Continental uplift through crustal hydration

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ABSTRACT

Isostatic surface uplift of large continental regions lacking deformation remains largely unexplained. Evidence from the eastern parts of the Cordilleran orogen in the western United States suggests that increased buoyancy in the lower crust supports the elevations of the High Plains and Wyoming craton. We suggest that hydration of the lower crust associated with the Laramide orogeny produced surface uplift by replacing dense mineral phases such as garnet with less dense phases such as amphibole and mica. Seismic and petrologic evidence from Wyoming and Montana is consistent with such changes. Comparable hydration in the Colorado Plateau is dated to the early Tertiary. Beyond establishing a newly recognized mechanism for broad continental uplift, such hydration suggests that interactions of subduction-derived fluids and the lithosphere can be more profound than previously envisioned.

INTRODUCTION

The diversity of tectonic and magmatic events in the western United States has spawned many explanations for the modern high elevation of the region in whole or in part, including dynamic topography, crustal thickening, magmatic underplating, mantle depletion, lower crustal flow, vertically non-uniform extension, delamination and/or foundering, mantle hydration, and lithospheric erosion. However, the Cenozoic rise of the western Great Plains (High Plains) and parts of the Wyoming craton to as much as 2 km above sea level has proven problematic. In general, seismic models, xenolith studies, and gravity residuals all support the concept that variations in continental crustal structure produce topographic gradients, yet insufficient crustal shortening or magmatism occurred that could plausibly be linked to all of the observed Cenozoic elevation variations. We propose that alteration of the lower crust in this region systematically replaced widespread dense mineral phases like garnet, pyroxene, and plagioclase with amphibole, mica, and other less dense phases and contributed to the observed increases in surface elevation.

Prior to the Laramide orogeny ~70 m.y. ago, areas east of the Sevier thrust front in the western United States lay near sea level (cf. Roberts and Kirschbaum, 1995). Most of these areas had been low lying for more than 150 m.y. Beginning in the early Cenozoic, large portions of the continental interior were affected by crustal shortening, magmatism, and thinning of mantle lithosphere, providing a wealth of means of elevating these areas. However, evidence for such events is absent in the High Plains, and they were probably insufficient in the Wyoming craton to produce the observed change in elevation. Given the scale of uplift across the Cordillera of the western United States (e.g., Eaton, 1987, 2008), it is likely that whatever process elevated the High Plains and Wyoming craton also impacted other areas.

WYOMING AND MONTANA TRANSECT

The deep crust can be best examined along a line from Montana across Wyoming and Colorado along which elevations increase by over 1 km and where xenolith suites are contained in Phanerozoic igneous rocks, the Deep Probe and Continental Dynamics–Rocky Mountain (CD-ROM) active source experiments were conducted, and upper mantle wave speeds are high (Fig. 1; e.g., Obrebski et al., 2011; Shen et al., 2013). Crustal xenoliths from several Tertiary magmatic centers along this profile vary from intermediate to mafic in composition and suggest that high-grade metamorphic lower crust formed via a combination of Archean to Mesoproterozoic orogenic and magmatic underplating and/or intraplating events (Barnhart et al., 2012).

Lower crustal xenoliths (Fig. 1) appear progressively more hydrated from north to south. Xenoliths near the Canadian border are well-preserved garnet-rich granulites (Barnhart et al., 2012). Xenoliths from 35 to 40 km depths in the ca. 50 Ma (inferred) Homestead kimberlite in central Montana (Fig. 1) are mafic garnet granulites (Mahan et al., 2012). One of the two studied xenoliths from this kimberlite (sample HS-1) displays extensive garnet, pyroxene, and feldspar retrogression to a secondary alteration assemblage of chlorite, albite, and calcite (Fig. 2A). This assemblage does not exclusively occur at depth, but the textures and extent of reaction suggest alteration in the deep crust prior to eruption. Using Hacker and Abers's (2004) method, at 500 °C and 1.0 GPa, calculated bulk densities of the unaltered and altered assemblages are 3.19 Mg/m^3 and 3.05 Mg/m^3, respectively. Farther south, hornblende–two pyroxene granulite xenoliths from the Quaternary Leucite Hills of Wyoming (Fig. 1) lack garnet, consistent with destabilization during hydration. Some high-temperature hydration of these rocks occurred in the late Archean (Farmer et al., 2005), but this does not preclude Cenozoic hydration as well.

The age of lower crustal alteration along this transect is constrained only to postdate ca. 1.7 Ga peak metamorphism. Carlson et al. (2004) documented extensive metasomatism in mantle xenoliths from Homestead, and argued for a Cenozoic age based on Sm-Nd tie lines for garnet and clinopyroxene, as did Facer et al. (2009) for mantle xenoliths from the Bearpaw Mountains farther north. The absence of significant metasomatism in the garnet-rich Paleozoic Stateline xenoliths (Farmer et al., 2005) and the absence of a lower crust with high P-wave speed (v_p) in that area today suggest a change in crustal mineralogy after the Paleozoic. In situ Th/Pb dates from monazite in a deep crustal xenolith from the Colorado Plateau (Fig. 2B) indicate a prolonged retrograde hydration event, from 91 to 58 Ma (Butcher, 2013), for sample RM-21 previously thought to have been altered solely in the Precambrian (Selverstone et al., 1999). Release of fluids from the subducting Laramide-age slab across the entire width of the Great Plains has been suggested both from the distribution of Cretaceous kimberlites and from numerical models of metamorphic reactions in this slab (Currie and Beaumont, 2011). Thus Laramide hydration of the lower crust is highly plausible.

Retrogression should lower both the density and v_p of the deep crust. The Homestead xenolith's hydrated mineralogy has a density ~5% (0.15 Mg/m^3) lower than the peak mineralogy of the same xenolith (Fig. 3). v_p also declines, though only by ~0.1 km/s for this particular specimen. Density and v_p co-vary quite closely across the full suite of xenoliths available (Fig. 3) with minimal suggestion of a substantial contribution from protolith variations (Fig. DR1 in the GSA Data Repository[1]). Thus we expect that any systematic increase in density-decreasing hydration reactions from north to south in this transect will be accompanied by a decline in v_p.

Although upper crustal and upper mantle v_p varies little along this line north of 41°N, they do vary within a prominent ~20-km-thick lower crustal layer, decreasing from ~7.7 km/s in Canada to ~7.2 km/s in central Wyoming to

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[1]GSA Data Repository item 2015128, supplemental figure comparing total FeO to modal garnet of xenoliths, is available online at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

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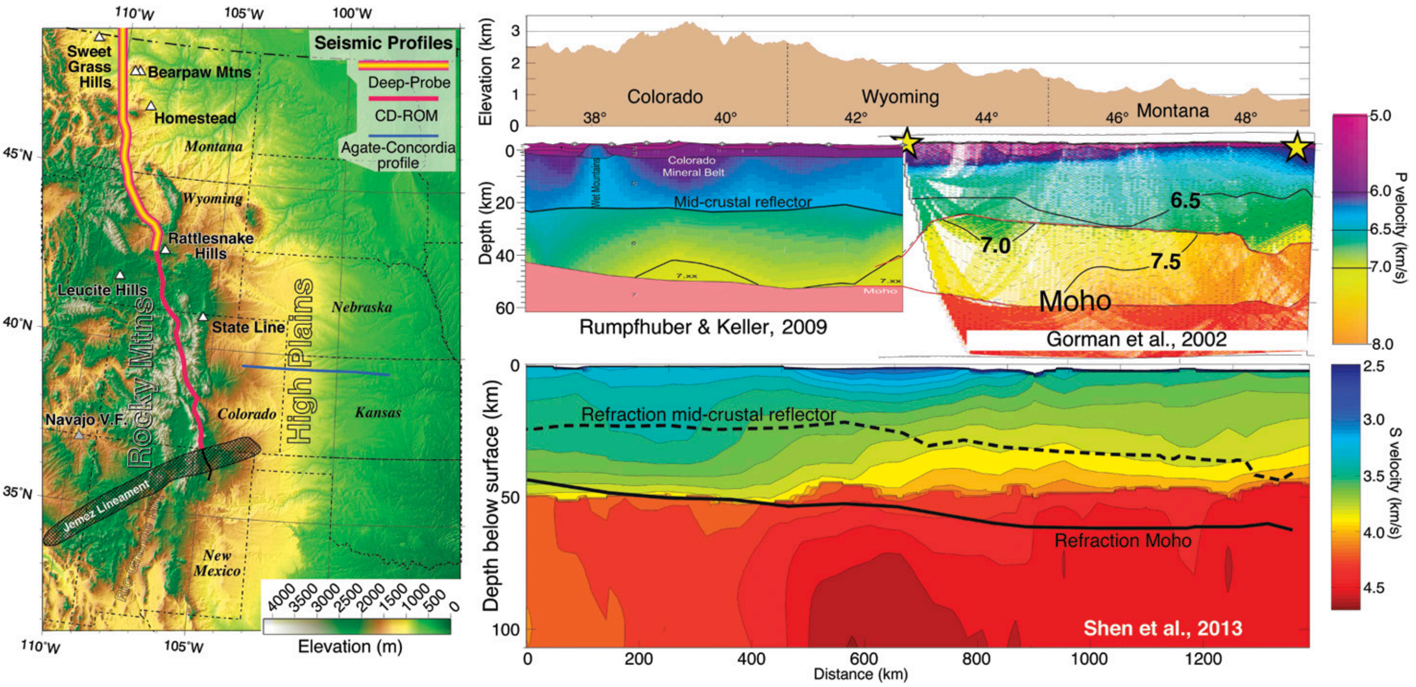

Figure 1. Left: Shaded topographic map of High Plains and Southern Rocky Mountains (United States) with locations of active source seismic profiles; Agate-Concordia profile was interpreted by Steeples and Miller (1989). Xenolith localities shown as triangles (white are used in Fig. 3, gray is locality with dated Laramide crustal hydration [Butcher, 2013]). Right: P-wave speed (v_p) profiles from Rumpfhuber and Keller (2009) and Gorman et al. (2002) and shear wave model of Shen et al. (2013; as made available at http://ciei.colorado.edu/Models/) along same profile. CD-ROM—Continental Dynamics–Rocky Mountain; V.F.—volcanic field.

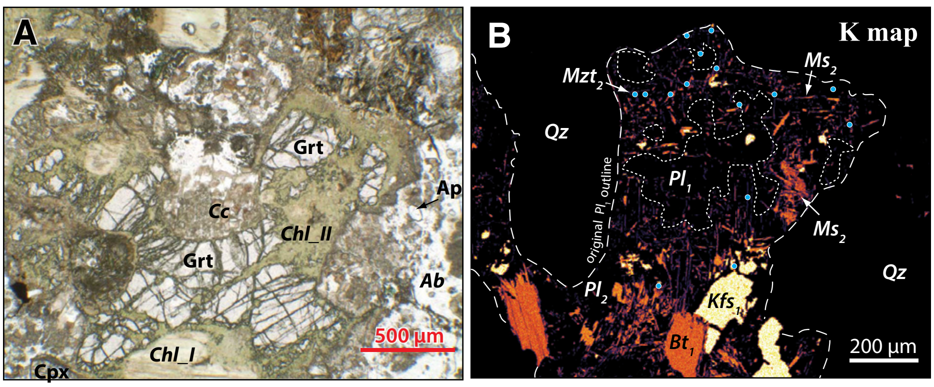

Figure 2. A: Microscopic plane-polarized image of altered Homestead xenolith. Grt—garnet; Cpx—clinopyroxene; Chl—chlorite; Ab—albite; Cc—calcite; Ap—apatite. B: Potassium X-ray map from Red Mesa xenolith RM-21 (Butcher, 2013) (Colorado Plateau); outline of early metamorphic plagioclase (Pl_1) shown with white dashed lines. This feldspar is partially replaced by albite (Pl_2) + phengite (Ms_2) + monazite (Mzt_2, blue dots). Other abbreviations: Qz—quartz; Kfs_1—potassium feldspar; Bt_1—biotite.

<7.0 km/s in southernmost Wyoming (Gorman et al., 2002). The high-v_p layer becomes indistinct from the rest of the crust at the southern end of the Wyoming craton (Rumpfhuber and Keller, 2009). This decline is well within the variation seen in lower crustal xenoliths (Fig. 3) and potentially reflects a decrease in layer density from 3.46 Mg/m^3 to 3.17 Mg/m^3. Assuming isostatic equilibrium, a decrease in density $\Delta\rho$ of a layer of thickness h will produce a change in elevation $\Delta\varepsilon = \Delta\rho h/\rho_m$, where ρ_m is the density of compensating asthenosphere (Fig. 3B). A reduction in density of 0.29 Mg/m^3 in the 20-km-thick lower crust should produce surface uplift of ~2 km, larger than the variation in mean elevation along this profile (Fig. 1), suggesting that some of the variation predates the Cenozoic. Projecting seismic observations onto the expected isostatic response to dedensification indicates that such density changes are needed only in a 10-km-thick layer (or, alternatively, that half the observed variation in the 20-km-thick layer need be Cenozoic) (Fig. 3B). Thus hydrous alteration of the lower crust, consistent with seismic constraints and xenolith observations, is capable of producing more than the modern elevation variation along this transect. This uplift is independent of any topography supported by crustal shortening of ~43–120 km in Wyoming (Bird, 1998; Chapin and Cather, 1983), equivalent to ~8%–30% crustal thickening over the area affected. The latter could contribute elevation increases of ~400–1700 m over the Cenozoic depending on density contrasts at the Moho and whether underlying dense mantle lithosphere was thickened; lower values would be expected if crustal décollements carry shortening out of the Laramide orogen (Erslev, 2005).

HIGH PLAINS

The High Plains lies east of the early Tertiary reverse faults that created the Southern Rocky Mountains. Previous gravity, refraction, and receiver function studies indicated that crustal thickness or density within the Great Plains decrease as elevation declines from ~2 km in the west to near sea level ~1000 km to the east (Bird, 1984; Sheehan et al., 1995; Steeples and Miller, 1989). The absence of thrust faults and widespread Cenozoic igneous rocks in the High Plains precludes whole-crustal tectonic or magmatic crustal thickening outside the Jemez lineament (Fig. 1) (e.g., Nereson et al., 2013). Rebound of crust thickened by sedimentation during late Mesozoic dynamic subsidence (Mitrovica et al., 1989) might contribute to modern elevations in some areas but not in the southern High Plains, where post-Triassic strata are absent. Elevation support from sublithospheric stress is inconsis-

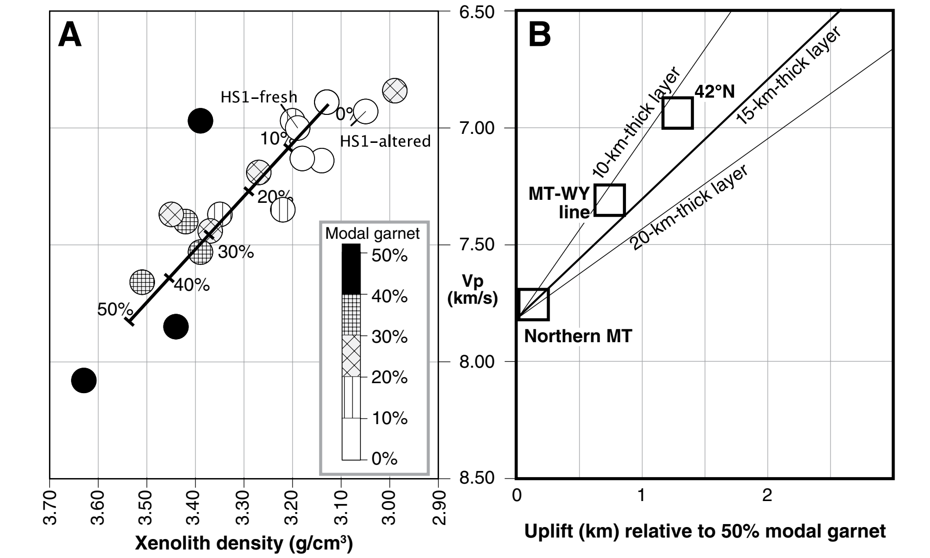

Figure 3. A: Scatter plot of P-wave speed (v_p) and densities from xenoliths in Wyoming-Montana corridor (western USA) (Fig. 1; data from Barnhart et al., 2012; Farmer et al., 2005; Mahan et al., 2012). Values calculated using relations of Hacker and Abers (2004) at 500 °C and calculated pressures (range of 0.8–1.5 GPa). Points shaded by mode of garnet, showing that the bulk of variation is due to garnet fraction. Labeled line is regression of modal percent garnet onto density and v_p. Sample HS1 as in text. B: Uplift as function of mean v_p assuming the regression to density at left if changes are in a layer 10, 15, or 20 km thick. Boxes are from lower crustal v_p profiles (Fig. 1) with elevations lowered by ~800 m. MT—Montana; WY—Wyoming.

tent with variations in gravity anomalies from the Rockies across the plains (Eaton, 1987; Sheehan et al., 1995) and the close association of changes in crustal thickness with topography (Steeples and Miller, 1989). Emplacement of lower crust by pervasive shear during the Laramide (Bird, 1984) predicts wholesale removal of mantle lithosphere from regions where isotopic evidence from Cenozoic volcanic rocks suggests that it has survived (Bird, 1988; Livaccari and Perry, 1993). Pressure-driven lower crustal flow has been rejected because of the scale of uplift and the implausibility of sufficiently low lower crustal viscosity (Bird, 1984). Removal of deep mantle lithosphere (e.g., Spencer and Chase, 1989) or its hydration (Humphreys et al., 2003) do not account for suggestions of thicker crust in the western plains or the more abrupt change in mantle wavespeeds in some seismic models (e.g., Obrebski et al., 2011) than in topography.

Ideally we would directly interpret seismic models derived from the EarthScope project to test our hydration hypothesis in the plains, but comparison of the active source models with a profile through the passive seismic (surface wave and receiver function) model of Shen et al. (2013) reveals differences critical to determining the characteristics of a lower crustal layer (Fig. 1). Although the Shen et al. model yields the same Moho depth as the active source profiles south of ~41°N (where a high-v_p crust is absent) and near the north end (where unaltered high-v_p crust is present), the active and passive seismic interfaces diverge between these areas. Within this same area, the Shen et al. model has a shallow, positive vertical wave-speed gradient in the upper mantle. This cannot reflect a thermal gradient in a lithologically uniform mantle because it would demand hotter sub-Moho mantle over colder, deeper mantle. Instead this gradient can be either a lithological gradient in the upper mantle or an artifact of the parameterization, both of which are consistent with our hypothesis. A lithologic variation would likely be mixed mafic and ultramafic rocks in the uppermost seismic mantle but would require an erroneous active source interpretation. If the parameterization is too limited by assuming a single discontinuity at Moho depth ranges, then a pair of discontinuities would likely place the single model discontinuity between the pair and generate a wave-speed gradient in the uppermost mantle to satisfy surface wave observations (Fig. 4 inset). Unfortunately a direct analysis of the Shen et al. model for the origin of topographic support (e.g., Levandowski et al., 2014) will mistakenly assign support from crustal hydration to the mantle.

Although further work is needed to adjust the passive seismic technique for possible high-wave-speed layers, we suggest that the appearance of fairly high (>0.005 km/s/km) gradients in the uppermost mantle (above 80 km depth) in the Shen et al. (2013) model may reflect the presence of partially hydrated lower crust similar to that inferred from the active source profiles. Such areas are darker in Figure 4 and include the Colorado Plateau, the Wyoming craton, and much of the High Plains. These gradients in the Colorado Plateau and the Wyoming craton, two of the highest areas least magmatically altered in the Cenozoic, support the hypothesis that fluid-induced phase changes within the lower crust helped produce the modern elevation of these regions. Lower-amplitude but similar gradients in the High Plains outside of the Jemez lineament (Fig. 1) suggest that lesser amounts of hydration might be responsible for uplift of these areas.

CONCLUSIONS

Xenolith and seismic observations in Wyoming and Montana are consistent with metasomatism of the lower crust; density variations in this lower crust can support topographic differences in excess of 2 km. Characteristics in passive seismic models from this area also occur in the Colorado Plateau, Wyoming craton, and High Plains, suggesting that similarly high-wave-speed lower crust might be present. Geochronology on a Colorado Plateau crustal xenolith suggests that hydration occurred in the Laramide. Dehydration of subducting lithosphere is thought to have extended far inland, providing fluids capable of hydrating the lower crust and uplifting broad regions otherwise unaffected by tectonism.

Cryptic crustal hydration would be reversing the long-term densification of lower crust in orogenic belts caused by isobaric growth of garnet (Fischer, 2002), a predictable consequence of lower crustal orogenesis (Williams et al., 2014). Textures predicted by isobaric densification are found in xenoliths from the Sweet Grass Hills (Mahan et al., 2012) and Stateline (Farmer et al., 2005) localities (Fig. 1). This might be a precondition for dedensification through hydra-

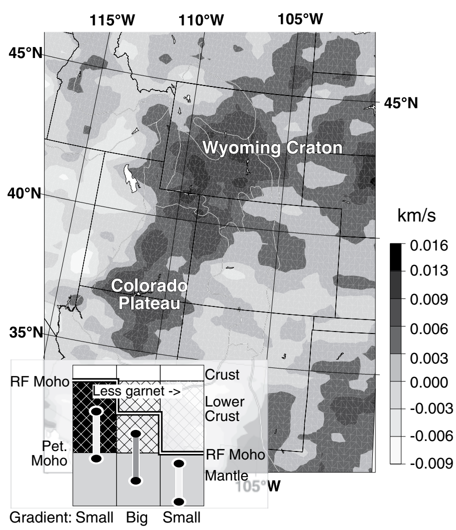

Figure 4. Gradient in shear wave speed of upper mantle in Shen et al. (2013) results from 8 km to 40 km below Moho. Darker areas are unusually large gradients requiring compositional variations in upper mantle or model errors which we infer reflect presence of partially hydrated lower crust. Inset shows that large vertical gradients are expected where the Shen et al. (2013) Moho (RF Moho, white outlined line) is within the lower crust and above the petrologic Moho (Pet. Moho).

tion because that requires considerable initial amounts of lower crustal garnet. Dedensification requires special circumstances allowing water to be introduced into garnet-rich lower crust after transiting the mantle lithosphere; associated hydration of the mantle lithosphere might be limited if fluids were largely transported in discrete channels (Nielson et al., 1993), but both crustal and mantle hydration could contribute to uplift. Thus the processes we propose for the western United States might represent a special class of crustal alteration representing a new means of uplifting broad regions of continental crust.

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