

New Insights from Legacy Seismic Data regarding Basalt Elevations and Variability on the Hanford Site



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ABSTRACT

Migration of groundwater contaminants in the Gable Gap area of the Hanford Site in southeastern Washington State is strongly influenced by the distribution and permeability of basalts that lie beneath an unconfined aquifer. Locally, folding and faulting of the Columbia River Basalt associated with the Yakima fold and thrust belt followed by erosion due to the Lake Missoula floods resulted in a complex basalt surface that represents either an impermeable lower boundary to the unconfined aquifer system or localized regions of increased permeability that potentially promote communication between the unconfined aquifer system and deeper, confined aquifer systems. Paleo-channels carved into the basalt by floodwaters are thought to provide preferential flow paths for groundwater contaminants. In 2011, a seismic landstreamer campaign was carried out to image the basalt surface and produced pre-stack depth migrated p-wave reflection images. The reflection images identified two large troughs that may represent paleo-channels and several areas of possible faulting. Here, the streamer data are re-analyzed using refraction travel-time and Rayleigh wave dispersion analyses to obtain images of compressional and shear wave velocities within the suprabasalt sediment sections and the upper basalt surface. The combined interpretation of reflection and seismic velocity images shows complexity in the basalt velocity and elevation, which varies by 50 m or more within the study area. These results, along with other ongoing geophysical investigations, will be used to inform the site geologic model and potentially guide placement of future boreholes needed to quantify vertical flow between the confined and unconfined aquifers.

INTRODUCTION

The Hanford Superfund site in southeastern Washington State is where plutonium was produced for use in U.S. nuclear weapons during World War II and throughout the Cold War. Liquid waste disposal associated with these activities has resulted in groundwater contaminant plumes within the unconfined aquifer emanating from the 200 East Area (Figure 1) (Hartman et al., 2009). The relatively impermeable nature of the underlying basalt compared to the overlying sediments supports the conceptual model of the basalt acting as a lower boundary of the local aquifer system (DOE/RL, 2012, Appendix E). In some regions, basalt elevations are higher than the water table, and flow is confined largely to the regions where Pleistocene flooding associated with Lake Missoula scoured channels through the basalt (Bjornstad et al., 2010). These channels and other local geologic structures are thought to provide preferential flow paths for contaminant plumes to travel north toward the Columbia River (Bjornstad et al., 2010). In addition, regional erosion of underlying basalt layers during Pleistocene flooding events exposed several of the deeper basalt units and interbeds to the unconfined aquifer above, which may allow vertical communication between the unconfined aquifer and deeper, confined aquifers within the Columbia River Basalt (CRB) group (Graham et al., 1984). Given these heterogeneities and potential impacts of contaminant migration to the Columbia River, a detailed examination of paleo-channel geometry is warranted as well as identification of regions that potentially support communication between deep, confined aquifer systems and the overlying unconfined system.

Surface-based seismic methods can provide useful information about subsurface structure as it relates to key hydrologic parameters (e.g., Hubbard and Linde, 2010). Seismic reflection methods utilize waves that reflect off subsurface boundaries to produce images of subsurface stratigraphy and structure. Seismic refraction methods utilize waves that are transmitted through the subsurface, turning back toward

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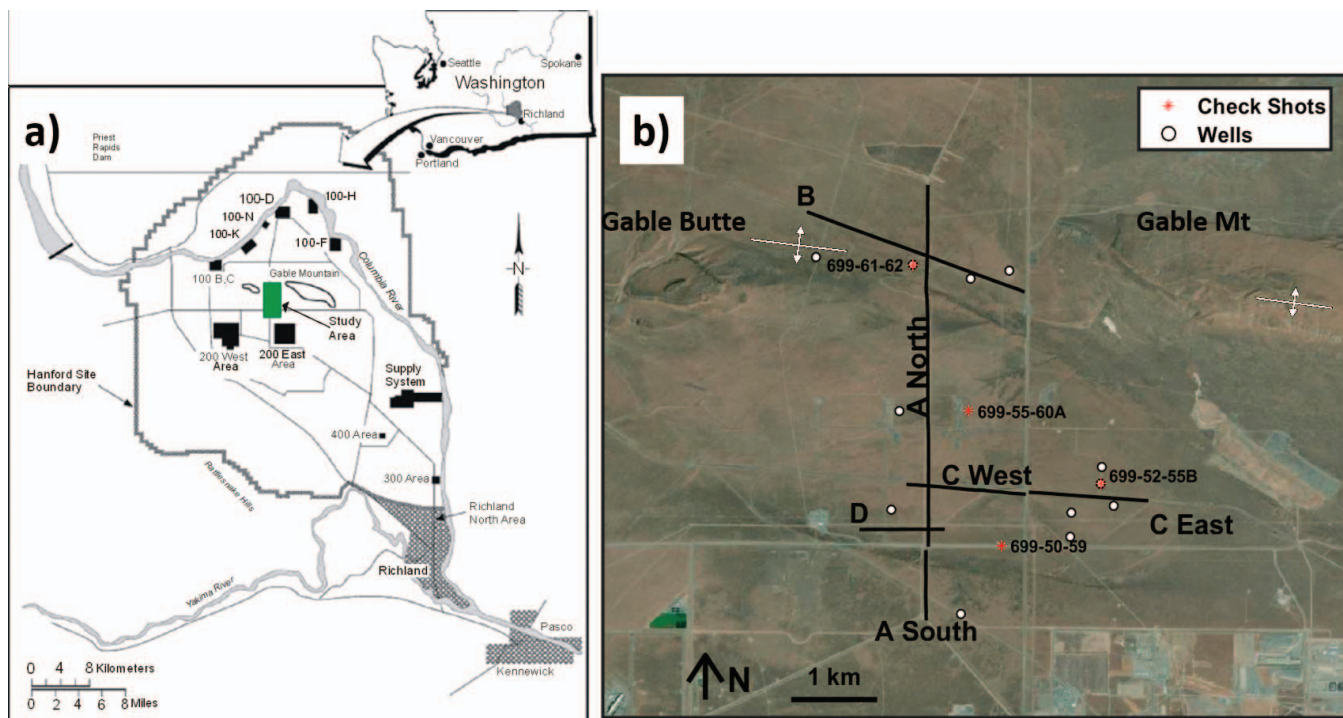


Figure 1. (a) Location of the Hanford Site in southeastern Washington (after Sunwall et al., 2011). The Gable Gap area is shown in green. (b) Satellite view of the Gable Gap area showing the seismic profile locations (black lines) and the locations of check shots (red stars) and wells (white circles with black outlines) used to validate seismic models. Gable Mountain Anticline is indicated by symbols.

the surface as they encounter faster-velocity materials at depth (Steeple, 2005). Travel times of refracted waves can be used to image the seismic velocity structure within the subsurface. It is most common to measure compressional wave velocity (V_p) with refraction methods. Shear wave velocity (V_s) can be measured with dispersive Rayleigh waves (Park et al., 1999). Seismic velocities vary systematically with lithology, porosity, saturation, and pressure (e.g., Mavko et al., 2009); thus, they are useful for identifying bedrock depths, the water table, and zones that are potentially more permeable.

Several seismic reflection surveys have been conducted in the Gable Gap area of the Hanford site. During 1979 and 1980, approximately 80 km of seismic reflection data were acquired as part of the Basalt Waste Isolation Project to identify deep storage targets for spent fuel (SSC, 1979, 1980). More recently, several shallow seismic investigations have been carried out to identify basalt elevations and structure for hydrologic investigations (Cummins, 2009; Hyde et al., 2011). Hyde et al. (2011) showed the utility of the seismic landstreamer for the rapid collection of seismic data on the Hanford site. Their results showed a rugged and faulted basalt surface along 12 km of seismic profiles. A subsequent landstreamer study (Sunwall et al., 2011) added 12 km of seismic profiles within the Gable Gap area and further supported a rugged and faulted

basalt surface. A 2012 study that integrated all of the existing seismic reflection data in the area and compared known basalt depths to the seismic interpretations concluded that the reflection data often overestimated basalt depths, possibly due to an inaccurate seismic velocity model used to convert travel times to depth (Williams et al., 2012).

In this study, 12 km of legacy landstreamer data (Sunwall et al., 2011) are analyzed to measure V_p and V_s from the refracted p-waves and the dispersive Rayleigh wave signals. The resulting velocity models provide an estimate of basalt depth that is independent from the pre-stack depth migrated (PSDM)-derived estimates. Comparing these new results to borehole observations and integrating them with the reflection images provides an updated interpretation of depth to basalt within the Gable Gap and 200 E areas that will be used to refine the site geologic model. Zones of interest are also identified where additional characterization is needed to evaluate the possibility of vertical communication between the unconfined aquifer and deeper aquifer systems.

Hydrogeologic and Geophysical Setting

The hydrostratigraphy in the Gable Gap area consists primarily of the sand- and gravel-dominated Hanford units overlying the faulted and folded CRB group

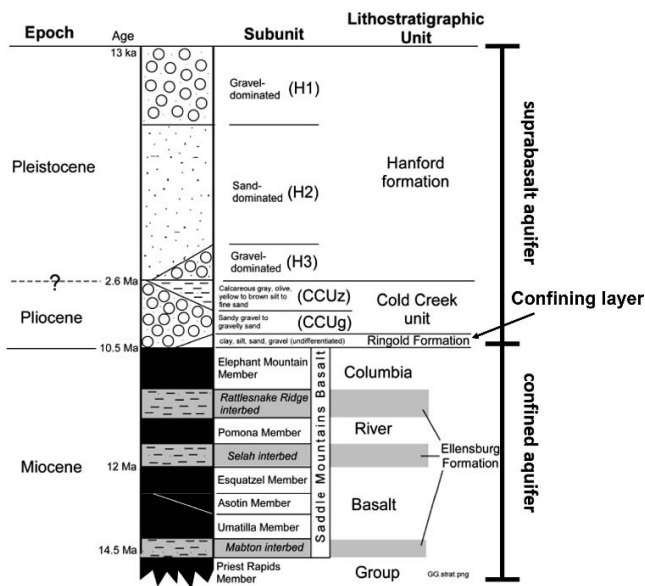


Figure 2. Generalized hydrostratigraphy in the Gable Gap area (after Bjornstad et al., 2010).

with sediment interbeds (Figure 2). Older suprabasalt sedimentary units in the area are the Cold Creek unit and the Ringold formation, which are sparsely distributed, having been eroded during Pleistocene flooding associated with Lake Missoula. The top of the basalt is considered to mark the bottom of an unconfined, suprabasalt aquifer. A confined aquifer also exists within the basalts and interbedded sediments.

The Gable Gap (Figure 1) lies within the Yakima fold and thrust belt. To the north of the study area, the asymmetric east-to-west Gable Mountain anticline (GMA) folds the CRBs and the interbedded Ellensburg formation. The northern dipping limb of this anticline is steep, whereas the southern limb has a relatively gentle dip. Second-order syncline and anticline pairs have been inferred to exist along the southern limb and throughout the study area (e.g., Ault, 1981). Previous studies have hypothesized that secondary normal and thrust faults may also exist along the southern limb of the GMA (e.g., Hyde et al., 2011).

The principal CRB unit of hydrologic interest in the Gable Gap area is the Elephant Mountain member, which is considered the base of the unconfined, suprabasalt aquifer (Bjornstad et al., 2010). However, erosion related to Pleistocene flooding of Lake Missoula has locally eroded the Elephant Mountain member, exposing older basalt units and sediment interbeds to the unconfined aquifer, and may allow vertical communication between the unconfined and confined aquifers in these areas (Graham et al., 1984).

Above the CRB, the unconfined suprabasalt aquifer is hosted in the flood-deposited Hanford formation. The Hanford formation contains both gravel- (H1 and

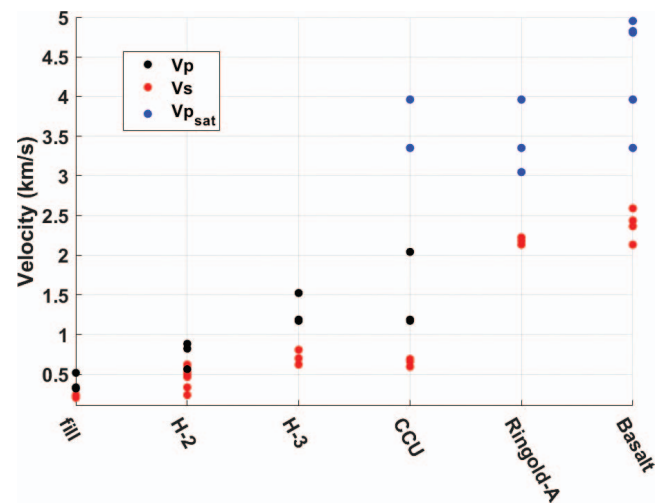


Figure 3. Summary of downhole seismic velocity measurements made in boreholes near the Gable Gap area (Rohay and Brouns, 2007). Unsaturated Vp for the Hanford (H-2 and H-3) and Cold Creek sediments are shown in black, and saturated Vp in the Cold Creek, Ringold (Rg), and basalt units are shown in blue. Vs for all units is shown in red.

H3) and sand- (H2) dominated units, and their distribution reflects flood dynamics. The older Ringold and Cold Creek units are river deposits containing gravels, sands, and silts. However, within the Gable Gap area, they were largely eroded during the Lake Missoula floods and are unlikely targets for refraction imaging.

Rohay and Brouns (2007) used check shots to measure Vp and Vs in three boreholes in the 200 East Area south of the landstreamer profiles (Figure 1a). Vp data were collected using a sledgehammer source, and Vs data were collected using a horizontal accelerated weight drop. The boreholes penetrated the Hanford units, the Cold Creek unit, the Ringold formation, and the CRB. Their measurements reflect unsaturated conditions for the Hanford units and the Cold Creek unit and saturated (below the water table) conditions for the Cold Creek unit, Ringold formation, and CRB (Figure 3). The influence of the water table can be seen for the Cold Creek formation; Vp in unsaturated Cold Creek is 1,200–2,000 ms⁻¹, while saturated Cold Creek ranges between 3,400 and 3,900 ms⁻¹. Since fluids do not support shear stresses, Vs is largely unaffected by fluid saturation.

The Rohay and Brouns measurements indicate that Vp should be a reliable means of distinguishing basalt from the Hanford formation. Unsaturated Cold Creek is also easily distinguished from basalt. Vp in saturated Cold Creek and the Ringold formation overlap with the basalt measurements, indicating that it will not be possible to differentiate them using Vp measurements. Additionally, Ringold Vs overlaps with basalt Vs, suggesting that even unsaturated Ringold cannot be

confidently distinguished from basalt on the basis of seismic velocity alone.

Legacy Seismic Data

The landstreamer is a rapid seismic acquisition system that consists of a string of geophones and an accelerated weight drop (AWD) towed behind a vehicle. The geophone spacing is fixed so that after each shot, the entire system is moved forward, and the source-receiver offsets remain constant. The streamer survey geometry is suitable for reflection imaging, compressional wave refraction imaging, and Rayleigh wave dispersion analysis.

Approximately 12 km of landstreamer data were collected in April and May 2011 by the Confederated Tribes of the Umatilla Indian Reservation and Montana Technological University. The streamer comprised 96 gimballed 30-Hz geophones spaced 2 m apart. The streamer was towed by a pickup truck, which also towed an AWD 6 m in front of the streamer. The AWD was a 227-kg steel ram lifted by a hydraulic pump and accelerated by an elastic band. The AWD was used to vertically strike a 60×60 -cm steel plate at 2-m intervals along each profile. This source produces high-amplitude seismic energy in the 5- to 150-Hz band and facilitates rapid data acquisition. The data set includes 5,564 shot gathers, and each shot record contains 2 seconds of data sampled every 0.5 ms. The minimum offset was 6 m, and the maximum offset was 196 m.

Sunwall et al. (2011) processed the reflection data and produced PSDM images. However, the raw data (Figure 3) show clear direct and refracted arrivals with apparent V_p that is consistent with suprabasalt sediments ($\sim 1,200 \text{ ms}^{-1}$) and the underlying basalt ($\sim 4,500 \text{ ms}^{-1}$). Dispersive surface waves are also evident and suggest suprabasalt V_s between 200 and 600 ms^{-1} . In this article, the first-arrival travel times and Rayleigh wave dispersion curves are used to image V_p down to basalt and V_s in the upper 15–30 m, respectively.

METHODS

Seismic Tomography

Seismic refraction methods utilize the travel time of the first arriving body wave for each source–receiver pair. Seismic velocities generally increase with depth, causing seismic rays to turn, or refract, back toward the surface as they propagate. At short source–receiver distances, the first arrival often represents the wave traveling directly from the source to the receiver, and the slope of the distance–time curve is the inverse of

seismic velocity. At longer distances, it is often observed that the travel-time curve becomes flatter, indicating that the waves have traveled through a higher-velocity medium. Seismic velocities can be measured directly from the slopes of the travel-time curves. However, when there is lateral variability in the subsurface, a more robust approach is required. Tomography is an iterative method where the subsurface is represented by many small elements of constant velocity. An initial velocity structure is chosen (usually an increase in velocity with depth), and travel times for the initial model are predicted. The difference between the predicted and the observed arrival times is used to find a model update that reduces the misfit. The process of predicting travel times and updating the model is iterated until the data misfit becomes acceptable.

Finding an appropriate update to the model requires solving an inverse problem that is generally poorly constrained. The data contain errors, some regions of the model may not contribute to the travel-time prediction, and solutions are non-unique (e.g., de Wit et al., 2012). The problem can be made stable by including smoothness constraints on the model, known as regularization. Regularization constraints place a penalty on model parameters that are very different from their neighbors. There is a trade-off between data misfit and model smoothness, and it is possible to find many velocity models that explain the data.

The conventional approach to regularizing the seismic tomography problem is to minimize the L2 norm (square root of the summed values) of the velocity gradient. This approach will not allow sharp boundaries, and because the L2 norm seeks to normally distribute velocity gradients throughout the model, it tends to produce smooth gradients in areas where the data do not constrain the model. An alternative approach is to minimize the L1 norm of the velocity gradient (sum of the absolute values), which allows sharp boundaries to develop and suppresses gradients in areas where the data do not constrain the model. In the Gable Gap area, the transition from suprabasalt sediments to basalt is likely a sharp velocity contrast for several reasons. First, borehole geophysical measurements indicate that the transition from suprabasalt sediments to basalt is sharp (Rohay and Brouns, 2007; Hyde et al., 2011; Figure 2). The raw seismic shots also contain several indicators of a sharp seismic boundary; the apparent basalt V_p measured on the raw data is $\sim 4,500 \text{ m s}^{-1}$, and the apparent sediment velocity is $\sim 1,200 \text{ m s}^{-1}$ (Figure 3). Much of the data also contain a converted shear wave phase with an apparent velocity of $\sim 2,200 \text{ m s}^{-1}$ (Figure 3). The presence of this phase indicates that V_s in the basalt is greater than the V_p of the material immediately above it (St. Clair and Liberty, 2019).

A two-dimensional tomography code (St. Clair, 2015) written in MATLAB was used to invert manually interpreted first-arrival observations. The code predicts travel times using the shortest path raytracing method (Moser, 1991), and the inverse problem is regularized with first-order derivative operators in the vertical and horizontal directions. The L1 constraint on model gradient is implemented with an iterative least squares algorithm (Ajo-Franklin et al., 2007). The high spatial density of the data results in many redundant raypaths; thus, travel times were interpreted on every second to third shot gather.

Travel-time tomography, like many geophysical methods, suffers from uncertainty and non-uniqueness. Uncertainty is due to noise in the data, the possible correlation of different model parameters, and non-uniform sensitivity of model parameters to the data. The same factors contribute to non-uniqueness of the solution. Because these issues are inherent in the method, the approach used to model the data needs to be consistent with the geologic setting, using information not contained in the travel times (e.g., the presence of converted phases) as a constraint for model selection and validation.

Non-uniqueness in the tomography solution occurs both because of errors and uncertainty in the input data and because many different velocity structures can have identical travel-time curves (e.g., Shearer, 1999). The choice of an L1 constraint on model smoothness favors models with a sharp boundary. The range of acceptable models is further limited by preferring solutions with the minimum amount of structure required to fit the data within estimated uncertainty. The picking error is estimated to be on the order of 1–3 ms; thus, the root mean square errors for the models should be ~ 3 ms. Finally, regions of the models are masked if they do not contribute to the data misfit (i.e., where no rays are present) using the derivative weight sum. The derivative weight sum represents the total length of raypaths that pass through a given model cell.

The data set contains several points where profiles intersect (Figure 1b). Comparing the velocity estimates and interpreted depths to basalt at these points can give some insight into the precision of the approach. Since each profile was inverted independently, the coincident measurements can be considered as independent observations. Finally, models are validated by comparing the predicted depth to basalt to nearby well observations and, where available, to velocity profiles derived from check shot data in nearby wells.

Rayleigh Wave Dispersion Analysis

Rayleigh waves are surface waves that have both vertical and horizontal components of motion. They

travel at phase velocities, which are slightly slower than shear waves, and the lower-frequency (longer-wavelength) components are sensitive to greater depths. A typical Rayleigh wave dispersion curve will have higher phase velocities at low frequencies and lower phase velocities at high frequencies.

The multichannel analysis of surface waves (MASW) approach (Park et al., 1999) was used in this analysis. A linear radon transform approach (Mikesell et al., 2017) was used to map the raw shot gathers into the frequency-phase velocity domain (Figure 4b–d), and the dispersion curves were manually interpreted. The dispersion curves were then inverted for one-dimensional Vs profiles that represent the average Vs structure across the width of the streamer aperture (196 m). The results are displayed as pseudo-two-dimensional Vs images with each one-dimensional model mapped to the midpoint of the streamer for the corresponding shot gather. Due to the high spatial density of the data set, the dispersion curves were interpreted for every second or third shot gather where noise level allowed a confident interpretation to be made. Some portions of the data were too noisy to interpret, and final images are masked to reflect the absence of data.

Like travel-time tomography, MASW requires an inverse solution to find the optimal Vs-depth profile that agrees with the measured data. The inversion is iterative and requires a method for predicting the dispersion curve for any given model. The propagator matrix approach described in Aki and Richards (2002) was used to predict dispersion curves, and first-order difference operators were used to constrain the inversions to be smooth. The Rayleigh wave phase velocity is sensitive to Vs, Vp, and density. Since the influence of Vp and density is small compared to Vs (Xia et al., 1999), the Vp/Vs ratio was fixed at 2, with density increasing linearly from 1,900 to 2,100 m kg^{-3} from the surface to a depth of 35 m. Models are parameterized as layers that increase in thickness from 1 to 3 m with increasing depth.

While the 30-Hz geophones used in this survey limit the amount of low-frequency content required for deeper MASW sensitivity, Ivanov (2008) demonstrated that dispersion curves can be interpreted down to about an octave lower than the natural frequency of the geophones. Frequency-phase velocity images for the data presented here show interpretable dispersion down to ~ 15 –20 Hz (Figure 3). To indicate where the models are sensitive to the data, the partial derivatives of data misfit with respect to model parameters are summed for each model layer, and layers that have little influence on data fit are masked. This procedure suggests that the dispersion data are most sensitive to the upper 10–20 m. Basalt depths are 50–100 m, so

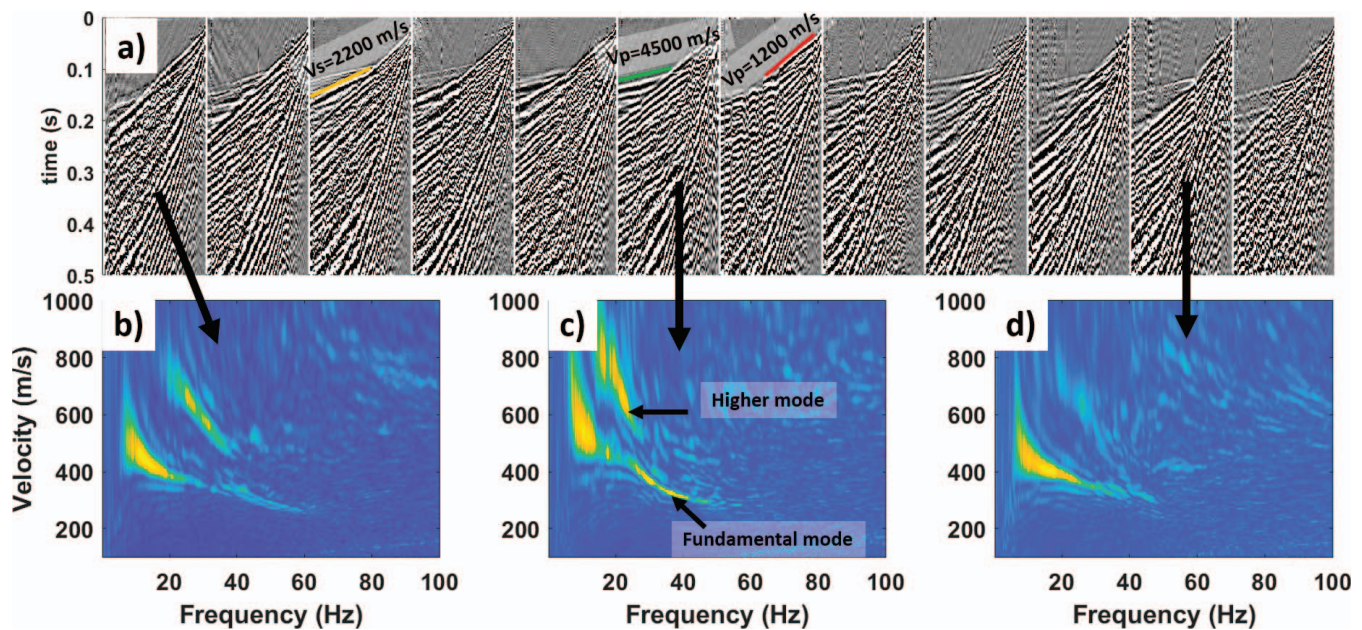


Figure 4. (a) Every 100th shot gathers along profile B, showing clear first arrivals with apparent V_p consistent with suprabasalt sediments ($1,200 \text{ ms}^{-1}$) and basalt ($4,500 \text{ ms}^{-1}$). Only 0.5 seconds of the 2-second record is shown. Many of the shot gathers in this data set also contain a converted phase with an apparent velocity around half that of the first arrival. This P-SV-P phase suggests a sharp transition from sediment to basalt. The data also contain dispersive Rayleigh waves suitable for imaging shallow V_s structure (b–d).

the data do not constrain basalt properties; however, they do provide information about shallow sediment properties.

The MASW approach produces one-dimensional models that smear geologic structure over the width of the geophone aperture of 196 m. Thus, the resulting images are unlikely to capture strong lateral changes in V_s but will highlight long-wavelength structure.

RESULTS

The refraction-generated V_p and MASW-generated V_s models for Line A North and Line B are presented (Figures 5 and 6) and compared to the PSDM images of Sunwall et al. (2011) as well as nearby borehole observations of basalt elevation and velocity profiles obtained from downhole check shots. Figure 1 displays the locations of all wells and check shots that are compared to the seismic results. Results for the other profiles indicated in Figure 1 are displayed in the supplementary information. <https://www.aegweb.org/eeg-supplements>

Line A North

Line A North (Figure 5) is an approximately 4.1-km, north-to-south profile (see Figure 1). It is the longest profile in the data set and crosses two troughs interpreted as paleo-channels (distances of 400–800 m and 2,000–3,000 m in Figure 5a). V_p in the suprabasalt

sediment section ranges from 400 and $1,500 \text{ m s}^{-1}$, and basalt V_p ranges between 3,000 and $4,500 \text{ m s}^{-1}$. The $3,000\text{-m s}^{-1}$ V_p contour is intermediate between sediment and basalt velocities and typically lies at the center of the steepest vertical velocity gradient; thus, this velocity contour was selected to interpret the transition from sediment to basalt.

Lower basalt V_p is apparent along the edges of regions where the depth to basalt is not well constrained (distances of 400–800 m and 2,000–3,000 m in Figure 5a). Here, the data lack the coverage necessary to adequately constrain basalt properties. In contrast, the lower basalt velocity near 3,400–3,500 m is well constrained and nicely correlates with a discontinuity in the PSDM reflection image (yellow colors representing $V_p \sim 2,500 \text{ m/s}$; Figure 5a). This area may represent a locally permeable basalt feature. There are two regions along A North where the refraction data do not constrain depth to basalt between profile distances of 300 and 1,000 m and 2,000 and 3,000 m. Here, either the basalt may be too deep to image with the streamer offset or the basalt velocity may be lower compared to other regions. It is also noteworthy that in these areas, the PSDM image shows a less continuous reflector, suggesting that the basalt may be fractured or otherwise damaged through erosion or faulting, supporting the interpretation that basalt V_p is lower.

There are three locations along this profile that are intersected by other lines, and the depth to the

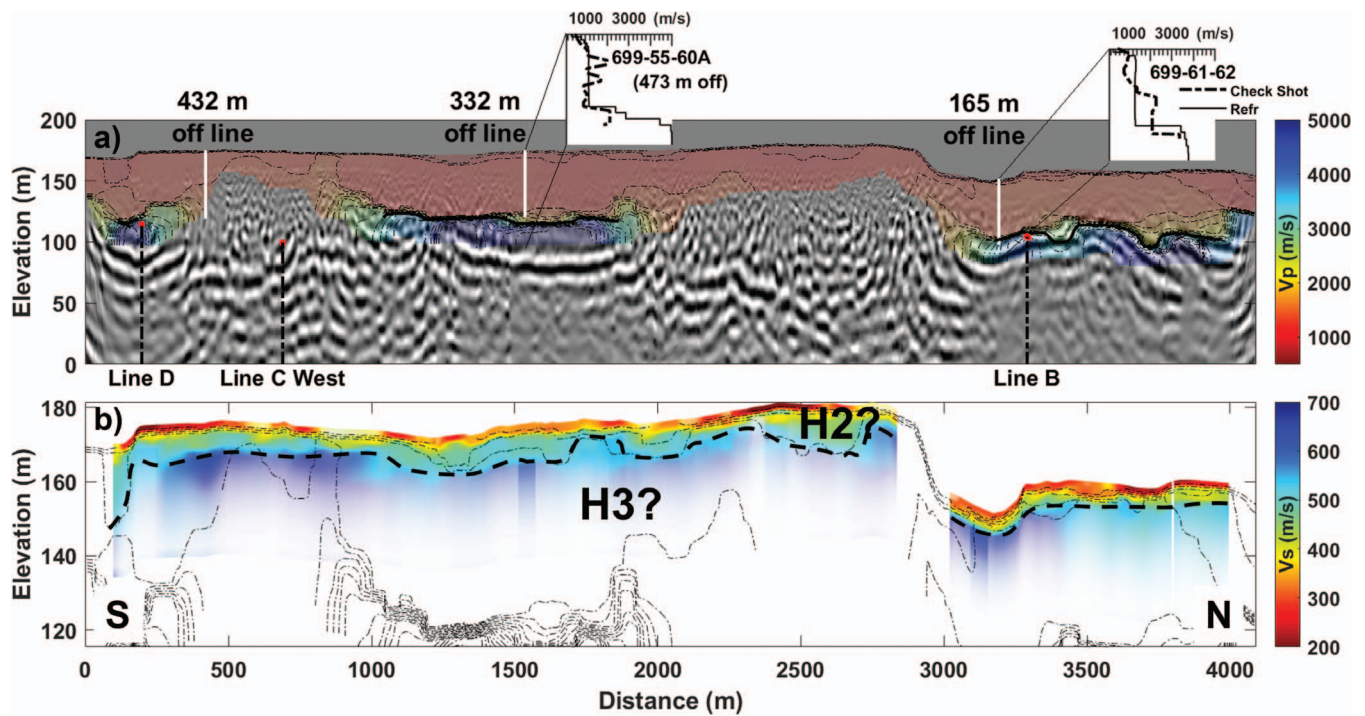


Figure 5. (a) Line A North Vp result overlaid on PSDM image. White lines indicate basalt elevations observed in nearby wells. Thick black line highlights the 3,000 m s⁻¹ Vp contour used to interpret basalt elevation. Red dots indicate interpreted depths to basalt on crossing profiles. Check shot Vp results compared to refraction-derived Vp are shown in insets. (b) The Vs image with Vp contours overlaid (thin dashed lines). Dashed black line is interpreted transition between the sandy H2 and the gravel-dominated H3.

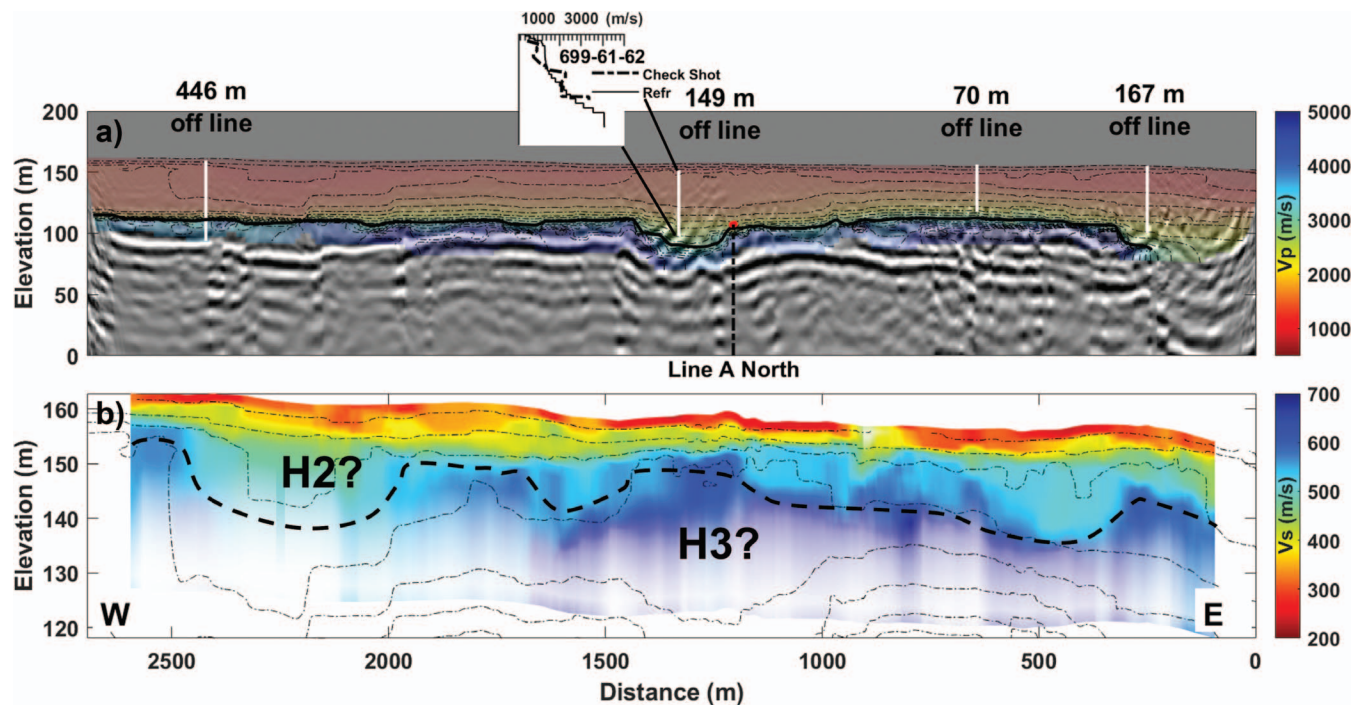


Figure 6. (a) Line B Vp result overlaid on PSDM image. White lines indicate basalt elevations observed in nearby wells. Thick black line highlights the 3,000 m s⁻¹ Vp contour used to interpret basalt elevation. Red dots indicate interpreted depths to basalt on crossing profiles. Check shot Vp results compared to refraction derived Vp are shown in insets. (b) The Vs image with Vp contours (thin dashed lines) overlaid. Dashed black line is interpreted transition between the sandy H2 and the gravel-dominated H3.

3,000-m s^{-1} contour along those profiles is shown by red dots. Where lines B and D cross, interpreted depth to basalt on all three profiles is very close. The 3,000-m s^{-1} contours are within 2 m at the Line D crossing and 4 m at the Line B crossing. Line C West has the depth to 3,000-m s^{-1} at approximately the same depth as the reflections in the PSDM image.

Three well observations of basalt elevation are within 500 m of Line A North to compare to the refraction result in Figure 5a. In two cases, the well observations correspond to locations where depth to basalt is well constrained by the refractions. Well 699-55-60A did not reach basalt; thus, this well represents a minimum basalt depth. The depth to basalt in well 699-61-62 and the 3,000 m s^{-1} are within 5 m, or ~ 10 percent, of the total observed depth. The difference may be attributed to uncertainty in the seismic result or to real variation in basalt elevation between the seismic profile and the well. A third nearby well cannot be compared to the Vp result, as the refraction data do not image basalt at that location.

Two check shot velocity profiles are also available along Line A North (Figure 5a). Well 699-55-60A is ~ 475 m to the east of the profile and did not reach basalt. It shows two thin, high-velocity ($V_p = 2,000$ m s^{-1}) layers that are not evident in the refraction result. Well 699-61-62, which did reach basalt at a depth of 54 m (Figure 5a, inset), indicates two Vp layers within the sediment section ($V_p < 1,200$ and $V_p = 2,000$ m s^{-1}). The refraction result is intermediate between these two velocities. The basalt depths and velocity estimates in well 699-61-62 closely match the refraction result.

Vs along this profile ranges between 150 and 650 m s^{-1} (Figure 5b). The 1,000-m s^{-1} Vp contour closely resembles the transition between Vs less than 500 m s^{-1} and Vs greater than 550 m s^{-1} , suggesting that this transition differentiates the H2 from the H3 formation.

Line B

Line B is an approximately 2.7-km-long northwest-southeast trending profile in the northern part of the study area (Figure 1). Here, a similar distribution of velocities is observed in the suprabasalt sediment section as Line A North. The top of the basalt shows a prominent dip at profile distance $\sim 1,350$ m, mirroring the structure depicted in the PSDM image (Figure 6a). Recovered velocities in that area are lower compared to adjacent sections. Combined with the discontinuous structure indicated by the PSDM image, it suggests that this is an area where the basalt surface is irregular and possibly fractured and permeable. On the eastern edge of the profile, either the basalt

is too deep for imaging or the velocity of the upper basalt is lower compared to elsewhere along the line.

Four well observations of basalt elevation are within 500 m of Line B (Figure 6a). The well projected onto profile distance 2,400 m is 446 m away and suggests that the refraction result underpredicts the basalt depth and agrees more closely with the PSDM image. The well is far away and may not represent the basalt depth at the seismic profile. Elsewhere along the line, the nearby wells are in better agreement with the refraction result.

Well 699-61-62 is 149 m to the south of Line B. It is evident that the refraction result represents a smoothed version of the velocity profile obtained from the check shot. The refraction result here shows a smooth transition into basalt velocities, which may explain why it overpredicts basalt depth and demonstrates the uncertainty inherent in the refraction method. It is also possible that basalt elevation is different at the well site than it is along Line B.

Vs along this profile shows a similar correlation to Vp structure, indicating that the H1 and H2 formations are not uniformly thick. Near the center of the profile, a high-velocity anomaly in both Vp and Vs suggests a thinning of the H2 layer.

DISCUSSION

Comparison to PSDM

Williams et al. (2011) compiled all the available seismic reflection data in the Gable Gap area and compared them to well observations and found that the reflection data consistently overpredicted the estimated depth to basalt. Much of the data that Williams et al. (2011) reviewed were time-stacked images converted to depth using a smooth-velocity model derived from check shots. Williams et al. (2011) suggested that the depth prediction errors were likely due to a poor time-to-depth-conversion velocity model. That argument does not apply to the PSDM images, as the PSDM approach simultaneously and iteratively produces a high-resolution velocity model directly from the data (e.g., Sheriff and Geldart, 1995). The final output is a reflection image in depth that can be directly compared to the refraction-derived Vp images.

Compared to the interpreted refraction Vp images, basalt depths predicted using the PSDM images do tend to be greater. However, this is not universally true. Along the northern-most end of Line A North (Figure 5a) and along Line C East (Figure S3), the refraction Vp results are remarkably similar to the PSDM images. Ignoring the depth discrepancies, a close similarity between topography along the basalt

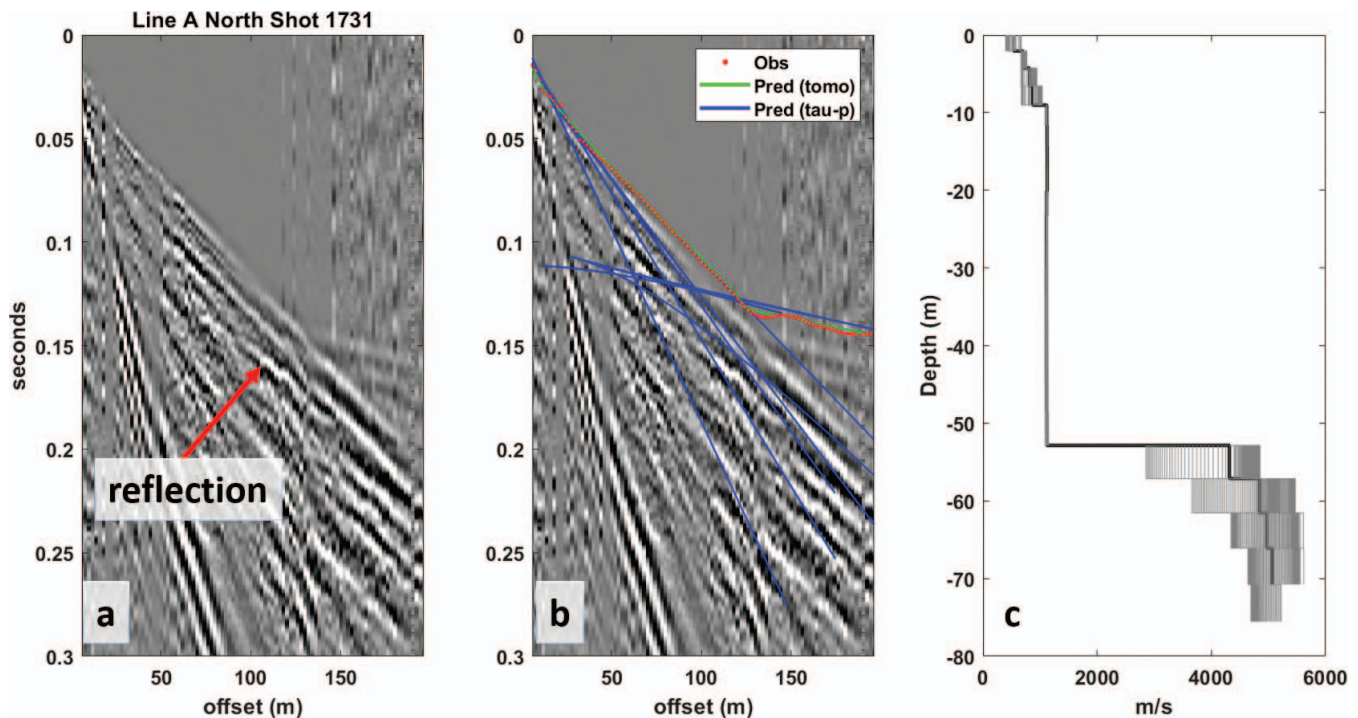


Figure 7. Shot gather 1731 along Line A North with a 200-ms agc. This shot gather corresponds to a profile distance of $\sim 1,362$ m in Figure 5a. (a) The raw data where a reflection can be seen at ~ 0.15 second and 100-m offset. (b) The same shot gather with picked travel times (red dots), predicted travel times for the tomography model (green line), and predicted arrival times of all phases generated by the tau-p approach (blue lines) for the extracted velocity model shown in panel c. The hyperbolic reflection generated by the steep velocity gradient at ~ 52 -m depth agrees with the reflection observed in the raw data. Black curve in panel c indicates the average vertical velocity gradient across the width of shot 1731, and gray lines show vertical velocity profiles extracted at 2-m intervals across the width of the streamer at shot 1731.

surface interpreted from the PSDM images and from the refraction V_p images was observed. This suggests that the structural complexity indicated by both approaches can be reliably interpreted even if the absolute elevations are uncertain.

Raw shot gathers contain visible basalt reflections and can be used to validate the V_p -derived basalt depth estimates. Travel times for a one-dimensional velocity profile extracted from the tomography result were predicted from Line A North in an area that shows relatively little lateral structure. The V_p was extracted along the aperture of a shot gather (196 m) centered at the midpoint between the source and the farthest offset receiver for shot gather 1731. This midpoint lies at profile distance of $\sim 1,362$ m in Figure 5a. Predicted travel times for all possible refracted and reflected phases were generated for the averaged, one-dimensional model using a tau-p approach (Shearer, 1999) and were overlain on the raw data. Figure 7a shows the trace-normalized shot gather with a 200-ms AGC applied. The basalt reflection is visible at ~ 0.15 seconds and 100-m offset. Figure 7b shows the predicted travel-time curves for the one-dimensional model (blue lines); refracted phases have

straight slopes, and reflected phases are hyperbolic. The reflection time predicted by tau-p closely matches the reflection observed in the shot gather. Figure 7b also shows the observed first-arrival times compared to the raytracing predictions, which closely match the tau-p predictions. This suggests that the discrepancy between the two approaches is related to the inherent non-uniqueness of both the V_p tomography and the PSDM approach and reinforces the need to validate geophysical results with multiple independent observations.

Coincident measurements of basalt elevation interpreted from the V_p results along Lines A North, B, and D suggest that the precision of the refraction method is on the order of ± 2 –4 m. This likely applies to regions where the basalt surface is laterally uniform. In areas where the basalt has rough topography, the refraction data will tend to smooth out the small-scale variations associated with sediment velocity transitions. A good example of this is the comparison of the V_p image and the check shot from well 699-61-62 (Figure 6a). The V_p tomogram shows a smooth gradient from sediment to basalt velocities, whereas 150 m away, the check shot shows a hard boundary. The re-

flection image shows rough topography on the basalt at this location. Since the refraction data can resolve only a smoothed version of the structure, the depth to basalt is poorly constrained.

Variations in Basalt Vp

The Vp tomograms show variations in basalt velocity that may indicate areas where the basalt is fractured and likely to be permeable. The highest basalt Vp ($\sim 5,000 \text{ m s}^{-1}$) on Line A North (Figure 5a) occurs where the PSDM image shows a strong coherent reflection, suggesting a smooth basalt surface (between $x = 100\text{--}300 \text{ m}$ and $1,200\text{--}2,000 \text{ m}$). On the edges of these sections, Vp is not as well constrained due to a lack of data coverage. However, basalt Vp is lower at transitions in reflection coherence. These areas may represent a rubbly basalt surface or a region of increased fracture density where communication between the unconfined and confined aquifers could potentially occur. However, well observations are needed to confirm this.

At profile distances greater than $3,000 \text{ m}$ for Line A North, Vp in the basalt is $\sim 3,500\text{--}4,500 \text{ m s}^{-1}$, and the reflection image indicates several discontinuities. Here, the low Vp along the top of the basalt is well constrained and may indicate fractured and permeable basalt. The combination of low basalt Vp and discontinuous structure in the PSDM image supports an interpretation of folded and/or fractured basalt.

Basalt Elevation Map

The $3,000\text{-m s}^{-1}$ Vp contour marks the transition from suprabasalt sediments to basalt and shows good agreement with well observations (Figures 5 and 6). There are a few sections of Line A North where the top of the basalt is too deep to be imaged with the 196-m streamer offset, and these are areas where paleo-channels carved through the basalt have been previously interpreted (Sunwall et al., 2011). Here, the PSDM images were used to interpret basalt elevation with the understanding that they may be biased toward overly deep estimates.

Combining the depth to Vp = $3,000 \text{ m s}^{-1}$ where it is well constrained with PSDM interpretations and well observations yields a two-dimensional visualization of the basalt surface in the Gable Gap area. Figure 8 shows a minimum curvature surface fit to the data masking everywhere in the study area that is more than 250 m away from an observation. It shows a complex basalt surface consistent with previous interpretations of paleo-channels and second-order syncline anticline pairs superimposed on the southern limb of the GMA.

Most of the variation in basalt elevation is supported by the seismic results; however, well observa-

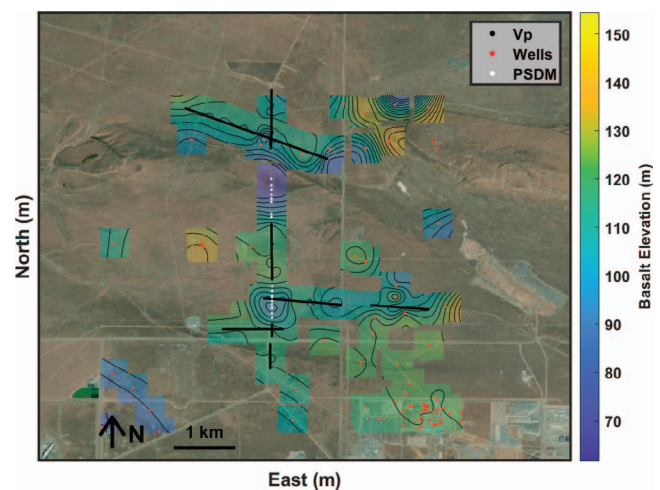


Figure 8. Minimum curvature surface fit to well data (red dots), interpreted depths to Vp = $3,000 \text{ m s}^{-1}$ (black lines), and basalt elevations interpreted from reflections on Line A North, where refractions do not constrain basalt elevation (white dots).

tions also show large variations over short distances. For example, two wells north of Line C East show elevation differences of 24 m over a distance of 190 m .

Variations in Shallow Vs and Vp

Both the Vp- and the Rayleigh wave-derived Vs images indicate lateral variations in shallow ($< 20 \text{ m}$ deep) sediment properties. These variations likely represent changes in grain size distribution or sandy versus gravelly deposits of the Hanford formation. For example, the dashed lines in Figures 5b and 6b indicate an interpreted boundary between the sandy H2 and the gravel-dominated H3 unit. Given the different grain size distribution in each unit, a difference in porosity and hydraulic conductivity is also expected, and variations in the distribution and thickness of these two units may impact vertical fluid infiltration and unsaturated flow within the vadose zone. Future studies with lower-frequency geophones could potentially image deeper Vs structure and provide the possibility of using Vp/Vs ratios to discriminate between dry and saturated conditions.

CONCLUSIONS

The seismic landstreamer is an effective tool for characterizing near-surface hydrostratigraphy at the Hanford Site. While previous work (Hyde et al., 2011; Sunwall et al., 2011) focused on the reflected wave field, this analysis demonstrates that the refraction and Rayleigh wave data can provide additional information on basalt elevations and properties.

The integration of reflection, refraction, Rayleigh wave data, and borehole information has revealed previously undetected features within the study area. The refraction data provided an updated estimate of basalt elevation and allowed the identification of low-velocity zones that coincide with discontinuities in the reflection images. These low-velocity zones may be indicative of fractures supporting vertical communication between the confined and unconfined aquifers. Integrating well observations with the seismic interpretations suggests that the refraction approach provided basalt elevation estimates that agree more closely with well observations compared to PSDM. Both methods support the interpretation of a highly variable basalt surface. Vs structure derived from Rayleigh wave analysis was broadly consistent with Vp structure in the upper 10–20 m, indicating variations in the distribution of Hanford formation units. This information, along with continuing geophysical investigations, will be used to refine the site geologic model.

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