Proceedings 18th NZ Geothermal Workshop 1996

LARGE SCALE NUMERICAL MODELS OF TVZ-TYPE GEOTHERMAL FIELDS

W.M. KISSLING

N.Z. Institute for Industrial Research and Development
Lower Hutt, NZ

SUMMARY – Several models for the large scale structure of a typical Taupo Volcanic Zone, New Zealand, geothermal field are presented. In these models, 'magmatic' water is injected over a region at 8 km depth at a rate of 20 kg/s, and heat is supplied to the same region at a rate of 400 MW. Meteoric water is entrained to form a buoyant plume which extends to the surface. The mass output, mean enthalpy, ratio of magmatic/meteoric waters and the outflow area of this plume provide constraints on both the size of the heat source and the large scale permeability distribution. The most satisfactory models have low vertical permeability (5-10 millidarcy) and small source regions (< 2 km radius). In most cases, these systems reach a steady state 10 - 40 thousand years after the heat source is turned on. In some models however, rather chaotic behaviour is seen, with the thermal output at the surface varying by a factor up to 3 on a timescale of a few thousand years.

1 INTRODUCTION

There has been much debate concerning the nature of the heat source which underlies the geothermal fields in the Taupo Volcanic Zone (TVZ), in the central North Island of New Zealand. Probably the most commonly held idea is that of an isolated magmatic body, or pluton, which is emplaced at depth and is cooled by circulation of the overlying groundwater. Cathles (1977) has carried out an extensive study of such systems, and finds that a single magmatic intrusion is not sufficient to supply the heat for a Wairakei-type geothermal system, suggesting that continuous replenishment must be taking place.

An alternative model is proposed by McNabb (1992), in which a saline 'hot plate' forms at depth as a consequence of the phase behaviour of high temperature NaCl - H_2O mixtures. This hot plate is widespread over the TVZ, providing the heat source for a TVZ - wide convective system. The regions where high heat and mass flows extend to the surface are the geothermal fields observed in the TVZ. This model naturally explains the characteristic 10 - 15 km spacing of the TVZ geothermal fields, and the cold downflows that are seen outside of the high heat flow regions (Studt and Thompson, 1969).

It is generally agreed (Bibby *et al.,* 1995) that the TVZ geothermal systems are the result of large scale convective flows. There have been a number of studies of such flows in the context of the TVZ. For example, Wooding (1978) considered the influence of the large scale permeability structure on convection above a widespread magma body, and found that a low Nusselt number and a three kilometre deep convecting layer could explain the observed heat flows in the TVZ. This and other similar work suffers from the defect that the temperature and pressure dependence of the thermodynamic properties of water are not fully accounted for. Straus and Schubert (1977) have remedied this and show that the onset of convection occurs for Rayleigh numbers much lower than would otherwise be expected.

Apart from the work of Cathles (1977) mentioned above, it appears that there has been little use of numerical simulation models to explore the properties of large scale TVZ-like convective systems. To do this, the models must extend to a depth of several kilometres. Such great depths of course imply high temperatures and pressures, and for this reason a 'supercritical' simulator must be used. Recently, several such codes have been developed (Hanano and Seth (1995), Hayba and Ingebritsen (1994), Kissling (1995)). In this paper the last of these codes is used to model a variety of large scale convective systems. The aim is to examine how the permeability structure and size of the source region influence these convective systems, and then draw some conclusions about the TVZ geothermal fields in light of this information.

2 THE TVZ GEOTHERMAL FIELDS

Any satisfactory model must be consistent with the available data, and the work presented here relies heavily on many fundamental observations concerning the TVZ geothermal fields. A good review of relevant geophysical results is given by Bibby *et al.* (1995), and Giggenbach (1995) has similarly given an extensive summary of TVZ geochemistry. Listed below are those results most pertinent to the present study. These are given without reference, and the reader is referred to the papers of Bibby *et al.* and Giggenbach for further details.

1. The typical spacing of the TVZ geothermal fields is 10 - 15 km, and is generally believed to be the result of a large scale convective system. The depth of the magmatic source is in the range **5** - 10 km.
2. The natural state heat flow from many geothermal fields is estimated to be in the range 400 - 600 MW, with **a** typical enthalpy of ~ 1000 KJ/kg.
3. The outflow region of a typical TVZ geothermal field is 10 - 25 km^2, and cold downflows are observed exterior to this.
4. In the shallow regions **(< 2 - 3 km)**, the horizontal permeability is greater than the vertical permeability. Representative values are $k_h = 5 \times 10^{-14}$ m^2 and $k_v = 5 \times 10^{-15}$ m^2. These values are understood to be representative of a fractured rock matrix.
5. The ratio of magmatic/meteoric waters is ~ 6±2 % for Western TVZ geothermal fields (Wairakei, Mokai and Waiotapu), and ~ 1435% for Eastern TVZ fields (Ohaaki, Rotokawa, Kawerau).

3 MODEL DESCRIPTION

The model that is developed in this paper is based on the simulator TOUGH2 (Pruess, 1991). **A** number of specialised Equation of State **(EOS)** modules are available for this code. In the present study, use has been made of two **EOS** modules. The first allows the modelling of flows containing dissolved NaCl and **air,** these being used **as** 'tracers' for the magmatic and meteoric waters respectively. Second, in some examples where the temperature is greater than **350°C** an EOS module for super-critical water (Kissling, 1995) has been used. This code can also be used with a tracer to follow the flow of meteoric and magmatic waters. To ensure compatability **of** the results, tests were performed which show that these two **EOS** modules produce the same results when run on identical problems.

The computational grid for the model is shown in Figure 1. The outer boundary is placed at **a** distance of 8 **km** to provide for an adequate inflow of meteoric water, and to be consistent with the typical spacing of 10 - 15 km seen for the TVZ geothermal fields. The 'magmatic' source is placed at a depth of 8 km, consistent with geophysical evidence. It was not considered practical to perform full **3-D** simulations of such **a** large region with the necessary rcsolution, and for this reason a cylindrical geometry has been adopted. The model has 961 **(31 x 3** 1) elements, which are graded in size to provide highest resolution near the r = 0 axis. The smallest elements are about 150 m across, the largest 300 m.

The boundary conditions for the model are also shown on Figure 1. In all cases, the source has been placed at a depth of 8 km, supplies.20 kg/s of water, 1 kg/s of 'tracer', and 400 MW of heat. These figures correspond approximately to those for the Wairakei system (McNabb, 1992), which we take as **a** typical TVZ system. The greatest depth to which meteoric water is permitted to circulate is also 8 km, and so no fluid enters the system from below the lower boundary in the region exterior to the source. Hydrostatic pressure, with the temperature fixed at **15°C,** is prescribed at the radial (r = 8 km) boundary, corresponding to observed conditions in the TVZ outside the geothermal areas. At the near-surface boundary (z = -200 m) constant pressure and temperature are prescribed. This provides for both the outflow of hot geothermal fluid, and the inflow of cold meteoric water. The initial conditions for the model are the same 'isothermalhydrostatic' conditions that are applied at the radial boundary.

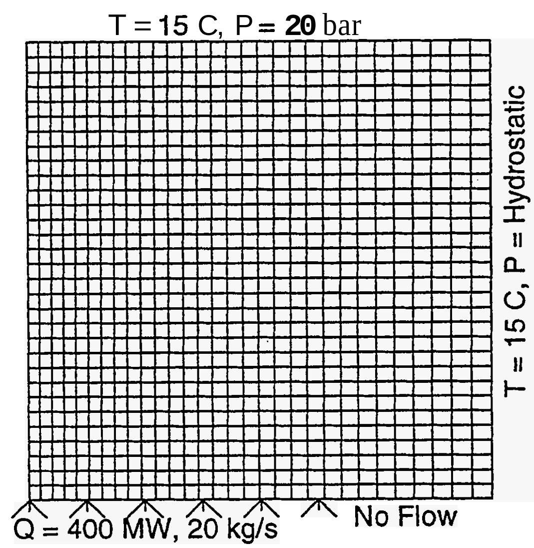

Figure 1: The computational mesh and boundary conditions of the Model. The region shown **is** 8 km deep **x** 8 km in the radial direction, and the radius of the source region is 5 km.

In this paper the main focus is on how the permeability structure influences the geothermal fields. To do this, four cases have been investigated, and these are listed in Table 1. All models presented have an homogeneous structure with the same horizontal permeability. Model **A** has a vertical permeability which is twice the horizontal permeability. In models B, C and D the vertical permeabilities are reduced by factors of 2, 10 and 20 respectively from that in model **A.** Model B is an homogeneous, isotropic system. The porosity in all models is set at **a** constant value of 0.2.

To investigate the effect of source size the inflow is distributed over radii ranging from 1 km to 8 km, for each permeability structure given in Table 1. The area of the source region, and hence the source strength per unit area, changes by nearly two orders **of** magnitude between these extremes.

In the discussion that follows particular models will be referred to in the form 'XR' where X represents Model **A,** B, C or D and R is the radius of the source region in km. For example, model B2 has an homogeneous, isotropic permeability structure with a source region of radius 2 km. For this purpose the source radii are given nominal values of 1, 2, 5 and 8 km for the various models, although the actual source radii differ slightly from these values.

Model	k_h (m^2)	k_v (m^2)	ϕ
A	5×10^{-14}	1×10^{-13}	0.2
B	5×10^{-14}	5×10^{-14}	0.2
C	5×10^{-14}	1×10^{-14}	0.2
D	5×10^{-14}	5×10^{-15}	0.2

Table 1: Permeability and porosity for the three models. k_h and k_v refer to the horizontal and vertical permeabilities, and ϕ is the porosity.

4 RESULTS AND DISCUSSION

In this section **a** comparison is made between the modelled geothermal plumes and the real geothermal fields of the TVZ. To do this, various features of the models will be considered in relation to the points noted in section 2 of this paper. First of all however, in section **4.1** the results from a single model are examined more closely. This will give some idea of the general behaviour of the models and make the following discussion more meaningful. In the second part of the discussion (4.2) the issues of permeability structure and source size are addressed.

4.1 A Typical Model

The essential process which occurs in all **of** the models presented here is the entrainment of cool meteoric water into the source region to form a buoyant column of hot fluid. This 'plume' extends to the surface, and is associated with a single TVZ geothermal field. For the purposes of this work, there is no 'outflow zone' - the geothermal fluid merely flows out of the system through the upper boundary. In a real field of course, the geothermal fluid will interact with the shallow groundwater and flow horizontally away from the upflow area. The models described here can be imagined as representing a real system in which the upper boundary is place below the outflow zone.

For this part of the discussion, model C5 has been chosen to show some typical results. The model was run for a period of approximately 120000 years. Figure **2** shows the mass and heat flowrates through the surface as a function of time. Following 'turn on' of the heat source, the mass flow rises to a peak value of over 1000kg/s before subsiding to a steady value of **530** kg/s. Similarly, the heat flow reaches about 700 MW before stabilising at 480 MW. The 'extra' 80 MW ovcr that supplied at the source comes from the enthalpy of the (cold) meteoric mater, and is not usable heat. The peak values are reached in about 13000 years, but steady conditions are not reached until about 40000 years after turn on. This initial transient is due to the displacement of overlying, initially stationary, cold water, and is a feature of all the models. The average enthalpy of the outflow fluid (heat flow divided by mass flow) increases from an initial 'cold' value of 65 KJ/kg to a final value of 905 KJ/kg, showing that it contains a steadily increasing fraction of hot fluid.

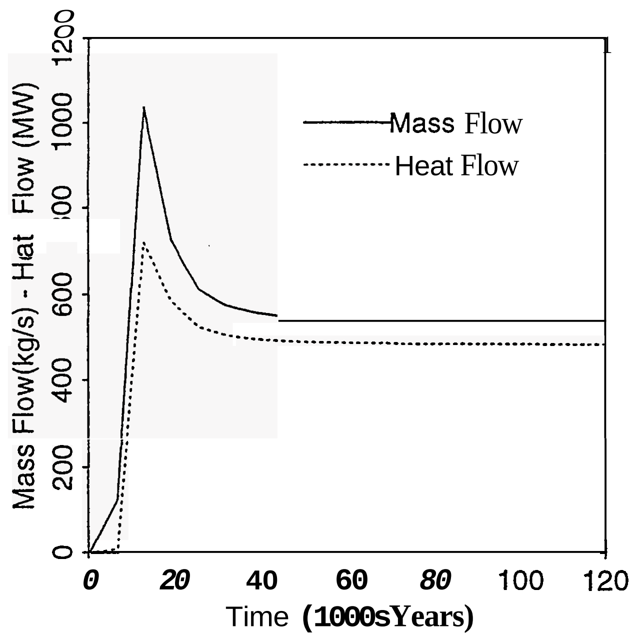

Figure 2: Mass and heat flowrates versus time for Model C5.

Figure **3** shows the **mass** flux vectors for model C5 at a time of 60000 years. There are several features of interest. Firstly, there is a large horizontal flow induced close to the heat source. This is due to the reduced density of the (hot) injected fluid, which results in a lower pressure region about the source. **This** flow extends to the boundary at 8 km radius. In this particular model, nearly all of the meteoric water enters the system through this side boundary, mainly below about 4 km depth. Other models have much greater inflow through the top boundary. The entrainment of the meteoric water into the geothermal plume is essentially complete at a depth of 7 km - 1 km above the heat source. Beyond this point very little mixing takes place, and the plume is almost isothermal in the vertical direction, with an almost fixed temperature profile in the radial direction (Figure 4).

Temperature contours for model C5 at 60000 years are shown in Figure 4. The maximum temperature for this model is 347°C. The entrainment **of** fluid at depth is obvious, with higher temperatures extending over the source region. The nearly isothermal nature of the plume is also evident, although more widely spaced contours at the edge of the plume show slight mixing is occuring there. In models with lower vertical permeability, there tends to be a larger vertical temperature gradient. Some interaction with cool fluids entering the top boundary can be seen at the upper boundary near the outer edge of the plume.

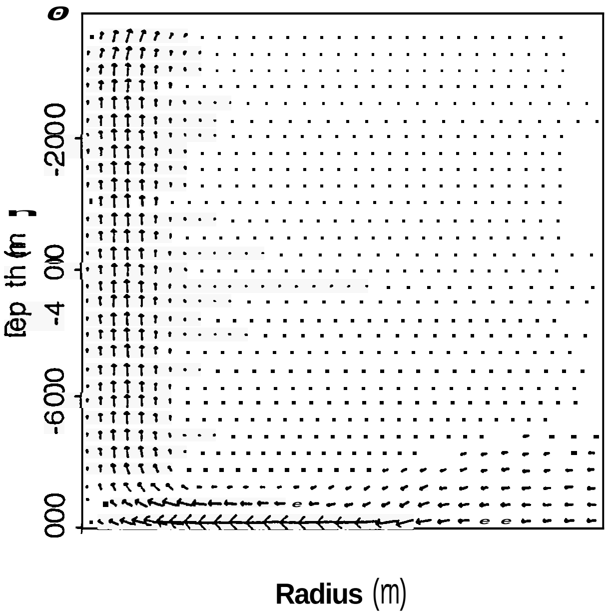

Figure 3: Mass flux vectors for model C5 at 60000 years.

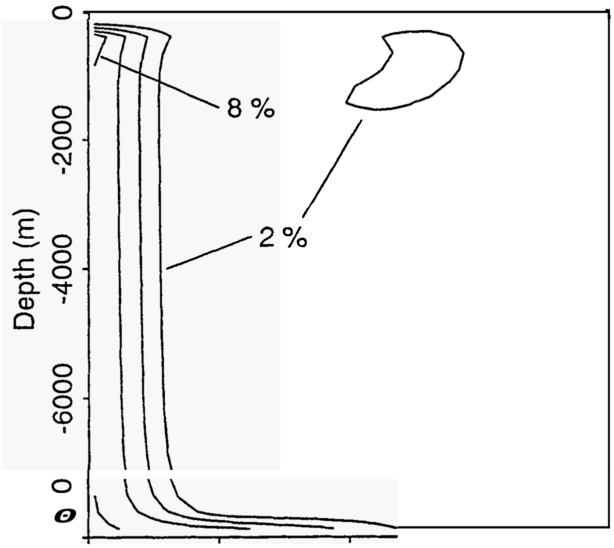

Figure 5: Contours of magmatic fraction (expressed as %) for model C5 at 60000 years. The contour interval is 2% magmatic fraction.

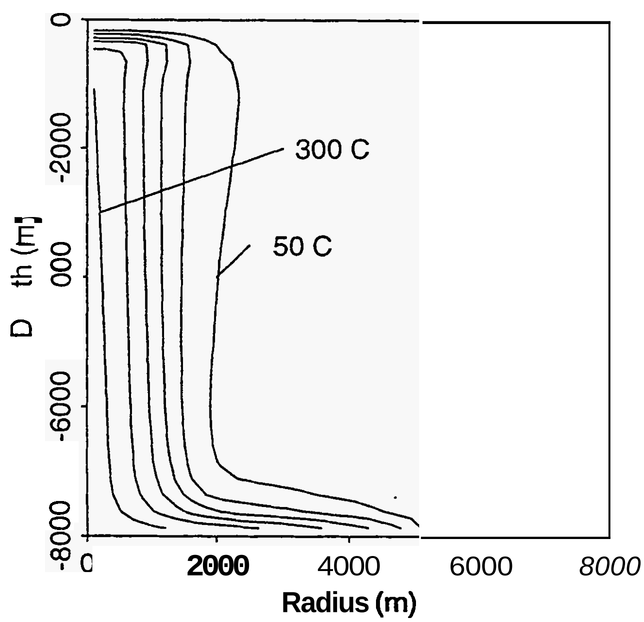

Figure 4: Temperature contours for model C5 at 60000 years. The contour interval is 50°C.

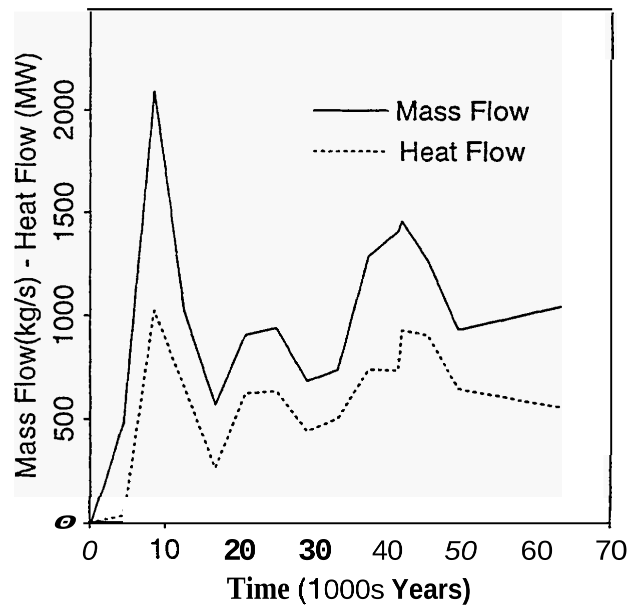

Figure 6: Mass and heat flow from the surface for model A5 for 65000 years. Following the initial transient, the mass and heat flows vary irregularly with a timescale of a few thousand years.

Figure 5 shows contours of the proportion of magmatic fluid for model C5 at 60000 years. The maximum value in this example is about 8%, although the average over the entire plume is about 4%. The 'pocket' of > 2% magmatic fluid at 5 km radius and 1 km depth is a transient feature left behind from the initial development of the plume. The magmatic fraction, like the temperature, depends strongly on the distance from the centre of the plume, but tends to be almost constant in the vertical direction. In some models boiling occurs in the upper plume, and gas components (particularly CO_2) and chlorides (which remain in the liquid phase) will be important in future, more detailed, modelling.

Most of the models reach a steady state similar to C5 in periods of 10-40000 years. In some cases however, this does not happen, and the mass and heat outputs are highly erratic on timescales of a few thousand years. Figure 6 shows the mass and heat outflows for model A5. This behaviour is observed for four models (A5, A8, B5 and B8), and has been reproduced with different TOUGH2 meshes and with different equation of state modules. These four models are characterised by relatively high vertical permeabilities and large source regions, and so give low enthalpy fluid outflows. At present the reason for this phenomenon is not clear, but because these particular models probably have little relevance to the TVZ geothermal fields (because of their low enthalpy), the subject will not be pursued any further in this paper.

4.2 Permeability Structure and Source Size

The main results of this study are summarised in Tables 2 - 5. These list the particular features of each of the models which are to be compared with observed properties of the TVZ geothermal fields. In order of appearence these are the rate of mass outflow at the surface, the mean enthalpy of this fluid, the proportion of magmatic or source fluid and the cross sectional area of the outflow.

Table 2 shows the mass flow rates from the surface for each of the models, and for each of the different source radii. The entry for model D 1 is missing because the temperature at the base of the plume exceeds 800°C, which is the upper limit for the 'super-critical' version of TOUGH2. There is a general trend to smaller flowrate as the source region decreases in size, and flowrates also reduce as the vertical permeability is decreased. Most of the figures are comparable to TVZ values, typically 350 - 700 kg/s. In addition to being highly variable (see Figure 6), models A5, **A8, B5** and B8 have peak flow rates that are very large, and for these reasons these models are not viable candidates for TVZ geothermal fields.

	Source Radius (km)			
Model	1.0	2.0	5.1	8.0
A	517	577	500-800	700-2000
B	451	517	700-1500	700-2000
C	348	398	530	680
D	-	351	490	650

Table 2: Mass flow rates (kg/s) through the surface for various models.

Table 3 shows the average enthalpy of the fluid output for each of the models, again for different source areas. The average enthalpy decreases as the source increases in size, and increases as vertical permeability is reduced. The two models with the lowest **(A8)** and highest (C1) enthalpies have central source temperatures of 186°C and 567°C respectively. For models C1 and C8, the base temperatures are 567°C and 287°C while the temperature at the upper boundary is 240°C in both cases. The far greater enthalpy for model C1 is due to shallow boiling in this model. For the TVZ fields, enthalpies in the range > 900 KJ/kg (about 210°C for pure water) are considered satisfactory, and so these results seem to exclude all of the 8 km source models. The high vertical permeability, small source models A1 and **A2** are also marginally excluded.

	Source Radius (km)			
Model	1.0	2.0	5.1	8.0
A	894	812	300-600	200-600
B	1023	908	400-600	250-600
C	1311	1163	905	730
D	-	1306	980	770

Table 3: Average enthalpy (KJ/kg) of outflow fluid for each model.

Table 4 gives values for the ratio of magmatic/meteoric water in the topmost single phase fluid in the plume. This has been calculated by taking the ratio of the 'magmatic' fluid input (21 kg/s) to the total fluid output, as listed in Table 1. The ratios depend strongly on position, and are typically twice the tabulated values near the centre of the plume. Most of the models are reasonably successful in this respect, giving 4-6 % magmatic fluid, which is in good agreement with measured values in the Western TVZ fields.

	Source Radius (km)			
Model	1.0	2.0	5.1	8.0
A	4.1	3.6	2 - 4	1 - 4
B	4.7	4.1	1 - 3	1 - 4
C	6.0	5.3	4.0	3.1
D	-	6.0	4.3	3.2

Table 4: Ratios (expressed as %) of magmatic/meteoric water in the geothermal plumes.

Next, we consider the outflow area of the computed geothermal fields, as shown in Table 5. This is taken to be the area over which there is an upflow through the upper boundary of the model. Taking the observed values of 10 - 25 km^2 in the TVZ, the two remaining models with source radii of 5 km are excluded, as are the A and B models with sources less than or equal to 2 km radius. Model D2 is a marginal case, but is probably acceptable given the wide range and inherent uncertainties in the measured outflow areas.

	Source Radius (km)			
Model	1.0	2.0	5.1	8.0
A	2.1	2.1	12-32	40-200
B	3.4	3.4	12-27	27-200
C	9.2	12	50	200
D	-	27	200	200

Table 5: Area of outflow (km^2) of the modelled geothermal fields.

There remain three models, namely C1, C2 and D2, which satisfy the conditions that have been imposed. Models with a small source region (< 12 km^2) and low vertical permeability (5 - 10 md) are favoured. These permeabilities are close to those derived for the upper regions of Wairakei (McNabb, 1992) and Mokai (Grant, 1983). It is perhaps only surprising that these values seem to be equally applicable at much greater depths.

Even more surprising is the apparent success of models with such small source regions. At first sight this seems to favour the localised cooling pluton models over the hot plate type models. However, it must be remembered that very simple permeabilty structures have been assumed in this work. In practice the permeability might be expected to decrease with depth, and to be anisotropic, and it remains to be seen if this result is robust even in that situation.

The highest temperatures in these models occur near the source region and are in the range 450°C to 570°C. This is somewhat less than the temperature expected for a magmatic source, which is probably at least 800°C. In this paper this difficult region has been ignored, and much more detailed chemical modelling under a variety of conditions will be needed to determine exactly what happens in this region.

5 CONCLUSIONS

In this paper, some models are developed to explore the large scale structure of the TVZ geothermal fields. Various models are examined to find the effect of varying the size of the source region and the permeability structure. No assumption has been made about the mechanism by which the magmatic fluid enters the system, the mere fact that it does is sufficient to draw some conclusions about the large scale structure. The model could therefore be equally applicable to 'cooling pluton' models or the 'hot plate' model of McNabb. Summarised below are the main results of this paper:

1. A number of different models for the large-scale structure of the TVZ geothermal fields have been investigated. These models have a fixed source depth of 8 km, and a fixed horizontal permeability of 50 md. The magmatic input is 21 kg/s of mass and 400 MW of heat. The models were run to a steady state where possible. The mass output, average fluid enthalpy, magmatic fluid fraction and outflow area are compared to observed values for TVZ geothermal fields.
2. The models which are most successful in reproducing the observed features of the TVZ geothermal fields have low vertical permeability (5-10 md) and small source regions (< 12 km^2). The highest temperatures, found in these models occur at the source and are in the range 450°C - 570°C.
3. If a steady final state is reached, this occurs in times of order 10000 - 40000 years.
4. Some 'largesource' models produce unsteady final states. The resulting geothermal fields are dynamic features, with variations in thermal output of a factor of 3 or more occuring on time scales of $\sim$ 5000 - 10000 years.

The results in this paper have been obtained with the magmatic source fixed at a depth of 8 km, and with the assumption that the rock properties (including the permeability) remain constant with depth. A further assumption is that of cylindrical geometry. In future modelling, a wide range of source depths (perhaps 5 - 10 km), with more general rock properties, fluid chemistry and with further data (for example temperature profiles) will be investigated in a 3 dimensional setting in the hope of identifying the most realistic models for the large scale structure of the TVZ geothcrmal fields.

6 ACKNOWLEDGEMENTS

The author would like to thank the New Zealand Foundation for Research, Science and Technology for funding this research. Thanks are also due to Steve White for helpful comments on the manuscript.

7 REFERENCES

Bibby, H.M., Caldwell, T.G., Davey, F.J. and Webb, T.H. (1995). Geophysical evidence on the structure of the Taupo Volcanic Zone and its hydrothermal circulation. ***J. Volcan. Geotherm. Res.*** 68, 29-58.

Cathles, L.M. (1977). An Analysis of the Cooling of Intrusives by Ground-Water Convection which Includes Boiling. ***Econ. Geol.*, 72(5)**, 804-826.

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

Giggenbach, W.F., (1995). Variations in the chemical and isotopic composition of the fluids discharged from the Taupo Volcanic Zone, New Zealand. ***J. Volcan. Geotherm. Res. 68,*** 89-116.

Grant, M.A. (1983). Mokai Reservoir Assessment. Applied Mathematics Division, DSIR, *Report 112.*

Kissling, W.M. (1995). Extending MULKOM to Super-Critical Temperatures and Pressures. ***Proc. World Geothermal Congress, Florence,*** 1687-1690.

Hanano, M. and Seth, M.S. (1995). Numerical Modelling of Hydrothermal Convection Systems Including Super-Critical Fluid. ***Proc. World Geothermal Congress, Florence,*** 1681-1636.

Hayba, D.O. and Ingebritsen, S.E. (1994). Flow Near the Critical Point: Examination of Some Pressure-Enthalpy Paths. ***Proc. 19th Stanford Workshop Geothermal Reservoir Engineering, Stanford,*** 83-89.

McNabb, A., (1992). The Taupo-Rotorua Hotplate. ***Proc. 14th New Zealand Geothermal Workshop,*** 111- 114.

McNabb, A., White, S.P. and McKibbin, R. (1993). Brine Circulation and Hot-Plate Formation. ***Proc. 15th New Zealand Geothermal Workshop,*** 275-277.

Pruess, K. (1982). Development of the General Purpose Simulator MULKOM, ***Report LBL-15500, Lawrence Berkeley Laboratory.***

Pruess, K. (1991) TOUGH2 - A general-purpose numerical simulator for multiphase fluid and heat flow ***Report LBL-29400,*** Lawrence Berkeley Laboratory.

Straus, J.M. and Schubert, G. (1977). Thermal Convection of Water in a Porous Medium: Effects of Temperature and Pressure Dependent Thermodynamic and Transport Properties. ***J. Geophys. Res.* 82(2)**, 325-333.

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(4)**, 673-683.