SEISMIC EXPLORATION IN GEOTHERMAL AREAS – EFFECT OF THE SURFACE LAYER

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SUMMARY – In an effort to improve the signal to noise ratio of seismic reflection data collected in the Central Volcanic Region, experimental seismic trials have been carried out near the Mokai geothermal field. The nature of the complex seismic wave propagation in the near surface layers was examined and the noise field characterised. We determined the elastic parameters of the near surface layer as well as variation of the seismic data quality with acquisition parameters, in particular, shot and receiver depth. High seismic attenuation is evident in the top 6 to 9m of the surface layer, as is a large P velocity contrast which is likely to be associated with the degree of water saturation. This velocity structure results in reverberation of P-wave refracted arrivals and substantial P-to S wave conversion. Changes in the shear wave velocity, probably related to changes in porosity, form what is in effect a very efficient waveguide for surface waves.

INTRODUCTION
It is notoriously difficult to collect good quality seismic data in the Central Volcanic Region (CVR), as is also found in other volcanic regions (e.g. Young and Lucas, 1988). Multichannel seismic reflection surveys have been carried out in the CVR since the 1950’s, both within active geothermal fields and on the margin of the fields, with varying success, depending on the exact location of the survey and the acquisition parameters that were used (e.g. Henrys, 1987, O’Connor, 1990). The recorded seismic data is usually characterised by a lack of identifiable reflections, the presence of strong coherent surface wave noise, and a narrow low frequency bandwidth. As a result the processing of the data becomes quite difficult, the final data quality affecting the degree to which the seismic reflection technique can be used to guide geothermal reservoir development.

Various seismic experiments have been carried out in the CVR to understand the nature of the noise problem (e.g. Henrys, 1987). Henrys highlighted the effect of the near-surface (top 10m) unsaturated layer on reflection seismic data collected in the Broadlands geothermal field. Institute of Geological and Nuclear Sciences Ltd (formerly DSIR Geology and Geophysics) has been carrying out experimental studies near the Mokai geothermal field to further understand these complex seismic wave propagation effects, to characterise the noise and to improve the signal to noise ratio of data collected in seismic reflection surveys. In the present study we have examined both the elastic parameters of the near surface layers and the variation of the seismic data with acquisition parameters.

Elastic parameters of the near-surface layers near Mokai

A detailed P and S-wave seismic reflection and refraction experiment was carried out on Forest Rd, east of the Mokai geothermal field. Twenty four 3-component geophones were placed with 2m and 10m spacing in two separate seismic surveys. Two seismic sources were used for each survey: small charges of explosive in 1m deep shot-holes (to preferentially generate compressional waves) and a shear-wave hammer source (described by Stephenson & Barker, 1989) to preferentially generate SH-waves. Refracted P- and S-waves are clear on the records (Fig 1), although the first arrival of the shear wave on channels close to the source is sometimes obscured by air-waves. First breaks of both P- and S-waves were determined and then used as input in an inversion for velocity structure using an iterative least-squares routine written by Woodward (1991). The layered structure shown in figure 2a was found to be sufficient for fit of the P-wave refracted arrivals. A shear-wave velocity model was also derived (figure 2b), although inversion using the shear-wave arrivals was less stable due to the quality of the arrivals.

The large contrast between the near-surface P-wave velocity and the velocity (in what, comparatively, is a half-space) below is striking. It is likely that this velocity contrast is linked to water saturation levels, an interpretation supported by direct measurements of the water table at the survey site. There are several important consequences of such a large impedance contrast, as noted by Henry (1987), * the critical angle for reflection is reached at a relatively small angle, and, from calculation of elastic reflection coefficients, P-to S wave converted waves can be expected to contain a large fraction of the transmitted and reflected energy for incident angles between 15° and 80°, the exact effect depending on the sharpness of the boundary and the porosity of the layers. This effect alone is sufficient to explain the lack of coherent reflections. P-wave transmission and reflection become optimal only at angles near normal incidence. Such wave conversion appears to be quite common in volcanic...
Fig. 1. Examples of seismic data recorded on Forest Rd, near Mokai geothermal field from (a) shear-wave hammer source, recorded on the transverse components of 3-component geophones, and b) explosive source, recorded on vertical components. Refracted S and P-wave arrivals are arrowed. The horizontal distance between each channel is 10m.

Fig. 2. (a) P-wave and (b) shear-wave velocity structure determined for the near surface at the experimental site. Velocities are shown in km/s. Peg numbers correspond to distance in metres.

Fig. 3. Records from surface geophones for shot depths between 3.7m and 18m (a) unprocessed (b) with lateral trace-balance applied. Each row has been scaled relative to the 18m deep shot. Low frequency surface Rayleigh waves are arrowed.
environments - vertical seismic profiling (VSP) experiments by Iversen et al. (1990) in volcanic rocks in Washington State showed that the strongest recorded downgoing energy represented P-to-S conversion from the near surface, these converted waves often creating a very complex wavefield.

The discrepancy between the velocity structure models (shown in figure 2) derived from the compressional and shear-wave arrivals is of interest. Although the P wave velocities are likely to be related to the level of water saturation, shear wave velocities are not remarkably influenced by the percentage of saturation, but are however affected by porosity. One possible interpretation of the shear-wave velocity structure would be a decrease in the porosity with depth.

In a second experiment, carried out in the same location as the refraction experiment outlined above, 24 geophones were placed in 24 separate holes, each 18m deep, with 20m spacing between holes. Strings of geophones (6 per string) were also placed on the surface in small arrays centred over the geophone holes. Explosive shots were then detonated at several locations, and at varying depths. Figure 3 shows the record from the 24 surface geophone groups (which were spaced 20m apart), for shots located at depths of 3.7m, 6m, 9m, 12m, and 18m and offset 20m from the nearest geophone group. All the records are plotted with amplitudes relative to the 18m deep shot, and shown with no processing, and also with lateral trace balancing applied. From the raw records (figure 3a) it appears that the attenuation of the energy from the shots at 3.7m and 6m was high relative to the shot at 18m - little to no energy can be seen on channels offset more than 120m from the 3.7m shot. Following trace balancing (figure 3b) low frequency (10-20 Hz) surface (Rayleigh) waves can be seen (arrowed) across the records of the shallow shots. In addition to the Rayleigh waves large amplitude reverberations of the first P-wave refracted arrival dominate the top 500 ms of the record. These reverberations are very apparent on the record from the 3.7m deep shot, and are still visible on the records from the 6m deep shot (figure 3b).

Spectral analysis of the data, shown in figure 4, was carried out using a window between 500 and 1000 ms on one of the far-offset channels, 440m offset from the shots. This particular window was chosen to avoid possible contamination from the visible Rayleigh wave and the direct refracted waves. Although no allowance has been made for the geophone response some qualitative observations can still be made. A large difference can be seen between the recorded amplitudes from the 3.7m/6m deep shots and that from the deeper shots - the maximum amplitude recorded from the 3.7m deep shot is less than 1/10th that recorded from the 18m deep shot, indicating high attenuation in the top 6-9m. Energy in the 30 to 50 Hz frequency band is thought to be associated with weak reflection events at 500-600 ms two-way-time, within the analysis window. Figure 4 shows that this particular frequency band suffers attenuation as the shot depth decreases from 12m to 9m depth. Although the analysis window was carefully chosen it is difficult to judge whether the signal to noise ratio is higher for deeper shots - there were no reflection events coherent enough for quantifying the signal to noise ratio.

Spectra of the records from the downhole geophones (figure 5) indicate similar attenuation to that recorded by the surface geophones. Although it is not apparent from figure 5 (due to the relative scaling) the amplitudes recorded on the downhole geophones are an order greater than that recorded on individual surface geophones, even before allowing for the effect of the free surface, further emphasising that there is high attenuation in the top surface layer.

The Rayleigh waves appear to be even more prominent on the records from the 24 downhole geophones (figure 6, mowed). This likely is because the surface geophone groups consisted of strings of 6 geophones which, spaced out, acted to suppress the surface wave, while no such suppression is possible with individual downhole geophones. That the downhole geophones still record the surface wave, even from shots at 12 and 18m depth, is at first surprising until the wavelength of the Rayleigh wave, 15 to 30m, is taken into account. Synthetic modelling was carried out to compare the expected elastic wavefield with these observed records. The input model involved horizontal plane layers with elastic parameters approximating that determined in the field study - a 9m thick layer overlying a half-space. The expected seismic response at 18m depth from point sources at 6m, 12m and 22m was calculated using routines written by R. Benites. The resulting wavefield, shown in figure 7, was converted from the frequency domain using a Ricker wavelet with a peak frequency of 15 Hz and band-limited to 40 Hz. It is noted that the synthetic modelling involved point sources rather than a true explosive source; one might expect less (converted) shear wave energy from a true explosive source. In addition the model used here involves plane-layers and so does not reconstruct the scattering effects that result from an irregularly layered system. It does however show a calculated response that closely resembles the observed data; the synthetic records show that the calculated Rayleigh modes from a shot at 12m depth appear to be still strong, but are not so apparent from a shot at 22m depth although higher frequency components are still visible. These higher frequency modes are however very weak on the observed data.

CONCLUSION

Both the observed and synthetic data highlight the effect the near-surface layer has on the seismic wavefield - the layer creates a very effective waveguide for elastic energy. Interestingly, the elastic parameters found for the surface layer near the Mokai geothermal field vary only slightly from that found by Henrys (1987) in the Broadlands geothermal field - it is likely that the wave propagation effects will be very similar across the whole CVR. In addition to the effects created by the large P- and S-wave contrast the records from downhole shots and downhole receivers indicate large attenuation in the top 6 to 9m of the surface layer, probably corresponding to the degree of water saturation.

Work characterising the various noise trains is valuable; effective acquisition parameters can only be designed once the complex effects resulting from the near-surface layer are understood. The work described here, and that of Henrys (1987), suggests for example that the shot depths in previous seismic reflection surveys have usually been too shallow. It also indicates that receivers are unlikely to be effective unless they are placed at depths well below 25m. An alternative approach is to involve geophone arrays on
Fig. 4. Spectra for records from surface geophones, for shot depths between 3.7m and 18m, with scaling relative to the spectrum from the 18m deep shot.

Fig. 5. Spectra for records from downhole geophones, for shot depths between 3.7m and 18m, with scaling relative to the spectrum from the 18m deep shot.

Fig. 6. Records from downhole geophones for shot depths between 3.7m and 18m (a) unprocessed (b) with lateral trace balance applied. Each row has been scaled relative to the 18m deep shot. The Rayleigh waves are mowed.
the surface, with the arrays designed around the worst wavelengths of surface wave. Knowledge of the elastic parameters and structure of the near-surface will, in time, lead to effective acquisition and deconvolution techniques which will act to suppress the worst reverberation.

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REFERENCES


Fig. 7. Synthetic records calculated using shot depths of 6m, 12m, and 22m for downhole geophones placed at 18m depth. The records are scaled relative to the record from the 6m deep shot. The trace spacing is 10m.