SGP-TR-150-25

PROCEEDINGS

TWENTIETH WORKSHOP
GEOTHERMAL RESERVOIR ENGINEERING

January 24-26, 1995

Stanford Geothermal Program
Workshop Report SGP-TR-150

PROCEEDINGS, Twentieth Workshop on Geothermal Reservoir Engineering
Stanford University, Stanford, California, January 24-26 1995
SGP-TR-150

INSTABILITIES DURING LIQUID MIGRATION INTO SUPERHEATED HYDROTHERMAL SYSTEMS

Shaun D. Fitzgerald & Andrew W. Woods

Institute of Theoretical Geophysics,
Departments of Applied Mathematics & Theoretical Physics,
Cambridge.

ABSTRACT

Hydrothermal systems typically consist of hot permeable rock which contains either liquid or liquid and saturated steam within the voids. These systems vent fluids at the surface through hot springs, fumaroles, mud pools, steaming ground and geysers. They are simultaneously recharged as meteoric water percolates through the surrounding rock or through the active injection of water at various geothermal reservoirs. In a number of geothermal reservoirs from which significant amounts of hot fluid have been extracted and passed through turbines, superheated regions of vapour have developed. As liquid migrates through a superheated region of a hydrothermal system, some of the liquid vaporizes at a migrating liquid-vapour interface. Using simple physical arguments, and analogue laboratory experiments we show that, under the influence of gravity, the liquid-vapour interface may become unstable and break up into fingers.

1 INTRODUCTION

For several decades fluids have been extracted from two-phase geothermal reservoirs for power generation. Important reservoirs include The Geysers, California and Larderello, Italy. As a result of the depletion of fluids that has arisen (Kerr 1991), the pressures within such reservoirs have decreased and some regions of the systems have become superheated. As reservoir pressures fall, the rate of supply of steam has been observed to decrease by as much as 50% per year (Enedy *et al.* 1993). In a number of reservoirs, cold fluid has been injected in order to recharge the system through the vaporisation of a fraction of this water as it migrates through the hot rock (Enedy, Enedy & Maney 1991; Enedy *et al.* 1993). In order to evaluate the efficiency of such reinjection programmes, the fundamental controls upon the rate of vaporisation of an advancing liquid front must be determined.

Pruess *et al.* (1987a) have shown that as liquid is injected at a steady rate from a vertical well, a relatively cold liquid-saturated region develops around the well. This region is separated from the hot superheated vapour in the far field by a migrating liquid-vapour interface. As the rock is invaded by cold water it cools and the thermal energy released is used to vaporize a fraction of the water. New vapour is thus formed at the liquid-vapour interface. Woods & Fitzgerald (1993) extended the work of Pruess *et al.* (1987a) by considering injection in various geometries and presenting analytical solutions for the long-time behaviour of the system. The results of both of these studies indicated that if liquid is injected from a vertical well then the total rate of vapour production increases with injection rate although the fraction vaporizing actually decreases. Figure 1 shows the relationship between the mass injection rate and the mass fraction of liquid which vaporises in a two dimensional injection geometry, as presented by Woods and Fitzgerald (1993).

Most studies of liquid injection into superheated geothermal systems have implicitly assumed that the migrating liquid-vapour interface is planar. However, such interfaces may become unstable and break up into fingers if a sufficient fraction of the water vaporizes. In this paper we build upon these preliminary results (Fitzgerald and Woods 1994) by assessing the important and sometimes dominant effects of gravity. Our analysis and analogue laboratory experiments provide important constraints upon models of vapour generation in superheated reservoirs. Our results are also of broader interest for understanding the dynamic state of geothermal systems. We also apply our results to interpret the effects of reinjection in the Geysers reservoir California, and the Lardarello field, Italy.

As well as being of major importance to the geothermal industry, our stability analysis may also lead to new insights into the formation of precious ore deposits (Cline, Bodnar & Rimstidt 1992) and also the intensity of fumarolic activity associated with crater lakes (Brown *et al.* 1989; Hurst & Dibble 1981).

After discussing the different structures of geothermal reservoirs, we present a physical discussion of the instability of a liquid-vapour interface, focusing upon the role of gravity in (de-)stabilising the interface, and identifying an upper bound upon the stability of the interface. We then present some new analogue laboratory experiments, which identify the development of this instability. Finally, we discuss the implications of our results to various geothermal reservoirs, and contrast our results with earlier models of reinjection.

2 MODELLING OF HYDROTHERMAL SYSTEMS

Hydrothermal systems are commonly thought to exist in a dynamic state in which fluid circulates within fractured, porous rock (Cathles 1977; Donaldson 1962; Dunn & Hardee 1981; Hurst & Dibble 1981; Parmentier & Schedl 1981; Grant, Donaldson & Bixley 1982; Wohletz & Heiken 1992). The pores and fractures of the systems are saturated with water or water and vapour. Typically one phase is continuous and to good approximation this determines the distribution of pressure within the system (Pruess & Narasimhan 1982). In the case of a two-phase system, if the distribution of pressure is similar to that of a liquid-filled (vapour-filled) system then the system is described as liquid-dominated (vapour-dominated).

The influence of the fractures upon the fluid flows within the reservoirs is highly dependent upon the relative resistance to flow within the fractures and porous matrix. The relative resistance to flow is dependent upon the fracture apertures, the permeability of the porous matrix and the spacing of the fractures. In many reservoirs, such as Kawah Kamojang in Indonesia (Wohletz & Heiken 1992), The Geysers in California, Larderello in Italy, Ahuachapan in El Salvador and Kawerau in New Zealand (Grant, Donaldson & Bixley 1982), the bulk of the fluid flows are believed to occur within the fractures. The flow of fluid in certain other reservoirs, such as the East Mesa reservoir in California is, in contrast, primarily within the porous matrix (Grant, Donaldson & Bixley 1982). Effects of the fractures upon the fluid flow within geothermal reservoirs are often inferred from detailed measurements of pressure and temperature within wells used for the extraction or injection of fluid (Fradkin *et al.* 1981; Goyal & Box 1990; Axelsson & Bodvarsson 1987). In many reservoirs such data suggests that the fracturing is highly pervasive (Wohletz & Heiken 1992) and that the heat and mass transfer occur over lengthscales of the same order as or larger than the interwell spacing. Therefore, for modelling purposes the rock may be treated as a homogeneous continuum with a uniform network of interconnected pores through which fluid may flow (Cathles 1977; Elder 1981; Pruess *et al.* 1987a). The volume flux per unit area $\mathbf{u}$ may then be related to the interstitial velocity $\mathbf{v}$ by

$$\mathbf{u} = \phi \mathbf{v} \qquad (2.1)$$

where ϕ is the void fraction (Bear 1972; Phillips 1991). In many situations of interest the fluid velocities through a geothermal reservoir are sufficiently low that viscous frictional forces are much greater than the inertial forces so that the interstitial Reynolds number, $R = \rho v d/\mu$ is small, $R \ll 1$ (Rubin & Schweitzer 1972; Pruess *et al.* 1987a); ρ represents the density of the fluid, μ the dynamic viscosity, d a typical pore size, and v the interstitial speed. The volume flux per unit area $\mathbf{u}$ is then given by Darcy's Law

$$\mu \mathbf{u} = -k\left(\nabla P - \rho \mathbf{g}\right) \qquad (2.2)$$

where ∇P is the applied pressure gradient, $\mathbf{g}$ is the gravitational acceleration and k is the permeability (Bear 1972; Rubin & Schweitzer 1972; Dullien 1992). For liquid flow through porous rocks of typical pore size $d \sim 10^{-5}$ m and liquid density 10^3 kg/m^3, the condition $R \ll 1$ is valid for interstitial speeds $v \leq 10^{-3}$ m/s.

If the fluid velocities v and the typical grain size D are sufficiently small then the timescale for fluid and solid to thermally equilibrate, D^2/κ, is much shorter than the timescale for the advection of heat across a grain, D/v, and the medium becomes locally isothermal. Thus for a grain size of 0.5 mm, the condition for local thermal equilibrium is the same as that for low Reynolds number flows, $v \leq 10^{-3}$ m/s since typically the thermal diffusivity of rock $\kappa \sim 2 \times 10^{-6}\,m^2/s$.

In the following analysis, we therefore assume that the rock may be modelled as an isotropic porous medium in which Darcy's law (2.2) is valid and in which the rock and fluid are in local thermal equilibrium. In this case we may adopt the results of Woods and Fitzgerald (1993), shown in figure 1 herein, which describes the mass fraction which vaporises as a function of the mass flow rate of liquid injected into the reservoir.

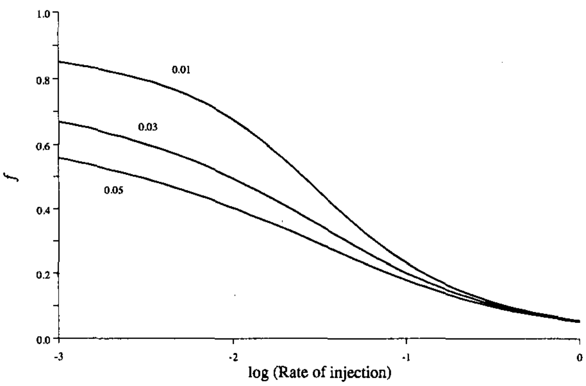

Figure 1 Mass fraction vaporising f as a function of dimensionless flow rate taken from Woods and Fitzgerald 1993. Curves are shown for 3 different porosities as labelled.

3 THE STABILITY OF A MIGRATING INTERFACE

We now present a simple physical argument to determine whether a vaporising liquid interface is stable. This simple analysis identifies the fundamental process by which an interface may become unstable, and provides a lower bound on the mass fraction which may vaporise without destabilising the interface.

The interface between two regions of fluid migrating through a porous medium can become unstable if the magnitude of the pressure gradient increases across the interface in the direction of the flow (Saffman and Taylor 1958; Homsy 1987). If liquid migrates towards a vaporising interface with Darcy velocity u, then the pressure gradient on the liquid side of the interface has the form

$$\frac{dp}{dz} = \rho_l g - \frac{\mu_l}{k} u \tag{3.1}$$

where g is the component of the gravitational acceleration in the directon of the flow, z, ρ_l is the liquid density and μ_l the liquid viscosity. If a fraction f of this liquid vaporises, then the mass flux of vapour ahead of the interface is $\rho_l f u$. Therefore, the velocity of the vapour ahead of the interface is

$$u_v = \frac{\rho_l}{\rho_v} f u \tag{3.2}$$

where ρ_v is the vapour density and the pressure gradient in the vapour region is

$$\frac{dp}{dz} = \rho_v g - \frac{\mu_v}{k}\frac{\rho_l}{\rho_v} f u \tag{3.3}$$

where μ_v is the vapour viscosity.

The pressure gradient therefore decreases (magnitude increases) in the direction of the flow across the interface whenever

$$(\rho_l - \rho_v) g > \frac{u}{k}(\mu_l - \mu_v f \frac{\rho_l}{\rho_v}) \tag{3.4}$$

Since the vapour density is much less than the liquid density, the left hand side of this expression is positive whenever the liquid overlies the vapour; such a configuration is therefore gravitationally unstable. Conversely, if the vapour overlies the liquid, then the system is gravitationally stable. The right hand side of (3.4) is positive whenever

$$\frac{\mu_l \rho_v}{\mu_v \rho_l} > f \tag{3.5}$$

In this case, the magnitude of the pressure gradient associated with the viscous stresses decreases in the direction of the flow, tending to stabilise the interface. The quantity $\frac{\mu_l \rho_v}{\mu_v \rho_l}$ has a value of the order 0.1-0.2 for a water-vapour system. Therefore expression (3.5) applies and the viscous pressure forces are stabilising whenever the mass fraction of liquid which vaporises, f, is sufficiently small

$$f < 0.1 - 0.2 \tag{3.6}$$

Otherwise the viscous pressure forces tends to destabilise the interface, as in the classical Saffman-Taylor instability (Saffman and Taylor 1958).

This result may at first appear surprising, since the relatively dense liquid is migrating into a region of less dense vapour; however, whenever, the mass fraction of liquid which vaporises is of order 0.1-1.0, the mass flux of vapour is comparable to the mass flux of liquid, and therefore the speed of the vapour is much greater than the speed of the liquid. Since the density ratio ρ_l/ρ_v is greater than the viscosity ratio μ_l/μ_v, this situation can then lead to an adverse viscous pressure gradient in the direction of flow.

For simplicity, let us introduce the dimensionless parameters

$$S = \frac{k\rho_l g}{\mu_l u} \tag{3.7}$$

and

$$F = f\frac{\mu_v \rho_l}{\mu_l \rho_v} - 1 \tag{3.8}$$

which denote the magnitude of the gravitational and viscous destabilisation of the interface. (Negative values correspond to a stabilising effect). If we assume that $\rho_v \ll \rho_l$, as is typically the case, then instability may be possible, (3.4), whenever

$$S + F > 0 \tag{3.9}$$

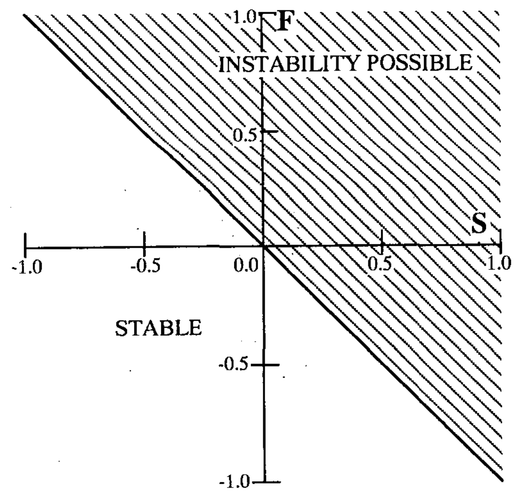

Figure 2 Schematic diagram of the various possible regimes in which a geothermal system may exist. If $S+F > 0$ then the interface may become unstable.

This expression identifies six possible stability regimes, as shown pictorially in figure 2. Two of the most interesting regimes are the cases in which (i) $S > 0$ and $F < 0$, $S + F < 0$, in which case the viscous pressure force can stabilise a gravitationally unstable situation; and (ii) $S < 0$ and $F > 0$, $S + F > 0$, in which case the viscous pressure forces may destabilise a gravitationally stable interface. Case (i) only develops when a small fraction of the descending liquid front vaporises, while case (ii) only develops when a large fraction of an ascending liquid front vaporises. Fitzgerald and Woods (1994) described the simpler system in which $g = 0$ and hence $S = 0$; this corresponds to the x-axis of figure 2.

Equation (3.9) provides a simple lower bound on the condition for instability of the vaporizing interface. In a more detailed analysis Fitzgerald & Woods (1994) have shown that in the absence of gravity long wavelength perturbations to a migrating liquid-vapour interface could be stabilized as a consequence of the compressibility of the vapour. In essence, newly formed vapour accumulates ahead of the interface, building up the pressure and suppressing the formation of an advancing perturbation on the interface. Also, short wavelength disturbances can be stabilized through the action of thermal diffusion. As a result, only perturbations of intermediate wavelength are unstable, and so perturbations of these wavelengths will grow on the interface.

4 EXPERIMENTS ON INSTABILITY

We have conducted a series of analogue laboratory experiments to examine conditions under which a migrating liquid-vapour interface becomes unstable. We injected liquid ether into packed beds of pre-heated sand, and recorded the evolution of the interface. Liquid ether was injected both from above and from below into a perspex vessel, of cross-sectional area $100cm^2$, and depth 40cm. The flow regime was recorded by video, and the mass fraction vaporising was calculated by comparing the mass supply rate with the observed rate of ascent of the interface. In the case of injection from below, from section 3, we expect that the liquid-vapour interface will remain stable. This is indeed the case, as may be seen in the photograph in figure 3(a). In our analogue experiments, the packed bed of sand was sufficiently fine-grained that the liquid and ether remained in thermal equilibrium. Thus, for a given descent speed and conditions under which the interface remains planar, the mass fraction vaporising follows the model of Pruess *et al.* (1987) and Woods and Fitzgerald (1993)

$$f = \frac{(1-\phi)\rho_s C_{ps}(T_s - T_b)}{(1-\phi)\rho_s C_{ps}(T_s - T_b) + \phi\rho_l(C_{pv}(T_s - T_b) + L)} \quad (4.1)$$

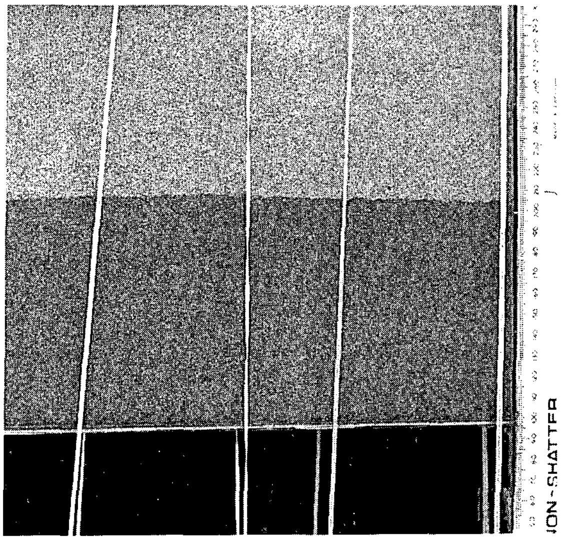

Figure 3 (a) Photograph of an ascending liquid-vapour interface. Liquid ether is supplied from below into a sand layer of approximately 50° C. The photograph shown was taken from an experiment conducted in a sand chamber of internal horizontal dimensions 30 cm by 1 cm. Although heat losses in this particular experiment were considerably greater than those conducted using the main apparatus described in the text, the morphology of the ascending interface was essentially the same in all experiments. The dark zone is the liquid-saturated region and the lighter zone the vapour-filled one.

as shown in figure 4, where C_p denotes specific heat capacity, T corresponds to temperature, L the latent heat of vaporisation, subscript s a property of the sand/rock, subscripts l and v correspond to properties of the liquid and vapour, and subscripts h and b to conditions in the hot superheated region and at the boiling temperature respectively. In the figure, we include theoretical predictions for the mass fraction vaporising in porous layers of porosity 0.4, 0.45 and 0.5, which represent bounds for the porosity in our laboratory experiments. The data were collected from a series of experiments in which the initial temperature of the sand bed was varied. The boiling point of ether is ~34° C.

When the liquid ether was injected downwards from above, the liquid-vapour interface became unstable, as suggested by the analysis shown in figure 2. The development of this instability is shown in the sequence of experiments in figure 3(b). The instability leads to the advance of non-linear fingers of liquid into the vapour zone, thereby significantly changing the vaporisation process from that predicted by equation (4.1). Indeed, we found that a broad two phase saturated zone develops as the interface descends, in contrast to the sharp interface which forms with an ascending interface.

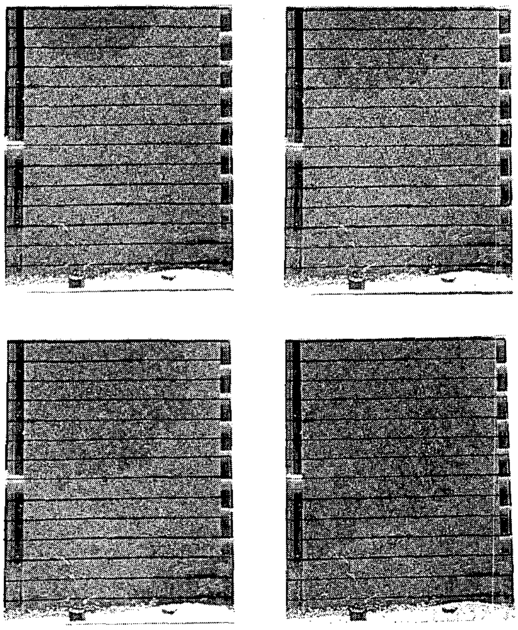

Figure 3(b) Sequence of four photographs of a descending liquid-vapour interface. Liquid ether is supplied from above into a sand layer of approximately 90° C. Photographs were taken at times 1, 3, 7 and 11 s after onset of the experiment. Fingers of liquid ether are observed to form and grow ahead of the uniformly liquid-saturated region. The dark zone is the liquid-saturated region and the lighter zone the vapour-filled one.

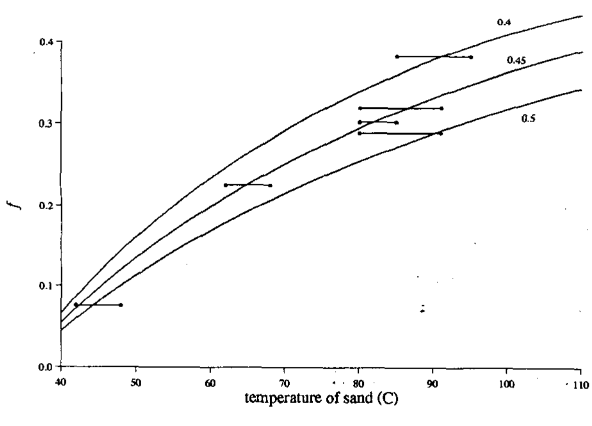

Figure 4 Mass fraction vaporising f as a function of sand temperature. Three curves are given representing the theoretical values for sand porosities of 0.4, 0.45 and 0.5 as indicated on the figure.

5 APPLICATIONS

The results of this analysis identify dynamic situations in which liquid zones may overlie and migrate into vapour-filled regions of rock. Although our model is very simple, and is based upon a number of simplifying assumptions, it is of interest to examine the consequences of our model, noting the rather severe restictions upon its applicability to a heterogeneous fractured reservoir.

For example, 22 years ago, in well SB10 at the Geysers reservoir, at points deeper than 350 m, the pressure gradient became significantly less than liquid-static, where the temperature was about 240° C and pressure about 30 bar (Truesdell & White 1973; White, Muffler & Truesdell 1971). Furthermore, below this point the fluid became more superheated. The well data of Enedy, Enedy & Maney (1991) and that of Enedy (1987) suggests that the permeability in the reservoir lies in the range $10^{-15} - 10^{-13} m^2$. Therefore, in the regions of lowest permeability, the simple analysis of section 3, combined with the relationship between the mass fraction vaporising and the injection rate (Woods and Fitzgerald 1993), predicts that the liquid-vapour interface would be stable for liquid velocities greater than about 2 m per year.

As a result of the net extraction of fluids from the Geysers reservoir, the pressures within the system have fallen (Kerr 1991) and values of 10 bar are now measured in various sections of the field (Enedy *et al.* 1993). This has increased the effective superheat of the reservoir. As a result, a greater fraction of the liquid may vaporize at a migrating liquid-vapour interface. For present conditions at The Geysers, we deduce that according to our analysis, a migrating liquid-vapour interface would be stable for liquid velocities greater than about 2.5 m per year in the regions of lowest permeability.

Conditions in various sections of the Larderello field in Italy are somewhat hotter. Temperatures of 300° C at pressures of around 30 bar are observed, (Pruess *et al.* 1987b), although we note temperatures as great as 394° C have been reported for the San Pompeo well (Truesdell 1991). The permeability variation within the field derived from well data indicates that values lie between $1-250 \times 10^{-16} m^2$ and hence our model predicts a migrating liquid-vapour interface is stable in the regions of lowest permeability for liquid velocities greater than about 0.2 m per year.

6 CONCLUSIONS

We have developed a simple model describing how liquid can migrate through a superheated geothermal reservoir. We have identified that if the liquid velocity is too low then the destabilizing force produced by gravity will exceed the resistance to motion within the liquid zone and consequently, the interface will become unstable. We have also identified that if the fraction of liquid which vaporizes is too high then the pressure gradient within the vapour region can exceed that in the liquid zone, thereby leading to instability at the interface. We have also demonstrated this instability with some analogue laboratory experiments in which large non-linear fingers develop.

7 REFERENCES

Axelsson, G. & Bodvarsson, G. 1987 Analysis of production data from fractured liquid dominated geothermal reservoirs in Iceland. *Trans. Geoth. Res. Counc.* **11**, 573–580.

Bear, J. 1972 *Dynamics of Fluids in Porous Media.* Dover.

Brown, G., Rymer, H., Dowden, J., Kapadia, P., Stevenson, D., Barquero, J. & Morales, L.D. 1989 Energy budget analysis for Poás crater lake: implications for predicting volcanic activity. *Nature* **339**, 370–373.

Cathles, L.M. 1977 An analysis of the cooling of intrusives by ground-water convection which includes boiling. *Econ. Geol.* **72**, 804–826.

Cline, J.S., Bodnar, R.J. & Rimstidt, J.D. 1992 Numerical simulation of fluid flow and silica transport and deposition in boiling hydrothermal solutions: application to epithermal gold deposits. *J. Geophys. Res.* **97**(B6), 9085–9103.

Donaldson, I.G. 1962 Temperature gradients in the upper layers of the earth's crust due to convective water flows. *J. Geophys. Res.* **67**, 3449–3459.

Dullien, F.A.L. 1992 *Porous Media Fluid Transport and Pore Structure.* Academic.

Dunn, J.C. & Hardee, H.C. 1981 Superconvecting geothermal zones. *J. Volcanol. Geoth. Res.* **11**, 189–201.

Elder, J. 1981 *Geothermal Systems.* Academic.

Enedy, S. 1987 Applying flowrate type curves to Geysers steam wells. *Proc. Stanford Geoth. Workshop* **12**, 29–36.

Enedy, S., Enedy, K. & Maney, J. 1991 Reservoir response to injection in the southeast Geysers. *Proc. Stanford Geoth. Workshop* **16**.

Enedy, S.L., Smith, J.L., Yarter, R.E., Jones, S.M. & Cavote, P.E. 1993 Impact of injection on reservoir performance in the NCPA steam field at the Geysers. *Proc. Stanford Geoth. Workshop* **18**.

Fitzgerald, S.D. & Woods, A.W. 1994 The instability of a vaporization front in a hot porous rock. *Nature* **367**, 450–453.

Fradkin, L.J., Sorey, M.J. & McNabb, A. 1981 On identification and validation of some geothermal models. *Water Resour. Res.* **17**(4), 929–936.

Goyal, K.P. & Box, W.T. 1990 Reservoir response to production: Castle Rock Springs Area, East Geysers, California, USA. *Proc. Stanford Geoth. Workshop* **15**, 103–112.

Grant, M.A., Donaldson, I.G. & Bixley, P.F. 1982 *Geothermal Reservoir Engineering.* Academic.

Homsy, G.M. 1987 Viscous fingering in porous media. *Ann. Rev. Fluid Mech.* **19**, 271–311.

Hurst, A.W. & Dibble, R.R. 1981 Bathymetry, heat output and convection in Ruapehu crater lake, New Zealand. *J. Volcan. Geoth., Res.* **9**(1), 215–236.

Kerr, R.A. 1991 Geothermal tragedy of the commons. *Science* **253**, 134–135.

Parmentier, E.M. & Schedl, A. 1981 Thermal aureoles of igneous intrusions: some possible indications of hydrothermal convective cooling. *J. Geol.* **89**, 1–22.

Phillips, O.M. 1991 *Flow and Reactions in Permeable Rocks.* Cambridge.

Pruess, K., Calore, C., Celati, R. & Wu, Y.S. 1987a An analytical solution for heat transfer at a boiling front moving through a porous medium. *Int. J. Heat Mass Transfer* **30**(12), 2595–2602.

Pruess, K., Celati, R., Calore, C. & Cappetti, G. 1987b On fluid and heat transfer in deep zones of vapour-dominated geothermal reservoirs. *Proc. Stanford Geoth. Workshop* **12**, 89–96.

Pruess, K. & Narasimhan, T.N. 1982 On fluid reserves and the production of superheated steam from fractured, vapour-dominated geothermal reservoirs. *J. Geophys. Res.* **87**, 9329–9339.

Rubin, A. & Schweitzer, S. 1972 Heat transfer in porous media with phase change. *Int. J. Heat Mass Transfer* **15**, 43–60.

Saffman, P.G. & Taylor, G.I. 1958 The penetration of a fluid into a porous medium or Hele-Shaw cell containing a more viscous liquid. *Proc. R. Soc. A* **245**, 312-329.

Truesdell, A.H. 1991 The genesis of vapour-dominated geothermal reservoirs with high-temperature zones. *Proc. Stanford Geoth. Workshop* **16**.

Truesdell, A.H. & White, D.E. 1973 Production of superheated steam from vapour-dominated geothermal reservoirs. *Geothermics* **2**, 154–173.

White, D.E., Muffler, L.J.P. & Truesdell, A.H. 1971 Vapour-dominated hydrothermal systems compared with hot-water systems. *Econ. Geol.* **66**, 75–97.

Wohletz, K. & Heiken, G. 1992 *Volcanology and Geothermal Energy.* California.

Woods, A.W. & Fitzgerald, S.D. 1993 The vaporization of a liquid front moving through a hot porous rock. *J. Fluid Mech.* **251**, 563–579.