Deep geothermal processes acting on faults and solid tides in coastal Xinzhou geothermal field, Guangdong, China

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Abstract

This paper investigated the deep fault thermal flow processes in the Xinzhou geothermal field in the Yangjiang region of Guangdong Province. Deep faults channel geothermal energy to the shallow ground, which makes it difficult to study due to the hidden nature. We conducted numerical experiments in order to investigate the physical states of the geothermal water inside the fault zone. We view the deep fault as a fast flow path for the thermal water from the deep crust driven up by the buoyancy. Temperature measurements at the springs or wells constrain the upper boundary, and the temperature inferred from the Currie temperature interface bounds the bottom. The deepened boundary allows the thermal reservoir to revolve rather than to be at a fixed temperature. The results detail the concept of a thermal reservoir in terms of its formation and heat distribution. The concept also reconciles the discrepancy in reservoir temperatures predicted from both quartz and Na-K-Mg. The downward displacement of the crust increases the pressure at the deep ground and leads to an elevated temperature and a lighter water density. Ultimately, our results are a first step in implementing numerical studies of deep faults through geothermal water flows; future works need to extend to cases of supercritical states. This approach is applicable to general deep-fault thermal flows and dissipation paths for the seismic energy from the deep crust.

1. Introduction

With elevated temperatures (one of the thermal water vent's temperature is as high as 98 °C), the Xinzhou geothermal field is known for its thermal water outflows associated with a deep faulting. Located in the coastal region of western Guangdong, it has a backdrop of the southern and south-eastern China geothermal belt, characterized by regional faults along the longitudinal, latitudinal, and northeastern-southwestern directions (Fig. 1a) (Lu and Liu, 2015).

The deep geological structure of the Xinzhou geothermal area has not been identified until recently. The Xinzhou geothermal area is classified as a medium-scale geothermal field (Tian, 2012). The hot spring waters are slightly alkaline and salty, and of CI-Na type (Table 1; Wang, 2013), mixed with the cold water and the seawater. The local area's ground waters are generally of HCO3-Ca, HCO3-Cl-Ca-Na, and SO4-Cl-Ca-Na types. The atmospheric precipitation is the source for the hot waters (Liang, 1993). However, the deep flow processes remain relatively unknown.

The thermal springs that occur in association with a fault zone have the potential to provide data on geochemical characteristics, physical states, and thermal outflows, all of which are useful in understanding deep flow processes and cyclic variations. In this study, daily and annual variations in the geothermal waters of Xinzhou geothermal field were recorded and modeled to understand the water flow processes of the system. The modeling study is considered as a vital tool used in understanding the underground geo-environments for the site.

Geothermal springs are commonly associated with fast flow paths that connect the deep crust to the surface (Curewitz and Karson, 1997; Zheng and Bennett, 2002; Sun and Lu, 2009) due to the horizontal impedance and vertical enhancement of the groundwater flow by a fault zone (Bredehoeft et al., 1992; Caine et al., 1996; Lu and Liu, 2015). Deep faults and fractured bedrocks, as infiltration and runoff channels, are critical control factors in the flow of hot waters to the surface (Wang, 2005; Gao et al., 2009). Faults and fractures are potential channels for seawater to intrude inland concurrently (Chen et al., 2001).
Fig. 1. Geological background map for the Xinzhou geothermal field in Yangjiang, Guangdong province: a. Regional tectonic map (Guangdong Province Geological Bureau Regional Geological Survey Brigade 1988; Chinese Academy of Sciences, 1959); b. Regional geological map for Xinzhou geothermal field (Geological information drawn from 1: 250,000 outline configuration diagram of Yangjiang city, 2004); and c. Water sampling sites. New Hole was a 1000 m deep scientific borehole (Wang et al., 2015). Cross section I-F shown in Fig.6.
As a realistic and emerging source, geothermal energy has attracted increased attention (Baioumy et al., 2015). Furthermore, geothermal energy is clean, renewable, and does not produce carbon dioxide (Smith, 2007; Younger and Gluyas, 2012; Craig et al., 2013). Studies on the potential of geothermal energy have implications in environmental geology. Deep and large faults are often linked to potential geothermal energy, as well as potential earthquakes (Skarbek and Rempel, 2016; Bai et al., 2013).

We studied the geothermal system in terms of the high temperature and high pressure states in the deep fault. We also investigated the upwelling of the geothermal water and the driving force along the deep fault in the Xinzhou geothermal field in order to find out the interplay of controlling factors. We did this in order to further understand how the deep system responds to changes of states. The results could potentially provide further insight into deep geothermal behavior in response to earthquakes.

The objective of this paper is to study the physical and chemical processes in the geothermal site via field site monitoring and numerical modeling. The site's description, conceptual model, and research methods are described (Sections 2 and 3), as well as the daily and annual variations in the outflows (Section 4). We also estimated the thermal reservoir temperature and calculated the reservoir depth (Section 5). The modeling study is presented (Section 6), and the findings are discussed and summarized (Sections 7 and 8).

### 2. Site description and conceptual model

#### 2.1. The regional geology

Geothermal fields in the Guangdong province are distributed in such a way that reflects the pattern of the regional deep faults in the southeast coast of China (Fig. 1a). Geophysical surveys show that the Enping-Yangjiang deep fault, located on the western side of the Pearl River estuary, reaches as much as 20 km in depth (Ren et al., 2011).

The Xinzhou geothermal field is located in the southwestern side of the Enping-Kaiping deep fault and the eastern side of the southern section of Yangjiang-Guangzhou-Conghua deep fault. The internal structure of the Xinzhou geothermal field is well developed, and the hot springs are exposed in the junction region of the two faults (Wang, 2013).

The Xinzhou geothermal area exposes the Yanshan II granite in the southern part of the geothermal fields and sees outcrops sporadically in the north. The granite forms the northwestern edge of the Xinzhou rock mass.

#### 2.2. Site description

The Xinzhou geothermal field is located on the northwestern part of the semi-circular Xinzhou basin. The Xinzhou basin is surrounded by low mountains on the eastern and southeastern edges. The altitude of the Xinzhou geothermal field is about 10–13 m (Fig. 1).

Quaternary sediments cover the Xinzhou geothermal field. The basement granites are buried on a shallow level and their outcrops can be seen in ditches and ponds. To the north of the site is the boundary of Precambrian-Cambrian light metamorphic elastic rocks (Fig. 1c).

Hot springs in Xinzhou are naturally exposed along the stream bed. Hot water flows from the boreholes and oozes from the pond bottoms. The Jia well is located in the middle of the field and the thermal water outflow can reach up to 98.4 °C, with this being the highest temperature among the springs.

The field site has a deep fault at a high angle (almost vertical at 97°), tipping to the south. The fault was initially revealed in drillings at an earlier time (Liang, 1993). Recently, the fault was further characterized by the geophysical method of Audio Magnetotelluric Sounding (AMT) (Wu, 2013) and was later confirmed in a 1000-m scientific drilling in 2014 (Wang et al., 2015).

The advanced geophysical method's apparent resistivity and impedance phase have revealed the faulting to extend to a depth about 10 km into the crust (Wang et al., 2015).

### 2.3. Conceptual model

The flow system is conceptualized as being well connected to the deep heat source through the deep fault F1 (Figs. 1 and 2). The groundwater is recharged through the rain water infiltration, and generally flows from the mountain to the ocean, while the seawater intrudes through the tidal zone to mix with the groundwater. The geothermal flow can create a relatively low pressure zone around the fall zone, thus affecting the groundwater flow system (Fig. 2a).

The local scale of the groundwater could affect the geothermal flow, in addition to the regional scale of the groundwater flow from the mountainside circulating to the ocean. However, the local groundwater tends to flow shallowly and travel a short distance before discharging into the nearby stream bed. Thus, the local flow can be assumed to have a relatively small effect on the geothermal flow in terms of the regional scale circulation.

The deep fault, perceived as a tension fault, serves as a fast flow path for the flow and heat transfer. It progressively gets hotter as it goes deeper into the crust. The hotter water from the greater depths drives itself up onto the ground surface because of its buoyancy.

The temperature in the deep ground can reach beyond a supercritical condition (374 °C at 22.5 MPa). In addition, the geothermal temperature at the Xinzhou area reaches the Curie temperature point of 585 °C (Zhang et al., 2000) at around 21 m in depth (Wu, 2013).
Rain water replenishes the groundwater and lowers the groundwater temperature in general. The groundwater interacts with rocks along its flow path, making it possible to trace the sources of waters.

The solid tides cause cyclical displacements of the crust and lead to pressure variations in the crust. The pressure changes cause physical state change in the geothermal water, resulting in the cyclical variations in outflows of the thermal spring wells.

3. Methods

3.1. Field work

The field work measured geothermal water outflows of wells for physical parameters. Water outflows were investigated from January 2013 to January 2014. Water temperatures were measured with mercury thermometers with 0.1°C resolution. Thermometers were calibrated against each other. Flow rates were measured with buckets and a stop watch. The well outflows provided direct access to geothermal water from the deep circulation for convenient water samplings and flow measurements. Because of potential mixing with shallow groundwater along the upward flow path to the ground surface, geothermal outflows may differ from those in the deeper grounds. Continual measurements using instruments for flow rates were not feasible because of the low tolerance of temperature for the built-in sensors in addition to the drifting that occurs on the readings.

The outflows were measured for flow rates and temperatures, in addition to a set of physical parameters such as the pH, Eh, dissolved oxygen, and electrical conductivity. In addition, the outflows were monitored for their annual variation and daily fluctuations. The annual variation is expected to yield information on rain infiltration, and the daily fluctuations are relevant to the solid tidal effect on the deep fault zone. The fluctuations or daily variations could potentially reflect the flow path, i.e., the fault zone or fracture zone. This is a new approach in terms of data interpretation for deep geothermal flow.

The geothermal water flow is affected by temperature. This warrants a non-isothermal consideration of the geothermal field.

3.2. Mathematical model

The geothermal processes in this study require mass and energy balance equations for fluid and heat flows in general multiphase (liquid, gas in this study) and multi-component systems. Fluid advection is described with a multiphase extension of Darcy’s law. Heat flow occurs by conduction and convection. Thermodynamic conditions are described on the assumption of local equilibrium of all phases.

The balances of the basic mass and energy can be expressed in a generalized form:

$$\frac{d}{dt} \int_{V_n} M^K \, dV_n = \int_{V_n} F^K \cdot n \, d\Gamma_n + \int_{\partial V_n} q^K \, d\Gamma_n$$

The integration is performed on an arbitrary volume $V_n$ enclosed by surface $\Gamma_n$. The $M$ in the summation on the left side represents the mass or energy in a unit volume, with $K = 1, \ldots, NK$ for mass components (water, air), $\mathbf{F}$ for mass flux and $q$ is the sink, or source. $n$ is the normal vector on the surface element $d\Gamma_n$ pointing to the inside of the $V_n$.

The generalized form of mass accumulation terms is summation of fluid phases:

$$M^K = \phi \sum_\beta S^K_\beta \rho^K_\beta X^K_\beta$$

The total mass for component $K$ is obtained from summation of fluid phases $\beta$. $\phi$ is porosity, $S^K_\beta$ is saturation of fluid phase $\beta$, $\rho^K_\beta$ is density of fluid phase $\beta$, and $X^K_\beta$ is the fraction of component $K$ in fluid phase $\beta$. Advection mass flux term $F^K_{adv}$ is the summation of fluid phases:

$$F^K_{adv} = \sum_\beta X^K_\beta \mathbf{F}_\beta$$

where $\mathbf{F}_\beta$ is the flux of fluid phase $\beta$ and is given by the Darcy’s law for multiphase flow:

$$\mathbf{F}_\beta = \rho^K_\beta \mathbf{u}_\beta = -k^K_\beta \frac{\rho^K_\beta}{\mu^K_\beta} (\nabla P^K_\beta - \rho^K_\beta g)$$
Here \( u_b \) is the Darcy velocity (volume flux) for fluid phase \( b \), \( k \) is absolute permeability, \( k_{rel} \) is relative permeability of fluid phase \( b \), \( \mu_b \) is viscosity. And \( g \) is the gravity acceleration. Fluid pressure (fluid phase \( b \)) \( P_b \) is given as

\[
P_b = P + P_{cb}
\]  

(5)

where pressure \( P \) for reference phase (generally taking gas phase) and capillary pressure \( P_{cb} (\leq 0) \).

Heat flux \( F^{NK + 1} \) includes conductive and convective components are given as

\[
F^{NK + 1} = -h \nabla T + \sum \beta h_b F_b
\]  

(6)

where \( h \) is thermal conductivity, and \( h_b \) is specific enthalpy in phase \( b \). \( T \) is temperature.

The mathematical model was implemented with the multi-phase flow simulator TOUGH2 from Lawrence Berkeley National Laboratory (Pruess et al., 1999). TOUGH2 is a non-isothermal flow and heat simulator, which has been widely used in the world. The simulation uses the equation of state (EOS3).

4. Daily and annual variations of geothermal flows

4.1. Daily flow rate and temperature

Flow variations reflect responses of the deep geothermal region to tidal waves. Monitored wells were selected based on the flow rates and accessibility. Daily monitoring was carried out on two wellsprings (the Jia well and the Dun well) (Fig. 1c) for variations in temperatures and outflow rates (Fig. 3). The Dun well has a smaller flow rate and a lower temperature, while the Jia well is among the highest in flow rates and temperature. The Dun well has a small flow rate of about 0.057 L/s, a temperature from 94.2 °C to 94.7 °C (Fig. 3b). Outflow temperatures were on an increasing trend and were the highest from 2 PM to 3 PM, while the flow rates were at a low level. From 1 AM to 2 AM, the outflow temperature tends to decrease, followed by another ascension when the flow rates are rising to the maximum of the day.

The Dun well’s outflow temperature changed in synchronization with flow rates (Fig. 3b). The flow rates change as the outflow temperatures change, implicating that outflow volume might be controlled by temperature. The solid earth tidal wave is also plotted in Fig. 3 to facilitate comparisons. The daily variations appear to be related to the tides. The flow and tides will be better matched with each other if we take into account the delayed responses of tidal fluctuations. A tidal peak corresponds approximately to a valley in solid earth tides (IERS, 2003; Ekman, 1989).

The Jia well, not far from the Dun well, shows much larger outflows and higher temperatures (as high as 98 °C) than the Dun well (Fig. 3a). There are two water venting holes in the Jia well, with one at half a meter above the other in a steel piping structure. In order to collect the real outflow rate data, we monitored the flow rates of the upper opening and temperatures for both the upper opening and lower outlet. The temperature of the upper opening is slightly higher than that of the lower outlet (Fig. 3a). The upper opening has lower pressure, resulting in a slightly higher temperature. This is consistent with the result in the modeling section (Fig. 10).
When pressure is lowered, the temperature rises slightly. An alternate explanation is that the upper opening is directly on the pipe wall, leading to a higher measurement in temperature. The lower opening is running through a tunnel-like passage of about 30 cm long of calcite and sodium-chloride salt, leading possibly to a reading of temperature lower than that of the upper opening. In addition, the lower opening is under higher pressure due to its lower position and the out-gushing venting of thermal water.

In contrary to the Dun well, temperatures of the Jia well are reduced between 2 PM and 3 PM, but increases in the morning of the next day. Although the trend seems to be the opposite of that of the Dun well, the variations are well synchronized with the changes of the solid earth tide.

4.2. Annual flow rates and temperature

The annual variations were monitored for the Jia well and the Dun well (Fig. 4). The temperature differs in summer and in winter for both wells. The fluctuation is about 2°C. The temperature and flow rate for the Dun well decreases from April to October and slowly rises after October, which results from the decline of rainfall. However, the flow rate of the Jia well decreases in May. Following this, the flow rate increases from October to December, followed by a slight decline past December. The outflow temperatures for the Jia well trends opposite to the flow rates.

The New Hole Well was a thousand-meter deep scientific drilling borehole. The hot water started to flow out on November 7, 2013.
7, 2013, causing slight decreases in the flow rates of the Dun well and Jia well. For the Dun well, the flow rate changed from about 0.58 L/s to 0.44 L/s after November 7, and rose back to 0.56 L/s on November 11. The same changes can be seen in the Jia well. The flow rates of the upper opening in Jia well were about 0.50 L/s before November 7 and after November 11. However, in between those two dates, the flow rate fell to 0.46 L/s. The drop in flow rate for both wells may indicate that the creation of the New Hole well had impact on flow rates of the other two wells, but the effect reduced in value after 3–4 days.

5. Thermal temperature at deep ground and groundwater circulation

Groundwater circulation of the deep ground is enhanced at a geothermal field. Geothermal water gets hotter and lighter in deeper ground and is driven up due to buoyancy. Temperatures of the reservoir fluid can be estimated based on temperature-dependent mineral fluid equilibrium.

Thermal reservoir temperatures of Xinzhou geothermal system were estimated by several commonly used thermometers. The geothermometers specifically include quartz (Verma and Santoyo, 1997), Na/K geothermometer (Giggenbach et al., 1983), and K-Mg geothermometer (Giggenbach, 1988). The estimated thermal reservoir temperatures are listed in Table 1.

The estimated K-Mg temperatures are generally lower than that of Na-K and Quartz temperatures, which can be explained due to mixing with the Mg-rich cooling water when flowing up to the ground surface. The Mg content of geothermal water decreases as temperature increases. Therefore water with elevated Mg concentrations would have been involved in low-temperature water-rock equilibriums, probably in a near surface environment.

According to the reservoir temperatures of quartz, we calculated the reservoir depth (Table 1) using the annual average temperature of Guangdong at 22.5°C, constant temperature zone at 13 m and geothermal gradient 2.49–4.5°C/100 m. Chalcedony was also used in validating the reservoir temperatures, yielding very close values to the ones from quartz (Fournier, 1977). Na-K combination would possibly yield higher values. Albite and K-feldspar release Na and K under high temperature. Considering that Xinzhou field is in the coastal region and possibly affected by seawater components, the applicability of thermometers is therefore complicated. Discussion of the limitations or applicability for these thermometers is beyond the scope of this work.

Field water samples were also evaluated in the Na-K-Mg1/2 triangular diagram (Fig. 5) for equilibrium states and different water types. The cation geothermometers (Na-K temperatures and K-Mg temperatures) were used to calculate the reservoir temperature. In the triangular diagram, the geothermal water samples from Xinzhou and Shenzao are plotted in the partially equilibrated or mixed area of the triangular diagram, while the cold water samples are plotted in the immature area. This indicates that Na-K temperatures and K-Mg temperatures can be applied to calculate the reservoir temperature, but the results might be affected by mixing.
Geothermal fluids under Xinzhou geothermal field most likely achieved only partial balance with wall rocks; there is also a chance that parent geothermal fluid is mixed by regional groundwater in the process of rising to the Earth surface. In comparison, two samples from Ruihai (Guo and Wang, 2012; the largest and most active hydrothermal area in Tengchong) were plotted around the full equilibrate line. It has been confirmed that the magma chamber exists below Tengchong, and the parent geothermal fluids makes it possible to reach the balance of water-rock interaction.

Fig. 7. Simulated results of regional shallow ground water flow at the Xinzhou thermal spring field: a. temperature field, b. pressure field, c. the density of geothermal water, d. temperature gradient, and e. pressure gradient. See Fig. 8 for the results of the corresponding deep region.

Fig. 8. Numerical simulation results of regional deep geothermal water flow at the Xinzhou thermal spring field: a. temperature field distribution, b. pressure field distribution, c. the density of geothermal water distribution, d. temperature gradient, and e. pressure gradient. See Fig. 7 for the temperature distribution of the corresponding shallow region.

6. Modeling study

Single continuum is assumed in generating the grid (Pruess et al., 1999). This rock model for granites does not distinguish fractured medium from porous medium. The reasons are twofold—one being our singular focus in the steady state flow process and the second due to lack of fracture characterization data. The Van
Genuchten-Mualem model (van Genuchten, 1980) is used for the relative permeability function. The parameters in van Genuchten notation are 0.47, 0.07, and 1.0, for the van Genuchten m, the liquid residual value, and the liquid saturation value, respectively. Capillary pressure is calculated with the van Genuchten function, and the maximum of negative capillary pressure is set at $1.0 \times 10^6$ Pa. The model parameters of permeability are summarized in Table 2.

6.1. The model settings and boundary conditions

The Xinzhou Hot Spring model covers the recharge area of upstream, the discharge area of downstream, the coastal area, and bottom part of the deep geothermal area (Profile I-I’ in Fig. 1b). Taking a full view of the recharge and discharge process, the rainfall infiltration segment is 11 km long, the low-lying stream segment is 9 km long, and the offshore portion is 10 km into the seawater from the coast. The transverse span of the stream reach is set to 222 m. The model is 7.2 km deep, including the high-temperature and high-pressure zone (Fig. 6a).

This site is conceptualized as a quasi three-dimensional model. The shallow groundwater can be simplified as mainly longitudinal flow advecting to the sea, and lateral-flow is simplified as being uniformly treated for its eventual convergence to longitudinal-flow as it flows towards the sea. Considering the key focus in simulation of deep geothermal processes, this approximation is deemed appropriate.

The simulated domain is divided into 99 columns along the stream flow, with 19–37 rows across the stream and 75 total layers, creating a total number of 176,493 gridblocks, including those for boundary conditions (Fig. 6). In the plan view, the upstream area has 37 rows and the downstream has 19 rows. The gridding is structured to effectively reflect specific simulation objects, such as fault zones, top layers and boundary nodes. The minimum interval of each column is 0.5–1 m in fault zone and 250 m on the boundaries on both sides. The maximum interval of seawater intrusion in the sea section and the low-lying stream segment is 1000 m. The interval decreases gradually from the sea to F1 and F2 fault zones, which is represented as 1.0 m wide dividing the opposite fault walls. In the vertical direction, from the surface layer to a depth of about 120 m, the grid is made up of small spacings; the smallest of which could be down to just 0.5 m thick. At the same time, to give consideration to the heat source, gridblocks are constructed down into 1 m thick sections. In order to prevent potential iteration convergence problems, the variation of adjacent grid spacings is generally less than a ratio of 1.75 for a smoother transition from side to side.

Boundary settings have taken full advantage of natural boundaries. The land part of the upper boundary serves both as a constant pressure boundary fixed at atmospheric pressure and as an infiltration boundary, set at an infiltration rate of 357.5 mm/y, which accounts for 15% of the average annual rainfall of the region (2383.2 mm/y). The low-lying stream of the estuary and the seabed both serve as the boundary for the fixed water level. The depth of sea at 10 km away from the coast is 10 m. The bottom boundary, at a constant temperature of $300^\circ C$, is set at a depth of 7200 m. The bottom temperature value is approximated according to the curie temperature interface (about $585^\circ C$) at a depth of about 21 km deep (Wu, 2013). Rainfall infiltration is given to the top gridblocks...
in the land section. The interface density is calculated by using upstream weighting. The upstream boundary, utilizing the local watershed divide, is set as a no-flow boundary. Marine boundary nodes were set at a fixed pressure, based on a freshwater head converted from seawater pressure at the center of each gridblock. The boundary nodes below the seabed are set as no-flow boundaries and are allowed to adjust their pressures. The marine boundary is set far away enough from the coast so that the seawater salinity is stable enough to form a waterhead to stop the fresh water from blending with the ocean. It is worth noting that the geothermal water is allowed to flow out according to the filtration conditions of the top boundary, so the boundary conditions are similar to those of natural geothermal waters under no exploration.

6.2. Model result and calibration

The model ran for a sufficient enough time to be considered to have a steady state. The modeling results are presented in shallow portion (Fig. 7) and deep portion (Fig. 8), in terms of temperature, pressure and water density. The profiles are drawn along the central gridblocks in the stream-flow direction.

The model was set up as a simplified 3D model for the deep geothermal system. The model scale is 30 km long and 7.2 km deep. In a planner view, the geothermal field is only about 500 m along the stream and about 75 m across. In this aspect, some simplification is necessary. From the regional flow direction, groundwater circulation would have led to deep circulation, which would quench the geothermal flux. However, the geothermal flow along the fault zone presents a cone of a higher pressure zone in the upper portion (Fig. 7b). This leads to the circulation of regional groundwater to a deeper depth than what a normal groundwater flow would otherwise with no geothermal effect involved. In the field site, the stream reach (from north to south) is narrow in the cross-sectional spread and just as wide as the size scale of the geothermal field. This means that the regional groundwater along the stream has to go through the thermal field. In this sense, a quasi 3D approximation for the flow along the regional direction is reasonable.

The flow in the thermal field is simulated as zero outflows occur from individual boreholes. This is equivalent to the boreholes being plugged. This corresponds to the flow field before drillings (Liang, 1993). This is warranted, given that the flow rates are equivalent to the waterhead for each water-venting borehole. For example, during the drilling of the 1000 m New Hole, the water column in the new borehole was cooler and created pressures equivalent to those surrounding rock walls of elevated temperature.

The model calibration was performed by adjusting the fault properties, specifically in permeability, by matching the temperature profile data for the geothermal field (Table 2; Fig. 9). The upper part was constrained by the 1000 m borehole data. The temperature data was field measured during drilling for a couple of scenarios: temperature recovering period for temperature measurements and flowing condition. The medium height of the thermal field is validated by the Currie temperature at corresponding depths, which is calculated using a certain range of thermal gradients. The bottom temperature is supported by a temperature interpreted from the Currie temperature interface.

7. Discussions and implications

7.1. Deep geothermal water

Deep geothermal water in states of high pressure and high temperature is distinctive in low density (Fig. 8). It has a tendency to move up owing to its low-density-inducing buoyancy. The regional deep fault extends deeply underground and becomes a fast flow path for the rising deep geothermal water, which changes the pattern of regional groundwater flow. Geothermal water owing to driving buoyancy can effectively flow to the shallow parts through the deep fault (Qi et al., 2017).

Density distribution of thermal waters can be demonstrated by the profile of a static water column (Table 3). Because of the low density at the bottom of the water column, the cumulative buoyancy of the thermal water could be so large that a small change in pressure could result in buoyancy fluctuation. The geothermal water around the fault has a density of about 700 kg/m$^3$ at the deep-seated area (Fig. 8c). The temperature is slightly higher in the deep fault zone compared to the outside at the same depth, where the pressure is relatively lower. This is the main reason why the density is abnormally low. The deep fault partially makes the pressure dissipate in the deep-seated area, resulting in significant drops in pressure.

At the middle and higher portion of the profile, the pressure distribution of geothermal water is lower in the hot spring region than in the other regions. This tendency is due to the regional effect caused by the deep fault (Fig. 8c). The affected region is about several kilometers deep and may extend up to just under the low-lying riverbed area, making it favorable to potential intrusion by the seawater. Regional groundwater forms a low pressure field in the hot spring area, which makes the regional water flow to the hot spring area at a depth of 1000 m. This result is obtained using the hydraulic conductivity field of granites approximated by a single continuum. Because the granites contain partial fractures, a more realistic two continua representation of the fractured granites could widely spread the low pressure zone in the hot spring area.

The temperature of the fault zone has a large gradient value at the shallow region within one hundred meters, while it has a smaller value in deep area (Fig. 8d). The fault and fractures are high in

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Table 2

<table>
<thead>
<tr>
<th>Layer No.</th>
<th>Rock types</th>
<th>Lithology</th>
<th>Porosity</th>
<th>$k_x$ (m$^2$)</th>
<th>$k_y$ (m$^2$)</th>
<th>$k_z$ (m$^2$)</th>
<th>Heat Conductance (W/m/$^\circ$C)</th>
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<tr>
<td>1</td>
<td>Q$^{sh}$</td>
<td>Fine sandy clay</td>
<td>0.25</td>
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<td>8.5E–14</td>
<td>8.5E–15</td>
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<td>Clayey fine sands</td>
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<td>Course sandy gravels</td>
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<td>8.5E–12</td>
<td>8.5E–13</td>
<td>1.9</td>
</tr>
<tr>
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<td>Q$^{mc}$</td>
<td>Sandy fine sands</td>
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<td>Strongly weathered granite</td>
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<td>1.5E–14</td>
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<td>Intermediately weathered granite</td>
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<tr>
<td>8</td>
<td>F2 Fault</td>
<td>Fault zone$^a$</td>
<td>0.20</td>
<td>5.9E–13</td>
<td>5.9E–13</td>
<td>5.9E–14</td>
<td>3.5</td>
</tr>
<tr>
<td>9</td>
<td>F1 Fault</td>
<td>Deep fault zone$^a$</td>
<td>0.35</td>
<td>1.3E–12</td>
<td>1.3E–12</td>
<td>1.3E–13</td>
<td>3.5</td>
</tr>
</tbody>
</table>

Note: a. Permeability data refer to site report (Wang et al., 2015), rock density 2100 kg/m$^3$ for Quaternary sediments, and 2650 kg/m$^3$ for granite; heat conductance takes common value. b. $k$ for fault zones calibrated by this model.
permeability, so geothermal water can ascend due to its own buoyancy. Taking all this into account, it follows that the fault and fractures are the fast flow channels for efficient geothermal flow ascending to the shallow layer.

The pressure gradient is believed to have led to the above temperature profile of higher gradients at the top and lower gradients at the bottom (Fig. 8e). The pressure gradient presents the highest value where the top part of a lightly weathered layer meets the bottom part of a heavily weathered shallow sedimentary layer. Due to the low permeability of lightly weathered granite and the dependence of geothermal flow on conduction, the pressure gradient is large. However, the strong water-transmitting ability of faults and fractures causes heat to transfer mainly through advection of geothermal water, resulting in a small geothermal temperature gradient at the bottom region (Fig. 8d). In comparison, the top region sees an abrupt steepness in geothermal temperature, a result of conduction dominant heat transfer.

### 7.2. Solid tides and pressure effect on the geothermal water

Tidal surface displacement moves the ground periodically and acts upon the crust accordingly at shallow and deep depths. Upward movement of the ground creates tension within the crust, corresponding to reductions in pressure within the rock. Downward movement compresses the crust, resulting in an increase of pressure (Métiévier et al., 2009). Tidal deformations vary in directions and magnitudes. However, they are predominantly vertical within the lithosphere and therefore are mostly radial and larger close to the surface. This means that tidal pressure within the lithosphere coincides with tidal surface displacement (e.g., Smith, 1974; Wahr, 1981). As a rule of thumb, the increments of pressure are around a magnitude of less than \(4 \times 10^{-3} \text{ bar} \) compared to earthquake stress drops of \(~1 \text{e}5–1 \text{e}7 \text{ bar}\) (Vidalé et al., 1998).

Field observations show changes in flow rate and temperature for geothermal water responding to solid tides. An increase in downward displacement pressure leads to lower flow rate and higher temperature for the Jia well (Fig. 3a); whereas for the Dun well, the opposite result of higher flow rate and lower temperature is true (Fig. 3b). When pressure decreases due to uplifting ground, Jia well responds with a higher flow rate and lower temperature, whereas Dun well responds with lower flow rates and higher temperatures (Fig. 3).

Pressure changes and the corresponding responses of the geothermal system were simulated using the site model discussed in Section 6. Pressure variation was implemented with an additional pressure which would be added or subtracted to the model domain. Specifically, the pressure was modified for the gridblocks in the whole domain, including the boundary gridblocks. The exception was the top boundary gridblocks, which remained at atmospheric pressure.

The bottom part of the fault zone would see a rise in water density as pressure increased and vice versa. This can also be said for the other bottom, lower and middle parts of the whole regional flow field.

Dun well reaches 200 meters deep and could provide insight into deep thermal processes (Figs. 3 and 10). The fault zone would have lighter waters at a greater pressure, leading to greater buoyancy and thus translate into greater flow rates for Dun well. As the buoyancy goes up, the thermal water acquires higher temperature; taking into consideration the delay of thermal water reaching the ground surface, this explains the peaking of pressure immediately after the bottoming out of the downward displacement.

The top layers have depth-sensitive responses quite opposite to the middle and lower parts of the fault zone (Fig. 10a). When pressure decreases in the uplift, the temperature increases and the thermal water gets lighter. The Jia well is shallow and could be affected by shallow groundwater. This could provide clues to the dynamics observed in Jia well. When the uplift nearly reaches the peak, the thermal temperature is at its lowest; when the uplift is down, the thermal temperature then reaches the highest temperature. Looking at the flow rate, it becomes higher and then peaks when uplift increases and reaches the top.

In the top layers, the temperature rises significantly when the pressure decreases (Fig. 10a). This could be a result of lowering pressure (Pruess et al., 1999).

### 7.3. Implications and geothermal reservoir

The simulation has shown that temperature increases rapidly at the top part and increase at a slower pace below the top (Fig. 9). The top part is believed to be affected by groundwater flow circulation which effectively removes heat by advection. At the deeper zone, temperature increases linearly; the heat is dissipated through upward thermal diffusion. In between lies the transition zone that marks the weakened dominance of groundwater circulation and top of the thermal diffusion zone supplying heat from the deep source down under. The so-called transition zone possesses properties of a geothermal reservoir.

The geothermal reservoir in the Xinzhou geothermal field, consisting of the fault zone and neighboring wall rocks, is implicated in temperature distribution and possibly aqueous chemistry. The reservoir temperature has its core temperature in the fault zone and bulk heat dissipating area around it. The reservoir occurrence is consistent with the predicted depth from the water chemistry of
field samples. The fault zone is at the core of the thermal reservoir and is consistent with the Na-K-Mg temperature. The fault wall, immediately adjacent to the fault, is the wing of the reservoir at a temperature predicted by quartz temperature (Table 1 and Fig. 9). In other words, this could imply that reservoir mineral-water interactions are kinetic in nature. In addition to the current heat and flow modeling, a reactive transport approach is warranted in a further study on reservoirs.

The calculated result indicates that invasive seawater can invade deep underground under certain favorable hydrodynamic conditions, and could seep further into the region where the hot spring geothermal water cycle occurs. Finally, the sea water intrusion could make the salinity of the hot spring water increase. This could theoretically explain why the hot springs have a relatively high salinity in coastal regions, caused by modern seawater intrusion facilitated by geothermal systems.

Temperature gradients are the largest at the shallow part and the smallest at the bottom part (Figs. 7d and 8d). The gravel layer has a larger gradient than the clay layer in the shallow area. Although the thermal conductivity coefficient of the gravel layer is a little higher than the clay layer, a greater porosity makes the gravel have a higher moisture content, leading to a less effective heat transfer. The general trend in the temperature gradient of deep geothermal water is smaller toward the bottom, and becomes the smallest at the bottom of the fault (Fig. 8a).

At the bottom of the fault, the temperature profile has shown a significant drop from the fault to the upper stream side at about 6500 m depth. This could indicate the lowest point that the water circulation has reached. The deep circulation point could have been affected by the bottom boundary which was set at 7200 m at depth. It would require an improved representation of a full three-dimensional (3D) regional groundwater flow to substantiate in future work.

The fault deep fault property was obtained from matching the temperature profile. A fault property representing regional flow in the fault zone and deep ground could provide data for further research in the study of physical states in the deep crust. A high porosity fault zone favors propagation of porosity waves, which could provide keys to prediction of earthquakes (Veveakis et al., 2017; Yarushina and Podladchikov, 2015; Yarushina et al., 2015).

8. Conclusions

We have conducted field measurements and numerical investigations to characterize the deep fault associated with coastal Xinzhou hot springs in Yangjiang, Guangdong Province, China. The numerical approach is useful in this case because it can accommodate heterogeneous hydraulic conductivity or permeability field, variable water density under thermal condition and complex boundary conditions. Field measurements for various physical and chemical parameters were carried out and waters were sampled for daily and annual quantifications.

Xinzhou hot springs are located near one of the large tension fault zones in the region and are thus favored for rooting deep down inside the crust. They have outflows measured from middle 60–98°C, large volumetric flow rates with a single well at about 840 m³/day and heat source estimated up to 200 °C at about 3.5 km down. The large flow rates and high temperature at the vents are believed to have resulted from deep faults.

The thermal water has flow rates of cyclic variations, which are amplified in the hot springs with deep faults at the Xinzhou geothermal field. These datasets can be useful in identifying deep faults or fault zones, which can serve as fast flow paths for high geothermal flux. However, detailed contribution or response from the deep fault has not been clearly known.

The daily and annual variations in flows might provide leads in better understanding the processes inside the crust, in term of identifying fluid behaviors such as pressure and temperature changes in response to seismic events. These changes are believed to be responses to solid earth tides. This might hold the key to further understanding conditions and processes in deep grounds.

The shortcoming of the modeling is in using the quasi 3D modeling to approximate the 3D complex system. The well-circulating geothermal flow may have been under-represented in terms of intensity; the results might have underestimated the permeability of the fault zone.

Our results show that a thermal root needs a fast flow path such as a fault zone several kilometers down into the crust to allow a significant large thermal outflow to reach out to the shallow ground. The fault zone with elevated temperature is providing a key driving force in powering the geothermal water up to shallower loci.

The fault zone is characterized in terms of hydraulic conductivity or permeability, which controls the geothermal water flow and heat transfer in the geothermal field. The model is calibrated against the temperature profile in the spring system. It was shown that the temperature distribution is quite sensitive to the permeability of the deep fault.

The deep fault property obtained from this study is conducive to further research studies in energy transfer and system response to perturbation in the deep crust. Specifically, the porosity wave propagates favorably in a high porosity zone, which is the characteristics of either tension or a tension-twist fault.

Our work represents the first of few studies with numerical experiments of the deep geothermal system. It provides a basis for further numerical studies of deep geothermal flow resulting from elevated heat conditions, including the effect of chemical species, dehydration and chemical precipitation. The results have implications for modeling of energy transfer in terms of porosity wave formation in seismic studies.

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