in the United States is elevated over 1 km above sea level. Thus this region cannot be thought of as thick crust floating in a uniform mantle isostatically; rather, the lithospheric mantle and/or the upper asthenosphere must vary in thickness or density across the region. Utilizing crustal thickness and density constraints we modeled the residual mass deficit that must occur in the mantle lithosphere and asthenosphere beneath the western Cordillera. A major hot spot broke out during a complex series of Cenozoic tectonic events that included lithospheric thickening, back-arc extension, and transition from a subduction to a transform plate boundary. We suggest that many of the characteristics that make the western Cordillera unique among extensional provinces can be attributed to the mantle plume that created the Yellowstone hot spot.

INTRODUCTION

Drawing on the growing understanding of mantle plumes and associated flood basalts (e.g., Richards et al., 1989; Sleep, 1990; Griffiths and Campbell, 1991; Hill et al., 1992) leads to the conclusion that the Yellowstone plume strongly influenced late Cenozoic tectonic events in the United States western Cordillera. Here we explore the probable consequences of the emergence of the hot spot about 17 Ma in northwestern Nevada, which was accompanied by voluminous basaltic volcanism on the Columbia Plateau and propagation of the northern Nevada rift hundreds of kilometers southward (e.g., Zoback and Thompson, 1978; Pierce and Morgan, 1992).

Unlike many other highly extended terranes, the northern Basin and Range province is elevated to an average 1.5 km above sea level. The western Cordillera crust is generally not thick enough to support its elevation isostatically, which implies that a deficit in mantle density has caused the uplift (e.g., Eaton, 1982; Morgan and Swanberg, 1985). The Great Basin (Nevada and western Utah) is the broadest part of the Basin and Range province, which began in an extensional back arc setting during Tertiary subduction off the west coast of North America (e.g., Eaton, 1979; Zoback et al., 1981). The total amount of crustal extension across the Basin and Range province estimated from surface geology is high (~50%-300% e.g., Hamilton and Myers, 1966; Zoback et al., 1981; Wernicke, 1992), and seismic probing of the crust has found thicknesses of 30-40 km (e.g., Prodehl, 1978; Catchings and Mooney, 1989). The mantle lithosphere is thought to be thinned beneath the Basin and Range province; a compilation of seismological models from studies using earthquake sources has yielded consistently thin lithospheric thickness measurements ranging from 50 to 70 km beneath the Basin and Range compared with 100-150 km beneath the Colorado Plateau (Iyer and Hitchcock, 1989). Interpretation of two-station earthquake-source data sampling mantle-lid thicknesses (from P waves) in the western United States has shown that the mantle lid (excluding the crust) beneath the Basin and Range province is thinner (20-40 km thick) than beneath the Colorado Plateau (35-50 km thick) and Great Plains (150-195 km thick) (Beghoul et al., 1993).

MANTLE PLUMES

While the deep origin of mantle plumes is controversial, their surface expression is dramatic and widely observed. Perhaps the two most ubiquitous manifestations of mantle plumes are voluminous volcanism and a broad regional topographic swell. The typical swell associated with plumes is ~1-2 km of uplift centered across a region ~1000-2000 km in diameter (e.g., Crough, 1979; Sleep, 1990), which is the result of isostatic compensation in the asthenosphere and lithosphere and, to a lesser extent, a dynamic pressure gradient of flow in the asthenosphere (below the limit of detection at Hawaii) (Sleep, 1990).

The interpreted anatomy of a mantle plume consists of a starting plume head generated during its initial ascent through the mantle, an active plume tail that continues to flow after the starting plume head contacts the lithosphere, and ponded plume-tail material that collects as the active plume tail continues to flow (e.g., Griffiths and Campbell, 1991). The initial contact of the starting plume head with the lithosphere heats a broad region because the plume head is much wider than its tail (e.g., Campbell and Griffiths, 1990; Duncan and Richards, 1991). The starting plume head and the active tail may drift independently in the asthenosphere, causing the active plume tail to contact the lithosphere away from the center of the starting plume-head swell (e.g., Griffiths and Campbell, 1991). When a hot-spot swell forms in continental lithosphere the uplifted crust may be raised far enough above the level of midplate compression from the ridge push force to be in a state of extension, which results in rift formation centered in the swell (Crough, 1983, Houseman and England, 1986). The magnitudes of the generated deviatoric stresses fall short of those necessary to cause complete continental breakup (Hill, 1991).

The buoyancy and swell associated with the starting plume head tend to move with the lithospheric plate (Sleep, 1990; Griffiths and Campbell, 1991). The heat associated with mantle plumes tends to be contained in the mantle near the plume for a considerable time. For example, Davies (1992) concluded that the thermal anomaly associated with the Hawaiian plume has not waned during its 43 m.y. existence because the swell does not decay monotonically with distance along the volcanic chain as would be the case in a thermal decline with age. That is, the material hotter than the normal mantle adiabat remains ponded below the lithosphere and above normal mantle because of the long time required for heat applied at the base of the lithosphere to conduct to the surface and because secondary convection within the thermal boundary created by plume material is inefficient.

Yellowstone Plume

During the past 16-17 m.y. the North American plate has moved southwest over the Yellowstone plume, leaving the Snake River Plain behind as its track (Morgan, 1972; Armstrong et al., 1975, Pierce and Morgan, 1992). The Yellowstone hot spot emerged with a burst of basaltic volcanism that formed the Columbia Plateau flood basalts to the northwest and the northern Nevada rift to the southeast (Zoback and Thompson, 1978), and contemporaneous silicic volcanism was centered on the McDermitt caldera in northwest Nevada. Magmatic activity on the eastern Snake River Plain was initially characterized by pulses of silicic volcanism (each of ~2-3 m.y. duration) progressing toward the present Yellowstone caldera.. Most of the silicic volcanic rocks decrease in age to the northeast (35 to 40 mm/yr) as a result of southwestward plate migration over the plume (Christiansen and Lipman, 1972; Armstrong et al., 1975). Extensive basaltic volcanism followed, covering the plain and persisting through Holocene time (e.g., Luedke and Smith, 1983). Seismic-refraction data indicate that much of the midcrust beneath the plain was replaced by intruded basalt (e.g., Sparlin et al., 1982). The location of the Yellowstone starting plume head and the start of the plume tail is subject to discussion because it may have encountered the subducting Juan de Fuca plate during its ascent, and part of the plume-head material could have been spread westward as far as the Juan de Fuca ridge by the descending plate (Hill et al., 1992).

WESTERN CORDILLERA MANTLE

Because of its relatively thin crust and high elevation, the western Cordillera must depend for isostatic support on mantle material of lower relative density beneath it. We compiled crustal thickness information for the western United States from seismic refraction surveys (Warren, 1969; Prodehl, 1978; Catchings and Mooney, 1989; Kohler and McCarthy, 1990; McCarthy et al., 1991) (Fig. 1) and corrected for upper-crustal density variation, by using a basin-stripped isostatic residual gravity map (Saltus, 1991). These gravity data give an indication of the relative density variation of the upper-crustal basement rocks because the effects of sedimentary basins and regional trends have been removed. Density perturbations of the upper-crustal rocks can be approximated with the relation □| |$(g/0.04L_{uc})$ (Simpson and Jachens, 1989), where g is gravity in mgal, and L_{uc} is an assumed upper-crustal thickness (15 km). Lower-crustal densities are not delimited by the data and were assigned a uniform value of 2.9 g/cm^3; the lower-crustal layer is too thin for even strong density variations within it to cause the large observed mass deficit. Combining the crustal-thickness and upper-crustal density data results in the approximate crustal contribution to uplift across the western Cordillera; this when compared with the topography, indicates the mantle contribution to uplift (Fig. 1). We assume regional isostasy because the free air gravity anomaly averages zero across the region (e.g., Thompson and Zoback, 1979).

We find the negative mass anomaly per unit area by dividing the study area into a grid of isostatic columns. These columns are calculated using a two-layer isostatic expression with a known crustal layer, and a second mantle layer that extends down to an assumed local iso-density 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 lithosphere (of unknown density) and the asthenosphere. Since we treat the crust as a known quantity and we know the topographic elevation, then the mass contained within each mantle column that is required to float it to its individual height can be calculated. The negative mantle mass anomaly per unit area of a given column (M_n) is found by comparing it to a reference column, which is the column in the study area that contains the greatest amount of mass (and hence has the lowest mantle buoyancy), as

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

where

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

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, ρ is density with the a and c subscripts denoting asthenosphere and crust respectively, E_n is topographic elevation, and H is the buoyant height of the hypothetical free surface of the asthenosphere (~2.4 km; Lachenbruch and Morgan, 1990).

On the basis of isostatic calculations, Figure 1 shows that a broad low-density anomaly underlies the northern Basin and Range province, extending from the Sierra Nevada in the west to the Colorado Plateau in the east. The peak of the anomaly corresponds approximately to the inferred location of the initial breakout of the Yellowstone plume at McDermitt Caldera (Zoback and Thompson, 1978), with the bulk of the anomaly to the south beneath the central Basin and Range province. The actual plume track (eastern Snake River Plain) does not appear as a low-density anomaly on this image because the crust there has been largely replaced by more dense basalt (e.g., Sparlin et al., 1982), which has caused subsidence.

We calculate that the sum of the mantle isostatic mass deficit is ~$3.6x10^{18}$ kg beneath the region investigated. Extension and thinning of the Basin and Range lithosphere allowed the asthenosphere to rise and take its place, causing some of the mass deficit. Earthquake-source methods that measure the variation in lithospheric thickness between the Basin and Range province and Colorado Plateau converge on a difference in the mantle-lid thickness that ranges between 15 and 50 km across the two provinces (Iyer and Hitchcock, 1989; Beghoul et al., 1993). If a reasonable density contrast between asthenosphere and lithosphere of 0.05 g/cm^3 is used (e.g., Thompson and Zoback, 1979; Lachenbruch and Morgan, 1990) then the mass deficit caused by lithospheric thinning amounts to ~$4.8x10^{17}$-$1.6x10^{18}$ kg, or 13%-44% of the total anomaly. If the entire anomaly is attributed to lithospheric thinning over an adiabatic asthenosphere, then a mantle lithospheric layer ~112 km thick would need to have been removed from the Basin and Range province across the study area to account for the entire observed mass deficit.

The calculated buoyancy flux of the Yellowstone plume is ~$1.5x10^3$ kg s^{-1} (Sleep, 1990). Thus the active Yellowstone plume tail could have generated ~20% (~$7.6x10^{17}$ kg) of the total mass deficit as ponded plume-tail material during the time (16-17 m.y.) it took to track from northern Nevada to its current position at Yellowstone. The time required for the Yellowstone plume tail to supply the mantle mass deficit not caused by lithospheric thinning (~$2.0x10^{18}$-$3.1x10^{18}$kg) ranges from 42 to 89 m.y., which is much longer than the probable age of the Yellowstone plume. Because the Yellowstone plume is probably about 16-17 m.y. old, we suggest that the plume had a starting head, the remnant of which causes a ~$1.2x10^{18}$-$2.4x10^{18}$ kg mass deficit beneath the investigated part of the western Cordillera. If the plume head was roughly symmetric, then these numbers extrapolate to a total mass deficit of about $1.7x10^{18}$-$3.4x10^{18}$ kg. Such a mass deficit is comparable to the expected deficit (~3-$8.5x10^{18}$ kg) from a small plume head (~400 km radius, Hill et al., 1992) that is 100-300 ^oC hotter than the surrounding asthenosphere.

DISCUSSION AND PROPOSED CENOZOIC TECTONIC HISTORY

The western Cordillera is elevated because of diverse processes that date back at least to the Laramide orogeny. The Cordillera is elevated more than 1 km above sea level from the Sierra Nevada and Cascades in the west, across the Great Basin and into the Rockies and Great Plains to the east. Compressional shearing and crustal thickening related to the horizontal subduction of the Farallon Plate most likely contributed to the thick crust and high elevation of the Rocky Mountains and eastern Colorado Plateau (e.g., Dickinson and Snyder, 1978; Bird, 1984), and the region may have been at high elevation at least since Eocene time (Gregory and Chase, 1992). The cessation of subduction and subsequent removal of the slab may have allowed asthenospheric contact with the western Cordillera lithosphere, causing part of the regional uplift that extends into the Great Plains. The period from 40 to 20 Ma represented a shift from low-angle subduction and back-arc compression to normal subduction and back-arc extension (e.g., Eaton, 1979; Zoback et al., 1981) and/or extensional collapse (e.g., Harry et al.,

1993) in the Basin and Range province (Fig. 2). The localization of the back-arc and back-transform extensional setting in the Basin and Range province probably caused the lithosphere to thin relative to the more stable Colorado Plateau and Rocky Mountains.

About 16-17 Ma the Yellowstone plume emerged, and accompanying basaltic volcanism began to dominate in the northern Basin and Range. Interpretations of Neogene cordilleran basalt compositions vary as to whether or not they are consistent with a mantle plume source; some workers have suggested that they are (e.g., Leat et al., 1988; Thompson et al., 1989; Fitton et al., 1991; Menzies et al., 1991), whereas others have found that isotopic signatures in places are more closely related to the underlying lithosphere (e.g., Lum et al., 1989; Lipman and Glazner, 1991; Bradshaw et al., 1993). The degree of lithospheric contamination of deep-sourced magmas remains controversial. A mantle plume acts more as a source of voluminous hot material than as a point source of heat, and it is likely that conducted heat applied broadly to the lithosphere by a plume head would cause some melting of the existing mantle lithosphere as well as supply more primitive melts directly from the asthenosphere.

If the Yellowstone plume behaved like most observed plumes, then it would have created a broad swell, centered approximately in northern Nevada. The present-day topography seems to bear the imprint of such a swell (Fig. 3), with the 1 km topographic contour marking a roughly circular region centered in northwest Nevada, bounded by the slope of the Sierra Nevada to the west, the Columbia River basin to the north, the northern-to-southern Basin and Range transition to the south, and verging into the thickened Colorado Plateau to the east. The diameter of the observed swell in the western Cordillera is smaller (~800 km) than the typical hot-spot swell (~1000-2000 km, e.g., Sleep, 1990; Hill et al., 1992), which may be the result of a smaller Yellowstone starting plume head caused by interaction with the subducting Juan de Fuca plate. The shape of the swell is likely to be asymmetric because previously thinned lithosphere would create more room for plume material to pond beneath it (Fig. 2). Thus, the strongest topographic inversion would be observed in regions that were thinned prior to plume-head contact. The newly elevated Basin and Range crust was put into a strong state of extension (e.g., Crough, 1983) with the northern Nevada rift forming at its center, and a new surge of broad extension began in the western part of the province during the middle Miocene. The stalk of the Yellowstone plume remained stationary as the North American plate tracked over it, introducing new basaltic volcanism to the eastern Basin and Range, as well as limited uplift resulting from plume-tail flow. The westward progression of the plate relative to the plume may have caused the broadening of the northern Basin and Range relative to the south and the most recent extension in the eastern part of the province (Pierce and Morgan, 1992). Further analysis of plume behavior in relation to the tectonic evolution of the Basin and Range province and its high-standing margins, the Sierra Nevada and Colorado Plateau, holds great promise.

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Figure 1. Color contour plot of negative mass anomaly in mantle beneath investigated part of the western Cordillera. Anomaly was calculated by using modified form of isostatic equation with crustal depth information from seismic-refraction profiles (shown as gray lines). Free-air gravity anomaly across region is near zero, which implies that area is in isostatic equilibrium. Thin crust of northern Basin and Range province associated with its high elevation requires underlying low-density upper mantle if isostasy is to be satisfied. We suggest that Yellowstone plume has supplied hot, low-density material to base of lithosphere causing broad mantle anomaly pictured.

Figure 2. Schematic Tertiary tectonic history of the western Cordillera. A: Horizontal subduction of Farallon plate thickens Cordillera lithosphere. B: Transition to normal-angled subduction creates back-arc extension in Basin and Range and thins lithosphere there. C: Yellowstone plume contacts lithosphere, ponding beneath thinnest regions. Northern Basin and Range province is centered above plume and thus is peak of topographic swell. Excess elevation triggers a new phase of large magnitude extension in Basin and Range which thins lithosphere relative to surrounding provinces. D: Subduction terminates, and extension continues in Basin and Range province causing subsidence relative to surrounding regions.

Figure 3. Current topography shown with path of Yellowstone plume. Dashed lines indicate suggested topographic expressions of starting plume head and active plume tail.

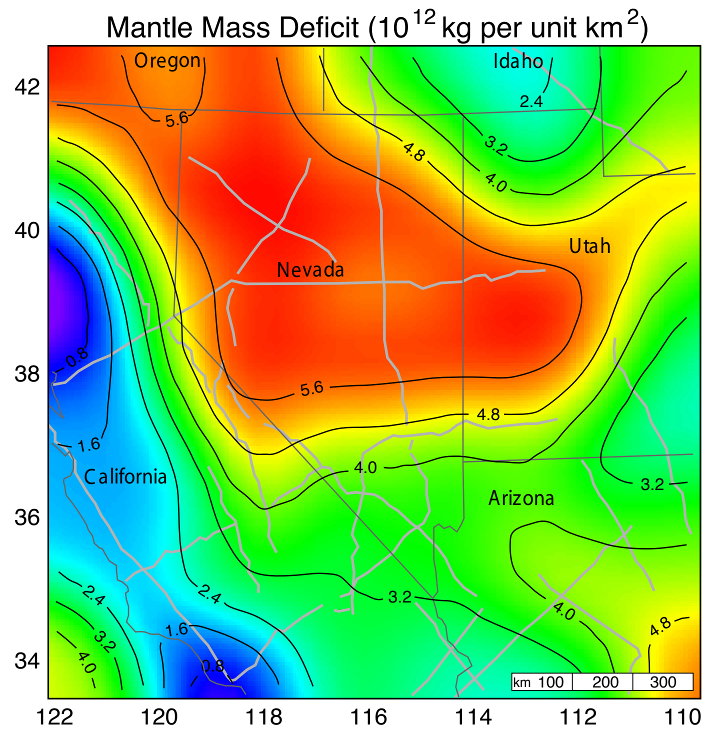
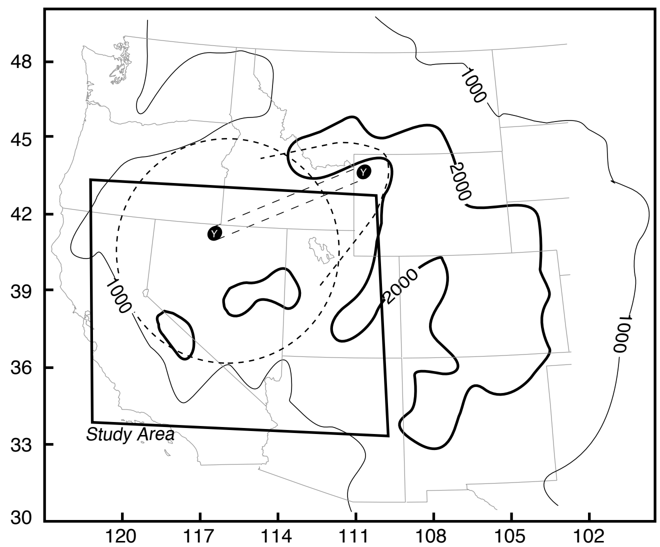

Mantle Mass Deficit (10^{12} kg per unit km^2)
Oregon
Idaho
Nevada
Utah
California
Arizona
2.4
3.2
4.0
4.8
5.6
0.8
1.6
42
40
38
36
34
122
120
118
116
114
112
110
km 100
200
300

Figure 1



48
45
42
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Study Area

Figure 2

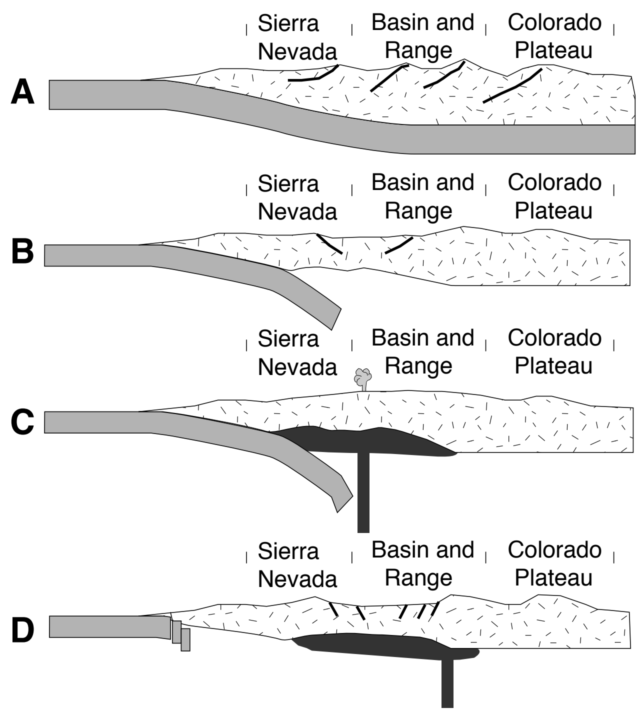
Sierra Nevada
Basin and Range
Colorado Plateau
A
Sierra Nevada
Basin and Range
Colorado Plateau
B
Sierra Nevada
Basin and Range
Colorado Plateau
C
Sierra Nevada
Basin and Range
Colorado Plateau
D

Figure 3