Recent Craton Growth by Slab Stacking Beneath Wyoming

Eugene D. Humphreys (corresponding author)

Department of Geological Sciences

1272 University of Oregon

Eugene, OR 97403-1272

541 346 5575

541 346 4692 (Fax)

genehumphreys@gmail.com

Brandon Schmandt

Department of Earth and Planetary Science

The University of New Mexico

Albuquerque, NM 87131

bschmandt@unm.edu

Maximiliano J. Bezada

Department of Earth Sciences

University of Minnesota

310 Pillsbury Drive SE

Minneapolis, MN 55455-0231

mbezada@umn.edu

24 Jonathan. Perry-Houts
25 Department of Geological Sciences
26 1272 University of Oregon
27 Eugene, OR 97403-1272
28 jperryh2@uoregon.edu

Seismic tomography images high-velocity mantle beneath the Wyoming craton extending to >250 km depth. Although xenoliths and isostatic arguments suggest that this mantle is depleted of basaltic component, it is not typical craton: its NE elongate shape extends SW of the Wyoming craton; xenoliths suggest that the base of Archean mantle was truncated from ~180-200 to ~140-150 km depth since the Devonian, and that the deeper mantle is younger than ~200 Ma. The Sevier-Laramide orogeny is the only significant Phanerozoic tectonic event to have affected the region, and presumably caused the truncation. Apparently, the base of the Wyoming craton was removed and young, depleted mantle was emplaced beneath the Wyoming craton during the Sevier-Laramide orogeny. We suggest that the Wyoming craton experienced a ~75 Ma phase of growth through a three-stage process. First, flat-slab subduction removed 40-50 km off the base of the Archean Wyoming craton. This was followed by emplacement of basalt-depleted ocean plateau mantle lithosphere of the Shatsky Rise conjugate, which arrived in the early Laramide. The geologic record of vertical motion in the Wyoming region suggests that the plateau's crust escaped into the Earth's interior at 70-75 Ma. Initiation of Colorado Mineral Belt magmatism at this time may represent a slab rupture through which the ocean crust escaped.

Keywords:

Wyoming craton, Shatsky conjugate, tomography, Laramide, Sevier, Colorado Mineral Belt

1. Introduction

Archean cratons of continents are stabilized and protected from tectonic and magmatic disruption by their highly viscous, basalt-depleted mantle lithosphere whose thermal negative

buoyancy is balanced by their compositional buoyancy (Jordan, 1978). The basis for this understanding is provided by seismic observations and isostatic arguments, and by xenolith samples of this mantle. However, the origin of the depleted mantle is not agreed upon, with three primary hypotheses being considered (Arndt et al., 2009; Lee et al., 2011, Snyder et al., 2014). Most prominent among these is the “slab stacking” hypothesis, in which subducted slabs of Archean ocean lithosphere under-accrete the continent. Alternatives include an origin as plume heads and as mantle depleted below volcanic arcs. In all these hypotheses a question arises as to the whereabouts of the basaltic complement to the depleted mantle (Arndt et al., 2009). For the plume and volcanic arc hypotheses, a common suggestion is eruption and erosion of the basalt. For slab stacking – where the basaltic oceanic crust would have transformed to eclogite – a sinking of the dense eclogite into the asthenosphere is required.

This paper addresses the well-studied Wyoming craton. As a result of USArray seismic coverage, it is among the best-imaged cratons, and it is unusual for having experienced recent and significant modification. As a consequence of this modification it stands high, and it provides xenolith sampling of its structure at times both before and after this modification. The record created by this relatively young activity sheds light on the processes of craton creation and destruction.

A range of geophysical, geochemical and geologic evidence, discussed below, leads us to suggest that the Wyoming craton experienced Sevier-to-Laramide-age basal erosion followed by the emplacement of highly depleted oceanic mantle lithosphere. The geologically-recorded history of vertical motions provides clues as to how the ocean crust was lost during this young slab-stacking event.

2. Geologic History

The Wyoming craton was created by addition of Neoarchean volcanic arcs to a continental core of Paleoarchean to Mesoarchean age (Chamberlain et al., 2003). Then, through a rapid succession of amalgamation events at ~1.8 Ga, this craton was added to the southern margin of the North American shield and placed deep within the continental interior by south-side terrane accretion (Whitmeyer and Karlstrom, 2007). Through the Proterozoic and until the late Cretaceous the Wyoming craton remained stable (Blackburn et al., 2012). Wyoming subsided slowly and, in a series of transgressions and regressions, was covered by a westward thickening 1.5-3 km of sediment (Peterson, 1988) deposited in shallow seas and broad flood plains. The late Paleozoic orogeny responsible for the Ancestral Rockies created major mountains and basins within the Proterozoic crust south of Wyoming, but left the Archean crust nearly unaffected (Erslev and Koenig, 2009). Another layer of sediment was deposited in the Cretaceous Interior Seaway, starting ~100 Ma and covering most of Wyoming with 1.5–5 km of sediment (Molenaar and Rice, 1988; Scott et al., 2009; Liu et al., 2011; Cook and Bally, 1975). Seaway subsidence is attributed to dynamic suction associated with a shallowing of Farallon slab dip (Mitrovica et al., 1989; Spasojevic et al., 2009). Unusually thick deposits were laid down in westernmost Wyoming, NW Colorado and eastern Utah (Roberts and Kirschbaum, 1995; Liu et al., 2011, Chapin, 2012) as a combined result of flexural downwarping driven by the ~east-verging Sevier overthrust system (DeCelles, 2004) and a more regional dynamic suction (Liu et al., 2011).

Then, with the first significant deformation in over a billion years, crystalline basement was thrust over its sedimentary cover as Wyoming crust was shortened by 50-60 km (Bird, 1998) during the Laramide orogeny (~75-45 Ma, Dickinson et al., 1988). The accommodating thrust

faults are distributed across Wyoming in a branch-like pattern that created closed and nearly closed basins. Sediment was stripped from areas of high relief, often exposing crystalline basement, and was deposited into the basins and then largely evacuated (McMillan et al., 2006).

Based on the occurrence of thrust faulting in the deep continental interior and the termination of California arc volcanism at ~80 Ma, the Laramide orogeny generally is attributed to flattening of subducted Farallon slab against the base of North America (Coney and Reynolds, 1977; Saleeby, 2003). This is supported by the xenolith record, which suggests basal erosion of cratonic western U.S. to depths of 135-150 km sometime between the Devonian and the Eocene periods, presumably during the Sevier and Laramide orogenies (Li et al., 2008; Eggler et al., 1988). The primary cause of slab flattening is thought to be subduction of the large ocean plateau created as a conjugate to the Shatsky Rise (Livaccari et al., 1981; Saleeby, 2003). The calculated path of the plateau takes it under southern California, the Colorado Plateau and Wyoming (Liu et al., 2008). During this time the Colorado Plateau and Rocky Mountain region began ascending to their current high elevations (Flowers et al., 2008; Karlstrom et al., 2012), and the region became magmatically quiet except along the Colorado Mineral Belt.

3. Mantle structure

Our goal is to understand the origin of the deep part of the Wyoming craton, and we make use of three approaches. Seismic tomography provides the best constraint on the shape of thermal structure in this region. Isostatic calculations provide constraint on lithospheric density and its time evolution, and when combined with the seismic constraints on temperature, it sheds light on the degree of basalt depletion. Xenoliths sample the mantle to depths ≥ 170 km at a few

locations, thereby providing information on composition and temperature at their time of eruption. The time spanned by the xenolith eruptions supply information on lithospheric evolution. We then consider our understanding of the deep Wyoming craton in the context of Late Cretaceous subduction of the Shatsky conjugate and flat-slab subduction. Subduction of this ocean plateau provides both a source of depleted mantle and a means for delivering it to the base of the Wyoming craton.

3.1 Mantle seismic structure

Regional upper mantle structure has been imaged well with both surface waves and body waves. We use the Rayleigh-wave tomography of Shen et al. (2013; online download), which they inverted jointly with receiver-functions to constrain Moho depth. Fig. 1 shows this model in our study area. To make these figures, we remove the mean velocity of each layer and divide the result by the average S-wave velocity for that layer to obtain a layer-wise velocity perturbations that can be compared to the body-wave tomography. The surface-wave model images a 400-km wide high-velocity ($V_s = 4.65\text{-}4.8$ km/s; in Fig. 1, $\delta V_s > 3\%$ fast) structure extending SW-NE across Wyoming and extending from ~60 to 180 km in depth. The deeper parts of this model are not well resolved: the published model extends to 150 km depth – considered a depth that is reliably resolved – whereas the online version provides the entire 200 km depth extent of the model. The NE-trending high-velocity mantle structure beneath Wyoming is the only major high-velocity upper-mantle structure imaged with surface waves west of the Great Plains. We associate the deep parts of this structure with the Shatsky conjugate.

The teleseismic P-wave tomography provides the best resolution on the deeper structure. We use the model of Schmandt and Humphreys (2010, Fig. 1). Resolution of body-wave tomography diminishes at depths above the average station spacing, which for our data is ~70 km. For deeper structure, relatively high resolution is achieved through the use of abundant data recorded over a broad area. Inversion also makes use of finite-frequency sensitivity kernels, although at the highest frequencies (~1 Hz) used for arrival-time estimation (by cross correlation), finite-frequency effects are of little significance. Lower frequency measurements are also included to reduce the reliance on regularization to stabilize the inversions and obtain smooth images. Resolution tests indicate that lateral and depth resolution is excellent over the 75-700 km depth range, with only minor amplitude degradation of the model amplitude at each node (Fig. S1). Visually there is little degradation of the input image. Synthetic resolution tests, however, do not address artifacts created by effects that are not included in the test (including seismic anisotropy, ray-path deviation from that in a 1-D Earth, and structure of scale smaller than our node spacing). Considering these limitations, the actual uncertainty in the imaged structure is greater than that represented by the resolution tests, and resolution estimation becomes somewhat subjective. Two lines of reasoning suggest that the actual resolution of the P-wave model is good across the upper mantle.

Frist, when the body- and surface-wave models are compared, it is clear that the two tomographic techniques image similar structure over the 75-160 km depth range where resolution is good in both models. The average amplitudes of the S-wave anomalies are about twice that of the P-wave anomalies, as is common (Schmandt and Humphreys, 2010, found $d\ln V_p/d\ln V_s = 1.8$ for the western U.S.). The main difference seen in the body- and surface-wave images over this depth range is the presence in the body-wave images of shorter wavelength

lateral structure, as expected for inversions based on teleseismic arrivals, whereas the surface-wave model resolves shorter wavelength depth variation in structure. The similarity of results between inversions using different data types with different sensitivities suggests that the imaged structures are well resolved beneath our study area in the 75-160 km depth range. Below 160 km depth we rely solely on the teleseismic P-wave images. The excellent results from the resolution tests at these depths indicate that there is sufficient ray-path coverage for good resolution. This, in conjunction with the corroborated good resolution above 160 km depth, is a strong indication that resolution based on the P-wave images is good at greater depth. The second indication of good body-wave resolution in the upper mantle of the Schmandt and Humphreys (2010) model is that the imaged velocity variations near 410 and 660 km depth correspond well with imaged deflections in the 410 and 660 km discontinuities (Schmandt et al., 2012). As a result of these two comparisons, we think that vertical resolution on major structures beneath Wyoming and nearby environs across the 75-400 km depth range of interest here is ~50 km.

The NE-trending Wyoming structure is imaged extending to depths of ≥ 250 km, with the structure beneath NE Wyoming extending to greater depths. The main difficulty in assessing the deepest reaches of this structure comes from the complications created by the presence of deeper high-velocity structures in the region that are thought to be fragments of Farallon slab that subducted since 60 Ma (Liu and Stegman, 2011; Schmandt and Humphreys, 2010).

To conclude this section, it appears that a large volume of the Wyoming upper mantle is as seismically fast and thick as other cratons. Since variations in mantle seismic velocity as great as those imaged must represent significant temperature variations (Schutt and Lescher, 2006; Cammarano et al., 2003, 2009), the steep sides of this structure indicate that it is not in thermal equilibrium with the adjacent mantle and hence its form must be far younger than the Archean

age of the overlying crust and uppermost mantle or maintained by active processes (Cooper et al., 2004). We add that, when addressing the possible presence of Shatsky conjugate ocean crust, results from receiver function imaging find no evidence for a preserved layer of a basaltic ocean crust within this high-velocity body (Levander and Miller, 2012; Lekić and Fischer, 2013; Hopper et al., 2014). If an oceanic crust had converted to eclogite, its seismic velocity would make it difficult to image, so the presence of an eclogitic layer cannot be ruled out from seismic velocity structure alone. Isostatic arguments, however, suggest that significant amounts of eclogite are not present.

3.2 Mantle Xenoliths

Kimberlite eruptions at two well-studied sites brought mantle xenoliths to the surface from depths > 150 km (Fig. 2a). The southern site erupted during the Devonian, well prior to the Sevier-Laramide orogeny, and the northern site erupted during the early Eocene, shortly after this orogeny. Since the only significant tectonic event to affect this region between these two eruptions was the Sevier-Laramide orogeny, these xenolith suites afford an opportunity to infer how this orogeny modified the mantle lithosphere (Eggler et al., 1988; Carlson et al., 1999). The southern site erupted through Proterozoic crust just south of the Wyoming craton. These samples define a cool Devonian geotherm (Eggler et al., 1988, Ashchepkov et al., 2013), originate from depths as great as ~200 km (Eggler et al., 1988), and incorporate Archean xenoliths from the greatest depths (Schulze et al., 2008). That is, these xenoliths define a Devonian craton of typical age, thickness and thermal structure (Artemieva, 2009) (Fig. 2b).

The northern site is exposed in the northernmost Wyoming craton near the Great Falls tectonic zone (Fig. 2a). Two xenolith populations are found there. A “low-temperature” suite defines a cool pre-eruptive geotherm (of $\sim 40 \text{ mW/m}^2$, Hearn, 2004) to depths of $\sim 130 \text{ km}$ (Hearn, 2004) to $\sim 140 \text{ km}$ (Eggler et al., 1988) (Fig. 2b) typical of a craton (Artemieva, 2009), and these xenoliths are of Archean age (Carlson et al., 2004). Xenoliths from greater depth, the “high-temperature” suite, are about as depleted as the low-temperature suite (Carlson et al., 1999; Carlson et al., 2004), but they are younger than a few hundred million years in age (Carlson et al., 1999; Carlson et al., 2004) and record temperatures $\sim 200\text{-}300^\circ\text{C}$ above that defined by the Archean geotherm (Hearn, 2004) (Fig. 2b). These samples come from depths estimated to cover the range $\sim 150\text{-}200 \text{ km}$ (Eggler et al., 1988) or $130\text{-}170 \text{ km}$ (Hearn, 2004). The common explanation for these findings is that Laramide-age flat-slab subduction truncated the craton base near 150 km depth, and that the deeper lithosphere is young, depleted peridotite of an undetermined origin (Eggler et al., 1988; Carlson et al., 1999; Carlson et al., 2004). The downward extrapolation of the Archean geotherm intersects the mantle solidus at $200 \pm 30 \text{ km}$ (Fig. 2b), implying that prior to its (young, presumed Laramide) truncation, this lithosphere had a craton-like thickness and thermal structure similar to that resolved in the southern (Devonian) site (Eggler et al., 1988). The only simple explanation for these xenoliths is the removal of the Archean lithosphere from a depth of $\sim 170\text{-}200 \text{ km}$ to a depth of $\sim 130\text{-}150 \text{ km}$, followed quickly by the emplacement of young depleted mantle that experience temperatures of $\sim 1250\text{-}1450^\circ\text{C}$ (Hearn, 2004) but was cool when it was emplaced against the overlying Archean lithosphere (to account for both its high seismic velocity and the preserved cool Archean geotherm). The location of this xenolith site is north of the deep high-velocity mantle structure that we associate

with the Shatsky conjugate, suggesting to us that the deep xenoliths from the northern site sampled more typical Farallon ocean lithosphere adjacent to the Shatsky conjugate.

Although the two well studied deep-sampling xenolith sites are far from one another and lie on the edges of the Wyoming craton, the fact that the Colorado site was cool and ~200 km thick during the Devonian eruptions, and the Montana site was cool (and therefore thick) shortly before the Eocene eruptions, seems to indicate the presence of a pre-Sevier-Laramide craton root extending to ~200 km beneath each site and, reasonably, beneath the intervening Archean craton as well. At present, the base of Archean North America occurs at a depth of ~130-150 km beneath central Montana (Eggler et al. 1988; Hearn, 2004). The other well-studied location where xenoliths define the post-Laramide base of the lithosphere is the Paleoproterozoic lithosphere of the central Colorado Plateau. There, the base occurs at a depth of ~135-140 km (Smith, 2010; Li et al., 2008). Assuming that flat-slab subduction was the cause of North America basal erosion (e.g., Li et al., 2008; Smith, 2010), the xenolith samples suggest that Laramide-age erosion removed basal North America lithosphere to depths of 135-150 km.

3.3 Isostasy

Wyoming's current elevation of 1.2-2.3 km is unusually great for cratonic lithosphere. If one were to estimate its elevation using a simple Airy isostatic calculation, the crustal thickness of Wyoming yields an elevation that is 1-2 km lower than that observed (Fig. S2). This implies a relatively buoyant mantle lithosphere. However, our interest in mantle density structure is not addressed well with this calculation since it does not incorporate information on other sources of lateral density variations that are known to be present (e.g., Gilbert, 2012; Levandowski et al.,

2014), and addressing the mantle buoyancy structure in particular is made difficult because the imaged structure is not the dominant factor dictating current-day topography in the region (Gilbert et al., 2003; Becker et al., 2013; Levandowski et al., 2014). This presumably is because the effects of cool temperatures (which largely creates the mantle seismic structure) are largely offset by the compositional buoyancy of this Archean and Paleoproterozoic mantle (Jordan, 1988; Lee et al., 2011; Bailey et al., 2012; and the mantle xenoliths discussed in the preceding section).

More insight is gained by comparing Wyoming's current elevation with its long-stable elevation near sea level prior to incursion of the Cretaceous Seaway (Spencer, 1996). Across most of the Rocky Mountains and Great Plains, the near-shore sediment of the Dakota Group was the first to be deposited as a result of the encroaching Seaway. This deposition began ~100 Ma (Scott et al., 2009), usually unconformably upon pre-Cretaceous rock. Because insignificant amounts of sediment were deposited between 100-125 Ma, we choose 125 Ma as a reference isostatic state. This time predates Seaway downwarping and the Sevier thrusting near the western margin of our study area. We then assess dynamic topography and changes in crustal buoyancy since that time to address changes in mantle lithosphere buoyancy. The results (Fig. 3) are smoothed at a wavelength of 150 km, roughly equivalent to the shortest flexural wavelength of this lithosphere (Lowry and Perez-Gussinye, 2011). At wavelengths of 150 km and longer, the primary crustal causes of increased surface elevation since 125 Ma are found to be the addition of Seaway sediment (accounting for up to 700 m of elevation), Laramide crustal thickening (up to nearly 250 m of elevation), post-Laramide crustal heating (up to nearly 500 m, Levandowski et al., 2014), and possibly lower-crustal hydration (Jones et al., 2015). After removing the effects of dynamic topography, sedimentation, erosion, crustal thickening and thermal expansion of the

crust, we conclude that about 0.4-1.4 km of Wyoming's current elevation can be attributed to a combination of increased mantle lithosphere buoyancy since 125 Ma and lower crustal hydration (Fig. 3i).

3.3.1. Current dynamic topography created by density structure below 200 km depth

We estimate current dynamic topography created by the density structure below 200 km using the Becker et al. (2013) coding of the Hager and O'Connell (1981) algorithm for viscous flow in a spherical Earth possessing only radial viscosity variations. Flow is driven by a density structure inferred by scaling the global Smean S-wave seismic model (Becker and Boschi, 2002) to density, using $d \ln \rho / d \ln V_s = 0.2$ (Stixrude and Lithgow-Bertelloni, 2011). We find that structure deeper than 200 km creates 73-50 m of dynamic subsidence (W-E) across our study area (Fig. 3b), and dynamic subsidence remains insignificant with plausible variations in radial viscosity structure. Considering that 125 Ma precedes the creation of the Cretaceous Seaway, we assume no significant dynamic topography at 125 Ma.

3.3.2. Elevation at 125 Ma

At 125 Ma, the regional landform across the Colorado Plateau, Rocky Mountains and Great Plains was one of low elevation (Fig. 3c) and relief, with the former Ancestral Rockies having been eroded to low hills and the highlands of the Sevier fold and thrust belt occupying what now is the Basin and Range (Molenaar and Rice, 1988, Blakey and Ranney, 2008). The shoreline crossed southern Wyoming with a shallow sea opening to the NNE (Blakey and Ranney, 2008; Sloss, 1988), suggesting a reference isostatic state for Wyoming near the sea level at that time. Fig. 3c shows our assumed initial elevation field, which we estimate from the 125 Ma paleo-

geographic reconstruction of Blakey and Ranney (2008). The disturbance on the west side of the area shown in Fig. 3 is the result of the early Sevier orogeny. Sea level at 125 Ma averaged about 150 m higher than its current level (Haq et al., 1988), so we add this amount to the elevation estimated at 125 Ma.

3.3.3. Post-125 Ma sedimentation and erosion

Deposition of sediment after 125 Ma elevated the surface by thickening the crust with low-density rock. We use the sediment isopach maps of Cook and Bally (1975) for Eocene and older deposits and McMillan et al. (2006) for younger deposits to infer net post-125 Ma sediment thickness (averaged over rectangular areas one-degree on a side). Rock density of this spatially varying sediment pile is estimated using the S-wave upper crustal model of Lin et al. (2014) and Brocher's (2005) velocity-to-density scaling relations. A mantle density of 3300 kg/m³ is assumed in our isostatic calculations. Fig. 3d shows the smoothed estimate of the resulting elevation gain. The major deep sedimentary basins are clearly visible.

To calculate the effects of the erosion of pre-125 Ma sediment from the Colorado Plateau and the Laramide uplifts, we use Fan and Carrapa (2014) and Reed et al. (2005) to estimate the amount of this sediment that was removed (again, averaging over areas one-degree on a side). The pre-erosion thickness of this sediment is estimated by extrapolating the mapped sediment thickness of the pre-125 Ma sediments Cook and Bally (1975). Sediment and mantle densities of 2550 and 3300 kg/m³, respectively, are assumed. Fig. 3e shows the combined effect on elevation of sediment deposited since our reference date of 125 Ma, and of the erosion of sediment older than 125 Ma. The erosion has a relatively minor overall effect, whereas post-125 Ma sediment accumulation is important. The overall RMS value of elevation gain is 330 m.

3.3.4. *Crustal thickening*

Except for the Sevier-age thrusting in westernmost margin of Wyoming, the only significant deformation in Wyoming occurred during the Laramide orogeny, and the amount of shortening was modest. We use the net displacements calculated by Bird (1998), which covers the time interval 40-80 Ma. From these displacements we calculate the strain field. To obtain change in crustal thickness, we assume that the current crustal thickness (from Shen et al., 2013) is created by straining an initial crust by the amount estimated from the Bird's displacement field, and use the resulting thickness difference to calculate an isostatic uplift assuming an average crustal density of 2700 kg/m^3 and a mantle density of 3300 kg/m^3 . Small values are obtained for crustal thickening, consistent with imaged crustal thickness variations, which show no apparent correlation with the distribution of Laramide thrusting (Gilbert, 2012; Yeck et al., 2014). The smoothed result is shown in Fig. 3f. At this level of smoothing, Rocky Mountain uplifts are as great $\sim 200 \text{ m}$, with a maximum uplift of $\sim 250 \text{ m}$ occurring at the overthrust belt in westernmost Wyoming.

3.3.5. *Crustal heating*

We use the values for crustal thermal uplift estimated by Levandowski et al. (2014) (Fig. 3g). This field is dominated by the heating in the Yellowstone area and across much of Colorado, largely over the Rocky Mountain region. We assume that the initial thermal structure was uniform over this cratonic region at 125 Ma, and that only warming has occurred to produce the observed thermal structure. We also make a simple eastward extrapolation of the Levandowski et

al. (2014) field from -102° to -99° in a smooth manner that follows the regional heatflow map (Blackwell et al., 1991). Estimated thermal uplift for Wyoming is 200-400 m.

3.3.6. *Summary of Isostasy*

Fig. 3h shows the combined effects on elevation increase since 125 Ma resulting from all crust and mantle contributions except those originating from a hydrated lower crust and from buoyancy changes in the mantle lithosphere. Ignoring crustal hydration, Fig. 3i shows our calculated increase in mantle lithosphere buoyancy since 125 Ma (represented as surface uplift). The result is a regional, long-wavelength uplift that increases from zero in the eastern Great Plains of North and South Dakota to about 0.4 km in NE Wyoming and to about 1 km in western Wyoming. In addition to this long-wavelength trend, there is a ridge of anomalous elevation extending from the very slow mantle beneath Yellowstone to the Colorado Rocky Mountains. In Wyoming, this ridge adds an additional 0.3-0.4 km of elevation. For comparison, the colored circles in Fig. 3i shows the nearly identical uplift calculated by Spencer (1996) along a traverse that crosses the Colorado Plateau, Colorado Rocky Mountains and the Great Plains. Also shown on this figure is the elevation predicted by Jones et al. (2015) for uplift due to lower crustal hydration. At least at the three sites discussed by Jones et al., their predicted uplift is equivalent to our elevation anomaly (squares in Fig. 3i). Assuming that their estimated uplift calculations are applicable across the Wyoming region, then it is possible that the mantle lithosphere in the Wyoming area has about the same buoyancy now that it has had for hundreds of million years.

Across most of Wyoming the mantle is imaged to be of relatively high velocity (and hence cool), extending in depth to 230-310 km (Fig. 1a). This mantle buoyancy has remained about constant or increased since before the region was affected by the Sevier orogeny (Fig. 3i).

However, the xenolith evidence from north-central Montana suggests that the mantle beneath ~150 km is young. The positive-to-neutral buoyancy of the deep high-velocity mantle (that we attribute to the Shatsky conjugate) implies that it is more depleted than normal Phanerozoic mantle. A plume origin and extraction of ~20 km of basalt from the mantle during creation of the Shatsky conjugate (Fig. S4) would create such highly depleted upper mantle. If the deep high-velocity mantle does have its origin as the Shatsky conjugate, the inferred non-negative mantle buoyancy precludes the presence of significant volumes of its eclogitic ocean crust.

4. The Shatsky conjugate

4.1 Shatsky conjugate motion

Most workers ascribe initial Sevier-Laramide slab flattening to the end-on subduction of a large ocean plateau created as a conjugate to the Shatsky Rise (Livaccari et al., 1981; Saleeby, 2003) at the Pacific-Farallon-Izanagi triple junction (Nakanishi et al., 1999). The shape of the Shatsky conjugate shown in Fig. 4 is made by reflecting the Shatsky Plateau about its short axis, orienting this plateau in direction of Farallon spreading at time of its creation, and removing the Ori volcanic construction on its NW side (being located west of the main volcanic trend, the conjugate Ori probably would have been on the Izanagi plate, not the Farallon plate). The motion of this plateau across the Pacific basin has been calculated from plate reconstruction models (Liu et al., 2008), and assuming this plateau remained attached to the Farallon Plate, its subduction would have begun ~90-95 Ma beneath southern California and follow a path similar to that illustrated in Fig. 4 (Liu et al., 2008).

By using the Shatsky Rise as a reference, the leading (oldest) part of the subducted conjugate would have been comprised of 20-30 km of basaltic crust (Korenaga and Sager, 2012) and its underlying basalt-depleted mantle. The initial volcanic construction on the Shatsky Rise (the Tamu massif) is described as the world's largest volcano (Sager et al., 2013), suggesting that the underlying mantle is the most depleted. Our ability to infer the shape of the Shatsky conjugate from the Shatsky plateau is limited on its NE and NW sides (Fig. S3. This and additional aspects of the reconstruction are provided in the Supplement.). This differs from its SW margin, where the shape of the Shatsky conjugate was created adjacent to the Pacific plate and its shape can be inferred from the Shatsky Rise.

The high-velocity Wyoming mantle deeper than 140 km is located near where Liu et al. (2008) place the Shatsky conjugate at ~64 Ma (Fig. 4), and its shape is similar to the inferred shape of this plateau. The most prominent and deepest part (>250 km) of the imaged high-velocity Wyoming mantle structure lies beneath NE Wyoming. Being on the leading (NE) end of the high-velocity structure, we associate this with the conjugate Tamu volcanic center. We modify the path estimated with plate reconstruction models (Liu et al., 2008) so that the Shatsky conjugate is located beneath the imaged high-velocity structure and with the inferred conjugate Tamu volcanic center positioned beneath the deepest high-velocity structure (Fig. 4). To make this path, we move the Liu et al. path north ~200 km and straighten out the predicted change in transport direction at ~80 Ma. These modifications are made to coordinate with the mantle image and match the geologic observations discussed in the following section (especially the location of the severely disrupted southern California crust described by Saleeby, 2003). Between these two constraints the actual path of the Shatsky conjugate is not well constrained, and other relatively minor variations to this path are possible. The positioning of the Shatsky conjugate

shown in Fig. 4 is one in which the ocean plateau rotated 8° clockwise so that the conjugate of the large Shirshov volcanic construction correlates with the imaged high-velocity structure beneath the SW Colorado Plateau. Levander et al. (2011) provide an interpretation of this high-velocity feature that involves a downwelling “drip” of the North American lithosphere, and this accounting does not require the Shatsky conjugate. We consider it possible that the deep part of this mantle structure is the mantle complement to the conjugate of the Shirshov massif; it also is possible that this high-velocity feature does not involve the mantle of the Shatsky conjugate at all.

4.2 Fate of the Shatsky conjugate crust

An unusual origin for the high-velocity Wyoming mantle deeper than ~150 km is implied by its basalt-depleted composition and young age (inferred from nearby xenoliths and isostatic modeling). The fortuitous arrival of mantle with these properties on the flat-subducting Farallon slab makes ocean-plateau under-accretion an attractive option. However, the fate of the plateau’s ocean crust then becomes an outstanding problem. This crust subducted beneath southern California, but seismic and isostatic evidence indicates that it is not present beneath Wyoming. During transit beneath North America at depths of ~140 km beneath the Colorado Plateau (Smith, 2010; Li et al., 2008) and Wyoming (Egger et al., 1988; Hearn, 2004), the Farallon and plateau basalt crust would eventually transform to eclogite. We use the geologically recorded history of vertical tectonics to infer when and where this transformation occurred, and for the subsequent fate of this crust.

The 90-80 Ma subduction of the Shatsky conjugate (Liu et al. Supplement, 2010; Fig. 4) removed much of continental lithosphere beneath southern California (Saleeby, 2003; Luffi et al., 2009), emplaced Pelona-type schists (Grove et al., 2003), and tectonically disrupted the volcanic arc (Saleeby, 2003). Farallon mantle lithosphere is sampled beneath the Mojave Desert with its crust presumably attached (Luffi et al., 2009), and basanite blocks are common in exposed schist outcrops that probably are pieces of Farallon ocean crust (Jacobson et al., 2011). If these are samples of Shatsky conjugate crust, they could account for some of the missing crust. However, the volume of ocean crust that might reside beneath California is small compared to that of the Shatsky conjugate crust. It appears that most of the Farallon oceanic crust, both off and on the Shatsky conjugate, was transported east of California with the Farallon slab.

Beneath the Colorado Plateau, the 85-60 Ma hydration of lower lithosphere under cool temperatures (Smith and Griffin, 2005), and hydration-enabled erosion of >10 km of basal North American lithosphere (Li et al., 2008), implies that Farallon slab was in contact with the base of the continent. Under these conditions the Farallon crust probably would be too cool (English et al., 2003; Currie and Beaumont, 2011; Menea et al., 2012) to overcome the activation barrier required for eclogite transformation (of ~600-800°C, van Hunen et al., 2004). Uplift above sea level and creation of km-scale relief in the SW Colorado Plateau at 80-70 Ma (Flowers et al., 2008; Flowers and Farley, 2012) corresponds to the time when the Shatsky conjugate passed beneath this region (Fig. 4, Liu et al., 2008), consistent with the idea that the Shatsky conjugate passed through this region with a basaltic crust. It may be possible that some of the oceanic crust was lost beneath the Colorado Plateau, as suggested by Usui et al. (2003) based on their interpretation of xenoliths, or near the Cheyenne Belt, consistent with the region's anomalously

dense crust (Crosswhite and Humphreys, 2002). It remains unknown, however, how much of the Farallon crust may have been transferred to the base of the North America interior during transit.

In contrast to SW Colorado Plateau uplift, sediments deposited into large basins in NW Colorado and southern Wyoming record major subsidence at 85-70 Ma (Cook and Bally, 1975; Roberts and Kirschbaum, 1995; Liu et al., 2011; Chapin, 2012). The Shatsky conjugate is thought to have arrived to this area at ~75-70 Ma (Fig. 4), i.e., towards the end of this anomalous subsidence event (Liu et al. Supplement, 2010, and Fig. S5). We infer that the subsidence indicates eclogitization of the Farallon ocean crust, and then Shatsky-conjugate crust, as they reached this location. Indeed, if the Shatsky conjugate passed by with basaltic crust, we would expect an uplift event similar in magnitude to that experienced by the SW Colorado Plateau. Thermal evolution models of flat-subducting ocean lithosphere predict that the crust would have attained temperatures of about 500-800°C (depending on conditions) in this vicinity (English et al., 2003; Currie and Beaumont, 2011; Menea et al., 2012).

A waning of sediment deposition in the NE Colorado and southern Wyoming basins indicates the region began to emerge at ~75 Ma in NW Colorado and ~70 Ma in south-central Wyoming (Roberts and Kirschbaum, 1995). This suggests a removal of the load that drove downwarping. If eclogitic ocean crust was the load, then the rebound may well represent the eclogite loss, presumably to the Earth's interior.

The Colorado Mineral Belt initiated at ~75 Ma (Fig. S6, Jones et al., 2011; Chapin, 2012 and references therein). These plutons are largely crustal melts, although heating probably necessitated mantle-derived magma. Accounting for asthenospheric access has been a problem because volcanism across most of the western U.S. had ended, which usually is attributed to flat-slab subduction. Jones et al. (2011) suggest small-scale mantle flow within a narrow mantle

wedge above a shallowly dipping slab, whereas Chapin (2012) suggests a tearing of the flat Farallon slab below the Colorado Mineral Belt. In our preferred reconstruction (Figs. 4 and S6), the SW margin of the Shatsky conjugate lies directly below the Colorado Mineral Belt at 75 Ma. We infer that an eclogitized Shatsky-conjugate ocean crust escaped in some fashion through a tear in the Farallon slab beneath the Shatsky conjugate, starting at ~75 Ma. Crustal mobility may have been a product of low viscosity induced by the presence of eclogitic quartz, crustal fabric, shear stress, the presence of water or recrystallization during transformation.

5. Summary and Discussion

A consideration of seismic tomography, isostasy, xenoliths, and plate tectonic history provides a case for the under accretion of the Shatsky conjugate beneath Wyoming at ~65 Ma. *Xenoliths* appear to record the local erosion of ~30-50 km of Archean lithosphere from basal North America, and the emplacement of young, depleted mantle beneath the Wyoming craton. *Isostatic* considerations of the Wyoming area indicate a current mantle buoyancy that is neutral to positive relative to times before 125 Ma, but that there was a phase of negative buoyancy beneath the region north of the current Colorado Mineral Belt that ended 75-70 Ma. *Seismic tomography* images high-velocity mantle to depths of ~230-310 km, deepest beneath NE Wyoming. The buoyancy of this high-velocity mantle seems to require that it is depleted of basaltic component. The creation of a large, cool and depleted volume of mantle lithosphere beneath a craton within the continental interior would normally be difficult to explain. However, in an unusual *tectonic* event, a large ocean plateau subducted beneath North America when and where it is needed. To deposit the ocean plateau where imaged, it must stop subducting at ~65

Ma. Fig. 5 illustrates the basic elements of this history. Cross sections of the inferred initial Farallon lithosphere that contains the Shatsky conjugate, and that of the initial and final North American lithosphere, are discussed in the Supplement and Fig. S4.

Our model has the Shatsky conjugate ocean plateau subducting with its crust but under-accreting Wyoming without its crust. If we assume that this model is correct, we look to the history of vertical motion along this plateau's path to understand where and how the ocean crust was separated from the depleted-mantle portion of the plateau. These motions suggests the ocean crust transformed to eclogite beneath NW Colorado and was subsequently lost starting ~75 Ma. Because most of this crust is missing, a loss to the deeper mantle seems likely. The coincident initiation of the Colorado Mineral Belt indicates a probable tear in the Farallon (Chapin, 2012), which could provide a pathway for this loss.

Debate continues on origin of cratons. In this paper we describe a young slab-stacking event in which an Archean-like piece of ocean lithosphere was made by a hotspot at a spreading center (creating a relatively highly depleted mantle lithosphere) and subsequently subducted beneath southern California. As in the Archean slab-stacking hypothesis, the oceanic crust has been lost. In our case, however, the geologic record of anomalous vertical motions in Colorado and Wyoming and the initiation of the Colorado Mineral Belt volcanic trend can be used to suggest that this loss may have occurred through a tear in the subducted Farallon slab.

The basement-cored uplifts characteristic of Laramide orogeny began ~75 Ma i.e., well after the subduction of the Shatsky conjugate and during our proposed Shatsky crustal loss event. It was at ~75 Ma that compressive stress in the Rocky Mountain region switched from east-oriented to NE-oriented (Bird, 2002). In the following ~10 m.y., during the time when we infer the Shatsky conjugate mantle lithosphere was being transported to beneath Wyoming, the

Wyoming thrust system developed (with the other Laramide thrusts initiating later) (Bird, 2002).
It is interesting to note that the Shatsky conjugate appears to have been transferred from the
Farallon lithosphere to the North America when the plateau encountered the Archean craton.

Acknowledgments

We thank Rick Carlson, Craig Jones, Roberta Rudnick, David Snyder, Stephan Sobolev and Jens
Tympel for useful and enjoyable conversations, Hersh Gilbert and Cin-Ty Lee for helpful
reviews, and Peter Shearer for handling the editorial responsibilities. We acknowledge support
from the CIDER program (EAR-1135452) and from NSF grants EAR-1052899 and EAR-
0952194 to EH.

References

- Arndt, N.T., Coltice, N., Helmstaedt, H., and Gregoire, M. (2009), Origin of Archean
subcontinental lithospheric mantle: Some petrological constraints. *Lithos*, 109(1), 61-71.
- Artemieva, I.M. (2009), The continental lithosphere: reconciling thermal, seismic, and petrologic
data. *Lithos*, 109, 23-46.
- Ashchepkov, I.V., Downes, H., Mitchell, R., Vladykin, N.V., Coopersmith, H., and Palesky,
S.V. (2013), Wyoming Craton Mantle Lithosphere: Reconstructions Based on Xenocrysts
from Sloan and Kelsey Lake Kimberlites, in D.G. Pearson et al. (eds.), Proceedings of 10th
International Kimberlite Conference, Volume 1, Special Issue of the Journal of the
Geological Society of India.

552 Bailey, I. W., Miller, M. S., Liu, K., and Levander, A. (2012), VS and density structure beneath
 553 the Colorado Plateau constrained by gravity anomalies and joint inversions of receiver
 554 function and phase velocity data, *J. Geophys. Res.*, *117*.

555 Becker, T.W., and Boschi, L. (2002), A comparison of tomographic and geodynamic mantle
 556 models, *Geochemistry, Geophysics, Geosystems*, *3*, 2001GC9000168.

557 Becker, T.W., Faccenna, C., Humphreys, E.D., Lowry, A.R., and Miller, M.S. (2013), Static and
 558 dynamic support of western United States topography, *Earth Planet. Sci. Lett.*, *402*, 234-246.

559 Bird, P. (1998), Kinematic history of the Laramide orogeny in latitudes 35°–49°N, western
 560 United States, *Tectonics*, *17*, 780-801.

561 Bird, P. (2002), Stress direction history of the western United States and Mexico since 85 Ma,
 562 *Tectonics*, *21*(3), 5-1.

563 Blackburn, T.J., Bowring, S.A., Perron, J.T., Mahan, K., Dudas, F.O., and Barnhart, K.R. (2012),
 564 An exhumation history of continents over billion-year time scales, *Science*, *335*, 73-76.

565 Blackwell, D.D., J.L. Steele, and L.C. Carter (1991), Heat flow patterns of the North American
 566 continent: A discussion of the DENAG geothermal map of North America, in Slemmons,
 567 D.B., Engdahl, E.R., Zoback, M.D., and Blackwell, D.D., eds., *Neotectonics of North*
 568 *America: Boulder, Colorado, Geol. Soc. Am., Decade Map Volume 1*.

569 Blakey, R.C., and Ranney, W. (2008), *Ancient landscapes of the Colorado Plateau*, Grand
 570 Canyon Assn., Grand Canyon, AZ, p. 156.

571 Brocher, T.M. (2005), Empirical relations between elastic wavespeeds and density in the Earth's
 572 crust, *Bull. Seism. Soc. Am.* *95*, 2081-2092.

573 Cammarano, F., Goes, S., Vacher, P., and Giardini, D. (2003), Inferring upper-mantle
 574 temperatures from seismic velocities. *Physics Earth Planet. Int.*, *138*, 197-222.

575 Cammarano, F., Romanowicz, B., Stixrude, L., Lithgow-Bertelloni, C., and Xu, W. (2009),
 576 Inferring the thermochemical structure of the upper mantle from seismic data. *Geophys. J.*
 577 *Int.*, 179(2), 1169-1185.

578 Carlson, R.W., Irving, A.J., and Hearn Jr, B.C. (1999), Chemical and isotopic systematics of
 579 peridotite xenoliths from the Williams kimberlite, Montana: clues to processes of lithosphere
 580 formation, modification and destruction, In *Proceedings of the VIIth International Kimberlite*
 581 *Conference* (Vol. 1, pp. 90-98).

582 Carlson, R.W., Irving, A.J., Schulze, D.J., and Hearn Jr, B.C. (2004), Timing of Precambrian
 583 melt depletion and Phanerozoic refertilization events in the lithospheric mantle of the
 584 Wyoming Craton and adjacent Central Plains Orogen, *Lithos*, 77, 453-472.

585 Chamberlain, K.R., Frost, C.D., and Frost, B.R. (2003), Early Archean to Mesoproterozoic
 586 evolution of the Wyoming Province: Archean origins to modern lithospheric architecture.
 587 *Canadian J. Earth Sci.*, 40, 1357-1374.

588 Chapin, C.E. (2012), Origin of the Colorado mineral belt, *Geosphere*, 8, 28-43.

589 Chapman, A.D., Kidder, S., Saleeby, J.B., Ducea M.N. (2010), Role of extrusion of the Rand and
 590 Sierra de Salinas schists in Late Cretaceous extension and rotation of the southern Sierra
 591 Nevada and vicinity, *Tectonics*, doi:10.1029/2009TC002597.

592 Coney, P.J., and Reynolds, S.J. (1977), Cordilleran benioff zones. *Nature*, 270, 403-406.

593 Cook, T.D., and A.W. Bally (1975), *Stratigraphic Atlas of North and Central America*, 272 pp.,
 594 Princeton Univ. Press, Princeton, N.J.

595 Cooper, C.M., Lenardic, A., and Moresi, L. (2004), The thermal structure of stable continental
 596 lithosphere within a dynamic mantle, *Earth Planet. Sci. Lett.*, 222, 807-817.

597 Crosswhite, J.A., and Humphreys, E.D. (2003), Imaging the mountainless root of the 1.8 Ga
 598 Cheyenne belt suture and clues to its tectonic stability, *Geology*, *31*, 669-672.

599 Currie, C.A., and Beaumont, C. (2011), Are diamond-bearing Cretaceous kimberlites related to
 600 low-angle subduction beneath western North America?, *Earth Planet. Sci. Lett.*, *303*, 59-70.

601 DeCelles, P.G. (2004), Late Jurassic to Eocene evolution of the Cordilleran thrust belt and
 602 foreland basin system, western U.S.A., *Am. J. Sci.*, *304*, 105-168.

603 Dickinson, W.R., Klute, M.A., Hayes, M.J., Janecke, S.U., Lundin, E.R., McKittrick, M.A., and
 604 Olivares, M.D. (1988), Paleogeographic and paleotectonic setting of Laramide sedimentary
 605 basins in the central Rocky Mountain region, *Geol. Soc. Am. Bull.*, *100*, 1023-1039.

606 Eggler, D.H., J.K. Meen, F. Welt, F.O. Dudas, K.P. Furlong, M.E. McCallum and R.W. Carlson
 607 (1988), Tectonomagmatism of the Wyoming Province, *in* J. Drexler and E.E. Larson, eds.,
 608 Colorado Volcanism, Colorado School of Mines Quarterly, Boulder, 25-40.

609 English, J.M, S.T. Johnson, K Wang, 2003, Thermal modeling of the Laramide orogeny: testing
 610 the flat-slab subduction hypothesis, *Earth Planet. Sci. Lett.*, *214*, 619-632.

611 Erslev, E.A., and N.V. Koenig (2009), Three-dimensional kinematics of Laramide, basement-
 612 involved Rocky Mountain deformation, USA: Insights from minor faults and GIS-enhanced
 613 structure maps, *Geol. Soc. Am. Bull. Memoir 204*, 125-150.

614 Fan, M., and Carrapa, B. (2014), Late Cretaceous–early Eocene Laramide uplift, exhumation,
 615 and basin subsidence in Wyoming: Crustal responses to flat slab subduction, *Tectonics*, *3*,
 616 509-529.

617 Flowers, R. M., and Farley, K.A. (2012), Apatite $4\text{He}/3\text{He}$ and $(\text{U-Th})/\text{He}$ evidence for an
 618 ancient Grand Canyon, *Science*, *338*, 1616-1619.

619 Flowers, R. M., Wernicke, B. P., and Farley, K. A. (2008), Unroofing, incision, and uplift history

620 of the southwestern Colorado Plateau from apatite (U-Th)/He thermochronometry, *Geol. Soc.*
621 *Am. Bull.*, 120, 571-587.

622 Gilbert, H. (2012), Crustal structure and signatures of recent tectonism as influenced by ancient
623 terranes in the western United States, *Geosphere*, 8(1), 141-157.

624 Gilbert, H.J., Sheehan, A.F., Dueker, K.G., and Molnar, P. (2004), Receiver functions in the
625 western United States, with implications for upper mantle structure and dynamics, *J.*
626 *Geophys. Res.*, 108.

627 Grove, M., Bebout, G.E., Jacobson, C.E., Barth, A.P., Kimbrough, D.L., King, R.L., Zou, H.,
628 Lovera, O.M., Mahoney, B.J., and Gehrels, G.E. (2008), The Catalina Schist: evidence for
629 middle Cretaceous subduction erosion of southwestern North America, *Geol. Soc. Am.*
630 *Special Papers*, 436, 335-361.

631 Grove, M., Jacobson, C.E., Barth, A.P., and Vucic, A. (2003), Temporal and spatial trends of
632 Late Cretaceous-early Tertiary underplating of Pelona and related schist beneath southern
633 California and southwestern Arizona, *Sp. Pap.* 374, Geol. Soc. Am., 381-406.

634 Hasterok, D., and Chapman, D. S. (2007), Continental thermal isostasy: 2. Application to North
635 America, *J. Geophys. Res.*, 112.

636 Hager, B.H., and O'Connell, R.J. (1981), A simple global model of plate dynamics and mantle
637 convection, *J. Geophys. Res.*, 86, 4843-4867.

638 Haq, B.U., Hardenbol, J., and Vail, P.R. (1988), Mesozoic and Cenozoic chronostratigraphy and
639 cycles of sea-level change, *Soc. Econ. Paleo. Min. Special Pub.* 42.

640 Hearn Jr., B.C. (2004), The Homestead kimberlite, central Montana, USA: mineralogy,
641 xenocrysts, and upper-mantle xenoliths, *Lithos* 77, 473-491.

642 Hopper, E., Ford, H. A., Fischer, K. M., Lekić, V., and Fouch, M. J. (2014), The lithosphere–
 643 asthenosphere boundary and the tectonic and magmatic history of the northwestern United
 644 States. *Earth Planet. Sci. Lett.*, 402, 69-81.

645 Jacobson, C.E., Grove, M., Pedrick, J.N., Barth, A.P., Marsaglia, K.M., Gehrels, G.E., and
 646 Nourse, J.A. (2011), Late Cretaceous–early Cenozoic tectonic evolution of the southern
 647 California margin inferred from provenance of trench and forearc sediments, *Geol. Soc. Am.*
 648 *Bull.*, 123, 485-506.

649 Jones, C.H., Farmer, G.L., Sageman, B., and Zhong, S. (2011), Hydrodynamic mechanism for
 650 the Laramide orogeny. *Geosphere*, 7, 183-201.

651 Jones, C.H., Mahan, K.H., Butcher, L.A., Levandowski, W.B., and Farmer, G.L. (2015),
 652 Continental uplift through crustal hydration, *Geology*, 43(4), 355-358.

653 Jordan, T.H. (1978), Composition and development of the continental tectosphere. *Nature*,
 654 274(5671), 544-548.

655 Jordan, T.H. (1988), Structure and formation of the continental tectosphere. *J. Petrology*, 1, 11-
 656 37.

657 Karlstrom, K., Coblenz, D., Dueker, K., Ouimet, W., Kirby, E., Van Wijk, J., Umhoefer, P., and
 658 Donahue, M. (2012), Mantle-driven dynamic uplift of the Rocky Mountains and Colorado
 659 Plateau and its surface response: Toward a unified hypothesis, *Lithosphere*, 4(1), 3-22.

660 Korenaga, J., and Sager, W.W. (2012), Seismic tomography of Shatsky Rise by adaptive
 661 importance sampling, *J. Geophys. Res.*, 117.

662 Lee, C.-T.A., Luffi, P., and Chin, E.J. (2011), Building and destroying continental mantle.
 663 *Annual Review of Earth and Planetary Sciences*, 39, 59-90.

664 Lekić, V., and K.M. Fischer (2013), Contrasting lithospheric signatures across the western
665 United States revealed by Sp receiver functions, *Earth Planet. Sci. Lett.*, *402*, 90-98.

666 Levander, A., and M.S. Miller (2012), Evolutionary aspects of lithosphere discontinuity structure
667 in the western U.S., *Geochem. Geophys. Geosyst.*, *13*, Q0AK07.

668 Levander, A., B. Schmandt, M. Miller, K. Liu, K. Karlstrom, R. Crow, C.-T. Lee, E. Humphreys
669 (2011), Continuing Colorado plateau uplift by delamination-style convective lithospheric
670 downwelling. *Nature*, *472*, 461-465.

671 Levandowski, W., Jones, C.H., Shen, W., Ritzwoller, M.H., and Schulte-Pelkum, V. (2014),
672 Origins of topography in the western US: Mapping crustal and upper mantle density
673 variations using a uniform seismic velocity model, *J. Geophys. Res.*, *119*, 2375-2396.

674 Li, Z.-X.A., C.-T.A. Lee, A.H. Peslier, A. Lenardic, and S.J. Mackwell (2008), Water contents in
675 mantle xenoliths from the Colorado Plateau and vicinity: Implications for the mantle
676 rheology and hydration-induced thinning of continental lithosphere, *J. Geophys. Res.*, *113*,
677 B09210, doi:10.1029/2007JB005540.

678 Lin, F.-C., Tsai, V.C., and Schmandt, B. (2014), 3-D crustal structure of the western United
679 States: application of Rayleigh-wave ellipticity extracted from noise cross-correlations,
680 *Geophys. J. Int.*, *198*, 656-670.

681 Liu, L. and M. Gurnis, M. Seton J. Saleeby, R.D. Müller, and J. Jackson (2010), The role of
682 oceanic plateau subduction in the Laramide orogeny, *Nature Geoscience*, *3*, 353-357.

683 Liu, L., and Nummedal, D. (2004), Late Cretaceous subsidence in Wyoming: Quantifying the
684 dynamic component, *Geology*, *32*, 397-400.

685 Liu, S., Nummedal, D., and Liu, L. (2011), Migration of Dynamic subsidence across the Late
686 Cretaceous United States Western Interior Basin in response to Farallon plate subduction,
687 *Geology*, 39, 555-558.

688 Liu, L.J., Spasojevic, S., and Gurnis, M. (2008), Reconstructing Farallon plate subduction
689 beneath North America back to the Late Cretaceous, *Science*, 322, 934-938.

690 Liu, L., and Stegman, D.R. (2011), Segmentation of the Farallon slab, *Earth Planet. Sci. Lett.*,
691 311(1), 1-10.

692 Livaccari, R.F., Burke, K., and Şengör, A.M.C. (1981), Was the Laramide Orogeny related to
693 subduction of an oceanic plateau?. *Nature*, 289, 276–278

694 Lowry, A.R., and Pérez-Gussinyé, M. (2011), The role of crustal quartz in controlling
695 Cordilleran deformation, *Nature*, 471, 353-357.

696 Luffi, P., Saleeby, J.B., Lee, C.-T.A., and Ducea, M.N. (2009), Lithospheric mantle duplex
697 beneath the central Mojave Desert revealed by xenoliths from Dish Hill, California, *J.*
698 *Geophys. Res.*, 114, B03202.

699 McMillan, M.E., Heller, P.L., and Wing, S.L. (2006), History and causes of post-Laramide relief
700 in the Rocky Mountain orogenic plateau, *Geol. Soc. Am. Bull.*, 118, 393-405.

701 Manea, V.C., Pérez-Gussinyé, M., and Manea, M. (2012), Chilean flat slab subduction controlled
702 by overriding plate thickness and trench rollback, *Geology*, 40(1), 35-38.

703 Mitrovica, J.X., Beaumont, C., and Jarvis, G.T. (1989), Tilting of continental interiors by the
704 dynamical effects of subduction, *Tectonics*, 8, 1079-1094.

705 Molenaar, C.M., and Rice, D.D. (1988), Cretaceous rocks of the Western Interior basin, in Sloss,
706 L.L., ed., *Sedimentary cover– North America Craton: U.S.*, Boulder, CO, Geol. Soc. Am.,
707 *Geology of North America*, v. D-2, 77-82.

708 Nakanishi, M., Sager, W.W., and Klaus, A. (1999), Magnetic lineations within Shatsky Rise,
 709 northwest Pacific Ocean: Implications for hot spot-triple junction interaction and oceanic
 710 plateau formation *J. Geophys. Res.*, *104*, B4, 7539-7556.

711 Patino-Douce, A., Humphreys, E.D., Johnston, A.D., 1990, Anatexis and metamorphism in
 712 tectonically thickened continental crust exemplified by the Sevier Hinterland, western North
 713 America, *Earth Planet. Sci. Lett.*, *97*, 290-315.

714 Peterson, J.A. (1988), Phanerozoic stratigraphy of the northern Rocky Mountains, *in* Sloss, L.L.,
 715 ed., Sedimentary cover—North America Craton: U.S., Boulder, CO, Geol. Soc. Am.,
 716 Geology of North America, v. D-2, 83-107.

717 Reed, J.C., Wheeler, J.O., and Tucholke, B.E., *compilers* (2005), Geologic Map of North
 718 America, Decade of North America Geology, Continental scale map 001, Boulder,
 719 Geological Society of America, scale 1:5,000,000.

720 Roberts, L.N.R., and Kirschbaum, M.A. (1995), Paleogeography of the Late Cretaceous of the
 721 Western Interior of middle North America—Coal distribution and sediment accumulation,
 722 *U.S. Geological Survey Professional Paper 1561*, 155 p.

723 Sager, W.W., Zhang, J., Korenaga, J., Sano, T., Koppers, A.A., Widdowson, M., and Mahoney,
 724 J. J. (2013), An immense shield volcano within the Shatsky Rise oceanic plateau, northwest
 725 Pacific Ocean. *Nature Geoscience*, *6*(11), 976-981

726 Saleeby, J. (2003), Segmentation of the Laramide slab—Evidence from the southern Sierra
 727 Nevada region, *Geol. Soc. Am. Bull.*, *115*, 655-668.

728 Schmandt, B., Dueker, K., Humphreys, E.D., Hansen, S. (2012), Hot mantle upwelling across the
 729 660 beneath Yellowstone, *Earth Planet. Sci. Lett.*, *331-2*, 224-236.

730 Schmandt, B., Humphreys, E.D. (2010), Complex subduction and small-scale convection

731 revealed by body-wave tomography of the western United States upper mantle, *Earth Planet.*
732 *Sci. Lett.*, 297, 435-445.

733 Schultz, D.J., Coppersmith, H.G., Harte, B., and Pizzolato, L-A. (2008), Mineral inclusions in
734 diamonds from the Kelsey Lake Mine, Colorado, USA: Depleted Archean mantle beneath the
735 Proterozoic Yavapai province, *Geochimica et Cosmochimica Acta* 72, 1685-1695.

736 Schutt, D.L., and Leshner, C.E. (2006), Effects of melt depletion on the density and seismic
737 velocity of garnet and spinel lherzolite., *J. Geophys. Res.*, 111.

738 Scott, R.W., Oboh-Ikuenobe, F.E., Benson Jr., D.G., and Holbrook, J.M. (2009), Numerical age
739 calibration of the Albian/Cenomanian boundary, *Stratigraphy*, 6, 17.

740 Shen, W., Ritzwoller, M.H., and Schulte-Pelkum, V. (2013), A 3-D model of the crust and
741 uppermost mantle beneath the Central and Western US by joint inversion of receiver
742 functions and surface wave dispersion. *J. Geophys. Res.*, 118, 262-276. Model download at
743 ciei.colorado.edu/Models (07/28/2015).

744 Sloss, L.L. (1988), Tectonic evolution of the craton in the Phanerozoic time (p. 45), *in* Sloss,
745 L.L. ed., *Sedimentary Cover- North American Craton: U.S.: Conterminous U.S.: Boulder,*
746 *CO, Geological Society of America, Geology of North America*, v. D-2, 25-51.

747 Smith, D. (2010), Antigorite peridotite, metaserpentinite, and other inclusions within diatremes
748 on the Colorado Plateau, SW USA: Implications for the mantle wedge during low-angle
749 subduction, *J. Petrology*, 51, 1355-1379.

750 Smith, D., and Griffin, W.L. (2005), Garnetite xenoliths and mantle–water interactions below the
751 Colorado Plateau, southwestern United States, *J. Petrology*, 46(9), 1901-1924.

- Snyder, D. B., Hillier, M. J., Kjarsgaard, B. A., de Kemp, E. A., and Craven, J. A., 2014,
Lithospheric architecture of the Slave craton, northwest Canada, as determined from an
interdisciplinary 3-D model, *Geochem. Geophys. Geosyst.*, *15*, 1895-1910.
- Spasojevic, S., Liu, L., and Gurnis, M. (2009), Adjoint models of mantle convection with
seismic, plate motion, and stratigraphic constraints: North America since the Late
Cretaceous, *Geochem. Geophys. Geosyst.*, *10*, Q05W02, doi:10.1029/2008GC002345.
- Spencer, J.E. (1996), Uplift of the Colorado Plateau due to lithosphere attenuation during
Laramide low-angle subduction, *J. Geophys. Res.*, *101*, 13,595-13,609.
- Stixrude, L., and Lithgow-Bertelloni, C. (2011), Thermodynamics of mantle minerals-II. Phase
equilibria, *Geophys. J. Int.*, *184*, 1180-1213.
- Usui, T., Nakamura, E., Kobayashi, K., Maruyama, S., and Helmstaedt, H. (2003), Fate of the
subducted Farallon plate inferred from eclogite xenoliths in the Colorado
Plateau, *Geology*, *31*, 589-592.
- van Hunen, J., van den Berg, A.P., and Vlaar, N.J. (2004), Various mechanisms to induce
present-day shallow flat subduction and implications for the younger Earth: a numerical
parameter study, *Physics Earth Planet. Interiors*, *146*, 179-194.
- Whitmeyer, S.J., and Karlstrom, K.E. (2007), Tectonic model for the Proterozoic growth of
North America, *Geosphere*, *3*, 220-259.
- Yeck, W.L., Sheehan, A.F., Anderson, M.L., Erslev, E.A., Miller, K.C., and Siddoway, C. S.
(2014), Structure of the Bighorn Mountain region, Wyoming, from teleseismic receiver
function analysis: Implications for the kinematics of Laramide shortening, *J. Geophys. Res.*,
119, 7028-7042.

Figure Captions

776

777 **Fig. 1.** Upper mantle seismic structure. **(a)** Depth slices. Top row shows Sv seismic perturbations
778 from the Rayleigh-wave tomography of Shen et al. (2013), interpolated onto the model nodes of
779 the teleseismic P-wave tomography model of Schmandt and Humphreys, 2010a (also shown is
780 the interpolated result at 75 km depth). The bottom two rows show the P-wave model at node
781 depths in the upper mantle to depths of 400 km. Note that the two models are shown the same
782 depths above 200 km.

783 **(b)** Cross sections. These three cross sections cross in NE Wyoming near the deepest part of the
784 high-velocity upper-mantle structure. The dashed line at 120-km depth separates the S-wave
785 model from the deeper P-wave model. The narrow panels show the S-wave model to depths of
786 160 km. The approximate Moho depth is shown. Moho depth variations show up as lateral
787 velocity variations near this depth.

788 We argue that the prominent high-velocity structure is a deep volume of relatively cool basalt-
789 depleted mantle within the NE part of the Shatsky conjugate, created beneath the conjugate
790 Tamu volcano.

791

792 **Fig. 2.** Xenolith sampling of the deep lithosphere.

793 **(a)** Map showing the sites of the two kimberlites that sampled to depths of ~200 km. Dashed
794 lines show the approximate boundaries of cratons, with the Wyoming craton nearest the center.
795 The Colorado (CO) site erupted in the Devonian just south of the Proterozoic-Archean suture.
796 The Montana (MT) site erupted in Archean crust during the Eocene, but lies very near the
797 Proterozoic Great Falls Tectonic Zone (GFTZ). Background image is the depth slice at 160 km
798 from the P-wave model shown in Fig. 1.

(b) Temperature-depth plot of xenoliths from the two sites shown in (a). Figure after Eggler et al. (1988). The Devonian Colorado xenoliths (green) define a cool geotherm to ~200 km depth, and the deeper population sampled depleted Archean mantle. The Eocene Montana xenoliths (orange) have two distinct populations. A depleted Archean “low temperature” population (blue stripes) lies on a cool geotherm (between 40-45 mW/m²), typical of cratons around the world (Artemiva, 2009) and samples to depths of ~140 km. The extension of this geotherm to mantle solidus temperatures (pink, from Artemiva, 2009) defines a pre-Laramide craton of 170-230 km thickness. A depleted “high temperature” population (red stripes) lies on a warm geotherm (> 45 mW/m²) and samples depths of ~160-200 km. Unlike the other xenoliths plotted on this figure, these xenoliths have an age younger than the Paleozoic. Apparently the cool, ~200-km thick craton was truncated at 140-160 km depth, with the deeper mantle being replaced by relatively young and depleted mantle that was created at relatively high temperature.

Fig. 3. Elevation contributions since 125 Ma (i.e., since the early Sevier orogeny). All significant contributions have been included except the changes in uppermost mantle density and lower crustal hydration, and hence the unpredicted elevation is an estimate for the changes in upper mantle buoyancy plus lower crustal buoyancy due to hydration. All units are in km. All fields have been smoothed with a low-pass Gaussian filter that reduces signal amplitude by 50% at a wavelength of 150 km. See section 3.3 for details.

(a) Actual elevation. The “Y” indicates the location of Yellowstone.

(b) Dynamic topography created by density structure deeper than 200 km. The result is an inconsequential 50-70 m of subsidence.

(c) Initial elevation, taken from the estimated elevation at 125 Ma.

(d) Elevation created by post-125 Ma sediment.

(e) Elevation in (d) plus elevation effects (reduction) created by erosion since 125 Ma.

(f) Elevation created by crustal thickening 40-80 Ma (mostly of Laramide origin).

(g) Elevation created by post-Laramide crustal heating (from Levandowski et al., 2014).

(h) Predicted elevation.

(i) Elevation anomaly (actual elevation – predicted elevation). The circles show the elevation anomaly calculated by Spencer (1996), and the squares show the uplift predicted by Jones et al. (2015) resulting from lower crustal hydration. Plausibly, much of the Wyoming elevation anomaly in (i) is caused by lower crustal hydration, which would suggest that Wyoming upper mantle currently is as buoyant as it was prior to the Sevier orogeny.

Fig. 4. Estimated shape and path of the Shatsky conjugate beneath North America. Map is modified after Liu et al. (2010). The base map is a palinspastic reconstruction for ~65 Ma; black dots represent outcrops of the Pelona-type schist. See the section 4 and the Supplement for additional details. The circled grey numbers show the time (in m.y.) when the main volcanic construction (conjugate Tamu massif) was at the indicated location, from Liu et al. (2010), with the grey line indicating their path. We modify this path (to the line with colored age numbers) so as to (i) match the geologic record that we associate with the passage of the plateau – especially its passage beneath the greatly disrupted southern California crust (indicated with the two short black lines) – and (ii) ending with the Shatsky conjugate being co-located with the major high-velocity upper-mantle seismic structure and with the conjugate Tamu massif located where we image the major deep high-velocity structure shown in the inset. For each time we show the contours for crust of 17 and 27 km inferred thickness, and at 65 Ma we also include the contour

for 10 km inferred thickness. The inferred location of the Shatsky conjugate at the time (75 Ma) it was adjacent to the Colorado Mineral Belt (CMB) is shown with a heavy outline. We have the Shatsky conjugate stopping beneath North America at ~65 Ma. The inset shows the tomography image at 195 km depth, from Fig. 1, and the outline of inferred Shatsky conjugate at 65 Ma.

Fig. 5. Inferred history of Shatsky conjugate ocean plateau subduction beneath North America. **(a)** The inferred path of the ocean plateau beneath North America. Ages (in Ma) are located at the Tamu massif. See Fig. 4 for additional description. **(b)** Cross section along the trajectory of the ocean plateau at 75 Ma, viewed looking toward the SE. Farallon ocean lithosphere is shown in shades of green. Thermal lithosphere, represented with a thickness of 100 km, is shown in light green. Depleted mantle is represented with dark green, where its thickness simply represents the amount of basaltic melt that has been removed during plateau construction. The “T” indicates the Tamu part of the plateau. Ocean crust is shown in black and red, where red indicates the presence of eclogite. North America lithosphere is shown in shades of brown, with the crust shown in light brown. Dashed line shows eventual location of the North American base, which is eroded to this depth during the Sevier/Laramide orogeny. CP is Colorado Plateau. At 75 Ma the Colorado Mineral belt (CMB) and the Laramide orogeny both initiate. We suggest that the eclogized crust of the ocean plateau sinks into the Earth’s interior through a tear in the Farallon slab located beneath the CMB, contributing to regional uplift and initiating CMB magmatism. **(c)** Cross section at 65 Ma. The Shatsky conjugate has lost most of its crust, and it accretes to North America beneath Wyoming and the Colorado Plateau near this time.

Fig. 1a

Fig. 2

Temperature ($^{\circ}\text{C}$)

400 800 1200 1600

(b)

Fig. 3

Fig 4

Fig 5