PROCEEDINGS, Twenty-Fourth Workshop on Geothermal Reservoir Engineering
Stanford University, Stanford, California, January 25-27, 1999
SGP-TR-162

AN ALTERNATIVE MECHANISM FOR THE FORMATION OF THE GEYSERS VAPOR-DOMINATED RESERVOIR

Rick Allis[1] and G. Michael Shook[2]
[1] Energy & Geoscience Institute, University of Utah, Salt Lake City, Utah, UT 84108
[2] Idaho National Engineering and Environmental Laboratory, Idaho Falls, ID 83415

ABSTRACT

Geological and geophysical evidence from The Geysers reservoir is consistent with both uniaxial tectonic extension and episodes of extension at times of magmatic intrusion. The creation of new fracture volume within an initial low porosity, liquid-dominated reservoir may cause and sustain vapor-dominated conditions. Simple, two-dimensional modeling of a dilating reservoir shows that a super-heated zone may underlie normal vapor-dominated conditions. These conditions form with either episodic or continuous dilation. Rapid venting or a sustained period of net mass loss from the reservoir are not essential. A small amount of meteoric recharge and steam loss at the surface causes the super-heated conditions to locally disappear, and these models are remarkably similar to the lateral transition from normal vapor-dominated to superheated conditions seen at The Geysers. Sustained vapor-dominated conditions still require relatively sealed boundaries, which in an extensional environment implies a self sealing mechanism on boundary fractures, ductile deformation, or movement on low angle faults. Additional modeling is needed to test in more detail the dilation hypothesis against known reservoir characteristics.

INTRODUCTION

The classic paper of White et al. (1971) describes the essential features of a vapor-dominated system. An initial, liquid-dominated system with a potent heat source and reduced liquid inflow begins to boil off more water than can be replaced, resulting in falling liquid levels. The increasing vapor zone resembles a heat pipe, with steam rising and condensing near the top of the reservoir, and condensate draining downwards. Subsequent numerical modeling has confirmed the physical requirements to form and sustain a vapor-dominated reservoir (Straus and Schubert, 1981; Pruess, 1985; Ingebretsen and Sorey, 1988; Shook, 1995). Low fluid pressures (30 – 100 bar) characteristic of large vapor-dominated reservoirs such as The Geysers, California, and Larderello, Italy, necessitate low permeability boundaries everywhere to prevent flooding from outside the reservoir. Some models invoke rapid venting of reservoir fluids over a short time (<100 years) and subsequent sealing of the vents to cause a relatively sudden transition to vapor-dominated conditions (Pruess, 1985; Shook, 1995). However, Ingebretsen and Sorey (1988) show that a slow transition is also possible. Very high-temperatures (>300°C) and low pressures such as those found in the northwest part of The Geysers and at depth in Larderello Field could also be consistent with either superheated conditions (Truesdell et al., 1993) or a residual, immobile brine in a two phase, vapor-dominated reservoir (Shook, 1995).

In view of the requirements for a long-lived heat source and a long-lived seal, large vapor-dominated zones are likely to be rare (Ingebretsen and Sorey, 1988). Magma intruded into the upper crust is likely to cool on time scales of 10^4 – 10^5 years (Cathles, 1981; intrusion volume typically <10 km^3) although, under special conditions of poor fluid circulation and intrusion characteristics, a time scale of 800,000 years is theoretically feasible (Cathles et al., 1997). A more likely scenario for a long-lived (i.e. $>10^5$ years) geothermal system with a significant surface heat flow is a combination of pulses of magma to sustain or rejuvenate a shallow heat source and/or tectonic factors such as fault movement to modify or sustain the reservoir permeability and as a result the fluid flow regime. In this scenario, a large vapor-dominated reservoir may be a relatively brief phase(s) of a long-lived, episodic, liquid-dominated geothermal system.

However, the available evidence for The Geysers reservoir does not support an ephemeral vapor-dominated phase. These conditions have probably been in existence since 0.25 Ma (Hulen et al., 1997a,b) and liquid-dominated geothermal conditions have been present at earlier times since the known felsite intrusion over 1 Ma ago (Sternfeld, 1989; Moore and Gunderson, 1995; Moore et al., 1998). The longevity of the reservoir in both its present vapor- and earlier liquid-dominated states raises many questions about the mode of heat and fluid flow through the upper crust in the region. In this paper we explore the possibility that new fracture volume as a result of tectonic extension and/or magmatic intrusions beneath the reservoir has helped sustain and possibly even caused the vapor-dominated conditions during the last few hundred thousand years. The enhanced fracture volume would decrease liquid saturation and improve the permeability and heat exchange characteristics of the reservoir rock. In the first part of the paper we review the evidence for extension at The Geysers. We then demonstrate with numerical modeling that relatively small amounts of increased fracture volume can cause vapor-dominated conditions in an initial low porosity, liquid-dominated reservoir.

EXTENSION AT THE GEYSERS

There is frequently a close association between geothermal activity, recent magmatism and crustal extension. Extension provides accommodation for magmas to rise into the upper crust and shallow magma bodies are usually considered to be the heat source for most large geothermal systems. Although there is not strong topographical evidence of extension at The Geysers, recent geological and geophysical studies have confirmed the earlier conclusions from geodetic and seismicity data that extension has been occurring in the Quaternary (Nielson and Nash, 1997; Stanley et al., 1997, 1998). These studies document a NE-trending extension zone between The Geysers reservoir and Clear Lake, where Quaternary north-northeast and northeast trending faults, fractures, lineaments and mineralized veins are common. The features are consistent with fault plane solutions from seismicity in The Geysers region which indicate uniaxial extension on a 105º azimuth below 1 km depth (Openheimer, 1986; Eberhart-Phillips, 1988). Subsurface geological evidence from within the reservoir is conflicting, with suggestions for randomly oriented fractures, north-northeast fractures, low-angle fractures, and a domed structural relationship to the felsite (Beall and Box, 1992; Thompson and Gunderson, 1992; Thompson, 1992) .

The rate of extension at The Geysers is poorly constrained. The present day west-northwest extension derived from geodetic measurements in the Geysers-Clear Lake area is 0.2μstrain/y (Prescott and Yu, 1986). This is equivalent to 2 mm/y of extension over the 10 km, northwest-trending length of the Geysers reservoir. If this extension were to be fully manifest as reservoir dilation, it would increase porosity by 5% over 0.25 Ma. However, although the geodetic extension direction agrees with the earthquake focal mechanisms, the geodetic network showed no significant dilatation. This does not preclude dilation in the Geysers reservoir area because it is a small part of the geodetic network and the network samples a short time period. In addition, dilation may occur intermittently, with most dilation closely associated with the earthquake cycle.

Repeated magmatic intrusions which are necessary to sustain The Geysers geothermal system may also cause dilation of the overlying reservoir rocks. Simple heat balance calculations based on the present day total natural heat flow from the reservoir being of the order of 50 MWth (25 MWth conductive, plus 25 MWth convective), and assuming this is representative of the last 0.5 Ma, implies a 10 km thick magma to have cooled by 500°C in the upper crust beneath the reservoir. Clearly there has been substantial magmatic intrusion since the original felsite intrusion, and some extension of the reservoir rocks seems inevitable at the time of subsequent intrusions.

The fracture volume used in published numerical models of The Geysers reservoir response to development is small, typically 0.01 – 0.06 (Williamson, 1992; Gunderson, 1992), so large amounts of dilation are not required to significantly affect the fracture volume. Indeed, it is surprising that the fracture volume is not larger in view of the thermal and tectonic history of the reservoir, and evidence at least at shallow depth, of carbonate vein dissolution. Presumably, loss of fracture volume also occurs over time (at least locally) due to vein-filling processes and rock recrystalization or ductile relaxation, particularly in the hotter, deeper parts of the reservoir (discussed by Gunderson, 1992).

The concept that faulting and rock dilation could cause reduced fluid pressure was proposed by Sibson (1987) as an explanation for much fault-hosted epithermal mineralization. Earthquake rupture termination at dilational jogs is believed to involve extensional fracturing, fluid pressure reduction, enhanced fluid flow, and local boiling with mineralization where temperatures are high enough. Sibson (1987) suggested that this type of event has occurred intermittently on a large scale in the dilational jogs of the southern San Andreas system, which are the sites of active magma-hydrothermal systems. Openheimer (1986) speculated that The Geysers-Clear Lake region could also be a site where dextral shear on the San Andreas steps northeastwards. If fluid flow from outside The Geysers reservoir is very poor, dilational events could cause a fluid pressure decline, and as shown by the modeling below, could cause or sustain the vapor-dominated state of the reservoir.

In contrast to the under-pressures found at The Geysers (relative to hydrostatic pressure from the surface), northern California is noted for pervasive over-pressures at shallow depth, especially in the Franciscan Complex (Berry, 1973; Figs. 1, 2). Both types of pressure anomaly are indicative of low permeability boundaries preventing fluid flow in the anomalous zones equilibrating with hydrostatic pressure. Tectonic compression due to the subduction of the Juan da Fuca plate beneath North America just north of The Geysers is the main cause of the high fluid pressures. If compressional tectonics are the cause of pervasive over-pressures in northern California, could local extension at The Geysers also be a major factor causing the sustained vapor-dominated conditions?

MODELING RESULTS

A numerical study was undertaken to evaluate the idea of reservoir dilation causing a flip to vapor-dominated conditions in the absence of discharge. The simulation domain is a simple 2-D model (100 by 40 grid blocks), 10 km in length and 1 km wide. This slice represents approximately 20% of the known width of the reservoir. The total model depth is 4 km, with the uppermost 1 km as the caprock and the lowest 1 km the bedrock. Reservoir permeability and initial porosity were taken as a constant 1×10^{-14} m^2 and 0.0125 respectively; caprock and bedrock properties were set to $k = 1\times10^{-18}$ m^2 and $\varphi = 0$. For the purposes of this preliminary study, the reservoir was taken as a porous (rather than fractured) medium, and pure water was the reservoir fluid. Relative permeability functions used are:

$$kr_{liquid} \equiv \left(\frac{S_{liquid} - 0.4}{1 - 0.4} \right)^{4.0}$$

$$kr_{vapor} \equiv \left(\frac{S_{vapor} - 0.05}{1 - 0.05} \right)^{2.5}$$

A hydrostatic pressure gradient was established assuming normal pressure from the ground surface. Temperatures were calculated for each layer as follows. Caprock and bedrock temperatures (i.e., above and below the geothermal reservoir) were set by assuming a conductive gradient of 230 ^{o}C/km, with surface temperature set at 10^{o}C. Reservoir temperatures were taken from a boiling point with depth relationship and the given pressure gradient.

A constant heat flux of 0.5 W/m^2 was added to the base of the domain, and the initial temperature was maintained at the upper boundary. This condition was simulated for 1000 years to establish a steady state initial condition. That condition is given in Figure 3. After a steady initial condition was achieved, the reservoir porosity was increased to 0.04 to simulate a dilation event. Rock density was increased at the same time to maintain energy conservation. Several recharge/discharge scenarios were also simulated, as discussed below.

In the base case, no recharge/discharge was simulated. Conditions after 30,000 years of simulated time are given in Figure 4. This figure shows the formation of a vapor-static pressure gradient (Fig. 4a) that resulted from the increase in pore volume. Various increases in pore volume were initially considered in order to obtain pressures similar to the initial state of The Geysers (3.4 MPa). We also note in passing that, based on some preliminary simulations not discussed here, it appears to not matter whether the pore volume increase occurs in a single step, or as an accumulation of several dilation events. Either scenario results in similar pressure and fluid distributions.

Figure 4b shows a temperature profile at t=30,000 years. Temperatures in the upper half of the reservoir (down to about 2300 m) are consistent with a vapor-dominated heat pipe. In the lower portion of the reservoir, however, a superheated region exists, with temperature gradients approximately conductive. The superheated region is also apparent in the saturation profile of Figure 4c. Liquid saturations in the normal vapor-dominated reservoir range between 0.45 and 0.5, but fall sharply to zero at 2250 m.

Two additional cases were simulated to evaluate the effects of recharge/discharge on reservoir conditions. An injection well was established in the upper left-hand grid block of the reservoir to simulate a fixed rate of recharge, and a second well was established in the upper center of the reservoir model for discharge. These well locations are meant to loosely simulate possible meteoric recharge from the southeast into The Geysers, with discharge in central Geysers region. The "recharge" well was set at a constant injection rate; the "discharge" well was placed on a wellhead pressure constraint of 10 bar. Over the time period simulated here (30,000 years) there is a net inflow to the reservoir for both recharge rates simulated, though discharge monotonically increases towards balanced (zero net) inflow/outflow. This clearly indicates lack of an equilibrium state over this time scale, as discussed below.

Although there is strong evidence for both natural meteoric water leaking into and natural steam losses from The Geysers reservoir, if the reservoir has been vapor-dominated for the last 0.25 Ma, then its equilibrium state is very sensitive to the net mass flow balance from the reservoir. For example a 1 kg/s net inflow/outflow of water to the model domain (20 km^3) would totally saturate/dry-out the 0.04 pore volume in less than 30,000 years. An outflow of 1 kg/s of steam amounts to a domain heat loss of 3 MW_{th}, or equivalent to around 15 MW_{th} from the whole 50 km^2 reservoir area. This is not an unreasonable steam loss rate given the extent of thermal activity shown in early photographs and descriptions of the original Geysers resort area (Koenig, 1992; Hodgson, 1992). Longevity of vapor-dominated conditions requires approximate mass flow equilibrium in the reservoir, or mass fluctuations to be counter-balanced by varying heat input, or varying reservoir pore volume.

In the first open-system case we modeled, a relatively small amount (0.5 kg/s) of cooler (35°C) water recharge was allowed. All other conditions were identical to those of the base case. Reservoir conditions after 30,000 simulated years are given in Figure 5. Reservoir pressures (Fig. 5a) in this case are somewhat larger than those observed at The Geysers due to increased pressure support from boiling recharge water. This is also sensitive to the relative permeability (and residual saturations) used in the model, and also due to the degree of dilation assumed from the base case. In any case, pressure gradients within the reservoir remain nearly vapor-dominated. Temperatures given in Figure 5b also reflect a vapor-dominated heat pipe, with temperature gradients of less than 0.01°C/m. This gradient is continuous throughout the vapor-dominated heat pipe. At depth furthest from the recharge/discharge, however, a superheated zone exists that is similar to (though not as pervasive as) that seen in the base case.

This superheated zone is also apparent in the saturation profile given in Figure 5c. The extent of this high temperature reservoir has been reduced through recharge from the "southeast", but prevails in the "northwest". The similarity between these simulated results and those postulated by Truesdell et al. (1993) is remarkable, and may reflect conditions existing in the high temperature reservoir in northwest Geysers. It should be noted, however, that any salt in the initial water will likely concentrate in the deeper portion of the reservoir, and the vapor pressure lowering effect would maintain two-phase conditions. This scenario will be explored in an extension of this work.

The amount of recharge was doubled to 1 kg/s in a second simulation. Results from this case after 30,000 years of simulated time are given in Figure 6. The pressure profile (Fig. 6a) in this case shows the effects of the recharge of cooler water, with significant relief on the isobars nearer the point of recharge. These effects are also seen in the temperature profile in Figure 6b. Elevated temperatures furthest from the point of recharge are the result of a more efficient heat pipe, increased pressure, and saturated conditions. The saturation profile given in Figure 6c also shows the effect of increased recharge. In this case, the superheated zone at depth has disappeared due to availability of recharge water.

In summary, a vapor-dominated heat pipe may develop in response to changes in pore volume, without any venting or recharge. If the pore fluid is fresh water, a superheated zone may develop at depth. That superheated zone will shrink or grow in size depending on the mass balance between liquid recharge and steam discharge at the surface. The results with a small meteoric inflow into one part of the reservoir appear to be in good agreement with the conceptual model of Truesdell et al. (1993). This conclusion, however, is based on the use of fresh water as the reservoir fluid, and may also be sensitive to the use of a porous medium in this study. Additional modeling studies are planned to consider the effects of salinity, fractured media and variations in the leaky caprock on this conceptual model.

CONCLUSIONS

Both tectonic evidence and the magmatic intrusion history implied by the longevity of the geothermal activity indicate that The Geysers reservoir is in a tensional environment. Dilation of the reservoir rock could have occurred intermittently, especially at the times of major local earthquakes and/or magmatic intrusion episodes beneath the reservoir. This dilation could have been responsible for causing the earlier liquid-dominated reservoir to become vapor-dominated, and could have assisted with sustaining these conditions despite periods of decreasing heat flow or increased meteoric recharge to the reservoir.

It is possible that the change from compressional tectonic stresses 1 – 3 Ma ago to an extensional regime with local dilation occurring in the Geysers also initiated the liquid-dominated phase of the geothermal system around the time the felsite pluton was intruded. Initially (> 1.2 Ma?) the Geysers area could have been similar to that now seen in the Wilson-1 well and at Sulphur Bank – significant fluid over-pressures and little fluid circulation despite high temperatures, with basaltic dikes being the main indicator of shallow magmatism. As tectonic stresses relaxed, the felsite intrusion occurred, and a liquid-dominated geothermal field became established, recharged by meteoric water. Progressive thermal alteration of the near-surface rocks and sealing of inflow channels could have limited near-surface permeability and fluid flows over time. Dilation of reservoir rocks at depth could have then caused a gradual transition to vapor-dominated conditions around 0.25 Ma.

Preliminary conclusions from simple 2-D modeling of a dilating reservoir are:

- vapor-dominated conditions can evolve in response to a dilation event, and do not require any venting or recharge. If no recharge occurs, it appears that a superheated zone may underlie a normal vapor-dominated reservoir.
- if any recharge occurs, superheated conditions tend to disappear near(er) the point of recharge, forming a normal vapor-dominated reservoir. A superheated zone may still exist away from the recharge. If sufficient recharge occurs, that zone may disappear entirely.

These results are based on the notion of pure (fresh) water in a porous medium, and are subject to change when those simplifications are removed from the model itself. These are planned changes, along with additional sensitivity studies on boundary conditions, in a following paper.

ACKNOWLEDGEMENTS

We are grateful for the comments and stimulating discussions with our colleagues which were triggered by the initial ideas and an earlier draft of this paper. In particular, we thank Jeff Hulen, Joe Moore, Dennis Nielsen, Mike Adams, and Joel Renner for their input. The research was carried out under contracts with the office of Geothermal Technologies of the U.S. Department of Energy.

REFERENCES

Beall, J.J. 1985. Exploration of a high temperature, fault localized, nonmeteoric geothermal system at the Sulphur Bank mine, California. GRC Trans. 9, 395-401.

Beall, J.J. and Box, W.T. 1992. The nature of steam-bearing fractures in the Southern Geysers reservoir. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 69-76.

Berry, F.A.F. 1973. High fluid potentials in California Coast Ranges and their tectonic significance. AAPG Bull., 57, 1219-1249.

Cathles, L.M. 1981. Fluid flow and genesis of hydrothermal ore deposits. Economic Geology 75th Anniversary Volume, 442-457.

Cathles, L.M., Erendi, A.H.J. and Barrie, T. 1997. How long can a hydrothermal system be sustained by a single intrusive event? Economic Geology, 92, 766-771.

Eberhart-Phillips, D, 1988. Seismicity in the Clear Lake area, California. Geol. Soc. Amer. Spec. Paper 214, 195-206.

Fournier, R.O. 1991. The transition from hydrostatic to greater than hydrostatic fluid pressure in presently active continental hydrothermal systems in crystalline rock. Geology, 18, 955-958.

Gunderson, R. 1992. Porosity of reservoir greywacke at The Geysers. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 89-96.

Hodgson, S.F. 1992. A Geysers album. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 19-42.

Hulen, J.B., Heizler, M.T., Stimac, J.A., Moore, J.N. and Quick, J.C. 1997a. New constraints on the timing of magmatism, volcanism, and the onset of vapor-dominated conditions at The Geysers steam field, California. Proc. 22nd Stanford Workshop, 75-82.

Hulen, J.B., Quick, J.C. and Moore, J.N. 1997b. Converging evidence for overpressures at peak temperatures in the pre-vapor-dominated Geysers hydrothermal system. GRC Trans., 21, 623-628.

Ingebretson, S.E. and Sorey, M.L. 1988. Vapor-dominated zones within hydrothermal systems: evolution and natural state. JGR, 93, 13635-13655.

Koenig, J.B. 1992. History of development at The Geysers geothermal field, California, Spec. Rep. 17, GRC, 1-18.

Moore, J.N. and Gunderson, R.P. 1995. Fluid inclusion and isotopic systematics of an evolving magmatic-hydrothermal system. Geochim. Cosmochim. Acta, 59, 3887-3907.

Moore, J.N., Anderson, A.J., Adams, M.C., Aines, R.D., Norman, D.I. and Walters, M.A. 1998. The fluid inclusion and mineralogic record of the transition from liquid- to vapor-dominated conditions in The Geysers system, California. Proc. 23rd Stanford Workshop, 211-218.

Nielsen, D.L. and Nash, G.D. 1997. Structural fabric of The Geysers. GRC Trans., 21, 643-649.

Oppenheimer, D.H. 1986. Extensional tectonics at The Geysers geothermal area, California. JGR, 91, 11463-11476.

Prescott, W.H. and Yu, S. 1986. Geodetic measurement of horizontal deformation in the northern San Francisco Bay region. JGR, 91, 7475-7484.

Pruess, K. 1985. A quantitative model of vapor-dominated geothermal reservoirs as heat pipes in fractured porous rock. GRC Trans., 9, 353-361.

Shook, M. 1995. Development of a vapor-dominated reservoir with a "high-temperature" component. Geothermics, 24, 489-505.

Sibson, R.H. 1987. Earthquake rupturing as a mineralizing agent in hydrothermal systems. Geology, 15, 701-704.

Stanley, W.D., Benz, H.M., Walters, M.A. and Rodriguez, B.D. 1997. Tectonic control on magmatism and geothermal resources in The Geysers-Clear Lake region, California: Integration of new geologic, earthquake tomography, seismicity, gravity, and magnetotelluric data. U.S. Geol. Surv. Open File Report 97-95, 39p.

Sternfeld, J.N. 1989. Lithologic influences on fracture permeability and the distribution of steam in the northwest Geysers steamfield, Sonoma County, California. GRC Trans., 13, 473-480.

Stanley, W.D., Benz, H.M., Walters, M.A., Villasenõr, A., and Rodriguez, B.D. 1998. Tectonic controls on magmatism in The Geysers-Clear Lake region: Evidence from new geophysical models. GSA Bull., 110, 1193-1207.

Straus, J.M. and Schubert G. 1981. Gravitational stability of water over steam in vapor-dominated geothermal systems. JGR, 85, 6505-6512.

Thompson, R.C. 1992. Structural stratigraphy and intrusive rocks at The Geysers geothermal field. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 59-64.

Thompson, R.C. and Gunderson, R.P. 1992. The orientation of steam-bearing fractures at The Geysers geothermal field. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 65-68.

Truesdell, A.H., Walters, M., Kennedy, M., and Lippmann, M., 1993. An Integrated Model for the Origin of The Geysers Geothermal Field. GRC Trans., 17, 273-280.

Walters M.A. and Combs, J. 1992. Heat flow in The Geysers-Clear Lake geothermal area of northern

California, USA. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 43-58.

Walters, M.A., Hulen, J.B. and Moore, J.N. 1997. Geology and oxygen-isotope geochemistry – Audrey-A1 geothermal well, Sulphur bank mine area, Lake County, California. Proc. 22nd Stanford Workshop, 473-482.

White D.E. Muffler, L.J.P. and Truesdell, A.H.1971. Vapor-dominated hydrothermal systems compared with hot-water systems. Economic Geology, 66, 75-97.

Williamson, K.H. 1992. Development of a reservoir model for The Geysers geothermal field. In: C, Stone, ed. Monograph on The Geysers Geothermal Field, Spec. Rep. 17, GRC, 179-188.

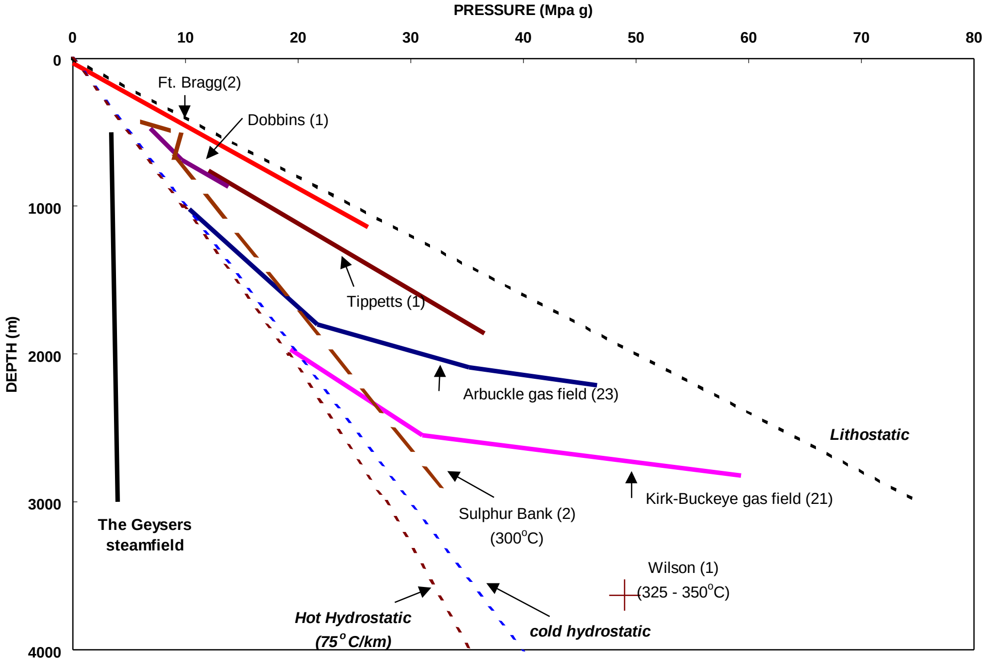


Figure 1: Compilation of Pressure-Depth trends from northern California. Most data are taken from Berry, (1973); the wells and fields are located on Fig. 2. The number of wells used to determine each trend is given in brackets. The Wilson point is from Fournier, (1991; geothermal exploration well Wilson-1), and the Sulphur Bank line is derived from static pressure and temperature data for geothermal exploration well Audrey A-1 (data for feedzone at 2890 m depth given in Walters et al., 1997; and Beall, 1985), and three points from Bradley-1 well (data from Berry, 1973). In both these wells, the bottom hole temperature was in the range 300 - 350°C, and there was evidence of formation waters (CO_2-rich chloride waters) in the feedzones. The degree of over-pressure at these two sites should be referred to the “hot hydrostatic” trend. The strong under-pressures (near vapor-static) at The Geysers are in marked contrast to the over-pressures found to the northeast and northwest of this reservoir.

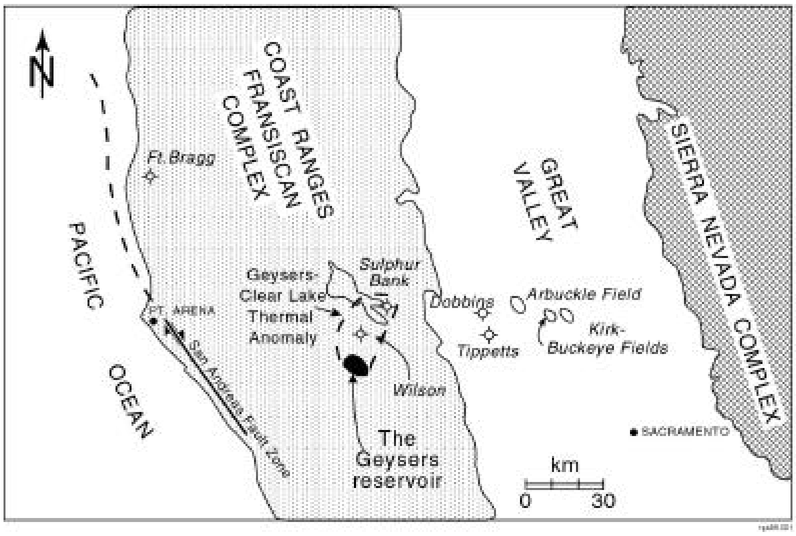
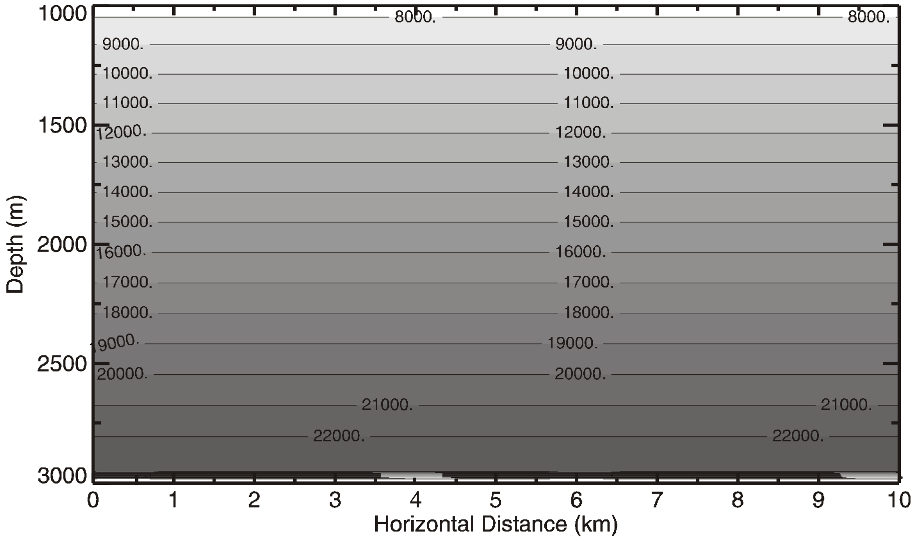
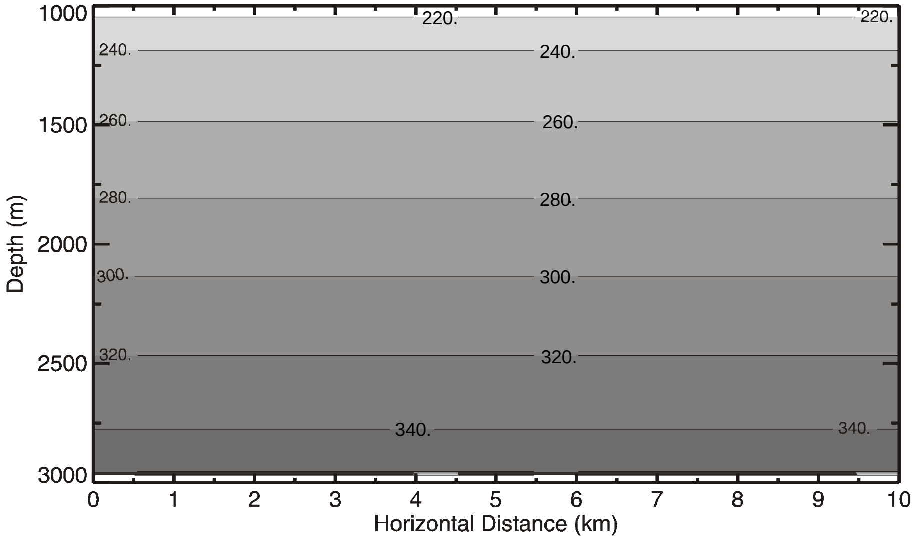


Figure 2. Map of northern California showing the location of The Geysers geothermal reservoir, and the other wells or gas fields plotted in Fig.1 with varying degrees of over-pressures. The boundary of the Geysers-Clear Lake thermal anomaly is taken from Stanley et al. (1998), and approximately coincides with a gradient of 75°C/km.

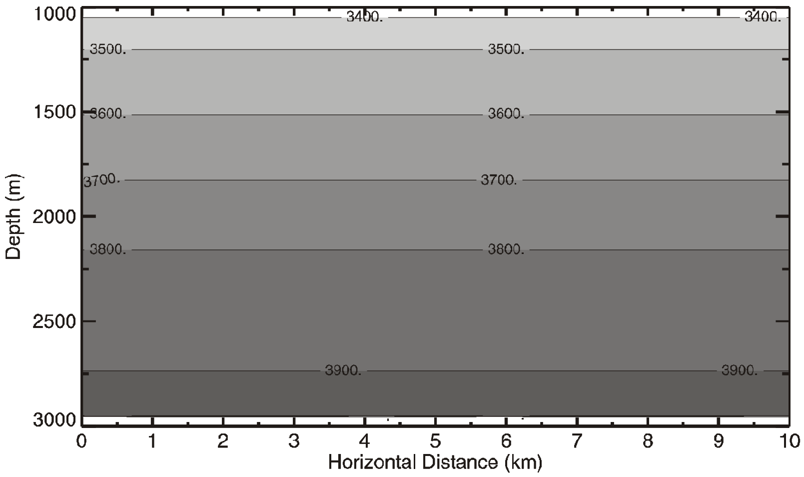


3a. Pressure

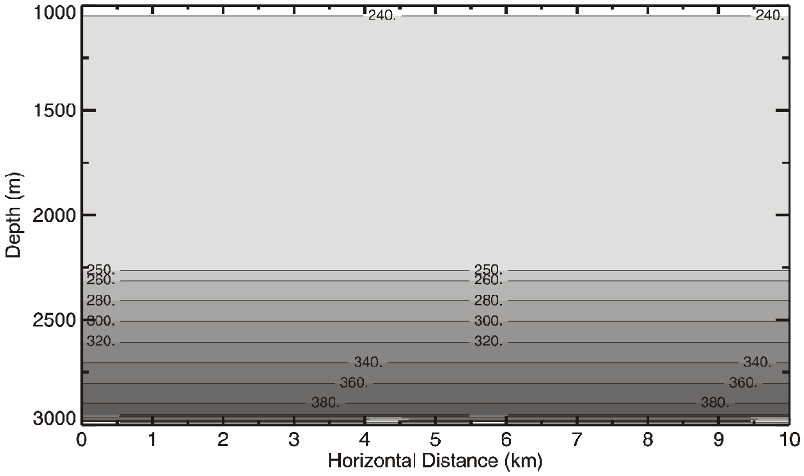


3b. Temperature

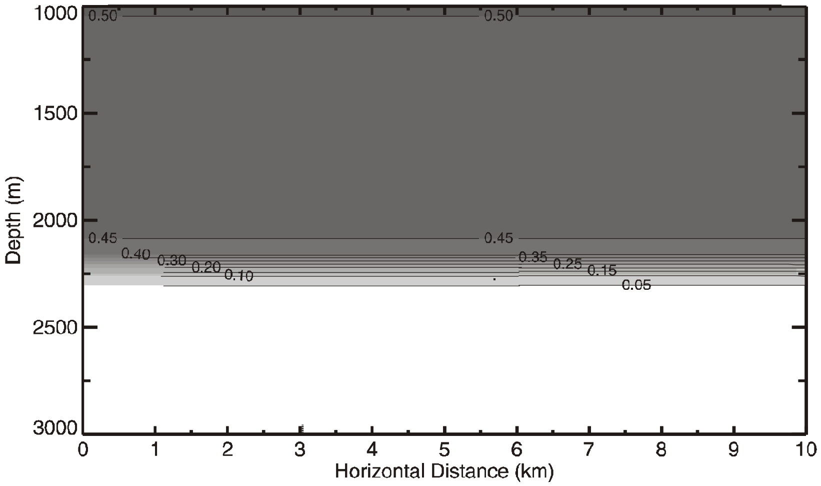
Figure 3. Initial reservoir pressure (kPa) and temperature (°C) for modeling study.



4a. Pressure

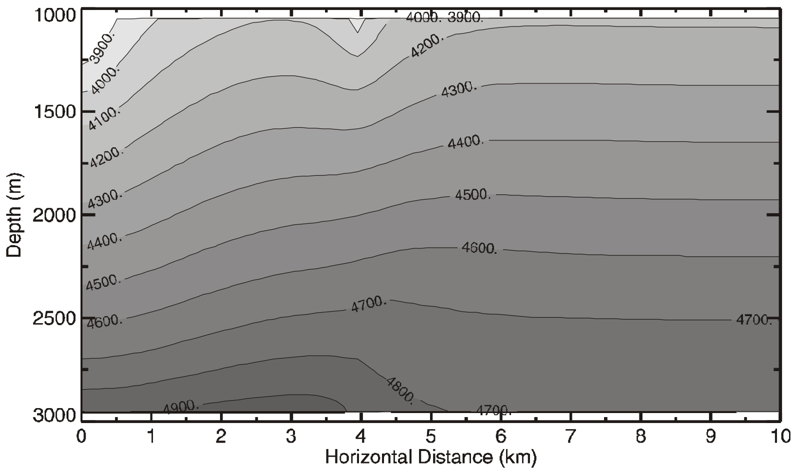


4b.Temperature

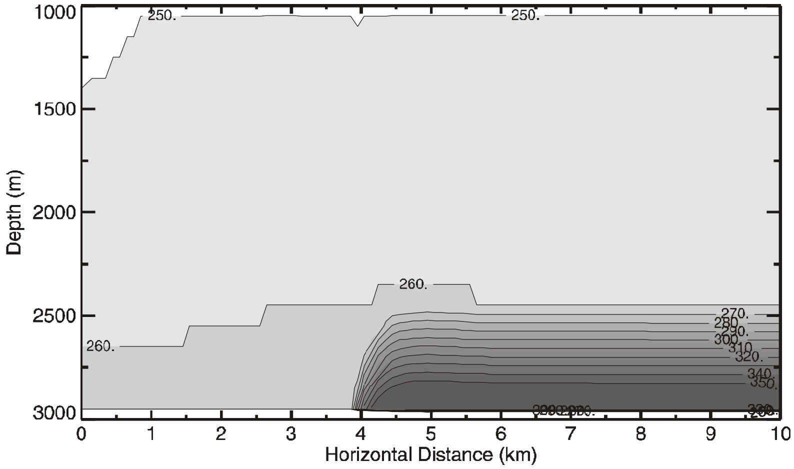


4c. Saturation

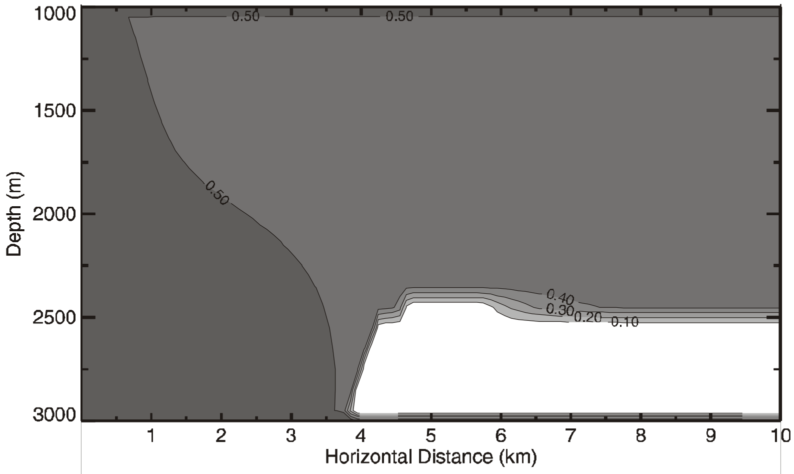
Figure 4. Reservoir conditions at t = 30,000 years for the base case (no recharge). Pressure in kPa and temperature °C.



5a. Pressure

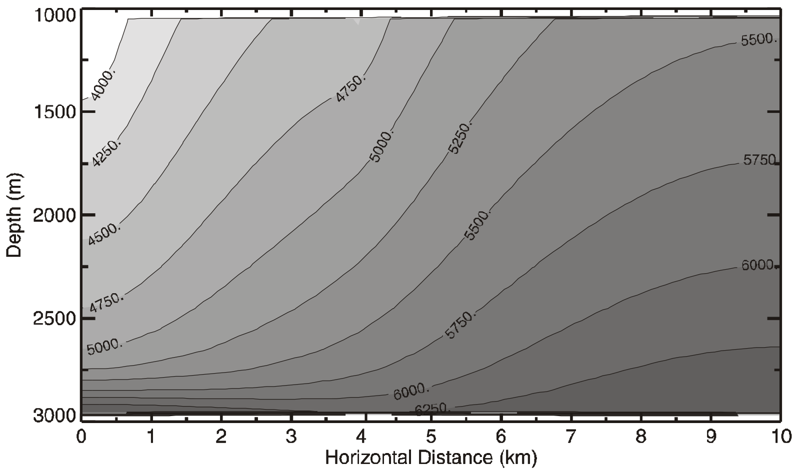


5b. Temperature

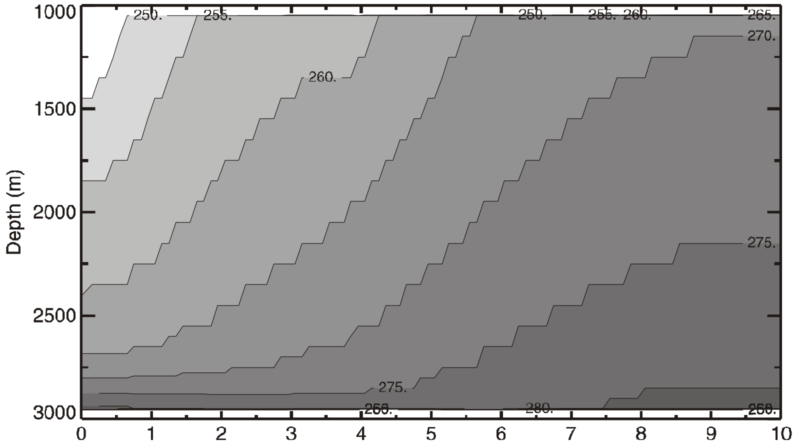


5c. Saturation

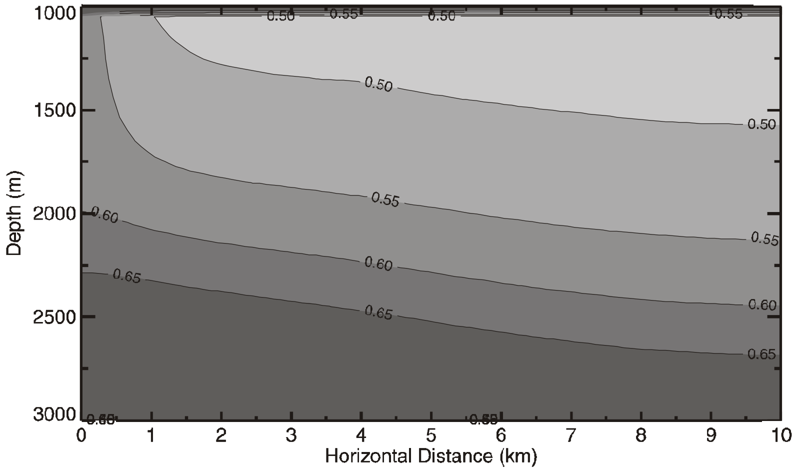
Figure 5. Reservoir conditions at t = 30,000 years for recharge case 1 (0.5 kg/s). Pressure kPa, temperature °C.



6a. Pressure



6b. Temperature



6c. Saturation

Figure 6. Reservoir conditions at t = 30,000 years for recharge case 2 (1 kg/s). Pressure in kPa, temperature in °C.