DETECTION AND QUANTIFICATION OF DEEP GROUNDWATER FLOW USING 3-D GEOTHERMAL MODELLING
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Summary
Temperature in the Earth’s upper crust is dominated by conductive heat transport. However, advection of heat associated with groundwater flow can alter the purely conductive regime significantly. We present a method which exploits this fact to quantify groundwater flow from precise subsurface temperature measurements. We propose to detect areas disturbed by advection from computed temperature residuals, i.e. differences between temperatures measured in boreholes and calculated from a three-dimensional conductive model. The transient effect of paleoclimate is taken into account by subtracting its present signal from the measured data. If all processes are correctly considered, the calculated residuals represent the advective influence on the thermal regime. In order to obtain a quantitative estimate of groundwater flow, it is necessary to model the hydraulic flow regime. For this purpose we extended a 3-D conductive model into a fully coupled hydrothermal simulation. The model is calibrated using both the measured heads and the temperature residuals, which ideally should nearly vanish. We demonstrate the potential of the method based on data from the German part of the Western Molasse Basin (northern Alpine foreland basin). Preliminary results show the potential of the method to detect advective temperature anomalies qualitatively. Obtaining quantitative estimates is a topic of current research.

Abstract text
Flow velocity and discharge rate of groundwater are of major interest in many geoscientific applications. However, it is particularly difficult to estimate these quantities in deep aquifers with normally little information. Groundwater flow can change the temperature field in the Earth's upper crust significantly. This fact is used to identify areas where the thermal regime is disturbed by water flow (e.g. Vasseur and Demongodin, 1995). We compared temperature-data (corrected for the diffusion of paleoclimate) from an extensive database of the GGA-Institute with results from a 3-D conductive simulation. The differences between these two data sets reflect well the assumed nature of the flow regime.

Development of a 3-D heat-transport model
A conductive 3-D reference model was set up in order to calculate residual temperatures. The study area is situated northeast of Lake Constance in southwest Germany. The depth range of interest extends from 500m to 1200m below the surface.
Setting up a model for computer simulations requires (1) defining a geologically consistent structural model, with a resolution on the order of the scale of the phenomena studied and (2) deriving a conceptual model from the structural model which transforms stratigraphic and lithological information into physical properties. The “true” real geological structures are simplified during this process so that the resulting model is a physically and geologically meaningful approximation.

![Fig.1] 3-D view of the structural model.
The structural model (Fig. 1), used for the heat flow calculation, is based on simplified geological information obtained from a drilling database. In some cases it was necessary to include further information from geologic maps and seismic explorations. The model area is part of a sedimentary basin containing mostly Tertiary sediments from the Alpine orogeny. The basin is wedge-shaped with a moderate dip towards SSE. In order to reduce artefacts the entire model is rotated in this direction (20° NE).

**The numerical model**

Simulation of flow and heat transport requires the solution of the following set of equations:

The steady-state nonlinear heat conduction equation:

$$\nabla \cdot (k \nabla T) = 0$$

(1)

Modifying this equation in order to account for the effect of advective transport yields the steady-state heat transport equation:

$$\nabla \cdot \left( k \nabla T - \rho_c c_f (T) \nabla v \right) + H = 0$$

(2)

Specific discharge $v$ is obtained from the flow equation:

$$\nabla \left( \frac{\rho g k}{\mu} \left( (\nabla h_0 + \rho_s \nabla z) \right) \right) + W = 0$$

(3)

Here $k$ is thermal conductivity, $T$ temperature, $\rho$ density, $z$ the vertical coordinate taken positive upward, $h_0$ hydraulic head at reference temperature $T_0$, $\mu$ dynamic viscosity, $g$ gravity, and $H$ and $W$ denote heat and fluid sources or sinks. Subscript $f$ denote fluid properties, $\rho_r = (\rho(T, P) - \rho_0) / \rho_0$ is relative density.

Both equations were solved in three dimensions using the finite difference code SHEMAT (Clauser, 2003). The thermal conductivity is considered isotropic, and the temperature dependence of the rock thermal conductivity is implemented according to an empirical expression from Zoth and Hänel (1988).

The model extends 130 km x 95 km in the horizontal with horizontal and vertical grid sizes of 1 km (130 x 95 cells) and 25 m (280 cells), respectively. To study the influence of the grid resolution on the simulation results we also ran a model with half the grid size. This is of particular interest in cases where we included vertical fault zones.

Specific heat flow (lower boundary condition) is estimated to 86 mW m$^{-2}$ for the total model area (in case of temperatures corrected for paleoclimate). Values for thermal properties were taken from different sources, mainly from measurements on cores from two drilling-sites, performed in the laboratories at GGA and RWTH. Additional data were taken from literature.

These data were assigned to the 18 stratigraphic basic units of the structural model. The model was then calibrated by varying these values within reasonable bounds. These best fitting values are shown in Fig. 2. The mean ground surface temperature (upper boundary condition) was determined using a linear regression of the borehole temperatures corrected for paleoclimate which was then interpolated on the surface.

GGA in Hanover has been maintaining and updating a temperature data base for Germany since 1977 (Pribnow and Schellschmidt, 2000). Most of the data are bottom hole temperatures (BHT). There are other types of subsurface measurements and also many undisturbed continuous logs. We used for this study: 596 BHT values, 453 re-sampled values from undisturbed continuous logs, 3081 values from 15 continuous logs re-sampled to 5 m interval (Fig. 3), and 162 other temperature measurements (mainly drill stem tests).
Results

The temperature residuals were calculated by subtracting the simulated temperatures from the measured paleoclimate-corrected temperatures. To avoid artefacts, no interpolation of the measured temperatures was used. To analyse this result we developed a 3-D inverse-distance weighting program that allows interpolating the residual temperatures with respect to the different qualities of the input data (Fig. 3).

![Fig.3.](image)

<Fig.3.> Correlation between the calculated temperature residuals (measured minus simulated temperature) and major faults. The horizontal cross-section is 200 m below sea level which is approximately 700 to 800 m below surface. The numbered crosses mark the 15 log locations and the circles the points with other temperature data. The size of a circle is proportional to the vertical distance from this layer.

![Fig.4.](image)

<Fig.4> Another horizontal cross-section at 200 m below sea level showing first results of the modelled flow field. Decadic logarithmic scaling (base 10) is used for magnitude of the specific discharge in m/a, so that -3.5 corresponds to 0.03 m d\(^{-1}\) and -5 to 0.007 m d\(^{-1}\).

The largest temperature residuals in Fig. 2 show an obvious agreement with the position of the well-known large Fronhofen fault zone in the western Molasse-basin (x = 46500 m, y = 61000 m). Other faults also display temperature anomalies, although in a smaller measure. This result is in line with the usually high permeability along such fault systems. Further, aspect, slope and position of the residual temperature field match reasonably well with those of the sedimentary layers and of the karstic aquifer in the north-west (Swabian-Alb). Further this anomaly lies at the position where the horizontal cross-section crosses the upper Jurassic karst aquifer.

The fully coupled heat-transport model was parameterized with head data taken from different maps. The most important parameter for this steady-state model is the permeability which is estimated and regionalized using information available in the literature and some rare data. We modelled a number of different scenarios with
different lateral boundary conditions (no flow or fixed head) and with or without high permeability faults zones. The model could be calibrated to a satisfying degree, using measured heads.

One preliminary result of the fully coupled heat transport model is that for the flow-regime thermal-convection can have an important influence. Free convection can take place if the Rayleigh number Ra for flow in porous media

$$Ra = \frac{\rho g \beta k D \Delta T}{\mu a} \cos \theta$$

(4)

is larger than a critical value $Ra_c$ which is estimated at $\approx 40$ (Bachu, 1999), where $\beta$ is the thermal expansion coefficient of water, $a$ is thermal diffusivity of the saturated porous medium, $\Delta T$ is the temperature difference over a vertical distance $D$, $\theta$ is the angle of the aquifer with the horizontal, all values were assumed for $\beta = 10^{-3}$ K$^{-1}$ and $\mu = 0.5 \times 10^{-3}$ N s m$^{-2}$. We will evaluate the influence of potential free convection using estimated Raleigh-maps, suggested by Pestov (2000). But the incomplete knowledge of the conditions in the deep underground (> 3000 m below the surface) is still a major problem.

The only flow-model able to reflect the measured thermal field in a satisfying manner has high-permeability vertical faults. All other models do not show a thermal anomaly at the correct positions. This leads to the hypothesis that the observed temperature anomalies are a combined result of a high-permeability karst aquifer crossing high-permeability vertical faults.

Discussion

Although detection of advectively disturbed regions is possible, a quantitative interpretation still poses some problems such as:

- different structures, such as high-permeability vertical faults and high-permeability aquifers will produce nearly the same anomaly;
- irregularities like fractures and faults can interrupt the aquifer and change the entire flow regime (Bethke, 1989);
- some of the flow patterns are difficult to understand, for instance because of the Jurassic limestone karst aquifer;
- although there is lots of data, there are still gaps, for instance regarding the deep groundwater head or the regional distribution of the permeability in the aquifer.

In spite of the difficulties, the method presented here can be used to quantify groundwater flow and discharge in situations where other methods fail (Fig. 4).

www.rwth-aachen.de/geom/Ww/geothermik/geothermik.htm

REFERENCES


