Evidence of Mantle-Based Deformation Across the Western US

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Key Points:

• Crustal seismic anisotropy reveals the coupling between the crust and the mantle
• Mantle vertical loads drive crustal flow in much of the western US
• Mantle loads can depress the Moho, thicken the crust, and cause isostatic compensation without any topographic expression

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Abstract
We investigate the role that upper mantle buoyancy anomalies play in determining the behavior of the crust. Recently, J. C. Castellanos et al. (2020) observed that the anisotropy of the Pacific NW crust correlates with the upper-mantle velocity structure and suggested that vertical loads in the upper mantle can displace the Moho and drive crustal flow on a regional scale. To provide further insight into this relation, we resolve the crustal anisotropy in regions where near-surface mantle-based deformation might be have occurred or is presently occurring. Specifically, we focus on the crust around the Rocky Mountains and around California since high-resolution tomographic images reveal the presence of mantle structures similar to the ones that are thought to be driving the crust in the Pacific NW. Our results reveal crustal flow driven by mantle vertical loading in both regions and suggest that this mechanism may be key in maintaining crustal isostasy during an orogeny.

Plain language summary
While it has long been hypothesized that the main forces driving tectonic deformation result from flow in the convecting mantle, there is new evidence suggesting that mantle vertical loading also plays an important role in crustal tectonics. Here, we illuminate the deformation field of the lower crust in different regions of the US to search for evidence of mantle forcing of crustal flow. Our findings reveal that mantle gravitational loads can uplift or downwarp the crust-mantle boundary and create lateral pressure gradients that drive the lower crustal away from upwellings and toward downwellings, causing thinning and thickening of the crust. This mechanism differs from conventional models in which crustal thickness variations are associated with topographic variations, and suggests that mantle vertical loading might be key to maintaining isostatic equilibrium and a main driver of crustal deformation. Our results bring us closer to understanding how crustal motions respond to mantle-derived forces and how these influence tectonic evolution processes.

1 Introduction

Unraveling the influence of deep geodynamic processes on the Earth’s surface stress field is critical for understanding the driving forces of tectonic deformation. To date, it is well-established that there are two main sources of stress in the lithosphere: (1) internal buoyancy forces arising from lateral density and thickness variations within the
crust and lithospheric mantle (Lachenbruch et al., 1985; Lachenbruch & Morgan, 1990),
and (2) vertical and horizontal basal tractions arising from buoyancy-driven mantle con-
vection below the lithosphere (Hager et al., 1985; Steinberger et al., 2001). Stresses are
continuous across plate boundaries and are not generated there. While substantial work
has been done to define the kinematics of these two sources, their relative contribution
on both the long-term stability of continents and their state of stress is largely unknown.
Here, we investigate how mantle-based stresses affect the dynamics of the lithosphere through
the analysis of crustal anisotropy.

In general, the difficulty of elucidating the origin of lithospheric stresses stems from
our imperfect knowledge of the physical properties of the crust and the lack of constrains
on the degree of coupling between the tectonic plates and the convective flow of the man-
tle. Over the last few decades, numerous studies have aimed at constraining the mechan-
ical structure of the crust. These efforts typically involve the modeling of the Earth’s to-

topographic response to tectonic loading (e.g., Wdowinski & Axen, 1992; Kaufman & Roy-
den, 1994) or the use of seismic data (e.g., Schutt et al., 2018) to derive estimates of crustal
viscosity and temperature. Findings show, for instance, that there can exist large com-
positional lateral variations across a single craton (e.g., Tesauro et al., 2014), and that
certain regions around the world have the conditions for the lower crust to act as a weak
viscous layer capable of accommodating the lateral pressure gradients within the litho-
sphere (e.g., Block & Royden, 1990; Bird, 1991). Methods aimed at constraining the de-
gree of coupling between the crust and the mantle have also been developed. These tech-
niques include the modeling the lithospheric stress field and the prediction of the tec-
tonic plate’s motion using mantle circulation models (e.g., Steinberger et al., 2001; Bird
et al., 2008). Although these investigations have made significant advancements in ex-
plaining a large part of the Earth’s surface observables, there are still many regions where
agreement between observed and predicted stresses is poor. These discrepancies may re-

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Seismic anisotropy – the dependence of seismic wave speeds on propagation direction and polarization – is a useful tool to detect the existence of coherently deformed structures and preferentially oriented anisotropic minerals (Babuska & Cara, 1991; Savage, 1999). As such, the characterization of seismic anisotropy represents an ideal observational method to map how stress is accommodated within the Earth. In a previous study, Lin et al. (2011) used stations from the EarthScope Transportable Array (TA) to investigate the crustal and mantle anisotropy beneath the western US. In their analysis, they confirm that the asthenospheric flow beneath this region is primarily controlled by a combination of North American plate motion and the sinking of the Juan de Fuca and Farallon slabs. But, more interestingly, they observed that mid- to lower-crustal seismic anisotropy is regionally coherent yet largely uncorrelated with the mantle anisotropy, suggesting that these two layers deformed independently. While the cause of the mid- and lower-crustal anisotropy is agreed to be caused by the crystallographic preferred orientation of anisotropic minerals, the mechanisms that underlie the creation of some of the observed patterns remain puzzling.

Today, it is well understood how sub-lithospheric vertical loads can create dynamic topography (e.g., Hager et al., 1985). This component of surface topography requires stress coupling across the viscous lower and middle crust. Here, we are interested in understanding the long-term effects of this stress on the continental crust, and how the response of the viscous crust affects the surface topography. Such response is recorded in a development of metamorphic fabric within the viscous crust. In the Pacific Northwest (PNW; dashed region in Figure 1), the mid-crustal fabric, as inferred from seismic azimuthal anisotropy, does not correspond to the mantle fabric, geologic province, tectonic history, crustal stress or structure (J. C. Castellanos et al., 2020). There is, however, a remarkable coordination between upper mantle seismic structure and mid-crustal anisotropy, in which crustal azimuthal anisotropy orients perpendicular to the seismically fast (and presumably negatively buoyant) upper mantle structures. This observation led J. C. Castellanos et al. (2020) to present a model in which mantle-derived vertical loads acting on the base of the crust will pull the Moho and surface down to attain isostasy, so that the mantle load is supported by the Moho and (mostly) the surface down warp. This condition creates a significant horizontal pressure gradient that drives viscous crustal flow toward the surface depression (Bird, 1991). With time, crustal flow will result in the surface flattening while the Moho down warp grows, with isostatic balance established between the man-
tle load and the Moho down warp. Because the density contrast across the Moho is small compared to that across the free surface, the magnitude of Moho down warp, and hence volume of crust driven to flow, can be quite large. A relatively small and isolated example is the well-developed local Moho depression of 15 km beneath the eastern end of the WA-OR state line (Gao et al., 2011; Gao, 2015). This Moho depression is centered above the seismically fast, laterally small (Stanciu & Humphreys, 2020), Wallowa mantle load. The imaged crustal anisotropy field is radial in form, centered on the Moho depression and mantle anomaly, and it extends 250 km from the anomaly center. The volume of crustal in the down warp is calculated to be 150,000 km$^3$, presumably supplied by crustal in flow (Wolff et al., 2008).

In most cases, an observed azimuthal anisotropy is thought to record the last major strain event. Since the correlation between mantle structure and azimuthal anisotropy is strong in parts of our study areas and in the PNW, and none of the other origins for the anisotropy are consistent with these observations, we proceed by assuming the mantle loading model is correct. Then, mid-crustal viscosity must be sufficiently low to have strained enough to create a strong and regionally coherent fabric. Just how the viscous middle continental crust develops rock fabric is not very well understood. The efficiency of mid-crustal flow depends on channel thickness, length scale of flow and rock viscosity, with rock flux rate being proportional to the cube of channel thickness and inversely proportional to viscosity and flow length scale squared $h^3/L^2$ (Kruse et al., 1991). Rock viscosities of $10^{19}$–$10^{21}$ Ps s are typical values inferred based on western US observations, in which flow lengths are $\sim$200 km and relaxation times of 10 m.y. are assumed (Kruse et al., 1991). The warm (Schutt et al., 2018) and wet (Jones et al., 2015; Behr & Smith, 2016) conditions prevalent throughout most of the western US would promote viscosity reduction, and under these conditions, an amphibole crystal-preferred orientation (CPO) of type II and III fabric would be preferred, for which the fast direction of anisotropy is subparallel to the flow direction (Ko & Jung, 2015).

To obtain a broader perspective on mantle-driven viscous crustal flow, we extend the analyses of Lin et al. (2011) and J. C. Castellanos et al. (2020) and use the dense station coverage provided by the TA, and a few other temporal networks, to reliably resolve the lateral variations of crustal anisotropy in regions where near-surface mantle-based deformation might have occurred or is presently occurring. In particular, we focus on areas around the Rocky Mountains and California since high-resolution tomographic
images (Schmandt & Humphreys, 2010) reveal the presence of seismically fast mantle structures similar to the ones that are thought to have driven crustal flow in the PNW (Figure 1).

2 Data and Methods

We cross-correlate the ambient seismic noise field recorded by every station of the TA with the one recorded by every publicly available broadband seismometer that operated within our two regions of interest (Figure S1). This process allowed us to extract clear fundamental mode surface waves traveling between different pairs of stations that were then used to construct the azimuthal anisotropy models. Because of the large differences in the crustal properties between the western and central US, we performed our surface wave study in two different frequency bands. The limits of these filters were determined by using a 3-D velocity model of the US (Figure 2AB; Shen & Ritzwoller, 2016), and computing the Rayleigh wave sensitivity kernels for a wide range of frequencies at each (x-y) coordinate of the domain. We then used a 3-D crustal thickness model of the US (Figure 2C; Schmandt et al., 2015) to define the depth of the middle and lower crust at each location and, with it, find the shortest and longest period in which the majority of the sensitivity kernel’s amplitude lies between the two layers (Figure S2). Here, we take the depth of the Moho as the base of the lower crust and the midpoint between the free surface and the Moho as the top of the middle crust. Figure 2EF shows the shortest period that is sensitive to the middle crust and the longest period that is sensitive to the lower crust, respectively. With this analysis, we determine that, in order to map the average anisotropic properties of the lower crust, the surface wave analyses must be made between the 18-31 s period band for the Rocky Mountain Complex, and between the 12-23 s period band for the California area.

To build the anisotropy models, we adopt a seismic beamforming scheme (J. Castel-lanos & Clayton, 2021). Within this framework, we use every station that is in our target areas and create several subarrays so that the coherent energy moving through each group of stations can be translated into a local phase velocity and direction of propagation (Figure S7). Here, the number of instruments that composed each subarray is variable but their radii, and hence their resolution, are close to constant as we set a 1-wavelength minimum and a 2-wavelength maximum threshold, thereby avoiding spatial aliasing (Brenguier et al., 2015; Nakata et al., 2016). To ensure the exclusive use of high-quality waveform
data, we only use cross-correlation functions that have a SNR larger than 5 and an off-
set of at least 3-wavelengths away from the geographic center of the subarrays. Once the
band-passed energy of all the virtual sources are beamformed, we collect all phase ve-
locity and backazimuth measurements that were made at each subarray and character-
ize the 20 wavefield’s azimuthal dependence using the generalized model of Smith and
Dahlen (1973) for surface waves in a weakly anisotropic media (Figure S9). This parametriza-
tion allows to extract a fast direction term and an amplitude term that are, in prin-
ciple, related to the anisotropic properties of the lower crust. Details of the cross-correlation
processing, the beamforming technique and the anisotropy characterization are described
in the supporting information.

3 Results

Figure 3AC shows the lower crust azimuthal anisotropy model for both target ar-
areas plotted on top of the mantle P-wave structure at 195 km depth. The bars are ori-
ented in the fast azimuth direction and their length is proportional to the magnitude of
anisotropy. In a similar representation, Figure 3BD, shows the SKS-derived upper man-
tle anisotropy measurements compiled by Becker et al. (2012). Below we analyze the re-
lation between these two anisotropy fields and discuss the mechanisms which may un-
derlie the creation of the lower crust anisotropy of both regions.

Mantle anisotropy beneath the Rocky Mountain region appears to be fully disso-
ociated with the anisotropy in the overlying crust. Nonetheless, and similar to that seen
in the PNW, the crustal anisotropy and upper mantle seismic velocity structures seem
to relate to one another (Figure 3A). Fast azimuth orientations tend to be perpendic-
ular to a SW-trending seismically fast upper mantle structure beneath Wyoming and parts
of Colorado and Utah, with near-zero amplitudes at almost its geographic center. Both
of these features suggests that the anisotropy of this region was created by crustal flow
roughly perpendicular to this trend.

Consideration of the tectonic history of the crust around Rocky Mountains offers
insight into the creation of its anisotropy field. After residing near sea level for 100s of
m.y., the swath of crust that now resides above a SW-trending zone of deep fast man-
tle beneath Wyoming became anomalously depressed in the mid-Cretaceous. This event
is evidenced by a series of deep marine basins that represent a local mantle loading along
this trend (S. Liu et al., 2011; Li & Aschoff, 2021). In this setting, and assuming that
the rheological conditions of the crust had sufficiently low viscosity, crustal materials could
have flowed towards the Moho depression and created the anisotropy that is observed
at present (Figure 4A). In an alternate scenario, subsequent uplift of the region repre-
sents an increase in mantle buoyancy that progressively elevated the continental inte-
rrior—including the depressed trend—into the broad western US uplift (Figure 2D). As
a result, the trend of Cretaceous basins became no longer a zone of anomalous topog-
raphy or crustal thickness (Shen & Ritzwoller, 2016), and the early crustal thickening
could have flowed away from the area of former Moho depression. This outflow would
have also developed the anisotropy that is observed at present. In either instance, the
low viscosity required for crustal flow probably was due to an abundance of water pro-
vided by flat-slab subduction (Jones et al., 2015; Behr & Smith, 2016). We note that
crustal strain is not related to a Laramide thrust fault, which was a small strain event
(Bird, 1998), and also faulting involved only a small fraction of the crustal volume. Po-
tentially complicating the simple interpretation of this crustal flow field, is the relatively
recent emplacement of slow mantle beneath NE Wyoming associated with Yellowstone
(Pierce & Morgan, 2009) and beneath central Colorado (Aslan et al., 2010; Lazear et al.,
2013). In each case, the mantle is thought to be buoyant not only because it is seismi-
cally slow but also because each area has experienced substantial young uplift. These
young structures might explain why crustal anisotropy is roughly radial around both of
these mantle slow volumes (Figure 4B).

California mantle anisotropy orientation south of the Juan de Fuca slab (JdF) ro-
tates clockwise from 70-90° in eastern California to 100-130° in western California, near
the San Andreas Fault (SAF). Compared to the mantle, the crustal anisotropy away from
the Isabella mantle anomaly (ISA) and south of San Francisco has a similar azimuth in
eastern California, but near the SAF the azimuth is further counter-clockwise at 110-
150° (Kosarian et al., 2011). Away from perturbations, anisotropy is expected to rotate
progressively toward the direction of simple shear, i.e., it should be slightly counter-clockwise
of the SAF. Assuming this behavior, there are two points to be made: (1) the crustal
anisotropy is ahead of the mantle (Barbot, 2020), suggesting either that the crust is driv-
ing the mantle or the crustal deformation zone is narrower (and more highly strained)
than that of the mantle (Figure 4C), and (2) the anisotropy near the negatively buoy-
ant ISA mantle anomaly appears to be a simple "sinker" perturbation to the SAF-related anisotropy (Figure 4A).

In northern California, the crustal anisotropy is coherent and organized around the southern JdF slab. The interpretation of these results is still not very clear, but the strong and complex crustal dynamics of the Mendocino triple junction area has been recognized and discussed by Furlong and Govers (1999) and K. Liu et al. (2012). Between the southern JdF slab and the ISA mantle anomaly, the margin-normal crustal anisotropy is nearly E-W in orientation. A straightforward interpretation is crustal flow away from the high-standing Great Basin and toward the low-standing Great Valley, much like that suggested for the Tibetan region (Xie et al., 2017). This flow occurs largely between the two seismically fast mantle features, perhaps owing to cooler lower crustal temperature and higher crustal viscosity above these cooler mantle structures (which comes up to the Moho; Stanciu & Humphreys, 2019).

4 Discussion and Conclusions

Including the observations of J. C. Castellanos et al. (2020), we have three areas that provide different tectonic settings for study. Yet their crustal anisotropy fields can be explained with a few simple kinematic processes. From these observations we can make some general conclusions about continental crust during an orogeny.

In the PNW and Rocky Mountains, crustal azimuthal anisotropy is roughly perpendicular to central seismically fast upper mantle structures (Castellanos et al., 2020; Figure 3A). The anisotropy is attributed to crustal flow that is approximately perpendicular to the location of seismically fast mantle. In each case there is no obvious topographic expression (the Wallowa Mountains are about 60 km south of the area of thick crust). For the Wallowa mantle anomaly, excitation is young (initiating ~16 Ma, associated with the Columbia River flood basalt event, Darold and Humphreys, 2013) and perhaps ongoing; in the Rocky Mountains, the anisotropy is thought to be ~60-80 Ma (Humphreys et al., 2015) and preserved since. For the crustal flow, we cannot distinguish between the inflow and outflow scenarios. Nonetheless, crustal flow appears to be poloidal in response to local buoyancy structure, i.e., the inferred flow is curl-free, centered on the buoyancy anomaly. Within the Great Basin, the long-wavelength elevation is relatively modest and horizontal pressure gradients are likely to be small. Nonetheless, the
west-to-southwest orientation of anisotropy is roughly aligned with the topographic gradient, and not aligned with the NW orientation of tectonic strain. This region has been elevated since before 70 Ma (Henry et al., 2012; DeCelles, 2004), yet crustal flow away from the Great Basin has not eliminated the regions high topography. This is attributed to the large volume of middle crust distributed over the wide area of high elevation, which would require a very long duration of time for mid-crustal flux to relax the topography. Recalling that relaxation time goes as flow length scale squared, the maintenance of high elevation is attributed to the ~700 km width of Great Basin high elevation. In contrast to the above examples, southern California crustal anisotropy is sub-parallel to the well-developed San Andreas shear zone and the crustal flow appears to be approximately toroidal, i.e. divergence-free, which is expected for a passive transform margin. The obvious anisotropy orientation perturbation near the ISA mantle anomaly is easily attributed to the superimposed poloidal flow, with crust converging on the site of mantle loading created by the negatively buoyant ISA lithospheric load (Figure 3C).

With respect to continental orogenies, we infer a few important consequences: (i) With mantle loading, isostatic balance is fundamentally maintained between the mantle load and compensating change in crustal thickness accommodated by Moho warping. However, if loading occurs more rapidly than crustal flow can supply compensating crust, a surface deflection will occur, such as the 80 Ma creation of basins in mechanism. With continued crustal flow and Moho adjustment, isostasy may be ultimately maintained without a surface expression and could be topographically invisible (J. C. Castillo et al., 2020; Crosswhite & Humphreys, 2003). This mechanism differs from tectonically-driven changes in crustal thickness, in which crustal thickness variations are associated with (and compensated by) topographic variations. (ii) Because the density difference across the Moho is small compared that of the free surface, modest mantle loads can create significant Moho warping and drive large amounts of crustal flow. This mechanism may be common in orogenies, and it is an effective way for straining lower crust to distances relatively far from the site of loading. In contrast, simple shear deformation (e.g., below the SAF or a Laramide thrust faulting) can accommodate large strain within the small volume of a shear zone, and anisotropy may be developed within only a relatively small volume of rock. (iii) The observed common occurrence in the western US of crustal anisotropy related to mantle loading suggests a widely distributed low-viscosity lower crust. Such crustal conditions may require an orogeny to supply crustal hydration.
or heating. Low crustal viscosity will tend to mechanically decouple the crust from hori-
horizontal mantle flow, in which case horizontal mantle flow may not play an important role
in crustal tectonics.

Figure 1. Upper mantle velocity structure of the western US and geography of the tar-
get regions. The main map shows a depth slice through the Vp tomography model at 195 km
(Schmandt & Humphreys, 2010). The dashed black rectangle delimits the region of study of
J. C. Castellanos et al. (2020), whereas the continuous black rectangles delimit the regions that
are focused in this investigation. The zoomed maps show the distribution of broadband seismic
stations used to resolve the crustal anisotropy beneath each region (red inverted triangles).
Figure 2. Crustal properties of the western US crust. (A-B) Show the mid-crust and lower crust shear wave velocity structure as derived from ambient seismic noise and earthquake tomography, receiver functions, and Rayleigh wave ellipticity (H/V) measurements (Shen & Ritzwoller, 2016). (C) Shows the crustal thickness as derived from teleseismic P-to-S receiver functions, and Rayleigh wave phase velocities and ellipticity from noise interferometry and earthquakes (Schmandt et al., 2015). (D) Shows the surface topography (Ryan et al., 2009). (E-D) Shows the shortest Rayleigh wave period that is sensitive to the middle crust (MC) and the longest Rayleigh period that is sensitive to the lower crust (LC), respectively. The average upper and lower period for the two target regions is shown in the bottom of each of the geographic borders.
Figure 3. Lower crust and mantle anisotropy for the two target regions. (A) Lower crust azimuthal anisotropy around the Rocky Mountains. (B) Station-averaged shear-wave splitting measurements around the Rocky Mountains (Becker et al., 2012). (C) Lower crust azimuthal anisotropy around California. (D) Station-averaged shear-wave splitting measurements around California (Becker et al., 2012). The bar orientation gives the fast direction of anisotropy, and the bar length is proportional to anisotropy amplitude. The background color represents the mantle P-wave structure at 195 km depth (Schmandt & Humphreys, 2010). In panels (A) and (B) the green arrows depict our preferred interpretation of crustal flow. The blue arrows in (C) depict the relative motions between the Pacific (PA) and North American plates (NA), the Juan de Fuca plate (JdF) and NA. The purple lines in (C) mark the plate boundaries. ISA stands for the Isabella mantle anomaly.
Figure 4. Proposed mechanisms that give rise to the observed crustal anisotropy. (A) Negatively buoyant mantle anomalies. The load of the mantle lithosphere creates vertical stresses on the Moho and pull the crust down, creating a lateral pressure gradient that drives Poiseuille flow in the ductile mid-lower crust towards the mantle anomaly. (B) Positively buoyant mantle anomalies. The hot and buoyant mantle rises and pushes the crust upward, creating a lateral pressure gradient that drives Poiseuille flow in the ductile mid-lower crust outwards from the mantle anomaly. (C) Plate interactions. The localized near plate boundaries drive the crust and the asthenospheric flow. (D) No crustal anisotropy. There are no mantle-buoyancy anomalies, crustal thickness variations or nearby plate boundaries. The mantle lithospheric strength isolates the crust from the sub-horizontal asthenospheric flow.

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