Shallow Damage Zone Structure of the Wasatch Fault in Salt Lake City from Ambient-Noise Double Beamforming with a Temporary Linear Array

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Abstract
We image the shallow structure across the East Bench segment of the Wasatch fault system in Salt Lake City using ambient noise recorded by a month-long temporary linear seismic array of 32 stations. We first extract Rayleigh-wave signals between 0.4 and 1.1 s period using noise cross correlation. We then apply double beamforming to enhance coherent cross-correlation signals and at the same time measure frequency-dependent phase velocities across the array. For each location, based on available dispersion measurements, we perform an uncertainty-weighted least-squares inversion to obtain a 1D $V_s$ model from the surface to 400 m depth. We put all piece-wise continuous 1D models together to construct the final 2D $V_s$ model. The model reveals high velocities to the east of the Pleistocene Lake Bonneville shoreline reflecting thinner sediments and low velocities particularly in the top 200 m to the west corresponding to the Salt Lake basin sediments. In addition, there is an ∼ 400–m–wide low-velocity zone that narrows with depth adjacent to the surface trace of the East Bench fault, which we interpret as a fault-related damage zone. The damage zone is asymmetric, wider on the hanging wall (western) side and with greater velocity reduction. These results provide important constraints on normal-fault earthquake mechanics, Wasatch fault earthquake behavior, and urban seismic hazard in Salt Lake City.

Introduction
Seismic tomography using surface waves from ambient-noise cross correlations has emerged as a useful tool over the past decades for imaging the Earth’s interior structure (e.g., Shapiro et al., 2005; Lin et al., 2009; Shen et al., 2013; Nakata et al., 2015; Li et al., 2016; Spica et al., 2016; Berg et al., 2020). By exploiting natural vibrations to extract empirical Green’s functions (e.g., Lobkis and Weaver, 2001), ambient-noise tomography has the advantages of not relying on the availability of ballistic active or passive sources. Combined with the recent development of portable low-cost temporary seismic instrumentation, it is now possible to acquire unprecedented high-resolution data in previously difficult or inaccessible regions providing new insight for seismic hazard assessment (Lin et al., 2013; Bowden et al., 2015; Nakata et al., 2015; Clayton, 2020; Castellanos et al., 2020), tectonic processes (Wang et al., 2019b), earthquake mechanics (Roux et al., 2016; Mordret et al., 2019; Li et al., 2019; Wang et al., 2019b), volcanic structures (Brenguier et al., 2016; Nakata et al., 2016; Wang et al., 2017; Ranasinghe et al., 2018; Wu et al., 2020), reservoir monitoring (De Ridder and Biondi, 2013), landslides (Thomas et al., 2020), critical zones (Keifer et al., 2019), and hydrothermal dynamics (Wu et al., 2017; Wu et al., 2019). In particular, it is well established that the shallow shear-velocity structure greatly affects the amplitudes of seismic waves (Tinsley et al., 1991), making shallow seismic imaging a critical component of seismic hazard assessment (e.g., Graves et al., 2011; Fu et al., 2017).

The Salt Lake City (SLC) (Utah, U.S.A.) metropolitan area is situated in the Salt Lake basin, a Cenozoic lacustrine and fluvial basin bounded by the normal-fault horsts of the Wasatch (east) and Oquirrh (west) mountains (Stokes, 1980). The upper few hundred meters of the basin comprise late Pleistocene, Lake Bonneville, unconsolidated sediments (Kowalewska and Cohen, 1998), which can dramatically amplify coseismic shaking (e.g., Johnson and Silva, 1981; Moschetti et al., 2017). With a population of ∼1.2 million and an ∼70% chance of a
In this study, we use data from a temporary linear nodal array that was deployed across 1700 South Street in SLC. The array cut across both the EBF and the Provo shoreline (Fig. 1), one of the major paleoshorelines of Lake Bonneville (Oviatt, 2015). First, we compute the cross-correlation functions and use vertical–vertical cross correlations to extract Rayleigh waves propagating along the line. A challenge in our analysis is the presence of non-diffusive noise in the ambient wavefield, likely relative to urban activities. To enhance the signal and measure phase velocities across the linear array, we use the double-beamforming method developed by Wang et al. (2019a,b), which has been successfully applied to both regional (Cascadia; Wang et al., 2019a) and local scales (San Jacinto fault; Wang et al., 2019b).

Data and Methods
A network of 32 nodal stations was deployed along 1700 South Street in SLC from 8 February to 12 March 2018 (station locations shown in Fig. 1; Lin, 2018). The linear array had an ~2.9 km aperture and spanned from 700 East to 1900 East Street. The availability of deployment locations—primarily local volunteer homeowners—resulted in some irregularity in station spacing with variation from ~10 to ~100 m.

Noise cross correlations
We compute ambient-noise cross correlations following Wang et al. (2019a). Examples of the vertical–vertical (Z–Z) cross-correlation record sections between a common station (virtual source) close to the array center and all other receiver stations across the array are shown in Figure 2, band-passed near 0.9 s (Fig. 2a) and 0.4 s (Fig. 2b). The observed Rayleigh-wave signal moveout is asymmetric, indicating stronger seismic energy propagating to the east. Besides the primary Rayleigh-wave
signals, secondary signals are also observed especially at shorter periods (Fig. 2b), likely due to the presence of persistent noise sources or active scatterers (Ma et al., 2013) in the vicinity of the study area. Although determining the nature of these secondary signals is out of the scope of our current study, future 2D dense array deployment in the area would allow us to study the radiation pattern in detail and determine if the phase is fault-zone related.

**Double-beamforming tomography**

To enhance the coherent Rayleigh-wave signals and measure phase velocities, we follow Wang et al. (2019a,b) and perform double beamforming. We form source and receiver beams across the array with a fixed 250 m beam width and 50 m beam center spacing. The beam width was chosen here to be large enough to include a sufficient number of stations (at least three stations in both the source and receiver beams). The beam width also intrinsically controls the lateral smoothing applied by the method. We adopt a less strict far-field criterion to remove all beam pairs with distance between the source and the receiver beam center smaller than either 1 or 1.5 wavelengths for periods between 1.1–0.7 and 0.6–0.4 s, respectively. Here, we use a 1 km/s reference velocity to estimate the wavelength. The slightly looser criterion for longer periods is because of the limited array aperture and hence the number of beam pairs passing the selection criterion. The application of this distance criterion can still result in including individual station pairs with distance smaller than 1 or 1.5 wavelengths, but we see no obvious disturbance to the observed Rayleigh-wave moveout when including those short-distance pairs. Following Wang et al. (2019a), we assume planar wavefront geometry and no off-great-circle propagation.

For each source–receiver beam pair, we first cut and taper the cross-correlation waveforms based on an empirically determined period-dependent maximum group velocity ($0.5$–$1$ km/s). Following Wang et al. (2019b), we then shift and stack the waveforms in the frequency domain (Fig. 3a) using different source ($u_s$) and receiver ($u_r$) phase slowness combinations, ranging from 0.4 to 5 s/km. We measure the phase slowness on both source and receiver sides using a two-step grid search (coarse and finer grid, Wang et al., 2019b) based on the maximum amplitude of the stacked waveform (beampower; example shown in Fig. 3b). To ensure that the grid search is picking the energy packet corresponding to the fundamental mode Rayleigh-wave signal, we require the dispersion measurements to be continuous across different periods. We first determine the phase slowness at periods longer than 0.9 s in which signals are clean and simple. Starting at 0.9 s, we then use the slowness from the period immediately above as the reference slowness. Only source and receiver slowness measurements within 25% of their corresponding reference slowness are accepted.

After performing the grid search for all source and receiver beam pairs across the array, the phase slowness at each location (i.e., each beam center) is determined by the mean of measurements at that location with different source-receiver beam combinations. Measurements outside of two standard deviations (st.dev.) are considered as outliers and removed. The uncertainty for each location is computed as the st.dev. of the mean divided by the square root of independent measurements (Fig. 4a), defined as the number of nonoverlapping beams (Wang et al., 2019a). To account for potential systematic biases due to an uneven source distribution, we set the minimum uncertainty

![Figure 2. Vertical–vertical cross-correlation functions between a virtual source station (gray reversed triangle in Fig. 1) and all the array receivers, band-passed (a) near 0.9 s and (b) near 0.4 s. Solid line depicts a reference velocity line (1 km/s). Rayleigh-wave signal moveout is observed in the record sections with the dominant energy propagating toward the east. The color version of this figure is available only in the electronic edition.](image-url)
Figure 3. Example of the waveform shift and stacking procedure and the 2D grid search for a beam pair at 0.5 s, in which the source beam center is located at -0.6 km and the receiver beam center is located at 1.25 km along the linear array (Fig. 2). (a) Stacked waveforms after shifting using different source ($u_s$) and receiver ($u_r$) phase slowness combinations. Dashed lines indicate the envelope function of each stacked waveform. (b) Beampower plot with varying source and receiver slowness. The black cross denotes the location of maximum beam amplitude. Iso-amplitude contours are plotted as black lines using 0.05 interval. The color version of this figure is available only in the electronic edition.
as 2% of the mean. In addition, we perform signal-to-noise ratio (SNR; Lin et al., 2009) calculations based on the stacked waveform and remove spurious measurements with SNR smaller than 5. We define SNR as the peak-stacked waveform amplitude within the signal window (velocity between 0.3 km/s and an empirically determined period-dependent maximum velocity between 0.5 and 1.1 km/s) divided by the root mean square of the noise window (20 s following the signal window). Considering the irregular station spacing and the significant beam overlap, we further smooth the phase slowness profile for each period. The smoothed phase slowness and uncertainty at each location is computed as the weighted average of the three neighboring points (the three locations are 50 m apart) and the standard error of the weighted average, respectively (Fig. 4b). The phase slowness and their uncertainties for all periods are combined and converted into phase velocity (Fig. 5a) and uncertainty (Fig. 5b) profiles before shear-velocity inversions are performed. The overall slower phase velocities to the west and faster phase velocities to the east likely reflect the thickening of sediments toward the center of the basin. On a smaller scale, a localized slow anomaly is observed near the surface trace of the EBF.

Figure 4. Phase slowness profiles for 0.5 s period. Phase slowness measurements not satisfying the SNR criterion are not shown. (a) Profile before smoothing and (b) profile after smoothing. The color version of this figure is available only in the electronic edition.

The anomaly is wider at shorter periods and narrower at longer periods (0.8–1.1 s) potentially related to the depth-dependent fault-zone damage structure.

Shear-velocity inversion
To invert for a 2D $V_S$ model across the nodal array, we first extract the Rayleigh-wave phase velocity dispersion curve at each location across the profile. Then, we invert each dispersion curve independently using the iterative weighted least-squares algorithm of Herrmann (2013) to obtain a 1D $V_S$ model. Following Wang et al. (2019b), we use a homogeneous starting model with fixed $V_p/V_S$ ratio equal to 1.75 and an empirical density calculated from $V_p$ using the relationship of Brocher (2005). Based on the geological setting...
in the area, we impose a monotonically increase constraint to the inverted 1D shear velocity models. We allow the inversion to iterate up to 80 times to get the final 1D $V_S$ model, but stop the iteration when the chi-square misfit does not improve by more than 1% compared to the previous iteration. On average, the inversion stopped after $\sim$18 iterations.

Examples of local dispersion curves and their inverted $V_S$ models are shown in Figure 6. The shear-velocity model for location A (Fig. 6a, 0.55 km distance) is generally slower compared to location B (Fig. 6c, 0.9 km distance). This is somewhat expected as location B is closer to the eastern basin edges and hence has thinner soft sediment. The observed phase velocity dispersion curves in general can be fitted well by the model predicted dispersion curves in which discrepancies are mostly smaller than the estimated measurement uncertainties. We only consider the top 400 m of the inverted models robust considering the overall depth sensitivity of our measurements (Fig. 7).

We combine all 1D $V_S$ models across the profile to create a 2D $V_S$ model (Fig. 8). The predicted phase-velocity profile (Fig. 5c) closely resembles the observed profile (Fig. 5a) with only noticeable differences where uncertainties are high.

The basin bedrock formations with shear velocities greater than 3 km/s suggested by previous geophysical studies in the area (Bashore, 1982; Hill et al., 1990; Mabey, 1992; Magistrale et al., 2009) are likely below our maximum resolved depth; a future study with a larger array aperture is needed to constrain the deeper basin structure.

The EBF is expressed in the period-dependent phase velocity (Fig. 6) and final $V_S$ profiles (Fig. 8) as a narrow asymmetric low-velocity zone. Large-scale seismogenic faults are well known to produce low velocities in their vicinity by breaking surrounding rock during coseismic shaking (e.g., Ben-Zion and Sammis, 2003); these features are termed “damage zones” and are typically on the order of tens to hundreds of meters wide depending on the size of the largest earthquakes (Faulkner et al., 2011), depth of the seismogenic zone (Ampuero and Mao, 2017), cumulative fault slip (Sagy et al., 2007; Perrin et al., 2016), and local rheology (e.g., Finzi et al., 2009; Molli et al., 2010; Thakur et al., 2020). In addition to the main mapped trace of the EBF, there are two other similar low-velocity zones (1000 to 600 m and 400 to 800 m relative to EBF), which likely correspond to other strands within the Wasatch fault zone; large-scale normal

Results and Discussion

The 2D shear-velocity model (Fig. 8) exhibits similar spatial patterns to the phase-velocity map (Fig. 5a). The overall trend in the $V_S$ profile is decreasing velocity to the west likely corresponding to sedimentary thickening toward the center of the basin (Radkins et al., 1989; Hill et al., 1990). The decreasing velocity may also represent changes in the overall composition of the shallow lacustrine deposits, which transition from younger and softer clay, silt, and fine sand in the west to older and mechanically stronger sand and gravel toward the east (McDonald and Ashland, 2008). The highest velocities ($>1.2$ km/s) in the model observed at depth in the east end of our model potentially mark the transition between shallow unconsolidated Quaternary sediments and deeper Tertiary strata, comprised of volcanic and plutonic rocks (Liberty et al., 2018b).
faults are expected to branch into “flower structures” at shallow depths due to the reduced normal stress (e.g., Twiss et al., 1992; chapter 5). Previous active-source imaging in the same area resolved 11 fault strands across a 4.5 km linear array, only one of which was mapped at the surface (Liberty et al., 2018b). We acknowledge that the observed localized low-velocity zones can also be related to other factors (Wang, 2001), including differences in sediment compaction rate and lithology (e.g., Olig et al., 1996), layering-induced anisotropy (e.g., Behera et al., 2011), and increased porosity, pore-fluid saturation (e.g., Shimeld et al., 2016). In addition, the low-velocity anomalies may represent remnant liquefaction areas from past earthquakes (e.g., Liberty et al., 2018a), and also be affected by interbedded smaller-scale structures such as colluvial wedges (e.g., Buddensiek et al., 2008) with increasing thickness proportional to past earthquake magnitude (e.g., Morey and Schuster, 1999). Additional ground truth information (e.g., drilling cores) will be required to distinguish these interpretations.

The increasing normal stress with depth also leads to a narrower damage zone with depth (e.g., Allam and Ben-Zion, 2012), either due to inhibited mode I fracture growth (e.g., Prudencio and Van Sint Jan, 2007) or due to enhanced rock healing with increasing temperature (Lyakhovsky and Ya-Hamiel, 2007). The main trace of the EBF observed here (Fig. 8) narrows with depth from ~600 m width at the surface to ~300 m width at the bottom of the resolved profile. The inferred damage zone is asymmetric with respect to the surface trace of the fault, with more and higher intensity damage in the hanging wall. This pattern is expected for normal faults and has previously been observed both geologically (Flodin and Aydin, 2004; Berg and Skar, 2005) and in numerical simulations of dipping faults (Xu et al., 2015). Because of the complicated damage pattern and the limited resolution of our tomographic image, it is difficult to constrain the precise dip of the EBF and other subsidiary strands from our result. However, the central portion of the damage zone is either vertical or dipping slightly to the west. If this is the case, it is further support for a listric structure to the Wasatch fault zone (Mohapatra and Johnson, 1998; Pang et al., 2020), listric structure to normal faults more generally (Wernicke, 1981; Davison 1986; Bose and Mitra, 2010), and is in line with free-surface orthogonality expectations from Andersonian faulting theory (Leung and Su, 1996).

The Provo shoreline is observed as a sharply bounded high-velocity region in the east within our 2D $V_S$ model (Fig. 8) with much higher velocity at all resolved depths (particularly at depth). Because of this relatively sharp lateral boundary, which persists to at least 400 m depth, we interpret this boundary as an additional fault strand related to the EBF or Wasatch fault zone more generally. Previously imaged non-fault shoreline boundaries in SLC have gradually thickened sediments westward from the shoreline (Liberty et al., 2018b). A recent geodetic study (Hu et al., 2018) indicates that the EBF is also hydrological boundary controlling surface deformation in the area.

**Conclusions**

We present new results of noise-based shallow imaging in Salt Lake Valley near the vicinity of the EBF using a dense linear array. We enhanced our noise cross-correlation signals and measured Rayleigh-wave phase velocities between 0.4 and 1.1 s period across the array using the double beamforming method.
Despite the less known and potentially complex noise wavefield associated with the inland metropolis, the observed phase velocity profile and 2D $V_S$ model constructed between the surface and 400 m depth are consistent with the geologic and the geotectonic models of the area. This indicates the ability of the method to produce reliable shallow crustal images even in an urban environment far away from the ocean microseism. We provide new constraints on the shallow structure of the EBF, which has an ~400 m damage zone that gets narrower with depth. The resolution achieved in this study enabled the imaging of this low-velocity fault-zone structure for the first time, because its spatial extent is smaller than the resolution limit of previous studies (e.g., Hill et al., 1990; Magistrale et al., 2009). Our results, along with other recent studies (Liberty et al., 2018a; McDonald et al., 2020; Pang et al., 2020), indicate that the fine-scale structure of the WFS is a complicated series of fault strands and their associated damage. The success of this study motivates future work in the area to better understand the structure and the seismic hazard associated with the WFS. The deployment of a network with wider aperture and in 2D is needed to study the WFS at greater depth, constrain its lateral variation, and better understand the SLC urban noise characteristics.

Data and Resources

Seismic data from this network (DOI: 10.7914/SN/9H_2018) will be available to download from Incorporated Research Institutions for Seismology (IRIS). Resources used for maps are available from the Utah Automated Geographic Reference Center (AGRC) (https://gis.utah.gov/) and the Utah Geological Survey (UGS). The surface wave inversion tool used in this study is part of “Computer Programs in Seismology” (Herrmann, 2013) and available from http://www.eas.slu.edu/eqc/eqccps.html. All websites were last accessed in October 2020.

Figure 7. $V_S$ sensitivity kernels for Rayleigh-wave phase velocity at 0.5, 0.7, and 1 s periods at two example locations. (a) ~0.65 km and at (b) 0.9 km from EBF.

Figure 8. 2D shear-velocity model constructed by piece-wise continuous 1D inversions. Shear-velocity contours are separated by 200 m/s.
Declaration of Competing Interests
The authors acknowledge there are no conflicts of interest recorded.

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