Estimating subsurface topography from surface-to-borehole seismic studies at the Rye Patch geothermal reservoir, Nevada, USA

Roland Gritto*, Thomas M. Daley, Ernest L. Majer

Center for Computational Seismology, Earth Sciences Division, Lawrence Berkeley National Laboratory, 1 Cyclotron Road, MS90-1116, Berkeley, CA 94720, USA

Received 30 May 2002; accepted 7 February 2003

Abstract

A 3-D surface seismic reflection survey, covering an area of over 7.7 km², was conducted at the Rye Patch geothermal reservoir (Nevada, USA) to explore the structural features that may control geothermal production in the area. In addition to the surface sources and receivers, a high-temperature three-component seismometer was deployed in a borehole at a depth of 1250 m within the basement below the reservoir, which recorded the waves generated by all surface sources. The objective of this study was to determine the subsurface structure of the reservoir based on this surface-to-borehole dataset. A total of 1959 first-arrival travel times were determined out of 2134 possible traces. Two-dimensional ray tracing was performed to simulate wave propagation from the surface sources to the receiver at depth. The ray tracing was based on a 2-D laterally homogeneous velocity model derived from results of a vertical-seismic-profile (VSP) experiment recorded in the same well. The method is an approximation where it is assumed that differences in travel time between the observed and modeled data are caused by structural deviations from a homogeneously layered model as estimated by the VSP profile, and thus are mapped into topographic changes at depth. The results indicate, to first order, the presence of two dominant geologic features. The first observation is consistent with the regional trend of the geologic units in the Basin and Range province with a north-south strike and dip to the west, as expected for this area west of the Humboldt Thrust Range. The second is a local feature in the form of an east–west ridge. The geometry of the structure is corroborated by results from a seismic-reflection survey, and by results of a gravity survey conducted in the area above the reservoir.

© 2003 CNR. Published by Elsevier Science Ltd. All rights reserved.

Keywords: Geophysics; Seismics; Surface-to-borehole survey; Nevada, USA
1. Introduction

Geothermal reservoirs are considered difficult seismic targets because of hydrothermal alteration and structural heterogeneity (Blackwell, 1985; Sorey, 1985). In the 1960s and 1970s, seismic experiments were started to determine the subsurface structure of Long Valley Caldera, California, and to more tightly constrain the geometry of the caldera floor (Pakiser et al., 1960; Hill, 1976). The results showed several sequences of shallow and deep reflectors interrupted by faulting, although no conclusive evidence for the presence of a hypothesized magma chamber was reported. These early studies revealed the problems associated with the application of seismic methods to geothermal areas. In the past 30 years, technological advances in seismic exploration have increased the impact of seismic surveys on hydrocarbon prospecting. Although 2-D and 3-D seismic methods have proven to be an integral part of modern oil and gas exploration efforts, the heterogeneous nature of geothermal reservoirs makes all seismic imaging more difficult (Black et al., 1991; Hill et al., 1985). In past years, seismic exploration efforts have been increasingly used in geothermal areas, yet it is still unclear how well these methods can be transferred from the petroleum to the geothermal industry.

Seismic surface and borehole experiments were recently conducted at the Rye Patch geothermal reservoir, Nevada, to determine the geologic structure of the (hypothesized) fault-controlled reservoir. The Rye Patch geothermal reservoir is located in Pershing County, Nevada, along the east side of Interstate 80, about 200 km northeast of Reno (see location in Fig. 1). Commercial development of the Rye Patch geothermal project started in the late 1980s and resulted in the construction of a 12 MW power plant and eight geothermal wells, of which seven were either too cold or non-productive. In the successful well (44-28), however, significant production at reservoir temperatures in excess of 200 °C was encountered. The eight boreholes were drilled within an area of less than 2.6 km², which indicated that distribution of reservoir fluids is most likely controlled by fractures and faulting with limited areal extent. In 1997, The Industrial Corporation (TIC), as the owner of the project, and Transpacific Geothermal Inc. (TGI), cooperated with the Lawrence Berkeley National Laboratory (LBNL) to evaluate and apply modern seismic-imaging methods for geothermal-reservoir definition under the US Department of Energy’s (DOE) Geothermal Program. As part of this effort, a vertical seismic profile (VSP) was acquired in 1997 to determine the seismic reflectivity of the reservoir horizons and to obtain reservoir velocity information. Because the results of the initial VSP profile indicated apparent reflections at depth (Feighner et al., 1998), the participants in the project decided to proceed with a 3-D seismic-reflection survey, which was acquired in 1998. As part of the seismic surface survey, an additional surface-to-borehole experiment was conducted, during which a three-component high-temperature geophone was installed in the original VSP well at a depth of 1250 m. This geophone recorded all seismic waves generated by the surface sources, creating a second dataset in addition to the seismic-reflection data. The locations of the 3-D surface survey and of borehole 46-28 containing the geophone at depth are indicated in Fig. 2. They coincide with the Rye Patch temperature anomaly, which is bounded by the
Humboldt City Thrust in the east and the Rye Patch reservoir in the west. Results of the 3-D seismic survey were presented by Feighner et al. (1999) and revealed possible faulting at depth based on surface seismic-reflection studies and surface-to-surface tomographic-travel-time investigations. In the current study, we present the results of the surface-to-borehole dataset, which was recorded with minimal extra effort during the acquisition of the surface-reflection survey and shows that it can provide additional valuable information about the reservoir structure at depth, confirming the results of previous studies.

2. Data acquisition and processing

The Rye Patch geothermal survey covered an area of approximately 7.7 km² and was designed with 12 north–south receiver lines and 25 east–west source lines. The
Fig. 2. Location map of the Rye Patch geothermal anomaly with area of 3-D seismic survey. The location of VSP Well 46-28 is indicated by the arrow (modified after GeothermEx, 1997).
source interval was 30 m, whereas the source line spacing was 120 m. Four Litton 311 vibrators were used in a squared array, with the source flag at its center. The source signal was a sweep with frequency bandwidth between 8 and 60 Hz. A detailed description of the data collection can be found in the contractor’s report (SECO, 1998).

A high-temperature, wall-locking, three-component geophone was installed in well 46-28 at a depth of 1250 m. The borehole geophone recorded all shots throughout the survey area, amounting to a total of 2134 traces. The locations of all sources as well as the boreholes are shown in Fig. 3. The gaps in coverage are caused by Interstate 80 and railroad tracks, which cross the survey area in a north–south direction.

The data quality is good, with a central frequency content of about 25 Hz for the first arriving waves. Fig. 4 shows a representative receiver gather of a source line about 1200 m north of well 46-28. In addition to the normal moveout across the source line, local and smaller variations in arrival time can be seen between source positions 10048 and 10063. These local variations in travel time will be mapped into topographic changes of the reservoir horizons at depth.

A total of 1959 first-arrival travel times were determined out of 2134 possible traces. Most of the picks are reliable because the well-sampled spatial moveout across the source lines facilitated the picking. However, in addition to the long...
source lines, “make-up lines” with shorter distances and a maximum number of nine sources per line were set up in between the original lines. The first-arrival picking was less reliable for these shorter lines.

3. Methodology

3.1. Ray tracing

In 1997, a VSP was recorded at the Rye Patch geothermal field in well 46-28 (Feighner et al., 1998). The resulting P-wave velocity profile between the depth of 120 m and 1265 m represents the best estimate for the distribution of velocities in the subsurface around the well. Based on these results, we derived a velocity function that represents a smoothed average of the VSP velocity profile. The function and its geologic interpretation are shown in Fig. 5. The prominent features of this velocity function are the high-velocity layer of 3500 m/s between 210 and 240 m depth, followed by a velocity inversion to approximately 2750 m/s down to a depth of 700 m. This upper interval represents the Tertiary sequence of sedimentary and volcanic rocks. Below this sequence lie the carbonates of the Triassic basement rocks, indicated by a velocity increase to about 6100 m/s. The productive zone of the reservoir is confined to a clastic layer of 60 m thickness at a depth of about 880 m within the

![Well 46-28](image)

Fig. 4. Common receiver gather for a source line north of well 46-28. The location of the receiver well is projected onto the data gather for reference.
This thin layer, however, is not resolved in the velocity function shown in Fig. 5. This velocity profile is subsequently extended to a 2-D homogeneous velocity model throughout the survey area. Based on this velocity model, 2-D ray tracing is performed to simulate wave propagation from the surface sources to the receiver at depth. Fig. 6 shows representative results of the ray tracing, where the velocity model is a 2-D representation of the function in Fig. 5. Sources are denoted by stars, while the receiver is indicated by a triangle at 1250 m depth. Fig. 6a represents the rays for a source line that runs in an east–west direction across well 46-28, while Fig. 6b shows a line running across the well in a north–south direction. The gaps in source coverage indicate the railroad tracks, Interstate 80, and an area in the vicinity of the well where the vibroseis sources could not operate.

The 2-D ray-tracing produces a total of 2134 rays, connecting the sources to the receiver at depth, and their associated travel times. None of the 2134 rays crosses the paths of other rays, which prevents the application of a tomographic inversion approach. It is therefore not possible to estimate lateral velocity variations within

Fig. 5. Velocity profile from the VSP survey in well 46-28 with interpretation of geologic strata. The depth is measured from the surface.
the layers. However, under the assumption that the original geologic sequences were deposited in layers, and that subsequent disturbances of the stratigraphy were caused by faults, the topographic structure of the reservoir can be mapped by comparing observed to numerically calculated travel times. In the current example, the velocity structure estimated from the VSP experiment is used as a reference model, and travel times are calculated for all surface sources. The observed and calculated travel times are compared for each source-receiver combination, and differences attributed to changes in elevation of the subsurface horizons. This method and its limitations will be explained in the next section.

---

**Fig. 6.** Velocity model and ray paths from source lines transecting well 46-28: (a) N–S direction; (b) E-W direction. The top of the velocity model is chosen to be equal to the elevation of the highest source position of the survey, which causes the sources in the figure to appear below the surface, because the depicted sources are all located below the highest source.
Although most geothermal reservoirs exhibit localized geologic heterogeneity caused by areas of hydrothermal alteration as well as volcanic deposition (i.e., basaltic lenses), these areas are confined to relatively small volumes (in the case of lenses) or thin sheets (in the case of hydrothermal alteration along fluid pathways). While the identification and mapping of faults at Rye Patch reservoir have proved difficult, several geophysical surveys, including surface magnetic, gravity, self-potential, and seismic reflection in conjunction with geological observations, suggest the existence of at least one east–west striking fault (GeothermEx, 1997; Teplow, 1999). The depth extension of the fault has been estimated to reach from a minimum depth of 1500 m in the Triassic carbonate basement upwards into the shallow parts of the Tertiary sediments, without producing evidence at the surface. The velocity profile in Fig. 5 suggests that a throw across this interface would juxtapose units with velocity differences up to 100% (≈3000 m/s versus 6100 m/s for sediments and basement, respectively). It is reasonable to assume that this high velocity contrast has a stronger effect on the travel times of seismic waves than the localized velocity heterogeneities described above. It is therefore assumed, to first order, that the large-scale structural features caused by faulting across the basement interface can be mapped by comparing the observed travel times of the current dataset with travel times computed for a horizontally-layered reference model. It should however be pointed out that this method will resolve large-scale features only, as smaller-scale features may be caused in part by local velocity heterogeneities.

### 3.2. Theory

Mapping travel-time deviations to elevation changes is a technique that has been used in seismic refraction studies in the past, and is also referred to as “seismic detailing” (Dix, 1952), “time-term method”, or “delay-time method” (Telford et al., 1990; Nettleton, 1940). The method is an approximation that can be applied to environments where a low-velocity layer is located above a high-velocity layer. Under the assumption that the ray path from source to receiver is known, any difference between the calculated and observed travel times is converted into a distance using the velocity model and applied as an elevation deviation in the boundary between the two layers. The same principle is applied in the current approach assuming that the top layer is represented by the 700 m thick sedimentary and volcanic Tertiary sequence, which can be approximated by an average velocity of 2750 m/s; whereas the Triassic carbonates of the basement are represented by a halfspace with a velocity of 6100 m/s (refer to Fig. 5).

Fig. 7 shows a schematic model of a low-velocity layer overlying a high velocity basement \(v_1 < v_2\). A geophone is positioned in a borehole at a total depth \(z = h_1 + h_2\), while a source is located at the surface at a distance \(x\) from the wellhead. The total travel time \(t_m\) from source to receiver can be expressed by ray theory as

\[
t_m = \frac{x}{c} + \frac{h_1 \cos \alpha_1}{v_1} + \frac{h_2 \cos \alpha_2}{v_2}
\]
where $\frac{1}{c}$ is the ray-parameter, which is constant along the seismic ray from source to receiver and

$$\frac{\sin \alpha_1}{v_1} = \frac{\sin \alpha_2}{v_2} = \frac{1}{c}$$  \hspace{1cm} (2)

If the boundary between the layer and the basement is perturbed by a difference in elevation of $\Delta h$ (refer to Fig. 7), the observed travel time becomes

$$t_o = \frac{x}{c} + \frac{(h_1 + \Delta h)\cos \alpha_1}{v_1} + \frac{(h_2 - \Delta h)\cos \alpha_2}{v_2}$$  \hspace{1cm} (3)

where Fermat’s principle has been invoked by assuming that $\alpha_1$ and $\alpha_2$ do not change. Thus, the difference in travel time between the unperturbed and perturbed case is

$$\delta t = t_m - t_o = \Delta h \left( \frac{\cos \alpha_2}{v_2} - \frac{\cos \alpha_1}{v_1} \right)$$  \hspace{1cm} (4)

and the elevation difference becomes
\[ \Delta h = \frac{\delta t}{\left( \frac{\cos \alpha_2}{v_2} - \frac{\cos \alpha_1}{v_1} \right)} \]  

(5)

Since Fermat’s principle states that the arrival time is stationary with respect to the perturbation of the ray path around the original one, it can be calculated along the unperturbed ray instead of the perturbed one (Aki and Richards, 1980). This principle is the basis for the current mapping approach, where the elevation difference \( \Delta h \) is computed along the unperturbed ray in the layered velocity model.

Once the elevation changes are computed for the two-layered model, they need to be migrated along the rays to be mapped at the locations where the rays cross the boundary between the sedimentary layer and the basement.

3.3. Seismic mapping and source elevation statics

Eq. (5) is used to map travel-time changes to changes in elevation at depth. Positive deviations denote source positions from which the actual waves travel faster to the receiver than in the ray tracing simulations. The interpretation is that the high-velocity basement is uplifted relative to the homogeneously layered velocity model used in the simulations. Similarly, negative deviations denote slower wave propagation than assumed in the simulations, indicating a thicker low-velocity layer on top of the basement (i.e., the basement is shifted downwards, refer to Fig. 7). The result of mapping the travel-time change \( \delta t \) to changes in topography of the interface between the carbonates and sediments at 700 m depth is given in Fig. 8a. The surface plot is generated using 1959 data points (representing the number of first-arrival times determined from the data) and smoothed over an area of 300 m by 300 m. The smoothing reduces the effect of local velocity heterogeneities that may be present in the reservoir but are not considered in the current approach. The results should therefore be interpreted qualitatively rather than quantitatively.

To first order, the interface reveals higher values in the east, gradually decreasing towards the western boundary of the survey area. This result mimics the dip in elevation of the surface sources throughout the survey area, as shown by the 3-D map in Fig. 3. It is therefore necessary to test whether the results are an artifact introduced by the surface topography of the survey area. A travel-time error can occur by using correct source locations with large elevation changes while applying a constant low-velocity model for the near surface layer because geologic processes often compensate for the travel time variation of this model. Source sites at higher elevations are usually exposed to stronger erosion, which removes low-velocity sedimentary layers, such that hard rock with higher velocities may be exposed. This effect compensates for the longer travel distance from the elevated source location to the receiver at depth. If, during the simulations, sources are placed at the correct elevations in conjunction with the use of a low-velocity surface layer, the travel times of the simulations may become too long relative to the observed travel times and the resulting positive travel-time deviations would be mapped into positive elevation changes at depth. The reverse effect may take place for lower elevations, where
thicker sedimentary fill can lower the values of the velocity below those assumed in the model.

Thus, a second simulation is performed to verify that the trend of the interface in Fig. 8a is not an artifact caused by the elevations of the source locations. During this test all sources are located on a plane equal to the elevation at the wellhead of borehole 46-28.
borehole 46-28. If the structure in Fig. 8a were caused by static problems with the source locations, it would disappear or change after the simulations with a flat source horizon. However, the resulting structure is almost the same, as shown in Fig. 8b. Although the magnitude of the elevation changes decreases slightly, relative to the results in Fig. 8a, the general features in both figures are still the same. Therefore, it is assumed that static time shifts associated with local inhomogeneities in the vicinity of the surface sources represent a secondary effect that can be neglected for the purpose of this study.

The overall trend of the interface in Fig. 8a seems to indicate a gradual dip to the west, which may reflect the regional tectonics of the Basin and Range province. The dominant structural mechanism in the Basin and Range is comprised of north–south trending faults dipping towards the west, which generate westward dipping units. In addition to the large-scale trend of the interface, Fig. 8a reveals a local-scale topography change in the central region of the survey area to the north of well 44-28. This region is dominated by an increase in elevation that is gradually decreasing towards north and south. It seems that wells 42-28 and 72-28 are located on the southern flank of this topography anomaly, which extends throughout the survey area from east to west. Thus, it appears that the analysis revealed two main features within the Rye Patch reservoir. The first is the expected dip of the geologic units towards the west associated with normal faulting on a regional scale, while the second is in the form of a broad east–west ridge possibly bounded by faulting to the north and south. The possibility of faulting associated with this anomaly will be investigated later on.

3.4. Error analysis

The surface maps in Fig. 8 are based on smoothed topography estimates, which average the underlying values that were computed based on Eq. (5). The maximum estimated topography value for the example in Fig. 8a was 498 m. This shift in deviation may appear high for the current method, particularly for large offsets, as it is stated that $\alpha_1$ and $\alpha_2$ do not change in Eq. (3). To determine whether the assumptions of the current method are violated, a numerical ray tracing experiment is performed.

A representative source line is chosen that covers the lateral extent of the farthest source positions in the field experiment, with the idea that the associated ray paths will produce the largest differences between incidence and refraction angles (see Fig. 7). The reference velocity model is the same as above: a 700 m thick low-velocity layer (2750 m/s) over a high-velocity basement (6100 m/s). The second velocity model consists of the same high-velocity basement but its interface is now uplifted by 500 m to a depth of 200 m, representing the extreme elevation changes encountered in the current results. Ray tracing is performed for both models and the resulting travel-time differences mapped into elevation changes. The result is presented in Fig. 9, where the reference model is shown with the interface at 700 m depth. The interface of the second model is indicated by the gray line at 200 m depth, while the ray paths for both models are superimposed on the figure. The
white line represents the estimated shift in elevation based on Eq. (5). It can be seen that the match is quite good considering the large offsets and differences in ray paths. The comparison between the estimated and modeled shift in elevation yields a standard deviation of 3.5% (17.7 m), well within the range of accuracy intended for the current analysis. The mapping approach did not therefore introduce errors large enough to be of concern for the final results.

The actual elevation changes of the basement horizon are likely to be smaller than the ones shown in Fig. 8, since all deviations from the assumed horizontally layered velocity model are mapped into elevation changes. Furthermore, the velocity model may not be a good representation at great distances from the borehole, and it is feasible that a deviation in travel time is caused by local velocity heterogeneities within a layer rather than by a shift in the boundary of the layered velocity model. However, it is not possible to estimate those local velocity changes, because this would require a tomographic method, for which the ray coverage in the current dataset is not sufficient. The error associated with the mapping procedure that was introduced by using a homogeneously layered velocity model cannot be accurately computed. Smoothing of the initial estimates has however reduced the smaller-scale effects of localized velocity heterogeneities, with the intention to increase the reliability of the larger-scale lengths in Fig. 8a. The estimated changes in elevation should therefore be considered as upper bounds for the actual values, while the uncertainty may be on the order of the intermediate-scale features in Fig. 8a, such as the ridge with an elevation of 100–150 m. With these considerations in mind, the structure of the interface will be investigated more closely.

Fig. 9. Velocity model and ray paths used to test the accuracy of the mapping approach for the survey geometry. The two sets of ray paths are computed for models with a basement interface at 700 m and 200 m (indicated by the gray line). The white line represents the estimated interface as determined with the current mapping method.
4. Interpretation and comparison with previous studies

A map view of the basement horizon elevation is provided in Fig. 10a. It can be seen that the 0 m elevation contour line runs through well 46-28, which is a confirmation that the smooth version of the velocity model shown in Fig. 5 is a good representation of the actual velocities in the vicinity of well 46-28. The map shows the contours of the broad ridge structure extending from east to west, following the general dip of the basement to the west.

In addition to the source locations shown in Fig. 3, four far-offset source locations were selected several kilometers outside the receiver array during the 3-D seismic survey in 1998, to obtain far-offset first-arrival data refracted along the basement interface that could be used to determine the deeper velocity structure. The far-offset shots were recorded by 10 receiver lines in the center of the survey. The qualitatively best datasets resulted from shot number 2, located 7.4 km NW of the VSP well 46-28, and shot number 4, located 5.2 km SSE of the well.

Fig. 10b shows the data for far-offset shot number 2 recorded by a receiver line in the western half of the survey area. The northern receivers recorded sharp first arrivals, but the signal is abruptly attenuated for receivers located in the central and southern part of the survey area. This pattern was consistent for all other receiver lines, allowing us to spatially map the boundary of observable arrivals. The two gray areas in Fig. 10a represent the receiver locations where first-arrival energy was clearly visible. The northern gray areas represent the arrivals of the data recorded from shot number 2 to the northwest, while the southern gray areas represent those of shot number 4 to the south. It is evident from the figure that the central area had weak or non-existing first arrival energy. A possible interpretation is the existence of near-vertical faults that scatter and attenuate seismic energy during its propagation across these features. Since the boundaries of the polygons match the outline of the elevated structure quite well, we propose that faults, bounding the ridge to the north and south, attenuated the seismic waves from the far-offset shots.

The location of a possible fault was interpreted by Teplow (1999), on the basis of 3-D seismic reflection data. The intersection of the fault with the clastic reservoir unit had a strike of N 76° W and a dip of 73° NNE. The intersection of this fault with the sedimentary-carbonate interface is indicated by the dark-gray line in Fig. 10a immediately south of well 44-28. In projecting the fault upward onto this interface, a constant strike and dip are assumed. The interpretation of the east–west extension of the fault was limited because of the poor continuity of reflected seismic energy in the east–west direction (Teplow, 1999). It can be seen that the strike-line of the fault coincides with the boundary of the southern zone that marks the transition from strong to weak first-arrival energy, and is located along the southern flank of the elevated structure indicated by the contour lines. Thus, a possible interpretation is that the ridge is a manifestation of the postulated fault.

The study by Teplow (1999) included a gravity survey of the Rye Patch geothermal field. The survey consisted of 334 stations along 19.8 km of mainly north–south trending profile lines (with one line traversing east–west) and was confined to the central region of the seismic survey. The Bouguer residual was computed by...
Fig. 10. (a) Contour map of the variations in elevation of the basement interface. Contour lines are in meters relative to the basement interface. The gray areas represent receiver locations where good first-arrival energy was recorded from far-offset shots during the 3-D seismic survey. The northern area recorded good first-arrival energy from shot number 2, located 7.4 km NW of the VSP well 46-28, while the southern area recorded good first-arrival energy from shot number 4, located 5.2 km SSE of the VSP well (see arrows pointing in direction of source locations). No strong first arrivals were recorded in the central region of the survey. The bold gray line represents the projection of a fault onto the basement interface that was interpreted by Teplow (1999) from 3-D seismic reflection data. (b) Seismic data recorded along a representative receiver line in the western part of the survey area. The source position was at shot number 2. Notice the abrupt change in amplitudes of the first arrivals, as indicated by the arrow.
calculating a best fitting plane to the local gravity data and subsequently subtracting it from the data (Teplow, 1999). The residual can be interpreted as a map of mass in excess of a uniform sloping plane that represents the top of the Triassic basement. Fig. 11 shows a surface plot of the Bouguer gravity residual as determined by Teplow (1999). Although the area of the gravity survey is smaller than that of the seismic survey, the features visible in the surface plot in Fig. 8 can also be observed in Fig. 11. On a large scale, the surface in Fig. 11 slopes towards west and south, while the area north of boreholes 42-28 and 72-28 shows an increase in residual gravity, similar to the ridge structure in Fig. 8. The area of the gravity survey, however, does not extend far enough north to determine the width of this elevated structure. Nevertheless, the structures in Figs. 8 and 10a reveal similar shapes on a large scale. The interpretation of the Bouguer residual map could include a densification of the reservoir or basement rocks resulting from hydrothermal mineralization, or an uplift of high-density basement rocks relative to the overlying lower density sediments.

The combination of seismic and gravity data could possibly suggest the presence of an elevated basement structure, while the process of localized densification may not be applicable to explain the seismic data. Although hydrothermal mineralization can increase seismic velocities relative to the surrounding host rock, it usually occurs along leaks from the production zone of the reservoir, preferentially along faults or other weak structures, producing one- or two-dimensional alterations. However, the volume of faster material needed to match the observed seismic travel time differences (see Fig. 8) is considerably bigger than that produced by hydrothermal alterations.
A feature similar to the elevated structure described above was reported by Feighner et al. (1999), and is shown in Fig. 12. The figure shows tomographic velocity estimates from two receiver lines along the eastern boundary of the survey area. Although the depth penetration for the tomographic study is limited as the turning rays propagate from surface sources to surface receivers, the ray coverage is good down to 500 m depth. The two vertical lines in Fig. 12 indicate the location of the 200 m contour line representing the center of the ridge structure at the eastern boundary of the survey in Fig. 10a. The tomographic estimates reveal an elevated structure of faster material in the center of the survey area. It should be noted that the elevated velocity contours at the margins of the images in Fig. 12 are an artifact of the ray geometry and do not represent actual subsurface structure. Both depth

Fig. 12. Velocity estimates of tomographic travel-time inversions for two receiver lines at the eastern boundary of the seismic survey area: (a) N–S receiver line, located above the maximum elevation of the basement interface in Fig. 10a; (b) N–S receiver line, located directly east of the maximum elevation of basement interface in Fig. 10a.
sections, however, indicate a broad range of elevated high-velocity material in the central and south-central section of the survey area. Furthermore, it appears that the high-velocity structure extends upward without breaking the surface, because the velocity contours seem to flatten out in the upper 50 m. On a large scale these results are in agreement with the seismic mapping results presented in Figs. 8 and 10a.

5. Conclusions

The 3-D seismic experiment conducted at Rye Patch geothermal field provided a series of datasets and methods to image and interpret the subsurface structure of the reservoir (surface reflection seismic, surface-to-surface seismic tomography, and surface-to-borehole seismic mapping). The addition of a borehole geophone at depth to record surface-generated seismic waves during the 3-D reflection survey provided an independent dataset at low cost and minimum technical and labor requirements.

Because most geothermal areas provide access to open boreholes during the developing stages of the reservoir, it is recommended that a VSP survey is conducted first, to obtain information about the velocity structure and the reflectivity of the subsurface. VSP results are generally extrapolated from the vicinity of the borehole into the surrounding area to provide a 2-D velocity model. Because of the heterogeneous nature of geothermal reservoirs, however, the error in extrapolating the VSP information can be minimized by conducting VSP surveys in multiple boreholes throughout the reservoir. The current study would have benefited from additional VSP data, which could have generated a more realistic velocity model. If it is determined for future geothermal exploration projects that a surface seismic-reflection survey may provide more detailed information about the reservoir structure, we recommend adding geophones to any available borehole within the survey area. These datasets collected at depth provide an independent, low-cost addition to the surface data and can help in the interpretation of the subsurface structure.

Seismic mapping is a robust method to convert travel-time differences to elevation changes, but it is an approximation that relies on a predefined reference velocity model. In the current study, the estimated elevation changes represent upper bounds of the actual changes, because the 2-D reference velocity model does not account for localized velocity heterogeneities. However, it was assumed that localized velocity heterogeneities have a smaller effect on seismic waves than the shift of the basement interface, where geologic units with a velocity contrast of about 100% are juxtaposed. Furthermore, large-scale smoothing of the seismic results should have reduced local effects of velocity heterogeneity in the data. The results confirm the regional structure of the Basin and Range province. The regional trend of the geologic units reveal a north–south strike and dip to the west, as expected for normal faulting encountered in the extensional regime on the western side of the Humboldt Thrust Range. Furthermore, a local disturbance of the regional trend is detected by an elevation of the interface between the carbonate basement and the overlying
sedimentary sequence. The structure, which resembles a ridge, strikes east–west and appears to be extending throughout the survey area. Previous studies corroborate the findings of the current work, because the boundaries of the elevated structure coincide with areas in which the first arrivals of seismic waves undergo a transition from strong to weak amplitudes (Feighner et al., 1999). One possible explanation could be faults bounding the ridge to the north and south. Such a fault is reported by Teplow (1999), on the basis of 3-D seismic reflection data, and is located along the southern flank of the elevated structure. Furthermore, gravity data reported in the same study indicate a residual gravity high that coincides broadly with the areal extent of the ridge in the central section of the Rye Patch reservoir. In addition, tomography results (Feighner et al., 1999) indicate an elevated high-velocity structure along the eastern border of the survey area. The synthesis of these results suggests the presence of a local structure resembling an east–west striking ridge, i.e., an uplift of the interface between the basement and overlying sediments. The actual elevation changes of the interface cannot be reliably estimated, because the calculated values are upper bounds that were subsequently smoothed on a large scale. Thus the presented results should be regarded qualitatively rather than quantitatively.

The basement elevation changes could be projected onto the clastic reservoir at 880 m depth if strike, throw, and dip of the present faults are known. Because this information cannot be extracted from the current data, we did not attempt to map the reservoir. However, the known faults at Rye Patch reveal a steep dip angle and a vertical extension that exceeds the elevation difference between the basement interface (700 m) and the reservoir (880 m). It can therefore be assumed that the trend of the interface mimics the structure of the reservoir below.

One way to assess the validity of the presented model of the Rye Patch structure could be to incorporate and test it with reservoir simulations. In general, we expect that the actual subsurface structure is similar to a combination of results derived from geophysical studies at Rye Patch reservoir, and that these surveys could thus be used as an example to explore other geothermal reservoirs.

Acknowledgements

The authors would like to acknowledge reviews by Lane Johnson, Valeri Korneev, and Daniel Hawkes. Thanks are also extended to Peter Malin and an anonymous reviewer for their insights and critical review, which enhanced the content and style of the paper. This work was supported by the Assistant Secretary for Energy Efficiency and Renewable Energy, Office of Geothermal Technologies, of the US Department of Energy under Contract No. DE-AC03-76SF00098. Data processing was performed at the Center for Computational Seismology, which is supported by the Director, Office of Science, Office of Basic Energy Sciences, Division of Engineering and Geosciences, of the US Department of Energy under Contract No. DE-AC03-76SF00098. The authors would also like to thank Presco Energy LLC for their permission to use the data compiled in Teplow (1999). Most graphics were produced with the help of Generic Mapping Tools (Wessel and Smith, 1998).
References