THE DEEP CONVECTIVE GEOTHERMAL SYSTEMS OF THE TAUPO VOLCANIC ZONE, NEW ZEALAND

H. M. Bibby, T. G. Caldwell, G. F. Risk

Institute of Geological and Nuclear Sciences, P O Box 1320, Wellington, New Zealand

ABSTRACT

Electrical resistivity surveys carried out over the last 30 years provide almost complete coverage of the Taupo Volcanic Zone (TVZ), an area of about 5000 km^2. Low resistivity anomalies in these data show the locations and distribution of some 23 separate geothermal fields through which the entire heat output (about 4200 MW(t)) of the TVZ is passed. Of the 23 fields, 17 have high natural heat outputs (>20 MW(t)); the remaining 6 may represent formerly active fields now in decline. In addition, there are at least two extinct fields. In the region between the geothermal fields, covering 94% of the total surface area, the heat flow is extremely small or zero.

A large-scale system of convecting fluids underlies the entire TVZ. Cold waters of meteoric origin are collected over the whole region and permeate downwards to be heated by the deep hot volcanic rocks. These hot waters rise under buoyancy and become concentrated in narrow plumes with cross-sectional areas of, on average, about 15 km^2. The geothermal fields are the near-surface expressions of these up-flowing plumes. Deep resistivity surveys suggest that the cross sectional areas of the plumes may increase with depth, consistent with the observed decrease in permeability. Chemical and detailed heat-flow studies indicate that once the hot geothermal fluids reach the surface they join the local drainage, and only a very small portion rejoins the deep circulating system.

KEY WORDS: Geothermal fields, Taupo Volcanic Zone, convection plumes, hydrothermal systems, resistivity.

INTRODUCTION

The Taupo Volcanic Zone (TVZ) of New Zealand (Fig. 1) is characterised by the high frequency of eruptions from both rhyolitic calderas and andesitic composite volcanoes (Wilson et al., in press) and by a large number of liquid-dominated geothermal fields. The total natural geothermal heat output of the TVZ, from the Bay of Plenty Coast to Tokaanu, 4200 ± 500 MW(t) (Bibby et al., in press), is discharged through these geothermal fields. The long-term energy output of the geothermal fields is at least twice that of the energy released through volcanic activity (Hochstein et al., 1993).

The setting and the distribution of the geothermal fields of the TVZ are also unusual. Unlike the andesitic arc environment, where the distribution of geothermal fields is aligned along the arc with spacings of the order of 100 km (e.g. Hochstein, 1991), the TVZ is characterised by a two-dimensional distribution of fields with average spacing of about 15 km, as we show here. Although the New Zealand setting is unusual, the TVZ provides an example of large-scale convective heat transport in the crust, features of which are applicable to many hydrothermal systems, regardless of their setting.

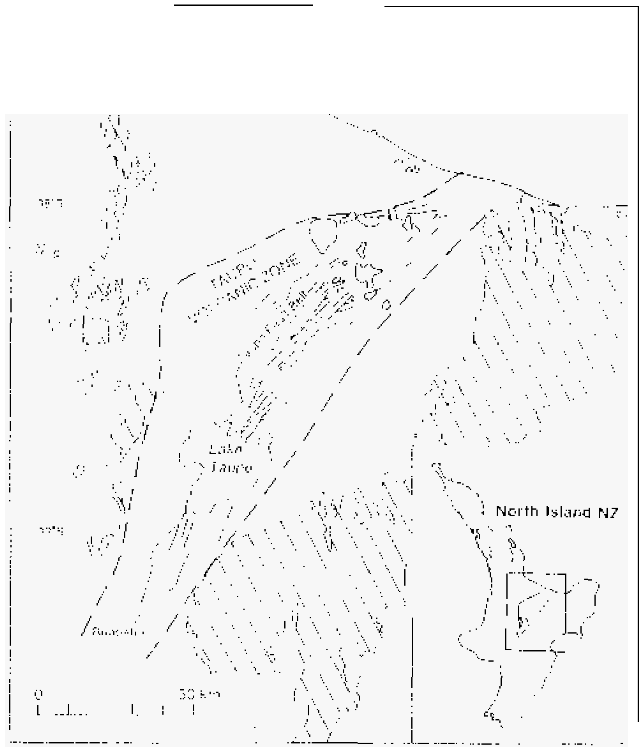

Figure 1. The Taupo Volcanic Zone, which is defined to enclose all the active volcanism that has occurred in the Central North Island over the last 1.6 My, contains all but one of the high temperature geothermal fields of New Zealand. Hatching to the east and west indicates outcropping Mesozoic greywacke rocks.

The data presented in this paper have been acquired over many years, and, during this time, our concepts of the deep convective hydrothermal systems have gradually become more refined.

RESISTIVITY MAPPING

The earliest geophysical investigations of the TVZ geothermal fields were mainly concentrated on determination of geological structure. In particular, gravity, seismic refraction and magnetic surveying were carried out (e.g. Banwell et al., 1957; Studt, 1958) in attempts to identify structural features believed to control the passages of the hot fluid to the surface. Electrical resistivity surveying techniques were first successfully applied to geothermal exploration in the mid 1960s (Hatherton et al., 1966; Macdonald, 1967). These techniques have proved to be the most effective geophysical tools for delineating TVZ geothermal fields.

The effectiveness of resistivity techniques relies on a number of factors, all of which combine to lower the electrical resistivity of

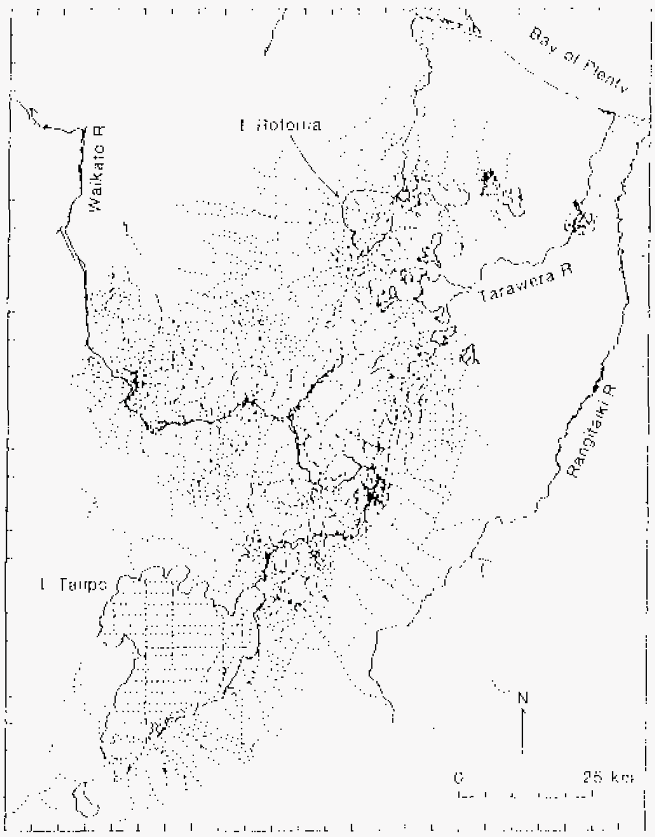

Figure 2. Distribution of the Schlumberger resistivity measurements made in the TVZ over the last 30 years. Many of the data have been collected during detailed investigations of individual geothermal fields.

a geothermal area with respect to its surroundings. Geothermal waters of the TVZ contain high concentrations of dissolved salts (typically 1000 ppm chloride), which provide a highly conducting electrolyte within the rock matrix. The electrical properties of this conductive fluid are also strongly temperature-dependent. For example, at 200°C, the resistivity of such a saline solution is only 10% of that at ambient temperatures (Quist & Marshall, 1968; Olhoeft, 1981). Consequently, waters of geothermal origin that have cooled during flow are less conductive than the same waters near their (hot) source. In addition, the rock matrix also makes an important contribution to ground conductivity. With prolonged passage of geothermal waters, the volcanic rock matrix undergoes hydrothermal alteration, which produces clay- and zeolite-rich products that can be very conductive. Typical resistivities of altered volcanic rocks are in the range of 10 - 50 Ωm at temperatures of 20 - 100°C (Caldwell et al., 1986).

These factors act together to reduce the electrical resistivity of a geothermal field, which allows the boundaries of the field to be mapped readily using appropriate resistivity mapping techniques. These techniques have proved highly effective in the young volcanic rocks of the TVZ. Since the initial success of electrical techniques for outlining the Wairakei geothermal field (Banwell & Macdonald, 1965), resistivity mapping has been routinely carried out both for reconnaissance and for providing detailed information on the boundaries of the hot water for drilling purposes (Risk, 1983). A database has been established in which all the reconnaissance resistivity values measured over the last 30 years are held.

The majority of the resistivity measurements in the TVZ have been made using the Schlumberger electrode array, with array spacing (AB/2) of 1,000 m and 500 m. Early measurements, made using the Wenner array with a spacing of 1800 ft (549 m) can be shown to be compatible with the 500 m Schlumberger data. The data reflect changes in the electrical resistivity to depths of about 200-300 m. Details of the measurement techniques used in New Zealand are given in Bibby (1988). Measurements have also been made in the many lakes of the TVZ. In shallow lakes a Schlumberger array, with the potential electrodes placed on the lake floor, has been used (Bennie, 1983). In Lake Taupo, where deep water (>100 m) prohibits the use of electrodes on the lake floor, a method using a continuously moving array has been specially developed (Caldwell & Bibby, 1992). These data have been measured using an array spacing of 500 m only.

In total, about 24,000 individual Schlumberger measurements have been made within the TVZ, distributed as shown in Fig. 2. One of tlic reasons fur the success of the resistivity mapping programme was the early recognition that measurements should not be confined to the environs of the geothermal fields but should extend between these areas to give a near-uniform coverage of the whole of the volcanic region. Many of the data shown iri Fig. 2 have been measured as part of the investigations of particular geothermal fields. In addition, a series of apparent resistivity maps, at a scale of 1:50 000, are being published (Geophysics Division, 1985; Bibby et al., 1997).

All the publicly available data within the TVZ were used fa compile Fig. 3 which shows contours of the apparent resistivity measured with [he 500 m Schlumberger array. Contouring was carried out by interpolating onto a 1 km grid rising the weighting techniques described by Bibby (1988). Contours are logarithmically spaced, using 3 contours per decade, at values of 5, 10, 20, *SO*, 100 . . Ωm. Contours have been limited tu regions where the density of data is sufficient to contour with confidence.

DISTRIBUTION OF GEOTHERMAL FIELDS

The eastern margin of the TVZ is clearly reflected in the apparent resistivity measurements as a steep gradient in resistivity. Values of 400-500 Ωm occur cast of the TVZ over most of the Kaingaroa Plateau. The rapid decrease in resistivity towards the TVZ reflects both the increasing depth to the highly resistive greywacke basement (Risk ct al., 1993) and the onset of geothermal phenomenn within the TVZ itself.

Near the coast, two zones of low resistivity are pi-eserit. In these zones, low resistivities reflect the presence of highly conductive sea water at shallow depths beneath the coastal plains. Within ttic TVZ there is a very strong correlation between the occurrence of geothermal waters at the surface arid the low-resistivity zones outlined in Fig. 3. With the exception of the coastal anomalies, each of the low-resistivity zones can be linked to geothermal activity, past or present. All hot springs and other manifestations of geothermal activity occur either within these low resistivity zones (geothermal fields) *or* result from near-surface drainage from the geothermal fields. Thus, Fig. 3 shows the distribution of geothermal fields within the TVZ. Note that at least one of the geothermal fields lies beneath the waters of Lake Taupo, although the contrast in the apparent resistivity is muted by the 100 m of high-resistivity lake water (Caldwell & Bibby, 1992).

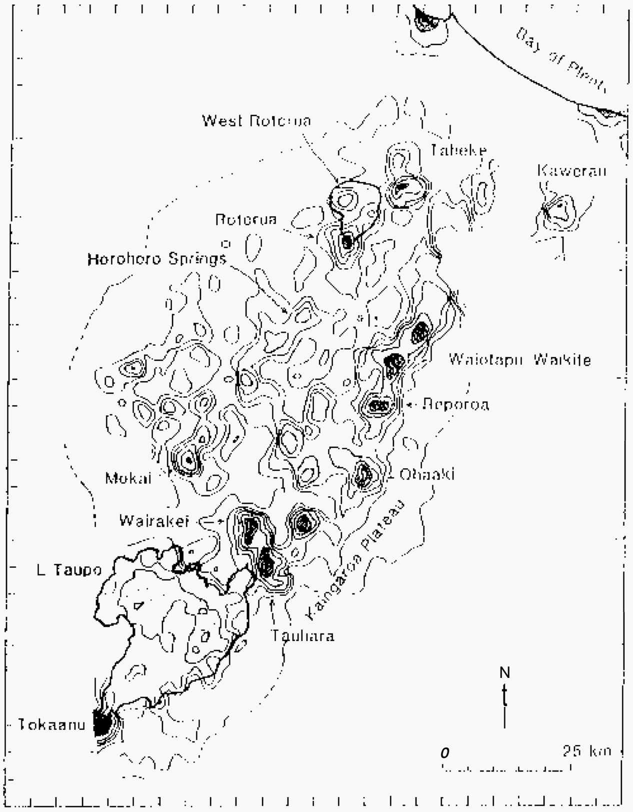

Figure 3. Contours of the apparent resistivity over the TVZ, drawn at logarithmic intervals (5, 10, 20, 50, 100 ... Ωm). Resistivity data have been measured using the Schlumberger electrode configuration with nominal spacing (AB/2) of 500m. Heavy shading outlines regions where resistivity is less than 10 Ωm; light shading represents 10 – 50 Ωm. In areas where no data exist within a 3 km radius, gaps have been left. This shortage of data occurs mainly around the fringes of the map.

The number of geothermal fields that can be identified in the TVZ depends on how some of the features with low heat output are classified (e.g. Horohoro Springs). Seventeen geothermal fields with heat outputs greater than 20 MW(t) have been identified within the TVZ along with 6 other fields with lower heat flows.

It is possible that some of the anomalies mark geothermal fields undergoing decline. It would also be expected that some fields may have become extinct. In such cases, although the highly conductive geothermal fluids may no longer be present, a low-resistivity signature would be expected as a result of the past hydrothermal alteration of the rock matrix. *It* would therefore be expected that extinct geothermal fields would exhibit intermediate values of resistivity over areas comparable to those of the presently active fields. The anomaly at Ohakuri is such a fossil field (Henneberger, 1986) and at least one other fossil field exists. In the western part of Lake Rotorua, a low-resistivity zone of about 12 km^2 (Fig. *3)* shows no signs of geothermal activity (Whiteford, 1992). This feature has been interpreted as a fossil geothermal field (Bibby et al., 1992).

Between the Tokaanu geothermal field in the south and the Kawerau field in the north, the geothermal fields are spread fairly uniformly across a zone up to 50 km in width and about 130 km in length. The average spacing between the fields is about 15 km over most of this zone, although the fields on the eastern side are more closely spaced than those in the west.

BOUNDARIES OF THE GEOTHERMAL FIELDS

Because the contours of apparent resistivity shown in Fig. *3* are a smoothed representation of the data, they give a misleading impression of the sharpness of the boundary between the hot and cold regions. For every geothermal field in the TVZ that has been investigated in detail, abrupt changes in resistivity have been observed for at least some portion of the field boundary. This is best illustrated by examining the apparent resistivity

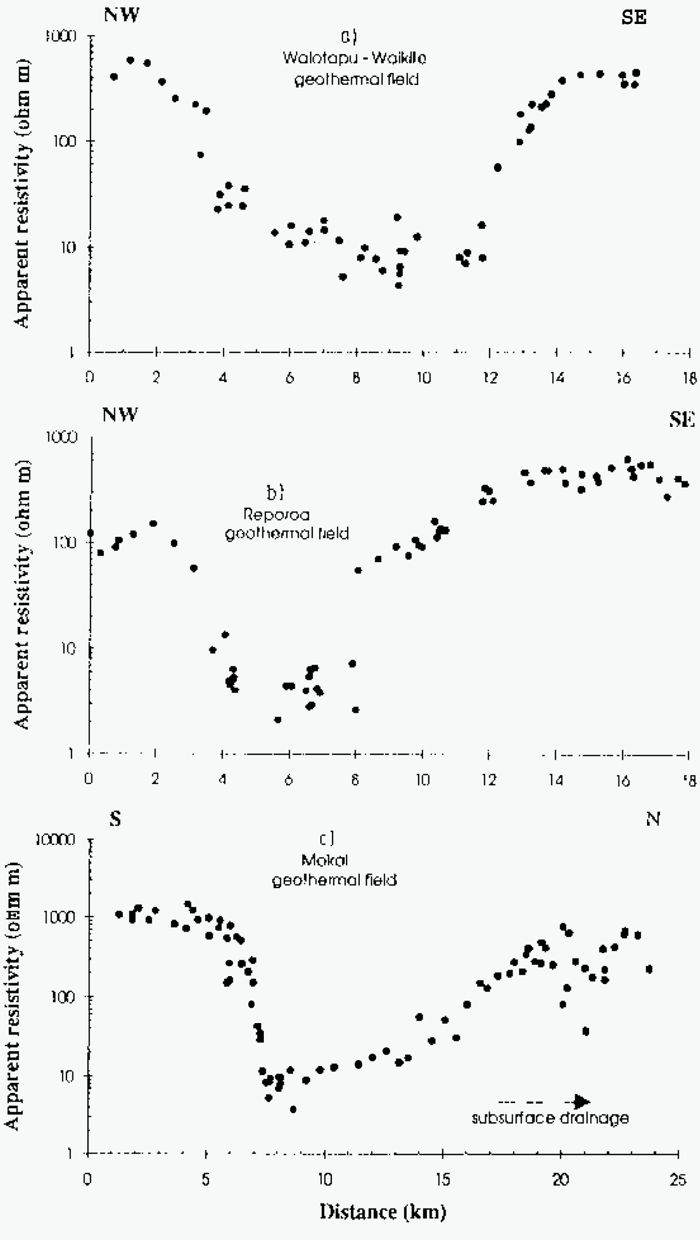

Figure 4. Cross sections showing the sharp changes in measured apparent resistivity that mark the boundaries of the geothermal fields in the Taupo Volcanic Zone.
a) The Waiotapu-Waikite geothermal field.
b) The Reporoa geothermal field.
c) Mokai geothermal field.
At the north side of the Mokai geothermal field the resistivity change is not sharp. In this area subsurface drainage carries the geothermal waters down-slope towards the major river of the area. This subsurface flow of geothermal waters masks the resistivity change associated with the edge of the deep convective plume. The main chloride springs occur some 7 km north of the up-flow zone

along profiles through typical geothermal fields. Three examples (Waiotapu-Waikite, Reporoa and Mokai) are shown in Fig. 4. In each case, the measurements show that the electrical resistivity changes by about an order of magnitude over a horizontal distance ranging between 200 m and 1 km. However, as is shown in Fig 4c, it is unusual for such an abrupt boundary to be found across all parts of the perimeter.

In order to delineate the horizontal extent of any field, it has become accepted practice in New Zealand (Risk et al., 1977) to define a boundary zone separating the geothermal field from its surroundings. The width of the boundary zone reflects the uncertainty in its delineation. It is implicit in the adoption of such a procedure that the boundary between hot and cold water is vertical or near vertical. That is, that the hot geothermal waters rise from depth as a near-vertical plume. Resistivity boundary zones have been determined for most of the TVZ fields being considered for exploitation. Where there are sufficient drillholes to provide a well distributed sample of temperature throughout a geothermal field, the zone where the steep horizontal temperature gradient at depth occurs correlates well with the boundary zone defined by resistivity (e.g. Kawerau - Allis et al., 1993; Ohaaki - Risk, 1983; Wairakei - Risk et al., 1984). Fig. 5 shows a comparison of the resistivity boundary and temperatures for the Wairakei geothermal field.

Apart from the Schlumberger mapping measurements, a variety of deeper-penetration resistivity techniques have also been used to investigate the boundary region (e.g. Risk et al., 1977). Deep-penetration, multiple-source, bipole-dipole measurements have been modelled to ascertain the shape of the hot-cold water boundary at depths down to about 2 km (e.g. Bibby, 1978; Mulyadi & Caldwell, 1979). These studies show that to a depth of 0.5 – 1 km, the interface between hot and cold water can be modelled to a good approximation by a vertical interface. At Ohaaki, Bibby (1978) showed that the interface flares outwards at greater depth. Similar results have been found for other

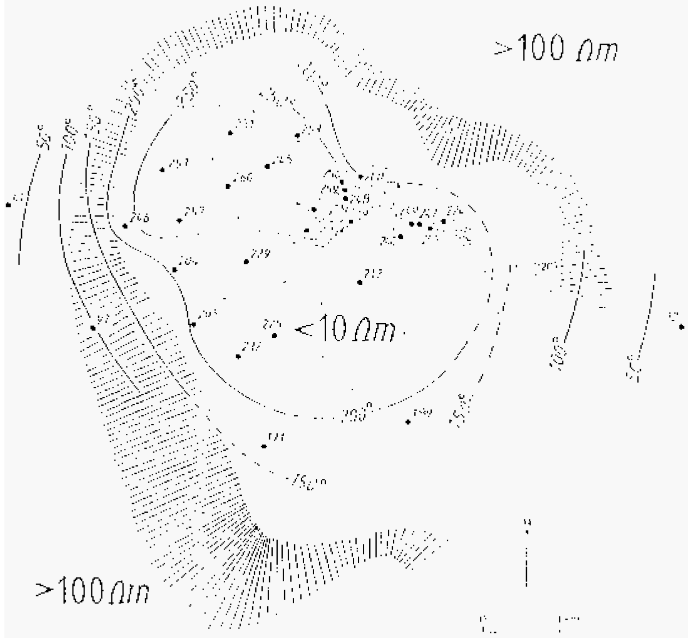

Figure 5. Wairakei geothermal field. The resistivity boundary, shown as the zone of hatching, is compared with the average temperatures measured in drill-holes (dots) between 0.7 and 1.0 km depth. The measurements demonstrate the correlation between the sharp changes in temperature and resistivity at the edge of the geothermal field (after Risk, 1983).

geothermal fields in the TVZ. Thus the areal extent of the hot-water zone seems to increase at greater depth. Bibby(1978) suggested that this increase in the cross-sectional area of the rising plume reflects a decrease in permeability with increasing depth. The cross-sectional area of a plume of constant flux will vary in inverse proportion with the permeability. This explanation is consistent with the simple models of convective plume flows through regions of variable permeability carried out by Wooding (1964).

Surface Drainage of Geothermal Waters

As geothermal waters approach the surface they become entrained into the near-surface hydrological system. Subsurface drainage can result in the geothermal waters reaching the surface a considerable horizontal distance away from the position of the convective plume at depth. Since these geothermal waters are highly conductive, the flow paths will be seen in the resistivity pattern. Such flows can cause ambiguities in the interpretation of resistivity data. The major river systems provide the natural collection points for both the cold groundwaters and the geothermal waters. Thus, where the TVZ geothermal fields occur on or near major rivers, for example Wairakei (Fig. 5), the drainage paths for the geothermal waters are short, and the low-resistivity anomalies associated with the geothermal fields reflect the position of the geothermal plume at depth. Consequently, the resistivity anomalies provide accurate locations for the deeper geothermal plumes.

In contrast, a number of geothermal fields in the TVZ do not occur on the major rivers. For each of these, the electrical resistivity pattern is more diffuse, with the low-resistivity regions reflecting both the plume and the near-surface drainage. The best example of this is the Mokai geothermal field (Bibby et al., 1984) where a tongue of low resistivity extends down the hydraulic gradient towards the Waikato River (Fig 4c). Other outflows can be recognised in the resistivity map (Fig. 3). The concept of a boundary has to be extended in areas where near surface fluid movements occur. In this situation, the boundary zone is used to delineate of the position of the deep geothermal plume, rather than reflecting the extent of the draining geothermal waters (e.g. Bibby et al., 1984).

DISCUSSION

Heat flow

All the known thermal springs of the TVZ have their origins in one or other of the geothermal fields identified by the resistivity surveying. All the heat output of the TVZ is concentrated thorough the geothermal fields. The average area of the resistivity anomalies associated with the 17 high-heat-output geothermal fields is 19 ± 6 km^2, and the average heat output about 250 ± 150 MW(t). The consistency in both the areas and heat outputs of the fields suggests that permeability at depth throughout the whole TVZ is nearly uniform.

Convective plumes and geothermal fields

Both the heat-flow pattern and the nature and distribution of the geothermal fields as revealed by resistivity surveying give strong support for the convective-flow model of the hydrothermal systems of the TVZ. This model of large-scale circulation has been developed over many years. Fundamental works include those of Donaldson (1962), Wooding (1963, 1964), Elder (1965, 1966) and McNabb (1965). High temperatures at deep levels within the TVZ give rise to large-scale convective flows driven by the buoyancy of the heated waters. The large-scale convection patterns require cold groundwater to be drawn down over most of the region. As this water becomes heated it is concentrated into narrow rising plumes. It is the upper portion of these convective plumes, where they interact with the surface hydrology, that are the geothermal fields. The heat output from a geothermal field thus is the heat accumulated over a region at depth that has an area more than 10 times that of the rising plume near the surface (McNabb, 1975). Models of convective flow in a porous media (e.g. Robinson, 1975) show that the ratio of plume spacing to the depth to the base of the plume is about 0.5. This suggests that, in the TVZ, meteoric waters penetrate to a depth of about 8 km. The depth of the seismogenic zone is of a similar order (Bibby et al., in press).

This model of convective heat transport does not require that each geothermal field is associated with a localised magmatic source of heat beneath it. The life span of individual geothermal fields in the TVZ has been estimated to be between 0.2 to 0.5 My (Browne, 1979). A geothermal field can be maintained only by continual replenishment, during its lifetime, of heat to the base of the convective system. In the convective-flow model of hydrothermal systems, heat replenishment may occur anywhere over the base of the convective region, which, in effect, is the total area of the TVZ. Thus, replenishment of heat, by intrusion or other mechanisms, may take place anywhere in the region.

The position of the geothermal fields appears to be stable over time. Any fossil geothermal field that was active in the past would have produced a zone of thermal alteration that remains as evidence of its former existence. If the position of the geothermal fields had changed with time, such residual signatures of the old fields would remain. Thus the lack of widespread evidence of past geothermal activity over most of the TVZ indicates that the circulating fluids have followed preferred paths that have remained fixed. It is still possible that the magnitude of the heat output may have waxed and waned. The sharp interfaces between the hot geothermal waters and the cold surrounding groundwaters also appear to have been maintained over the lifetime of the geothermal systems. Boundary regions with large horizontal temperature gradients, similar to those observed in the TVZ, are characteristic of laboratory models of convective hydrothermal plumes (e.g. Wooding, 1964). The sharp boundaries of the geothermal fields are maintained by the dynamics of the convecting hydrothermal plumes.

Although recirculation within closed convection cells was a feature of many of the early conceptual models of the large-scale hydrothermal systems, there is evidence suggesting that when thermal waters in the rising plume reach the surface, they are mostly discharged to the surface drainage, and only a small fraction joins the down-flowing waters to recirculate around the cell. A measure of the recirculation can be obtained by comparing heat-flow calculations made by physical measurements with those calculated from chemical fluxes. A systematic comparison was made using data from the Wairakei geothermal field (Ellis and Wilson, 1955). From the increase in the chloride flux in the Waikato River past Wairakei and Tauhara geothermal fields, Ellis and Wilson (1955) calculated a total heat discharge of 600 MW(t), of which approximately 180 MW(t) could be attributed to the output of the wells. The total natural heat output of Wairakei and Tauhara, estimated from independent physical measurements, was shown by Allis (1980) to be 460 MW(t). Thus, almost all of the hydrothermal plume is discharged into local surface groundwater, with only a small amount of the fluid becoming cooled sufficiently to be able to recirculate through the hydrothermal convection cell. The systems are essentially 'once through'. The model of non-recirculating hydrothermal flow is now widely accepted and is a tacit assumption made in many of the estimates of the heat output of geothermal fields.

CONCLUSION

The near-surface resistivity distribution within the TVZ is now well known, with only a few areas where data have not been measured. These data provide a detailed picture of the distribution of the geothermal fields, and use of these data has contributed to the very high success rate of exploration drilling. Indeed, all wells drilled within the low-resistivity anomalies have encountered high temperatures (>160°C) within 1 km of the surface.

The resistivity data provide a valuable perspective on the processes involved in heat transport within the TVZ. These data also show that only a small number of extinct geothermal fields are present. The relationship between the location of geothermal fields and possible sources of heat, such as magma bodies at depth, has often been debated. Over the last 0.5 My there has been widespread volcanic activity with evidence of both extrusive and intrusive volcanism. If a geothermal field existed over each intrusive magmatic body it would follow that geothermal fields would be widespread over the TVZ. The small number of fossil fields that have been observed and the long life span of the active geothermal fields suggest that the TVZ hydrothermal systems are essentially stable on a time scale of the order of 200,000 years. A single intrusive event of the size observed at the surface is an order of magnitude too small to supply the total energy discharged by a typical TVZ geothermal field.

The stability in position and the long life of the geothermal fields provide further support for the model of large-scale hydrothermal convection throughout most of the TVZ. Continuing volcanic activity and deep intrusions replenish the heat supply to the base of the convective region but do not alter the paths taken by the convecting fluids. However, variations in the heat supply may influence the rate at which heat is discharged at any one field. Large-scale convection systems have considerable inertia, which appears to stabilise the convective flows despite intermittent disruption by volcanism.

REFERENCES

Allis, R.G. (1980). Changes in heat flow associated with exploitation of Wairakei geothermal field, New Zealand. *N.Z. J. Geol. Geophys.*, 24, pp. 1-19.

Allis, R.G., Christenson, B.W., Nairn, I.A., Risk G.F. and White, S.P. (1993). The natural state of Kawerau geothermal field. *Proc. 15th N.Z. Geothermal Workshop*, Auckland, pp. 227-233.

Banwell, C.J. and Macdonald, W.J.P. (1965). Resistivity surveying in New Zealand thermal areas. *Proc. 8th Commonwealth Mining and Metallurgical Congress*, Australia and New Zealand, paper 213, pp. 1-7.

Banwell C.J., Cooper, E.R., Thompson, G.E.K. and McCree, K.J. (1957). *Physics of the New Zealand thermal region.* Bulletin 123, New Zealand Department of Scientific and Industrial Research. Wellington, 109pp.

Bennie, S.L. (1983). *Geophysical investigations of the Ngatamariki geothermal area.* Report 192. Geophysics Division, D.S.I.R., Wellington, New Zealand.

Bibby, H.M. (1978). Direct current resistivity modeling for axially symmetric bodies using the finite element method. *Geophysics*, 43, pp. 550-562.

Bibby, H.M. (1988). Electrical resistivity mapping in the Central Volcanic Region of New Zealand. *N.Z. J. Geol. Geophys.*, 31, pp. 259-274.

Bibby, H.M., Dawson, G.B., Rayner, H.H., Stagpoole, V.M. and Graham, D.J. (1984). The structure of Mokai geothermal field based on geophysical observations. *J. Volcan. Geotherm. Res.*, 20, pp. 1-20.

Bibby, H.M., Dawson, G.B., Rayner, H.H., Bennie, S.L. and Bromley, C.J. (1992). Electrical resistivity and magnetic investigations of the geothermal systems in the Rotorua area, New Zealand. *Geothermics* 21, pp. 43-64.

Bibby, H.M., Bennie, S.L., Graham, D.J. and Rayner, H.H. (1993). Electrical resistivity map of New Zealand: 1:50000, Sheet V16 - Rotorua. Institute of Geological & Nuclear Sciences geophysical map: 1.

Bibby H.M. Caldwell, T.G., Davey, F.J., and Webb, T.H. (in press). Geophysical evidence on the structure of the Taupo Volcanic Zone and its hydrothermal circulation. *J. Volcan. Geotherm. Res.*

Browne, P.R.L. (1979). Minimum age of the Kawerau geothermal field, North Island, New Zealand. *J. Volcan. Geotherm. Res.*, 6, pp. 213-215.

Caldwell, T.G., Pearson, C. and Zayadi, H. (1986). Resistivity of rocks in geothermal systems: A laboratory study. *Proc. 8th N.Z. Geothermal Workshop 1986*, pp. 227-281. University of Auckland Geothermal Institute.

Caldwell, T.G. and Bibby, H.M. (1992). Geothermal implications of resistivity mapping on Lake Taupo. *Proc. 14th N.Z. Geothermal Workshop 1992*, pp. 207-212, University of Auckland.

Donaldson, I.G. (1962). Temperature gradients in the upper layers of the earth's crust due to convective water flows. *J. Geophys. Res.*, 67, pp. 3449-3460.

Elder J.W. (1965). Physical processes in geothermal areas. In: *Terrestrial heat flow,* W.H.K. Lee (Ed). Geophysical monograph 8, American Geophysical Union, pp. 211-239.

Elder J.W. (1966). *Heat and mass transfer in the earth, hydrothermal systems.* Department of Scientific and Industrial Research Bulletin 169, 115pp.

Ellis, A.J. and Wilson, S.H. (1955). The heat from the Wairakei-Taupo thermal region calculated from the chloride output. *N.Z. J. Sci. Tech.*, 36, pp. 662-631.

Ellis, A.J. and Mahon, W.A.J. (1977). *Chemistry of geothermal systems.* Academic Press, London, 392pp.

Geophysics Division (1985). Sheet U17 - Wairakei. Electrical resistivity map of New Zealand 1:50 000. Wellington, New Zealand, Department of Scientific and Industrial Research.

Hatherton, T.; Macdonald, W.J.P.; Thompson, G.E.K. (1966). Geophysical methods in geothermal prospecting in New Zealand. *Bulletin volcanologique* 29: 484-498.

Henneberger, R.C. (1986). Ohakuri fossil epithermal system. In: *Guide to the active epithermal (geothermal) systems and precious metal deposits of New Zealand,* Henley, R.W., Hedenquist, J.W., and Roberts P.J. (editors). Monograph series on Mineral Deposits 26, pp. 121-128. Gebruder Borntraeger, Berlin.

Hochstein M.P. (1991). Geothermal prospects of Sumatra (overview). *Proc. 13th Geothermal Workshop 1991*, pp. 219-222. University of Auckland.

Hochstein M.P., Smith, I.E.M., Regenauer-Lieb, K. and Ehara, S. (1993). Geochemistry and heat transfer Processes in Quaternary rhyolitic systems of the Taupo Volcanic Zone, New Zealand. *Tectonophysics*, 223, pp. 213-235.

Macdonald, W.J.P. (1967). *A resistivity survey of the Taupo - Waiotapu area at fixed spacing (1800 ft).* Department of Scientific and Industrial Research, Geophysics Division Report 46, Wellington, New Zealand.

McNabb, A. (1965). On convection in a porous medium. *Proc. 2nd Australian conference on hydraulics and fluid mechanics*, C161-171.

McNabb, A. (1975). *Geothermal physics.* Department of Scientific and Industrial Research, Applied Mathematics Technical Report 32. Wellington, New Zealand. 37pp.

Mulyadi, A. and Caldwell, G. (1979). Analysis of the resistivity boundary structure of the Broadlands geothermal field. *Proc. N.Z. Geothermal Workshop 1979*, pp. 298-306, University of Auckland.

Olhoeft, G.R. (1981). Electrical properties of rock. In: *Physical properties of rocks and minerals,* Y.S. Touloukian, W.R. Judd, and R.F. Roy (*Editors*) pp. 257-330. McGraw-Hill, New York.

Quist, A.S. and Marshall, W.L. (1968). Electrical conductance of aqueous sodium chloride solution from 0 to 800°C at pressure to 400 bars. *Phys. Chem.* 72, pp. 684-703.

Risk, G.F. (1983). Delineation of geothermal fields in New Zealand using electrical resistivity prospecting. *Proc. 3rd biennial conference of the Australian Society of Exploration Geophysicists*, pp. 147-149.

Risk, G.F., Groth, M.J., Rayner, H.H., Dawson, G.B., Bibby, H.M., Macdonald, W.J.P. and Hewson, C.A.Y. (1977). *The resistivity boundary of the Broadlands geothermal field.* Report 123, Geophysics Division, Department of Scientific and Industrial Research, Wellington.

Risk, G.F., Rayner, H.H., Stagpoole, V.M., Graham, D.J., Dawson, G.B., and Bennie, S.L. (1984). *Electrical resistivity survey of the Wairakei geothermal field.* Report 200, Geophysics Division, D.S.I.R., Wellington, New Zealand.

Risk, G.F., Bibby, H.M. and Caldwell, T.G. (1993). DC resistivity mapping with the multiple-source, bipole-dipole array in the Central Volcanic Region, New Zealand. *J. Geomag. Geoelectr.* 45, pp. 897-913.

Robinson, J. (1975). *Modelling of convecting fluids.* Department of Scientific and Industrial Research, Applied Mathematics Technical Report 35. Wellington, New Zealand. 32pp.

Studt, F.E. (1958). Geophysics reconnaissance at Kawerau, New Zealand. *N.Z. J. Geol. Geophys.*, 1, pp. 219-246.

Studt, F.E. and Thompson G.E.K. (1969). Geothermal heat flow in the North Island of New Zealand. *N.Z. J. Geol. Geophys.*, 12, pp. 673-683.

Whiteford, P.C. (1992). Heat-flow measurements in the sediments of Lake Rotorua. *Geothermics*, 21, pp. 75-88.

Wilson C.J.N., Houghton, B.F., McWilliams, M.O., Lanphere, M.A., Weaver, S.D., and Briggs, R.M. (in press). Volcanic and structural evolution of Taupo Volcanic Zone, New Zealand: a review and synthesis. *J. Volcan. Geotherm. Res.*

Wooding, R.A. (1963). Convection in a saturated porous medium at large Rayleigh number or Peclet number. *J. Fluid Mechanics*, 15, pp. 527-544.

Wooding, R.A. (1964). Mixing-layer flows in a saturated porous medium. *J. Fluid Mechanics* 19, pp 103-113.