NUMERICAL MODELING OF HYDROTHERMAL CONVECTION SYSTEMS INCLUDING SUPER-CRITICAL FLUID

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Key words: geothermal systems, hydrothermal systems, magma intrusion, natural convection, numerical modeling, super-critical fluid and Pruess, 1990; Hayba and Ingebritsen, 1994). We have developed a numerical simulator to model high-temperature hydrothermal convection systems which may include super-critical fluid in the vicinity of the magmatic pluton and purely heat conductive cooling of the pluton. In this paper, we describe the numerical simulator and present an example of modeling of such a high-temperature hydrothermal system.

General description of the simulator
All results presented here were obtained using Technical Software and Engineering's geothermal simulator known as SIMFoS (Fully Implicit Geothermal Simulator). SIMFoS is a general purpose three-dimensional model for mass and heat transfer in single- and multiple-porosity porous media, based on a fully-implicit (or optionally-sequential) finite-difference formulation. The model basically consists of two equations expressing conservation of mass of water/saturated and conservation of energy in porous media. The equations and solution procedures are the same as those described by Coats (1977). The procedure employs a variable substitution approach, where pressure and temperature are chosen as the solution variables for single-phase water and steam, and pressure and saturation are chosen as those for two-phase water and steam. The matrix equations comprising of 2x2 unknowns per grid-block are optionally solved using the D–ordered direct method (Price and Coats, 1973), the line successive overrelaxation method, or one of several preconditioning and acceleration schemes which employs the conjugate gradient type approach (Vinson, 1976; Cheshire et al., 1980, Oppe et al., 1988) for nonsymmetrical matrices.

The model simulates thermal, viscous, gravity and capillary forces, taking into account reservoir heterogeneity, geometry, variation in fluid properties, a host of different production–injection schedules, surface separator, two–phase flow behavior in production wells, heat losses from the under–burden, over–burden and adjacent side strata, aquifer influx from the surrounding regions, and other pertinent factors. Additional capabilities of SIMFoS include the treatment of double porosity (fractured reservoirs) systems using Warren and Root type (e.g. Gilman and Kosem, 1983) and Multiple Interacting Continua (MINC) concepts (Pruess and Narasimhan, 1985), and n–component flow (mass transport) suitable for tracer tracing applications; it has an efficient solution procedure for MINC flow (Seth and Hanano, 1995). The local grid refinement feature affords a more detailed resolution in the areas of interest. Heat transfer by conduction only, is considered for non–porous rock bodies. SIMFoS has been validated with several analytical results and also with the results of other reservoir simulators under subcritical conditions (e.g. Stanford Geothermal Program, 1980).

Improvement of the simulator
We improved SIMFoS to treat super–critical fluid up to 800°C and 1000 bars. Our goal was to keep the method simple, computationally stable, and easy to implement into any existing geothermal simulator. A careful study of the fluid properties of pure H2O reveals that there are two principal characteristics which pose computational problems:

(1) Fluid behavior around the critical point:
The properties change extremely rapidly in the vicinity of this region which renders the computational task highly unstable. Fluid saturations can undergo a change of 100% with a slight change in the temperature, pressure, or both.

(2) Rapid change in the properties across the subregions 2 and 3:

Sub regions identified as 2 and 3 in Fig 1 represent the region with the super–critical fluid (e.g. International Formulation Committee, 1967). The boundary denoted as \( \mathcal{B}_c \) divides these two regions. There is a steep density gradient across this boundary with a much higher density in the region 3. Such a steep change in the density manifests itself in the form of a large pressure change as a point traverses across this boundary, which in turn limits the time step size in numerical simulation. Studying the natural state of high–temperature hydrothermal convection systems entails simulation periods of several tens of thousands to a few million years, a reasonably large time–step size (of the order of a few tens to a few thousand years) is a practical necessity.

Apart from the above complexities in the fluid properties, there is nothing in the basic mathematical formulation of a conventional geothermal simulator which limits its applicability in all pressure and temperature ranges.

To numerically simulate fluid flow, under such a thermodynamic condition, Cox and Pruess (1990) employed a two–dimensional table of densities and internal energies as functions of temperature and pressure in a critical region, from the formulation developed by Haur et al. (1984). They also employed a bi–linear interpolation schemes to estimate values between table values. On the other hand, Hayba and Ingebritsen (1994) used pressure and enthalpy as variables to facilitate computational problems caused at the critical point. They used a look–up table for density, viscosity and temperature as a function of pressure and enthalpy. They used a bi–cubic interpolation.

To perform accurate numerical simulations, a code verification is very important. The original version of SIM.FIGS has been validated under conditions below the critical point, as described above. However, as described by Cox and Pruess (1990), it seems to date that no adequate experimental data is available for the code verification for the near–critical flow in porous media. Thus, the modified version of SIM.FIGS has not been verified yet above the critical point except for the conventional mass and energy balance checks. It is a very important task to be studied further.

**AN EXAMPLE RUN**

Our final goal of numerical modeling using the improved version of SIM.FIGS is to model the natural state of fluid circulation that includes heat conductive cooling of the magmatic intrusives following intrusion in order to understand the complete picture of the high temperature hydrothermal systems like Kakkonda. As a first step of such modeling, we employed a simplified symmetrical two–dimensional cross–sectional porous model for an example because of its simplicity. Fig. 3 shows its grid mesh, permeability distribution and bottom–boundary temperature distribution. To account for deep fluid circulation, we chose the depth and horizontal length of the model to be 6 km and 18 km, respectively. The thickness of the cross section was 1 km. Permeability distribution was basically two–layered, a relatively permeable shallow layer and a less permeable deep layer, based on the above–mentioned permeability structure of Kakkonda, Japan. All runs were carried out on a PC 486/66 and Sun Sparc Station 2 and 5 computing systems.

Boundary conditions at both sides were closed to mass and energy. A boundary condition at the top surface was permeable and heat conductive, i.e. it was open to mass and energy with a constant temperature, 1°C and constant pressure, 1 bar. The bottom boundary was closed to mass but was heat conductive at a constant temperature as shown in Fig 3. Since, no mass and heat input were imposed from outside of the model, the system was heated by an intrusive of 800°C at the deep left corner (Fig 3) and by the bottom boundary.
The initial condition was static with linear temperature distribution, 15°C at the top surface and 600°C at the bottom, excluding the intrusive; the temperature of the entire intrusive was 800°C. The initial pressure was 1 bar at the surface with a hydrostatic profile which corresponded to the initial temperature field. Thermal conductivity and rock heat capacity were 2 W/m-K and 2.6 MJ/m³K, respectively throughout the model. Porosity of the model was 10% excluding the intrusive. We assumed both porosity and permeability of the intrusive to be zero.

We ran the model up to 90,000 years. Changes of mass and heat in-place in the model over 90,000 years are shown in Fig. 4. As seen in Fig. 4, the heat in-place of the whole system once increased until 10,000 years due to heat gain by heat conduction from the bottom boundary, and then it decreased due to the out-flow of hot fluid and an in-flow of cold surface water to and from the surface. The mass change is inversely proportional to the change of heat in-place. The mass and heat changes approached a quasi-steady state at around 40,000 years, though there still remained a small fluctuation.

Fig. 4 shows the cross-sectional temperature distribution at 10,000 years. As seen in Fig. 5, convective hot up-flow develops at around X=4 and 5 (horizontal node number), around 1.8 km to the right, from the upper left corner; c.f. Fig. 3). Also, convective down-flow is notable at around X=1 and 2 (0 to 0.6 km to the right from the upper left corner). This pair of convective up-flow and down-flow forms a local hydrothermal convection. Fig. 6 is a temperature profile at X=5 (1.8 km to the right from the upper left corner) above 3 km depth. As seen in Fig. 6, there are two temperature layers of approximately 250°C in the shallow zone and approximately 350°C in the deeper zone. Thus, a two layered temperature structure perfectly similar to the situation of the shallow and deep reservoir in Kakkonada, Japan (e.g. Hanano and Takahashii, 1993; Hanano, 1995) is well reproduced. This implies that the model may represent a real high temperature geothermal reservoir.

Fig. 7 shows the cross-sectional temperature distribution at 90,000 years. In Fig. 7, convective up-flow is seen at around X=3 and 4 (around 1.4 km to the right from the upper left corner), and convective down-flow is seen at around X=1 (0 to 0.4 km to the right from the upper left corner), and X=7, 8 and 9 (around 3.2 km to the right from the upper left corner). Also, the low permeability area from X=14 to 22 (9 to 18 km to the right from the upper left corner) serves as the areal recharge zone, i.e., an areal convective down-flow zone, though its descending velocity is very small. Thus, there developed several roles of relatively small scale hydrothermal convection and one role of large scale hydrothermal convection to great depth. Super-critical fluid was widely present below 3.5 km depth.

Another notable point in Fig. 7 is the cool down of the intrusive. The intrusive at X=1 to 4 (0 to 1.6 km to the right from the upper left corner) and Z=7 to 12 (vertical node number; 3 to 6 km depth; c.f. Fig. 3) which was originally 800°C at the start of the run, cooled down to the order of 400–700°C; its temperature was much lower than that in 10,000 years (c.f. Fig. 5).

CONCLUSIONS

We have developed a numerical simulator capable of simulating geothermal reservoirs which may contain super-critical fluid up to 800°C and 1000 bars, to study deep high-temperature geothermal resources. To overcome computational problems associated with rapid changes in fluid properties around and within a certain region above the critical point, we employed interpolation in the vicinity of the region. The example run clearly demonstrates that we are able to model not only the total picture of the natural state of deep high-temperature hydrothermal convection systems but also thermal structure and purely heat conductive cooling of high-temperature magma intrusives.

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REFERENCES


Fig. 1 Illustration of subregions on the pressure–temperature diagram (Japan Society of Mechanical Engineers, 1981).

Fig. 3 Grid mesh, permeability distribution and bottom boundary temperature distribution.
Fig. 5. Temperature distribution at 10,000 years.

Fig. 7. Temperature distribution at 90,000 years.