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Physical state of the western U.S. upper mantle

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Using observed P wave images of the western U.S. upper mantle, which show lateral variations of up to 8%, and existing scaling relations, we infer that the low-velocity mantle is hot and partially molten to depths of 100-200 km, and that the high-velocity upper mantle is subsolidus. Most the high-velocity upper mantle within a few hundred kilometers of the coastline appears to be relatively dense, suggesting that it is relatively cool (i.e., a thermal lithosphere). This is expected for features associated with the subducting Juan de Fuca and Gorda slabs, and the high velocity upper mantle beneath the Transverse Ranges has been attributed to the sinking of negatively buoyant mantle lithosphere. Other high-velocity mantle structures near the continental margin are consistent with this interpretation. In contrast, the generally high elevations of the continental interior imply a buoyant upper mantle there, an inference that holds for both the high- and the low-velocity upper mantle. The only reasonable way to produce the high-velocity low-density upper mantle is through basalt depletion, thereby creating mantle of increased solidus temperature and decreased density. We distinguish a marginal domain, within ~250 km of the Pacific coast, from an interior domain. This is based on the inferred upper mantle compositional difference and regional associations: beneath the marginal domain, upper mantle structures trend parallel to the surface physiography and young tectonic structures, whereas upper mantle structures beneath the continental interior trend northeasterly. This northeast orientation is discordant with the young tectonic structures, but aligns with young volcanic activity. The high lateral gradients in observed upper mantle seismic structure found throughout the western United States imply high lateral gradients in the associated temperature or partial melt fields. Because these fields diffuse on time scales of less than a few tens of millions of years, the imaged upper mantle structure is young. The following upper mantle processes are hypothesized to account for these findings and inferences. Away from the plate margin, small-scale upper mantle convection driven by partial melt-induced buoyancy of hot upper mantle leads to the production and segregation of melt and the creation of compositional variations. The heterogeneous upper mantle P wave structure of the elevated continental interior is largely a consequence of partial melt variations that are modulated by the compositional variations, and throughout this region we infer high temperatures and low densities. Near the plate margin, relative plate motions force upper mantle flow, although upper mantle flow driven by the positive buoyancy of melt and the negative buoyancy lithosphere is important locally.

Introduction

The western third of the United States currently is experiencing both marginal and intracontinental tectonic and volcanic activity. Most of the geologic activity near the western U.S. margin can be understood in the context of a continental margin undergoing transition from a subduction to a transform setting. In contrast, deformation and volcanism in the continental interior, especially in the Great Basin and the Rocky Mountains (see Figure 1 for geographic locations), do not appear to be simple consequences of plate interaction. The great distance to which volcanic and tectonic activity occurs within the continent suggests that upper mantle processes not directly associated with plate interaction are primarily responsible for this activity.

The upper mantle of the western United States is slow on average, delaying P wave arrivals by ~0.6 s compared to global average [*Romanowicz*, 1979; *Dziewonski and Anderson*, 1983; *Humphreys and Dueker*, 1994]. This slow upper mantle is part of an extensive low-velocity volume associated with the East Pacific Rise, and beneath the western United States low velocities extend to 200-400 km depth [*Helmberger et al.*, 1985; *Montagner and Tanimoto*, 1991; *Su et al.*, 1992]. The base of this low-velocity zone beneath the western United States is defined by a high positive velocity gradient at a depths of between 150 and 250 km [*Hales*, 1991; *Iyer and Hitchcock*, 1989]. Teleseismic P waves, in addition to being late on average, reveal great lateral variations in seismic velocity within this region, especially above ~200 km (Figure 2, from *Humphreys and Dueker* [1994]). The patterns of the imaged upper mantle structure, along with elevation and regional tectonic style, are used to define two domains in the western United States: a marginal domain within ~250 km of the Pacific coast, where upper mantle structures parallel major tectonic belts and average elevations are not great; and a broad interior domain, where average elevations are greater than 1 km and uppermost mantle structures trend northeasterly, which generally is discordant with the young tectonic structures and concordant with young volcanic activity. A zone of transition exists between these two domains that includes the Walker Lane Belt and the high Sierra Nevada.

Our primary goal in this paper is the estimation of western U.S. upper mantle temperature, partial melt content and composition fields, by considering the relatively detailed P wave velocity (V_P) image of the upper mantle [*Dueker et al.*, 1993; *Humphreys and Dueker*, 1994] in conjunction with regional geophysical studies and the young geologic record. Our reasoning is summarized here. Physical causes for the great magnitude in imaged lateral variations of upper mantle P wave structure are limited to changes in phase (of which partial melt is considered the most important), temperature, and composition. Although our ability to infer these properties is limited, the depth and degree to which western U.S. upper mantle is seismically slow implies that this mantle is unusually hot, and supersolidus temperatures are suggested for the slowest upper mantle. In addi-

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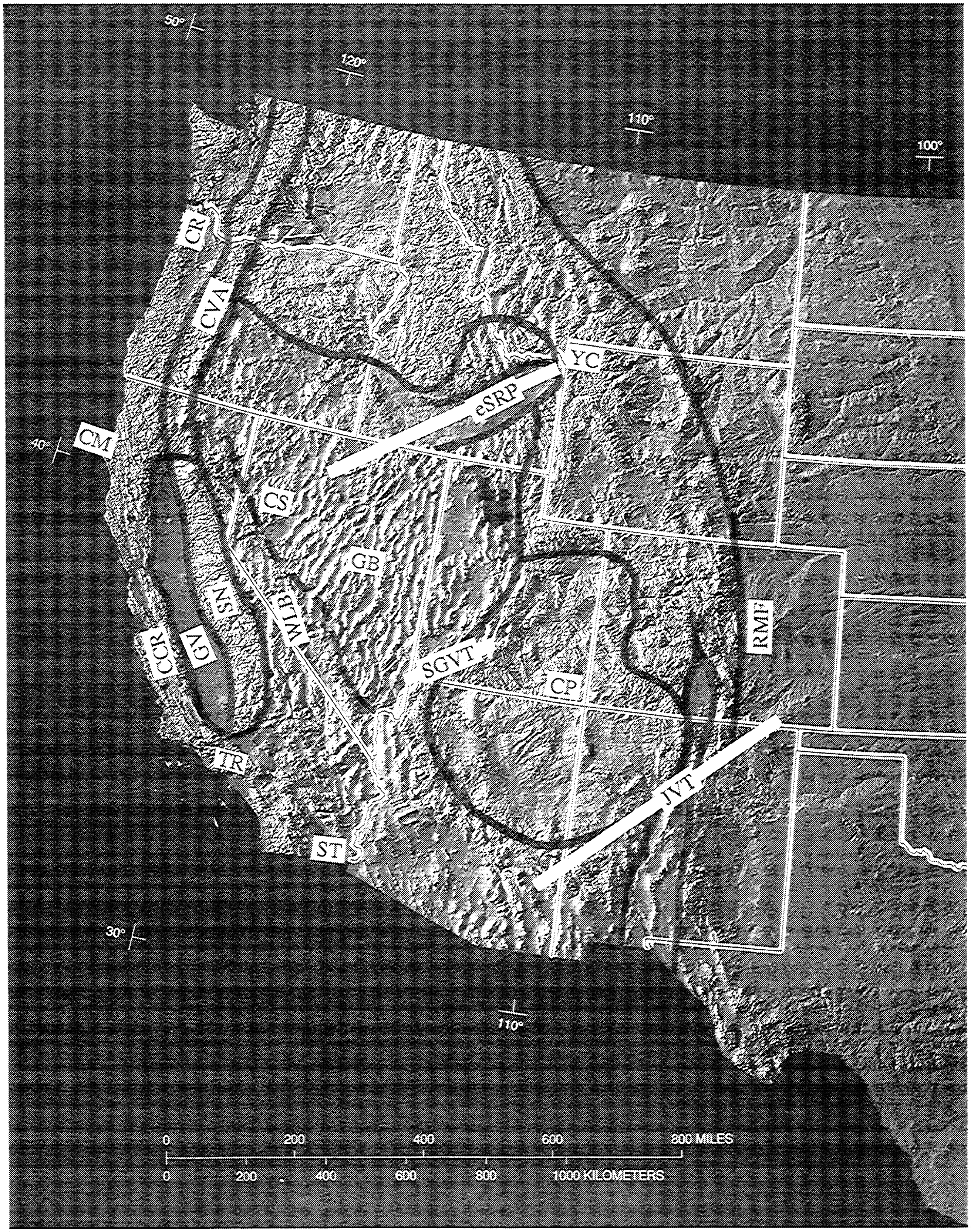

Fig. 1. Western U.S. topography, tectonic, and geographical features. Provinces (bold lines): Cascade Volcanic Arc (CVA), California Coast Ranges (CCR), Coast Ranges (CR), Colorado Plateau (CP), Great Basin (GB), Great Valley (GV), Rocky Mountain Front (RMF), Salton Trough (ST), Sierra Nevada (SN), Transverse Ranges (TR) and Walker Lane Belt (WLB). Eastern margin of Walker Lane Belt is shown with dashed line. Geographic locations: Cape Mendocino (CM), Carson Sink (CS) and Sevier depression (SD). Volcanic areas: eastern Snake River Plain (eSRP), Jemez Volcanic Trend (JVT), St. George Volcanic Trend (SGVT) and Yellowstone Caldera (YC).

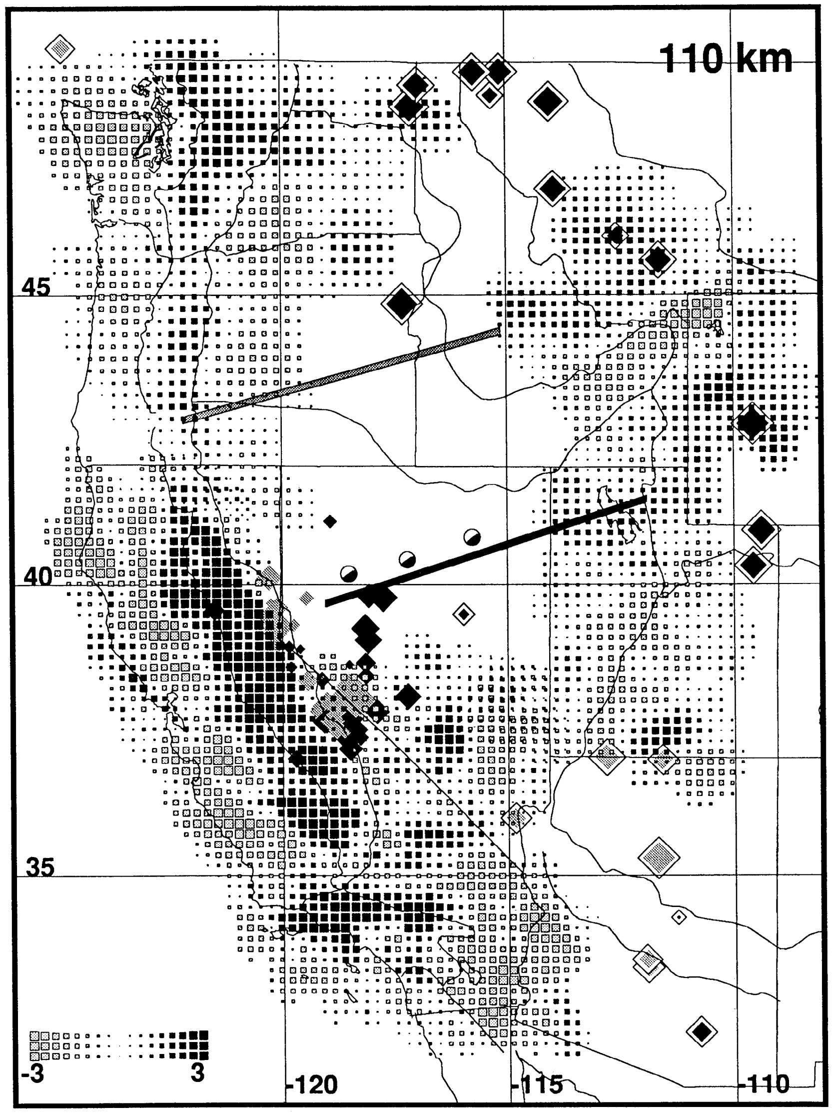

Fig. 2. *P* wave velocity structure at a depth of 110 km [*Humphreys and Dueker*, 1994; *Dueker and Humphreys*, 1990] with other indicators of upper mantle *P* wave velocity structure. Black represents fast upper mantle, and gray represents slow. Symbol size is proportional to velocity perturbation, with velocities of absolute value greater than 3% clipped at this level. In addition, the following information is included to qualitatively extend the regional patterns of the seismic structure: (1) diamond symbols with a white border denote average International Seismological Centre residuals [*Dziewonski and Anderson*, 1983], (2) unbordered diamond symbols denote average station residuals from the northern Nevada array, (3) half-solid circles in northern Nevada denote the ~1 s early arrivals from southeast azimuths [*Koizumi et al.*, 1973], and (4) black and gray lines in the northern Great Basin represent fast and slow surface wave paths, respectively [*Priestley and Brune*, 1982].

tion, uplift of the western United States requires a massive influx of buoyancy since the times that shallow seas covered much of the area during the Cretaceous. Thermal (i.e., warming) and compositional (i.e., basalt depletion through melt removal) buoyancy are the two reasonable means of decreasing the upper mantle density. Based on the similar spatial alignments between tectonic structures and upper mantle structure in the marginal domain (Figures 1 and 2), the high-velocity upper mantle there appears to be relatively dense and therefore attributed to the relatively low temperatures of the lithosphere. In contrast, the generally high elevations of the interior domain imply buoyant upper mantle there, an inference that holds both for the high- and the low-velocity upper mantle, leading us to conclude that the relatively great *P* wave velocities of the interior domain upper mantle result from basalt depletion.

The temperatures we infer for the western U.S. upper mantle are greater than those commonly associated with upper mantle. Because plumes have a temperature of ~200°C in excess of average upper mantle [e.g., *Sleep*, 1990], they are a source for the hot upper mantle. But upper mantle temperature variations of ±200°C also can be caused by convective processes unrelated to deep mantle plumes [*Tackley et al.*, 1993; *Anderson*, 1994]. Both origins for hot upper mantle have been suggested as the cause of western U.S. volcanic and tectonic activity (e.g., a set of plumes [*Suppe et at.*, 1975]; hot upper mantle associated with the East Pacific Rise [*Cook*, 1969; *McKee*, 1971]). In either case, upper mantle temperatures temperatures ~200°C in excess of those typically inferred for the upper mantle (i.e., 200°C hotter than an assumed average mantle potential temperature of ~1300°C [*McKenzie and Bickle*, 1988]) could cause partial melting to depths of a few hundred kilometers [*Tackley and Stevenson*, 1993], resulting in depressed seismic velocities, increased melt production, and decreased upper mantle densities due to thermal expansion and basalt depletion. A layer of upper mantle exceeding solidus temperatures would be expected to be dynamically unstable to convective overturn, creating large volumes of melt and compositionally depleted upper mantle with horizontal wavelengths of 2-3 times the thickness of the convectively unstable layer [*Tackley and Stevenson*, 1993].

We interpret much of the upper mantle structure in the context of the above discussion: Away from the plate margin, small-scale upper mantle convection driven by partial melt-induced buoyancy leads to upper mantle flow, melt segregation and the creation of important upper mantle compositional variations. Depleted volumes, such as are inferred beneath the central Great Basin, are relatively high in V_P compared to volumes where significant melt fractions are present. Near the plate margin, heterogeneous upper mantle V_P structure is a result of cold and dense lithosphere (much of it subducted oceanic plate) placed within the asthenosphere. In this region, plate motions (subduction and transform) control most of the upper mantle flow. However, the sinking of dense lithosphere is also responsible for some tectonic activity near the transform margin.

SCALING RELATIONS

The relations used below to scale V_P and density to upper mantle temperature, composition, and partial melt content are discussed in this section and summarized in Table 1. A more extensive treatment of density scaling with respect to mineralogy is given in *Cordery and Phipps Morgan* [1993].

Scaling of V_P

Temperature. The relation $\partial V_P/\partial T \approx -0.5$ m s^{-1} K^{-1} is commonly used to scale temperature (T) to V_P [*Anderson and Bass*, 1984; *Creager and Jordan*, 1986; *Duffy and Anderson*, 1989], yielding ~160°C temperature decrease per percent V_P increase. *Sato et al.* [1989] present experimental evidence for $\partial V_P/\partial T \approx -2.0$ m s^{-1} K^{-1} over the range 100-200°C just below the solidus. *Karato and Spetzler* [1990] question the validity of using Sato et al.'s results at seismic frequencies. However, current experimental data do not exclude an activation of a near-solidus weakening mechanism at seismic frequencies, and such a V_P reduction mechanism may be possible.

Composition. Variations in upper mantle composition are primarily due to varying degrees of basaltic melt removal. The influence of melt extraction on V_P is relatively minor, with 10% basalt extraction causing ~1% increase in V_P [*Jordan*, 1978; 1981].

Partial melt. From experimental and theoretical analysis of the seismic velocity variations induced by the presence of partial melt, a linear scaling relation $(V_P - V_0)/V_0 = -A \times \phi$ relating partial melt fraction ϕ and velocity perturbation $V_P - V_0$ from initial

TABLE 1. Scaling Relations

Relation			Units	Comment
Scaling of V_P				
$\frac{\partial V_P}{\partial T}$	$\approx$	-0.5	m s^{-1} K^{-1}	
		(−2.0)		using *Sato et al.* [1989]
V_0	$\approx$	8.0×10^3	m s^{-1}	
$\frac{1}{V_0}\frac{\partial V_P}{\partial T}$	$\approx$	-6.25×10^{-5}	K^{-1}	
		$(-25. \times 10^{-5})$		using *Sato et al.* [1989]
$\frac{1}{V_0}\frac{\partial V_P}{\partial \phi}$	$\approx$	-1.25	—	
$\frac{1}{V_0}\frac{\partial V_P}{\partial R}$	$\approx$	0.1	—	
Scaling of ρ				
ρ_0	$\approx$	3.3×10^3	kg m^{-3}	
$\frac{1}{\rho_0}\frac{\partial \rho}{\partial T}$	$\approx$	-2.5×10^{-5}	K^{-1}	
$\frac{1}{\rho_0}\frac{\partial \rho}{\partial \phi}$	$\approx$	-0.10	—	at 75 km depth; basalt segregated
		(-0.20)	—	at 75 km depth; basalt not segregated
$\frac{1}{\rho_0}\frac{\partial \rho}{\partial R}$	$\approx$	-0.09	—	depth > 70 km
$\rho - V_P$ Scaling				
$\frac{(1/\rho_0)(\partial \rho/\partial T)}{(1/V_0)(\partial V_P/\partial T)}$	$\approx$	0.4	—	
		(0.1)		using *Sato et al.* [1989]
$\frac{(1/\rho_0)(\partial \rho/\partial \phi)}{(1/V_0)(\partial V_P/\partial \phi)}$	$\approx$	0.08	—	basalt segregated
$\frac{(1/\rho_0)(\partial \rho/\partial R)}{(1/V_0)(\partial V_P/\partial R)}$	$\approx$	-0.09	—	depth > 70 km

ϕ is in situ melt fraction.

R is volume fraction of upper mantle removed through basalt segregation.

velocity V_0 holds well for small melt fractions, where typically $1<A<3$ [*Stocker and Gordon*, 1975; *O'Connell and Budiansky*, 1977; *Mavko*, 1980; *Schmeling*, 1985; *Sato et al.*, 1989]. The value of scaling factor A depends on melt geometry, with aspect ratios much smaller than unity yielding larger values of A. Using the theoretical results of *Schmeling* [1985], $A=3$ corresponds to aspect ratios of $0.1-0.01$. Values of $A>3$ are possible if the melt aspect ratio is very small, such as if melt films are common. Filmlike melt geometries have been observed in lab experiments of partial melting [*Waff and Faul*, 1992; *Faul et al.*, 1994].

Scaling of Density

Temperature. Temperature is related to density through the thermal expansion coefficient α, which for upper mantle material is $\alpha \approx -2.5\times10^{-5}$ K^{-1} [*Stacey*, 1977; *Duffy and Anderson*, 1989]. This yields 0.25% decrease in density for a 100°C increase in temperature.

Composition. The creation of compositionally depleted upper mantle through basaltic melt extraction causes a density decrease. Mantle olivine is of magnesium number $Mg\# = X_{Mg}/(X_{Fe}+X_{Mg}) \approx 90$ (i.e., Fo_{90}). Upper mantle containing olivine of this general composition produces melt of $Mg\# \approx 75$ [*Roeder and Emslie*, 1970]. Thus, the creation of 1% melt would increase upper mantle olivine $Mg\#$ by ~ 0.15. Using $\partial\rho/\partial Mg\# = -12$ kg m^{-3} $Mg\#^{-1}$ and reference upper mantle density $\rho_0 = 3300$ kg m^{-3} [*Sheehan and Soloman*, 1991; *Akimoto*, 1972], calculated residuum density changes by $(1/\rho_0)(\partial\rho/\partial R) \approx -0.05$ for R being the volume fraction of rock removed by basalt segregation. At depths greater than about 70 km garnet is present, and the consumption of garnet during melting also decreases the bulk density of the upper mantle. This has an effect on upper mantle density that is nearly as great as changing the $Mg\#$ of olivine, for a given degree of basalt removal [*Jordan*, 1978, 1979, 1981]. Thus $(1/\rho_0)(\partial\rho/\partial R) \approx -0.09$ is predicted below depths of about 70 km (i.e., residuum density would decrease by ~1% due to the extraction of 10% melt).

Partial melt. At 40 km depth, basaltic melt is ~ 400 kg m^{-3} less dense than unmolten upper mantle [*Niu and Batiza*, 1991]. Thus the presence of 1% partial melt decreases bulk density by $\sim 0.12\%$. Because melt is more compressible than the crystal matrix [*Stolper et al.*, 1981; *Agee and Walker*, 1988], this difference diminishes with depth. Extrapolating with depth, this difference goes to zero at a depth of a 200-300 km [*Agee and Walker*, 1988].

If melt were created but not segregated from the matrix, buoyancy contributions arise from both the depleted residuum and the presence of melt. At ~75 km depth each of these contributions is approximately equal, yielding $(1/\rho_0)(\partial\rho/\partial\phi) \approx -0.2$ (for the case of melting without melt segregation).

RELATED GEOPHYSICAL STUDIES

Many of the relevant aspects of the seismic structure of the western United States are shown in Figure 2, which shows the general form and relative amplitude of the upper mantle structure (see *Dueker et al.* [1993] and *Humphreys and Dueker* [1994] for a more complete presentation). Further constraint on the physical state of the upper mantle is provided by other geophysical studies, which are summarized in this section.

Additional Seismic Studies

As discussed in the introduction, the upper mantle of the western United States is a portion of an extensive low-velocity feature associated with the East Pacific Rise [*Montagner and Tanimoto*, 1991]. This low-velocity region extends east of the region imaged in Figure 2 to the near the Kansas-Colorado border [*Lee and Grand*, 1993]. The North America craton lies east of the elevated western United States, and upper mantle velocities there are high to depths of several hundred kilometers [*Grand and Helmberger*, 1984; *Grand*, 1987].

Within the portion of the elevated western United States represented by Figure 2, upper mantle structure is dominated by two northeast-oriented low-velocity volumes, one beneath the Yellowstone volcanic trend and another beneath the St. George volcanic trend. In addition to these aligned low-velocity structures, a similar northeast-oriented low-velocity structure exists beneath with the Jemez volcanic trend in New Mexico [*Parker et al.*, 1984; *Spence and Gross*, 1990; *Davis et al.*, 1993]. Like the St. George and Yellowstone volcanism, Jemez volcanism is young (predominantly younger than 5 Ma along its entire length [*Smith and Luedke*, 1984]). These three parallel upper mantle and young volcanic alignments have a spacing of ~500 km.

Heat Flow

Western U.S. heat flow (Figure 3) is high on average, although considerable regional variability occurs [*Morgan and Gosnold*, 1989]. A strong similarity is apparent between upper mantle V_P and heat flow, with regions of low-velocity upper mantle usually exhibiting high heat flow. Volcanism younger than 16 Ma [*Smith and Luedke*, 1984] occurs primarily in these regions as well, suggesting that recent crustal intrusions of magma have contributed significantly to the heat flow in these areas. The frequent association of young volcanism and high heat flow with low-velocity upper mantle suggests that these zones of low velocity are partially molten; similarly, the occurrence of high-velocity upper mantle beneath areas of relatively low heat flow and very little young magmatism suggests that high-velocity upper mantle is largely devoid of partial melt.

In the Great Basin, low heat flow is observed in central Nevada (i.e., the Eureka low [*Lachenbruch and Sass*, 1977]). This is a region where relatively high upper mantle velocities occur and post-Miocene volcanism is rare [*Smith and Luedke*, 1984]. High heat flow occurs in northern Nevada (i.e., the Battle Mountain high [*Lachenbruch and Sass*, 1977]), in a region where the southwest extension of the eastern Snake River Plain volcanic trend is underlain by low upper mantle velocities, and post-Miocene volcanism is abundant. The northeast-trending St. George volcanic trend [*Smith and Luedke*, 1984], which crosses the Colorado Plateau-Great Basin transition, is associated with a region of high heat flow (Figure 3) and lies above the upper mantle low-velocity trend that extends from southern Nevada to central Utah. Near the western margin of the Colorado Plateau, heat flow is high (~90 mW m^{-2}) [*Bodell and Chapman*, 1982] and the lower crustal and upper mantle is anomalously electrically conductive [*Porath et al.*, 1970; *Porath*, 1971]. Relatively high heat flow values (~70 mW m^{-2} [*Saltus and Lachenbruch*, 1991]) are observed in the southeastern Sierra Nevada, where young volcanism occurs and where relatively low-velocity upper mantle is imaged. Very low heat flow is observed for the eastern Great Valley and western-to-central Sierra Nevada [*Lachenbruch and Sass*, 1977; *Saltus and Lachenbruch*, 1991], where high upper mantle velocities occur. In the California Coast Ranges south of Cape Mendocino, heat flow is moderate to high [*Lachenbruch and Sass*, 1980], upper mantle velocities are low and young volcanic activity is present. In southern California, high heat flow occurs in the Salton Trough region and low heat flow in the Transverse Ranges region [*Lachenbruch et al.*, 1985], again corresponding well with imaged low and high upper mantle velocities, respectively.

Heat flow in the marginal portion of the Pacific Northwest [*Blackwell et al.*, 1990] is high in the Cascades, beneath which the high-velocity, steeply dipping slab is imaged, and low in the Coast Ranges, where generally slow upper mantle is imaged (presumably beneath a shallowly dipping slab there [*Humphreys and Dueker*, 1994]). Although this relationship between heat flow and upper mantle velocity is opposite from that found elsewhere in the western United States, this heat flow pattern is

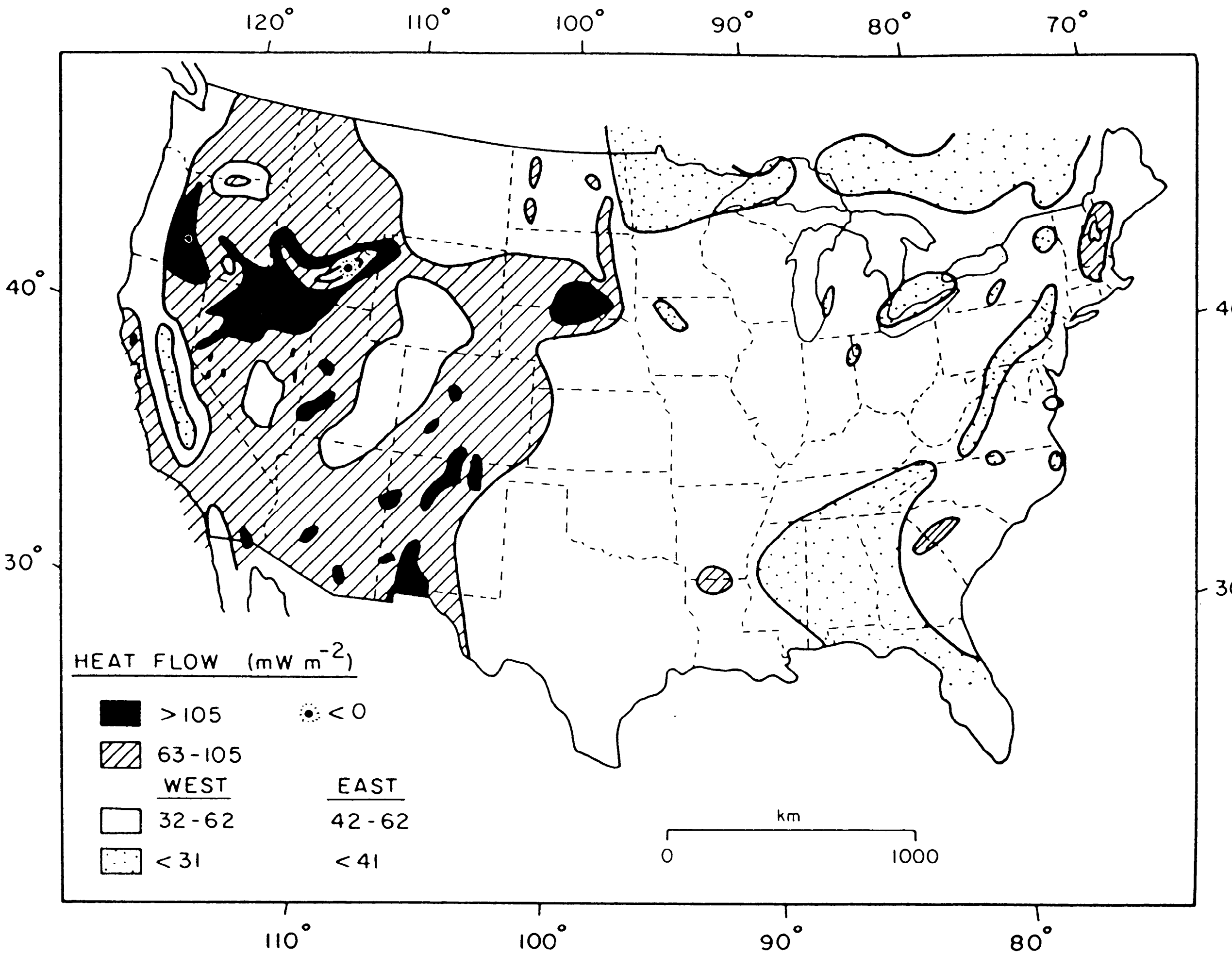

Fig. 3. Heat flow [from *Morgan and Gosnold*, 1989].

expected as a natural consequence of ongoing subduction beneath the Pacific Northwest [*Blackwell et al.*, 1990].

Isostasy

In this section we estimate the density structure of the western U.S. upper mantle, concluding that much of this mantle is anomalously buoyant, including the high-velocity upper mantle beneath large portions of the Great Basin. This conclusion is supported by the uplift history; much of the uplift involves crust that has not been tectonically thickened since the time when it stood near sea level in the Cretaceous.

Most of the western U.S. interior lies at elevations above 1 km, and elevations in excess of 2 km are common, making this region one of the great plateaus on Earth. Figure 4 [from *Simpson et al.*, 1986] presents elevation smoothed so that the short-wavelength structures associated with faulting and flexural support are suppressed. This figure also represents the crustal thickness predicted by Airy isostasy with a simple crustal model. Figure 5 [after *Braile*, 1989; *Mooney and Weaver*, 1989] presents "observed" crustal thickness. Crustal thickness is not constrained by seismic data in many regions, and even in regions where seismic data are available, its thickness and density structure is sometimes controversial. Where seismic data do not constrain crustal thickness, thickness estimates in this figure appear to be influenced by the ideal of maintaining Airy isostasy. Even so, the elevations of the western U.S. interior are much greater than can be maintained by its crust. Figure 6 shows the difference between the observed and predicted crustal thicknesses. Alternatively, this can be viewed as a map of long-wavelength elevation anomaly, with respect to observed crustal thickness. Most of California lies near its predicted elevation, whereas the western U.S. interior is ~1 km too high and most of the cratonic region is ~1 km too low. Variations in crustal density may be responsible for some elevation fluctuations (*Braile* [1989]; also discussed below), but the broad regions of greatly anomalous elevation must be caused by buoyant upper mantle there. For example, to account for elevations that are ~1.0 km higher than predicted by crustal thickness, the upper mantle could be 1% buoyant to depths of ~100 km.

In the Great Basin, where both high- and low-velocity upper mantle are found, a more developed case can be made that the generally high elevations of this region are due to a buoyant upper mantle (and that major variations in crustal density create only relatively modest variations in elevation). Predicted elevations there are 1-2 km too great (Figure 6), with the most anomalous elevations actually occurring over the regions of inferred high-velocity upper mantle in the central Great Basin. Topography is depressed ~0.6 km in the regions near the Sevier depression and the Carson Sink, when compared to the elevations of the nearby Great Basin. A high-velocity layer that locally attains thicknesses of ~9 km is found near the base of the crust beneath the Sevier depression [*Pechmann et al.*, 1984; *Smith et al.*, 1989] and the Carson Sink [*Priestly et al.*, 1982; *Thompson et al.*, 1989]. At $V_P \approx 7.6$ km/s, these layers are seismically more similar to the upper mantle ($P_n = 7.7$–8.0 km/s [*Hearn et al.*, 1991]) than to the overlying crust. This layer is thought to be transitional between the crust and mantle (basaltic to ultramafic in composition) [*Thompson et al.*, 1989] and is

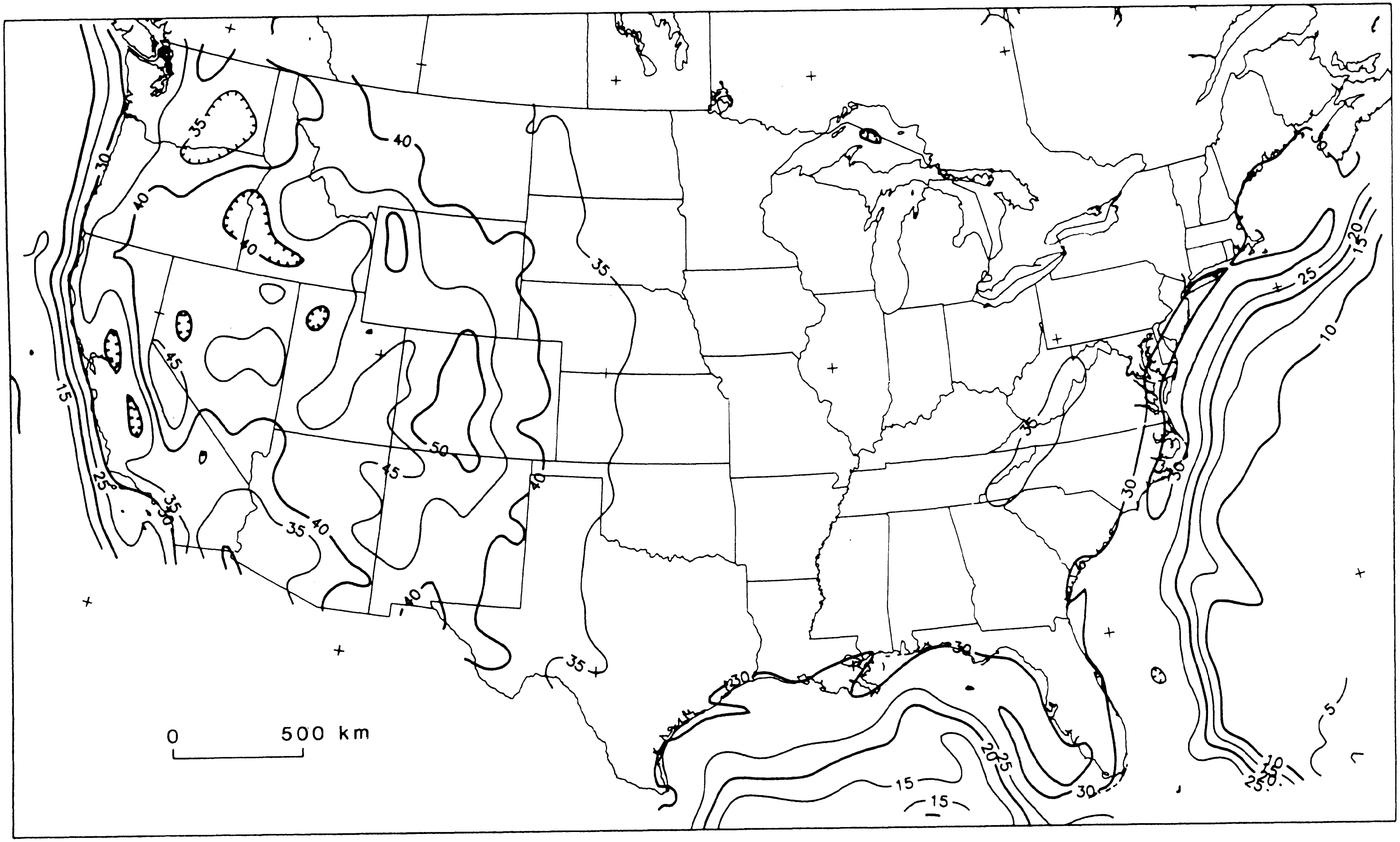

Fig. 4. Long-wavelength elevation or predicted crustal thickness (contours are in km of crustal thickness). This map [from *Simpson et al.*, 1986] presents elevation (filtered to suppress wavelengths shorter than 200 kilometers). This is equivalent to predicted crustal thickness by assuming Airy isostasy and a simple crustal density model. Crustal thickness is calculated by assuming Airy isostasy, using a density contrast across the Moho of $0.35 \times 10^{-3}\,\mathrm{kg\,m^{-3}}$, a density for near-surface rock of $2.67 \times 10^{-3}\,\mathrm{kg\,m^{-3}}$, and an isostatically balanced crustal thickness at sea level of 30 km. By subtracting 30 km from the values in the figure, and multiplying by 0.35/2.67, one obtains the smoothed elevation. The most important point, for our purposes, is that a great thickness of crust is predicted for the elevated western U.S. interior.

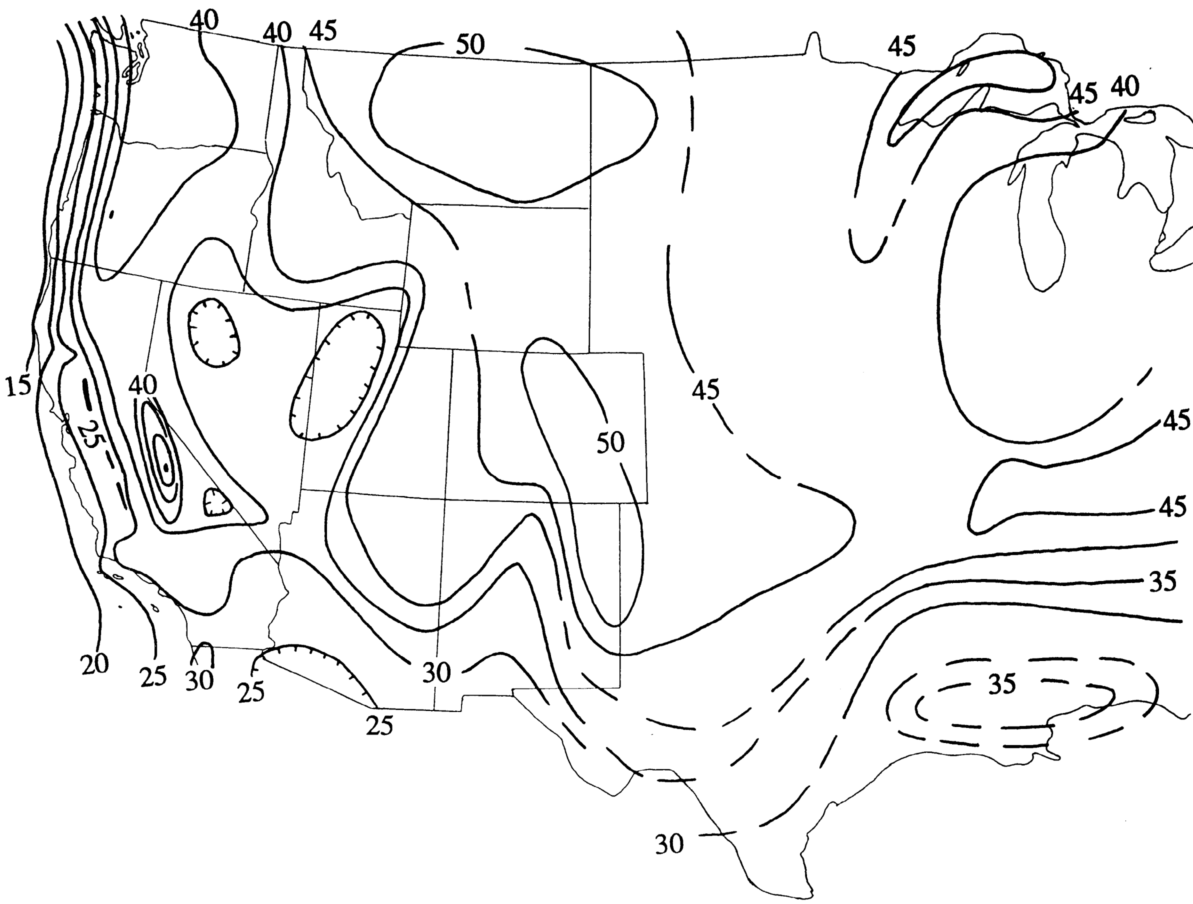

Fig. 5. Seismically inferred crustal thickness. This map is after the maps of *Braile et al.* [1989] and *Mooney and Weaver* [1989], modified so as to remove internal inconsistencies and make contours compatible across map boundaries. In areas where seismic control is weak, inferred crustal thickness appears to be guided by the ideals of Airy isostasy. Even so, in the elevated western United States, the observed crustal thickness is much thinner than predicted by crustal Airy isostasy alone (Figure 4). In some descriptions of the Great Basin crust, igneous underplating results in a uniform crustal thickness of ~35 km [e.g., *Thompson et al.*, 1989]. This layer is not included in this figure. Thickness of the Sierra Nevada crust also is controversial, with some researchers preferring a much smaller crustal root [e.g., *Jones et al.*, 1994].

interpreted as the result of magmatic underplating at the base of the crust that has occurred during the mid-Cenozoic [*Gans*, 1987], possibly associated with crustal extension [*Priestly et al.*, 1982]. The choice of whether to include this layer with the crust or the mantle varies with researcher. As shown in Figure 5, the thickness of crust above this layer varies considerably. If this layer were incorporated with the crust, a nearly uniform crustal thickness of 35 km results [*Thompson et al.*, 1989]. The isostatic effect of replacing 5 km of lower crust with basalt would be the depression the surface 0.3-0.6 km, and 5 km of crustal thinning would cause ~0.7 km of depression [*Lachenbruch and Morgan*, 1990]. Hence, the modest elevation differences within the Great Basin can be explained by known crustal density structure, whereas the generally high elevations of the Great Basin must be attributed to an upper mantle that averages more buoyant, for instance, than California upper mantle (Figure 6). Furthermore, the seismically heterogeneous upper mantle beneath the Great Basin appears to be roughly uniformly buoyant.

Similar reasoning holds qualitatively for the high-standing continental interior as a whole, though greater regional variations in crustal structure make such isostatic accountings more difficult. In particular, local depressions occur where high-velocity "lower crust," similar to that just discussed, is inferred (Walker Lane Belt [*Jones et al.*, 1992], south-central Washington [*Catchings and Mooney*, 1988], eastern Snake River Plain [*Sparlin et al.*, 1982]). Also, elevations are anomalously high (Figure 6) over both high- and low-velocity upper mantle. This is clear in Colorado, where variations in crustal thickness are smooth and not great [*Sheehan et al.*, 1992], yet the upper mantle structure is very heterogeneous and shows little resemblance to the surface tectonic patterns [*Lee and Grand*, 1993; K. Dueker, unpublished data, 1994].

In support of the arguments for a relatively buoyant high-velocity upper mantle is the fact that the current high elevations of much of the continental interior involve crust that was near sea level prior to the Cretaceous [*Sahagian*, 1987]. The elevation of the Colorado Plateau is especially relevant because it resides ~2 km higher than its pre-Laramide elevation near sea level and has not been strongly deformed [*Elston and Young*, 1991]. The tectonic and magmatic history experienced by the crust throughout most of the western U.S. interior is more complex. However, the crustal thickening (largely Cretaceous) and collapse (Tertiary) that have dominated crustal evolution since early Cretaceous times [*Coney and Harms*, 1984] have not created an unusually thick crust, nor have the geologic events diminished average crustal density (igneous addition to the crust will increase its density). Hence upper mantle density must have decreased significantly since pre-Cretaceous times.

PHYSICAL STATE OF THE UPPER MANTLE

Lateral variations in V_P structure are imaged at ~8%, and if seismic structure is confined above 200 km, seismic contrasts are ~10% [*Humphreys and Dueker*, 1994]. Because composi-

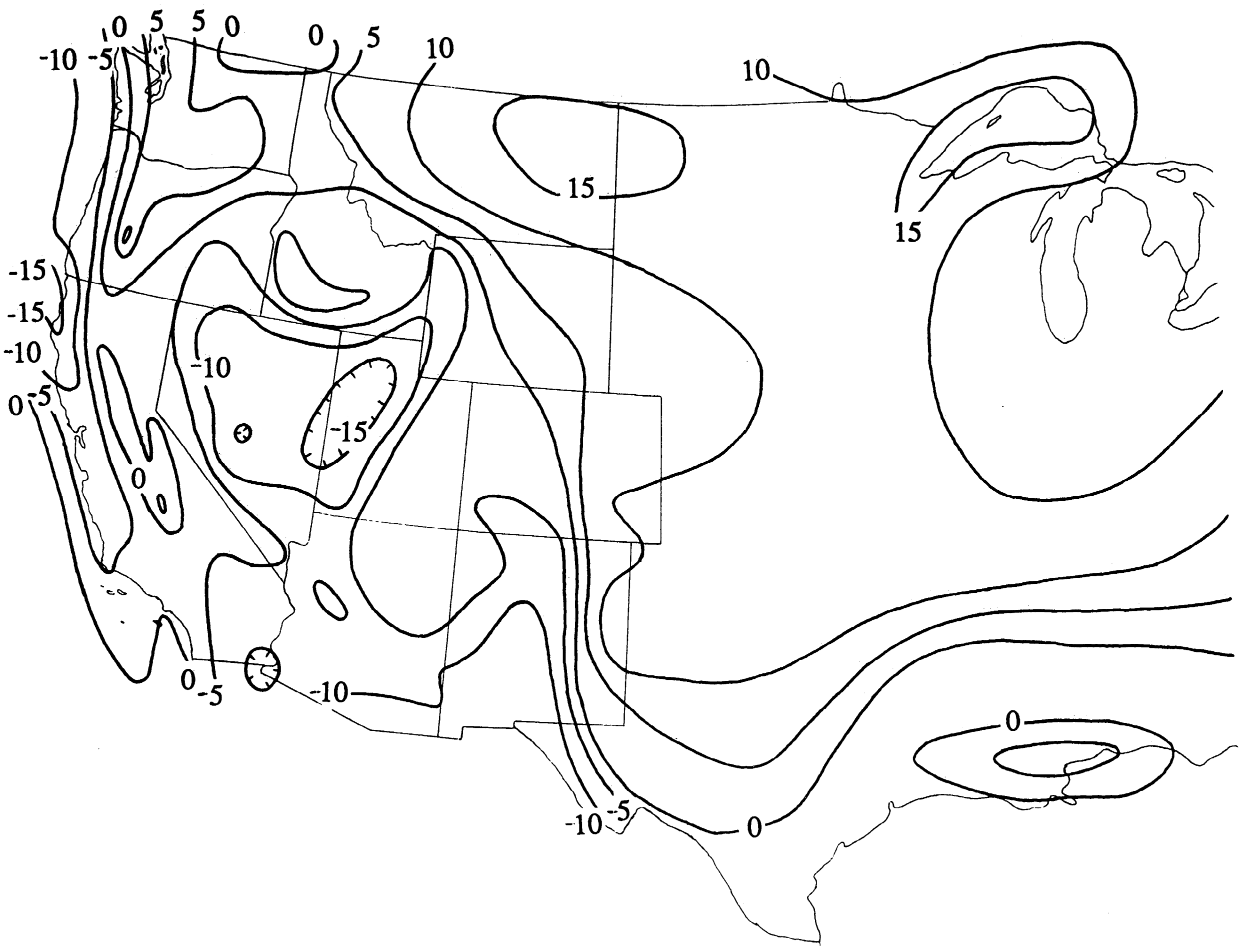

Fig. 6. Anomalous crustal thickness or elevation (contours are in km of crustal thickness). This map indicates the difference between observed crustal thickness (Figure 5) and thickness predicted by Airy isostasy (Figure 4). Assuming a crustal structure described in Figure 4, dividing excess crustal thickness shown in this figure by about -8 yields an elevation anomaly. Lateral variations in crustal density can account for only a small portion of this anomaly.

tional variations probably account for no more than ~1% V_P variation (Table 1), temperature and partial melt variations are primarily responsible for the observed structure. Maximum lateral temperature variations in the upper mantle are limited to ~500°C by considering the juxtaposition of mantle at solidus temperatures of ~1300°C [*McKenzie and Bickle*, 1988] against a lithosphere with Moho temperatures of ~800°C (a typical Great Basin estimate [*Lachenbruch and Sass*, 1977]). Using $\partial V_P/\partial T = -0.5$ m s^{-1} K^{-1} (Table 1), 600°C can explain only ~3% variation in V_P. The fraction of partial melt required to cause an additional 4% of V_P depression ranges from about 1.5 to 4%, depending on the geometry taken by the melt [*Schmeling*, 1985]. This suggests the presence of >1% partial melt in the lowest velocity portions of the upper mantle. Using instead $\partial V_P/\partial T = -2.0$ m s^{-1} K^{-1} for the 200°C range in temperature just below the solidus [*Sato et al.*, 1989] and $\partial V_P/\partial T = -0.5$ m s^{-1} K^{-1} for cooler temperatures, 500°C temperature variation can cause a 7% variation in V_P. Thus with knowledge only of lateral variations in V_P structure and with limited understanding of the relation scaling V_P to T, we conclude that at least a portion of the V_P structure is caused by the presence of partial melt. Because actual V_P variations probably are greater than those imaged, and because 500° of lateral temperature variation probably is unreasonably great, significant fractions of partial melt are likely to be present.

Instead of attributing as much structure to temperature as possible, we now consider holding temperature constant and attribute all of the imaged upper mantle V_P structure to variations in the fraction and geometry of partial melt. If melt fractions >3% are considered implausible (e.g., 3% is much greater than thought possible by *McKenzie* [1984; 1985]), it is not possible to account for an 8% variation of V_P solely with partial melt unless the melt volumes have aspect ratios of 0.01-0.001 (Table 1). Hence it appears that some of the V_P structure is caused by subsolidus temperature variations. The fact that V_P amplitudes are underestimated by some amount only strengthens the need for partial melt.

To better define the volumes of subsolidus and supersolidus upper mantle, additional information must be considered. In principal, information on density offers important constraints on the upper mantle temperature, partial melt, and compositional variations. To produce a given increase in V_P, the required temperature changes cause much greater increases density than do the required changes in partial melt content (Table 1). Knowledge of density also provides information on composition because basalt depletion increases V_P significantly while actually decreasing density (Table 1).

Marginal-Domain High-Velocity Upper Mantle

In this subsection we discuss evidence that the high-velocity upper mantle of the marginal domain is lithospheric in origin (i.e., is relatively cool). High-velocity upper mantle beneath the Pacific Northwest is attributed to the subducted oceanic lithosphere of the Juan de Fuca and Gorda plates. The high velocity structure beneath the Transverse Ranges in southern California has been interpreted as descending lithosphere, based on the magnitude of the seismic anomaly [*Bird and Rosenstock*, 1984; *Humphreys and Hager*, 1990] and on the pattern of southern California tectonics, which requires a positive density anomaly beneath the Transverse Ranges [*Bird and Baumgardner*, 1984; *Sheffels and McNutt*, 1986; 1987; *Humphreys and Hager*, 1990]. The required density variations are predicted only with the use of $V_P - T$ scaling similar to $\partial V_P / \partial T \approx -0.5$ m s^{-1} K^{-1} (i.e., scaling such as suggested by *Sato et al.* [1989] results in too little density anomaly).

Unlike the other portions of the marginal domain, research bearing on the density of high-velocity upper mantle beneath central California is quite limited. However, simple isostatic considerations suggest that the high-velocity upper mantle there is also relatively dense. Crustal density variations would need to be great to cause the southern Great Valley, of crustal thickness 25-35 km (Figure 5), to lie 0.5-1 km below the Coast Ranges, of crustal thickness near 25 km. The isostatic gravity field is consistent with such a density structure; high gravity occurs above high-velocity upper mantle and low gravity occurs to either side (although *Simpson et al.* [1986] associate this field to crustal structure). Also, unlike the recent uplift of the Sierra Nevada and Coast Ranges, tectonic uplift of the southern Great Valley has been only 200-400 m since 20 Ma [*Loomis and Glazner*, 1986]. Upper mantle processes probably have caused the uplift to the east and west of the southern Great Valley. A thinned mantle lithosphere has been suggested in the slab-free zone beneath the Coast Ranges [*Lachenbruch and Sass*, 1980; *Furlong et al.*, 1989], and the uplift of the Sierra Nevada probably is related to processes active in the continental interior such as removal of mantle lithosphere or foundered subducted slab [*Crough and Thompson*, 1977; *Jones et al.*, 1994; *Biasi and Humphreys*, 1992] and the tilt or flexural response to normal faulting on the eastern front of the Sierra Nevada [M. Anders, personal communication, 1992].

Thus along the length of the marginal domain, evidence either suggests or is consistent with high-velocity upper mantle being relatively cold and dense, i.e., thermal lithosphere.

Interior-Domain High-Velocity Upper Mantle

The above discussion about interior-domain upper mantle density indicates that this mantle is buoyant compared to the upper mantle of the marginal domain, in spite of the fact that upper mantle velocities there are similar to marginal velocities in average value and range of value. Furthermore, because elevation anomalies (Figure 6) in the Great Basin are similar above both high- and low-velocity upper mantle, the high-velocity upper mantle is not dense compared to the adjacent low-velocity upper mantle. Hence this high-velocity upper mantle cannot be attributed solely to low temperatures. Only compositional variations due to basalt depletion can account for the buoyancy of the high-velocity upper mantle of the interior domain.

By comparing the high-velocity Great Basin upper mantle to California upper mantle of similar velocity, a rough estimate for the amount of Great Basin basalt depletion can be obtained from isostatic arguments. From Table 1, a depletion of 10% basalt decreases the density of the residuum by ~1% and increases V_P by ~1%. In order to compare the California and Great Basin upper mantle of the same velocity, the depleted upper mantle must be slightly hotter in order to offset this V_P increase resulting from basalt depletion. After making this correction, a net increase in buoyancy of ~1.3% occurs with a 10% depletion of basalt (i.e., $(1/\rho_0)(\partial\rho/\partial R)|_{V_P} = -0.13$). Using this relation, the creation of 1 km of surface uplift requires the segregation of a basalt layer ~8 km thick (e.g., 8% depletion over a 100-km column of upper mantle). Considering the uncertainties, the 1 km (or slightly more) of elevation excess for the Great Basin areas underlain by high-velocity upper mantle compared to California areas underlain by high-velocity upper mantle (Figure 6) requires the segregation of basalt that would create a layer averaging 5-10 km thick. Some of this basalt can be accounted for in regions where seismic studies suggest a basaltic underplating of the Great Basin crust, with thicknesses locally attaining values of 9 km [e.g., *Thompson et al.*, 1989]. Furthermore, geochemical studies of the voluminous mid-Tertiary western United States rhyolites show that these rocks were largely products of basalt fractionation [*Perry et al.*, 1993], indicating that much of the basalt that was differentiated from the upper mantle has evolved into more typical continental crust.

Low-Velocity Upper Mantle

In the introduction of this section we infer 1-3% partial melt in the low-velocity volumes [see also *Humphreys and Dueker*, 1994]. Melt fractions this great are thought to be unstable to segregation [e.g., *McKenzie*, 1984, 1985] or to convective overturn of the upper mantle (especially if occurring in a thick layer [*Tackley and Stevenson*, 1993] such as is inferred beneath the western United States). This is consistent with the observations that high heat flow and volcanism are common above the regions of low-velocity upper mantle. The rate at which melt leaves the upper mantle depends on factors that control melt percolation rate or convective instability (such as filmlike melt geometries [*Waff and Faul*, 1992; *Faul et al.*, 1994], a decrease in melt buoyancy with depth [*Stolper et al.*, 1981; *Agee and Walker*, 1988], melt viscosities that are greater than expected, and the presence of upper mantle mineral assemblages that do not form interconnected melt networks [*Nakano and Fujii*, 1989]). If large fractions of partial melt have segregated from the upper mantle during earlier times [e.g., *Gans*, 1987; *Perry et al.*, 1993], then upper mantle partial melt variations in the interior domain could be primarily the result of variations in upper mantle fertility (i.e., not temperature), with infertile volumes being relatively devoid of partial melt, seismically fast and compositionally buoyant. The fact that the high-velocity upper mantle of the Great Basin appears not to be dense relative to the low-velocity upper mantle suggests that this is the case.

DISCUSSION

Figure 7 shows inferred structure of the western United States at a depth of 100 km. Mantle of *P* wave velocity lower than the regional average is assumed to be partially molten, whereas the higher-velocity mantle is assumed to be subsolidus. This distinction is guided by the distribution of young magmatism and arguments based on V_P–T and $V_P - \phi$ scaling developed above. A cross section through this structure is shown in Figure 8, running from the Pacific plate off south-central California to the central United States. High-velocity upper mantle is represented as being of chemical (depleted) or thermal (cool) origin. As discussed below, most of this structure is thought to have been created after the Laramide orogeny. However, lithosphere immediately beneath the crust is older than this. Figure 8 shows a layer of pre-Tertiary thermal lithosphere beneath California and an irregular layer of Precambrian upper mantle beneath the continental interior. We show the Precambrian lithosphere extending to depths that, in places, include the garnet stability field (~70 km) [*Bennett and DePaolo*, 1987; *Perry et al.*, 1987; *Livaccari and Perry*, 1993]. A transition to the thicker lithosphere of the stable craton occurs near the Colorado-Kansas border [*Helmberger et al.*, 1985; *Grand*, 1987; *Lee and Grand*, 1993].

A general youth of the imaged upper mantle structures is inferred, in part, because the diffusion of temperature and the migration of partial melt towards the surface, which act to diminish the magnitudes and gradients of the corresponding

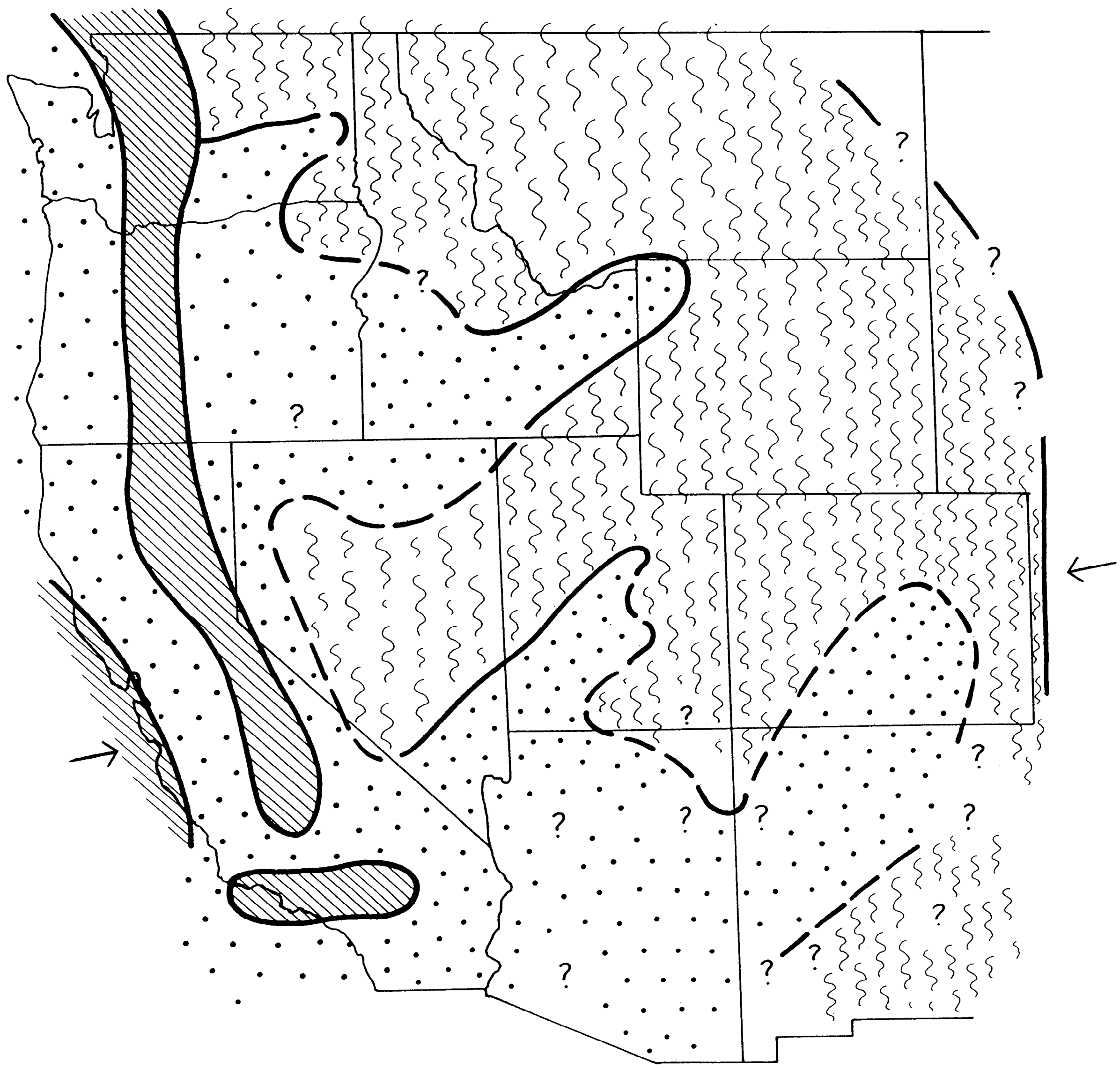

Fig. 7. Map showing hypothesized upper mantle structure of the western United States at a depth of 100 km. The heavy solid and dashed line separates regions of inferred partial melt (stippled) from regions mostly devoid of partial melt. This line is dashed where its location is approximate, and queries indicate regions where knowledge of upper mantle structure is especially poor. Subsolidus regions are divided further into thermal boundary layer (diagonally ruled pattern), which is thought to be subducted oceanic slab beneath a marginal domain, and volumes of basalt-depleted upper mantle beneath the interior (wavy pattern). The interior domain is characterized by generally high elevations and a northeast orientation to the low-velocity upper mantle structures, which are associated with regions of volcanism younger than about 15 Ma. The marginal domain is characterized by lower elevations and high-velocity upper mantle structures that trend parallel to young tectonic structures. The Walker Lane Belt (near the California-Nevada border) is a tectonic zone that parallels upper mantle structure and also commonly expresses young volcanism; as such it appears to be transitional between the marginal and interior domains. The approximate location of the upper mantle transition between the interior domain and the platform/craton is shown near the Colorado-Kansas border, based on an observed strong gradient in teleseismic delays of *S* waves [*Lee and Grand*, 1993] and *P* waves (K. Dueker, unpublished data, 1994). The northward continuation of this transition is estimated using the line of equal elevation. Arrows indicate cross section shown in Figure 8.

seismic field, probably proceed on time scales of less than a few tens of millions of years for the dimensions of imaged upper mantle structures. Furthermore, the upper mantle structure correlates well with young volcanic activity in the interior domain and with young tectonic activity in the marginal domain. These associations suggest that convective processes control the physical state of the upper mantle, which, in turn, influence ongoing crustal tectonic and magmatic activity.

Tectonism and Magmatism

The processes that have created the marginal domain are thought to be controlled by plate-margin activity. Tectonism

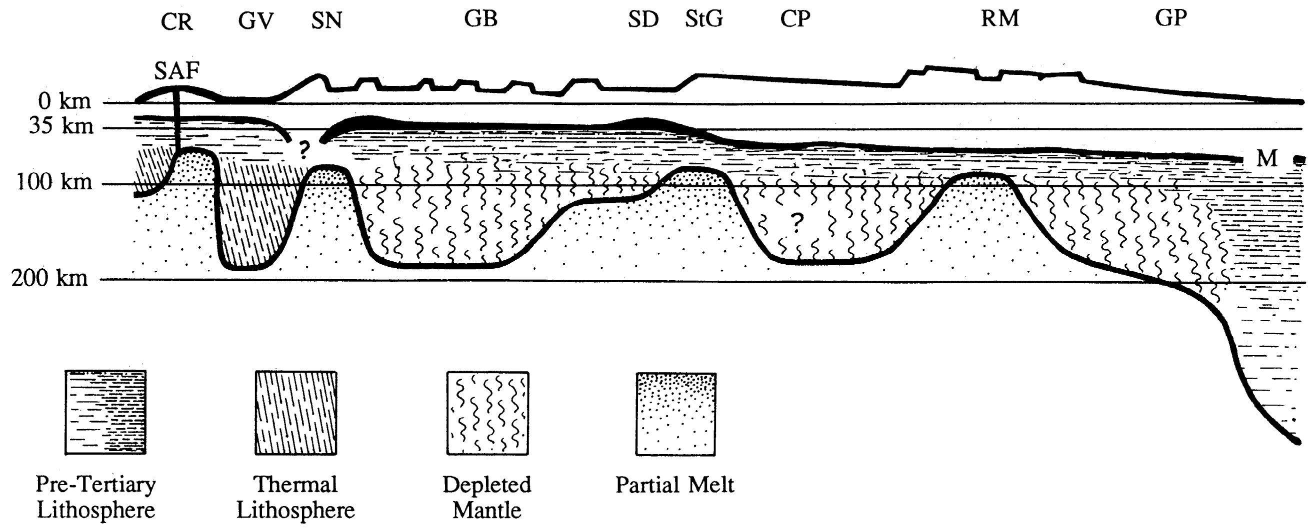

Fig. 8. Cross section of hypothesized upper mantle structure of the western United States, showing various types of upper mantle. Cross section (general location shown on Figure 7) runs from west of central California to east of Colorado, crossing the Coast Ranges (CR), San Andreas fault (SAF), Great Valley (GV), Sierra Nevada (SN), Great Basin (GB), southern Sevier Depression (SD), St. George volcanic trend (StG), Colorado Plateau (CP), Rocky Mountains (RM), and Great Plains (GP). Elevation (greatly exaggerated) is shown for reference. Vertical exaggeration beneath the surface is about 1.5. Moho (M) defines the base of the crust. The crust generally thins to the west. Lenses of basalt-rich rock recently added near the base of the crust are shown beneath the Sevier Depression and the western Great Basin. The horizontal-ruled pattern shows a layer of pre-Tertiary North America lithosphere lying just beneath the Moho. Along the line of the cross section, this lithosphere is Precambrian in age east of the Sierra Nevada (shown with the relatively dense pattern), and younger to the west (shown with the less dense pattern). Beneath this lithosphere lies a heterogeneous upper mantle that is post-Laramide in age. Shown with a diagonally-ruled pattern are post-Laramide structures near the continental margin. This mantle is thought to be dense thermal lithosphere, which is probably subducted oceanic slab. Beneath the continental interior the post-Laramide upper mantle is thought to be hot and buoyant compared to global average. Shown with a stippled pattern are large volumes of partially molten mantle extending to depths considerably greater than 100 km. This mantle is buoyant and often is associated with regions of volcanism younger than about 15 Ma. The wavy pattern represents post-Laramide upper mantle that is relatively depleted of basalt. Because of the depletion, this mantle is buoyant but remains subsolidus.

and magmatism in the Pacific Northwest is dominated by subduction, and imaged upper mantle structures are generally consistent with the setting. South of the Cascadia subduction zone, tectonism is dominated by the strike-slip faulting associated with the transform setting. Based on the distribution of transform-accommodating deformation, plate interaction extends across California and into western Nevada, with diminishing importance to the east [*Atwater*, 1970]. Although right-lateral slip on faults that parallel the San Andreas fault dominates the tectonics in this region, deformation involving zones with important components of normal and thrust faulting are common (e.g., the Salton Trough [*Elders et al.*, 1972], the Walker Lane Belt [*Stewart*, 1988], the Transverse Ranges [*Meisling and Weldon*, 1989], and the Coast Ranges [*Namson and Davis*, 1988]). Thus the marginal domain is a region where plate-margin interactions are the primary causes of crustal tectonics and upper mantle flow, although deviations occur that indicate the activity of smaller scale, local processes.

Tectonic activity east of the marginal domain is dominated by normal faulting that commonly accommodates extension away from regionally elevated areas [e.g., *Zoback et al.*, 1989]. This activity is insensitive to the major variations in upper mantle structure (except that known zones of shear, the Walker Lane Belt [*Stewart*, 1988] and the Lake Mead shear zone [*Smith and Sbar*, 1974], occur within zones of low-velocity and presumably relatively weak lithosphere). Within the continental interior the upper mantle structure correlates most strongly with zones of young magmatism. This region has a high average heat flow, is magmatically active, and stands high in spite of its rather normal crustal thickness. These relations suggest that the upper mantle is currently hot and buoyant, and that crustal extension is a result of gravitational collapse above the elevated upper mantle. We conclude that partial melt buoyancy drives convection in the continental interior, giving rise to volcanism, and that sinking lithosphere drives convection and some local tectonic activity in the marginal domain.

Apparently, the physical state of the upper mantle controls much of the tectonic and magmatic activity of the western United States; however, the current physical state is itself thought to be a product of previous tectonic and magmatic events. Events of particular importance are the Sevier/Laramide orogeny, the emplacement of hot upper mantle immediately following the Laramide, and perhaps the overriding of the Yellowstone plume. The area of current anomalous elevation (Figure 6) is essentially the same area that experienced the most recent and pronounced of the Phanerozoic compressional events to have occurred in the western United States, the Sevier/Laramide orogeny. This suggests that the Sevier/Laramide orogeny created the conditions necessary for the current activity, including those necessary for the emplacement of buoyant upper mantle. The Laramide orogeny is thought to have involved a progressive encroachment of low-angle subduction beneath the western United States [*Coney and Reynolds*, 1977; *Bird*, 1988]. An episode of ignimbritic eruption (the ignimbrite "flareup" of *Coney* [1978]) followed Laramide activity. These eruptions require the transfer of great quantities of basalt across the Moho [*Perry et al.*, 1993]. Most reasonably, this occurred with the withdrawal of the Laramide-aged slab from beneath the western United States, exposing the base of North America lithosphere to asthenosphere [e.g., *Christenson and Lipman*, 1972]. The presence of the upper mantle structure beneath the western U.S. interior (Figure 2) in regions thought to be occupied once by the

Laramide-aged slab suggests that this structure was created within the asthenosphere that ascended into the volume vacated by the Laramide-aged slab.

To explain the upper mantle structure, the high elevation of the western United States interior, and the history of magmatism, we suggest the following. Nearly horizontal subduction of the Laramide-aged slab mechanically thinned the lithosphere to a thickness of perhaps 100 km (which would be thinned further by more recent extension). The subsequent removal of the slab left an isostatic column that would be relatively buoyant and high standing. The asthenosphere probably was (and still is) hot enough to produce partial melt to depths of ~200 km. This is consistent with the upper mantle structure of East Pacific Rise, which defines a broad low-velocity (i.e., hot) volume to ~200 km depth [*Su et al.*, 1992]. Segregation of basalt from portions of the upper mantle creates most of the seismic structure imaged beneath the interior domain, and causes additional uplift by decreasing the net density of the isostatic column. A rapid heating of the crust occurred, primarily advectively through magma ascent, giving rise to the ignimbritic activity. The sudden initiation of ignimbritic magmatism occurred behind two propagating fronts, one that advanced to the south across the Great Basin, and another that advanced west-northwest across the southern Basin and Range [e.g., *Armstrong and Ward*, 1991]. This suggests that removal of the sub-horizontal slab occurred by breaking and pulling away from its northwestern and southeastern margins, such as could occur by buckling and downwelling along a northeast-oriented trend within the central portion of the slab (perhaps associated with the Farallon-Vancover break [*Severinghaus and Atwater*, 1990]),

The presence of a broad and large-amplitude geoid anomaly centered on Yellowstone [*Milbert*, 1991] and the anomalous helium being released there [*Craig*, 1993] suggest that a plume associated with Yellowstone has an important influence on North America lithosphere in this region. The geologic association of Yellowstone with the eastern Snake River Plain suggests a similar cause of volcanic and tectonic activity in this region as well [e.g., *Anders et al.*, 1989]. The St. George and Jemez volcanic and low-velocity upper mantle trends parallel the Yellowstone trend. The relatively young volcanism associated with these trends, and the uplift of western United States as a whole, have been attributed to plume activity [*Suppe et al.*, 1975] associated with the absolute motion of North America. This gives an alternative explanation to the overriding of hot upper mantle associated with the East Pacific Rise for the activity of the western U.S. interior. Because evidence can be found for both Laramide-related and plume-related processes, both of these events may have been (and may continue to be) important to the distribution and nature of western U.S. volcanism and tectonism. Of course, these two processes could be related: plume activity may have organized upper mantle flow patterns; and the large volumes of low-velocity mantle may be the result of long-standing plume activity.

Length Scales

Western U.S. upper mantle structure (Figures 2 and 7) shows three characteristic length scales: (1) Where structures are well resolved, they commonly extend to about 150-200 km depth [*Dueker et al.*, 1993; *Humphreys and Dueker*, 1994]. Although most western U.S. upper mantle structure is not well resolved in depth, observed delay times and test inversions are consistent with upper mantle structures of this depth extent throughout most of western United States. Exceptions include the Juan de Fuca and Gorda slabs, which extend to greater depths. We think that ~200 km represents the base of the active upper mantle beneath western United States, above which the upward mobility of partial melt or the sinking of mantle lithosphere create young structures of high seismic contrast. (2) Western U.S. upper mantle consists of three domains, two in regions that we image (i.e., the marginal and interior domains), and a third domain beneath the craton [e.g., *Grand*, 1987]. The marginal domain is about 250 km wide, which probably represents the horizontal distance over which plate-boundary interactions [*England et al.*, 1985] dominate over the locally driven instabilities of the interior. This approximate length scale applies to both the transform boundary of California and the oblique subduction boundary of the Pacific Northwest. A transition zone that includes the Walker Lane Belt displays characteristics of both the margin (significant shear tectonism [*Argus and Gordon*, 1991; *Pezzopane and Weldon*, 1993]), and the interior (margin-normal extension and young magmatism). The width of the western U.S. interior domain is not clear because high-resolution images of the upper mantle are not available for the eastern part of this region. Regional-scale seismic evidence indicates that the interior domain is separated from the cratonic domain near the Colorado-Kansas border [*Grand*, 1987; *Helmberger et al.*, 1985; *Lee and Grand*, 1993]. (3) Imaged structures of the upper mantle domains commonly are elongated and express a wavelength of ~250 km beneath the continental margin and ~500 km beneath the western U.S. continental interior. The latter value may represent a characteristic wavelength for asthenospheric upwelling in the upper few hundred kilometers beneath this region.

SUMMARY

The upper mantle teleseismic P wave delay times are the most important information upon which our description of the physical state of the western U.S. upper mantle is based. Also important are large-scale aspects of the elevation and crustal structure, which are used to estimate upper mantle density variations, and the tectonic and volcanic history of the region since ~100 Ma. We conclude the following. (1) The low-velocity volumes of upper mantle are caused by the presence of hot, partially molten mantle, which in places extends to depths greater than 100 km. (2) The high-velocity volumes beneath the continental interior are caused by basalt depletion, which has created a buoyant and infertile residuum. (3) The high-velocity volumes near the continental margin are largely the result of thermal lithosphere emplaced into the low-velocity upper mantle. (4) The upper mantle structures postdate the Laramide (i.e., are younger than ~50 Ma).

We make a distinction between marginal and interior domains, based on the inferred compositional difference in high-velocity upper mantle, on associations between surface tectonics and upper mantle structure in the marginal domain, and on associations between volcanism and upper mantle structure in the continental interior. In the interior domain the dominant physical process is thought to be the ascent of partial melt in NE-trending zones which are associated with young volcanism and high heat flow. High elevations are common in the interior, and tectonism typically is extensional, occurring at relatively slow rates. Extension there usually is oriented normal to the regional topography and is discordant with upper mantle structure. These relations suggest that uplift provided by buoyant upper mantle has induced a regional gravitational collapse of the elevated crust. In the marginal domain, plate interactions (transform and subduction) appear to dominate deformation processes. However, deformation patterns deviate significantly from what would be expected of simple margin-related processes, suggesting that locally created forces associated with sinking high-density (high-velocity) upper mantle are important locally.

Many aspects of the physical state are not well constrained because of uncertainties in the scaling relations used and in the structure of the crust and upper mantle. It is expected that knowledge, much of it yet to be gained, on the V_P, V_S, Q_P, Q_S and anisotropy structure of the upper mantle will help distinguish the physical properties of the upper mantle, as it is expected that a better understanding of the density structure of the crust will allow for a better-constrained estimate of upper mantle densities. This broader geophysical data set will permit more strongly constrained assessments of the processes that have created the upper mantle.

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