Subsidence at The Geysers geothermal field, N. California from a comparison of GPS and leveling surveys

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Abstract. Between 1994 and 1996 three GPS surveys were conducted in The Geysers region of northern California. Our aim was to constrain models of the stresses and strains induced by geothermal power production in that region. Each survey spanned The Geysers geothermal field and consisted of typically 40 monuments. These monuments had been previously employed in a series of first order leveling surveys during the 1970's. This earlier study had determined that The Geysers region was subsiding, with a maximum rate of 0.048 ± 0.0055 m/yr between 1973 and 1977. In order to be able to directly compare the leveling and GPS surveys we transform them to the same reference frame using the GEOID 96 geoid model. For the period 1977–1996 we determine a maximum susbsidence rate of 0.047 ± 0.002 m/yr. We then model this subsidence using a series of point sources of contraction and find their optimal configuration by applying the random cost method and the F–test. The minimum volume strain that we find consistent with this subsidence is approximately 5×10^{-4}. Such a strain cannot be explained by a thermoelastic mechanism, but does seem to be consistent with poroelastic deformation and a quasi-static reservoir bulk modulus, K, of $\leq 3.6 \times 10^{9}$ Pa.

Introduction

The Geysers geothermal field is situated in the coast ranges of northern California. It is the largest producer of geothermal power in the world. At its peak in the mid-1980's some 2 GW of power were generated here, entailing the extraction of vast quantities of steam. Power production has since declined as steam pressure within the reservoir has fallen from an initial 3.5 MPa to as low as 1.2 MPa by 1988 [*Barker et al.*, 1992].

The steam producing reservoir itself is a highly fractured volume of Franciscan greywacke and Quaternary silicic intrusives, the latter known as the felsite, capped by a 1–3 km layer of, low permeability, metamorphic melange [*Thompson*, 1992]. The top of the reservoir lies on average 1 km or so below sea level, but ranges from 0.3 km above sea level in the central south east of The Geysers field to 2 km below sea level (bsl) at the edges. The bottom of the reservoir is poorly constrained, but is estimated to be typically 2 – 3 km bsl and as much as 4 km bsl in the central north west part of the reservoir [*Williamson*, 1992]

The region is presently deforming as evidenced by the fact that it is one of the most seismically active regions in northern California [*Hill et al.*, 1990]. Vertical surface deformation was measured, during the 1970's, by a series of first order leveling surveys across The Geysers [*Lofgren*, 1981]. The land surface above the geothermal field was observed to be subsiding. The maximum relative subsidence with respect to a chosen fixed site (monument Y 626), located some 20 km from the reservoir, was 0.192 ± 0.022 m between 1973 and 1977; a mean rate of 0.048 ± 0.0055 m/yr. The greatest subsidence appeared to be centred on the area of most active steam extraction during that time.

Lofgren [1981] suggested a causal relationship between the decline in reservoir steam pressures and surface subsidence. However, Denlinger *et al.*[1981] noted that the modest reduction in steam pressure, $\Delta P \approx 1 \times 10^{6}$ Pa, combined with the large bulk modulus determined from seismic data, $K_d \approx 3 \times 10^{10}$ Pa, was not consistent with the observed subsidence. They considered the strain to be due to a combination of thermoelastic and poroelastic deformations where the major component was thermal. The origin of these thermoelastic strains are as follows. Most of the reservoir water is stored as a liquid phase within the rock matrix porosity. The liquid water is first flashed into steam and then extracted via the reservoir fracture network. This phase change absorbs large amounts of heat and so lowers the reservoir temperature. The cooling reservoir contracts and this is observed at the surface as subsidence.

Resurveying with GPS Receivers

In 1994, 1995 and 1996 a number of the existing leveling monuments were surveyed using GPS receivers. Each survey included 30 – 40 locations which were generally occupied twice; positioning errors at one standard deviation were typically 3 – 5 mm horizontal and 15 mm vertical. However, GPS and leveling survey heights are not directly comparable. Leveling measures elevation with respect to a geoid based, orthometric, datum, whereas GPS elevation is with respect to an idealised ellipsoidal reference frame. In this instance the 1977 leveling survey was adjusted to the NGVD 29 datum and the GPS heights were determined relative to the WGS 84 reference ellipsoid. For comparison to be made between the leveling and GPS results they first have to be transformed into the same coordinate system. This transformation can be directly determined given simultaneous leveling and GPS surveys. In this case though, more than 10 years of deformation separates the two methods.

The problem of converting geodetic reference frames is the subject of ongoing research at the National Geodetic Survey (NGS). The present state of the art geoid model is GEOID 96 [*Milbert and Smith*, 1996]; this refers the height of the NAVD 88 geoid with respect to the NAD 83 ellipsoid and so allows conversion between these two reference frames. An additional transformation from NGVD 29 to NAVD 88 coordinates is required for the Geysers leveling data and is achieved by applying NGS's VERTCON model. The difference between the NAD 83 and WGS 84 ellipsoids is insignificant ($< 10^{-6}$m for this study) compared to the errors in the GPS data, and so no further transformation is necessary. It should be noted that the GEOID96 and VERTCON models are themselves inexact and so increase our data uncertainties. For the short baselines considered here the combined transformations introduce

Paper number 97GL51792.
0094-8534/97/97GL-51792$05.00

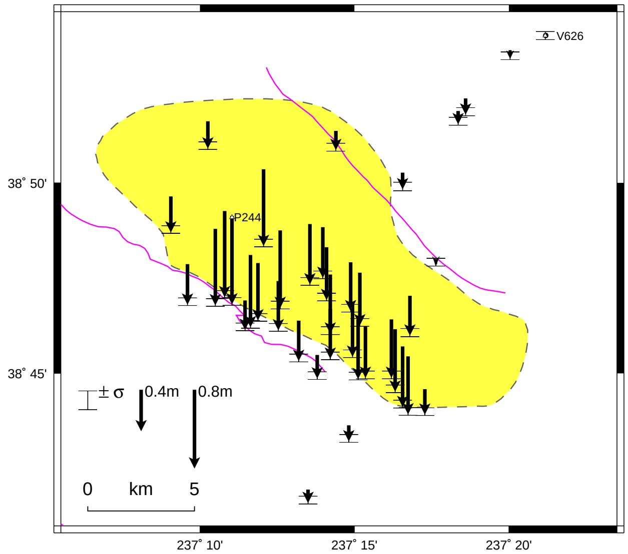

Figure 1. Subsidence between the 1977 leveling and 1996 GPS surveys, holding V626 fixed. Note that the tail–end of each arrow marks the associated monument location and that the $\pm\sigma$ error bars are drawn to the same scale as the subsidence vectors. Reservoir boundary from the California Division of Oil, Gas and Geothermal Resources.

a more or less constant error of 0.03 m to each height.

Subsidence Between 1977 and 1996

The height changes between 1977 and 1996, relative to site V 626, are shown in figure 1. Here the arrow lengths scale with elevation change, those pointing southwards indicating subsidence. The error bars show the range $\pm 1\sigma$. We chose V 626 as our fixed point because it is the furthest site from the geothermal field and because little significant vertical motion was observed there during the earlier leveling surveys. The subsiding monuments seem to be well bounded by the known extent of the geothermal reservoir. Little relative subsidence is observed for sites outside the reservoir to the north-east and south-west which appears to confirm the reference frame transformations and our choice of fixed point.

The maximum measured subsidence is 0.90 ± 0.04 m, consistent with a rate of 0.047 ± 0.002 m/yr. This is at monument P(1)244, approximately 2 km north of the site of maximum subsidence observed between 1973 and 1977 (T1244 not surveyed in 1996, hence not shown in figure 1).

Modeling the Subsidence

Subsidence indicates volume contraction within the reservoir, consistent with both the poroelastic and thermoelastic deformation mechanisms suggested by Lofgren [1981] and Denlinger *et al.* [1981] respectively. It seems appropriate, therefore, to attempt to model the surface deformation with volume change at depth. Our models are built from idealised point sources of volume change, often referred to as Mogi sources [*Mogi*, 1958], imbedded within an elastic half space. These give a good approximation to roughly equi-dimensional bodies, within the crust, undergoing uniform volume change. The location and intensity of the Mogi sources are found by optimising the fit of the predicted subsidence with that observed, using the random cost method described in Murray *et al.* [1996]. In reality, the volume contraction is distributed throughout the reservoir. The point contraction sources serve simply to identify the locations of maximal volume change.

An optimally located single Mogi source achieves a poor fit with the data and is located far deeper than we would expect for a source associated with the geothermal reservoir. Increasing the number of sources to two gives a much better fit to the data and both sources are located at depths more consistent with volume change within the reservoir. The improvement in fit yields a large reduction in the normalised sum of squared residuals. Here residual refers to the difference between observed and predicted elevation change at a site. The normalised sum of squared residuals, NSSR, provides a measure of how well we fit all the data.

We can carry on adding Mogi sources to our model and improving the fit to the data. However, each additional source must yield a statistically significant improvement in fit otherwise we're merely building an unnecessarily complex model. To measure the significance of the reduction in NSSR versus the increase in complexity due to the addition of an extra Mogi source we used an F–test. For the case of going from one to two sources a significance greater than the 99th percentile was found. Similarly the addition of a third Mogi source yielded an improvement with a probability greater than the 95th percentile. The addition of a fourth Mogi source caused an increase in the reduced χ^2 value (see Table 1) indicating that no significant improvement was achieved.

Table 1 gives the normalised sum of squared residuals and reduced χ^2 for the best fitting one, two, three and four Mogi source models. The location and volume reduction of each of these sources are also given. Note the large reduction in NSSR in going from one to two sources and the slight increase with going from three to four sources.

Figure 2 shows the residuals for the best fitting, three Mogi source, model. The Mogi sources themselves are represented by circles, the radii of the circles scale with the volume reduction. Note that the residuals are, in general, comparable

Table 1. *Location and intensity of the Mogi sources for the optimised cases of 1, 2, 3 or 4 sources and the associated normalised sum of squared residuals, NSSR, and the Reduced* χ^2*,* χ^2_R*. Depth, Z, given as below sea level after correcting for the median monument elevation of 850 m above sea level.*

°N	°W	Z (m)	Δ V (m^3)
1 Mogi Source: NSSR = 667.9, $\chi^2_R = 19.644$			
38.8063	122.7927	8970	$-3.18 \pm 0.19 \times 10^8$
2 Mogi Sources: NSSR = 97.59, $\chi^2_R = 3.2530$			
38.8218	122.8049	3270	$-7.06 \pm 0.21 \times 10^7$
38.7671	122.7340	2910	$-3.45 \pm 0.16 \times 10^7$
3 Mogi Sources: NSSR = 44.63, $\chi^2_R = 1.7165$			
38.8227	122.8065	3050	$-6.42 \pm 0.14 \times 10^7$
38.7622	122.7274	1720	$-1.55 \pm 0.07 \times 10^7$
38.7869	122.7576	1690	$-9.53 \pm 0.75 \times 10^6$
4 Mogi Sources: NSSR = 42.24, $\chi^2_R = 1.9200$			
38.8240	122.8091	3080	$-7.10 \pm 0.25 \times 10^7$
38.7622	122.7274	1720	$-1.55 \pm 0.07 \times 10^7$
38.7872	122.7581	1740	$-9.82 \pm 0.78 \times 10^6$
38.8367	122.8286	160	$+6.05 \pm 2.09 \times 10^6$

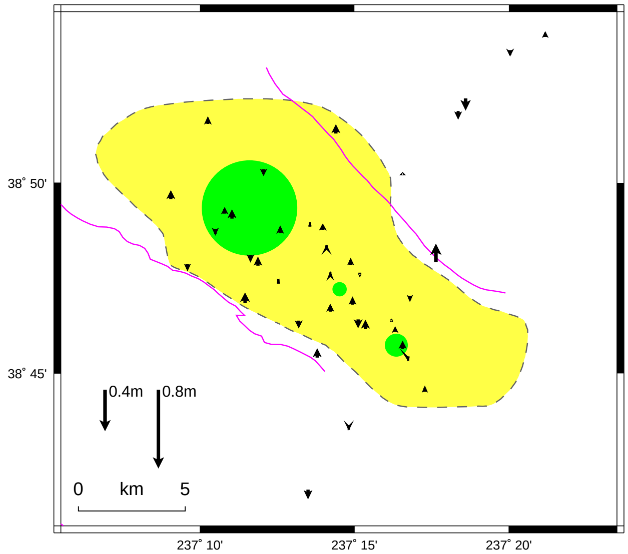

Figure 2. The best fitting Mogi source model and associated residuals. The mapped Mogi source radii are proportional to the source volume change.

to the combined measurement and conversion errors in the observed subsidence, *i.e.* are about the same as the error bars in figure 1. This would suggest that our best Mogi type model is apparently as good a solution to the data as could be expected from any type of model. It should be noted though that the true volume strain in the reservoir is continuous, not localised to discrete points, and hence there are other solutions, using distributed volumetric strain, that will give equally good results.

Deformation Mechanisms

The areal extent of the geothermal reservoir is fairly well constrained and covers an area of, at most, 10^8 m^2 (100 km^2). The reservoir thickness is less well constrained but is generally estimated to be less than 2×10^3 m or so [*Majer et al.*, 1992]. contraction of the best fitting triple source model (and the other models too) is about 10^8 m^3. Hence we see a minimum volume strain, ϵ_{kk}, of approximately 5×10^{-4}. An absolute lower bound on the volume change is the volume of subsidence itself. This we determine by finding a best fitting subsidence *surface* which we constrain to fall rapidly to zero outside of the reservoir boundary. The result is a minimum volume change of 7.7 $\times 10^7$ m^3, indicating that any physically reasonable model will give a similar minimum value of ϵ_{kk}.

For the case of thermoelastic strain we can relate ϵ_{kk} to a reservoir temperature change, ΔT, via the volumetric coefficient of thermal expansion, α_v, as: $\epsilon_{kk} = \alpha_v \Delta T$. Values of $\alpha_v \approx 3 \times 10^{-5}$ °C $^{-1}$ have been measured for the reservoir greywackes for temperatures of 250 °C [*Taylor et al.*, 1982]. Hence for the previously cited minimum volume strain of 5×10^{-4} a temperature change of some 17 °C, between 1977 and 1996, would be required.

We can place an upper bound on how much The Geysers has cooled by following a similar approach to that outlined by Segall and Fitzgerald [1997]. We assume, that for time periods on the order of tens of years, the unexploited geothermal system would maintain a steady temperature and pressure, and that, initially, all the reservoir fluid was in its liquid state. Records, from the California Division of Oil, Gas and Geothermal Resources, show that 1.4×10^{12} kg of steam have been extracted from The Geysers between 1977 and 1996. The energy loss associated with the simple removal of reservoir water at ambient temperature does not produce a temperature drop in the remaining reservoir; reservoir cooling is caused by the flashing of water to steam. For the reservoir temperature of approximately 240 °C [*Truesdell et al.*, 1992], the enthalpy of evaporation is about 1.8×10^6 Joules kg^{-1} [*Henley et al.*, 1984]. This means a net cooling of the reservoir of some 2.5×10^{18} Joules has occurred. In addition 4.3×10^{11} kg of liquid water, at about 40 °C, were injected into the reservoir over this period. We assume this injectate was rapidly heated to the ambient reservoir temperature of 240 °C. For a specific heat capacity of at least 4.2×10^3 Joules Kg^{-1} °C^{-1} a further net cooling of 3.6×10^{17} Joules is predicted. The energy associated with the increase in steam filled porosity within the reservoir during this time is, at most, 7.2×10^{16} Joules and can therefore be neglected.

In total 2.86×10^{18} Joules have been removed from the Geysers between 1977–1996. The greywacke's density, ρ, is 2.7×10^3 kg m^{-3}, its specific heat capacity, c, at 250 °C is 1000 Joules kg^{-1} °C^{-1} [*Taylor et al.*, 1982]. So for a reservoir volume of 2×10^{11} m^3 a maximum temperature change of some 5.3 °C would be expected. Clearly even this upper bound value of ΔT cannot explain the observed subsidence.

We turn now to the case of poroelasticity. Here we relate ϵ_{kk} to the pore pressure change in the geothermal reservoir, ΔP, by: $\epsilon_{kk} = \Delta P\left((1/K) - (1/K_s)\right)$ [*Nur and Byerlee*, 1971]. The parameter K is the quasi-static bulk modulus of the reservoir and surroundings and K_s is the bulk modulus of the mineral grains that make up the rock. For The Geysers reservoir of graywacke and silicic intrusives we set $K_s \approx 3.7 \times 10^{10}$ Pa, the value for quartz [*Mavko et al.*, 1996]. The steam pressure reduction within the reservoir fractures between 1977 and 1996 is at most about 2×10^6 Pa [*Barker et al.*, 1992]. It is assumed that this reduction gives an upper bound on the value of ΔP throughout the reservoir. So for the observed subsidence to be caused by pore-pressure changes we require a quasi-static bulk modulus, $K \leq 3.6 \times 10^9$ Pa.

However, O'Connell and Johnson [1991] observed seismic velocities for The Geysers reservoir of $V_p \approx 4.8 \times 10^3$ m s^{-1} and $V_s \approx 2.8 \times 10^3$ m s^{-1}. When we use these velocities to determine the dynamic bulk modulus, K_d, a value of $K_d = 3.4 \times 10^{10}$ Pa is found. This roughly agrees with laboratory measured values for graywacke, $K_{lab} = 3.7 \times 10^{10}$ Pa [*Kern*, 1982], and is an order of magnitude greater than K, the quasi-static bulk modulus required to explain the subsidence as a poroelastic process.

For quasi-static strains a porous solid has an effective bulk modulus, K_{eff}, given by:-

$$\frac{1}{K_{eff}} = \frac{1}{K_r} + \frac{\phi}{K_\phi} \qquad (1)$$

[*Mavko et al.*, 1996]. Where ϕ is the porosity and K_ϕ is the bulk modulus of the pore structure. The Geysers has a fracture porosity of about 1 – 2 % [*Barker et al.*, 1992]. Measurements of natural rock fractures give $K_\phi \approx 3 - 4 \times 10^7$ Pa [*David et al.*, 1994]. Here, the appropriate value for K_r, the bulk modulus of the unfractured rock matrix, is 3.7×10^{10} Pa [*Kern*, 1982]. Hence $K_{eff} \approx 1.6 - 3.5 \times 10^9$ Pa appears to agree with our subsidence derived value of K.

This discrepancy between bulk moduli, depending on whether we derive it from seismic or subsidence data can be explained by noting that the effective media approximation (1) relies on certain assumptions. These are that the deformation wavelength is large not only with respect to the smallest pore length scale but also with respect to the pore spacing. Fracture spacing in The Geysers reservoir is cited as 50 – 200 m [*Barker et al.*, 1992]. For subsidence strains with wavelengths of some 10^4 m these criteria are met. However, seismic arrival times are determined from elastic waves with

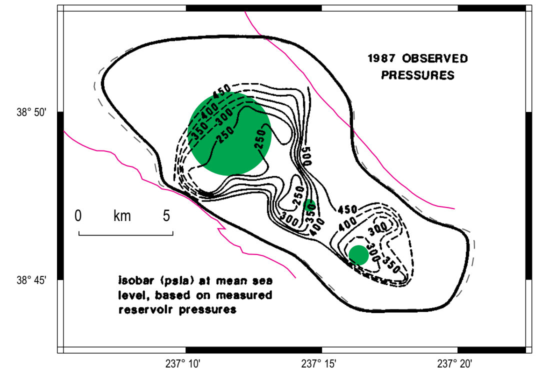

Figure 3. The best fitting Mogi sources with, superimposed, measured pressure lows as of 1987 [Williamson, 1992]. 1 psia = 6895 Pa

wavelengths on the order of 10 to 100 m and so are not consistent with effective media assumptions. Simply put, elastic waves with length dimensions smaller than the fracture spacing can find a *fast path* through and therefore exhibit travel times equivalent to unfractured rock.

There will also be some component of inelastic deformation contributing to the low quasi–static bulk modulus, K, inferred from the subsidence. However, the inelastic properties of The Geysers reservoir rocks are unknown and so we restrict ourselves here to purely elastic phenomena.

We conclude that we cannot explain the major part of the observed subsidence at The Geysers as a thermoelastic contraction. We can explain it as a poroelastic contraction. This requires a low effective bulk modulus for the reservoir, consistent with its being fracture dominated. The apparent discrepancy in reservoir stiffness determined from seismic arrival time data is due to the wide spacing of the fractures compared to the seismic wavelengths employed. The possibility that some other mechanism, not considered here, may also be able to account for the observed subsidence cannot be dismissed. However, in support of a poroelastic explanation figure 3 shows the close correlation in location and relative intensity between the pressure lows that had developed by 1987 [*Williamson*, 1992] and our best fitting centres of contraction.

Acknowledgments.
The fieldwork involved in this study was not always pleasant, so my thanks go to all those who helped me in the field:- Carl Chang, Doug Dodge, Patti (Maria-Giovanna) Guatteri, Mark Holloway, Monique Jaasma, Shelley Kenner, Dave Lessick, Ian Manger, Mark Murray, Susan Owen, Mike Poland and Ellen Yu. I'm also grateful for the help given by the employees of Calpine, C.C.P.A., N.C.P.A. and Unocal, by Ali Khan of DOGGR, and by Roger Denlinger who passed on the thermophysical data. This research was supported by the D.O.E. Office of Basic Energy Sciences.

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(Received March 11, 1997; revised May 12, 1997; accepted May 16, 1997.)