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How Laramide-Age Hydration of North American Lithosphere by the Farallon Slab Controlled Subsequent Activity in the Western United States

EUGENE HUMPHREYS,[1]

Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403

ERIN HESSLER,[2]

Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403

KENNETH DUEKER,

Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071

G. LANG FARMER,

Department of Geological Sciences, University of Colorado, Boulder, Colorado 80309

ERIC ERSLEV,

Department of Earth Resources, Colorado State University,Fort Collins, Colorado 80523

TANYA ATWATER,

Department of Geological Sciences, University of California, Santa Barbara, California 93106

Abstract

Starting with the Laramide orogeny and continuing through the Cenozoic, the U.S. Cordilleran orogen is unusual for its width, nature of uplift, and style of tectonic and magmatic activity. We present teleseismic tomography evidence for a thickness of modified North America lithosphere ≥200 km beneath Colorado and >100 km beneath New Mexico. Existing explanations for uplift or magmatism cannot accommodate lithosphere this thick. Imaged mantle structure is low in seismic velocity roughly beneath the Rocky Mountains of Colorado and New Mexico, and high in velocity to the east and west, beneath the tectonically intact Great Plains and Colorado Plateau. Structure internal to the low-velocity volume has a NE grain suggestive of influence by inherited Precambrian sutures. We conclude that the high-velocity upper mantle is Precambrian lithosphere, and the low-velocity volume is partially molten Precambrian North America mantle.

We suggest, as others have, that the Farallon slab was in contact with the lithosphere beneath most of the western U.S. during the Laramide orogeny. We further suggest that slab de-watering under the increasingly cool conditions of slab contact with North America hydrated the base of the continental lithosphere, causing a steady regional uplift of the western U.S. during the Laramide orogeny. Imaged low-velocity upper mantle is attributed to hydration-induced lithospheric melting beneath much of the southern Rocky Mountains. Laramide-age magmatic ascent heated and weakened the lithosphere, which in turn allowed horizontal shortening to occur in the mantle beneath the region of Laramide thrusting in the southern Rocky Mountains. Subsequent Farallon slab removal resulted in additional uplift through unloading. It also triggered vigorous magmatism, especially where asthenosphere made contact with the hydrated and relatively thin and fertile lithosphere of what now is the Basin and Range. This mantle now is dry, depleted of basaltic components, hot, buoyant, and weak.

[1]Corresponding author; email: gene@geops2.uoregon.edu
[2]Present address: Illinois State Water District, Champaign, IL.

Introduction

ROCKY MOUNTAIN TECTONISM, widespread Tertiary volcanism, and uplift of the western United States are important geologic events occurring within the interior of North America with no obvious cause. As such, placing these activities in a plate tectonic context appeals to unusual behavior, and their causes remain controversial. The ~70–45 Ma Laramide orogeny is central to discussions of this activity. This orogeny is attributed, in some fashion, to rapid subduction of the Farallon slab beneath North America. Coney and Reynolds (1977), noting a cessation of Sierra Nevada arc magmatism and an east-migrating front of magmatic initiation immediately prior to and during the Laramide, suggest Farallon slab flattening beneath the western U.S. Slab-flattening models are made attractive by the occurrence of similar behavior in areas where the Nazca "flat slab" subducts beneath South America (Cahill and Isacks, 1992), and by their ability to associate Rocky Mountain contraction, western U.S. uplift, and the evolving pattern of magmatism to plate tectonic processes. In particular, volcanism is attributed to propagation of subduction-related arc-like volcanism associated with slab flattening followed by roll back (Coney and Reynolds, 1977). Uplift, which cannot be entirely supported by crustal thickening (Sheehan et al., 1995; Spencer, 1996), is attributed to mechanical thinning of the lithosphere by the Farallon slab (Humphreys, 1995; Spencer, 1996).

In this paper, we present evidence that lithosphere beneath much of the Colorado Rocky Mountains is ≥200 km thick and beneath the New Mexico Rocky Mountains is >100 km thick, making arc-like magmatism and lithospheric thinning unlikely explanations for volcanism and uplift. The fact that no convincing explanation has emerged for the Laramide orogeny and broad uplift of the western U.S. suggests the activity of important processes that have remained outside the domain of normal consideration.

We present an explanation for western U.S. uplift, Laramide contraction, and voluminous mid-Tertiary magmatism. We expend the flat-slab model by incorporating slab de-watering to hydrate North America lithosphere under the relatively cool conditions that would be expected during slab contact (Dumitru et al., 1991; Spencer,1996). It is the hydration of North America lithosphere to which we attribute decreased density and uplift of thick western U.S. lithosphere as far east as the Great Plains, and melting beneath the Colorado Rocky Mountains. Melt ascent advectively heated and weakened the lithosphere, allowing contractional failure beneath the Rocky Mountains and creation of seismically slow velocities beneath the Colorado Rocky Mountains to depths of ≥200 km. Post-Laramide volcanism in the Rocky Mountains and Basin and Range is attributed to slab removal and the resulting asthenospheric contact with the base of the hydrated lithosphere. The exceptionally vigorous magmatism in the Basin and Range is thought to be a result of its relatively thin and fertile lithosphere. Buoyancy in magmatically active areas now is supplied by thermal expansion and basalt depletion of the mantle.

Regional lithospheric setting

Figure 1A is a map of upper mantle seismic velocity structure for the western United States and nearby regions. The eastern part of our study area lies over the western margin of a high-velocity mantle "cratonic root" (Jordan, 1979) that extends deeply beneath the Archean (in central Canada) and Paleozoic (in the north-central U.S.) crust (Grand, 1994; van der Lee and Nolet, 1997). In contrast, most of the western U.S. overlies the northern extent of a large, low-velocity volume that loosely correlates with the East Pacific Rise (Grand, 1994). The contrast betweenthe high-velocity cratonic mantle and the low-velocity western U.S. mantle is among the greatest to be found. Yet the 100–200 km wavelength mantle structure imaged beneath the western U.S. (Fig. 1A) is very heterogeneous, with strong local velocity contrasts that are as great as that seen across the continent as a whole (Humphreys and Dueker, 1994a).

Our seismic investigation spans the transition between the western U.S. upper mantle and the cratonic upper mantle. This transition is not simple (Lee and Grand, 1996; Henstock et al., 1998; van der Lee and Nolet, 1997). Using S-wave data recorded by the same broadband seismometers that we use in our P-wave study, Lee and Grand (1996) found mantle beneath central Colorado to be very slow. In New Mexico, Slack et al. (1996) and Spence and Gross (1990) found low velocities beneath the Jemez volcanic trend that are more prominent that any low-velocity structure beneath the Rio Grand Rift. Our image of P-wave velocity (Vp) provides a resolved image of upper mantle structure in the area of transition to the craton, and helps define the structure internal to this transition.

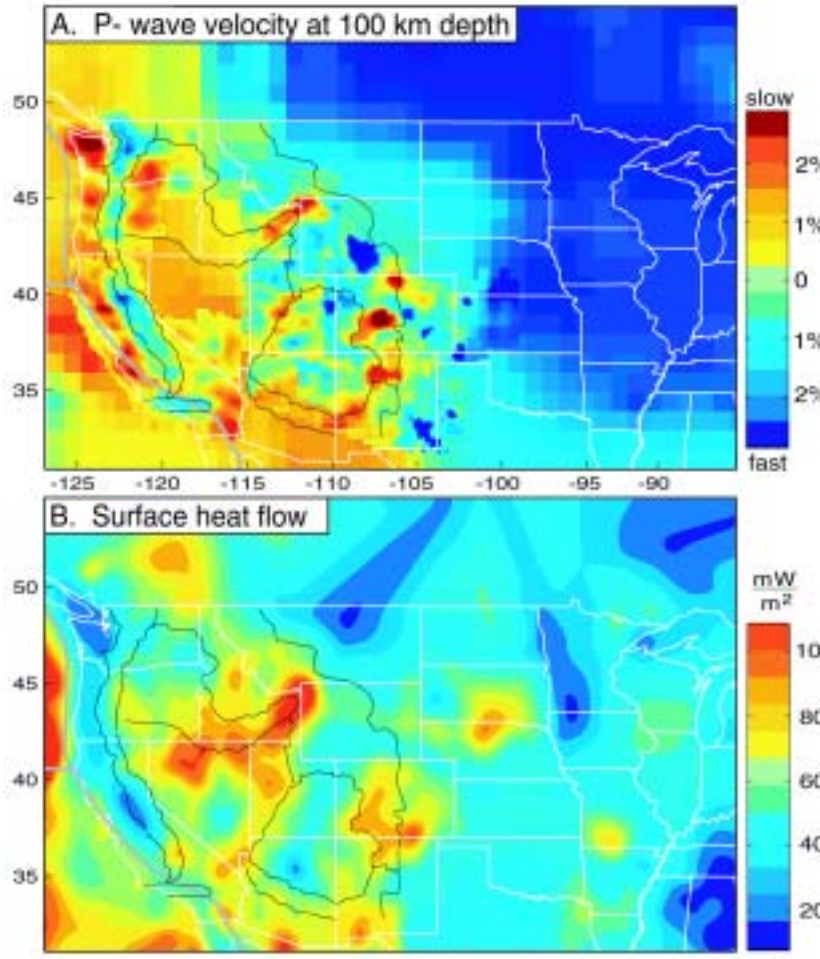

FIG. 1. Seismic velocity and surface heat flow of the western U.S. region. Black lines delineate physoigraphic provinces, and grey lines show plate boundaries. A. Composite image of seismic velocity structure at 100 km depth. The continental-scale image is from the multi-bounce S-wave modeling of Grand (1997), and velocity has been scaled by 0.5 to provide a background estimate of P-wave velocity. Resolution of this image is ~300 km. Overprinting this image are five P-wave inversions of regional array data: the Washington–Oregon, California–southern Nevada, and Idaho–Utah–western Wyoming images of Humphreys and Dueker (1994a), the northern Arizona–New Mexico image of Slack et al. (1996), and the Colorado–New Mexico image from our study. To adjust for differences in average velocity beneath each study area, results from individual inversions have been shifted by the addition of a constant to the delay times of each data set. Adjustments are determined by minimizing differences in velocity where inversions overlap, and with the background image. Because most western U.S. upper mantle is slow, most of the regional images have been made slower. Time adjustments are: northern Arizona–New Mexico, 0.3 s; California–southern Nevada, 0.4 s; eastern Idaho–Utah, 0.2 s; Washington–Oregon, 0.4 s; Colorado, 0.0 s. B. Surface heat flow. Data have been interpolated to a rectangular mesh and smoothed to eliminate wavelengths greater than about 250 km. The strong similarity of the pattern of heat flow with upper mantle seismic velocity and with Cenozoic volcanism (Christenson and Yeats, 1992) suggests that near-vertical ascent of mantle melt at depth (red areas in A) leads to volcanism and high heat flow at the surface. Data are from the heat flow data repository at www.heatflow.und.edu.

Geologic overview

Proterozoic accretion of ~NE-trending arc-like terrains onto the Archean Wyoming province (Karlstrom and Bowring, 1993) created a heterogeneous continental landmass possessing a strong northeasterly grain that has expressed itself repeatedly through geologic time (Henstock et al., 1998; Karlstrom and Humphreys, 1998). Mantle model ages and equilibration depths of xenoliths indicate that western U.S. crustal provinces were underlain by mantle lithosphere of corresponding ages, and that this association persists to the present in most areas (Livaccari and Perry, 1993). Following a period of stability, rifting of the continent occurred at ~600 Ma along a zone that trends north through central Nevada (Burchfiel et al., 1992). Once established, the northerly-oriented continental margin progressively developed over preexisting structures (Karlstrom and Humphreys, 1998).

Paleozoic subsidence of the young continental margin outboard (west) of a slope-bounding flexural hingeline indicates lithospheric thinning and subsequent creation of thermal lithosphere (Sleep and Snell, 1976) to a thickness of ~125 km beneath what now is the eastern half of the northern Basin and Range province (Bond et al., 1989). This lithosphere is expected to be more fertile than the basalt-depleted mantle of the unthinned Proterozoic lithosphere east of the hingeline. The short wavelength of hingeline flexure implies a weak lithosphere at ~600 Ma along the zone that now defines the western margin of the Colorado Plateau (Bond et al., 1989). In comparison, flexure east of the hingeline was relatively long in wavelength and minor in amplitude (Stewart and Suczek, 1977; Jordan, 1981), expressing a more rigid continent beneath the Colorado Plateau and Rocky Mountains at this time. Kimberlites sampled a lithosphere that was thick beneath the eastern Rocky Mountains near the Colorado-Wyoming state line during the Silurian and Devonian (Eggler et al., 1987). Following, the Pennsylvanian-aged Ancestral Rocky Mountain orogeny faulted and uplifted crust across most of New Mexico, Colorado and Utah, probably resulting from subduction or collision along the southeast (Ye et al., 1996) margin of the continent. Conspicuously, Archean lithosphere was avoided in this orogeny. The effects of this orogeny on the lithosphere are not clear, although it is the first event capable of strongly affecting the Colorado Rocky Mountains and Colorado Plateau lithosphere since the Precambrian.

The next (and final) event to strongly affect the Colorado Plateau and southern Rocky Mountains (including the Archean Wyoming province) was the Laramide orogeny, although the lithosphere of southern New Mexico was modified at ~200 Ma by the development of a NW-trending magmatic arc (Reynolds, 1980). Presumably, southern New Mexico lithosphere was less than ~90 km thick over the duration of arc magmatism—i.e., it was thin enough to permit slab dehydration at ~90 km depth to cause asthenospheric melting.

Laramide thrust faulting and crustal shortening occurred as the relatively undeformed Colorado Plateau moved northeast relative to stable North America (Hamilton, 1989; Varga, 1993). Magmatism was widespread but sparse over most of the western U.S. interior during Laramide time (Christenson and Yeats, 1992). In the region of Laramide-age basement-cored uplifts, the most magmatically active area was the Colorado Mineral Belt, trending northeast from near northeastern Arizona across the range of the Colorado Rocky Mountains (Mutschler et al., 1987).

Subsequent to the Laramide orogeny, the Colorado Plateau—Rocky Mountain—Great Plains region experienced minor orogenic collapse and widespread, moderate levels of magmatism, much of it distributed in the vicinity of the Colorado Mineral Belt; exceptions to this were the occurrence of large-volume magmatism starting at ~37 Ma in the San Juan Mountains area of southwestern Colorado, and modest but obvious extension of the Rio Grande Rift. The NE-trending Jemez volcanic trend, which corsscuts the N-trending Rio Grande Rift, has become magmatically active in the last ~10 m.y. (Christenson and Yeats, 1992). Much of highest Colorado still preserves the erosional surfacecreated by the end of the Laramide, and the high Rocky Mountain valleys are primarily a result of Pliestocene glaciation (Small and Anderson, 1998). Information on the timing of uplift (discussed below) indicates that important contributions to uplift occurred both during and following the Laramide orogeny.

Thus, poorly understood Laramide-age processes acted to drive tectonism and magmatism far inboard of the continental margin, and now, ~50 m.y. after the presumed causative event, the resulting high elevations show no signs of falling. Subduction-related Pelona-type schist of Laramide age crop out in southeastern California and southwestern Arizona, and indicate that North American lithosphere was relatively thin in these areas during the Lara-

mide. Colorado Plateau xenoliths from ~140 km depth (Smith, 2000), a ~200 km thick lithosphere beneaththe southern Rocky Mountains (this study; Dueker et al., 2001) and ~250 km thick lithosphere beneath the central U.S. craton (Grand, 1994) suggest a Farallon slab with a dip of ~15 between southwestern Arizona and the central Colorado Plateau, perhaps shallowing beneath the southern Rocky Mountains.

The relative stability of the Rocky Mountains and the Colorado Plateau stands in marked contrast to the pervasive and vigorous magmatism and extension that occurred south and west of this region, in the areas now occupied by the Basin and Range (Coney and Harms, 1984; Christensen and Yeats, 1992).

The net result of Cenozoic activity has been the creation of a high-standing landmass overmost of western U.S. that is composed of distinctive tectonic provinces. These include the greatly extended Basin and Range crust to the west and south of the relatively coherent Colorado Plateau and southern Rocky Mountains, which themselves bound the tilted and intact Great Plains. The Rocky Mountains are distinguished from the Colorado Plateau by higher average elevation and greater magnitude of crustal shortening and mountain building during the Laramide orogeny. The Colorado Plateau and Rocky Mountain lithosphere in Utah and Wyoming currently is much stronger than that of the Basin and Range, as revealed by flexural strength studies (Forsyth, 1985; Lowry and Smith, 1995).

Evidence for "flat-slab" subduction during the Laramide orogeny

The blocky, thrust-bounded Rocky Mountain uplifts are the most obvious manifestation of the Laramide orogeny. A wide range of causes have been suggested for this faulting, including basal contact of the Farallon slab. Without additional evidence, it is not possible to argue that, among the various models, Farallon slab contact was the actual cause for Rocky Mountain faulting.

More convincing is the evolution of western U.S. magmatism preceding, during, and after the Laramide orogeny. Pre-Laramide volcanism migrated east across the western United States, reaching as far east as western South Dakota during the Laramide orogeny, and leaving the once-vigorous Sierra Nevada arc quiescent. Jerrard (1986) calculated that an increasingly rapid Farallon subduction rate and younging slab age (Engebretson et al., 1988) should cause an extreme shallowing of subduction angle. This supports the Coney and Reynolds (1977) suggestion that magmatic propagation resulted from progressive Farallon slab shallowing. During the Laramide orogeny, the lithosphere was refrigerated from below in the Great Basin (Dumitru et al., 1991) and the central Colorado Plateau (Riter and Smith, 1996; Smith 2000), an occurrence attributed to the presence of the Farallon slab at the base of North America. Similarly, basal cooling of the Sierra Nevada by the Farallon slab has been used to explain the very low heat flow currently observed throughout most of the Sierra Nevada (see Fig. 1B) (Dumitru, 1990; Saltus and Lachenbruch, 1991), and the creation of a lithosphere there of ~200 km thickness by the end of the Laramide orogeny(Farmer et al., 2002). In addition, low mantle heat flow into the crust at ~60 Ma is required to explain the magmatism of Great Basin mid-crust, which was entirely crustal in origin (Patino-Douce et al., 1990). This interval with cool crust was followed by the extremely energetic "ignimbrite flareup" (Coney, 1980), whose initiation propagated regularly across both the southern and northern Basin and Range (Christensen and Yeats, 1992).

The ignimbritic magmatism represents a sudden change from conditions with low mantle heat flow and no evidence for magma transfer into the crust, to very hot conditions with great volumes of basalt transferred into the crust (Johnson, 1991; Perry et al., 1993). This remarkable magmatic transition seems to require the cooling effectsof the Farallon slab contact with North America during the Laramide, and a progressive Farallon slab removal and exposure of basal North America to hot asthenosphere to explain the mass and thermal budget of the ignimbrite flareup (Humphreys, 1995). Although the crustal welt created by Sevier-age shortening would tend to drive continental extension (Coney and Harms, 1984), it does not appear to be the cause of magmatism: where resolved in central Nevada, magmatism preceded extension (Gans et al., 1989); and the magmatic front was nearly perpendicular to the regional direction of extension, which seems unlikely if magmatism was a simple consequence of extension (Armstrong and Ward, 1991). The ignimbrite flareup was nearly over before the "slab-free window" began to open in conjunction with the growth of transform margin along California (Dickenson and Snyder, 1979), and thus creation of this window was not the cause of ignimbritic activity.

The spatial pattern and timing of vertical motions across the western United States also indicate Farallon slab flattening and subsequent removal from beneath North America. The only available explanation for pre-Laramide subsidence—as evidenced by the Cretaceous "inland seaway"—is that dynamic suction associated with the load of the flattening Farallon slab pulled the western half of the U.S. down (Mitrovica et al., 1989; Gurnis, 1992). Although there is strong evidence for signicant contributions to the uplift both during and after the Laramide orogeny, this history is not well constrained. The inland seaway retreated during the early part of the Laramide orogeny as the Rocky Mountains rose. Gregory and Chase (1994) and Wolfe et al. (1998) interpret leaf morphology data to indicate that elevations had attained heights similar to those found today by the end of the Laramide orogeny. This conclusion is supported by oxygen isotopic evidence (Dettman and Lohmann, 2000). Other authors, however, conclude that most of the uplift occurred following the Laramide orogeny. Pederson et al. (2002) suggest 1.8 km of isostatic Colorado Plateau surface uplift since pre-Laramide times, with much of this occurring after the Laramide orogeny. Significant post-Laramide uplift of the Colorado Plateau is inferred by Sahagian et al. (2002), who report an accelerating uplift to the present. Similarly, much of the uplift and tilting of the Great Plains appears to have occurred since the middle Miocene (Heller et al., 2003). Thus, although the cause and history of uplift is not resolved, it is remarkable that the portion of the western U.S. thought to have experienced Farallon slab contact (based on the extent of Laramide magmatism and tectonism) is the broad area now standing high.

Previous suggestions for the causes of uplift and contraction

Bird (1984, 1988) suggested that the Farallon plate mechanically removed western U.S. mantle lithosphere and transported Basin and Range lower crust to beneath the Great Plains, thereby providing the required buoyancy across the western U.S. However, the isotopic character of post-Laramide magmas suggests preservation of North America mantle lithosphere beneath crust of the same age (e.g., Bennett and DePaolo, 1987), thereby rendering unlikely the wholesale removal of North America lithosphere during the Laramide (Livaccari and Perry, 1993). MacQuarry and Chase (2000) have suggested that uplift was caused by lower-crustal flow from beneath crust thickened during the Sevier orogeny to beneath the Colorado Plateau, Rocky Mountains, and Great Plains. Although this model can explain the persistence of uplift, the proposed flow would tilt the Colorado Plateau to the east and decorrelate the lower crust with the upper crust. We know of no evidence for eastward Colorado Plateau tilting, and in the central Colorado Plateau, lower-crustal xenoliths from either side of an inferred suture (probable Yavapai-Mazatizal) shows that the suture is preserved in the lower crust (Wendlandt et al., 1993; Selverstone et al., 1999). Furthermore, no westward flow occurred as the Basin and Range fell below the height of the Colorado Plateau, and earthquakes are observed deep in the lower crust beneath the Colorado Plateau (Wong and Humphrey, 1989), implying that this crust currently is strong. This is consistent with flexural strength studies (Forsyth, 1985; Lowry and Smith, 1995). All this points to a cool, viscous, and largely intact lower crust beneath the Colorado Plateau.

Because crustal thickening is inadequate to account for much more than half of present-day uplift (Sheehan et al., 1995), and because >1 km of Colorado Plateau and Great Plains uplift opccurred in the near absence of crustal shortening (at least as expressed at the surface), it commonly is thought that the mantle load beneath western North America has been reduced by some process related to Laramide or post-Laramide activity. Spencer (1996) suggested ~1.2 km of Colorado Plateau uplift was caused by reduction of mantle buoyancy, which he attributed to a mechanical thinning of North America lithosphere by a Farallon slab in contact with North America, and the subsequent removal of the dense slab. Heller et al. (2003) suggested uplift was driven by a combination of slab unloading, a young influx of relatively buoyant sublithospheric mantle, and processes associated with Rio Grande rifting.

Most suggestions for the cause of Rocky Mountain contraction seem unlikely in the context of a thick mantle lithosphere. Maxson and Tickoff (1996) appealed to plate buckling in response to collisional activity at the western plate margin, but the amplitude and wavelength of such buckling is not consistent with a thick lithosphere. Bird (1984, 1988) and Egan and Urquhart (1993) invoked a shear thickening of the lower crust driven by eastward transport of the mantle lithosphere. The in-place nature of the mantle lithosphere indicates that this mechanism is improbable (Livaccari and Perry,

1993). Livaccari (1991) suggested that the north-trending zone of thick crust created by the Sevier orogeny provided an eastward push on the Colorado Plateau and Wyoming that drove contraction in the Rocky Mountains. However, this upper-crustal push could not deform a thick lithosphere without extensive lower-crustal decoupling, which seems especially difficult across the broad and strong Colorado Plateau. Also, an east-directed push would not account for apparent right-lateral Laramide deformation in the New Mexico Rocky Mountains (Karlstrom and Daniel, 1993; Cather,1999), and the generally ENE-WSW orientation of slip vectors on Laramide-age faults in Colorado, Utah, and Wyoming (Erslev, 1993).

Seismic Investigation

We invert a composite data set for upper mantle velocity structure beneath portions of the southern Rocky Mountains, Great Plains, Colorado Plateau, and Rio Grande Rift. These data consist of teleseismic arrival times from several line arrays deployed in the New Mexico area, the 1992 Rocky Mountain Front (RMF) seismic deployment, and the 2000 Continental Dynamics of the Rocky Mountains (CDROM) line-array deployments across the Wyoming—Colorado and Colorado—New Mexico borders (Fig. 2A). The older data from New Mexico are from four previously published seismic studies, which were collected as part of a collaborative effort by the University of California, Los Angeles, the U.S. Geological Survey, the Los Alamos National Laboratory,and the University of Karlsruhe to study the upper mantle associated with the Rio Grande Rift (Halderman and Davis, 1991; Davis et al., 1993; Parker et al., 1984; Spence and Gross, 1990; Slack et al., 1996). These data were recorded by vertical-component seismometers that provided teleseismic data at frequencies near 1 Hz, from which high-quality P-wave travel-time picks were made by those involved in the prior studies. Use of the P-wave travel-time residuals from the RMF and CD ROM arrays, which are derived from broad band seismometers provided by the IRIS PASCAL program, are new. Discussion of the new data and their processing are found in the Appendix, as is discussion of their joint-data inversion. In Figure 2, we also show the average delays for Yellowstone array stations. Although data from these stations are not used in tomographic inversion, the station delays help define central Wyoming as a relatively fast upper mantle province.

Seismic results

Tomographic inversion of the combined data set provides an image of *V*p structure in the upper ~450 km of mantle beneath Colorado, New Mexico, and adjacent portions of Arizona, Kansas, Texas, Utah, and Wyoming. We approximate this volume by dividing it into 18 30 km thick layers, and dividing each layer into blocks 25 × 25 km in EW and NS dimension. We enforce horizontal smoothing so that the effective horizontal resolution is ~75 km.

Figure 2 shows the observed station-average delays, the crustal corrections calculated from independent information on elevation, sediment basin thickness (Woodward, 1988), and crustal thickness (Keller et al., 1998), and the station-average delays that result from application of these corrections. The effect of the crustal corrections is seen to be relatively minor. The delays corrected for crustal structure are the data input to the tomographic inversion.

In addition to the estimated station corrections, event, array, and station statics are calculated and applied during inversion, as described in the Appendix. The reduction in data scatter resulting from the application of statics is small. Data inversion produces structure that reduces data RMS by 54% (from 0.39 s to 0.18 s) for 79% of data variance. The percent residual reduction is comparable to those found in other western U.S. studies (Humphreys and Dueker, 1994a). The travel times that remain unexplained may result from structure of a smaller scale than can be resolved, and may also be affected by a heterogeneous anisotropy structure beneath the area (Schutt and Humphreys, 2001). The residuals are much greater than picking error, and they are not a consequence of the relatively sparse station spacing, which would result in a better fit to the data.

As described below, a range of structures explains the data equally well. All of these structures are similar in map-view to the delay pattern seen in Figure 2D, but differ in the depth distribution of structure. Figure 3 shows a tomographic model for upper mantle *V*p structure produced under the constraint that as little structure as possible lies beneath 200 km. The profile lines shown on the bottom panel of Figure 3 indicate the profiles through this structure shown in Figure 4. This is our best overall representation of the actual structure based on testing explained below. Large-magnitude

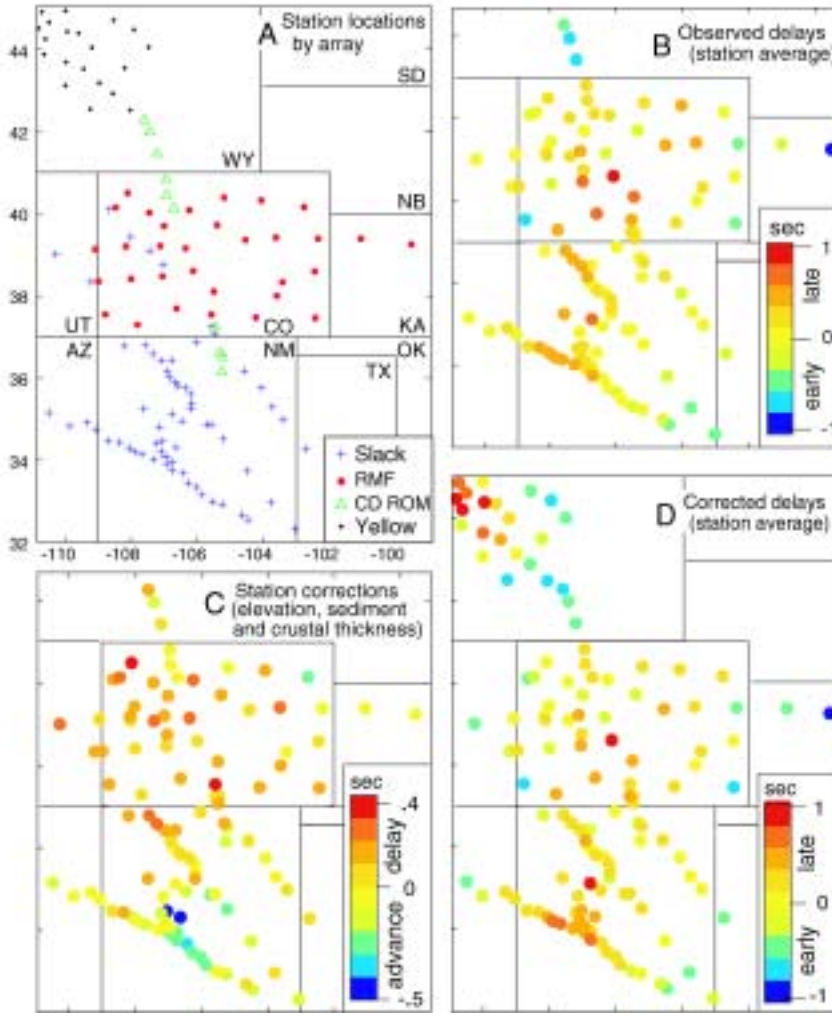

FIG. 2. Travel-time information for seismic stations. A. Station locations for the three seismic arrays used in our study and the Wyoming portion of the Yellowstone array. B. Average observed travel time delay to each station. C. Station travel time corrections for the combined effect of elevation, sediment, and crustal thickness. D. Average corrected travel times to each station. The corrected travel times are the data used in tomographic inversion. These times represent the effects of variations in upper mantle velocity, with early arrivals having propagated through relatively fast upper mantle. The basic pattern of mantle structure is easily seen; regions of slow wave propagation lie beneath central New Mexico, central Colorado, and Yellowstone (in northwest Wyoming) and faster regions lie to the east and west, and in central Wyoming. Arrivals to the easternmost station are very early, indicating the presence of a large thickness of fast upper mantle below this site.

variation in *V*p is seen, and imaged structures define low-velocity volumes beneath the Rocky Mountains in Colorado and beneath the Jemez volcanic trend in New Mexico, and high-velocity volumes beneath the Great Plains (i.e., beneath eastern New Mexico and western Kansas and Texas) and the Colorado Plateau. Correlations between imaged mantle structure and regions defined geologically are strong. Visually, imaged structures vary in wavelength from ~100 km to 800 km. The longer wavelength structures define a north-trending volume of low-velocity mantle beneath the southern Rocky Mountains tec-

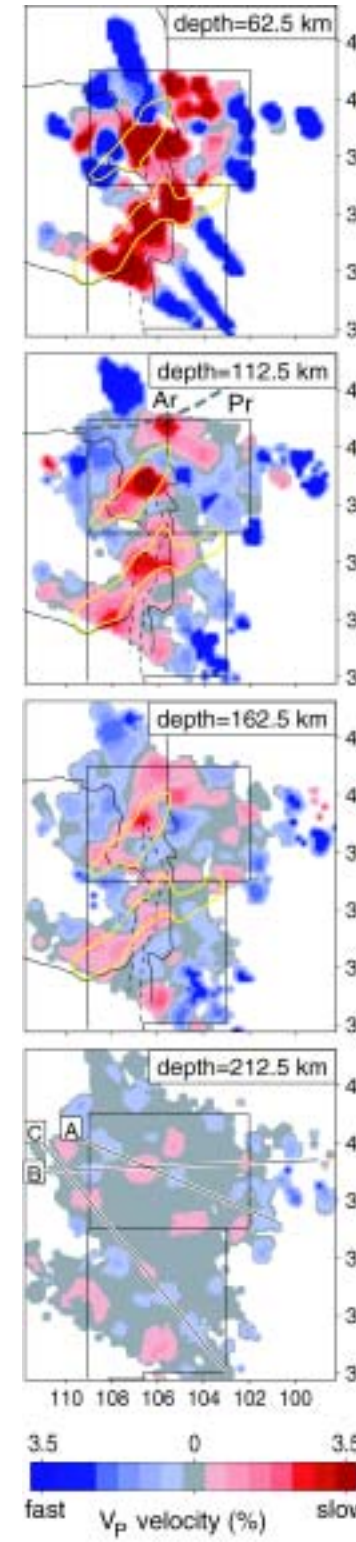

FIG. 3. Tomographic image of P-wave velocity structure. Prominent volumes of mantle with slow propagation speed are imaged beneath the Jemez volcanic trend and beneath west-central Colorado. The profile lines in the bottom panel locate three profiles shown in Figure 4; the bottom left three panels in Figure 4 are profiles through this image. Lines on the top three panels show the main province boundaries. Color symbols: solid black = Colorado Plateau to the west and Laramide front to the east (Miller et al., 1992); the Rocky Mountains lie between these two lines; dashed black line = Rio Grande Rift, a young rift that trends north through the Rocky Mountains; solid yellow = Colorado Mineral Belt to the north (Miller et al., 1992), which is largely of Laramide age, and Jemez volcanic field to the south (Christiansen and Yeats, 1992), which is younger than the Rocky Mountains and Rio Grande Rift. The suture between Archean (Ar) and Proterozoic (Pr) crust is shown with a dashed grey line on the second panel down.

tonic province. Long-wavelength *V*p structure generally correlates with the S-wave velocity structure imaged by Lee and Grand (1996). But, because the *V*p data set permits resolution of smaller-scale upper mantle structures, transitions between low and high velocity to the east and west of the Rocky Mountains are more distinct and detailed in form. On a scale of 200–400 km, an overall NE-oriented grain is observed, which includes the Jemez volcanic trend and the Colorado Mineral Belt. These trends largely are comprised of smaller, more equidimensional bodies. Features of wavelengths shorter than ~100 km may exist, but remain poorly resolved. Before interpreting these results, we examine resolution of the tomographic image.

Resolution of imaged structures

As is typical of teleseismic studies, horizontal resolution is good, and uncertainty is greatest in the vertical direction. The relatively poor depth resolution tends to create vertically blurred structures of diminished *V*p amplitude. Because of these problems, much of our effort has been directed toward understanding the possible depth limits of structure.

Two distinct aspects of data quality affect inversion quality: errors in estimating arrival times and quality of the ray set (the number of rays and their geometric diversity). Timing errors usually pose no serious problem because travel times are picked with relatively high accuracy, and these errors tend to be symmetrical in their distribution. The negligible effects of timing errors are confirmed by adding random noise to the travel-time data at twice the magnitude of the estimated picking errors, and

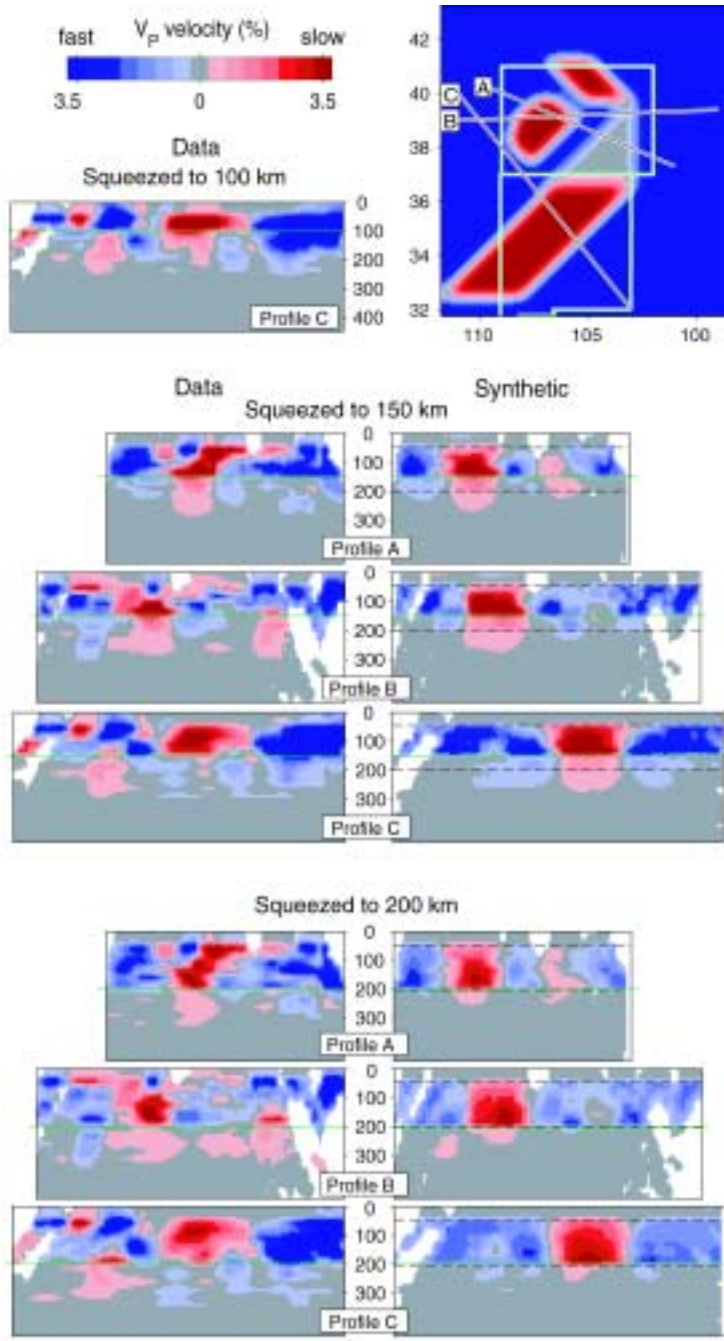

FIG. 4. Profiles of P-wave velocity structure resulting from inversions of real data (left column) and synthetic delays (right column) squeezed within various depth ranges. Map view of the synthetic structure is shown in the upper right panel. The magnitude of the synthetic structure is ±5% in velocity, and it extends from 50–200 km in depth (shown with dashed lines in profiles). Inversion is forced ("squeezed") to be within the depth range from 50 km to the depth indicated with green lines, and then this constraint is relaxed. Only if information exists for significant amounts of structure to greater depth will structure be imaged below the green line.

noticing no significant effects on the modeled structure.

With our ray set, as is typical in teleseismic tomography, imperfect imaging is almost entirely the result of having insufficient ray coverage to resolve the structures of interest; that is, the model typically has degrees of freedom that the available data cannot resolve. As a result, a suite of models exists that account for the data equally well. The practical effects of this are: (1) additional constraints must be provided on the inversion in order to obtain a single model as a representative image; and (2) without all acceptable models available for investigation, it is difficult to know just what model features are required by the data and what class of possible features are inconsistent with the data.

The effects of imperfect resolution often are examined with tests in which the actual ray set is traced through a test structure (e.g., a single anomalous block or a checkerboard pattern) to generate a set of hypothetical delays, and these delays are inverted as data. These tests are useful in demonstrating relative differences in resolving power, but their use is limited because each test represents the investigation of but a single, preconceived structure. It is especially difficult to use such tests to argue that other classes of acceptable structure do not exist.

In this paper, we are especially interested in understanding how deep the structure must extend, i.e., we would like to know what is the shallowest structure that can explain the travel-time delays. "Squeezing" experiments are used to test the need for structure beneath specified depths (Saltzer and Humphreys, 1997). The general principle is to hypothesis test for the acceptability of a depth constraint by testing if any information in the data is significantly in contradiction with the constraint. For example, data can be inverted in a normal fashion except that structure is constrained to the upper 200 km; then this constraint is relaxed and inversion is continued with additional iterations. The first part of this procedure has the effect of finding a least-squares best model in the upper 200 km, and yields the travel-time residuals with respect to this model. With continued iteration following relaxation of the depth constraint, the residuals (which are the times that cannot be explained by the model in the upper 200 km) are inverted for structure in the upper 450 km (in this instance) and the resulting model update is added to the original model. If all pre-relaxation residuals were zero, then no new structure is created after relaxation; if pre-relaxation residuals are random values, then essentially no new structure is created. Only if a better model exists (in a least squares sense) is the post-relaxation structure different from the pre-relaxation structure, and this is recognized by the inclusion of significant structure beneath 200 km. This is a useful test because we are specific about the hypothesis (i.e., does any structure need to exist beneath 200 km to satisfy the data?), but we are unspecific about the structure being tested (i.e., we let the inversion find the best structure above 200 km without need to know in advance about the nature of this structure).

We have inverted our data under a suite of squeezing constraints. We also have created a simple test structure that is similar to the structure imaged with the actual data, calculated delay times using the ray set of the actual data but with travel times determined with the test structure, and inverted these delays to compare the resulting imaging behavior to that of the actual data. These tests are called synthetic tests. Synthetic tests with the depth constraint located deeper than the base of the test structure produce imaged structure that is falsely "streaked" down to the depth of the constraint, and once the constraint is removed there is no tendency to falsely image structure beneath the constraint depth. Figure 4 shows squeeze tests that have been applied to the actual data and to synthetic delays. The synthetic test structure extends from 50 to 200 km and has the lateral distribution shown at the top of Figure 4. Squeezing experiments with depth constraints of 150 and 200 km have been used to test the need for structure beneath these depths. The synthetic tests show that when the depth constraint is placed above the base of the test structure, coherent structure is produced beneath the depth constraint once the constraint is relaxed (e.g., Fig. 4, right column, box labeled "Squeezed to 150 km"). This indicates that the residuals resulting from the depth-constrained inversion do contain information that structure exists at depths greater than the depth constraint.

Inversions of actual data show behavior similar to that of the synthetic test examples. The three profiles chosen in Figure 4 cross structures with the strongest lateral variations, where the best information on depth resolution occurs. Figure 4 shows that with a depth constraint of 150 km, structure deeper than this is produced beneath the prominent low-velocity structure in Colorado (Profiles A and B) once the depth constraint is removed. This down-

ward streaking behavior is remarkably similar to the synthetic test case (shown to the right of the data case in Fig. 4). When the depth constraint is 200 km, only a small amount of streaking occurs, and the streaking is very similar to that seen in the synthetic case (in which the test structure extends to 200 km). We conclude from these tests that lateral variation in structure is required to depths of ~200 km beneath Colorado. The behavior is different for Profile C, which crosses a strong low-velocity structure beneath the Jemez lineament near Jemez caldera. In this case, streaking below the depth constraint is clear only when the squeezing depth is brought up to 100 km (uppermost cross section shown in Fig. 4). From this behavior, we conclude that lateral variation in structure is required to depths slightly greater than 100 km beneath central New Mexico.

Dueker et al. (2001) presented a tomographic image of this region that made use of the same data that we use. However, they did not specifically address the depth extent of imaged features, and their images have deep structures that our tests show are not required by the data.

Discussion

Upper mantle physical state

The main structures imaged are low-velocity volumes that extend approximately vertically to depths of >100 km beneath central New Mexico and to 200 km beneath central Colorado. Locally these features correlate strongly with the areas of major Tertiary magmatism (Christiansen and Yeats, 1992) and high heat flow (Fig. 1), and in a more regional sense they correspond with the Proterozoic portion of the southern Rocky Mountains (Figs. 1 and 3). These associations suggest that the low-velocity mantle is partially molten, that the creation of the low-velocity mantle is related to the Laramide orogeny, and that the Archean lithosphere resisted melting.

To address these issues more thoroughly, we use relations that relate upper mantle seismic velocity to physical state. Possible causes for variation in *V*p are variation in temperature, partial melt content, and composition. The compositional change considered most important is that owing to basalt depletion. Because this has very little effect on seismic velocity (Humphreys and Dueker, 1994b; Jordan, 1979), the low seismic velocities are thought to result from the combined effects of relatively high temperature and the presence of interstitial melt. If attributed solely to temperature, the imaged variations in *V*p imply temperature variations of >500°C (using velocity-temperature scaling relations of Karato, 1993 and assuming average Qp >100). This temperature range probably is greater than reasonable for mantle at 100–150 km depth, which suggests that the very low seismic velocities require the presence of some melt. For instance, ~1% melt could be responsible for the slowest 4% of the imaged velocity range (using the *V*p-melt relations of Hammond and Humphreys, 2000), leaving a more reasonable 2–4% range in *V*p to be attributed to temperature variations. A seismically based argument for interstitial melt is strongest beneath Colorado at depths >150 km, where large variations in seismic velocity occur, yet large temperature variations at these depths are unlikely. Dry melting at these depths is not thought to be possible, implying melting in the presence of fluid such as water (Ito et al., 1999).

Hydration hypothesis

Our hypothesis for regional evolution of the western United States for the period 25–100 Ma, illustrated in Figure 5, is a modification of the flat-slab model proposed by Coney and Reynolds (1977). It includes the hydration of North America by the flat-subducting Farallon slab to provide a subduction-related mechanism for the western U.S. evolution of elevation, tectonism, and magmatism.

Uplift. Regional changes in elevation require negative buoyancy to be broadly distributed beneath the Cretaceous inland seaway prior to the Laramide orogeny, followed by transformation to positive buoyancy beneath the same region during and after the Laramide orogeny. The regional pre-Laramide downwarp is attributed to the suction that supported the shallowing Farallon plate, as discussed by Gurnis (1992) and Mitrovica et al. (1989). Uplift during the Laramide orogeny is attributed to a combination of crustal thickening in the Rocky Mountain area, a regional unloading caused by the younging of the Faralon slab, and lithosphere de-densification owing to the creation of low-density hydrous minerals. Water (and other volatile components) is supplied by dehydration of the Farallon slab and associated sediments, but during flat-slab subduction the creation of hydrous conditions does not lead to mantle melting because the vapor-rich mantle is never exposed to a hot asthenospheric mantle wedge; instead, it is kept cool by its contact with North American lithosphere and the water reacts to create hydrous phases. Although phase relations are

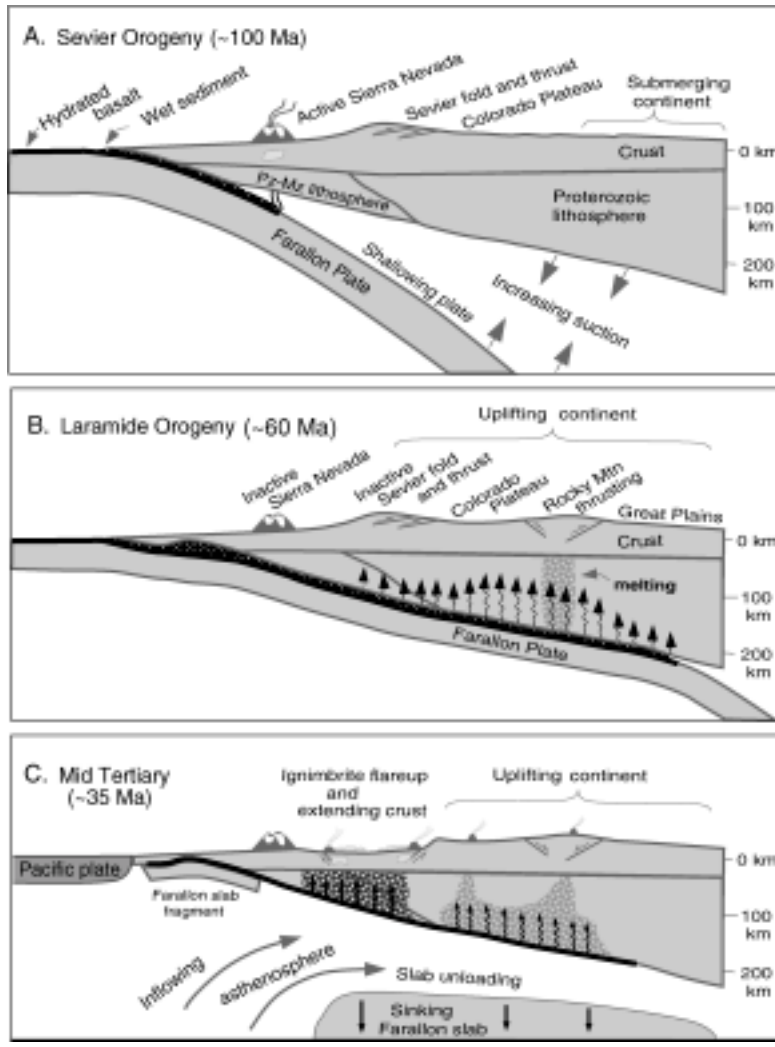

FIG. 5. Hydration hypothesis for Laramide and mid-Tertiary magmatism and tectonism and uplift of the western U.S. Black lines and pattern indicate elements important to lithospheric hydration: hydrated basalt layer (heavy black line), subducting hydrated sediment (black pattern), and fluid ascent (wavy arrows). A. At 100 Ma, Sierra Nevada arc magmatism and Sevier thrusting were active. The continental interior submerges in response to increasing suction (Mitrovica et al., 1989). Paleozoic subsidence of the continental margin following continental rifting at ~600 Ma (Bond et al., 1989) indicates that Phanerozoic mantle cooled onto the base of Proterozoic lithosphere. Lithosphere of the continental interior is shown ~200 km thick, as resolved from teleseismic seismology (Dueker et al., 2001; Fig. 4). B. Rapidly subducting and younging Farallon plate has ascended to make contact with North America, quenching Sierra Nevada magmatism and causing eastward propagation of magmatism into the Rocky Mountains (Coney and Reynolds, 1977), and Rocky Mountain thrusting. Rocky Mountain magmatism and tectonism normally would not be expected because of the thick lithosphere. We invoke basal hydration of western U.S. lithosphere under the increasingly cool conditions resulting from salb contact (Dumitru et al., 1991) to: (1) de-densify western U.S. mantle; (2) cause water-rich melting and early Laramide-age Rocky Mountain magmatism; and (3) weaken the lithosphere beneath the Rocky Mountains, enabling mantle shortening which allowed crustal thrusting. C. Farallon slab removal exposes the hydrated base of North America to asthenosphere, causing widespread magmatism that was especially intense where the lithosphere was relatively fertile and thin (Humphreys, 1995); this now is the Basin and Range. Slab removal unloads North America lithosphere, and lithospheric heating decreases the density of North America, both contributing to uplift.

not well understood under the high-pressure and low-temperature conditions that we are suggesting, serpentine is a candidate water-bearing mineral because it is stable to depths of 150–200 km if kept cooler than ~600°C (Ulmer and Trommsdorff, 1995). Phlogopite also may be created, and even free water is possible under pressure-temperature conditions that may be expected where slab remains in contact with deep lithosphere (Pawley and Holloway, 1993). Geochemical evidence for lithospheric hydration is abundant, including the common occurrence of unusually potassic Tertiary basalts (that require mantle melting involving potassium-rich amphibole or phlogopite), phlogopite-bearing mantle xenoliths within the Tertiary basalts, and hydrous melting of Precambrian mantle lithosphere (Mutschler et al., 1987 and references therein; Lopez and Cameron, 1997).

Additional uplift following the Laramide orogeny probably has several causes. Uplift would result directly from slab removal. Lithospheric warming caused by the increasing temperature at the base of the lithosphere could also contribute to uplift, assuming that the associated cooled asthenosphere is swept away. In areas with significant magmatism, an advective steepening of the geotherm would cause additional lithosphere warming, and basalt removal from the mantle would create a significant compositional buoyancy (Humphreys and Dueker,1994b), although water removed with ascending basalt would eliminate the buoyancy contribution provided by the hydrous phases. It also is possible that the asthenosphere emplaced beneath North America during slab removal was unusually hot and buoyant, as is suggested by the fact that western U.S. low-velocity mantle is the northern part of a pronounced low-velocity anomaly that occupies a large volume of upper mantle beneath the East Pacific Rise (Grand, 1994). Post–middle Miocene uplift (Pederson et al., 2002; Heller et al., 2003), which may be accelerating (Sahagian et al., 2002), and the occurrence of Jemez magmatism only in the last ~10 m.y. (Christiansen and Yeats, 1992), may be related to the anomalous East Pacific Rise upper mantle or perhaps to a deeper source (Heller et al., 2003).

Upper-mantle velocity structure. In the above discussion on upper-mantle physical state, we conclude that the imaged low velocities result from the presence of interstitial melt and from relatively high temperatures. We envision that mantle melting began during the Laramide orogeny, at least beneath the Colorado Mineral Belt. This melting must have been a consequence of activity at the base of the lithosphere, i.e., at depths of ~200 km, because partial melting initiated by activity within the interior of stable mantle lithosphere does not seem possible. We attribute melting to a reduction in the solidus temperature that resulted from hydration of the North America lithosphere by water transported beneath the continent by the subducting Farallon slab.

The gross trend of the low-velocity anomalies define a N-S structure, with higher velocities present to the east and west. More regional imaging supports this view (Fig. 1). Grand (1994) imaged high-velocity mantle to the east, beneath the craton, and Humphreys and Dueker (1994a) found generally higher velocities beneath the Colorado Plateau to the west. The N-S trend of the zone of upper mantle thought to be partially molten suggests that pressure played a role in enhancing water release from the subducting slab as it attained a depth of ~200 km, or that an enhanced transfer of water into the North America lithosphere occurred in the Rocky Mountain area. The higher upper-mantle velocities in central Wyoming that are implied by early arrivals there (Figs. 1A and 2D) may indicate that the Archean upper mantle was relatively undisturbed by the Laramide orogeny. The northeast trend of the smaller-scale low-velocity zones, and their locations beneath what wwre thought to be Proterozoic sutures, suggests that the preferential melting was a consequence of inherited structures (Karlstrom and Humphreys, 1998) such as the composition of the suture zones being more fertile than other portions of the lower lithosphere.

Magmatism. Because lithosphere is think in our study area, Laramide and pre-Laramide magmatism cannot be attributed to normal subduction processes, i.e., to water-induced melting of asthenosphere resulting from water release from a subducting slab at ~100 km. We attribute Laramide-age partial melting to solidus depression resulting from the presence of volatiles at the base of the continental lithosphere (see also James and Sacks, 1999 for a similar discussion applied to the Andes). Presumably, most melting occurred during initial slab contact because prolonged contact would have cooled the base of the lithosphere by conduction. This could explain the eastward propagation of magmatism across the western U.S. prior to and during the Laramide orogeny that, at a given location, was followed by reduced or absent magmatic activity

(Coney and Reynolds, 1977). This behavior includes the Colorado Mineral Belt magmatism, which experienced concentrated activity early in the Laramide orogeny (Mutschler et al., 1987), followed by relative quiescence.

The great intensity of the "ignimbritic flareup" (Coney, 1980) and its N-to-S propagation across what now is the northern Basin and Range, are not expected for simple slab rollback. Rather, we attribute magmatism to unusually high melt productivity where, upon slab removal, the asthenosphere made contact with hydrated lithosphere that was relatively thin and fertile. As suggested above, northern Basin and Range basal lithosphere would be more fertile because it is cooled asthenosphere accreted during lower Paleozoic margin subsidence, and hence it was not involved in prior magmatism (see Fig. 5). Basin and Range lithosphere was thinner because it was younger and, for the southern Basin and Range, it was recently heated by arc magmatism. The importance of being thin is that advective heating by magma invasion of the lithosphere is required for rapid lithospheric heating (conduction being a slow process), and asthenospheric melt (see Ito et al., 1999, for discussion of vapor-induced melting at depths below the dry solidus), and hence more melt would be available beneath thinner lithosphere.

Karlstrom and Humphreys (1998) suggested that the Jemez trend is a relatively fertile Precambrian suture, similar to the Colorado Mineral Belt. Although this might account for its relatively high magmatic productivity, it does not explain why magmatism here has occurred only in the last 10 m.y. (Christiansen and Yeats, 1992). Dueker et al. (2001) and Heller et al. (2003) suggested that a recent influx of warm mantle may be responsible for the magmatism and young uplift of the region.

Contraction. In our model, Rocky Mountain contraction occurred as the Colorado Plateau was driven northeast (Hamilton, 1989) by basal tractions derived from Farallon slab contact with North America (Bird, 1988). Mantle strain was concentrated in lithosphere that was thermally weakened by magmatic ascent. The distribution of partially molten mantle during the Laramide orogeny is thought to be similar to that imaged beneath the Rocky Mountains (Fig. 3). Crustal strain arranged itself on available pre-existing zones of favorably oriented weakness located near volumes of concentrated mantle strain. The NE-SW orientation of contraction and the northerly orientation of the large-scale zone of weakened mantle would require oblique shortening across this mantle. With upper-crustal strain occurring approximately above the mantle strain, decoupling across the lower crust for distances greater than ~100 km is not required in our study area. This may not hold true for the Rocky Mountains in Wyoming, however, where most of the mantle lithosphere probably remained strong through the Laramide orogeny. Our seismic data, where present (see Fig. 2), show no evidence for a weakened lithosphere except in the Yellowstone area, and Laramide-age magmatism was strongly limited to the Black Hills and Absaroka Mountains near the northeastern and northwestern corners of Wyoming, respectively. Similarly, heat flow (Fig. 1B) indicates little crustal heating away from Yellowstone. These observations suggest that the Wyoming lithosphere may have remained strong over distances of several hundred kilometers, and hence that lithospheric shortening may have occurred at locations quite distant from the areas of crustal shortening (which are distributed across most of Wyoming).

Conclusions

General agreement exists that Rocky Mountain and Basin and Range tectonism and magmatism are consequences of Late Cretaceous and Eocene compression associated with rapid subduction, followed by post-orogenic collapse. Disagreement and confusion exist regarding the nature of the slab-continent coupling that drove contraction, with edge forces (e.g., Maxson and Tickoff, 1996), potential energy of Sevier welt (e.g., Livaccari, 1991), and basal tractions arising from flat-slab contact with a broad area of lithosphere (e.g., Bird, 1988) having been suggested. Explanations for the fundamental cause of western U.S. Tertiary magmatism include a migrating volcanic arc (e.g., Coney and Reynolds, 1977), and extension-driven lithospheric thinning and plume impact (Fitton et al., 1991; Parson et al., 1994). Cause for post-orogenic collapse has focused on a transition in boundary condition at the widening transform margin (e.g., Atwater, 1970), gravitational potential energy (e.g., Sonder and Jones., 1999), and lithospheric weakening owing to magmatic heating (e.g., Armstrong and Ward, 1991).

The growing body of evidence for relatively thick North America lithosphere (Sierra Nevada, Farmer et al.,2002; Great Basin, Wang et al., 2002; Colorado Plateau, Smith, 2000; and Rocky Mountains, Dueker et al., 2001; this study) is challenging most

of these explanations, and the rather disconnected nature of the explanations for activity in the different western U.S. provinces is disconcerting to those who view the Cordilleran orogen as a coherent process.

We integrate several published suggestions—most importantly that of the flat slab—with constraints provided by the imaged velocity structure beneath the Rocky Mountains and adjoining Great Plains and Colorado Plateau to hypothesize that water provided by the subducting Farallon slab hydrated the base of a thick western U.S. lithosphere, providing regional buoyancy, limited magmatism, and lithospheric weakening localized to the Rocky Mountain area; Farallon slab removal allowed asthenospheric contact, which initiated vigorous magmatism, especially where hydrated lithosphere was relatively fertile or thin, and contributed additional buoyancy. Relatively great potential energy of the continental interior and the evolving western-margin boundary condition (from a compressional to an extensional subduction setting; Jerrard, 1986) and then to trans-tensional transform (Atwater and Stock, 1998) allowed extension to occur where magmatism had greatly weakened the lithosphere.

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Appendix: Seismic Data and Analysis

Seismic data

The 1992 Rocky Mountain Front (RMF) and the 1999 Continental Dynamics Rocky Mountain (CDROM)deployments used broadband three-component (CMG3-ESP and STS2) seismometers made available by the program for Array Seismic Studies of the Continental Lithosphere (PASSCAL). Seismometers were distributed over the area of Colorado and portions of adjoining states, as shown in Figure 2A. About 4000 teleseismic P-wave arrivals were collected, from which delay times were determined. Most back azimuths are represented, although NW and SE back azimuths are most common.

Routine data processing of the RMF data for travel-time residuals involved deconvolution of the instrument-11-responses and phaseless 0.5-2 Hz bandpass filtering. Waveforms for each event were time shifted using theIASPEI91 Earth model, and relative deviations from these predicted arrival times were sought. These data exhibit unusually large variations in waveform across the array. As a result, cross-correlation methods were not very effective at determining relative time lags precisely. Instead, we estimated travel-time delays by visual alignment on the first prominent phase. Picking uncertainty is estimated to be ~0.15 s. The resulting delays then had their mean subtracted on an event-by-event basis as a first-order correction for source mislocation and mantle heterogeneities outside the modeled region. In this form, the travel-time data for each event exhibit up to 3 seconds of travel time variation.

The broadband residuals have a root-mean-squared (RMS) value of 0.43 seconds, which is larger than most other western U.S. data sets (Humphreys and Dueker, 1994a). Figure 2B shows the average delay to each station.

Delays were then corrected for elevation, sediment thickness, and crustal thickness variations, obtaining a travel time residual for each ray (Fig. 2D). Sediment thickness estimates are from Woodward (1988), and crustal thicknesses from the receiver function analysis determined at each RMF site and from refraction studies (Keller et al., 1998; Dueker et al., 2001; Rumple, pers. commun., 2002). Crustal corrections for each site are shown in Figure 2C. The crust-corrected residual RmS of the data are essentially unchanged by the application of station corrections, implying that sediment and crustal thickness variations are not a cause for the large travel-time variance of the data. This indicates that the mantle beneath the study area is very heterogeneous.

Inversion

We then seek the blocks' slowness perturbations δs_b from an assumed depth-varying (one-dimensional) velocity structure by solving

$$\delta t_r = \sum_b l_{br}\delta s_b + \delta t_e + \delta t_s + \delta t_a ,$$

where δt_r is the travel time residual for the r-th ray. The length of the r-th ray in the b-th block is l_{br}, and δt_e, δt_s, and δt_a are event, station, and array statics, respectively. A least squares approach is used to solve (1). Several similar algorithms are available to obtain a least squares solution, and because the various algorithms weight the data and model blocks differently, the solutions differ. Two inversion algorithms are commonly used: SIRT(a Jacoby iteration method; Humphreys and Clayton, 1988) and LSQR (a conjugate gradient method; van der Sluis and van der Vorst, 1987). We use the SIRT algorithm, employing 10 iterations and incorporating a nearest-neighbor model covariance to provide a specified degree of model smoothness. In comparative tests of the two inversion algorithms using teleseismic data, no significant differences were found (Saltzer and Humphreys, 1997). The slowness values of blocks with three or fewer hit counts are downweighted in the inversion, which results in a model parameter weighting that is similar to LSQR in the poorly hit regions (Nolet, 1993).

The assumed depth-dependent velocity structure is used to guide the rays with Snell's Law, and because teleseismic rays are steeply inclined, ray position is not sensitive to reasonable variations in the assumed velocity structure. Tests of teleseismic data show that updated ray tracing through the imaged structure produces no visible difference (Saltzer and Humphreys, 1997), even when these data have encountered strong lateral velocity gradients in the upper few hundred kilometers of the Earth. Event and array statics were calculated as described in Humphreys and Dueker (1994a).

Event and array statics were calculated as described in Humphreys and Dueker (1994a). Event statics effectively adjust the average delay of each event so that the delays for each event are as consistent with the entire data set as possible (in a least

squares sense). Array statics are analogous. These static time shifts are higher-order estimates to the event arrival time than were estimated (and applied) by subtracting the mean. Station statics are meant to correct for the travel time effect of structure close to each station. However, it is not possible to distinguish clearly between the effects of local structure and deeper structure. For our stations, we have good estimates of crustal delays from Dueker et al. (2001), Keller et al. (1998), and Rumple (pers. commun., 2002). We take a conservative approach to applying station statics, similar to Humphreys and Dueker (1994a) and Saltzer and Humphreys (1997). After applying half the SIRT iterations—at a point where imaged structure accounts for delays well and residuals are relatively small—we calculate and apply each station static at 60% of the station's average residual.

In our inversions, we impose a nearest-neighbor model smoothness and find the model of minimum model energy; the inclusion of these constraints results in a unique model. Minimum model energy is implicit to SIRT. Among all models that best explain the data and satisfy any imposed constraints, inversion finds the model of least energy. The other commonly used constraint, one of maximum smoothness (minimum Laplacian [e.g., VanDecar and Snieder, 1994]) produces similar images, but structures become somewhat less localized. SIRT converges on a solution iteratively, so any smoothing relation enforced between iterations becomes a constraint in the normal sense. The application of a smoothness constraint imposes the specified covariance between nearby model parameters (Meyerholtz et al., 1989). We impose nearest horizontal neighbor model covariance by using a nearest horizontal neighbor smoothing function between iterations. No vertical smoothness is imposed because teleseismic rays inherently smooth structure in this direction.