Bulletin of the Seismological Society of America, Vol. 106, No. 5, pp. 2395–2401, October 2016, doi: 10.1785/0120150173

Ⓔ

Short Note

Characterizing Earthquake Location Uncertainty in North America Using Source–Receiver Reciprocity and USArray

by Janine S. Buehler and Peter M. Shearer

Abstract The Comprehensive Nuclear-Test-Ban Treaty community often uses calibration events with well-determined origins to improve absolute locations of nearby seismic events by accounting for the biasing effects of unknown velocity structure, but the number of these ground-truth events is limited. To provide additional constraints, source–receiver reciprocity allows us to use seismic stations as calibration events with known locations. The dense and uniform spacing of the USArray transportable array stations makes them ideal to measure the spatial coherence of mislocation vectors across North America and hence to assess how close calibration events (or stations) need to be to target events to improve locations for a given region. We use a grid-search approach for the station "relocations," using both teleseismic earthquakes and simulated regional events. Our results show that the mislocation vectors are spatially coherent for scales up to 500 km in many regions, but that in some places, such as regions that can be associated with strong velocity anomalies in the upper mantle, mislocation vectors exhibit large changes over short distances.

Online Material: Figures showing station location and dislocations and travel-time residuals.

Introduction

To comply with the Comprehensive Nuclear-Test-Ban Treaty (CTBT), verification programs need to be able to accurately detect, locate, and classify seismic events. The treaty requires an absolute location error ellipse of an 1000 km^2 area or less to manage onsite inspection efforts (e.g., Kennett and Ringdal, 2001). In practice, it can be quite challenging to accurately locate events, due to uncertainties in the seismic-velocity structure and arrival-time picks (e.g., Billings *et al.*, 1994). Often travel times are computed through a simple 1D velocity model (Husen and Hardebeck, 2010), at least for a fast initial hypocenter estimate. To improve location accuracy, one of two approaches (or a combination of both) is typically chosen: the 1D structure is replaced with a more complicated and hopefully more accurate 3D velocity model (e.g., Ryaboy *et al.*, 2001), or location-dependent empirical corrections are applied to the travel times from the 1D model (e.g., Myers and Schultz, 2000; Nicholson *et al.*, 2008).

The International Monitoring System relies on calibration events to obtain source-specific station corrections (e.g., Bondar *et al.*, 2001) to improve location accuracy. If the location of an event is precisely known, for example a quarry blast or an earthquake that can be very well located with a dense array, the deviations to predicted travel times can be analyzed, and corrections can be computed for each event-station path. However, these corrections are only valid for events close to the ground-truth calibration event, because we cannot expect that the corrections are coherent over large distances because of heterogeneous velocity structure. But how close does the calibration event have to be to the unknown target earthquake location to yield improved locations?

The availability of ground-truth seismic sources is limited (e.g., Yang *et al.*, 2004). To provide more calibration events, Shearer (2001) showed that stations can be relocated using arrival-time data in a similar fashion to events because of source–receiver reciprocity and that stations therefore can be used as calibration events in areas without known source locations. Here, we follow this approach and investigate the coherence of mislocation vectors computed by relocating USArray stations. Because these receivers have a known location (just like quarry blasts), they can then be used as calibration events and empirical travel-time corrections estimated from the difference between their true locations and those obtained from relocation. By relocating the regularly spaced USArray stations, we can obtain a measure of how coherent the mislocation vectors are with interstation distance and therefore how close calibration events need to be to ground-truth events to obtain a desired level of location accuracy. In addition to improving earthquake locations, station mislocation vectors may indicate features of local near-surface velocity structure and/or anisotropy if we assume that the delay times are most likely accumulated in the crust and upper mantle below the station. Cleary (1967) suggests that azimuthal variations in source terms are likely caused by features in island arcs in subduction zones.

Here, however, we only focus on the coherency of the mislocation vectors and the potential use of stations as calibration events and extend the work of Shearer (2001) to USArray stations: we first relocate all the stations with a grid-search approach using teleseismic arrivals alone and analyze the coherence of the mislocation vectors as a function of interstation distance. In addition, we estimate mislocation vectors from regional events using an upper-mantle *Pn* velocity model and compare the coherence between the stations. Finally, we discuss these results in terms of their implications for using stations as reference events for the CTBT community.

Data and Station Relocation

Because of source–receiver reciprocity, station relocation is similar to earthquake location, with the added advantage that we already know the positions of the stations. The corresponding disadvantage is that the source locations are only approximately known, and hence reciprocity does not completely hold. However, errors in earthquake locations and origin times will bias the computed station locations only if there is a systematic azimuthal dependence in the resulting residuals (Shearer, 2001). In addition, any bias will be similar for all stations within a local region if the events are all far away, as is the case for the teleseismic data that we analyze here. Because we focus in this study on the spatial coherence of the station mislocation vectors, and not on how precisely we can locate the stations with advanced velocity models or techniques, the accuracy of the earthquake locations is relatively unimportant.

We use arrival-time picks made by seismic analysts by the Array Network Facility (ANF). These picks can be downloaded from their website in monthly tables. We collect all the teleseismic *P*-arrival picks recorded at USArray stations between April 2004 and February 2014 from the ANF catalog. We restrict our analysis to earthquakes with epicentral distances between 30° and 90° to avoid complications from triplications and the core (Fig. 1). To avoid mislabeled picks, we only use times with residuals smaller than 4 s with respect to the ak135 1D velocity model and U.S. Geological Survey (USGS) hypocenter locations. This results in ~2600 picks per station on average. We do not correct for station elevation as this affects all the data for a station in the same way.

To relocate a station, we perform a grid search around the known station location, similar to many earthquake location approaches (e.g., Richards-Dinger and Shearer, 2000).

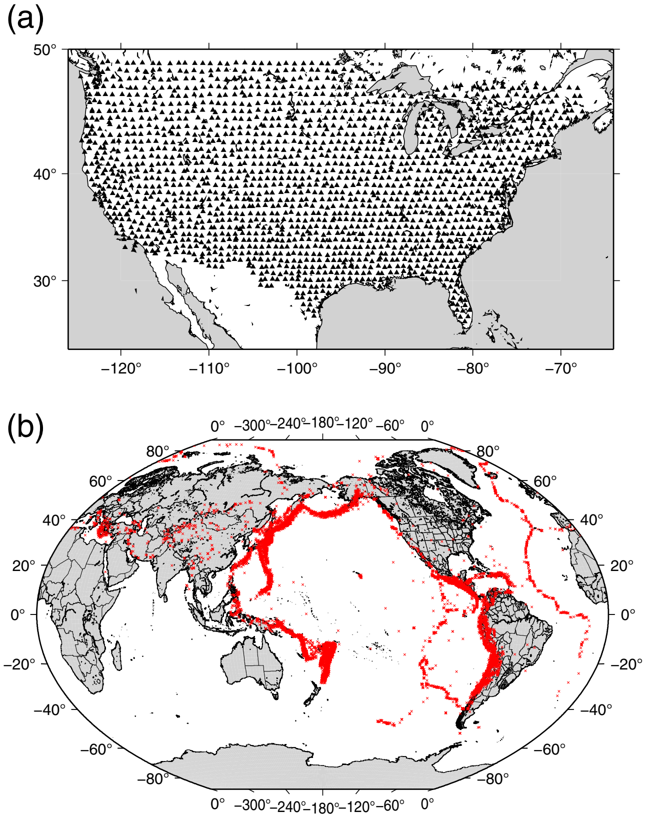

Figure 1. (a) USArray stations (triangles). (b) Events (crosses) used in this study. The color version of this figure is available only in the electronic edition.

First, we compute travel-time residuals for each event on small circles around the known station location, in 0.02° range intervals out to 0.26° and 18° intervals in azimuth. We find the location with the smallest misfit. We then repeat the process with a finer grid from this new location to find the final best-relocated location. To assess the misfit at each trial location, we compute both the L1 and L2 norms of the residuals, after removing the median or mean, respectively, from the residual distribution to account for local crustal changes below stations. We then compute the azimuth and length of the mislocation vector from the known station location to the location with the smallest norm.

The station relocation process can be influenced by the data distribution, because earthquakes are generally not uniformly spread around the stations. We therefore also compute summary ray paths for each receiver by averaging residuals based on 10° bins in range and azimuth. We only compute mislocation vectors for stations that have data from at least 10 source bins and a smaller than 90° gap in azimuth.

Station Location Uncertainty from Teleseismic Earthquakes

Figure 2 shows the computed mislocation vectors for each USArray station, that is, the vector from the true station

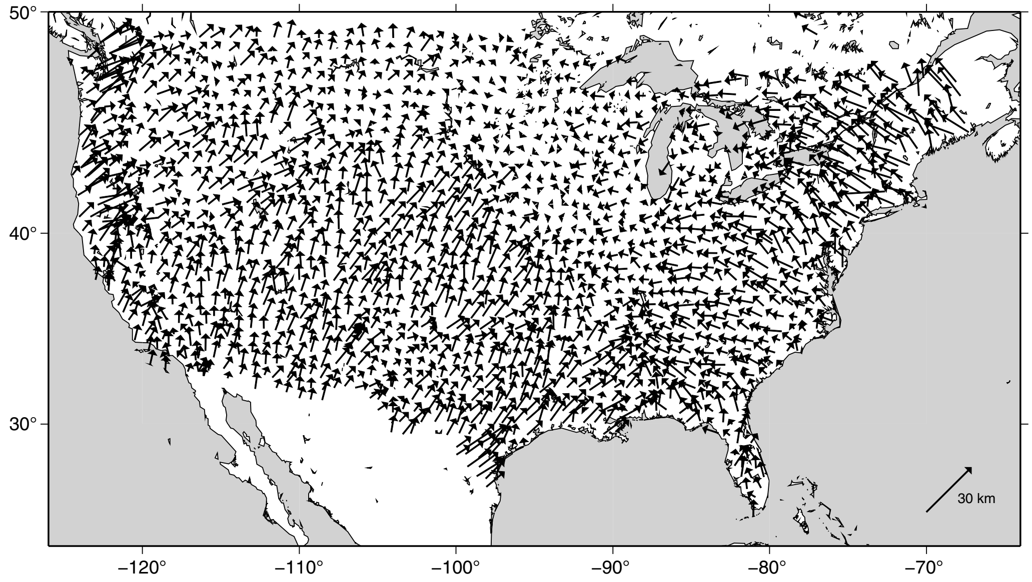

Figure 2. Station mislocation vectors showing the offset from the true station location and that estimated by locating the station using teleseismic *P*-wave arrival-time picks from distant earthquakes. Arrows are highly exaggerated in length (see 30-km reference vector at lower right).

location to the computed station location. Most of the vectors do not exceed 13 km, and the median vector is 6.25 km long. Assessing the spatial coherence of the station mislocation vectors is of primary importance if they are used as calibration events. Over small scales, the mislocation vectors in most regions are spatially coherent, which means that the location accuracy of a station (or a real earthquake) could be improved by locating it relative to a nearby station. However, in some areas the station mislocation vectors change rapidly, indicating that the use of calibration stations (or events) will be problematic and could even worsen the location accuracy of a target event. The most significant feature in the mislocation vector map is the ~90° change in direction in a relatively narrow belt around longitude 90° W. To the west and east of the boundary, the directions of the mislocation vectors vary more smoothly.

The great coherence of mislocation vectors in the heterogeneous western United States suggests the residuals are influenced over larger scales than that of localized velocity variations in the crust or upper mantle beneath a station. This may include not only anisotropy (e.g., Bokelmann, 2002) and geometric effects, but also systematic picking errors, uneven ray distribution, and errors in hypocenter location. The ANF analyst picks are associated with the hypocenter solutions from the USGS in the ANF catalog for all teleseismic events. Hence, the picks and hypocenters are not self-consistent as in the study of Shearer (2001) (the picks are not used to determine the hypocenter in the catalog). Therefore, we also perform the relocation procedure with a consistent set of picks and locations from the Engdahl–van der Hilst–Buland (EHB) bulletin from the International Seismological Centre. Unfortunately, the EHB catalog is only available through 2008, when USArray was still in the western half of the United States. Testing with the two data sets (i.e., only using the events that are in both catalogs and using EHB locations with ANF picks and vice versa) shows that difference in the two results is mainly caused by different hypocenter depths or origin times, and the different earthquake latitude/longitude coordinates have only a minor effect. For example, we find that events from the southwest are generally deeper in the ANF catalog compared to the EHB catalog. As discussed earlier, such a dependence on azimuth with a 180° periodicity can bias the analysis. Consistently too deep hypocenters (or late origin times) in the south would lead to station mislocation vectors pointing northward. This effect could be partially responsible for the great coherence in the western half of the ANF picks mislocation vector image.

To assess the influence of this effect in more detail, we remove the average residual for each earthquake across the United States prior to the relocation process. These new relative residuals should be less sensitive to hypocenter errors. Removing a source term for each earthquake is not straightforward, because the transportable array (TA) stations move every two years. Hence, we cannot just remove the average residual from these stations (which would result in large differences in source terms for events at the same location recorded at different subsets of stations at different times because of receiver-side structural changes), because we would not be able to resolve changes across the whole continent, but only the footprint of the recording stations at the time of their operation. To have consistent relative event residuals, we therefore use the residuals from the permanent stations (the USArray Reference Network, see Ⓔ Fig. S1, available in the electronic supplement to this article) to compute the source terms. These permanent stations cover the whole continent at larger station spacing than the TA. Unfortunately, there are

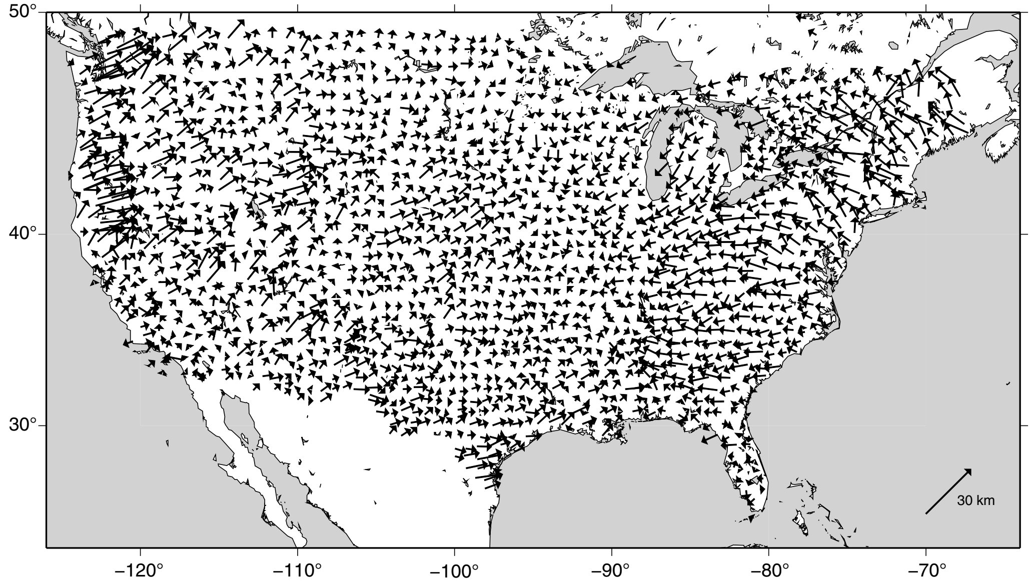

Figure 3. Same as Figure 2, but with earthquake source terms to remove systematic location bias.

usually only picks available for a subset of the reference network. To have a spatially uniform residual distribution across the United States for each earthquake, we therefore interpolate the residuals at the available permanent stations to all TA stations and then compute the average of the interpolated values. With this approach, we should avoid any bias from local receiver-side structure to bias the individual source terms. However, picks at the reference network only appear to be available in the ANF catalog starting in 2008, making it impossible to obtain a source term from reference network residuals for events from 2004 to 2007. For this simple test, we therefore interpolate the source terms at similar locations to earlier events (this is not ideal and assumes the offset in hypocenter location stays consistent throughout the years). Figure 3 shows that this effort results in smaller mislocation vectors, but the overall pattern stays similar (the correlation coefficient between the two images, computed as the sum of the dot products of the two vector fields divided by the square root of the product of the sum of each vector field dotted with itself, is 0.74).

Figure 4 shows the divergence of the mislocation vector field in Figure 3 to highlight the regions of greater incoherency. Although we do not focus on tectonic implications in this article, we will briefly highlight a few features. In addition to the large patch of incoherence that is located between the northeastward pointing mislocation vectors in the western United States and the southwestward pointing vectors in the east, we observe smaller patches that include regions around the Wyoming basin, the Cascadia back-arc, and coastal areas. In these regions, calibration stations or events need to be close to the target event to obtain good location accuracy. The areas of incoherency seem to be mostly associated with consistent velocity anomalies in the top 200 km of the upper mantle. Enhanced incoherency above velocity anomalies is to be expected, as mislocation vectors will either point toward or away from them depending on the sign of the perturbation. In contrast, we expect to observe long and coherent mislocation vectors above boundaries of strong velocity contrasts, for which travel times from one direction are consistently delayed or advanced. The patches of incoherence in the Cascadia back-arc coincide with the locations of high-velocity blocks from the subducting Juan de Fuca slab imaged by body-wave tomography (Schmandt and Humphreys, 2010; Obrebski *et al.*, 2011; Schmandt and Lin, 2014). In the central slab region where the velocity anomaly is diminished in body-wave tomographies, the mislocation vectors are smaller. Likewise, there are patches of incoherency in the vicinity of the low-seismic velocities near the Yellowstone hotspot, in the vicinity of the large high-velocity anomaly beneath Wyoming, and in the central Great Basin. Further east, the areas of strong divergence outlines the high-velocity anomaly associated with the North American craton (e.g., Burdick *et al.*, 2014). In addition, the mostly southwest direction of mislocation vectors throughout the North American craton agrees with the *P*-wave anisotropy in the upper mantle as inferred by Bokelmann (2002). Bokelmann (2002) also imaged coherent fast directions throughout the craton but a strong change in direction at its boundary.

In Figure 5, we plot the difference in mislocation vector as a function of interstation distance. As expected, the mislocation vector difference increases with distance. For calibration events within 100 km of the target event, the median and 90th percentile vector differences are ~4 and 9 km, respectively. For calibration events 500 km away from the target event, the corresponding differences are ~5 and 13 km. The dashed lines show the median and 90th percentiles for the individual station mislocations. An improvement in the 90th percentile is only achieved for station separations of less

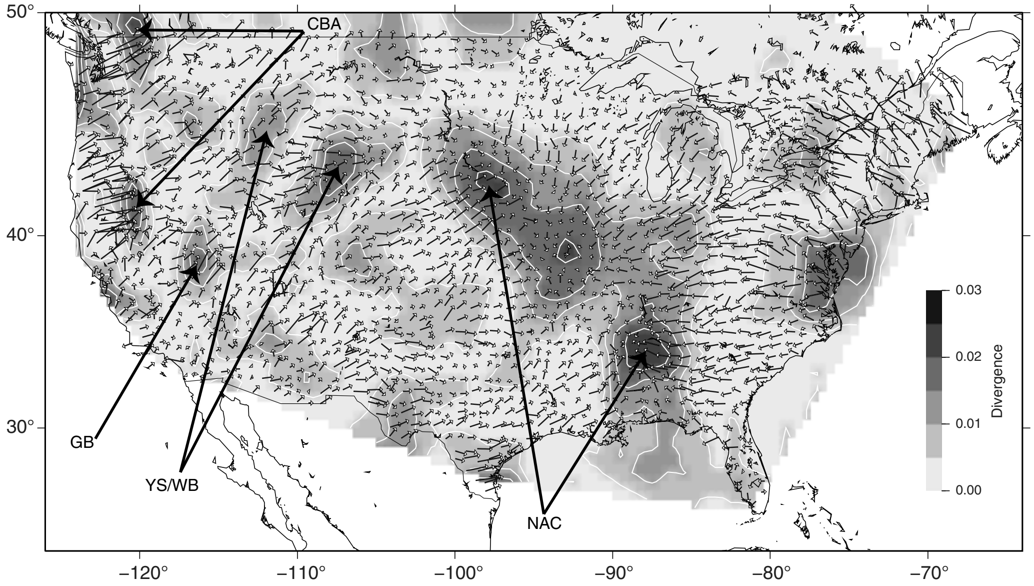

Figure 4. Absolute value of the divergence of the mislocation vector field plotted in Figure 3 to indicate regions with sharp spatial changes in mislocation vectors. Black arrows point to areas with great incoherency that can be associated with velocity anomalies in the upper mantle. CBA, Cascade back-arc; GB, Great Basin; YS, Yellowstone; WB, Wyoming basin; NAC, North American craton.

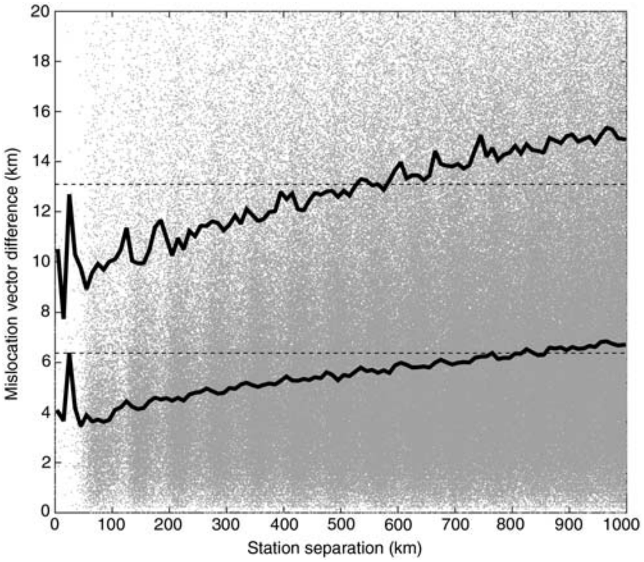

Figure 5. Difference in station mislocation vector pairs as a function of the true interstation distance. The solid lines show the 50th and 90th percentiles of the data. The dashed lines show the 50th and 90th percentiles of the individual station mislocation vectors. Most of the interstation distances are 70 km or larger, resulting in only a few points at shorter ranges.

than 500 km; at larger distances, the use of a calibration event could actually worsen the location accuracy. These results roughly agree with the global station location results plotted in Shearer (2001) for station separations of 300 km and less, but the USArray results show a continued increase in the vector differences at larger distances. Figure 5 provides an overall guide to the expected improvement that one can obtain in location accuracy using nearby calibration events, but as shown in Figure 4, the teleseismic location biases vary greatly across the United States.

Station Location Uncertainty from Regional Simulated Events

The CTBT community often uses regional arrivals in addition to teleseismic data to locate smaller events. Because the regional azimuthal earthquake coverage is not ideal for a majority of the USArray stations, we conducted a synthetic experiment to assess the likely mislocation vector coherence at regional scales. We used our upper-mantle *Pn* model (Buehler and Shearer, 2014) to compute synthetic *Pn* residuals at each station from simulated earthquakes at a 5° radius around each station. We then proceeded in a similar way as with the teleseismic data described in the previous paragraphs to relocate the station.

Figure 6 shows the mislocation vectors derived from the *Pn* velocity model. The mislocation vectors again show sharp changes in locations at the boundaries of strong velocity anomalies, for example in the Sierra Nevada or the Snake River Plain. Because of the generally more uniform uppermost mantle in the east, mislocation vectors are smaller in the eastern part of the continent. Figure 6 shows that, when working with regional events, the calibration stations need to be closer to the target and on average closer than 300 km to be a useful correction. This makes sense because in general we expect the mislocation vectors from regional data to be less coherent than those from teleseismic data, because of the strong uppermost mantle velocity variations over relatively short scales. Figure 7 also implies that the absolute location errors are less for regional phase data than teleseismic data. However, these uncertainties are likely underestimated

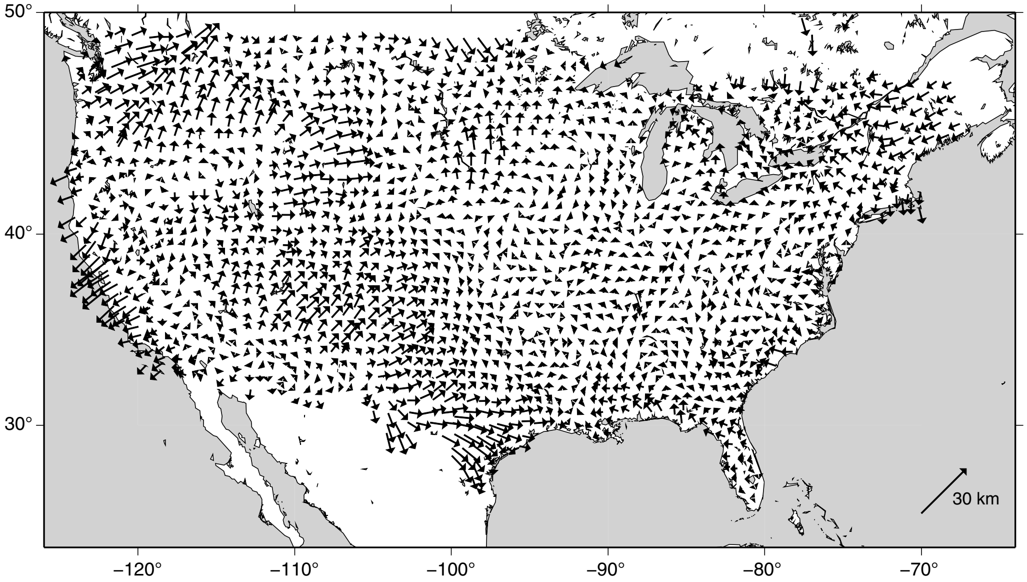

Figure 6. Station mislocation vectors estimated from synthetic *Pn* times based on the uppermost mantle *Pn* velocity model of Buehler and Shearer (2014).

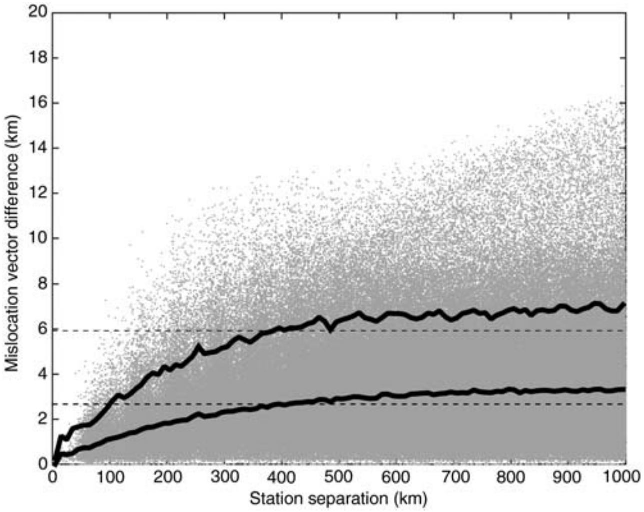

Figure 7. Difference in station mislocation vector pairs as a function of the true interstation distance for the regional synthetic example.

because our synthetic experiment ignores crustal velocity and thickness variations, and the *Pn* velocity variations are damped to some extent by tomography regularization.

We use the crustal thickness estimates from Buehler and Shearer (2014) to adjust the travel times depending on event location for a simple test on the influence of Moho depth changes. Crustal thickness changes in North America appear to be smoother than seismic-velocity variations in the crust and uppermost mantle, with generally thinner crust in the western United States and thicker crust to the east (Buehler and Shearer, 2014). Hence, the influence of crustal thickness changes on the mislocation vectors will depend on the epicentral range of the earthquakes. In Ⓔ Figures S3–S5, we show three examples of simulated regional experiments, with earthquakes at 3°, 5°, and 8° away from the station, to demonstrate the increased effect of the crustal thickness variations with an increase in range. At larger ranges, the crustal thickness variations dominate the behavior of the mislocation vectors, whereas at shorter ranges they are mostly influenced by the uppermost mantle velocity structure (as crustal thickness is approximately constant).

Conclusions

Because of source–receiver reciprocity, stations can be used as reference events in locations without known source locations, such as quarry blasts. We demonstrate how relocating a station network can provide valuable information on the coherency of the mislocation vectors and therefore on how close the reference events need to be to the new target event to improve its location. It is clear that coherency varies with tectonic regions in North America, and that the sharpest changes seem associated with structural changes in the upper mantle. We observe the most significant break of coherency near 90° W, with more subtle mislocation vector changes to the east and the west. In many regions, the mislocation vectors are spatially coherent for scales up to ~500 km.

Data and Resources

Maps were generated with the Generic Mapping Tools of Wessel *et al.* (2013). Arrival-time picks used in this study can be downloaded from the Array Network Facility (ANF) at http://anf.ucsd.edu/tools/events/ (last accessed June 2015) in monthly tables.

Acknowledgments

We thank two reviewers and Editor-in-Chief Diane Doser for their constructive comments. This research was supported by National Science Foundation Grant EAR-1358510.

References

Billings, S., M. Sambridge, and B. Kennett (1994). Errors in hypocenter location: Picking, model, and magnitude dependence, *Bull. Seismol. Soc. Am.* **84,** no. 6, 1978–1990.

Bokelmann, G. H. (2002). Convection-driven motion of the North American craton: Evidence from *P*-wave anisotropy, *Geophys. J. Int.* **148,** no. 2, 278–287.

Bondar, I., X. Yang, R. North, and C. Romney (2001). Location calibration data for CTBT monitoring at the Prototype International Data Center, *Pure Appl. Geophys.* **158,** nos. 1/2, 19–34.

Buehler, J. S., and P. M. Shearer (2014). Anisotropy and V_P/V_S in the uppermost mantle beneath the western United States from joint analysis of *Pn* and *Sn* phases, *J. Geophys. Res.* **119,** no. 2, 1200–1219.

Burdick, S., R. D. Van der Hilst, F. L. Vernon, V. Martynov, T. Cox, J. Eakins, G. H. Karasu, L. Tylell, L. Astiz, and G. L. Pavlis (2014). Model update January 2013: Upper mantle heterogeneity beneath North America from travel-time tomography with global and USArray transportable array data, *Seismol. Res. Lett.* **85,** no. 1, 77–81.

Cleary, J. (1967). Azimuthal variation of the longshot source term, *Earth Planet. Sci. Lett.* **3,** 29–37.

Husen, S., and J. L. Hardebeck (2010). Earthquake location accuracy, *Community Online Resource for Statistical Seismicity Analysis*, available at http://www.corssa.org (last accessed June 2015), 10 pp., doi: 10.5078/corssa-55815573.

Kennett, B., and F. Ringdal (2001). Locating seismic events in the CTBT context, *Pure Appl. Geophys.* **158,** nos. 1/2, 7–18.

Myers, S. C., and C. A. Schultz (2000). Improving sparse network seismic location with Bayesian kriging and teleseismically constrained calibration events, *Bull. Seismol. Soc. Am.* **90,** no. 1, 199–211, doi: 10.1785/0119980171.

Nicholson, T., D. Clarke, and J. Townend (2008). Regional earthquake location using empirical traveltimes in a region of strong lateral velocity heterogeneity, *Geophys. J. Int.* **175,** no. 2, 560–570.

Obrebski, M., R. M. Allen, F. Pollitz, and S.-H. Hung (2011). Lithosphere–asthenosphere interaction beneath the western United States from the joint inversion of body-wave traveltimes and surface-wave phase velocities, *Geophys. J. Int.* **185,** no. 2, 1003–1021.

Richards-Dinger, K. B., and P. M. Shearer (2000). Earthquake locations in southern California obtained using source-specific station terms, *J. Geophys. Res.* **105,** no. B5, 10,939–10,960.

Ryaboy, V., D. Baumgardt, P. Firbas, and A. Dainty (2001). Application of 3-D crustal and upper mantle velocity model of North America for location of regional seismic events, *Pure Appl. Geophys.* **158,** nos. 1/2, 79–103.

Schmandt, B., and E. Humphreys (2010). Complex subduction and small-scale convection revealed by body-wave tomography of the western United States upper mantle, *Earth Planet. Sci. Lett.* **297,** no. 334, 435–445.

Schmandt, B., and F.-C. Lin (2014). *P* and *S* wave tomography of the mantle beneath the United States, *Geophys. Res. Lett.* **41,** no. 18, 6342–6349.

Shearer, P. M. (2001). Improving global seismic event locations using source-receiver reciprocity, *Bull. Seismol. Soc. Am.* **91,** no. 3, 594–603, doi: 10.1785/0120000238.

Wessel, P., W. H. F. Smith, R. Scharroo, J. Luis, and F. Wobbe (2013). Generic mapping tools: Improved version released, *Eos Trans. AGU* **94,** no. 45, 409–410.

Yang, X., I. Bondar, J. Bhattacharyya, M. Ritzwoller, N. Shapiro, M. Antolik, G. Ekstrom, H. Israelsson, and K. McLaughlin (2004). Validation of regional and teleseismic travel-time models by relocating ground-truth events, *Bull. Seismol. Soc. Am.* **94,** no. 3, 897–919, doi: 10.1785/0120030148.

Institute of Geophysics and Planetary Physics
Scripps Institution of Oceanography
University of California, San Diego
La Jolla, California 92093-0225
jsbuehle@ucsd.edu

Manuscript received 30 June 2015;
Published Online 16 August 2016