GL01877

ELECTRICAL RESISTIVITY ANOMALIES AT NEWBERRY VOLCANO, OREGON; COMPARISON WITH ALTERATION MINERALOGY IN GEO COREHOLE N-1

Geothermal Resources Council, Transactions, V. 10 p. 247-252

Phillip M. Wright and Dennis L. Nielson, 1986

Earth Science Laboratory
University of Utah Research Institute
Salt Lake City, Utah 84108

ABSTRACT

Corehole N-1, drilled on the south flank of Newberry volcano in Oregon by GEO Operator Corp under a cooperative agreement with DOE, encountered about 15 separate horizons between 2800 ft. and 4000 ft. that are good electrical conductors as shown by an induction log. These conductors correlate with horizons of altered basaltic, andesitic and lactitic ash and tuff in a lithologic section composed predominately of basaltic andesite flows. X-ray diffraction and scanning electron microscope analyses show the dominant alteration type to be a calcium smectite and we believe that the alteration is low temperaure in origin. Surface electrical geophysical surveys have detected a widespread resistivity low in the Newberry area. A portion of this resistivity low is believed to be associated with the high-temperature hydrothermal system in Newberry caldera, whereas other portions of the low appear to be caused by the altered horizons on the flanks of the volcano. Delineation of the high-temperature system by electrical surveys may be difficult or impossible because of effects from the altered rocks.

INTRODUCTION

Newberry volcano has been the site of intensive geothermal exploration activity and scientific studies of geothermal processes in a young volcanic environment. As a result of this interest, a great deal of information is currently available through the efforts of the U.S. Geological Survey (USGS) as well as those of private exploration companies. As part of a DOE-sponsored geothermal-gradient drilling program in the Cascade range, additional information and core samples have recently become available. This paper uses data from corehole N-1, drilled by GEO Operator Corp. under the cooperative program with DOE, to describe the resistivity structure under the south flank of Newberry volcano. We discuss low-resistivity zones in GEO N-1, evaluate the chemical and physical processes that have produced these zones, and present caveats concerning the interpretation of electrical resistivity surveys in volcanic environments. In particular, can resistivity anomalies produced by geothermal systems be distinguished from resistivity anomalies developed as part of the volcanic evolution of the system?

Newberry is one of the largest Quaternary volcanoes in the continental United States. It is a shield volcano, located east of the axis of the High Cascades volcanic province of central Oregon. Shield volcanoes are characterized by basaltic eruptions, but Newberry has also experienced important eruptions of silicic rocks forming domes and flows and depositing significant volumes of ash and ash-flow tuff. The volcano has a summit caldera which is six to eight miles across, and MacLeod and Sammel (1982) believe that the caldera collapse was produced by emptying of a magma chamber during tephra eruptions. In addition, eruptions of basaltic cinders are common, and the flanks of the volcano are covered with small cinder cones (MacLeod et al., 1982).

A high-temperature geothermal system exists at Newberry as demonstrated by corehole Newberry-2 (Fig. 1) which was drilled by the USGS (Sammel, 1981; MacLeod and Sammel, 1982). This hole was drilled within the summit caldera and reached a bottom-hole temperature of 265°C at a depth of 932 meters (3058 ft). Corehole GEO N-1 was drilled in 1985 on the south flank of the caldera (Fig. 1) approximately three miles from the southern caldera margin (UURI, 1986). The hole was drilled to help determine the depth of the rain curtain, or the extent to which cold, meteoric water penetrates the Newberry volcanic pile. GEO N-1 intersected a sequence principally composed of basaltic-andesite lava flows and associated pyroclastics. However, there were notable sequences of rhyolitic ash and ash-flow tuff, some of which will be described below. Core was collected from 487 ft to TD at an unknown depth greater than 4000 ft. In addition, a suite of geophysical logs was run in the hole providing an opportunity for investigation of log interpretation in this young volcanic environment. Data and samples are in the public domain only for depths 4000 ft and less.

Surveys designed to measure the electrical resistivity of rock at depth are a common tool in geothermal exploration. Rock resistivity is lowered by saturation with the high-temperature, commonly briney fluids in hydrothermal systems, yielding a contrast with rocks of higher resistivity exterior to the system. Surface electrical surveys at Newberry have been described by Bisdorf (1983), Fitterman (1983) and Fitterman et al. (1985). Figure 1 shows the results of a time-domain electromagnetic (TDEM) survey reported in the latter two references. Interpretation was done using a layered-earth (one-dimensional) model with either two or three layers. The indicated resistivity structure includes an uppermost layer 90 to 170 m (295-558 ft) thick of resistivity 850 to 5000 ohm-m and a second layer 470 to 650 m (1542-2133 ft) thick of resistivity 220 to 500 ohm-m. These two layers overlie the conductive horizon whose parameters are shown on Figure 1. These data indicate that electrically conductive horizons, having a

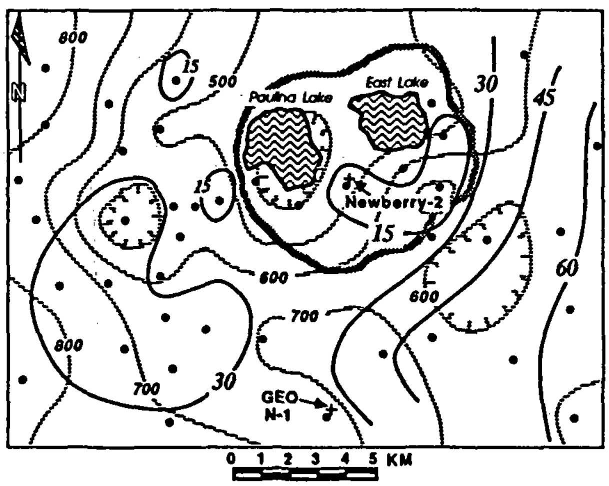

FIGURE 1
Time domain electromagnetic survey of the Newberry volcano (Fitterman, 1983; Fitterman et al., 1985) showing interpreted depth in meters to horizons of low resistivity (shaded contours) and interpreted resistivity of the horizons in ohm-m (solid contours). Station locations are shown by dots. The bold line shows the outline of the Newberry caldera.

resistivity about one order of magnitude lower than the layer above, underlie a substantial portion of the Newberry area, continuing well outside the caldera. The areal limits to this low-resistivity anomaly were not defined by the reported surveys. The cause of such a widespread electrical conductor and its implications in prospecting for extensions of the high-temperature hydrothermal fluids discovered in USGS Newberry-2 are topics of natural curiosity which form the basis for this paper.

RESISTIVITY DATA FROM GEO NEWBERRY COREHOLE N-1

Geophysical logs were run in GEO N-1, and the public-domain portions of these logs are available from Rocky Mountain Well Log Service in Denver, Colorado. Logs that are available include caliper, temperature, natural gamma, spontaneous potential (SP), 16-in short normal resistivity, induction, acoustic velocity and acoustic fraclog. Only the induction log is discussed here.

The electrical section in GEO N-1 can be divided into two broad regimes. The upper, higher resistivity regime includes the interval 490 ft, where the logs begin, to 1250 ft, which shows an average resistivity of about 70 ohm-m with a few thin horizons of higher resistivity (to > 500 ohm-m) and lower resistivity (±25ohm-m). Between 1250 and 1780 ft, there are no electric logs. Between 1780 ft and 2800 ft, resistivity values average about 50 ohm-m with thin horizons as high as 300 ohm-m and as low as 25 ohm-m. The lower regime comprises the interval 2800 ft to 4000 ft. In the lower regime, there is a marked decrease in average resistivity and an increase in the occurrence of highly conductive horizons, many of which show resistivity values lower than 10 ohm-m.

Figure 2 shows a selected portion of the induction log along with a simplified lithologic log and data from a separate temperature survey completed about one week later than the electric logs. The induction log was selected for analysis because it illustrates the conductive horizons of interest and because it is expected to measure values more representative of the true formation resistivity than the 16-in short-normal log. The parameter of the induction log is electrical conductivity in siemens per meter (S/m), which is the reciprocal of resistivity in ohm-m. Deflections to the left on this log indicate higher conductivity, i.e. lower resisitivity. The log shows scales of both conductivity and resistivity for reference. Conductive zones can be seen near 2830 ft, 2890 ft, 3110 ft, 3350 ft, 3430 ft, 3470 ft, 3490 ft, 3600 ft, 3670 ft, 3710 ft, 3830 ft and 3880 ft. These conductive horizons are observed to correlate with altered ash and tuffaceous units.

MINERALOGY OF LOW RESISTIVITY ZONES

Several zones which exhibit low resistivity on the geophysical well logs were chosen for mineralogical study. These zones are summarized in Table 1 and are described in more detail below.

2825-2835 ft. A 15 ohm-m anomaly is coincident with a sequence of basaltic ash and cinders which was probably deposited on the flanks of a cinder cone. X-ray diffraction analysis shows that the sample contains approximately 5% smectite. The only other mineral detected was plagioclase, which was a phenocryst phase in the ash and cinders. A scanning electron microscope (SEM) image of this sample (Fig. 3) shows the incipient formation of Ca-smectite from the volcanic glass. The smectite forms a complete cover on some of the

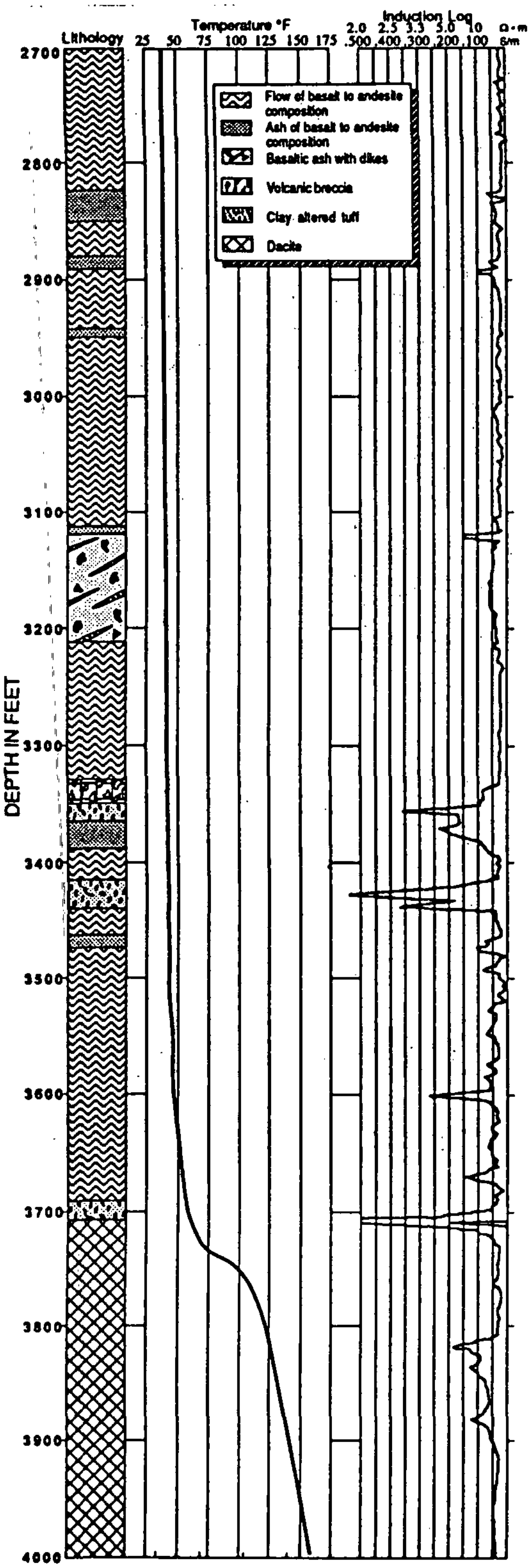

FIGURE 2
Lithologic, temperature, and induction log for a portion of well GEO N-1.

TABLE 1

Depth (ft)	Resistivity Ω-m	Temperature °F	Lithology
2825-2835	15	43	Basaltic cinders and ash - Sample @ 2830 ft shows plagioclase phenocrysts and basaltic glass altering to smectite (5% sm)
2888-2889.5	10	49	Compact clay-altered felsic ash - Sample @ 2889 ft containing 37% Ca-smectite with plagioclase phenocrysts and volcanic glass
3113-3120	6	56	Densely welded ash-flow tuff altered to 25% smectite + halloysite (7Å) + cristobalite + plagioclase
3702	1.5	144	Smectite-rich (49%) altered ash with cristobalite and K-feldspar

FIGURE 3
SEM image of basaltic ash from 2830 ft in GEO N-1 showing the formation of Ca-smectite.

glass to a thickness of less than one micron. Other areas have widely separated points at which crystals of smectite have nucleated but have not yet coalesced to form a continuous layer. Above 2825 ft is a vesicular flow breccia which extends to 2823 ft, with a dense basaltic-andesite flow above that. This flow shows very minor vesiculation and some fractures which contain small amounts of calcite and clay. Below the anomalous cinder unit is a sequence of basaltic debris that contains coarse blocks of basalt.

<u>2888-2889.5 ft.</u> A 10 ohm-m anomaly is associated with a compact, clay-altered felsic ash. This ash is well bedded and sorted and was deposited in a lacustrine environment. X-ray diffraction shows that the rock contains 37% Ca-

smectite with plagioclase as a relict phenocryst phase. Below 2889.5 ft is a section of basaltic cinders which continues to about 2901 ft, where a dense porphyritic basalt flow is encountered. The cinders are very porous and do contain some clay alteration along fractures. Above 2888 ft is a cinder sequence which is very similar to that described in the 2825-2835 ft interval.

3113-3120 ft. This is a 6 ohm-m zone which is correlated with an ash-flow tuff interval. A portion of the ash-flow tuff is densely welded; however, the entire sequence is altered to an assemblage which consists principally of clay minerals. X-ray diffraction analysis at 3117 ft shows that the clay phases consists of 27% Ca-smectite and a 7 Å-halloysite. α-cristobalite and plagioclase are also present, with plagioclase representing an unaltered phenocryst phase. Above this unit is a vesicular basalt flow which shows no signs of alteration. Below the felsic ash, a basalt flow or intrusion contains numerous small vesicles, but permeability is quite low. The absence of alteration above and below the ash-flow tuff interval and the pervasive alteration of the unit itself, even when part of it is densely welded, suggests the possibility of formation by hydromagmatic activity. This phenomena is associated with the interaction of magmas with ground waters, geothermal waters or surface waters during eruption (Sheridan and Wohletz, 1983; Walker et al., 1984). The material is erupted wet and depositional units are altered even though overlying and underlying units may be unaffected.

3694-3708 ft. This is a felsic ash sequence which has an associated resistivity anomaly of 1.5 ohm-m over a short interval. At 3702 ft there is a clay-rich zone which could have been developed as a weathering horizon within the felsic ash. X-ray diffraction analysis shows that this layer is composed of 49% smectite with α-cristobalite and K-feldspar. There is no glass detectable in the X-ray pattern. SEM analysis shows that the clay is a Ca-smectite (Fig. 4). Much of the clay is pseudomorphous after volcanic glass, while the material which fills voids is crystalline with a texture similar to that seen at 2830 ft. A sample at 3708 ft shows the above assemblage accompanied by unaltered volcanic glass, demonstrating that the alteration reaction has not gone to completion. Smectite constitutes 24% of this sample. The resistivity increases to about 7 ohm-m as a result of the reduction of the smectite content. Above this sequence is a basaltic volcanic breccia. Below the ash unit is a dense dike or sill of dacitic composition. Again, neither the overlying nor underlying units contain more than minor amounts of clay.

DISCUSSION

Fluids present in a rock constitute the medium in which electrical current is carried. The drilling showed that the water level in the vicinity of GEO N-1 is at a depth of about 1600 ft below the surface. Efforts to collect samples of formation fluid were not successful. However, chemical analyses exist for a warm well in the Little Crater Campground (Mariner et al., 1980). Waters in this well are at a temperature of 35.5°C and are slightly saline at 700 ppm total dissolved solids. For this study we assume that this analysis is representative of the fluids encountered in the lower portions of GEO N-1. The upper portions of N-1 are being flushed with cold meteoric waters, and they are likely to be less saline than the deeper waters.

FIGURE 4
SEM image of felsic ash from 3701 ft in GEO N-1 showing textural character of Ca-smectite.

The mineral transformations responsible for the clays which produce the low electrical resistivity anomalies in GEO N-1 are low-temperature reactions. The alteration could have taken place in either the lacustrine or in the goundwater environment at temperatures observed in GEO N-1. Principally, the reactions involve the transformation of volcanic glass, both basaltic and rhyolitic, to clay minerals of the smectite family. Depending on the chemistry of the volcanic glass and the composition of the smectite produced, the alteration reaction takes the general form

$$\text{glass} + H_2O \rightarrow \text{smectite} + M + SiO_2,$$

where M represents cations, principally Ca^{++}, Na^{+}, and K^{+}. Additional products of the reaction can be other clays, such as kaolinite, if the fluids reacting with the glass are acidic, or zeolites if they are basic (Slaughter and Earley, 1965). The 7 Å-halloysite noted at 3117 ft may have developed in this manner.

Evidence for the low-temperature, non-geothermal, nature of these reactions is two-fold. First, the alteration is concentrated in specific stratigraphic horizons. Overlying and underlying rocks show only minor or no alteration effects. Second, mineral assemblages characteristic of hydrothermal alteration are not present. The upper temperature limit of smectite is approximately 160°C, and this is where the mineral essentially disappears in USGS Newberry-2 (Bargar and Keith, 1984). In many geothermal systems, mixed-layer illite-smectite (I/S) will form by 100°C (Hulen and Nielson, 1986; Fig. 14). No I/S clay has been found in our investigation of the Newberry core. In short, there is no evidence that the rocks in GEO N-1 have experienced post-formation temperatures higher than they are experiencing at the present time.

Low frequency electrical conduction in rocks is generally some combination of three types: (1) ionic conduction due to movement of ions in the fluids filling the pore spaces, (2) surface conduction due to movement of ions adsorbed on the surfaces of minerals, and (3) electronic conduction in metallic minerals. Because the rocks from GEO N-1 contain no metallic conducting minerals, the third mechanism may be discounted. There is no simple way to separate completely the effects of ionic and surface conduction for the rocks found in GEO N-1. For ionic conduction, Archie (1942) gave the classic relationship among the bulk resistivity of the rock, the resistivity of the saturating fluid and the porosity. If the calculated resistivity of 7 ohm-m at 50°F for the water from a warm well at Little Crater Campground (Mariner et al., 1980) is used along with an estimated porosity of 5% percent in the massive flows, the rock resistivity should be 2800 ohm-m. The measured resistivity is 50 ohm-m. We conclude that surface conduction is important even in the rocks in GEO N-1 which contain no clays. Ions are sorbed on the surfaces of pores and cracks, loosely held there by unsatisfied charges along boundaries in the glass and phenocryst minerals. These ions are mobile enough to migrate under an electric field, thus enhancing the conductivity.

Certain minerals, including the clays and zeolites, show strong tendencies to adsorb ions from the saturating fluids, and these ions are also able to migrate under an electric field. This surface conduction effect in clay-bearing rocks has been well studied in the petroleum environment (Waxman and Smits, 1968; Daniels et al., 1977), but relatively little is known about the effects of clay minerals in volcanic rocks. It seems certain, however, that the low-resistivity values observed in the smectite-altered horizons in GEO N-1 are due primarily to surface conduction effects from this ion-adsorbing mineral.

It is also apparent that the combined effects of the conducting horizons between about 2800 ft and 4000 ft (and perhaps deeper) in GEO N-1 cause the resistivity low observed on surface electrical surveys in the vicinity of GEO N-1. The average resistivity for the interval 2800 ft to 4000 ft is determined from the logs to be about 28 ohm-m, ignoring possible anisotropic effects. This is in good agreement with the lower-layer resistivity interpreted by Fitterman (1983) and Fitterman et al. (1985). The depth to the conductive horizons in GEO N-1 is also in reasonable agreement with interpretations of the surface surveys (see Fig. 1).

The temperature log of GEO N-1 shows an abrupt increase below about 3260 ft. Meteoric water must circulate quite freely to this depth, because temperatures above are effectively clamped at the mean annual air temperature, near 50°F. Below about 3500 ft, the temperature gradient is reasonably uniform and is anomalously high at about 115°C/km, based on temperature data taken about 5 days after last circulation. There is no guarantee, however, that this high gradient represents heat flow from depth because hydrologic effects could still be present. The resistivity effects of increasing temperature and possibly increasing salinity below 3260 ft can be seen by comparison of the log responses in the relatively massive, uniform flows above this depth (40-50 ohm-m) and below this depth (25-35 ohm-m). Effects of abruply increased temperature and possibly salinity below the base of the rain curtain at 3260 ft may likewise influence the conductivity of the altered horizons. We observe that the conductive zones are not only more plentiful but also more highly conductive below 3300 ft. Frequency of occurence of the altered horizons is mainly controlled by lithology, but their conductivities are controlled by intensity of alteration as well as by the temperature and salinity of the pore fluids. Intensity of alteration may itself be controlled to some extent by the salinity and temperature of the pore fluids. There may be some possibility of using considerations such as these to map the base of the rain curtain with an appropriate electrical geophysical technique.

It is not presently clear how surface electrical anomolies due to low-temperature alteration of the type found in GEO N-1 could be separated from anomolies due to the high-temperature hydrothermal system found in USGS Newberry-2. It may be possible to separate the two causes of decreased resistivity using induced polarization (IP). IP effects from membrane polarization are caused by adsorption of ions on minerals, and the effects increase as cation exchange capacity (CEC) of the minerals increases (Keller and Frischknecht, 1966). Smectite-group clays are known typically to have a much higher CEC than the chlorites, and smectite was observed to decrease downhole as temperature increased in USGS Newberry-2 whereas chlorite-smectite and chlorite increased (Bargar and Keith, 1984). Moreover, the conductivity of high-temperature alteration phases may be lower than the conductivity of low-temperature phases in some circumstances, and one can conceive of the electrical expression of the hydrothermal system around USGS Newberry-2 as consisting of lower resistivity with higher IP response at shallower depths and higher resistivity with lower IP response at greater depths. From the low-temperature alteration phases found in GEO N-1 we would expect low reisistivity and high IP response. We are, however, in the realm of speculation. Research work is needed to clarify these issues.

CONCLUSIONS

The implications of the results reported here for electrical geophysical prospecting in the Newberry area are obvious. Surface electrical surveys may be misleading in terms of delineating subsurface hydrothermal systems. There is some possibility that electrical surveys could help map the thickness of the rain curtain in this area, but a relationship among alteration, electrical conductivity and the base of the rain curtain is not established. Separation of geothermal resistivity anamalies from those due to low-temperature alteration is a possibility that remains to be explored. To the extent that other volcanos in the Cascades have the same type of ash and tuff horizons susceptible to alteration, electrical geophysical surveys should be used with caution.

ACKNOWLEDGEMENTS

The work reported here was done under U.S. Department of Energy Contract No. DE-AC07-85ID12489. The SEM photographs were made by Thom Little of Terra Tek. Beth Orvin typed the manuscript and Pat Daubner made the illustrations. We are all grateful to all of them. More information on the Cascades Deep Geothermal Gradient Drilling Program can be obtained from Susan Prestwich, DOE/Idaho Operations Office, Idaho Falls, Idaho.

REFERENCES

Archie, G. E., 1942, The electrical resistivity log as an aid in determining some reservoir characteristics: Trans. AIME, v. 146, p. 54-62.

Bargar, K. E., and Keith, T. E. C., 1984, Hydrothermal alteration mineralogy in Newberry 2 drill core, Newberry volcano, Oregon: U.S. Geological Survey Open-File Report 84-92, 50 p.

Bisdorf, R. J., 1983, Schlumberger soundings near Newberry Caldera, Oregon: U.S. Geological Survey, Open-File Report 83-825.

Daniels, J. J., Scott, J. H., Blackmon, P. D., and Starkey, H. S., 1977, Borehole geophysical investigations in the south Texas uranium district: Jour. Research U.S. Geol. Survey, v. 5, n. 3, p. 343-357.

Fitterman, D. V., 1983, Time-domain electromagnetic soundings of Newberry Volcano, Deschutes County, Oregon: U.S. Geological Survey Open-File Report 83-832.

Fitterman, D. V., Neev, D. K., Bradley, J. A., Grose, C. T., 1985, More time-domain electromagnetic soundings of Newberry Volcano, Deschutes County, Oregon: U.S. Geological Survey, Open-File Report 85-451.

Hulen, J. B., and Nielson, D. L., 1986, Hydrothermal alteration in the Baca geothermal system, Redondo dome, Valles caldera, New Mexico: Journal of Geophysical Research, v. 91, p. 1867-1886.

Keller, G. V., Frischknecht, F. C., 1966, Electrical methods in geophysical prospecting: Pergamon Press Inc., 519 p.

MacLeod, N. S., and Sammel, E. A., 1982, Newberry volcano, Oregon: A Cascade range geothermal prospect: Oregon Department of Geology and Mineral Industries, Oregon Geology, v. 44, p. 123-131.

MacLeod, N. S., Sherrod, D. R., Chitwood, L. A., 1982, Geologic map of Newberry volcano, Deschutes, Klamath and Lake Counties, Oregon: U.S. Geological Survey Open-File Report 82-847.

Mariner, R. H., Swanson, J. R., Orris, G. J., Presser, J. S., and Evans, W. C., 1980, Chemical and isotopic data for water from thermal springs and wells of Oregon: U.S. Geological Survey Open-File Report 80-737,

Sammel, E. A., 1981, Results of test drilling at Newberry volcano, Oregon: Geothermal Resources Council Bulletin, v, 10, n. 11, p. 3-8.

Sheridan, M. F., and Wohletz, K. H., 1983, Hydromagmatism: basic considerations and review: Journal of Volcanology and Geothermal Research, v. 17, p. 1-29.

Slaughter, M., and Earley, J. W., 1965, Mineralogy and geological significance of the Mowry bentonites, Wyoming: Geological Society of America Special Paper 83, 116 p.

UURI, 1986, Open file release of data from the GEO Operator corehole N-1 at Newberry Volcano, Oregon.

Walker, G. P. L., Self, S., and Wilson, L., 1984, Tarawera 1886, New Zealand - A basaltic Plinian fissure eruption: Journal of Volcanology and Geothermal Research, v. 21, p. 61-78.

Waxman, M. H., Smits, J. J. M., 1968, Electrical conductivities in oil-bearing shaly sands: Amer. Institute of Min., Metallurg., and Petrol. Eng., Trans., v. 243, p. 107-122.