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Key Points:
- Geothermal systems in the Taupo Volcanic Zone are long-lived and stable, which requires non-uniformity.
- Topography can provide this non-uniformity and explain the locations of 14 of 19 geothermal systems in our fluid and heat flow model area.
- Topographically driven convection can extend to several kilometers depth, even with moderate surface relief.

Supporting Information:
Supporting Information may be found in the online version of this article.

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Topography as a Major Influence on Geothermal Circulation in the Taupo Volcanic Zone, New Zealand

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Abstract
Geothermal systems are an important resource in the global effort to reduce fossil fuel use. They are formed by convection of water heated at depth, but without external controls convection is unstable and upflows move or dissipate. In the central Taupo Volcanic Zone in New Zealand, the geothermal systems have been in the same locations for more than 50 kyr. We created heat and fluid flow models with uniform geology and basal heat which showed that the known topography localizes convective upflows. Fourteen out of 19 known geothermal systems in the model area were correctly located beneath topographic lows. Even at 3 km depth, inclusion of topography in our models improved agreement between the locations of upflows and low-resistivity zones derived from 3-D inversion of magnetotelluric data. Therefore, while topography is not the only factor that influences geothermal convection, its contribution is on a larger scale than is generally assumed.

Plain Language Summary
Geothermal systems are a clean, renewable energy source. They are created by water that is heated at several kilometers depth and rises to the surface in distinct “plumes.” We don’t know why the plumes rise where they do or why they seem to stay in the same place for hundreds of thousands of years. We used specialized computer code to look at whether topography can keep these plumes in place. With uniform geology and heat flow at depth, our simplified models show that high topography can drive water downwards even at 3 km depth. The geothermal systems form between these highs, in valleys and topographical lows. Faults and localized heat sources will also be playing a part, but topography is known accurately and can explain the locations of 14 of the 19 geothermal systems in the Taupo Volcanic Zone, New Zealand. Before, it was thought that topography mainly affected nearby shallow fluid flow. Future geothermal exploration and modeling of warm water extraction should take these results into account.

1. Introduction
Geothermal energy is an important resource in the global effort to reduce fossil fuel dependence. However, our understanding of why geothermal systems form where they do and remain there is limited. Geothermal systems are formed by convective fluid flow within the outermost brittle part of the Earth’s crust, requiring an underlying heat source, fluids, and circulation pathways (Rowland & Sibson, 2004). Evidence from numerous systems in the Taupo Volcanic Zone (TVZ) suggests that their locations are stable for hundreds of thousands of years (Arehart et al., 2002; Miličič et al., 2018). Heat sources at depth can initiate convection, but without other effects the resulting upflows move around over time or dissipate completely (Kissling & Weir, 2005).

Fluid flow can be focused by subsurface hydraulic pressure gradients, causing preferential recharge in areas of high topography and discharge in areas of low topography (Garven, 1995). Examples of topographically driven fluid flow are found worldwide and have been observed in the Basin and Range, USA (Person et al., 2008), the Swiss Alps (Bodri & Rybach, 1998), the Alberta Basin, Canada (Bachu, 1995), Citaman in Indonesia (Hochstein, 1993) and Waimangu Volcanic Valley in New Zealand (Simmons et al., 2020). In the high-relief Swiss Alps and western USA, topographic effects have been observed to 2–5 km depth (Bodri & Rybach, 1998; Luijendijk et al., 2020; McIntosh & Ferguson, 2021). In general however, it is assumed that topographic forces are significant only at shallow depth, and are small compared to buoyancy forces that drive fluids vertically through fractures toward the surface (Chi & Xue, 2011).
Controls on fluid flow in the crust have been investigated with numerical modeling, where geological complexity can be simplified to identify other processes that can have an effect. For example, high-relief surface topography has been shown to drive regional-scale convective fluid flow (Forster & Smith, 1989). In fracture-dominated systems, low-relief topography (<600 m) has been shown to enhance geothermal circulation, focus upflow zones, or distort a hydrothermal plume (Ratouis & Zarrouk, 2016; Wisian & Blackwell, 2004). However, low-relief unfractured terrain with moderately elevated heat flux results in temperatures that are laterally consistent over tens of kilometers (Forster & Smith, 1989). Here, we explore the effects of low-relief terrain with uniform geology, and heat flux at a higher level more typical of geothermal systems. We use the characteristics of the extensively studied Taupo Volcanic Zone as an example (e.g., Bertrand, et al., 2012; Bibby et al., 1995; Rowland & Sibson, 2004; Wilson & Rowland, 2016).

2. Geologic Setting

The Taupo Volcanic Zone (TVZ) is an actively rifting volcanic arc (Wilson & Rowland, 2016). Heat flux in the TVZ is 10 times higher than for average continental crust and is focused through the central silicic ~6,000 km² (Figure 1) (Bibby et al., 1995). Active arc-volcanism occurs to the southwest, beyond Lake Taupo, and to the northeast, continuing offshore into the Tonga-Kermadec arc (Figure 1; Bibby et al., 1995; Seebeck et al., 2014; Wilson & Rowland, 2016). Geothermal activity manifests through 23 individual geothermal systems in the central TVZ that discharge an estimated 4.2 GW of heat (Figure 1; Bibby et al., 1995).

The lithology and geologic structures in the central TVZ are dominated by volcanic and tectonic activity that occurred more than 61,000 years ago (Wilson & Rowland, 2016). Mesozoic metasedimentary basement is overlain by up to 3 km of volcanic and volcanioclastic rocks (Downs et al., 2014). Steps in the basement surface have been created by faults and by seven major caldera-forming eruptions between ~350 and 280 ka (Figure 1) (Alcaraz et al., 2012; Wilson & Rowland, 2016). These eruptions and many minor ones deposited the rhyolitic ignimbrites and less extensive pyroclastic units and lavas that form the dominant stratigraphy encountered beneath central TVZ geothermal systems (Downs et al., 2014). Geothermal wells have also intersected a (cold) rhyolitic intrusion and buried andesitic volcanic rocks (Chambefort et al., 2014).

Within the basement, focal depths of earthquakes indicate that on average rocks are brittle to ~6–7 km depth (Hurst et al., 2016; Stagpoole et al., 2015). Above this brittle-ductile transition, rocks can host fracture networks and convective fluid flow. Permeability measurements from samples located near the top of the greywacke basement are on the order of 10⁻¹⁴ to 10⁻¹³ m², and vary from 10⁻¹⁴ to 10⁻⁹ m² in the overlying volcanic deposits (Milicich et al., 2018). Topographic variation across the central TVZ is moderate, with local changes of less than a few hundred meters (LINZ, 2012; Rowland & Sibson, 2004).

The central TVZ is geographically divided into two parallel northeast-elongate basins separated by a high-standing, fault-controlled range known as the Paeroa Block (Downs et al., 2014). To the northwest, the Taupo Fault Belt exhibits a classic rift morphology and is active seismically (Downs et al., 2014). It is associated with cold down-welling fluids and an absence of geothermal systems (Rowland & Sibson, 2004). To the southeast, the Taupo Reporoa Basin has very limited geomorphic evidence for faulting and little topographic relief (Wilson & Rowland, 2016). The TVZ geothermal systems are predominantly within the Taupo Reporoa Basin, or are to the northwest of the Taupo Fault Belt (Figure 1). Only a few of the geothermal systems appear to be associated with major faults, for example at Waiotapu, Te Kopia, and Orakei Korako (Villamor et al., 2017).

A conceptual model of TVZ geothermal systems has evolved from extensive research undertaken over several decades. The systems appear to be long-lived and stable, with estimated lifetimes of at least 50,000 to 250,000 years (Rowland & Simmons, 2012). Heat transport in the brittle part of the crust is dominated by convection of meteoric fluid, with only a minor magmatic component (Giggenbach, 1995; Stewart, 1978). Recent magnetotelluric (MT) surveys designed specifically to image the brittle part of the crust have shown low-resistivity “plumes” (interpreted as upwelling fluids) rising through the crust and connecting with the surface locations of geothermal fields (Bertrand et al., 2012, 2015; Heise et al., 2016). Although some of these imaged plumes are quasi-vertical, others are inclined and offset from the geothermal expression at the surface, for example at Ohaaki (Bertrand et al., 2013).
The underlying heat source that drives convective heat transport in the brittle TVZ is still in question. End-member hypotheses are (1) a “hot-plate” of partial melt below the brittle-ductile transition that extends beneath the entire central TVZ (Kissling & Weir, 2005; McNabb, 1992), or (2) localized intrusions associated with individual geothermal systems (Chambefort et al., 2014; Wilson & Rowland, 2016). Numerical simulations have shown that localized heat sources in combination with faults could explain several aspects of the dynamics of geothermal circulation in the TVZ (Dempsey et al., 2011; Kaya & Widyanti, 2015; Kissling & Weir, 2005). One shallow intrusion has been drilled that was emplaced 710 to 650 ka that could be such a heat source (Chambefort et al., 2014), however a lack of hydrothermal alteration in the overlying...
Whakamaru ignimbrite (∼349 ka) shows that this intrusion is not the heat source for the present-day hydrothermal system (Arehart et al., 2002). The time-scale of cooling for such an intrusion in the crust (if not replenished) is on the order of tens of thousands of years (Kissling, 1999), and cannot explain the longevity of the TVZ geothermal systems. In addition, there are few obvious crustal structures such as faults or caldera boundaries that would focus the convective upflow from such bodies.

3. Model Design

Our models of the central TVZ focus on a 25 by 34 km area that encompasses six geothermal systems (purple rectangle in Figure 1). Using a subsection of the TVZ allows detailed topography to be included in the model without unfeasibly long computational times. The area selected is representative of the wider TVZ with varied topography and lithology, as well as faults and groundwater catchment areas (Milicich et al., 2018; Pearson-Grant et al., 2017; White & Reeves, 2017). The model was later extended to encompass the entire central TVZ at a coarser resolution (green rectangle in Figure 1).

The newly developed Waiwera code (Croucher, 2020a, 2020b; Croucher et al., 2018) was used to simulate heat and fluid flow in the brittle crust to a depth of 5 km, avoiding supercritical conditions that may occur deeper and cannot be reproduced by the code. Initial model results were consistent with equivalent models generated using TOUGH2 software (Pruess et al., 1999), which is the most commonly used simulator for geothermal reservoirs (Burnell et al., 2015; O'Sullivan et al., 2001). The modern Waiwera code was preferred as it is more stable during phase changes and was found to be computationally faster for these models.

The model was comprised of a grid with horizontal blocks of 500 m by 500 m, and 250 m vertically. There were up to 24 layers vertically depending on the top surface elevation, resulting in 73,071 grid blocks. The model was fully saturated with the top surface set using values from the National Water Table model (Westerhoff et al., 2018, Figure 2). The maximum height of the model surface was 805 masl and the minimum was 268 masl. The water table surface was a muted reflection of topography on average only 18 m below it, and therefore the numerically complex vadose zone was neglected. The top surface boundary was fixed at atmospheric pressure (10^5 Pa) and temperature (20°C). The model included a uniform basal heat source of 700 mW/m^2 (Bibby et al., 1995) (Figure 2a).

A range of models were computed that each ran for 1 Myr to simulate steady state conditions. The models explored the following (see Table 1):

- Bulk permeability was varied to find the range within which convective upflow occurs (models R1 to R5)
- A two-layer geological model was created with lower permeability in the bottom layer representing the basement greywacke (Alcaraz et al., 2012) (models R6 and R7)
- Representation of topography was explored by refining vertical grid block sizes from 250 to 100 to 50 m (Figures 2b–2d) (Table S1)
- The model limits were extended to include nearly the entire central TVZ (green rectangle in Figure 1), using the same model parameters as above but with an increased horizontal grid block size of 1 km by 1 km (models T1 to T4)

Each of these suites of models were run with and without topography for comparison (Table 1).

Modeled temperature results were compared with geothermal system locations inferred from electrical resistivity data (Bibby et al., 1995; Bertrand et al., 2012). Model fit was determined using the χ^2 test (SciPy, 2020) to check for a relationship between elevated and non-elevated temperatures (factor 1) and low and high resistivity (factor 2) for each model. A high χ^2 statistic means a stronger relationship and therefore a better model fit. Shallow DC resistivity data (Stagpoole & Bibby, 1998) was compared with temperatures at −125 masl. Deep MT resistivity data, where it was available (Bertrand et al., 2012), was compared with temperatures at −3,000 masl; it is not as spatially complete as DC resistivity data (Figures 3b and 3c). Elevated temperature was defined as greater than 30°C near the surface following New Zealand’s Resource Management Act (Ministry for the Environment, 1991), and as >100°C at −3,000 masl due to the geothermal gradient. Low resistivity zones were defined as areas of less than 30 Ωm following Rowland and Simmons (2012; Figure 2). Electrical resistivity detects a contrast between the highly conductive material
within geothermal fields in the TVZ and the resistive cold-water saturated volcanic rocks that surround the fields (Bibby et al., 2005).

4. Results and Discussion

Our results show that topography localizes convective fluid flow in an otherwise homogeneous medium. Upwelling model plumes reproduced the locations and trends of geothermal systems in the TVZ far better than expected (Figure 3a). The $\chi^2$ fit of elevated temperatures to low resistivity zones was greatly increased when topography was included compared to when it was not (Table 1, models R4 v R1). This held true even with a coarse, TVZ-scale model (Table 1, T3 v T1; Figure 4). Topographically driven upflow can explain the locations of 14 of the 19 known geothermal systems in our study area.

At 3 km depth, the $\chi^2$ goodness of fit between modeled temperatures and MT-derived system locations increased from 3 to 10 when topography was incorporated. Northeast-southwest and northwest-southeast trends formed, consistent with TVZ faulting (McNamara et al., 2019) and geothermal system locations (Figures 3b and 3c). There is a maximum of 537 m between the highest and lowest points in the model, which

Figure 2. TVZ model with three times vertical exaggeration. Grid blocks correspond to basement (gray) and volcanic deposits (tan). (a) Top surface follows the groundwater table and is overlain by the DC resistivity map (Stagpoole & Bibby, 1998). Black spheres correspond to hot springs, blue lines to rivers and lakes, black outlines to geothermal systems and purple box to the regional model area. Heat is input uniformly across the base of the model at 700 mW/m² (Bibby et al., 1995). Representations of water table elevation in the regional model can be seen with vertical block sizes of (b) 250 m, (c) 100 m, and (d) 50 m.
is equivalent to a 5.3 MPa difference in driving pressure. At −125 m, this is 66% of the maximum pressure, whereas at −3,000 m it is only 14%. It appears that when all else is equal, a topographic change of <600 m over 20 km distance is enough to focus large-scale convective upwellings beneath topographic lows.

There is a fairly narrow permeability range where convective upflow occurs with the high temperatures observed in the TVZ. If permeability is too high, fluid circulates rapidly and the model becomes uniformly cooler than 100°C (Table 1, model R2). If permeability is too low, hot fluid is trapped at depth and does not convect (Table 1, model R5). The permeability must therefore be greater than 10^{-15} m^2, but less than 10^{-13} m^2 (Table 1, models R3, R4). This is consistent with measurements and model estimations in New Zealand (Dempsey et al., 2011; Kaya & Widyanti, 2015; Kissling & Weir, 2005; Milicich et al., 2018; Ra-

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**Table 1**

*Fit of Model Temperatures to Resistivity Data Under a Range of Conditions. Models Without Topography (Red) are Shown as a Comparison to Those With Topography (Blue).*

| Model number | Model description | Topography | Permeability (m²) | \(\chi^2\) DC fit | \(\chi^2\) MT fit |
|--------------|-------------------|------------|------------------|------------------|------------------|
| R1           | Regional          | N          | 1.E-14 Baseline  | 0.04             | 3                |
| R2           | Regional          | Y          | 1.E-13 Baseline  | 108              | -                |
| R3           | Regional          | Y          | 1.E-14 Baseline  | 299              | 1                |
| R4           | Regional          | Y          | 5.E-15 Baseline  | 283              | 10               |
| R5           | Regional          | Y          | 1.E-15 Baseline  | -                | -                |
| R6           | Permeability contrast | N          | 1.E-14 Baseline  | 12               | 6                |
| R7           | Permeability contrast | Y          | 1.E-14 Baseline  | 210              | 25               |
| T1           | TVZ-scale         | N          | 1.E-14 Baseline  | 10               | 0.8              |
| T2           | TVZ-scale         | Y          | 1.E-14 Baseline  | 50               | 8                |
| T3           | TVZ-scale         | Y          | 5.E-15 Baseline  | 46               | 12               |
| T4           | TVZ-scale         | Y          | 1.E-14 Baseline  | 46               | 7                |

*Note.* Darker colored bars correspond to DC resistivity and temperatures at −125 m. Lighter colored bars correspond to MT resistivity and temperatures at −3000 m. Models R1-7 are regional models, while T1-4 are of the central TVZ (Figure 1) and have a coarser horizontal resolution of 1 km instead of 500 m.
Our results suggest that permeability is on the lower end of the modeled range, especially at depth. To explore parameters other than permeability, a value of $1 \times 10^{-14} \text{ m}^2$ was used as this is consistent with measured permeability and is where convective upflow occurs but computation time is still efficient.

The locations of upflow zones do not appear to be controlled by the deeper basement surface. Incorporating a permeability contrast between basement and cover rocks at varying depths representative of the TVZ (Alcaraz et al., 2012) only slightly changed the model fit (Table 1, models R4 v R7). Some individual features...
were better represented however, for example the plume at Ohaaki that extends from northwest to southeast as it shallows (Bertrand et al., 2013, Figure 3c).

4.1. Study Limitations

We deliberately created a highly simplified uniform model with a hot-plate heat source to explore the effects of topography on geothermal circulation. The TVZ is however highly variable in terms of rock properties, faulting, and inferred heat input at depth (Chambefort & Bignall, 2016; Rowland & Sibson, 2004; Wilson et al., 1995). These variations may affect the locations, stability, and intensity of geothermal systems (Chambefort et al., 2014; Kissling et al., 2009, 2018; Ratouis & Zarrouk, 2016). Our study does not preclude these mechanisms, which may dominate or act in tandem with effects of topography. Indeed, topographical offsets can be a result of faulting, but we have not included that explicitly here. One thing to note however is that rock properties, subsurface faulting and deep heat input are poorly constrained, while the water table is based on high-resolution digital elevation models of surface topography (LINZ, 2012; Westerhoff et al., 2018) and its effects can explain over 60% of geothermal system locations.

Figure 4. Numerical model results over the central TVZ area as shown in Figure 1. Red contours correspond to 30°C model temperature overlain on DC resistivity map (Bibby et al., 1995). Geothermal systems are outlined in gray. Mokai (MK), Te Kopia (TK), Horohoro (HO), Tikitere (TI), and Rotoma (RT) are geothermal systems that are not in topographic lows.
A topographically driven model cannot explain the locations of five of the geothermal systems. Tikitere, Mokai, Horohoro and Rotoma are in moderate topography, while Te Kopia is in a topographic high. Tikitere, Mokai and Horohoro are however close to modeled upflow zones (Figure 4). At Mokai, shallow lateral flow does displace surface features ∼5 km to the north into a topographic low at Waipapa Stream (Bibby et al., 1984).

Modeled convective flow patterns formed quickly and remained stable as long as topography remained consistent. Topography does change with time; however, the geology of the TVZ generally predates the formation of its geothermal systems (Rowland & Simmons, 2012; Wilson & Rowland, 2016). In addition, many of the geothermal systems coincide with the Waikato River or its tributaries (Ratouis & Zarrouk, 2016; Rowland & Sibson, 2004), which is thought to have remained along a narrowly defined route since at least the Whakamaru events 340–350 ka (Wilson & Rowland, 2016).

To run these models, geologic detail had to be balanced against computational time. Refining vertical grid spacing (Figure 2) increased computational time while slightly improving model fit to both shallow DC and deep MT resistivity data (Table S1). Overall patterns in upflow zones did not change and we therefore believe that even our coarsest models show a valid correlation between topographically driven upflow and geothermal system locations.

5. Conclusions

Our study shows that topographically driven convection can explain the locations of most geothermal systems in the central TVZ. Fourteen out of 19 known geothermal systems were correctly located beneath topographic lows within the model region. While we show the importance of topography in an area of high heat flow, presumably it is also an important consideration in lower temperature geothermal systems. However, this would require additional work to determine.

Evidence indicates that geothermal systems in the TVZ are long-lived and stable. Although this can be achieved in many ways (e.g., crustal structure, localized heat sources at depth), we show that the known topography is sufficient to localize geothermal upflows in an otherwise homogeneous medium. Topographic lows in the TVZ have remained relatively consistent over hundreds of thousands of years, potentially explaining the apparent stability of the geothermal systems.

Even at 3 km depth, inclusion of topography in our models improved agreement between the locations of convective upflow and low-resistivity zones derived from 3-D inversion of magnetotelluric data. This suggests that the influence of topography can be significant at depth, even with the moderate topographic relief of the TVZ. Therefore, while topography is not the only factor that controls or influences geothermal convection, its contribution is not as localized as is generally assumed and can be included precisely.

Data Availability Statement

Data are available through Bibby et al. (1995), V. M. Stagpoole and Bibby (1998), and Bertrand et al. (2012).

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