MODELING THE EFFECTS OF DIRECT USE ON THE TAURANGA LOW-TEMPERATURE GEOTHERMAL SYSTEM, NEW ZEALAND

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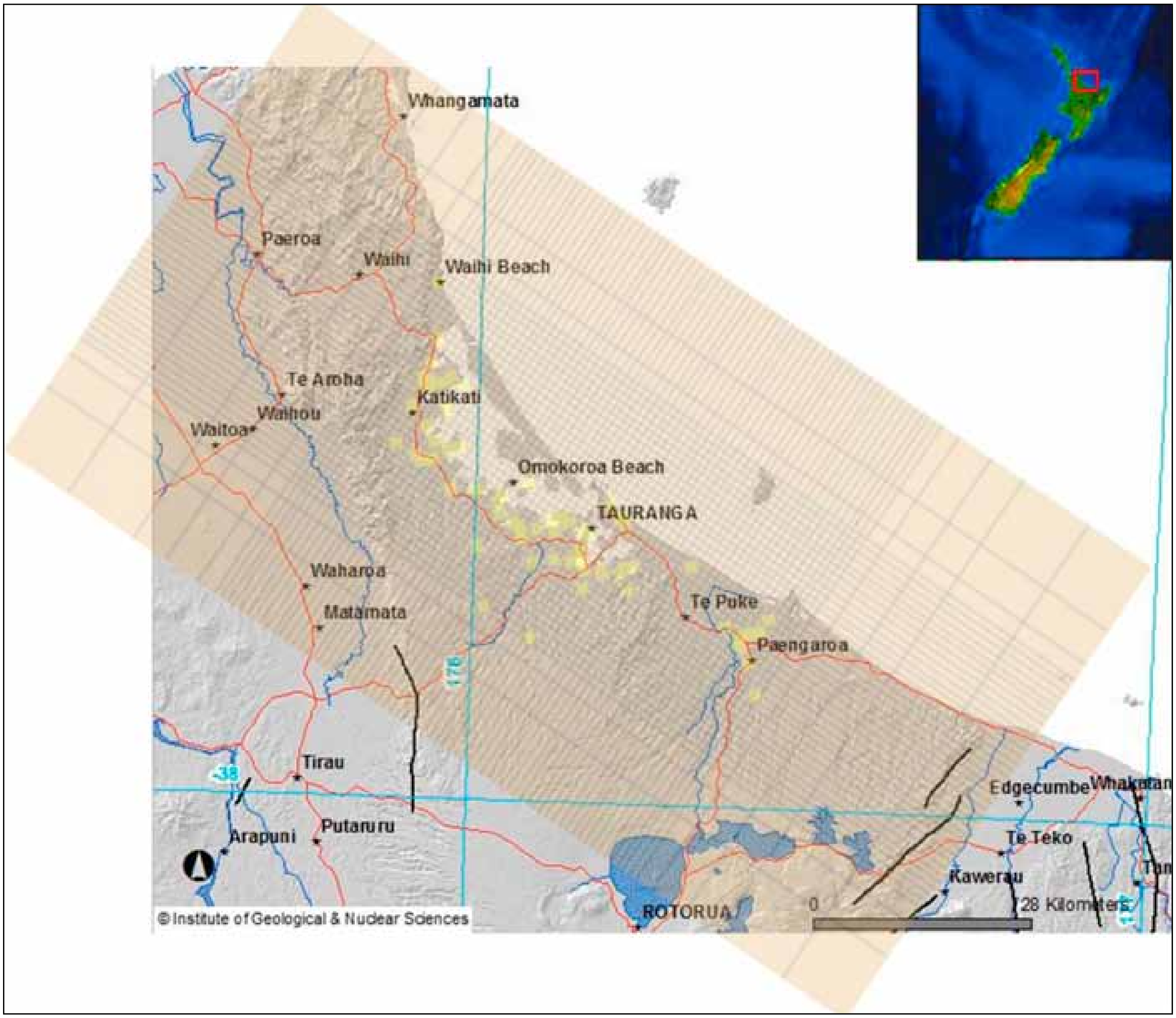

Figure 1. Location map of Tauranga geothermal field. The brown grid shows the model extent and dimensions. Yellow dots represent cells with well data. Black lines correspond to major faults. Inset map: Location of Tauranga within New Zealand.

ABSTRACT

Tauranga, on the north coast of the North Island of New Zealand, is the site of a fairly extensive low-temperature (60˚C at 800 m) geothermal resource that is currently used for hot pools, swimming baths, domestic use, greenhouses and tropical fish growing. As the population of the area grows and interest in direct use of geothermal resources increases, the system comes under increasing demand. In this study, a TOUGH2 heat and fluid flow model of the Tauranga geothermal field is used to determine the extent of the system, and the possible effects of withdrawing hot fluid from the area.

The TOUGH2 model covers a 70 by 130 km area and extends to 2 km depth. Modeled temperatures matched measured well temperatures using surface heat flow rates to constrain the heat input at depth. The high temperature gradient observed in the top 500 m was replicated using a low thermal conductivity of 1.05 W/m˚C in the shallow Tauranga Formation sediments. A good match could be obtained over the majority of the field using a homogeneous 2-layer model and two zones of basal heat influx. The model shows that heat flow is conductive to the northwest, but convective to the southeast. The geothermal system appears to be stable over long periods of time in its natural state.

When warm water is extracted, the pressure of the system re-equilibrates within a few months. However, there is a permanent decrease in temperature. After extraction has ended, the system takes hundreds of years for the temperature to return to its natural-state levels. Therefore it is important that these systems are carefully managed, and that modeling is carried out to ensure that they are not over-produced.

INTRODUCTION

Geothermal systems play a vital role as an energy source in New Zealand. 19% of total primary energy is geothermal, and 13% of electricity generation is from high-temperature geothermal sources (Ministry of Economic Development, 2011). Low temperature resources are particularly of increasing interest; in 2010 ~10 PJ of energy was used for industrial, commercial, agricultural and residential direct uses, an increase of 35% since 1990 (Ministry of Economic Development, 2011). It is therefore important to ensure that these systems are used and maintained in an effective and sustainable manner. TOUGH2 numerical modeling is often used to assess high-temperature geothermal systems (O'Sullivan et al., 2001), but here we apply it to the Tauranga low-temperature geothermal field to assess its energy potential and the effects of withdrawing hot fluid.

TAURANGA GEOTHERMAL FIELD

Tauranga is located on the north coast of the North Island of New Zealand (Figure 1). It is a city with approximately 120,000 people, making it the sixth largest urban center in New Zealand (Statistics New Zealand, 2012). It is bounded to the west by the Kaimai mountain range and to the east by the Pacific Ocean. The Tauranga area itself is relatively flat, other than Mount Maunganui (252 m) which is situated on the spit just north of Tauranga town (Figure 1).

Geologic Setting

Tauranga is located close to the subduction zone between the Pacific and Australasian plates. It is situated in the Tauranga Basin, a tensional graben formed about 2-3 million years ago (Davis & Healy, 1993). The basin sits within the Coromandel Volcanic Zone, a north-northwest trending zone that was highly active in the Miocene-Pliocene (Briggs et al., 2005). Volcanism commenced at ~18 Ma (Adams et al., 1994) but shifted to the Taupo Volcanic Zone between 1.9 and 1.55 Ma (Briggs et al., 2005). During this time at least 21 dacite-rhyolite domes or dome complexes and three defined ignimbrite formations were emplaced (Briggs et al., 2005). The remnant heat from these domes is thought to be the source of the warm water system at Tauranga (Reyes, 2008).

In a large part of the Tauranga area, the volcanics have been overlain by relatively young sediments. The Minden rhyolite domes remain some of the most dominant landforms, but these have been overlain inland by sediments dated at ~6.5 ka (Davis & Healy, 1993). Tidal sediments are somewhat younger, between 3.4 and 0.7 ka (Davis & Healy, 1993). Sediments thicken seawards (Simpson and Stewart 1987), reaching a thickness of 300 m under Matakana Island, but disappearing to the west of our study area (White et al., 2009). There are major faults to the south and west of our study area, but none within it (Figure 1) (Briggs et al., 2005).

Geothermal System

The Tauranga geothermal field is a significant low-enthalpy resource. There are a number of springs with water at between 22 and 39°C, and temperatures of up to 60°C have been measured in wells drilled to 800 m depth (White et al., 2009). These low-enthalpy fluids are used primarily for bathing, but also for domestic use, greenhouses and tropical fish growing (White, 2009). Tauranga is a popular tourist destination and hot pools and commercial swimming pools are found throughout the area, while Highway Fisheries in Papamoa, to the southeast of Tauranga city, is a major grower of ornamental and tropical fish. Therefore the geothermal field plays a significant role for the area, and its long-term stability and further potential are of interest to the region's inhabitants and authorities.

TOUGH2 SIMULATION

We used the Petrasim interface to TOUGH2 to create a numerical model of heat and fluid transfer in the Tauranga area. TOUGH2 simulates multicomponent, multiphase flow in porous media. Full details can be found in Pruess (1991).

The Model

We created a model to encompass the entire Tauranga area and some distance beyond (Figure 1). It covers 70 km by 130 km and extends down to 2 km depth. It is orientated to the northwest to fit the geographical extension of the field and to cover the locations of warm-water wells (Figure 1). Over the warm water area the spacing is 1 km by 1 km, but beyond this it has a spacing of up to 10 km by 10 km to ensure that the warm-water area of interest in the center is not affected by the boundary conditions (Figure 1). The model comprises two rock types: sediments overlying volcanics. Initially the sediments were 150 m thick throughout, but later a more realistic stratigraphy was added where the contact dipped eastward so that the sediment thickness was 50 m to the west of the model but 300 m to the east (Figure 2).

The model was run with fairly simple initial conditions for two million years, to represent the age of the Tauranga Basin (Davis & Healy, 1993). Initially the interior and boundaries of the model were set at atmospheric pressure (101 kPa) and mean annual air temperature (12°C) (NIWA, 2011). It was fully water saturated but the uppermost layer was 100% air to represent the atmosphere (Figure 2) and had a very large volume so that the atmospheric conditions were fixed. This allowed recharge into the system to be simulated at realistic rates. Recharge was injected into the second layer at 129 mm/yr (Figure 2) to simulate 10% of the mean annual rainfall (NIWA, 2011). Vertical boundaries were set as no-flow. As geochemistry suggests that there is minimal flow of geothermal fluids from depth (Reyes, 2008), heat was input into the base of the model at varying rates until an optimal fit was found between model temperatures and measured ones.

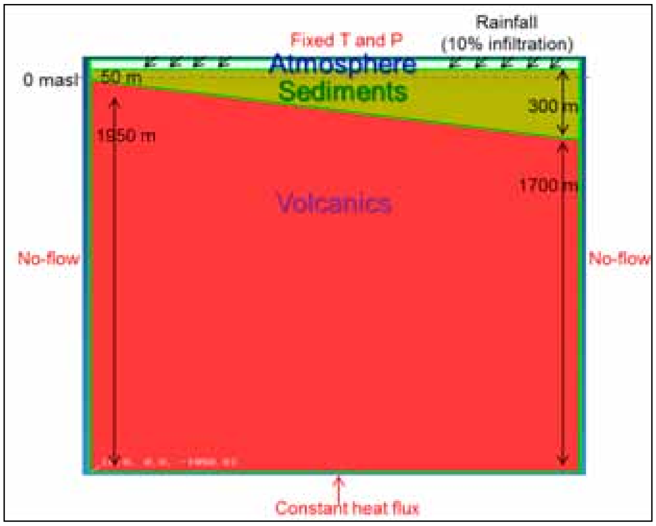

Figure 2. Cross-section of model showing stratigraphy and boundary conditions.

Constraints

A number of modeling constraints have been measured in or near the Tauranga area. Well temperatures were used as the primary constraint on the model. Geochemistry was used as a guide for boundary conditions. Basal heat flux, permeability and thermal conductivity were based on surface measurements but were varied to minimize the misfit between modeled and measured data. Density, specific heat capacity and porosity were set at measured/typical values.

Well Temperatures

The Tauranga area has been drilled extensively for groundwater studies, providing lithological and temperature information (White et al., 2009). More than 150 wells tap warm groundwater in 500 km^2 area around Tauranga (Simpson, 1987). Between 1960 and 2005 the temperature was measured in 73 wells. In 17 of them temperature profiles were recorded with depth, while the rest were measured at a single depth. The measurements were recorded at between 149 and -738 masl, from the surface to 752 m depth. Temperatures varied between 12 and 56˚C, with the majority at between 20 and 40˚C (White et al., 2009). In general deeper measurements were hotter (Figure 3). These well temperatures were used as the primary constraint for the TOUGH2 model. Other data in the area is also in agreement, with temperatures generally 35-45˚C at 600 m, but sometimes over 55˚C (Simpson, 1987).

Geochemistry

Geothermometry from the nearby Hauraki Fault suggests that temperatures are up to 160˚C (Reyes, 2008). Geochemical analysis shows that geothermal fluids in the Tauranga area are mainly heated groundwater with minor seawater in the north and minor magmatic volatiles in the south nearest to the Taupo Volcanic Zone (Reyes, 2008). Seawater intrusions have been noted around Mt Maunganui (Simpson & Stewart, 1987), although they are now thought to be at minimal levels (White, 2005).

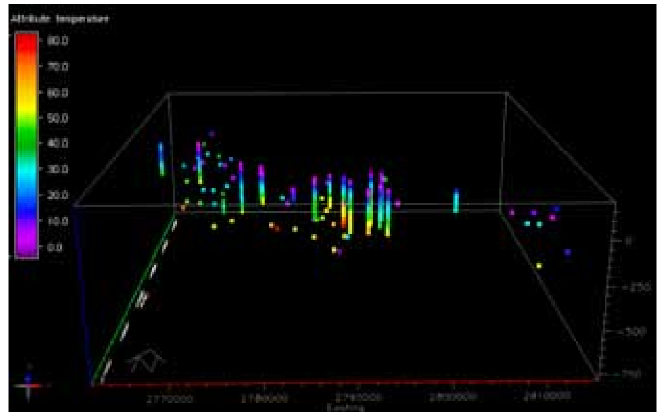

Figure 3. Temperature measurements from groundwater wells in Tauranga (White, Meilhac, et al. 2009) were used as the primary modeling constraint.

HEAT FLUX

Surface heat flux has been measured across the Tauranga area a number of times. As basal heat flux is a major variable in the model but is unconstrained, we used the surface heat flux as an initial guess for the basal flux rather than as an output of the model.

Average heat flow over the Tauranga area is measured as 88 ± 16 mW/m^2 (Simpson, 1987). In several distinct areas (Maketu, Mt Maunganui and around Tauranga Harbour edge) heat flow is above 120 mW/m^2, up to 336 mW/m^2. At one site a heat flow of 55 mW/m^2 was measured, but just 8 km to the southeast a heat flux as high as 200 mW/m^2 was recorded (Studt & Thompson, 1969). In the nearby Hauraki rift zone surface heat flux has been measured at between 80 and 90 mW/m^2 (Reyes, 2008). This means that there is considerable variability in the surface heat flux over relatively small areas. To prevent the model from becoming complicated beyond the level that the information can support, the average of 88 mW/m^2 was used across the base of the whole model initially and varied to refine the fit of the model temperatures to measured data.

Permeability

Permeability is difficult to constrain, but some work has been done in the Tauranga area. Outcrops show that volcanic rocks exhibit variable permeability and are fractured, allowing them to transmit fluid but not freely (Simpson, 1987). In general, the shallow groundwater system is fed by recharge in sediments while the deeper system contains considerably older fluids and is only recharged slowly by vertical seepage (Petch & Marshall, 1988).

In the Tauranga group sediments, permeability estimates in the Hamilton area (100 km away) range from 5 x 10^{-13} m^2 in the silts and sands to 9 x 10^{-12} m^2 in the coarse sands (Petch & Marshall, 1988). Bulk permeability is up to 8 x 10^{-11} m^2 (Heu, 1985). As sediments are typically less permeable than this (Bear, 1972) and the layer is thought to be a confining cap (Simpson, 1987), the upper value of 5 x 10_{-13} m^2 was used.

Estimates of shallow permeability in volcanic rocks in Auckland (200 km away) are similar to those for Tauranga sediments. They range from 2 x 10^{-13} to 8 x 10^{-11} m^2 for volcanic rocks that have little or no scoria, and from 8 x 10^{-12} to 8 x 10^{-10} m^2 for volcanic rocks with significant amounts of scoria (Harding et al., 2010). The bulk permeability is likely to be significantly lower than this (O'Sullivan, personal communication, December 2010) because our model extends to some depth (Ingebritsen & Scholl, 1993). Harding et al. (2010) did not find any evidence of significant horizontal/vertical anisotropy.

Other rock properties

A number of measurements have been made in the Tauranga area that provides extremely useful information for heat and fluid flow models. For the Tauranga Formation sediments, the thermal conductivity has been measured at 1.05 W/m˚C (Simpson, 1987). This was therefore used in the model, although other values were also tried. Typical values of 2,500 kg/m^3 and 0.1 were used for the rock density and porosity respectively. For the volcanic rocks in the Tauranga area, more measurements have been made and so a larger range of properties have been constrained. Taking an average of all of these values gives a density of 1,890 kg/m^3, a porosity of 0.42 and a thermal conductivity of 1.26 ± 0.05 W/m˚C (Simpson, 1987). These were the values used in the TOUGH2 modeling.

Results

Modeling shows that with a fairly simple model of two rock layers and just two different zones of heat influx, a good match can be obtained to most of the well data (Figure 4). In the shallow sediments, a permeability of 5 x 10^{-13} m^2 provides the best match; lower permeability results in the model wells being slightly too cold. However, the model appears to be fairly insensitive to this parameter. In the volcanic rocks, permeability of more than 5 x 10^{-16} m^2 results in convection throughout the system. This would result in fairly large variability in well temperatures that is not observed, particularly in the northwest. Therefore a permeability slightly less than that suggested by the literature is required to match well temperatures with model data.

As the temperature of the field is fairly low, conduction is a major source of heat transfer. This means that the basal heat flux is very important. The average value of 88 mW/m^2 (Simpson, 1987) gives a good match to well data in the northwest of the model, but fluids in the southeast wells are generally hotter than model temperatures (Figure 5a). With a heat flux of 120 mW/m^2 to the southeast as suggested by surface measurements, the match is greatly improved (Figure 5b). Modeling suggests that this results in conduction to the northwest, but some convection to the southeast.

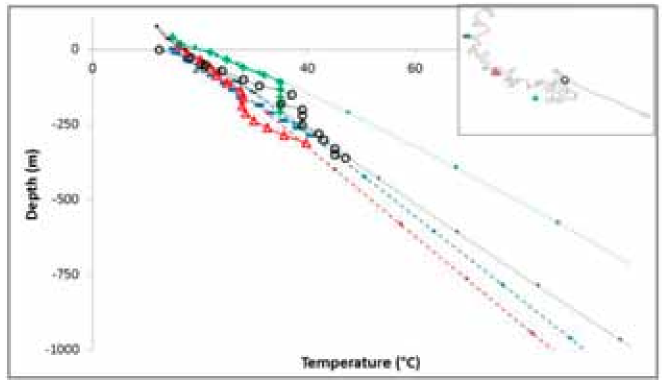

Figure 4. Match between measured (solid lines, large symbols) and modeled (dashed lines) data. The inset map shows the locations of the wells within Tauranga.

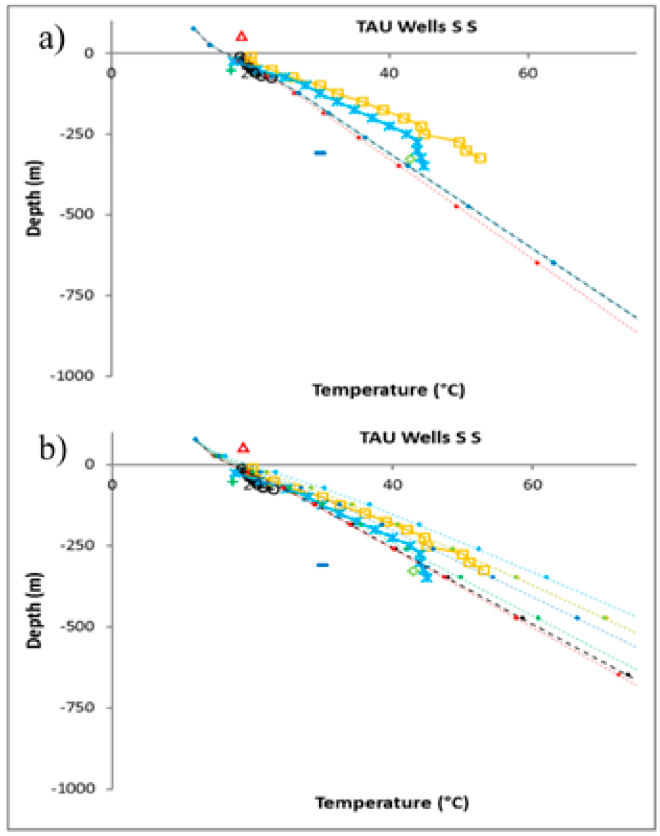

Figure 5. Model results to the southeast with varying basal heat flux. Dashed lines represent model results; solid lines and large symbols represent measured well data. a) Uniform heat injection of 88 mW/m^2. b) Heat injection of 88 mW/m^2 to the northwest and 120 mW/m2 to the southeast.

The model shows that the system is fairly sensitive to the thermal conductivity of the rock. Measured thermal conductivities are relatively low (Simpson, 1987), and these provide the best match to the data. With a higher thermal conductivity, the temperature does not increase quickly enough with depth. With a lower thermal conductivity, the shallow rock reaches very high temperatures. Therefore the measured thermal conductivities of 1.05 W/m˚C in the sediments and 1.26 W/m˚C in the volcanics appear to be fairly widespread within the system. In the sediments there is an unusually high thermal gradient of ~120˚C/km in most of the field which can be explained by this low thermal conductivity.

Although the well data did not include any colder areas to the south or west, the well to the north allows the northern boundary of the geothermal system to be identified to within 100 m (Figure 6), as northwest of Katikati but southeast of Waihi Beach (Figure 1). Adding more data to the south and west would allow the full extent of the geothermal field to be determined, important for understanding the system and its potential capacity. Changing the contact between the volcanics and sediments from horizontal to the more realistic dipping to the east does not affect many of the results, but does improve the fit in some cases.

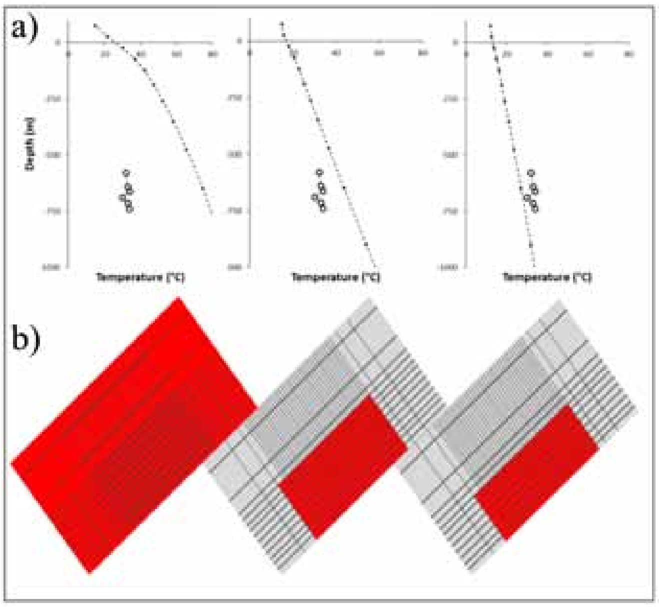

Figure 6a.) Model results (dashed lines) compared to data (symbols) in the furthest north well at Waihi Beach (Figure 1). Reducing the area of heat input from left to right (b), significantly improves the fit, allowing the northern extent of the system to be determined to within 100 m.

Modeling allows the energy and fluxes contained within the Tauranga system to be determined. It suggests a total energy of 228 MW within the system, but spread over 2,360 km^2. The average fluid flux is just 2.6 x 10^{-9} kg/m^{2}s at the surface over the area of heat input. The maximum fluid flux is 1.5 x 10^{-6} kg/m^{2}s, with a heat flux of 595 mW/m^2. This maximum heat flux is slightly higher than surface measurements, but within an order of magnitude. These model results suggest that there is significant energy potential within the system, but that it is widely distributed throughout the area.

Calculating errors throughout the model allows us to identify the areas that are most poorly represented. The average error is 27%, with 70% of errors less than 25% (Figure 7). This is acceptable, particularly as many measurements are single values recorded in open wells during different times of year and they are fairly small numbers so errors are proportionally larger. However, in the center of the field (W2018) there is an error of 186% (Figure 7), possibly due to topographic effects, localized variations in depth to the heat source, rock properties or measurement error.

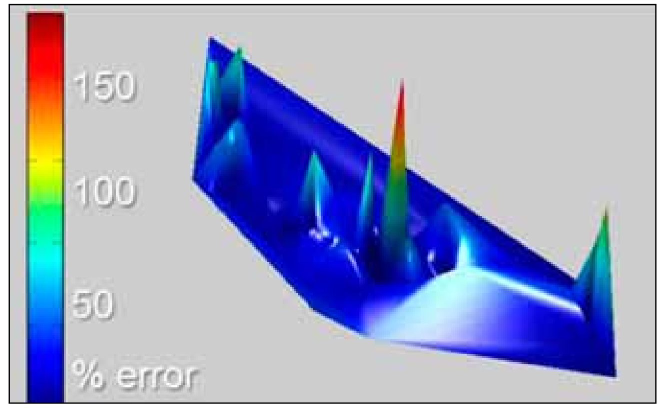

Figure 7. Plot of misfit between model temperatures and well measurements.

PRODUCTION

As there is a significant amount of energy within the Tauranga geothermal system but it is spread over a wide area, over-utilization could definitely become a problem. Therefore we used the TOUGH2 model to study the effects of withdrawing fluid. We started by simulating a production well in the center of the model for 100 years. The depths of production were 75 m (near the surface), 125 m (at the contact between volcanics and sediments), 350 m (within the volcanics) and 650 m (within the volcanics near the depth of the deepest well). Rates varied from 4 to 40 kg/s in the one cell, so from 4 x 10^{-6} to 4 x 10^{-5} kg/m^{2}s. This is up to an order of magnitude greater than the maximum modeled fluid flux and four orders of magnitude greater than the average.

Modeling a medium production rate at different depths shows that this should be sustainable (Figure 8). For production at 8 kg/s (8 x 10^{-6} kg/m^{2}s) with shallow production, the temperature decreases steadily over 100 years but the pressure is only minimally affected (Figure 8). In contrast, for deep production the temperature remains stable but the pressure decreases. The pressure decrease is very rapid however; after 2 months the system has restabilized but at a lower value (Figure 8). This suggests that shallow production, from within the sediments, would cause the system to continually cool, whereas deeper production from within the volcanics could affect surface features as the pressure drops, but would then be more stable over the long term.

For the volcanic-sediment interface (125 m depth), a range of production rates shows that, as expected, the higher the production rate the greater the decrease in temperature and pressure (Figure 9). Again, the pressure restabilizes after a few months but at a lower level, while the temperature decreases steadily by as much as 10°C for the highest withdrawal rate, and by at least 2°C for a withdrawal rate on the same order of magnitude as the maximum modeled. This is a decrease of between 6 and 30%.

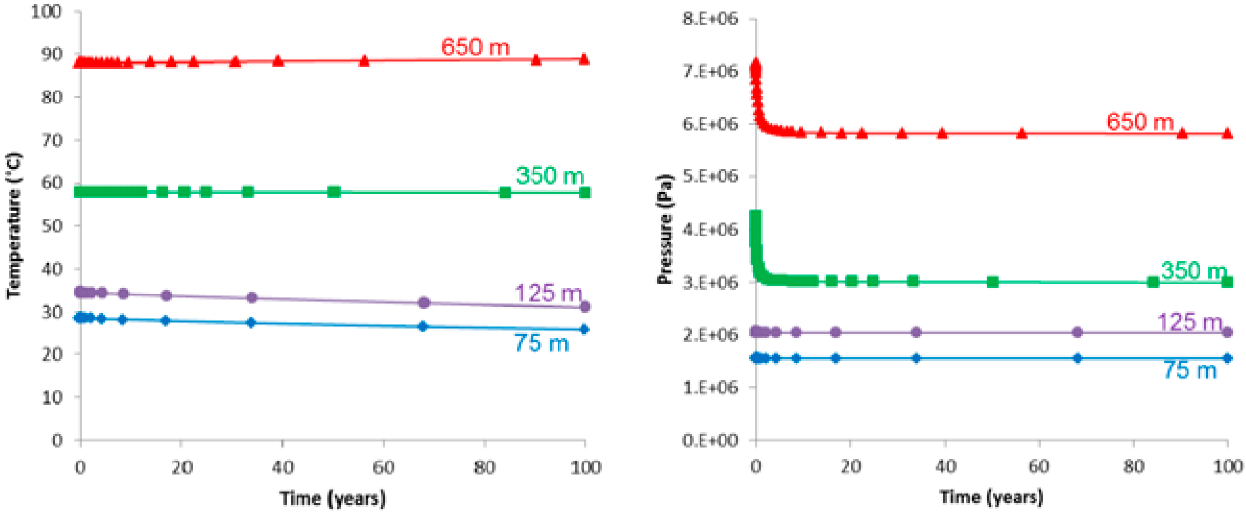

Figure 8. The effects of withdrawing fluid at 8 kg/s over a 1 km² area. Numbers represent the depth of production.

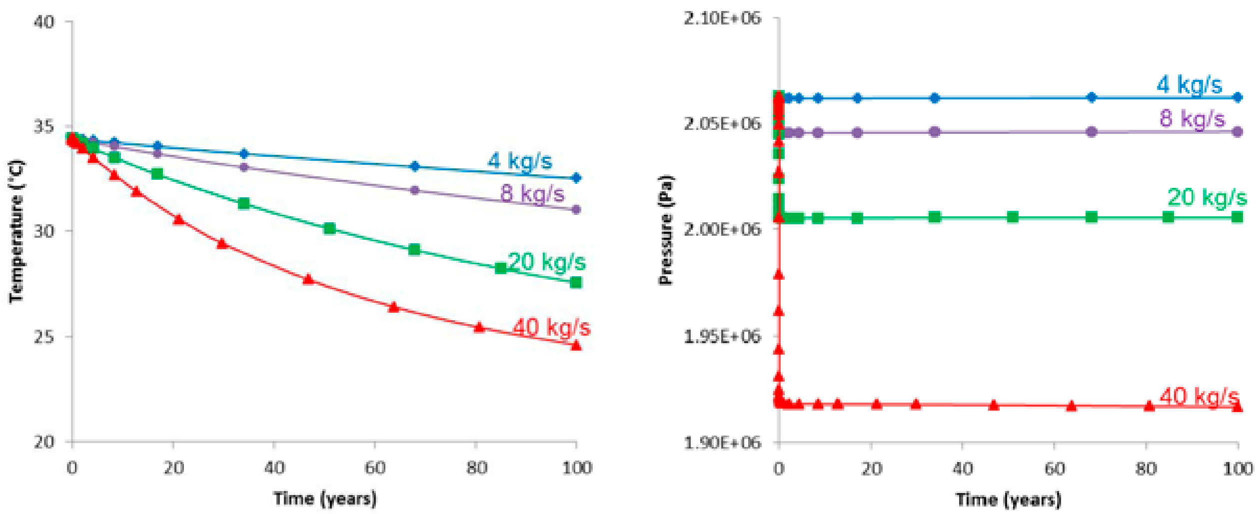

Figure 9. Effects of withdrawing fluid from the sediment-volcanic interface. Numbers represent the production rate over a 1 km² area.

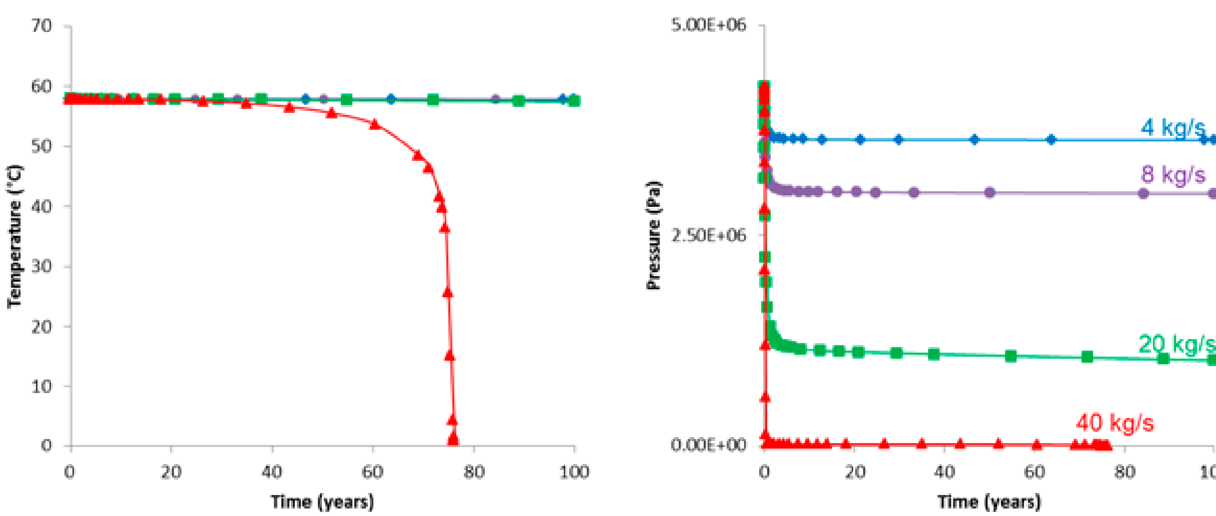

Figure 10. Effect of withdrawing fluid at 350 m depth. Numbers represent withdrawal rates. For the highest withdrawal rate, the system dies after less than 80 years.

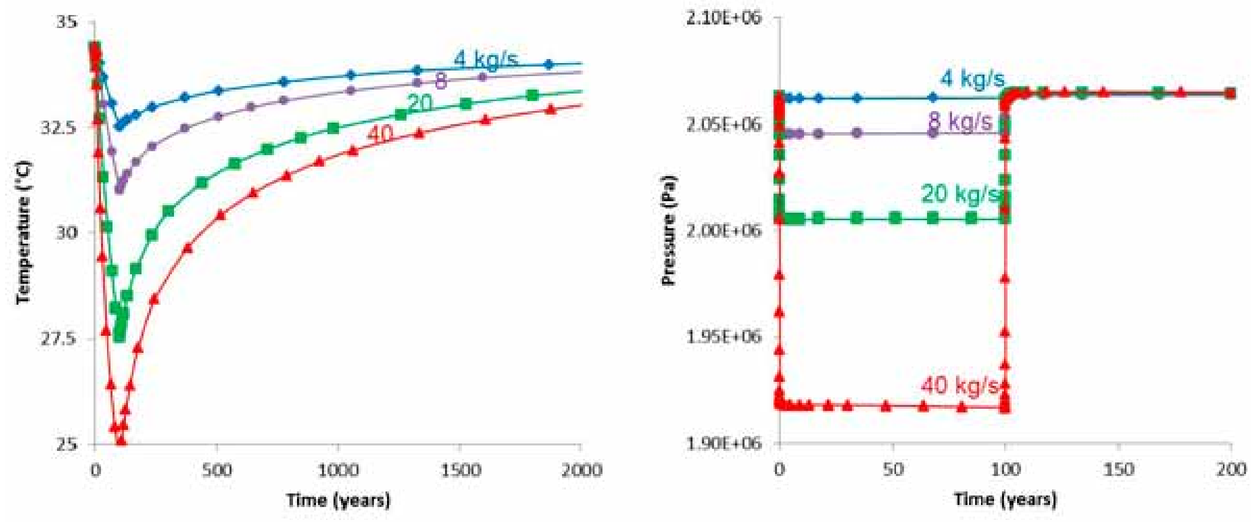

Figure 11. Simulated recovery of the system after production. Numbers show the rates of production at 125 m depth. Withdrawal was modeled for 100 years and was then turned off.

For the deeper wells, within the volcanics, the effect of high withdrawal rates can be even more severe. At 350 m depth, the temperature remains stable but the pressure drops significantly for rates of 4 – 20 kg/s (Figure 10). For 20 kg/s, the pressure decreases by more than half which would definitely affect surface features. At a withdrawal rate of 40 kg/s, the system essentially dies as it cools down and dries up (Figure 10). Therefore it is important to manage withdrawal from these types of systems, and to decide on the maximum induced variations that would be acceptable.

RECOVERY

Another important aspect of the system to understand is its recovery after withdrawal has ended. The model was therefore run for another 10,000 years after switching off the well. It shows that the system does recover, but very slowly (Figure 11); after 100 years less than 25% of the temperature loss has been recovered. After 1,000 years the temperature is half-way back to background levels, but it takes a full 10,000 years for the system to re-approach its natural state. However, the pressure again re-stabilizes after only two months, and at the original levels. This suggests that the Tauranga system is stable before, during and after production, but heat is essentially lost permanently, and it would take the system a very long time to recover from over-production.

CONCLUSIONS

The Tauranga geothermal field is a low-temperature system that contains ~225 MW over more than 2,300 km^2. Modeling allows the northern extent of the field to be determined, and shows that the low thermal conductivity measured in the Tauranga sediments is the best explanation for the relatively high thermal gradient measured in wells in the area. The heat flux was found to be the main constraint on the model, although a simple two-zone model with 88 mW/^{2} to the northwest and 120 mW/m^2 to the southeast results in a good match between measured and modeled temperatures. The average modeled fluid flux above the heat source is just 2.6 x 10^{-9} kg/m^{2}s, while the maximum fluid flux is 1.5 x 10^{-6} kg/m^{2}s with a maximum heat flux of 595 mW/m^2. This suggests that there is significant energy within the system, but that it is generally very diffuse and therefore only appropriate for direct use.

Modeling production scenarios shows that for rates twice that of the maximum modeled as naturally occurring within the system, shallow wells cause a constant decrease in temperature, while deep wells result in a rapid drop in pressure that then re-stabilizes at a lower level. For a withdrawal rate ten times modeled, production from the deep wells results in the entire system dying. Modeling recovery suggests that it is very slow, on the order of thousands of years. The pressure appears to be stable, but the effect of withdrawal on the temperature of the geothermal system is essentially permanent. Therefore it is vital that these systems are well managed to ensure that fluid withdrawal is sustainable.

FUTURE WORK

There are a number of steps that we hope to achieve to improve this model. Initially, the misfit between measured and modeled well temperature data needs to be addressed by varying topography, local rock properties and/or heat flux. There may also be more well data that can be included in the model, particularly to the southern and western extents of the currently modeled warm water area.

Having improved the model and recalibrated it, we hope to simulate more production scenarios. Well locations, depths and approved withdrawal amounts from the local authorities will allow us to assess current and future usage rates and their potential long-term effects. We will also model reinjection scenarios based on actual data. From this we will be able to deduce whether the system is cooling, and if currently approved rates are sustainable. We then hope to add some additional wells to see if the current system capacity can be increased for direct use.

EDITOR'S NOTE

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