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# THE TAUPO-ROTORUA HOT-PLATE

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**ABSTRACT - A** qualitative model of the hydrothermal systems in the Taupo Volcanic Zone originating from a common hot-plate is presented and tested for viability against data from the Wairakei Geothermal Field and various available geophysical and chemical measurements. The concept of a deep dense stably-stratified hot brine layer forming the hot plate is presented as an inevitable consequence of the phase properties of the $H_2O$-NaCl system at high pressures and temperatures.

## 1. INTRODUCTION

The central North Island volcanic zone contains about fifteen geothermal fields lying in a thirty kilometre wide and one hundred kilometre long strip stretching from Turangi to Kawerau. Detailed studies at Wairakei, Broadlands and Kawerau reveal these structures to be buoyant plumes of hot chloride water rising in cold ground-water. They have a cross-sectional area of 15 to 20 square kilometres, a spacing between plumes of about 15 kilometres, a plume temperature beneath a superficial boiling zone of about 250 C and a sodium chloride content about one tenth that of sea water. These features suggest the whole structure is a system of convective plumes rising from one gigantic hot-plate about 30 kilometres wide, stretching from Taupo to Rotorua and bounded on the south-east by the Kaingaroa Plateau. We are concerned in these notes with the size, depth and nature of this hot-plate and in particular, with the role of chlorides in controlling its temperature and heat output.

## 2. DIMENSIONS OF HOT-PLATE

The lateral dimensions of this thermal activity are outlined in Healy's map (1964) of volcanic vents and hot springs in the Taupo Volcanic Zone (Figure 1), and also by the magnetic anomalies (Gerard & Lawrie, 1955) generated by the rhyolitic domes which have risen off this hot-plate in the last million years. Studt & Thompson (1969) found anomalously low natural thermal gradients in the cold ground-water throughout the same zone and essentially produced a map of the down-flow recharge region over the hot-plate.

The depth to the hot-plate is more difficult to ascertain. Evidence for a crustal thickness of between 5 and 10 kilometres is provided by the spacing of major fault lines parallel to the edge of the Kaingaroa plateau. Furthermore, petrological studies of Ewart et al. (1971, 1975) provide evidence of rhyolites melting at temperatures from 700°C to 780°C and pumices derived from magmas at temperatures between 780°C and 860°C and mean depths of formation in the range 5 to 7 km. Various bounds on the depth to the hot-plate derived by McNabb (1975) from thermal profiles through the plume at Wairakei gave similar values between five and ten kilometres compatible with the above.

## 3. CONVECTION SYSTEM

The concept of a down-flow of cold ground-water over the whole of the Taupo Volcanic Zone (TVZ) onto a hot plate where it is heated and convected to the surface in a number of plumes and discharged as geothermal activity is consistent with the following data and analysis. The magmatic water content of hydrothermal waters was estimated by Wilson (Ellis & Wilson, 1960) to be at most 10 per cent, so that most of the water discharged is of meteoric origin and enters the system at the surface. The mass discharge of water via this single-transit convection system is of the order of 200 kg/s for each plume and 3000 kg/s for the whole TVZ. This is about 1/20 of the rainfall over the zone. Likewise, the total energy and salt outputs from the whole system are of the order $10^9$ cals/sec and 5 kg/sec respectively.

Pressures measured in the Wairakei field before substantial exploitation were correlated with depth (McNabb & Dickenson, 1975) and were found to lie about half way between cold and hot hydrostatic. The convective driving force in the hot column was 40 per cent of what it would have been if cold hydrostatic conditions had prevailed there, and the corresponding driving force in the cold region under the same gradient was 60 per cent of what it would have been with hot hydrostatic pressures prevailing there. Assuming the vertical permeabilities inside and outside these convective plumes are the same and a viscosity for cold water 10 times that for the hot, the ratio of the area of hot up-flow in the plume to the area of cold down-flow is roughly 3/20. This figure gives an area of about 150 $km^2$ of hot-plate per plume and a distance of about 12 km between plumes.

A further but somewhat enigmatic link between the hydrothermal activity and the rhyolitic magmas erupted from the TVZ derives from the observations that $10^{23}$ cal of thermal energy has passed out through the geothermal plumes in the last million years, assuming a level of activity prevailing throughout the life of the system at its present rate, and this is the same amount of thermal energy as is required to melt the rhyolites and tephras produced from the TVZ in the same period. This seems to imply some mechanistic coupling between the hydrothermal activity and the deep volcanic magma generating processes.

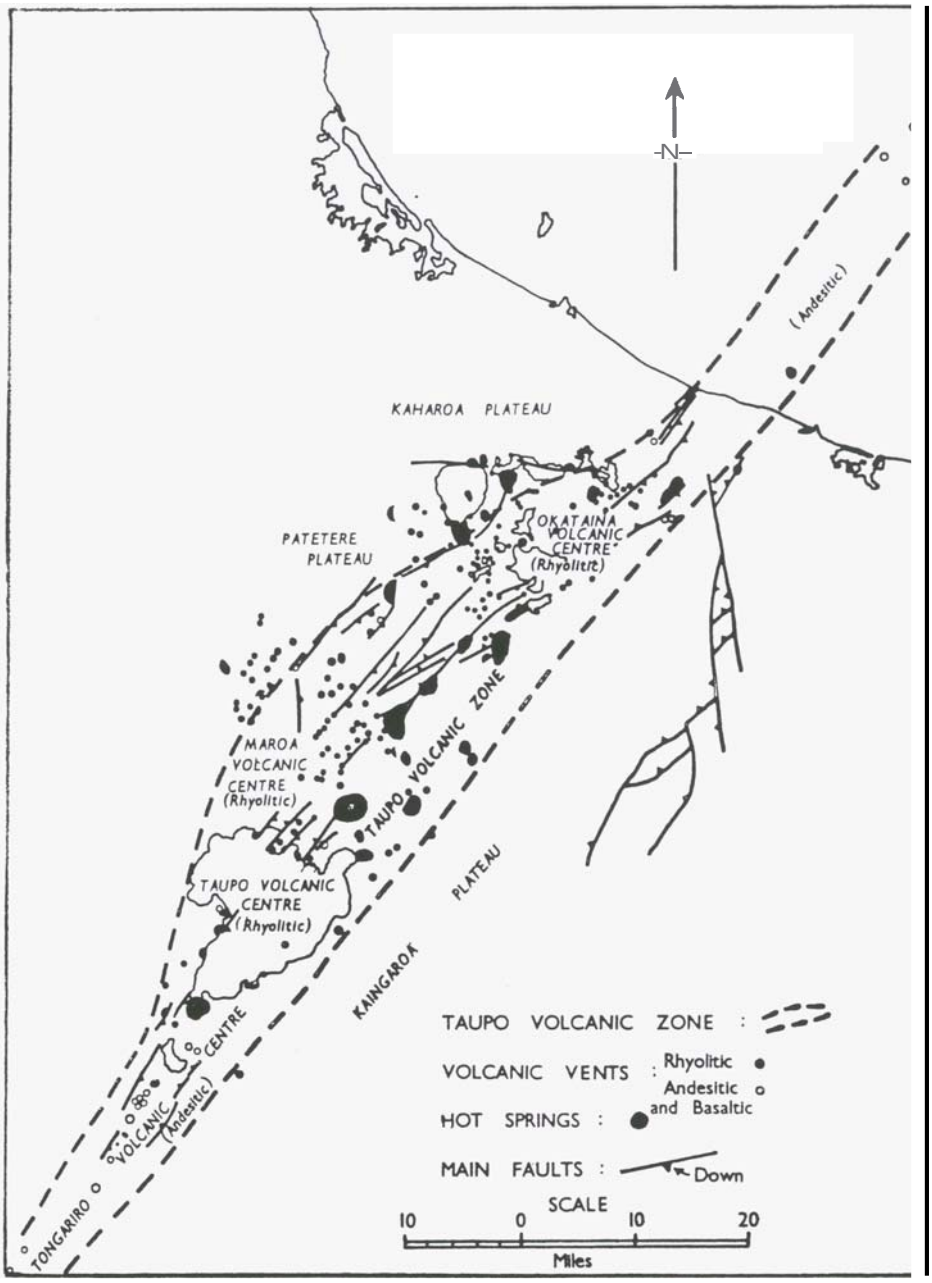


Figure 1 Location of volcanic vents and hot springs in the Taupo Volcanic Zone. (After Healy, 1964.)

## 4. TEMPERATURE OF HOT-PLATE

The temperatures below half a kilometre in the centre of the plume at Wairakei before exploitation were found to be approximately 250°C and increasing at about 20°C per kilometre. Such a temperature-depth profile points to hot-plate temperatures below the critical temperature for water, whereas one might have expected hot-plate temperatures at least twice this value if we are contemplating molten magmas as the heat source. How can this be?

A resolution of this problem can be found in the properties of high temperature - high pressure brine solutions (Sourirajan & Kennedy, 1962). Wairakei emits about half a kilogram of salt per second and since it is highly unlikely that the rocks of the TVZ were originally at least half a percent salt, this must be predominantly of magmatic origin.

Imagine a steady influx of magmatic steam entering the ground-water through the hot-plate beneath Wairakei carrying 2 per cent by weight of salt. A 10 percent contribution of such magmatic water to the plume would be sufficient to provide the sodium chloride in the output.

Now suppose the plume temperature were say 600°C near a hot-plate at a depth of 7 km. We see from Figure 2 that such a plume would start to form a dense brine phase as the fluid rose above 5 km depth.

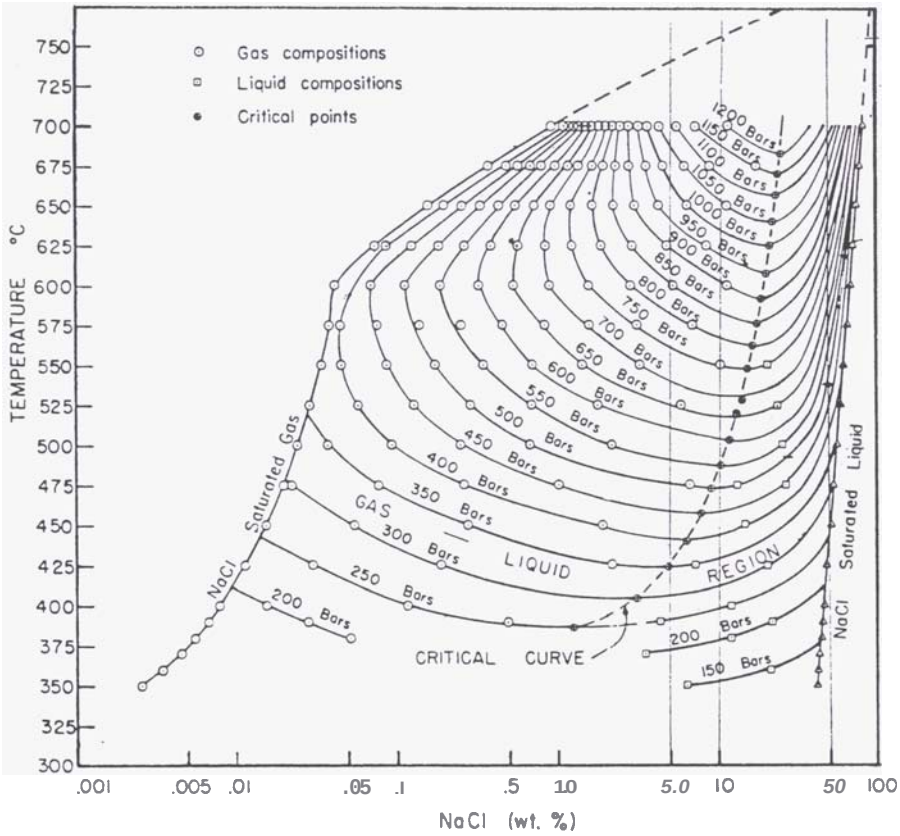


**Figure 2** Isobaric curves in the gas-liquid region in the system $H_2O$-NaCl. (After Sourirajan & Kennedy, 1962.)

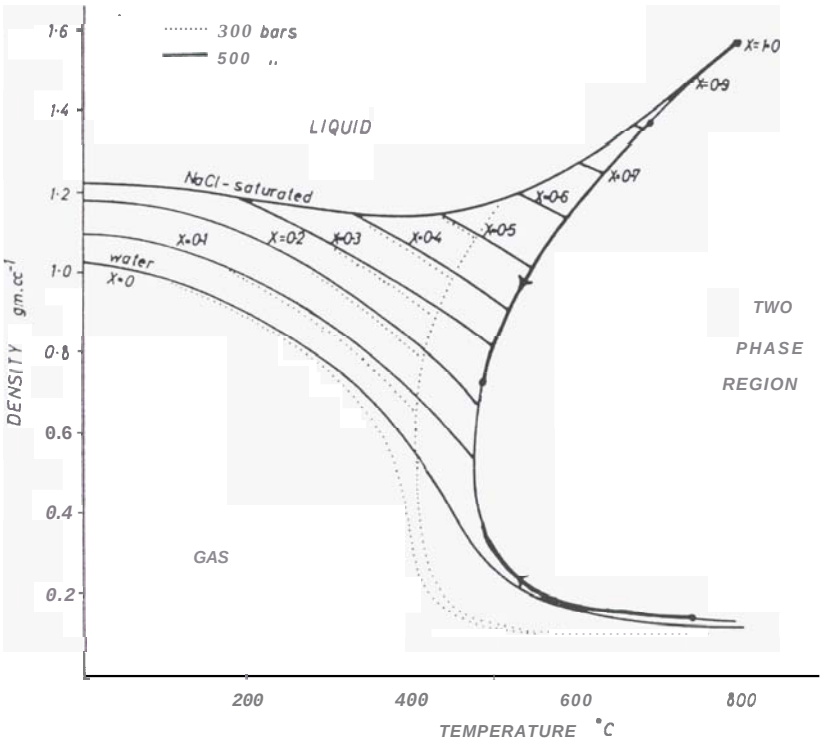


**Figure 3** Estimated densities of salt solutions at pressures of 300 and 500 bars. (After Henley & McNabb, 1978.)

The density of this second fluid phase as seen from Figure 3 (Henley & McNabb, 1978) is such that it would fall towards the hot-plate causing salt to be trapped and accumulated in the system. This salt transport problem prevails whenever plume temperatures are supercritical at shallow depths. One must conclude then that salt brought into the system when high hot-plate temperatures prevail remains in the system and accumulates over the hot-plate as a dense brine layer. The data in Figure 2 also shows that it is possible for a stably stratified hot dense brine layer to form at the bottom of the TVZ and thermally insulate the ground-water from molten magmatic temperatures. One can imagine the top of this purely conductive layer being at temperatures in the range 350°C to 400°C and the bottom being perhaps as hot as 700°C. Fluid pressures would be in the range 500 to 700 bars, depending on the depth of the hot-plate. If conduction of thermal energy through this layer is the dominant transport mechanism, and uniform conductivities prevail throughout, it could be assumed to sustain a linear temperature gradient with depth. A thickness of 1/4 km from magma to the top of the brine layer would give a conductive energy flux per km2 of nearly $10^6$ cal/s. If the material over the hot-plate is fractured greywacke with a porosity of 1 per cent, it would take about $10^4$ years chloride output to fill this layer. Although this is a surprisingly long time, it is small compared to the life of the system and it gives the regime a certain stability.

## 5. CHLORIDE FLUX CONTROL MECHANISMS

There are potential control mechanisms in this system able to adjust the temperature of the hot-plate to accommodate the flux of chloride through it. One such mechanism involves heated ground-water picking up chloride from the hot-plate as it passes over it by a double-diffusive mechanism described by Griffiths (1981) and releasing part or all of it again to a dense brine phase as it rises towards the top of the plume whenever the fluid is too hot, thus allowing the depth of the brine layer to increase and cool the convection process, or decrease the chloride content and heat the process if it is too cool. This mechanism requires the plume to contain a countercurrent of dense brine near its boiling surface, a condition not found at Wairakei.

Griffiths' (1981) assessment of chloride fluxes at Wairakei due to the double-diffusive mechanism assumes porosities of 0.2. Although this is a reasonable figure for near surface tephras, in the deep system (depths > 3 km) one is more likely to encounter original basement material - possibly greywacke, and in such material, a porosity of 1 per cent due to fracture structure would be as much as one could hope for. This would seem to reduce Griffiths flux estimates by a factor of 10. However, the convection processes in the TVZ involve a factor of 10 in the viscosity contrast between hot and cold water, a feature not incorporated in Griffiths work and one which leads to a corresponding increase in the area of hot-plate per plume. Perhaps this increase in hot-plate area compensates for the reduced porosity; at the least it would seem to be a feature worthy of further study.

Another possible control mechanism, perhaps a little in the realms of fantasy, supposes that a gradually thickening brine layer over a period of a thousand years or two leads to more rapid melting of basement material by a basalt magma. Such a magma is postulated by McNabb (1975) as the primary hot-plate for the rhyolitic magmatism in the TVZ. This process of intermittent blanketing of the basement material from cooling ground-water, followed by more rapid melting, could be a trigger for shallow rhyolitic eruptions. It would at least imply a coupling between the melting processes involving basement rock and the heat released through the hydrothermal activity above.

## ACKNOWLEDGEMENTS

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