THREE-DIMENSIONAL P-VELOCITY STRUCTURE OF THE SUMMIT CALDERA OF NEWBERRY VOLCANO, OREGON

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ABSTRACT

A three-dimensional high-resolution seismic study of the summit caldera of Newberry Volcano, Oregon, was conducted by the U.S. Geological Survey using an adaptation of the method applied by Mercresian et al., (1984). Preliminary interpretation of the traveltimes residuals reveals a ring of high P-velocity material coinciding with the inner ring fault system of the caldera in the upper 2 km. A zone of lower P velocities extends deeper than 2 km in the center of the caldera.

INTRODUCTION

Newberry Volcano is a young shield volcano located in central Oregon about 50 km east of the axis of the Cascade volcanic chain (Figure 1). MacLeod and Sammel (1982) describe the geology of Newberry Volcano. The flanks of the shield are veneered with basalt and basaltic-andesite cinder cones and flows, the youngest having 0-14 ages of 5800 years. A 6-km diameter caldera is present at the summit and has been the locus of recurring rhyolitic volcanism during the last 6000 years. The youngest rhyolite is 1350 years old and vented near the southern edge of the caldera. The continuing rhyolitic volcanism in the summit caldera has made it an attractive target for geothermal exploration and suggests that silicic magma may be present in the shallow crust beneath the caldera.

Two geothermal wells drilled in the caldera encountered high temperatures at shallow depths. A 930-m well drilled by the U.S. Geological Survey (USGS) had a bottom hole temperature of 265°C (Sammel, 1981). A 424-m well drilled nearer the caldera ring faults by Sandia National Laboratories encountered a similar but slightly hotter temperature profile. This pattern led Black et al. (1984) to suggest that thermal waters were flowing away from the caldera ring fault system. The locations of magmatic intrusions, and the likely heat source for the thermal water, are poorly understood. They may be near the ring faults, or deeper, with the ring faults serving as conduits for the water.

Figure 1. Location of Newberry Volcano and major Cascade Range volcanoes: volcanoes (open triangles), previous seismic refraction lines (dashed lines), and shotpoints used in the current study (stars).

The USGS has conducted several seismic studies of Newberry Volcano to elucidate its structure and to locate magma chambers. Stauber et al., (1985) did a teleseismic P-residual study to examine the crust beneath the volcano to a depth of 40 km with a spatial resolution of 5-10 km. Figure 2 is a northwest-southeast cross section of the P-velocity model obtained. The main feature is the zone of high P-velocities beneath the caldera extending from within 10 km of the surface to 25 km depth. Stauber et al. (1985) show that these high velocities severely limit the fraction of magma present in this volume to less than a few percent. They interpret the high P-velocities to be caused by numerous subvolcanic mafic intrusions, lodged in the crust as the volcano grew.

Next, to examine the shallow structure of the volcano, the USGS conducted an east-west seismic refraction experiment across the volcano. A preliminary interpretation of the seismic refraction profile (Stauber and Berge, 1984) is shown in Figure 3. This interpretation

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Figure 2. P-velocity cross-section through Newberry Volcano from teleseismic P-residual study of Stauber et al. (1985). P-velocity in km/s.

was used to design the three-dimensional experiment described below. The iso-velocity contours in the upper 6 km rise beneath Newberry Volcano so that the velocity is about 10% higher beneath the volcano than in adjacent areas at the same elevation. The high P-velocity zone detected in the teleseismic experiment continues downward from this shallow high-velocity feature.

EXPERIMENT DESCRIPTION

To examine the three-dimensional structure of the upper 5 km beneath the caldera, we used a technique similar to that applied at the Mont Dore volcano, France, by Mercier et al. (1984). In this technique, a two-dimensional array of seismic recorders is deployed over the target volume. Explosions are detonated in several directions at a distance selected to give strong reflected or refracted phases traveling upward through the target volume toward the array (Figure 3). By illuminating the target volume in this manner from many different directions and measuring the travel-times, the three-dimensional velocity structure of the volume can be found.

We deployed 120 portable seismographs in an array 12 km in diameter centered on the caldera of Newberry Volcano (Figures 3 and 5). The average station spacing was about 1 km and we expect a spatial resolution of about 1 km³. We placed shotpoints at two distances from the array in as uniform an azimuthal distribution as possible (Figure 1). Raypaths for the Pg phase from the closer shots, 30 to 40 km from the array, penetrate to a depth of about 6 km (west shotpoint in Figure 3). The more distant shots,

Figure 3. Preliminary P-velocity model for the western part of the east-west seismic refraction profile through Newberry Volcano (Stauber and Berge, 1984). The iso-velocity contours are at uniform intervals between the labeled contours. Units are km/s. Newberry Volcano spans the section under the title "RECORDING ARRAY". Examples of the raypaths used in the present study are shown--Pg raypaths on the left and the wide-angle mid-crustal reflection on the right.

80 to 90 km away, produced a strong wide-angle reflection from the mid crust (east shotpoint in Figure 3).

Arrival-times were measured for both phases by filtering the seismic traces with a narrow-band filter (centered on the dominant frequency of the desired phase) and then measuring the time of a peak or trough in the first cycle. An example of 21 narrow-band filtered Pg traces from one shot are shown in Figure 4. Uncertainties in the arrival-time measurements are less than a few hundredths of a second.

We compute the traveltime residuals with respect to a laterally homogeneous earth structure. Over the array dimension, the expected traveltime versus distance curve for the Pg phase departs from a straight line by less than a hundredth of a second. Consequently, we computed residuals by fitting lines of the following form to the measured Pg arrival times:

$$T_{ij} = T_j^0 + S \cdot D_{ij} + R_{ij}$$

where T_{ij} is the observed traveltime at the i th station for the j th shot, T_j^0 is a constant for the j th shot, S is the slowness, and D_{ij} is the distance between the j th shot and the i th station, and R_{ij} is the traveltime residual due to velocity deviations from

laterally homogeneous structure. T_j (intercept time) and S (slope of traveltime linear fit) were computed by a least-squares algorithm using all of the measured Pg times from the six closer shots. Residuals, R_{ij} , were then computed by subtracting the predicted arrival times, $T_j + S \cdot D_{ij}$, from the observed times.

The residuals have a simple geometric interpretation. If the raypath for a particular observation passes through a region with lower than average velocity for that depth, the arrival-time will be later than predicted and the residual will be positive. Conversely, higher than normal velocities produce negative residuals. As the direction to the shot is varied, rays passing through the anomalous region reach the surface at varying positions. The deeper the anomalous region, the greater the change in position of the traveltime anomaly. By observing how the residual pattern varies with direction of approach of the seismic waves one can qualitatively infer the locations of anomalous regions. The process can be made quantitative by using a linear inversion technique.

DATA AND QUALITATIVE INTERPRETATION

Pg residuals for four shots are plotted in map view in Figure 5. The first feature to note is the large magnitude of the residual variations, up to several tenths of seconds—much larger than the uncertainties of the measurements. The north and south sides of the array are relatively late (positive residuals) and the central and western parts of the array are relatively early. This pattern indicates that the average velocities beneath the northern and southern parts of the array are lower than beneath the central and western areas.

Within the caldera, the anomaly pattern varies systematically with azimuth. In Figure 5A (for a shot from the west) a partial ring of negative residuals occurs near the inner ring fault system on the north, west, and south edges of the caldera; extending from the center of the caldera toward the east is a zone of positive residuals. In Figure 5B (for a shot to the southwest) this pattern has rotated such that the later arrivals are in the northeastern part of the caldera with the early arrivals near the ring faults completely surrounding them. As the direction to the source rotates further, to just east of south (Figure 5C), the residual pattern in the caldera again rotates with it. The late arrivals now extend away from the caldera to the north-northwest. The early arrivals near the ring faults now extend from the west through the south, east, and northeast part of the caldera. Finally, in Figure 5D (shot to the east) the late arrivals are located along the northwest edge of the caldera with the early arrivals in the south and east.

Figure 4. Reduced time plot of 21 narrow-band filtered seismograms for Pg. The first large trough, indicated by the arrow near the center trace, was timed in this case.

This pattern of late arrivals on the opposite side of the caldera from the source and a ring of early arrivals around the rest of the caldera can be interpreted in the following way. A ring of high-velocity material exists in the upper 2 km all the way around the inner ring fault system of the caldera producing the early arrivals observed there. In the center of the caldera and extending to a depth of greater than 2 km is a zone of low-velocity material which casts a "shadow" of late arrivals on the side of the caldera opposite the shot. The widths of these early and late arrival zones are on the order of 2 to 3 km and the magnitude of the residual variation is 0.15 to 0.25 s. These dimensions require P-velocity variations of 20 to 40%.

The presence of a ring of high-velocity rock in the upper 2 km around the ring fault system is consistent with an interpretation of gravity data made by Griscom and Roberts (1983). They suggest that a ring of short-wavelength gravity highs around the ring faults are the result of a ring of high density intrusive rocks in the upper

Figure 5. Pg residuals for 4 shots recorded at the Newberry caldera array. The values are in units of hundredths of seconds and are contoured at 0.1 s intervals. Areas with residuals greater than 0.1 s and less than -0.1 s are indicated by the shading. The heavy dark lines show the caldera ring fault system. Dark arrows show the direction of wave propagation across the array.

few km. Higher density and higher P-velocity are expected in crystalline intrusive bodies due to lower porosity and lower glass content than the surrounding extrusive rocks. This effect is clearly demonstrated for the rift zones of Kilauea volcano by Ellsworth and Koyanagi (1977) and Broyles et al. (1979). A zone of frequent small intrusions in the upper few km around the ring fault system at Newberry Volcano is an attractive candidate for the heat source of the hot well water found in the caldera.

The source of the rhyolitic magmas erupted in the caldera is not addressed with this ring-intrusion model. The teleseismic P-residual experiment ruled out the presence of a magma chamber below 5 km depth with dimensions of 5 km or greater. However, the deeper portion of the low-velocity zone located in the center of the caldera by the present study is a possible candidate for a small silicic magma chamber. A formal inversion of the present Pg and wide-angle reflection data is needed to evaluate this hypothesis (and is underway).

CONCLUSIONS

Qualitative analysis of the traveltime residual patterns obtained by a three-dimensional seismic imaging experiment at Newberry Volcano indicates that P-velocity anomalies on the order of 30%, are present in the upper 5 km of the volcano. In particular, a ring of high-velocity material in the upper 2 km around the caldera ring fault system is inferred to exist and coincide with a density anomaly inferred by Griscom and Roberts (1983). A low-velocity zone in the center of the caldera is inferred to extend somewhat deeper than the high-velocity ring and is a possible source region for the high-silica rhyolitic magmas which have erupted in the caldera several times in the last 6000 years.

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