Crustal structure of the Ruby Mountains metamorphic core complex, Nevada, from passive seismic imaging

Mairi M. Litherland and Simon L. Klemperer
Stanford University, Department of Geophysics, 397 Panama Mall, Stanford, California 94305, USA

ABSTRACT

We use data from a temporary passive seismic array to illuminate the deep structure of the Ruby Mountains metamorphic core complex (RMCC). Despite decades of geologic mapping and geophysical exploration, the relative importance of lateral crustal flow, diapirism, and brittle faulting in the formation of the RMCC has remained unclear. Our Ruby Mountains Seismic Experiment (RMSE) utilized 50 passive broadband stations from 2010 to 2012 spaced at 5–10 km along three ~100-km-long intersecting profiles as part of the EarthScope Flexible Array program. Common conversion point stacks of our P-wave receiver functions show a fairly flat Moho at 32 ± 2 km depth throughout most of the study area but reaching 40 km in a narrow, north-south crustal welt 20–50 km west of the exposed RMCC. Our shear-wave splitting analysis shows that fast directions of polarization rotate clockwise from west to east across our study area, broadly matching regional studies and models that placed the anisotropy below the lithosphere. However, because our north-south crustal welt coincides with W-E polarizations, the observed splitting may include a component of crustal anisotropy. We integrate our observations with older magnetotelluric and seismic refraction and reflection data to support a model of asymmetric crustal flow during formation of the RMCC at the edge of a preexisting orogenic plateau.

INTRODUCTION

Motivation: The Metamorphic Core Complex “Controversy”

Metamorphic core complexes (MCCs) are exposures of deep, ductilily deformed crust in both oceanic and continental settings around the world; the rise of these complexes to the surface is linked to extensional processes (e.g., Whitney et al., 2013; Platt et al., 2015). Three distinguishing features of an MCC are (1) a domed metamorphic assemblage in the footwall beneath (2) an unmetamorphosed hanging-wall assemblage, separated by (3) a shallowly dipping mylonitic shear zone (Coney, 1980). Since the widespread recognition of MCCs, workers have proposed numerous mechanisms for the formation of these structures; these mechanisms include gravitational collapse of over-thickened crust (Coney and Harms, 1984), crustal weakening resulting from regional magmatism (Armstrong and Ward, 1991), and a combination of elevated temperatures and thick crust (Buck, 1991). A related mechanism that has been applied to explain the development of the RMCC involves the diapiric rise of a thermally buoyant “gneiss dome” associated with regional magmatism and/or crustal thickening (Whitney et al., 2004; Rey et al., 2009), at its simplest dominated by symmetric vertical crustal flow (Fig. 1A). A contrasting model, which has also been applied to the RMCC, is dominated by large-scale horizontal extension and vertical thinning due to asymmetric uplift along originally low-angle extensional faults rooted in the lower crust or even below the Moho (Wernicke, 1981; Howard, 2003; Fig. 1B). Still other explanations—for example, “rolling-hinge” models (Buck, 1988; Wernicke and Axen, 1988; Lavier et al., 1999)—combine dominantly horizontal and dominantly vertical motions at different levels of the crust (Fig. 1C).

The relative importance of asymmetric, subhorizontal material movement versus symmetric vertical flow is unclear from geologic mapping. Surface observations alone cannot distinguish which processes occur in the deep crust. Application of different models to the same surface exposures leads to substantially different estimates of local and regional extension (Howard, 2003), presenting uncertainty in pre-extensional crustal thickness and surface elevation (e.g., Konstantinou et al., 2012). All classes of models can yield a smooth or flat Moho, presuming that lower-crustal flow and/or magmatic additions to the lower crust would remove any differential crustal thickness resulting from laterally variable degrees of crustal thinning (cf. Coney, 1987). However, some recent numerical experiments have explored the role of pre-existing Moho topography and the possibility of preservation of the topography during and after core-complex formation (Rey et al., 2010; Whitney et al., 2013).

The Ruby Mountains metamorphic core complex (RMCC) and its northeastern continuation, the East Humbolt Range, are located in northeastern Nevada, within the northern part of the Basin and Range Province of the western United States (Fig. 2). The RMCC has long been studied as an outstanding example of the metamorphic core complexes (MCCs) that have played a major role in Basin and Range extension. The RMCC lies between the Albion-Raft River–Grouse Creek MCC to the north and the Snake Range MCC to the south, which together comprise the central part of a band of MCCs spanning the Cordillera from Canada to Mexico (Coney, 1980; Armstrong, 1982; Sullivan and Snake, 2007; Fig. 2C). Despite extensive geological and geophysical investigation, the mechanisms for formation of the RMCC and exhumation of rocks from ~30 km depth are still debated, including the relationship between low-angle normal
faulting at the surface and deeper-crustal ductile flow (e.g., Colgan et al., 2010; Henry, et al., 2011; Konstantinou et al., 2012).

Here we present common conversion point (CCP) images of the RMCC derived from our passive seismic deployment; these images suggest that asymmetric deep-crustal flow was an integral part of core-complex formation.

Geologic Background

The Basin and Range Province extends from the Sierra Nevada on the west to the Colorado Plateau on the east, and from the Snake River Plain on the north into Mexico to the south. In our study area (Fig. 2), Precambrian basement includes the Archean Wyoming Province in the northeast and the Paleoproterozoic Mojave Province in the southeast; earlier studies placed the transition between these two regions near the RMCC (Rodriguez and Williams, 2008; Premo et al., 2010), but others suggest the boundary lies to the north of the RMCC (Nelson et al., 2011). The basin transitions into juvenile Phanerozoic crust across the 87/86Sr = 0.7060 line conventionally taken as the Neoproterozoic continent-ocean boundary (Wooden et al., 1998). The entire region experienced passive-margin sedimentation until the Late Devonian, when the Antler orogeny emplaced the Roberts Mountain allochthon to the west of the Ruby Mountains (e.g., Dickinson, 2006). The Cretaceous Sevier orogeny thicken a large area of crust known as the Sevier hinterland near and to the west of the RMCC (Coney and Harms, 1984; Hodges et al., 1992; Long, 2012; Fig. 3A).

Major rock uplift of the RMCC began in the Late Cretaceous and continued through the Oligocene (McGrew et al., 2000; Figs. 3A and 3B). Regional exhumation by erosion was likely limited to only ~1.5–3.5 km in our study area (Long, 2012), and the lack of evidence for significant surface-breaking extension before Middle Miocene time suggests that early rock uplift was localized within the lower plate of the developing RMCC (Colgan and Henry, 2009; Colgan et al., 2010; Lund Snee et al., 2016). The exhumed core of the northern and central RMCC was intruded by Cenozoic plutons between 42 and 29 Ma (Wright and Snoke, 1993; MacCready et al., 1997; Howard et al., 2011; Fig. 3C). This period of mantle-derived magmatism and crustal melting coincided with a southward sweep of volcanism across the region (Armstrong and Ward, 1991; Brooks et al., 1995). The largest of the plutons, the Harrison Pass Pluton (HPP in Fig. 3C), was intruded at 36 Ma (Barne et al., 2001) at the boundary between the high-grade metamorphic rocks of the northern RMCC (NRM/EH in Fig. 3) and the lower-grade southern RMCC (SRM in Fig. 3).
Figure 2. Map of Ruby Mountains Seismic Experiment (RMSE); red triangles indicate RMSE station locations; dark-blue triangles are Earthscope Transportable Array stations. (A) Topographic map with locations of previous geophysical projects. Dashed lines—Consortium for Continental Reflection Profiling (COCORP) and Wyoming and Arizona active seismic profiles. Dotted lines—magnetotelluric (MT) transects. Solid line—Highway I-80. (B) Simplified geologic map of Ruby Mountains metamorphic core complex (RMCC) and surrounding area. A1–A1’, A2–A2’, B–B’, and C–C’ (further defined in Fig. 8B) mark the common-conversion point profiles shown in Fig. 7 (C) Index map showing regional geology: SR—Snake Range metamorphic core complex; ARG—Albion–Raft River–Grouse Creek metamorphic core complex; RM—Ruby Mountains metamorphic core complex; OR—Oregon; ID—Idaho; NV—Nevada; UT—Utah; CA—California. Parts (B) and (C) after Colgan et al. (2010).
The lower plate of the RMCC is separated from the unmetamorphosed Paleozaic–Mesozoic stratigraphic section of the upper plate by a ~1-km-thick mylonitic shear zone that lies beneath a now ~20° west-dipping detachment fault (e.g., Snoke, 1980; MacCready et al., 1997). The detachment strikes ~N-S along the western flank of the RMCC, notably cutting both the high-grade northern RMCC and the low-grade southern RMCC, as well as the intervening HPP (e.g., Colgan et al., 2010). As is common with core-complex detachments, the RMCC mylonitic shear zone is overprinted by brittle deformation, with a total slip of 50–70 km (Howard, 2003). Extensional exhumation along the mylonitic shear zone may have begun by end-Eocene (Fig. 3C) but is thought to have been most active in Late Oligocene and Early Miocene, from 29 to 20 Ma (Wright and Snoke, 1993; MacCready et al., 1997; Howard et al., 2011). Brittle slip along this detachment occurred primarily in the Middle Miocene, 16–12 Ma, accommodating ~15–20 km of horizontal extension (Colgan et al., 2010, Fig. 3D). From the Late Miocene to the present, only modest extension has occurred along the high-angle normal faults bounding the Ruby Range, including the steeply east-dipping Ruby Valley fault that bounds the east side of the Ruby Mountains horst and contributes to its modern topographic expression (Fig. 3E).

The multi-stage history and 3D structure of the RMCC complicate interpretations, including the often-attempted construction of 2D cross sections (e.g., Howard, 2003; Colgan et al., 2010). The northern RMCC experienced its highest-grade metamorphism, up to ~9 kb and ~800 °C, during Mesozoic crustal shortening, but these deeply seated rocks were partially exhumed as early as the Late Cretaceous (Hodges et al., 1992; McGrew and Snee, 1994; McGrew et al., 2000; Hallett and Spear, 2015). Surface-breaking faulting may have been minor prior to large-magnitude fault motion in the Miocene (Lund Snee et al., 2016). Hence, in the RMCC, there appears to be a large time gap between much of the lower-crustal flow recorded in rocks now at the surface and surface-breaking normal faults such as the detachment that ultimately exposed the Ruby Mountains in its footwall. The Ruby detachment fault runs along the western flank of the entire RMCC (Snoke, 1980), but whereas the northern RMCC has been exhumed from ~30 km depth, the southern RMCC was never buried beyond stratigraphic depths (<10 km) (e.g., Colgan et al., 2010).

Previous Geophysical Studies

Numerous geophysical studies precede our work in this area (Fig. 2A), but no previous seismic study has imaged the core of the RMCC. Seismic-reflection profiles include a 1983 west-east Consortium for Continental Reflection Profiling (COCORP) crustal-scale transect immediately south of the Ruby Mountains across Overland Pass (Hauser et al., 1987; Figs. 1A and 4D). The COCORP study imaged a mostly smooth Moho at ~30 km depth (Klemperer et al., 1986) beneath strong subhorizontal layering in the deep crust that was interpreted as being characteristic of highly extended terranes (Allmendinger et al., 1987; Gans, 1987). A University of Wyoming 15-km-long reflection...
Table. Metadata for Ruby Mountains Seismic Experiment and Transportable Array stations. Please visit http://doi.org/10.1130/GES01472.S1 or the full-text article on www.gsapubs.org to view the Supplemental Table.

Receiver Functions

We calculated P-wave receiver functions (RFs) at each station using an iterative time-domain deconvolution technique (Ligorria and Ammon, 1999). The P-wave RFs model the Earth response to nearly vertically arriving P-waves, allowing us to image wave-speed discontinuities within the Earth. We selected all events with magnitude >5.5 at epicentral distances of 30°–90° that were recorded on any of our stations or the four nearest TA stations (O11, O12, N11, and N12). After removing the mean and applying a 5% Hann taper, with a bandpass filter of 0.02–2 Hz, we rotated the seismograms from the north-east-vertical coordinate system to the radial-tangential-vertical coordinate system for our RF calculation. We used a relatively high-frequency Gaussian filter width of 5 to permit imaging of crustal features potentially as thin as the shortest wavelengths, ~1 km (Ligorria and Ammon, 1999).

In the iterative time-domain deconvolution technique, a candidate RF is forward modeled in each iteration and convolved with the actual vertical-component data, yielding an estimate of the radial horizontal data. This is compared to the actual radial horizontal data and iterated until the change in error is less than 1.0 x 10^-4 or until 200 iterations are completed. Events with less than 60% fit were automatically discarded. Additionally, all RFs were visually inspected, and events with excessive noise or ringing were discarded. Receiver functions can be found in the Supplemental Data.

Most of our stations located on bedrock had at least 100 RFs with reasonable misfit (e.g., stations A04 and A06, Figs. 5 and 6; Supplemental Table [see footnote 1]), whereas the worst station (B13) within the Neogene sedimentary basins had just eight acceptable RFs. The lack of interpretable RFs is likely due to interference from multiple reverberations within the basin, violating our processing assumption that no multi-pathing is present in our seismograms.

For the basin stations, we therefore deconvolved the recorded horizontal component from a matched vertical component taken from the same event at a station recorded on bedrock. Because we used earthquakes ~3000–10,000 km distant from our array, which has a lateral dimension of only ~150 km, the earthquake source-time function is effectively uniform across our array. This procedure successfully minimized basin reverberations at some stations (e.g., C05 in Fig. 5B) but not all. We tried, but did not find useful, either a notch filter to remove the dominant basin reverberation frequency (following Yu et al., 2015) or stacking Ps, PpPs, and PpSs multiples (e.g., Wilson and Ast, 2005).
Figure 4. Previous geophysical data superimposed on our new interpretations. Colored images are resistivity models of Wannamaker and Doerner (2002) from magnetotelluric (MT) profiles (A) through Secret Pass ~30 km NE of our RMSE profile B–B′; (C) through Harrison Pass, 0–20 km distant from Ruby Mountains Seismic Experiment (RMSE) C–C′; and (D) past Bald Mountain, 10–25 km south of Consortium for Continental Reflection Profiling (COCORP) line 4 (Hauser et al., 1987). (B) Generalized wide-angle seismic wave speed measured along Ruby Valley (Fig. 2A) immediately east of the Ruby Mountains metamorphic core complex (RMCC), with broad gray lines marking depth intervals of rapidly increasing wave speed (Stoessel and Smithson, 1998). Magnetotelluric (MT) and seismic data in (A), (B), and (C) are overlain by receiver-function converters (solid black lines) labeled as in Figure 7 from this study, from RMSE profiles B–B′, A2–A2′, and C–C′, respectively. MT data in (D) are overlain by cartoon of crustal reflectivity along COCORP line 4; coarse stipple is Phanerozoic section; gray shading represents pervasive lower-crustal layering (after Hauser et al., 1987). RMSE profiles B–B′, A2–A2′, and C–C′ are placed so that the intersections of B–B′ with A2–A2′ and of C–C′ with A2–A2′ align vertically (dashed lines). In (B), the red dashed line represents approximate modern depth of ~6 kb metamorphism, from outcrop in northern RMCC to >10 km below surface in southern RMCC (~3 kb at surface).
Figure 5. Receiver-function gathers binned by slowness $p$ (proxy for earthquake distance $\Delta$) for (A) station A04 (installed on metamorphic bedrock), one of our best stations, and (B) station C05, typical of our stations installed in Quaternary deposits over Tertiary formations west and east of the Ruby Mountains metamorphic core complex. Note different scale in (A) and (B) for trace count, the number of traces stacked in each slowness bin. $Ps$ is the direct conversion of the incident $P$ wave front at the Moho to an $S$ wave in the crust. $PpPs$ and $PpSs$ arrivals are “multiples” with additional $P$ or $S$ ray paths in the crust. Dashed lines in (A) and (B) are theoretical arrival times for a 30-km-thick crust.

Figure 6. Receiver-function gathers binned by back azimuth for (A) station A04 (Fig. 5) and (B) station A06, ~20 km SW of A04, also on the western flank of the Ruby Mountains metamorphic core complex. Dashed lines—interpreted primary conversions; $Y$—mid-crustal converter; and $M$—Moho (also labeled $Ps$) and multiples $PpPs$ and $PpSs$. Note $Y$ is only visible from southerly azimuths. Both $Y$ and $Ps$ (and its multiples) show travel time varying with back azimuth, evidencing lateral variations and Moho dip. Vertical line at back azimuth 225°/45° separates arrivals from the NW that we stacked to produce Ruby Mountains Seismic Experiment profile $A1-A1'$ and arrivals from the SE that we stacked to create profile $A2-A2'$. 
We used CCP stacking (Dueker and Sheehan, 1997) to sum receiver functions along their respective back-azimuthal paths. Our CCP stacks (Fig. 7) were calculated in 3D, then projected onto 2D profiles along each of our three lines (A, B, and C; Fig. 8B) as well as individually at D stations and TA stations that did not fall along our lines. We used a block size of 2 km horizontally and 1 km vertically. Errors were calculated by performing 200 bootstrap resamples; bins with a standard error greater than 0.2 were discarded. Projection of data orthogonal to our lines of section aggregated image rays with piercing points spanning ±20 km at Moho depths and ±15 km in the mid-crust. For the time-to-depth conversion, we use a 1D model modified from IASP91 (Kennett and Engdahl, 1991) to be typical of the wave-speed structure of the RMCC (Satarugus and Johnson, 1998; Stoerzel and Smithson, 1998; Fig. 9A). Our time-to-depth conversions were performed with a 1D wave-speed model because modest changes in wave speed do not have a noticeable effect on CCP migration, and while changes in Vp/Vs ratio can have a larger effect, local changes in Vp/Vs are even less well known than the wave-speed structure (see Supplemental Fig. S1).

Because of the location of profile A on the western flank of the RMCC, and because our back-azimuthal coverage is dominated by earthquakes from the NW and the SE (Fig. 8A), events from the NW sample a different lower crust than events from the SE (Fig. 6). To better image the local changes in crustal structure along the RMCC, we therefore processed line A separately as line A1, using only events from the NW, and line A2, using only events from the SE. Although this results in fewer RFs per station, it allows us to see differences between the crust directly beneath the RMCC and that immediately west of it. Finally, the four CCP profiles are presented in Figure 7 with horizontal smoothing applied over ~6 km. The locations of earthquakes we used are shown in Figure 8A, and their piercing points are shown in Figure 8B.

To provide additional constraints on Moho depth as well as obtain an estimate of Vp/Vs ratio, we performed H-k stacking for each station. H-k stacking takes advantage of the different moveouts of Ps and the later phases present in receiver functions, PPpSs and PPsSs + PsSs (Zhu and Kanamori, 2000). A grid search is made in H (Moho depth)-k (Vp/Vs ratio) space for the maximum amplitude of the stack of the phases along the travel-time curve predicted for each. We use a ratio of 7:2:2 for the three arrivals when stacking. However, the H-k stacking method routinely produces several local maxima; so we cannot always be certain which arrival to pick. H-k stacking results for all our stations are shown in Supplemental Figure S2 (see footnote 3).

SKS Splitting

Shear waves traveling through an anisotropic material split into orthogonal fast and slow components corresponding to the fast and slow axes of the anisotropic fabric. Polarization directions of core-refracted teleseisms (SKS phase) are commonly interpreted as arising from lattice-preferred orientations of olivine in the upper mantle, although highly anisotropic schists in the crust are sometimes a significant factor in shear-wave splitting (e.g., Godfrey et al., 2000). We measured shear-wave splitting of the core-refracted SKS phase from all events of M ≥ 5.5 at epicentral distances of 90°–130° using SplitLab software (Wüstefeld et al., 2008) to find the best-fitting single-layer model with horizontal anisotropy beneath each of our stations. We also searched for best-fitting two-layer models (Silver and Savage, 1994), but we were unable to find any that were a statistical improvement over our one-layer models. This may in part be due to the relatively poor back-azimuthal coverage available from only two years recording; but it may also indicate that the lower crust and upper mantle were deformed along the same azimuth even if perhaps at different times.

Prior to the splitting calculation, events were individually windowed and bandpass filtered from 0.02 Hz to 0.2 Hz, and events without a clear SKS arrival were discarded. We calculated the fast polarization direction φ and time delay τ, using both the minimum-energy method and the rotation-correlation method. In the minimum-energy method, we searched for the rotation of the seismogram that minimizes energy on the transverse component (Silver and Chan, 1991). In the rotation-correlation method, the seismogram is rotated to maximize the correlation of the two components (Wüstefeld and Bokelmann, 2007). Results were assessed based on whether they produced a linear particle motion, the magnitude of the computed error, and whether the results from the two methods were comparable; each result was assigned a value of “good,” “fair,” or “poor” (Fig. S3 [see footnote 2]). Poor results were discarded. We then stacked the results of the minimum-energy method at each station, with good results given twice the weight of fair results, to produce a single measurement at each station (Fig. 10 and Supplemental Table [see footnote 1]).

RESULTS

Receiver Functions

We identify and label key features on our CCP receiver-function images, requiring consistency with existing maps and cross sections (Howard, 2003; Colgan et al., 2010). From known surface locations of the main range-bounding faults of the Ruby Mountains, we follow the amplitude and polarity trends (i.e., color variation) in the CCP image down to the west and east on profiles B–B' and C–C' (F1 and F2 in Fig. 7).

Next we identify intra-crustal converters with possible correlations across multiple profiles. In the mid-crust east of the Ruby Mountains, a converter X in Fig. 7 dips ~20° east, from 20 to 30 km depth on B–B' and from 15 to 25 km on C–C'. Although some stations in this region have few high-quality RFs due to reverberations within Ruby Basin, feature X is particularly clear on C–C'. In the northern portion of our survey, profiles A1 and B show a subhorizontal mid-crustal converter at ~19 km depth (Y in Fig. 7). Converter Y dips ~10° SSW on A2 [directly beneath the RMCC] from 10 km at the northern end of the profile to 20 km depth midway along the profile. Converter Y ends abruptly midway along profile A1, and instead, we label a separate weaker converter, dipping ~20° SSW in the southern Ruby Mountains, as Z. We do not see converter Y on line C–C' across the southern Ruby Mountains.
Figure 7. Common conversion point (CCP) stacks along Ruby Mountains Seismic Experiment (RMSE) lines B, A1, A2, and C with no vertical exaggeration; all depths are measured below station elevations. Colored lettering (A, B, C, etc.) correlates with colors of profiles in Figure 8B. Profiles B and C are viewed from the southwest, and A1 and A2 are viewed from the northwest. Individual D stations and Transportable Array (TA) stations located in Fig. 2A are viewed from the SW (stacked along an NW-SE line that parallels the piercing-point distribution) and are shown beside the line closest to them. Smoothed topography along CCP projection lines is shown above each profile (vertical exaggeration x15); area beneath exposed Ruby Mountains metamorphic core complex is shaded. Red and blue triangles indicate station locations and elevations of RMSE and TA stations, respectively. Vertical dashed lines show intersections of profiles (see Figs. 4 and 8B). Horizontal dotted fiducial lines are placed at 15 and 30 km. Colors in CCP images represent $P_s/P$, with red (blue) colors indicating shear-impedance (product of $P$-wave speed and density below, $S$-wave speed and density above) increasing (decreasing) downwards. Solid black lines are converters discussed in the text. $M$ is the Moho; $m_1$, $m_2$, $m_2a$, and $m_2b$ are segments where the Moho converter is discontinuous. $F_1$ and $F_2$ are normal faults interpreted from surface geology and breaks in the CCP images. $X$, $Y$, and $Z$ are mid-crustal converters.
Figure 8. (A) Locations of earthquakes used to create the common conversion point (CCP) images, also showing separation of sources for A1 and A2 profiles. (B) Area of Figure 2, with outline of Ruby Mountains metamorphic core complex (RMCC) and East Humbolt Range (gray shade) for reference. White triangles show station locations; dots are piercing points of receiver functions at 30 km depth; broad swaths with colors keyed to Figure 6 show region of receiver functions stacked to produce each CCP profile in Figure 7: blue—B, red—A1, yellow—A2, and green—C. A1 and A2 represent events recorded on the A-line with back azimuths from the NW and SE, respectively. Line A was located on the western flank of the RMCC; so A1 images structure beneath the deepest part of Huntington and Lamoille valleys and profile A2 best shows structure directly beneath the RMCC. N-S gray line shows region of proposed crustal welt m1-m2 from Figure 7.

Figure 9. (A) Synthetic receiver function (RF) with 6.1 km/s upper crust to 15 km depth, 6.1–6.6 km/s gradient lower crust, and 7.8 km/s mantle, with Moho at 30 km and Vp/Vs = 1.76 everywhere. Bold numbers in lower panels are P wave speed in km/s in each layer; numbers in parentheses indicate Vp/Vs. (B) Same as (A) except with 1-km-thick Quaternary basin at 2.5 km/s with Vp/Vs = 1.96 replacing crust at surface. (C) Same as (B) except with 3 km 3.5 km/s Neogene basin with Vp/Vs = 1.96 (with gradient at base) additionally replacing upper-crustal material. Dashed lines show Moho and crustal multiple delay times calculated for model (A).
The Moho, identified at ~30 km depth in many previous studies (Klemperer et al., 1986; Satarugsa and Johnson, 1998; Stoerzel and Smithson, 1998), is unmistakable on all our profiles. On A2, directly beneath the Ruby Mountains, a clear Moho signal (M on Fig. 7) extends across the entire line at depths ranging from ~30 km depth in the northern RMCC to ~34 km in the southern RMCC, with an apparent dip of ~10° SSW north of Harrison Pass (intersection of A–A′ with C–C′). All depths are given below surface; thus, Moho depths correspond to crustal thickness. On our B and C lines, the Moho converter is also very clear, and consistent with the image on line A: the Moho has an apparent dip of ~8° ESE along B–B′ and is subhorizontal on C–C′ beneath the Ruby Mountains. Minor differences between images at their intersections are likely due to cross-line projection of data up to ~10 km, as well as the ~6 km lateral smoothing applied independently to each profile in its in-line direction.

We see no change in Moho character or depth directly beneath the Ruby Mountains along profiles B and C, even though the region of high topography crossed by these profiles is significantly wider (~20 km) than the lateral smoothing of our CCP images (~6 km). East of the Ruby Mountains, the Moho appears continuous but with minor undulations. Although our images do not show any meaningful crustal root beneath the Ruby Mountains, comparison of profiles A1 and A2, recorded at the same stations but with different back azimuths, suggests the amplitude of the Moho conversion is generally larger beneath the basin surrounding the RMCC (A1–A1′) than beneath the Ruby Mountains (A2–A2′). If there is variation in the incidence angle of the events used to construct the CCP image, Moho dip and crustal anisotropy could also cause amplitude variation. However, we do not observe Moho dip on profiles B and C beneath the RMCC and do not observe variation in amplitude by back azimuth when looking at our B and C stations individually, as would be expected if there was regional crustal anisotropy. Additionally profiles A1 and A2 are both constructed using events from the full range of incidence angles (30°–90°; Fig. 8); therefore, we suspect the change in Moho-conversion amplitude has a geological origin.

In contrast to the region directly beneath the RMCC, on both the B and C lines, the Moho is complex west of the RMCC. On C–C′, the Moho converter deepens to ~40 km depth and then returns to the more typical ~30 km depth farther west, forming an ~20-km-wide apparent crustal welt or Moho trough (m1 and m2, Fig. 7). At a similar location on B–B′, we recognize the same west-dipping Moho segment m1; however, here there are two east-dipping segments, m2a and m2b. The similar geometry of m2a (B–B′) and m2 (C–C′) suggests that these converters are laterally equivalent.

In order to test whether the apparent crustal welt on the west side of the RMCC could simply be a consequence of imaging challenges arising from the Cenozoic basin fill underlying Lamoille (B–B′) and Huntington (C–C′) valleys, we forward modeled 1D receiver functions using RFTool (Helffrich, 2006). Our modeling parameters were based on simple wave-speed models based on the known subsurface geology (Satarugsa and Johnson, 2000; Colgan et al., 2010). These synthetics (Fig. 9) show that the Cenozoic basins are unlikely to...
create more than 2 km of apparent Moho relief, far smaller than the ~9 km relief shown in profiles B and C (Fig. 7). It would take implausibly large Vp/Vs ratio variations (1.76–1.86 in Fig. S1 [see footnote 2]) only above the proposed crustal welt, changes incompatible with significant basin effects (1.96 in shallow basins; Fig. 9) to produce the apparent Moho trough. While the Vp/Vs results from our H-k data do show variation (Fig. S2 [see footnote 2]), these changes are not systematically located above the crustal welt. In addition, industry seismic-reflection profiles show that Ruby Valley (east of RMCC) and Lamoille and Huntington valleys (west of RMCC) have similar thicknesses (~4 km) and west-east extents (~15 km) of Tertiary sedimentary rocks (Satarugsa and Johnson, 2000), yet neither profile B nor C shows comparably complex Moho geometry beneath Ruby Valley. Hence we believe that the imaged crustal welt represents a real geologic feature west of the RMCC that requires explanation.

Receiver functions acquired at an individual station can be stacked to form a CCP image; although due to the absence of crossing ray paths from different directions, such single-station stacks are not as reliable as densely instrumented profiles. Nonetheless, the CCP images formed at our individual stations D1–D6 and the closest TA stations N12, O11, and O12 show Moho images consistent with the closest (albeit ~30 km distant) portions of profiles A, B, and C [Fig. 7]. Station D4 shows two possible Moho converters and may be imaging the Moho trough seen on profiles B and C at a similar distance west of the RMCC. Because of the importance of the crustal welt to our interpretations, we also separately plot TA station N11, which is directly above the welt. Both CCP imaging and H-k stacking support the existence of the converter m2b and a deeper m2a-m1 converter in this location (Fig. S4 [see footnote 2]).

**SKS Splitting**

We compare our SKS splitting results to fast-polarization azimuths predicted by two mantle-flow models. Our SKS splitting results broadly agree with previous regional studies that show delay times ~1 s and polarization directions that rotate clockwise from west-east in the west of our field area to NW-SE in the east (Fig. 10A). This rotation has previously been noted as the NE part of a smooth circular or parabolic pattern in the orientation of regional fast polarization directions (Savage and Sheehan, 2000; Walker et al., 2005; Zandt and Humphreys, 2008; West et al., 2009).

In Figure 10B, we show our SKS results, with dot color representing azimuth and size representing splitting time, superimposed on the location of the crustal welt. Our greater density of stations speculatively suggests a possible discontinuity in splitting direction: stations A00–A05, B01–B06, C01–C06, D01, and D06 (mostly NW of the RMCC) show an average splitting direction of ~93° (blue colors in Fig. 10B), while all others (to the SE) have an average splitting direction of ~120° (red colors). Few stations show intermediate values (very light or white colors in Fig. 10B).

**DISCUSSION**

**Interpretation of Converters**

Our CCP images show a clear subhorizontal Moho at ~30–32 km depth throughout most of our field area, as well as intermittent mid-crustal converters. This is consistent with previous seismic studies that show a consistent Moho between 30 and 34 km depth (Klemperer et al., 1986; Satarugsa and Johnson, 1998; Stoerzel and Smithson, 1998; Fig. 4). In this context, the gentle SSE apparent dip of the Moho directly beneath the Ruby Mountains on our CCP images along lines B and A2 is surprising but could represent a SSE increase in Vp/Vs and/or lower wave speeds (Fig. S1 [see footnote 2]) rather than a change in crustal thickness. Unlike the ~10-km-thick crustal welt, these modest apparent thickness changes beneath the Ruby Moun-
tains (~4 km) are smaller than our ability to resolve trade-offs between thickness and velocity.

Both mid-crustal converters we image beneath the Ruby Mountains (Y and Z on A2–A2′, Fig. 7) dip to the south. The clearer of these features, Y, likely corresponds to a previously imaged, rapid wave-speed increase from 6.1 to 6.3 km/s to 6.6 km/s (Stoerzel and Smithson, 1998; Fig. 4B) and perhaps also with a prominent kilometer-thick band of reflectivity at 12 km depth beneath Secret Pass (Valasek et al., 1989; Fig. 4B). We also recognize the possible correlation between Y and the top of the conductive lower crust (Wannamaker and Doerner, 2002; Fig. 4A). The possibility that Y represents a downward increase in metamorphic grade in rocks of relatively uniform composition seems unlikely given the often gradational nature of such transitions and the small expected wave-speed change at any one isograd, including amphibolite to granulite (Hacker et al., 2003). Although the seismic wave speeds of ~6.6 km/s below Y are unlikely to only represent mafic lithologies, these wave speeds could correspond to a region of mafic intrusions into more felsic granulites. If converter Y represents the top of a zone of mafic intrusion into the crust, the volume of intrusion must be more marked in the northern Ruby Mountains, because Y is not imaged in the south part of A-A′ or on C-C′. At least the northern part of converter Y is shallower on A2 (along the axis of the Ruby range) than on A1 (Lamoille Valley) by ~5 km, so that our hypothesized zone of mafic intrusions is also thicker beneath the Ruby Mountains than west and east of the exposed RMCC. Previous seismic studies have been interpreted as showing moderate magmatic additions (Satarugsa and Johnson, 1998) or major intraplateing (Stoerzel and Smithson, 1998) during the era of core-complex exhumation, with the latter interpretation including up to 12–15 km of mafic rocks under the northern Ruby Mountains and up to 8–7 km under the southern Ruby Mountains (although these thicknesses could include preexist-
ing material). We interpret Y, and by inference the boundary between both seismic and magnetotelluric lower crust in this region, as the top of a zone of intensively intruded and modified Precambrian crust, significantly thicker beneath the northern RMCC. The southward dip of Y (and Z) is similar to the average metamorphic field gradient from north to south along the Ruby range.
Research Paper

Earthquake sources clearly shows both converters m1/m2a and m2b (Fig. S4 [see footnote 2]). The Moho converter is the most continuous, highest amplitude converter we image. Seismic amplitudes are often hard to interpret because of potentially large differences in station coupling and near-surface attenuation between stations. However, the difference in amplitude of the Moho Ps converter between profiles A1 and A2 most likely arises in the lower crust and/or upper mantle because the profiles were recorded at identical seismographs. The change in Moho amplitude, weaker beneath the Ruby Mountains, is consistent with a more mafic lower crust beneath the Ruby Mountains than beneath the adjacent basins, hence also consistent with a localized region of crustal underplating and intrusion beneath the exposed core complex.

Although the Moho is relatively flat beneath the Ruby Mountains and its flanking valleys, farther west we image complex Moho topography (Figs. 4 and 7, converters m1 and m2). This narrow (~20-km-wide) but thick (up to 10 km) welt on the base of the crust has not previously been recognized (e.g., Heimgartner and Louie, 2007; Gilbert, 2012); however, previous analyses may not have had the capacity to detect this structure. Only a single station of the Earthscope USArray data (N11, Fig. 2) lies above our interpreted welt, but it clearly shows both converters m1/m2a and m2b (Fig. S4 [see footnote 2]). However, simple 1D body-wave interpretations of Moho depth (e.g., Crotwell and Owens, 2005) are complicated by the presence of the three converters—m1, m2a, and m2b (Fig. 7, B–B’ and Fig. S4 [see footnote 2]) and also rely upon earthquake sources ≥30° distant with ray paths that pierce the Moho ≥10 km NW and SE of station N11, thereby failing to image the greatest crustal thickness. Nor would a deep and narrow welt be easily identifiable on gravity maps, because the modeled gravity anomaly would be ~100 km wide but only ~10 mGal. Intriguingly, the western end of the COCORP 4 reflection profile, close to the possible southwest extension of the crustal welt to RMSE station D4 (Figs. 1A and 7), is located over the deepest reported wide-angle reflection/refraction Moho and some of the deepest Moho reflections along the COCORP 40°N transect (Klemperer et al., 1986).

Based on our interpretations of CCP profiles B–B’ and C–C’, we can map the axis and lateral extent of the crustal welt (Figs. 8B and 10B). The axis of greatest Moho depth is aligned approximately north-south, seemingly distinct from the SSW-NNE trend of Neogene Basin and Range faults and their topographic expressions in the Ruby, Adobe, and Cortez ranges (Fig. 2). The axis of the welt lies beneath the Elko Hills and Cedar Ridge exposures of the upper Paleozoic miogeocline (Fig. 2; Colgan et al., 2010) and so does not represent a simple depression of the Moho beneath the Huntington and LaMoille Neogene depocenters. We suggest instead that this topographic welt is related to the genesis of the RMCC and is a key asymmetric element of the modern crustal configuration. In addition to the asymmetric Moho topography, the intra-crustal converter X that we only recognize on the east side of the Ruby Mountains may bear on asymmetric formation of the RMCC.

A key uncertainty in our interpretations (Fig. 11) is whether the crustal welt is derived from Moho topography existing prior to core-complex formation (Figs. 11B2 and 11C2) or whether it is new Moho topography created during core-complex formation (Fig. 11D) or subsequent high-angle Basin-Range normal faulting (Fig. 11E). Our preferred interpretation (Figs. 11B2 and 11C2, elaborated in the following section) is that the crustal welt is a remnant of pre-existing Moho topography that may have served to localize formation of the RMCC. Alternate possibilities follow from the geometric resemblance of converters m2 to the base of a half-graben formed above a normal fault represented by converters m1. Figure 11D is drawn to show m1 as a ductile shear zone accommodating extension in the upper mantle, decoupled by bulk pure shear in the lower crust from normal faulting in the upper crust (converter F1 and its hypothesized westward prolongation) (cf. “decoupled stretching model” of Klemperer, 1988). This model is not easily testable but begs the question of why, during formation of the RMCC, the mantle would have deformed discontinuously with heave –10 km along a single shear zone (m1), as well as by the wholesale flow suggested by the unusually high temperatures required to produce the Harrison Pass and other plutons. Because the Moho returns to its regional level farther west, m1 operating as a shear zone would paradoxically act to thicken the crust locally within the overall region of extension. In contrast, Figure 10E is drawn showing m1 as the down-dip prolongation of a west-dipping, range-bounding fault bounding the Ruby Mountains, nominally formed post–10 Ma. Connecting m1 at the Moho to the western basement exposures in the Ruby Mountains would require a normal fault dipping ≥65° WNW, but m2 dips ~25° ESE (Fig. 7), and if rotated from an initial horizontal by normal faulting, requires a non-physical initial vertical dip for the fault. Additionally the observed thickness of the crustal welt would require a throw of ~10 km (Figs. 7 and 11E), precluded by well penetrations of Paleozoic sedimentary rocks at <1.25 km depth only 4 km west of the Ruby range front (Colgan et al., 2010). We conclude that these interpretations of Moho topography, as created from a previously flat Moho by local crustal thickening during Cenozoic extension (Figs. 11D and 11E), face major or insuperable problems.

In Figure 11, we show all recognized converters, F1, F2, X, Y, m1, m2, and M, irrespective of age. Crustal flow and uplift (pink shading in Fig. 11) of the rocks now exposed within the RMCC are thought to have begun in the Eocene at least in the East Humboldt Range and northern RMCC (Fig. 3B), but F1 was dominantly active, and F2 only became active, in Late Miocene (Figs. 3D and 3E) (Coney and Harms, 1984; Colgan et al., 2010; Lund Snee et al., 2016). As drawn, cartoons C–C’ show inflow of lower crust (pink), which would require some element of shallower return (outward) flow to balance the section. Another possibility is that flow in the middle to lower crust of the southern RMCC only became important after the Middle Miocene. The latter view is consistent with the greater uplift and greater mafic intraplating and underplating observed in the northern RMCC.
Figure 11. Models for core-complex formation based on (A) previous work and (B–E) our observations. (A) Re-drawn from Whitney et al. (2013) (A1) symmetrical core complex; pink area represents hottest crust, with mid-crustal partial melt; (A2) asymmetrical core complex formed at the edge of an orogenic plateau. Parts (B) and (C) each show two different interpretations of the converters identified in the common conversion point images (Fig. 7). (B1) Line B and (C1) line C show a symmetric interpretation as in (A1) but fail to explain the Moho complexities of m1 and m2. (B2) Line B and (C2) line C show an asymmetric interpretation of core-complex formation as in (A2), with crustal flow localized at preexisting crustal thickness change at m1–m2. (D) and (E) show two additional potential explanations that were rejected, in which regional asymmetry and crustal welt were formed (D) only during core-complex formation and (E) only after core-complex formation. RMCC—Ruby Mountains metamorphic core complex. All cartoons are true scale with no vertical exaggeration.
Core-Complex Formation

Recent modeling studies of core-complex formation have included focus on both symmetric and asymmetric elements, whether as successive temporal stages of development (Tirel et al., 2008) or as manifestations of pre-extensional structure (Whitney et al., 2013; Figs. 11A1 and 11A2). In the numerical models of Tirel et al. (2013) and Rey et al. (2009), deformation of a nearly uniform crust yields models of core complexes with near-uniform crustal thickness, albeit in some cases with asymmetric intra-crustal detachments and flow (Fig. 11A1). Only the numerical models with initial laterally varying crustal thickness (such as the edge of an orogenic plateau) seem to develop and retain a modified asymmetric Moho structure (Rey et al., 2010; Whitney et al., 2013; Fig. 11A2). Core complexes, as defined only by the juxtaposition of ductilely deformed rocks against a brittle upper crust across a high-strain zone, can form in a wide range of geothermal gradients (Whitney et al., 2013; Platt et al., 2015). However, the class of core complexes typified by the RMCC, which occurs in extensional provinces with regionally flat Moho and extensive intra-crustal melting, appears to require high initial lithospheric temperatures, perhaps over 800°C at the Moho, to allow sufficiently low mantle viscosity that the Moho can easily deform to accommodate lower-crustal flow (Tirel et al., 2008). High temperatures are also required to produce a partially molten lower crust able to flow to fill a zone of extension (Rey et al., 2009), with sufficient buoyancy to allow internal re-distribution of masses (Rey et al., 2010). These elevated temperatures allow geologically rapid attainment (or maintenance) of isostatic equilibrium and the removal of lateral thickness contrasts at all but the shortest wavelengths (Kuszniir and Matthews, 1988; McKenzie et al., 2000). Hence, even a step change in crustal thickness between two adjacent blocks will evolve into a modest crustal welt after flow at intermediate and long wavelengths (Fig. 11A2; cf. Rey et al., 2010, their fig. 3b). We therefore suggest that the crustal welt we image is a short-wavelength remnant of a preexisting long-wavelength, crustal-thickness variation.

Guided by the above considerations, we attempt to match characteristic features of numerical thermo-mechanical models of core-complex development to key aspects of our CCP images. A symmetric model cartoon (Fig. 11A1, e.g., Whitney et al., 2013) superimposed on the cartoon of our CCP images along profiles B–B’ and C–C’ (Figs. 11B1 and 11C1; converters taken directly from Fig. 7) matches our observations in the upper crust but does not explain our observed Moho variation (m1 and m2). An asymmetric model (Fig. 11A2, after Rey et al., 2010; Whitney et al., 2013) has an identical upper crust but also can match our key observation of a Moho welt west of the RMCC (Figs. 11B2 and 11C2). However, we note that numerical experiments are very sensitive to model setup and detailed parameter choices, so that it is possible to create core-complex–like numerical flows at crustal thickness changes either within the thicker crust (Rey et al., 2010, their figure 2b) or within the thinner crust (Whitney et al., 2013, their figure 13). Hence, our observation of a crustal welt lacks detailed predictive power about the preexisting crustal structure from which the RMCC formed. For example, the models of Whitney et al. (2013) and Rey et al. (2010) use a uniform crustal lithology, but our interpretation suggests a stronger mafic lower crust beneath the RMCC (Fig. 12). Modeling (e.g., Beumont et al., 2006) shows that lower-crustal heterogeneities cause tectonic as well as gravitational forcing of crustal flow, but our observations do not allow us to distinguish whether the presumed mafic lower crust of the RMCC participated in flow or simply fed heat into the crust during emplacement.

Coney and Harms (1984) hypothesized that the MCCs of the Basin and Range developed along the line of a crustal welt within the Sevier hinterland and that the core complexes symmetrically thinned this narrow axis of over-thickened crust by a factor of two. However, reconstruction of large crustal thickness directly beneath the core complexes relies on assumptions of high-magnitude extension across the core complexes (Coney and Harms, 1984; Konstantinou et al., 2012), and more recent depictions of the Sevier hinterland suggest a broad plateau (“Nevadaplano” of DeCelles, 2004) underlain by thickened crust (Long, 2012; Cassel et al., 2014). It is now widely accepted that crustal buoyancy forces due to differential crustal thicknesses are a major driving force for extension of the Basin and Range (e.g., Sonder and Jones, 1999), including perhaps the Cretaceous high-grade deformation recorded in the basement of the RMCC (Hodges and Walker, 1992). The observation of locally large exhumation of originally deeply buried rocks (Hodges et al., 1992; McGrew et al., 2000) from beneath a low-relief plateau that experienced low magnitudes of regional exhumation (Gans and Miller, 1983; Gans et al., 1990; Crafford, 2005; Van Buer et al., 2009; Konstantinou et al., 2012; Long, 2012) can be explained by mid-crustal flow during extension (Gans, 1987; Hodges and Walker, 1992; Cassel et al., 2014). We suggest that focused mid-crustal flow in the absence of strongly localized extension created the RMCC. Although mid-crustal flow can be driven by any lateral variation of gravitational potential energy, the Moho topography we observe today suggests that the driving forces were largely related to variable crustal thickness formed during the Sevier orogeny.

The north-south axis of our crustal welt suggests it is the remnant of originally ~north-south structure that drove east-west crustal flow such that...
proposed by MacCready et al. (1997). The gradual change in fast-polarization directions of SKS splitting across the RMCC (Fig. 10) has previously been interpreted as part of a larger circular anomaly due to complex mantle flow around a mantle upwelling (Savage and Sheehan, 2000) or downwelling (West et al., 2009). Our dense array of stations hints at a more abrupt transition along the west side of the Ruby Mountains, with the fast-polarization direction becoming orthogonal to the crustal welt. Additional data are required to test the strength of this spatial correlation, but the map pattern of polarization directions is consistent with west-east crustal flow associated with the crustal welt, superimposed on the regional lithospheric pattern.

Finally we note that analogous crustal welts may have been imaged beneath other core complexes. “Moho offsets” and “mantle shear zones” have previously been interpreted beneath the North Sea as due to post-Caledonian Devonian crustal collapse (the “paired shear zone” model of Fossen et al., 2014) that is linked to core-complex formation, overprinted by Permo-Triassic rift formation (the “decoupled stretching” model of Klemperer, 1988). Marine deep-seismic reflection profiles along the Norwegian margin show an “intra-mantle bundle of reflections” and “crust/Moho offsets” that span 20–30 km at the Moho and extend ~10 km below the Moho (Gabrielsen et al., 2015), similar to the scale of our crustal welt defined by converters m1 and m2. Other possible examples are minor crustal roots at the margins of the Dabie Shan core complex, eastern China (Yuan et al., 2003) and the Nyainqentanglha core complex, eastern Tibet (Tian et al., 2015). The Nyainqentanglha Shan also exhibit fast polarization directions in the crust that are orthogonal to the extension direction and clearly distinct from the anisotropic fabric of the upper mantle (Zhang et al., 2013). Clearly not all seismic profiles across core complexes image such features (Klemperer et al., 1986) and Hauser et al. (1987) found that the Moho is smooth beneath the Snake Range metamorphic core complex in east-central Nevada, but our data across the RMCC demonstrate the value of identifying Moho topography, or its absence, as a tool for understanding the evolution of core complexes.

Because crustal flow decouples shallow and deeper structure, many salient features of metamorphic core complexes can only be observed through crustal-scale geophysical imaging. Our new CCP mapping of the exposed RMCC and adjacent areas shows unexpected Moho topography along a north-south axis oblique to and 30–80 km distant from the axis of the Ruby Mountains. Together with intra-crustal receiver-function converters and a possible shift in the fast polarization direction of shear-wave splitting, our results suggest that the RMCC formed primarily by asymmetric lower-crustal flow. Our interpretation of the formation of the RMCC (Figs. 3, 11, and 12) invokes crustal thickening during the Sevier orogeny as a cause of high-grade metamorphism, together with subhorizontal mid-crustal flow not strongly expressed at the surface, to create the large lateral changes in metamorphic grade now exposed in the Ruby Mountains and East Humboldt Range. For the most part, Mesozoic crustal thicknesses were maintained until Paleogene southward-sweeping magmatism triggered a new phase of crustal melting and plutonism, which led to creation of the RMCC by allowing renewed mid- and lower-crustal flow into regions of localized extension. The largest volume of mafic intrusion into the lower crust appears to be directly beneath the northern Ruby Mountains, but the modern thickness is likely due to focused crustal flow as much as to focused intrusion. The region of major crustal flow in the across-strike dimension of the RMCC may be broader than the exposed core complex, spanning ~100 km between the residual crustal welt (m1 and m2) to the west and the eastern limit of the inferred shear zone on the east side of the RMCC (where converter X approaches the Moho). Our results suggest that the RMCC formed at a transition between the foreland and hinterland of the Sevier orogen and that flow focused by this boundary played an important role in the RMCC’s formation.

ACKNOWLEDGMENTS

We dedicate this paper to George Thompson, who died 12 May 2017. George’s persistent questioning of our ideas, his deep insights born of his 60-year study of the Basin-and-Range Province, and his mentorship of both authors made this a better paper. This project was funded by National Science Foundation grant EAR-0844386 as part of the Earthscope Flexible Array and by the Petroleum Research Fund of the American Chemical Society grant 48467-AC1. Thanks to Incorporated Research Institutions for Seismology (IRIS) Portable Array Seismic Studies of the Continental Lithosphere (PASSCAL) for providing instrumentation and support, especially Deryn Webb and Greg Chavez for help in the field and Mouse Reusch for help with data archiving. We thank the Bureau of Land Management, U.S. Forest Service, and all of the private landowners whose generous assistance allowed siting of our instruments. Numerous volunteers and graduate students assisted with array installation, maintenance, and retrieval, making this experiment possible. IRIS interns, Kelsey Schlitz and Eva Golos, helped with SKS-splitting processing and analysis. Chris Castillo assisted with figure preparation. Reviews (one by John Platt and one anonymous) and detailed suggestions by Associate Editor Craig Jones helped us improve the paper.

REFERENCES CITED

Law, R.D., et al., eds., Channel Flow, Ductile Extension and Exhumation in Con-


