PROCEEDINGS, Thirty-Eighth Workshop on Geothermal Reservoir Engineering
Stanford University, Stanford, California, February 24-26, 2014
SGP-TR-202

Seismicity and Subsidence: Examples of Observed Geothermal Deformation Synergies from New Zealand

Chris J. Bromley

GNS Science, Wairakei Research Centre, Private Bag 2000, Taupo 3352, New Zealand

c.bromley@gns.cri.nz

Keywords: subsidence, induced seismicity, triggered seismicity, surface deformation

ABSTRACT

The patterns of triggered seismicity and subsidence effects from conventional geothermal operations are often irregular in their timing and location, and therefore difficult to predict, particularly in the absence of detailed knowledge of the local subsurface stress conditions, rock properties and permeability structure. Reinjection is usually identified as the principal cause of triggered seismicity, and ground subsidence effects are usually attributed to pressure decline from production. The real situation is more complex, however, and the two processes of subsidence and seismicity can be closely entwined. Both processes are products of subsurface stress changes acting on clays, rocks or fault surfaces, usually exhibiting anomalous geo-mechanical properties. Adaptive mitigation for adverse effects from either process calls for a coordinated approach, using injection management to control induced stress and strain changes, while considering the potential adverse effects of both deformation processes. This paper reviews several geothermal cases, and shows that interlinked mechanisms, involving temperature and chemical changes, as well as transient pressures, are implicated. Interlinked mechanisms such as 'slow-deformation' events or earthquakes, and 'seismicity-induced' subsidence, are probably more common in geothermal settings than we previously thought.

Examples from New Zealand of triggered seismicity favor a mechanism associated with the indirect effects of increased fluid flow. The flow is driven by pressure gradients through a fracture network, but seismic failure is triggered only on pre-existing, favorably-oriented fracture-networks, and can occur throughout the fracture network affected by moving fluid. The triggering mechanism can be local temperature, pressure or chemical transients, or local stress perturbations, unlocking asperities on stressed fractures.

Some subsidence and seismicity mechanisms require consideration of the transition between brittle and ductile behavior across a range of temperatures, pressures and rock types. Settlement can also originate from shaking of seismic origin and non-linear stress-strain relationships such as yielding. To simulate such interactions and deformation processes, what is required is a better conceptual understanding of deformation processes, and truly inter-coupled Thermal-Hydraulic-Mechanical-Chemical (THMC) modeling. Some of the more fundamental rock properties used in traditional reservoir simulation, such as permeability, porosity and stress state, which are usually treated as constant parameters in history matching and subsequent scenario predictions, are, in reality, time-variables, and this needs to be incorporated into the inter-coupled modeling.

1. INTRODUCTION

New Zealand's geothermal systems are mostly located in high-temperature, volcano-tectonic settings. Rates of tectonic deformation in these settings are relatively high, especially across the Taupo Volcanic Zone (TVZ), a region of active rifting, volcanism and seismicity, located in the central North Island, above the subducting Pacific Plate (Figure 1a). Background levels of natural micro-seismicity (MEQ) are relatively high, but locally quite variable and patchy. Felt earthquakes, that are of shallow origin and of magnitude range M_L 2 to 4, are relatively common, and a familiar experience to many inhabitants. Rates of anomalous ground surface deformation are also relatively high in some locations for a variety of reasons.

The rifting process stresses the brittle crust (about 12 km thick) which generally deforms seismically, that is, by brittle failure. The underlying wedge of mantle material (from ~12 km to ~150 km depth) is so hot, it deforms a-seismically (that is, by plastic or ductile deformation). The deeper subducting Pacific Plate (>150 km depth) is cooler and more rigid, and therefore again hosts tectonic earthquakes, but the effects of these, at the surface, are attenuated by the hot ductile layer. The rifting process also causes horizontal stretching of the ground surface (or extension) by about 5 to 20 mm/yr and regional subsidence averaging 3 mm/yr. The recent rates of tectonic movement in the TVZ, using data from continuous GPS stations (GEONET.org.nz) and repeat satellite radar information (differential InSAR), are described in Samsanov *et al.* (2011), and Hole *et al.* (2007). An example of a differential InSAR image along with changes in vertical and horizontal relative movement at cGPS sites in the northern TVZ is shown in Figure 1b. An example is given in Figure 2 of the history of relative changes in height observed at several cGPS sites at central TVZ sites (surrounding geothermal fields near Taupo).

Within the upper crust (at 5-12 km depth), and beneath some of the volcanoes, caldera structures and geothermal systems, there are inferred to be zones of partial melt or magma. The evidence is largely based on interpretation of MT surveys (eg Bertrand *et al.*, 2013), and natural-state geothermal modeling. Such hot bodies are inferred to exhibit ductile deformation including periods of inflation and contraction during and after episodes of magmatic fluid injection. They also exhibit anomalous seismic velocity, density, chemistry and thermal properties. Some region-wide models of the TVZ represent the hot ductile zone as a conductively-heated continuous layer, with plumes of circulating water rising above it and forming convective geothermal systems (eg Kissling *et al*., 2005).

The spreading rift and associated graben structures are in-filled (usually to about 1-3 km depth) by young erupted volcanic rock and sedimentary material which is generally of lower density, higher porosity and weaker strength. The shallower deposits in particular are generally poorly consolidated. The basement is mapped or inferred using gravity surveys and borehole geology, and consists mostly of greywacke, with some cooled intrusives and buried lavas (eg andesite). Such formations are relatively strong. If the temperatures are below about 370-400 °C, they deform under tectonic stress by fracturing, often generating MEQ and thereby creating or maintaining permeable faults and associated structures. These sustain natural fluid flow by counteracting the natural process of fracture sealing through hydrothermal mineral deposition and clay alteration. If temperatures are much higher the deformation is likely to be ductile or plastic, that is, aseismic.

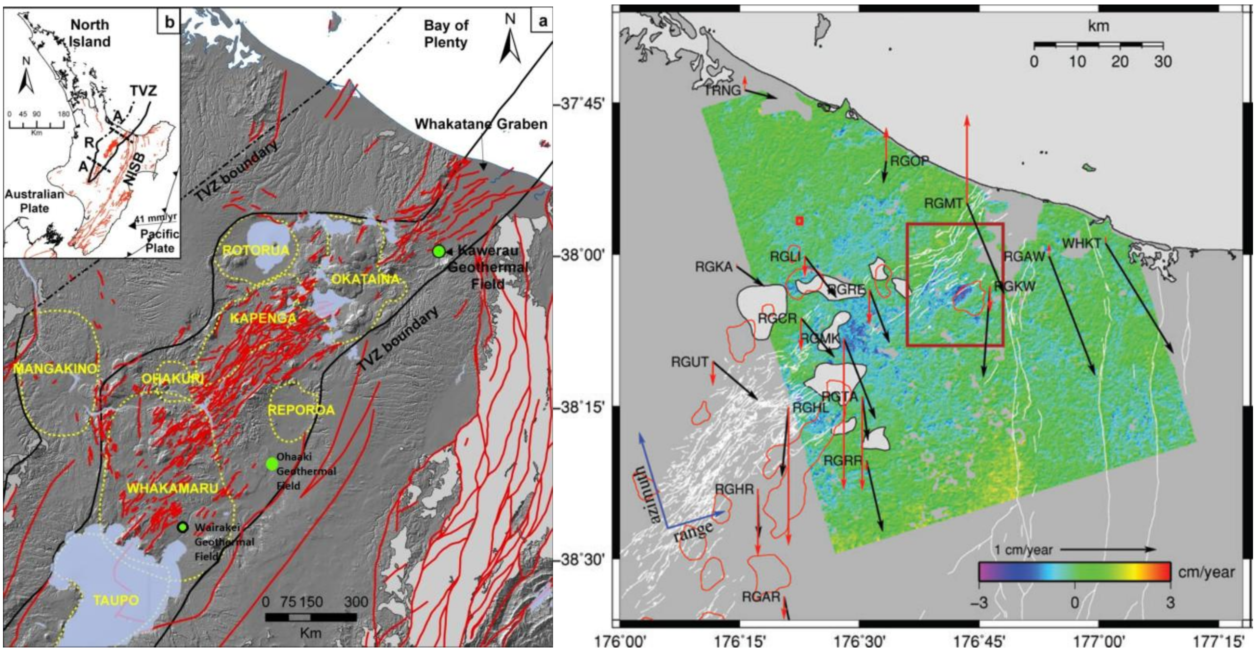

Figure 1: a) Map of Taupo Volcanic Zone, from GNS Science, showing mapped faults (red), inferred caldera boundaries (yellow) and outcropping basement greywacke (light grey); scale bar is 0-20 km; b) Recent tectonic deformation in Bay of Plenty region from InSAR deformation image (blue=subsidence), and continuous GPS vectors (black= horizontal, red=vertical), relative to the Tasman Plate (Auckland), for 2006-2010, from Samsonov et al. (2011).

Part of the TVZ infill consists of a sedimentary sequence of mudstones, siltstones and sandstones, for example, the *Huka Falls Formation* at Wairakei and Ohaaki (Figure 1a) or the older *Tahuna Formation* at Kawerau. These formations are found beneath many central and northern TVZ geothermal fields at intermediate depths of up to 600m, and they usually act as capping layers or aquitards to fluid flow. They are generally more porous (30% to 60%) and up to two orders of magnitude more compressible than the basement rocks (typically 5-10% porosity). They are therefore much more susceptible to compaction following pressure decline. The low permeability of the aquitard provides a delaying mechanism whereby pressure and temperature changes diffuse relatively slowly through them, postponing the subsidence effect.

Subsidence, as an environmental issue, has been thoroughly dealt with, especially in New Zealand, through submissions to geothermal resource consent hearings, operational consent conditions and monitoring outcomes. The issue of assessing and dealing with induced seismicity in fractured geothermal reservoirs is being addressed through international cooperation by the IEA-Geothermal Implementing Agreement (Bromley and Mongillo, 2008) and the International Partnership in Geothermal Technology (IPGT). Protocols, white papers, and predictive tools are under development though working group collaboration. It is suggested here, however, that there is perhaps a need to consider more carefully the data and mechanisms that link the two processes, particularly when building comprehensive Thermal-Hydraulic-Mechanical-Chemical (THMC) models for predicting future effects.

2. SUBSIDENCE

Subsidence leveling surveys have been conducted routinely across all the New Zealand producing geothermal fields, and the results from Wairakei, Tauhara, Ohaaki and Kawerau, where accumulated subsidence anomalies of more than 1m (or rates greater than 50 mm/yr) were measured, have been described in, for example, Bromley (2006), Allis *et al*, (2009), Bromley *et al* (2009), and Brockbank *et al*, (2011).

At Wairakei and Ohaaki, measurements on samples from relatively shallow deposits of mudstone, volcanic pyroclastics, hydrothermal eruption debris, or hydrothermally-altered breccia, show locally anomalous properties, i.e., relatively high porosity and clay content, and compressibility that can be about ten times greater than that of surrounding or deeper sediments. When tested, samples from such deposits sometimes reveal a property called 'yielding', whereby the compressibility increases a further 5- to 10-fold when subjected to an effective vertical stress change exceeding the yield stress. This yield stress can be within the range of in-situ vertical stress (that is, overburden weight minus pore pressure) experienced by these deposits over the duration of production-induced pressure decline. Weak, yielding formations have been found to be responsible for several subsidence bowls (1m to 15m of accumulated subsidence) within the Wairakei, Tauhara and Ohaaki geothermal fields (Bromley et al, 2013). These very weak formations consist of hydrothermally altered Huka Falls Formation and hydrothermal eruption breccia, at 30m to 300m depth. They have porosities of up to 65% and compressibility ranging from 2 E^{-3} MPa^{-1} to 130 E^{-3} MPa^{-1}.

Local subsidence anomalies in geothermal areas, such as at Crown Road, Tauhara, are often relatively small in area (<1 km^2) and sometimes their cause is attributable to the effects of shallow groundwater level decline acting on a deposit of hydrothermally-altered material (Bromley and Currie, 2003). In this particular case, after further investigation drilling, a 30m to 200m deep breccia deposit was found. It consists of an in-filled 11 ka hydrothermal eruption crater (Bromley et al, 2013). The results of comprehensive K_0 compressibility testing of core from within the Wairakei and Tauhara subsidence bowls (Pender *et al*., 2013) showed the effects of anomalous clay yielding at effective yield stresses within the range of in-situ vertical stresses. The anomalous deformation process is largely inelastic and mostly permanent.

At shallow levels (up to 100m depth) sometimes there exists a deposit of weak, unconsolidated alluvium. Such deposits vary significantly in thickness across relatively short horizontal distances and may have formed in buried topographic depressions such as paleo-river channels. They may be responsible for some elongated local subsidence anomalies, such as observed at Kawerau Geothermal Field. Several of these elongated anomalies located adjacent to the Tarawera River have relatively steep edges, implying a shallow origin, and appear to be connected, implying a source related to a buried channel. Based on borehole geology, this inferred channel appears to have been eroded into the underlying 320 ka Matahina ignimbrite at about 30-100m depth. A deep paleo-river channel could have been the result of a period of rapid fluvial erosion after the volcanic ignimbrite episode. Over time, this and subsequent erosional channels, may have been refilled by flood deposits, peats, reworked volcanic ash, and other sediments accumulated by a meandering river. Such alluvial deposits could include sequences of unconsolidated sands, clays and organic material, which are potentially highly compressible and subject to significant consolidation or settlement by a variety of mechanisms. Examples of mechanisms include vibration from earthquakes or heavy vehicles, oxidation and drying of organic material (peat), or compaction through declining ground water pressure. Locally, the Tarawera river bed has been steadily declining through erosion (at about 40 mm/yr), and the resulting average river level decline has transmitted to the surrounding groundwater, causing a similar long term decline, and providing a plausible natural mechanism for long-term subsidence.

Potential subsidence mechanisms at Kawerau, particularly those of relatively shallow origin, were also addressed in a thesis by Mackenzie (2012). Modeling of the historic and predicted subsidence contribution from deep geothermal operations, and the issue of subsidence mechanisms and effects has been an ongoing subject at four recent public resource-consent hearings (for Bay of Plenty Regional Council) to consider planned incremental increases in geothermal fluid extraction and reinjection.

An interpretation of the observed long-term (pre-production) subsidence rates at Rotokawa by Powell et al, (2011), suggests that natural-state (pre-1997) subsidence (2 mm/yr) could have been caused by mass removal from the reservoir through hydrothermal alteration. The main natural alteration process is depletion in silica, creating an increase in porosity. The silica is presumably re-deposited elsewhere in veins or discharged from hot springs. When accompanied by weakening of the surrounding rock matrix, this process leads to collapse of some (but not all) of the newly created pore space. The resulting formation consolidation causes surface subsidence. The deformation may take the form of small movements along hard-soft bedding-plane boundaries (typical of a volcano-sedimentary sequence) or reactivation of fault structures and the surrounding fault-gouge/damage zones within competent basement rock. This can occur either seismically or a-seismically (plastic deformation or creep). Post-1997 deformation at Rotokawa has largely been dominated by the local effects of shallow reinjection which peaked in 2005, causing a radiating ring of inflation (tumescence of up to +50 mm from rising injection aquifer pressure), followed by a central core of subsidence from the temporary effects of injection cooling (up to -50 mm/yr), combined with longer-term production-zone reservoir pressure decline.

2.1 Subsidence from Earthquakes and 'Creep'

Large earthquakes have been reported to cause co-seismic deformation, in the form of subsidence, within geothermal areas. These can be located up to several hundred kilometers from the earthquake epicenter. One example is the 2010, M 8.8, Maule earthquake just off the coast of Chile. It apparently caused subsidence of about 50 mm to 150 mm across five undeveloped geothermal fields, located high in the nearby Andes volcanic chain (Pritchard *et al*., 2013). Over a period of several weeks, spanning the earthquake, differential InSAR images reveal co-seismic (or immediately post-seismic) deformation across these volcanic-geothermal areas (containing hot springs, thermally altered ground and fumaroles). The anomalies were about 15 to 30 km long and 10 to 15 km wide, oriented north-south, parallel to the Andes chain. Although the authors postulate that the mechanism is related to a loss of geothermal fluid from earthquake shaking, through increased spring discharges to rivers, an alternative explanation is that the subsidence was largely caused by co-seismic shaking and consolidation of weakened, high-porosity, hydrothermal clay alteration. These thermal clays may alternate with stronger layers of deposited silica, which resist normal overburden consolidation, but pancake when fractured by large amplitude vibration from surface waves. If the pore-space is liquid-saturated, then a reduction in porosity (from subsidence) would be accompanied by increased hot-spring discharge. If vapor-filled, the increase in pore pressure from the subsidence would be dissipated by temporary increases in steam discharge from fumaroles. If the pore-space is largely air-filled, that is, above the groundwater table, then settlement could occur without expulsion of fluid.

At the Cerro Prieto geothermal field in Mexico, the subsidence history over 20 years was interpreted by Glowacka *et al.* (2000) in terms of the possible local effect of tectonic extension (several mm/yr), as well as pressure drawdown associated with production. Average subsidence rates within the field and across the nearby Imperial Fault (8 km NE of the production field) were of a similar magnitude (~100 mm/yr). Continuous monitoring of deformation across the fault revealed relative vertical movement that was not constant but concentrated during 'creep events' lasting several days and separated by months of quiescence. The timing of these events did not coincide with production or injection changes, nor with local micro-seismicity. In this case the principal subsidence driving mechanism, long-term, was still inferred to be deformation associated with geothermal fluid pressure drawdown, but the effects were interpreted to be intermittent strain release along the Imperial Fault regulated by visco-elastic processes. In seismology, another term for this phenomenon might be 'slow-slip-event' earthquakes. Such events have been well documented along the margin of the TVZ where the subducting Pacific Plate is sliding in a stick-slip fashion beneath the Tasman Plate.

At Kawerau, subsidence mechanisms can be grouped into three main types: a) deep tectonic movement (major earthquakes plus rifting of about 3 mm/yr); b) geothermal reservoir changes (pressure and temperature decline) amounting to about 10-20 mm/yr; and c) local compaction of buried shallow sediments by up to 50 mm/yr. In 1987, the M 6.3 Edgecumbe earthquake, which was located at 8 km depth and about 20 km NE of Kawerau, caused significant co-seismic subsidence within the geothermal field. Over a 4 month period, spanning the earthquake, about 250 ($\pm$ 120) mm of subsidence was recorded relative to a reference base station outside and NW of the field. This represents approximately 25% of the maximum accumulated subsidence that has been recorded since 1970 in local Kawerau anomalies. Some of the earthquake-induced subsidence can be attributed to local fault movement (that is, deep tectonic adjustment to stress changes), but some can also be attributed to differential settlement of sediments during the seismic shaking of the main-shock and after-shocks. This is based on the observed differences of up to 50 mm in subsidence between adjacent benchmarks, suggesting control by local variations in subsoil consolidation properties.

3. INDUCED SEISMICITY

Triggered (or induced) seismicity effects from New Zealand geothermal operations have been variable and irregular. At some developed systems increased levels of triggered micro-earthquakes have been detected, following changes to injection strategy, while at other systems there were no significant changes to normal levels of natural seismicity following injection or production changes. Understanding the reasons for these differences will help estimate the potential seismic risks, and predict the outcomes (beneficial or adverse) of different development strategies for different locations. Therefore, some of the primary objectives of ongoing induced seismicity research are to illuminate probable mechanisms, identify zones of potential fracture permeability enhancement from micro-seismic locations, and help provide possible mitigation options, if (and when) induced seismicity magnitudes and felt event rates exceed acceptable values (Bromley and Majer, 2012).

Pressure drawdown from large scale production of fluids at the first two geothermal projects commissioned in the late 1950's (Wairakei and Kawerau) did not initially cause any reported effects, in terms of the rate of felt earthquakes originating from within the fields (unlike subsidence, which started within a few years of operation commencement). Over time, however, an increased focus on fluid reinjection, the most commonly attributed cause of induced seismicity, has, in places, had an impact on local levels of micro-seismicity. In the mid 1980's (at Wairakei), initial injection trials were shallow (<0.5 km) and triggered some local low-magnitude micro-seismicity across a wide area, but only at high effective stimulation pressure (~5 MPa). Elsewhere in New Zealand, other, low-pressure, reinjection projects soon followed, with several changes in injection strategy over time (including injection depth, location, in-situ temperature, pressure and flow-rate). The projects included: Ohaaki (1989, deep to shallow), Kawerau (1992, shallow to deep), Rotokawa (1997, shallow to deep), Ngawha (1998, deep), Mokai (2000, shallow to deep), and Ngatamariki (2013, deep). A range of different induced seismicity responses has accompanied each of these changes in injection strategies.

At Ohaaki and Ngawha, natural seismicity rates are relatively low, and local induced seismicity (M>2) has not been detected, despite Ngawha's 100% peripheral injection to ~1km depth and Ohaaki's 70% peripheral shallow injection (<1km depth). At Kawerau, natural rates of seismicity are high (average two felt events/month), but there have been no obvious triggered events associated with production or injection changes, including an expansion and transition in 2008 from shallow (0.4km) infield to mostly deep (2km) peripheral injection. At Wairakei, Rotokawa, Ngatamariki and Mokai (adjacent systems), natural seismicity is moderate, but locally variable. Increasing deep reinjection (since 2006) has, in places, triggered moderate levels of micro-seismicity within inferred fault zones conducting fluid between injection and production sectors. The maximum magnitude recorded has been M_L 3.5 (at Rotokawa), but most are well below M_L 2.5, and felt seismicity effects have not been an issue with the local rural inhabitants, who are familiar with similar-sized natural events. There is some evidence of cooling contraction increasing permeability with time, and micro-seismicity appears to be constrained by fault-controlled flow inhibitors.

In addition to the permanent seismic array network across New Zealand, maintained for public information (www. GEONET.co.nz), there are several, specifically geothermal, local seismic array networks, consisting of 5 to 13 instruments. These have been established by the operators (Contact-Energy and Mighty-River-Power) at Wairakei, Rotokawa, Ngatamariki, Mokai and Kawerau geothermal fields. They monitor for locally seismicity, both induced and natural. The semi-permanent local networks are boosted, at times, by short-term studies using temporary deployments.

The results of monitoring at Rotokawa from 2008 to 2012 were recently presented in Sherburn et al, (2013) and Sewell et al, (2013). More than 1000 events of M_L>0.8 and 50 of M_L>2 are located in the field, ~70% within a sharply bounded volume of about 1.5 km^3 located between SE reinjection and NW production sectors. A NE trending central bounding structure, the 'Central Field Fault', appears to have limited the cross-strike flow of injected fluids driven by the pressure gradient between these sectors, whilst allowing fluids to disperse along-strike, within fault-controlled permeable fractures, and presumably throughout the volume of reservoir marked by increased induced seismicity. The bulk of this seismicity appears to be related to changes in flowrate into one of the more permeable of the four available injection wells (RK20). The base of the zone of induced seismicity and stimulated permeability is at about 3.5 to 4 km depth. This is thought to indicate the probable limit of natural fracture permeability within the reservoir, where conductive heat-flow predominates, temperatures climb rapidly, and high-temperature ductile conditions are inferred to commence. The original temperature difference between the injected fluids and the receiving formation at 2-2.5 km depth was ~200^{o}C, so most of the induced MEQ near the injection wells are thought to have been triggered by thermal stress transients related to cooling of fracture surfaces. Some evidence of cooling contraction and widening of fracture apertures is revealed by an observed improvement in injectivity over time (t^n dependence, where n=0.4 to 0.7), as discussed in Grant *et al.* (2013). Permeability is sufficient that pumped injection is not required, so the pressure increase in the injection sector is relatively low (~1 bar). Never-the-less, injection-induced pressure transients that propagate through the interlinked fracture network may be responsible for triggering of seismicity at greater radial distances, particularly if the reservoir rock (predominantly andesite and basement greywacke) is naturally critically stressed and favorably-oriented fractures are close to failure.

At Wairakei, Sepulveda *et al* (2013) recently reported on the results of a local 9 to 13 station borehole seismometer network that has been deployed since 2009. This is in addition to long term monitoring at a station ("Poihipi") that has been in place since 1997.

The results have been used to try to assess the levels of natural and induced seismicity, and constrain the depth of fluid circulation occurring on stressed fracture systems. For Wairakei, most of the infield seismicity occurs within borehole depths of about 1.5-2 km, and at about 4-6 km depth along a NE trending inferred fault zone beneath the Te Mihi upflow of high temperature fluid; 95% is < 6.5 km depth, labeled the top of the brittle-ductile transition zone. Outside the field, particularly along the nearby Taupo fault belt, the most common MEQ depths are at 3 km and 6 km (95% < 9.6 km depth). This was inferred by the authors to indicate an up-doming of the high-temperature ductile zone beneath the geothermal system.

A common MEQ observation across the geothermal systems is their b-slope calculation (the Gutenberg-Richter linear relationship on a histogram plot of the log of the cumulative number of events versus magnitude). At Kawerau, for 20 years of monitoring, across an area of 634 km^2, the calculated b-slope is 1.46 over a magnitude range of M 2.2 to 3.2. This is relatively high compared to normal tectonic settings (b-slope=1). There is an even steeper decline in the number of events of M >3.5, implying a natural deficit of larger events in this high temperature setting. Although the depths are poorly constrained for many of the events, the majority in the Kawerau field occur within the upper 5 km, whereas seismicity occurring outside the field (for example, along the Okataina fault belt 20 km to the NW, Figure 1a) is predominantly at 5 to 12 km depth and exhibits a normal tectonic b-slope of 1. At Rotokawa, for seismicity inside the system, the b-slope also has a high value of 1.44, for M = 2.2 to 3.5 (Sherburn *et al*, 2013) and depths are similarly constrained to <4 km. When considered together with the results from Wairakei, these observations support the inference that MEQ originating from within high temperature TVZ geothermal systems, whether natural or induced, are likely to be shallower in depth. They are consequently constrained in terms of the relative frequency of large magnitude events, either because of geometric constraints on the potential size of stressed fractures that might be available for triggered failure, or arguably because thermally-derived stress triggers (from cooling) might operate across a reduced range of fracture dimensions compared to pressure-derived triggers. Either way, the consequences are comforting in terms of geothermal induced seismicity risk from larger events.

The vexing question as to whether observed seismicity is natural or induced cannot yet be resolved for most of these cases. Triggering mechanisms such as stress perturbations caused by temperature and pressure transients, or chemical dissolution and deposition, can operate throughout the permeable fracture network where fluids circulate, so induced events cannot be arbitrarily distinguished on the basis of proximity to injection wells. Natural triggering mechanisms, related to seismic swarm behavior and observed clustering of events in time and space, obviously operate throughout the seismically active parts of the TVZ. They are also likely to be prevalent within and adjacent to most of these geothermal systems.

4. DEFORMATION LINKAGES

In places, it is clear from the examples given above, that the two processes of induced seismicity and subsidence are closely entwined. Both can be products of subsurface stress changes, which act in various ways on clays, rocks or fault surfaces exhibiting anomalous geo-mechanical properties. Interlinked mechanisms, involving temperature changes and the effects of chemical dissolution/deposition, as well as transient pressures, are implicated. Consideration of source depth might unravel some of these.

Mossop (2001) specifically recognized the potential linkages between mechanisms responsible for ground subsidence and induced seismicity at The Geysers Geothermal Field, California. Poro-elastic stressing due to steam pressure decline and associated thermal contraction due to cooling, are triggering mechanisms for shear reactivation on fractures, leading to micro-seismicity and subsidence. Conversely, however, in 1997 and 2003, subsidence rates slowed within a 1-2 km radius of injection wells in response to increasing cold water injection at 1-2 km depth, while the micro-seismic event rate increased. The dominant mechanism, in this case, is thought to be increasing steam pressure from boiling of injected fluid.

Rutqvist *et al* (2013) also observed a linkage between seismicity and deformation during the 2011-2013 northwest Geysers EGS fracture stimulation project. Although they were able to simulate, using coupled THM modeling, the extent of the calculated fracture stimulation zone, and match the field observations over the first few months of injection, they also observed that surface deformations and MEQ evolution showed more heterogeneous behavior than their model had predicted. This was inferred to be a result of more complex geology, such as minor faults and fracture zones that are not accounted for in the model, but may be important for analysis of long term energy production, as well as predicting future deformation.

4.1 Compressibility and Phase Change

An important consideration in conceptual models of the underlying mechanisms behind seismic failure or deformation processes within high temperature geothermal systems is the effect of fluid phase changes. Intuitively, it would seem that any boiling process linked to pressure decline, or any re-saturation of a steam-zone caused by injection or natural liquid recharge, should cause a change in the compressibility of the fluid in the pore space, and therefore a change in the effective compressibility or seismic failure criteria for the host formation. In reality, however, a change in pore fluid compressibility does not directly change the associated formation strength; what might change is the pore fluid pressure and temperature, which will, in turn, affect the deformation process through a change in effective stress or vertical load (in the case of subsidence), or shear stress (in the case of MEQ). These fluid property changes are accounted for in a multi-phase reservoir simulator such as TOUGH2, and there is no need to incorporate in a simulation model changes in strength parameters when such phase changes occur. However, indirect effects that accompany phase changes may cause rock strengthening or weakening over time. These include acid-condensate hydrothermal alteration into weak clay, or boiling deposition of a stronger mineral such as quartz. Hence, there is a need for fully-coupled THMC modeling.

4.1 Creep and Vibration

Natural deformation events within geothermal systems, such as creep, may be similar to 'slow-slip-event' earthquakes (SSE). These have recently been observed on 'hot & sticky' subduction plate interfaces. Such deformation processes are probably quite common in high temperature geothermal settings, particularly in tectonically active areas (eg. Figure 2). Also, as noted above, shallow consolidation can be induced by vibration from passing seismic Rayleigh waves. Seismically active regions coinciding with shallow geology consisting of weak sandy sediments are therefore susceptible to seismic-induced subsidence.

4.2 Liquefaction

The process of liquefaction is commonly associated with consolidation of saturated sandy sub-soils and sediments particularly from relatively large earthquakes (M_L>5). Ground subsidence is also frequently associated with liquefaction observations. The 1987, M_L 6.3, Edgecumbe earthquake, near Kawerau geothermal field, is a local TVZ example. Some underground explosions (equivalent in energy release to magnitude M_L>2) have also induced soil liquefaction at radial distances of less than 1 km (Wang et al., 2006). These authors present a simple, empirically-deduced, liquefaction limit relationship (for saturated, unconsolidated sandy sub-soils) using the distance to an earthquake's hypocenter, and its magnitude:

$$\log R\max = 2.05(\pm 0.1) + 0.45M$$

where, *R* max is the liquefaction limit in meters (the maximum distance from earthquake hypocenter to a potential liquefaction site) and *M* is the earthquake magnitude. Hence, for shallow seismic events within the magnitude range of 2 to 4, and hypocenter distances of 1 km to 7 km respectively, this relationship suggests that vibration from seismically-generated surface waves may be of sufficient energy intensity to cause local liquefaction, and therefore consolidation, of saturated sandy sub-soils. Hence, locally felt earthquakes of this magnitude range, which are quite common in many parts of the TVZ, might produce surface waves and vibration of sufficient amplitude to cause ongoing settlement of susceptible formations at rates that, when averaged over several years, are comparable to subsidence from other mechanisms.

5. CONCLUSIONS

Examples of geothermal injection triggered seismicity in New Zealand favor a mechanism associated with the indirect effects of increased fluid flow. These are driven by pressure gradients through a fracture network, but trigger seismic failure only on pre-existing, favorably-oriented fracture-networks, through local temperature, pressure or chemical transients, or by local stress perturbations unlocking asperities on stressed fractures.

Some subsidence and seismicity mechanisms require consideration of the transition between brittle and ductile behavior across a range of temperatures and rock types. Settlement can also originate from shaking of seismic origin and non-linear stress-strain relationships such as yielding.

Injection management is the preferred adaptive tool for mitigating adverse effects from all of these processes. The purpose is to control or limit induced stress and strain changes without losing the long-term benefits of permeability enhancement.

The potential of joint micro-seismicity and deformation modeling and interpretation is significant. As a field management tool to optimize make-up and step-out drilling, as well as to better predict the effects of changes in injection strategy, such advances could prove invaluable. Further work on joint geophysical imaging, joint deformation interpretation, and advanced data analysis (e.g. seismic tomography, MEQ focal mechanisms, shear-wave splitting, THMC modeling, etc.) is eagerly anticipated.

REFERENCES

Allis, R., Bromley, C.J., Currie, S.: Update on subsidence at the Wairakei-Tauhara geothermal system, New Zealand, *Geothermics*, 38(1): (2009), 169-180.

Bertrand, E.A., Caldwell, T.G., Hill, G.J., Wallin, E.L., Bennie, S.L., Cozens, N., Onacha, S.A., Ryan, G.A., Walter, C., Zaino, A., Wameyo, P.: Magnetotelluric imaging of upper-crustal convection plumes beneath the Taupo Volcanic Zone, New Zealand, *Geophysical Research Letters*, 39(2), L02304, doi:10.1029/2011GL050177 (2012).

Brockbank, K.; Bromley, C.J.; Glynn-Morris, T.: Overview of the Wairakei-Tauhara subsidence investigation program, *Proceedings*, 36th Workshop on Geothermal Reservoir Engineering, Stanford University, Stanford, CA (2011), p 68-74.

Bromley, C.J.: Predicting subsidence in New Zealand geothermal fields: a novel approach, *Transactions* (Geothermal Resources Council), 30 (2006), 611-616.

Bromley, C.J., Currie, S.: Analysis of subsidence at Crown Rd, Taupo: a consequence of declining groundwater. *Proceedings* 25th New Zealand Geothermal Workshop, (2003), p113-120.

Bromley, C.J., Majer, E.L.: Geothermal induced seismicity: risks and rewards, *Proceedings* 34th New Zealand Geothermal Workshop, 19-21 November (2012), 7 p.

Bromley, C.J., Mongillo, M.A.: Geothermal energy from fractured reservoirs: dealing with induced seismicity. *OPEN energy technology bulletin*, 48: (2008), 7 p.

Bromley, C.J., Currie, S., Manville, V.R., Rosenberg, M.D.: Recent ground subsidence at Crown Road, Tauhara and its probable causes, *Geothermics*, 38(1): (2009), 181-191.

Bromley, C.J., Brockbank, K., Glynn-Morris T., Rosenberg, M., Pender M., O'Sullivan, M., Currie S.: Geothermal subsidence study at Wairakei-Tauhara, *Journal of Geotechnical Engineering*, Volume 166, Issue 2, Special Issue on Renewable Energy, (2013), 22p.

Glowacka, E., Gonzalez, J. & Nava, A.: Subsidence in Cerro Prieto Geothermal Field, Baja, California, Mexico, *Proceedings* World Geothermal Congress, Kyushu-Tohuku, Japan, (2000), p591-596.

Grant, M.A., Clearwater, J., Quinão, J., Bixley, P.B., Le Brun, M.: Thermal stimulation of geothermal wells : a review of field data, *Proceedings*, 38th Workshop on Geothermal Reservoir Engineering, Stanford University, Stanford, CA (2013), 7p.

Hole, J.K., Bromley, C.J., Stevens, N., Wadge, G.: Subsidence in the geothermal fields of the Taupo Volcanic Zone, New Zealand from 1996 to 2005 measured by InSAR, *Journal of Volcanology and Geothermal Research*, 166(3/4): (2007), 125-146.

Kissling, W.M. and Weir, G.J.: The Spatial Distribution of the Geothermal Fields in the Taupo Volcanic Zone, New Zealand, *Journal of Volcanology and Geothermal Research*, 145, (2005), 136-150.

Pender, M., Ramsay, G., Glynn-Morris, T., Lynne, B., Bromley, C.J.: Rock compressibility at the Wairakei–Tauhara Geothermal Field, New Zealand, *Proceedings* of the Institution of Civil Engineers, *Geotechnical Engineering*, 166(2), (2013), 224-234.

Powell, T.: Natural subsidence at the Rotokawa Geothermal Field and implications for permeability development, *Proceedings*, 33rd New Zealand Geothermal Workshop, Auckland, (2011), 4p.

Pritchard, M.E., Jay, J.A., Aron, F., Henderson, S.T., Lara, L.E.: Subsidence at Southern Andes volcanoes induced by the 2010 Maule, Chile earthquake. *Nature Geoscience* 6, (2013), p632-636.

Mackenzie, H.T.: Determining the origin of localised subsidence features in the Kawerau Geothermal Field, Bay of Plenty, New Zealand, MSc thesis, Canterbury University, http://hdl.handle.net/10092/6470.

Mossop, A., Seismicity, subsidence and strain at The Geysers Geothermal Field, PhD Dissertation, Stanford University, (2001).

Rutqvist, J., Dobson, P.F., Garcia, J., Hartline, C., Jeanne, P., Oldenburg, C.M., Vasco, D.W., Walters, M.: The Northwest Geysers EGS Demonstration Project, California: Pre-stimulation Modeling and Interpretation of the Stimulation, *Mathematical Geology*, DOI 10.1007/s11004-013-9493-y (2013).

Samsonov, S., Beavan, J., Gonzalez, P.J., Tiampo, K., Fernandez, J.: Ground deformation in the Taupo Volcanic Zone, New Zealand, observed by ALOS PALSAR interferometry, *Geophysics Journal International*, 187, (2011), 147 – 160.

Sepulveda, F. Andrews, J., Alvarez, M., Montague, T. and Mannington, W.: Overview of Deep Structure Using Microseismicity at Wairakei, *Proceedings*, New Zealand Geothermal Workshop, Rotorua, New Zealand (2013).

Sewell, S. M., Cumming, W., Bardsley, C.J., Winick, J., Quinao, J., Wallis, I.C., Sherburn, S., Bourguignon, S., Bannister, S.: Microseismicity at Rotokawa Geothermal Field, 2008 to 2012, *Proceedings*, 35th New Zealand Geothermal Workshop, Rotorua, (2013), 8p.

Sherburn, S., Bourguignon, S., Bannister, S., Sewell, S., Cumming, W., Bardsley, C., Quinao, J., Wallis, I.: Microseismicity at Rotokawa Geothermal Field, 2008 to 2012, *Proceedings*, 35th New Zealand Geothermal Workshop, Rotorua, (2013), 5p.

Wang, C., Wong, A., Dreger, D.S., Manga, M.: Liquefaction Limit during Earthquakes and Underground Explosions: Implications on Ground-Motion Attenuation, *Bulletin of the Seismological Society of America*, Vol. 96, No. 1, (2006), 355-363.

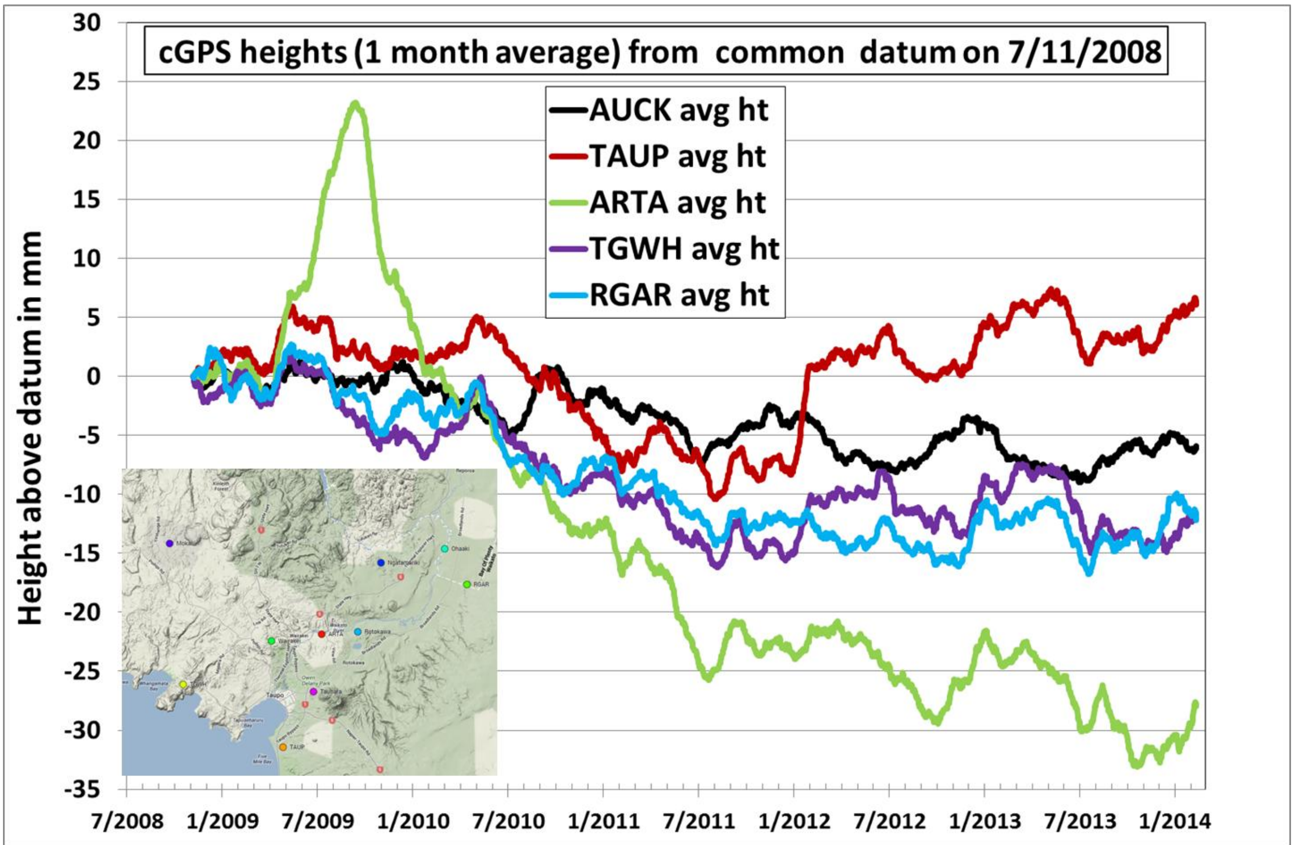

Figure 2: Changes in continuous GPS heights (1 month moving average) from several GEONET sites surrounding the Wairakei, Tauhara, Rotokawa and Ohaaki geothermal systems. Some consistent regional variations in deformation (~7 mm) across central TVZ, are presumed to be transient tectonic 'creep' events of several months duration. Also shown is an outfield injection-induced elastic response to a brief pressure increase in 2009 (at ARTA), and long term differences in subsidence rate.