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THE INFLUENCE OF GROUNDWATER FLOW ON THERMAL REGIMES IN MOUNTAINOUS TERRAIN

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ABSTRACT

Active circulation of cool groundwater in mountainous terrain can cause an advective disturbance of the thermal regime. This factor complicates interpretation of data collected in geothermal exploration programs. An isothermal free-surface model has been developed which provides qualitative insight into the nature of an advective disturbance as it is affected by topography, permeability and climate. A fully coupled model of fluid and heat transfer is being developed for quantitative study of idealized mountain hydrothermal systems.

INTRODUCTION

Geothermal exploration in the Pacific Northwest is hampered by the advective disturbance of thermal regimes. Active circulation of cool groundwater in this region of steep topography and abundant precipitation can complicate interpretation of temperature gradients measured in shallow boreholes (La Fleur, 1983). Furthermore, shallow groundwater flow may suppress high-temperature surface manifestations of a geothermal resource and mask an underlying hydrothermal flow system (Duffield, 1983). In regions of low relief, and/or low permeability, advective heat transfer is minimized and high-temperature thermal phenomena such as geysers, fumaroles and hot springs are more likely to occur. Geothermal exploration in such environments is simplified due to the close association of surface phenomena with underlying free-convection cells, thermal sources and possible reservoirs.

Conventional exploration strategies often use data from thermal springs to draw inferences on the nature of the thermal regime. In mountain hydrothermal systems, the character of thermal springs is determined largely by topographically-induced groundwater flow systems which are difficult to relate to possible exploration targets. For example, at Meager Creek, in the Coast Mountains of British Columbia, a localized thermal anomaly has been identified in close proximity to a thermal spring (Lewis et. al., 1978). However, despite a significant exploration effort, the relationship between thermal springs, possible reservoirs and thermal sources has yet to be resolved.

An improved understanding of the interaction of groundwater flow and thermal regimes in mountain systems will help refine exploration strategies. To this end, we have initiated a model study of the mechanism of coupled heat and fluid transfer in mountainous terrain. A first step in our study has involved formulating a method to simulate isothermal groundwater flow in mountain systems. The isothermal model has been developed and is being incorporated in a fully-coupled model of fluid and heat transfer. In the following discussion, results of isothermal simulations demonstrate our approach to modeling groundwater flow in mountain systems and provide qualitative insights into the nature of advective disturbance of thermal regimes in mountainous terrain.

MODELING APPROACH

An idealized isothermal groundwater flow system in mountainous terrain is shown in Figure 1. In order to emphasize the relatively steep slopes, the system has been drawn without vertical exaggeration. We are interested in the bulk flow of groundwater through the mountain massif rather than the details of flow in thin surficial deposits. Thus, the top boundary is viewed as the bedrock surface with vegetation and surficial deposits removed. Impermeable vertical and basal boundaries allow recharge and discharge only across the top boundary.

In taking the bedrock surface as the top boundary of the flow system, we assume that overlying surficial deposits act as a thin high-permeability skin on the bedrock surface. Precipitation either crosses the bedrock surface as recharge, or is lost as runoff. We lump overland flow, subsurface flow through surficial deposits and evapotranspiration in a single runoff term. These processes are strongly influenced by the variable magnitude, frequency and duration of precipitation events. However, in the context of large-scale mountain systems, a steady-state approach is reasonable. We define an available infiltration rate to represent the maximum rate of recharge expected for specified climatic, geologic and topographic conditions (Figure 1). Depending upon the nature of the flow system, recharge rates may be less than this maximum. If available infiltration exceeds recharge, additional runoff is generated (Figure 1).

Groundwater recharge is controlled by hydraulic gradients near the bedrock surface. Hydraulic gradients are governed by the permeability and geometry of the overall flow system and reflect the capacity for flow through the system. Low gradients and low permeability can produce recharge rates less than available infiltration rates. High permeability and high gradients may enable a flow system to accept all the available infiltration.

Saturated and unsaturated regions of flow occur along the bedrock surface. They are separated by a water table which meets the bedrock surface at a point of detachment (POD). Note that the POD does not necessarily coincide with the boundary between recharge and discharge (shown as the hinge point in Figure 1). Where the water table coincides with the ground surface, the saturated zone is recharged directly by incident precipitation. Elsewhere, recharge at the bedrock surface is transferred to the water table by unsaturated flow. Significant lateral flow in the unsaturated zone may cause patterns of flux across the water table to differ from patterns of recharge at the bedrock surface. The nature and magnitude of lateral flow in unsaturated regions of mountain flow systems is not well known. In our modeling approach, we use the simplest assumption of one-dimensional vertical flow in the unsaturated zone. This approach implies that fluid flux rates across the water table correspond exactly to recharge rates at the bedrock surface, allowing us to neglect the unsaturated zone. Thus, the water table is the effective upper boundary of the entire flow system.

In using the water table as the upper boundary of the flow system, we require information on water table elevations in mountainous terrain. Such data are sparse and assumptions must be made to estimate realistic water table elevations. One possibility is to arbitrarily fix the water table position with reference only to surface topography. This method can result in flow systems which erroneously have recharge rates in excess of available infiltration rates. In order to maintain reasonable recharge patterns we adopt a free-surface modeling approach which does not require explicit definition of the water table. This method provides a constraint on water table estimates by incorporating available infiltration rates in the computations. For example, if recharge accepted by the system is everywhere less than available infiltration, the water table will be at the bedrock surface. If the capacity of the system to accept recharge exceeds available infiltration, the water table lies at some distance below the bedrock surface.

A vertical, planar/axisymmetric model of steady-state groundwater flow in two-dimensional mountain systems has been developed using a free-surface method similar to that of Neuman and Witherspoon (1970). Linear triangular elements and one-dimensional line elements are used. Line elements have been included to allow simulation of thin, high-permeability fault zones.

The free-surface method involves iterative solution of the groundwater flow equation using a finite element mesh which can deform in the vertical direction. For initial and subsequent iterations, a mixed boundary condition is specified on the top surface (Figure 1). Fixed heads are specified where the water table coincides with the bedrock surface. Elsewhere, an available infiltration rate is assumed. The solution proceeds by solving this initial problem and computing the corresponding head distribution. The resulting head values are used to extrapolate a new water table position and the finite element mesh is deformed to conform to the shape of the new upper boundary. At each iteration, the POD is moved along the top surface to ensure a balance of recharge and discharge. The iterative procedure is repeated until a stable water table configuration is determined.

Conventional techniques in free-surface models track the movement of the point of detachment (POD) along the bedrock surface. Groundwater recharge is assumed to occur only above the POD, which is implicitly assumed to coincide with the hinge point. In developing our modeling approach, it was observed that the POD does not necessarily coincide with the boundary between recharge and discharge (hinge point) in steep terrain with relatively low permeability. Thus, recharge can occur downslope of the POD where the water table coincides with the bedrock surface in recharge areas (Figure 1). This situation cannot be handled by conventional methods which may produce solutions with unrealistic water table configurations.

We have adopted the conventional approach of tracking the movement of the point of detachment (POD) along the top surface. However, we do not assume that the POD marks the lowest point of recharge. Rather, we identify the POD as the uppermost point where hydraulic heads are known to equal the surface elevation (Figure 1). Although recharge rates are not initially known for the region between the POD and hinge point, they can be computed following solution of the flow problem. Recharge rates are computed at each iteration using hydraulic heads in the vicinity of the bedrock surface. Computed recharge rates are compared to available infiltration rates at each iteration. If recharge rates exceed infiltration rates, additional water table adjustments are made within the iteration.

The accuracy of each solution is estimated by performing a fluid flux balance. A total balance of fluid crossing all system boundaries is computed, in addition to a balance of fluxes across only the top surface. Because vertical grid lines in the deforming mesh are rarely orthogonal to the sloping top boundary, computation of fluid flux normal to the surface using hydraulic heads can be inaccurate. After the final iteration is complete, the problem is reformulated in terms of stream functions and solved for the velocity potential distribution. This technique enables direct computation of fluxes normal to the water table. Differences between boundary fluxes obtained from the hydraulic head and stream function solutions can be used to identify regions where the finite element mesh may be too coarse. If discrepancies are significant, the mesh is refined and the problem is solved again using the new mesh. Contour plots of the stream function solutions are particularly useful in visualizing patterns of groundwater flow in mountain systems and relating them to the thermal problem.

ISOTHERMAL GROUNDWATER FLOW SYSTEMS

Isothermal fluid flow in mountainous terrain is strongly controlled by topography, climate and the distribution of rock mass permeability. The effect of these factors on water table configurations and groundwater flow patterns can be demonstrated by modeling a series of idealized systems. In this discussion we neglect the influence of thermal bouyancy on fluid flow patterns. This effect will be addressed upon completion of the fully coupled model.

Figure 2a shows the flow system and water table configuration computed for the homogeneous isotropic rectilinear system described in the previous section. A uniform available infiltration rate is arbitrarily assigned as 20 percent of a reference hydraulic conductivity K_0. In regions where recharge accepted by the flow system exceeds available infiltration, the water table lies below the bedrock surface (Figure 2a). Increased infiltration rates would produce higher water table positions with corresponding POD's at higher elevations on the bedrock surface.

Streamlines shown in Figure 2a demonstrate the pattern of flow. In nonisothermal systems, heat within a streamtube is transferred by fluid motion parallel to streamlines. If thermal dispersion is neglected, heat is transferred between streamtubes by thermal conduction. A qualitative impression of advection-dominated heat transfer can be obtained from the density of streamlines and their proximity to a thermal source. Thermal gradients are enhanced where streamlines downgradient of a heat source are closely spaced. Conversely, thermal gradients are reduced upgradient of a thermal source and in regions where streamlines are more widely spaced. As groundwater flow rates decrease, thermal conduction between streamtubes increases.

A reservoir at depth is simulated by including a region of increased hydraulic conductivity in the rectilinear system of Figure 2a. Figure 2b shows the flow system for a basal reservoir, 400 meters thick, with hydraulic conductivity 50 times that of the homogeneous isotropic system (K_0). The pattern of flow is significantly altered by the reservoir, suggesting that cool fluids might penetrate more deeply into the system. Convergence of streamlines within the reservoir implies that thermal anomalies in the mountain massif and in the discharge region might be more localized than in the previous example (Figure 2a). The increased permeability at depth causes a maximum water table decline of 200 meters and an increased net flux through the system of 20 percent.

The effect of surface topography on mountain flow systems is illustrated in Figure 3. Figure 3a represents a vertical section of a concave ridge with volcano-like form. A vertical section of a convex ridge which might be found in glaciated crystalline terrain is shown in Figure 3b. The total relief and overall geometry of these systems are the same as in the previous case. Again, the reference homogeneous, isotropic hydraulic conductivity K_0 is assumed with uniform available infiltration rates equal to 20 percent of K_0.

Water table configurations and net flux through each flow system are affected by the surface topography. The relatively flat upland of the convex system (Figure 3b) results in a more extensive recharge area than is found in the other topographic types (Figures 2a and 3a). This effect results in greatest flux and the highest water table elevation in the system with convex topography. The smallest net flux and lowest water table elevation occur in the concave system where the extent of the recharge area approximately equals that of the discharge area (Figure 3a). Convergent streamlines in the restricted discharge area of the convex system suggest that convex topographies will produce localized surface thermal anomalies of greater magnitude than those in concave topographies. The general nature of this effect has been identified by Smith and Chapman (1985) in modeling of systems with low topographic relief.

Anisotropy and heterogeneity in rock mass permeability affects both water table configurations and flow patterns. Figure 4a shows how the flow system of Figure 3b is modified by increasing the vertical hydraulic conductivity by a factor of 5 relative to the horizontal hydraulic conductivity, which is held at K_0. This anisotropic system is intended to simulate uniform sub-vertical fracturing. Figure 4b shows how the flow system of Figure 3b is modified by a vertical fault zone 100 meters wide with hydraulic conductivity 5 times that of the surrounding rock mass (K_0). Overall patterns of flow through each system are similar. However, streamline densities in the discharge regions of each system are markedly different. The vertical fault and vertical anisotropy contribute to an increased density of streamlines in discharge areas suggesting an enhanced thermal expression at the surface. Note that relatively minor geologic features, such as the 100 meter wide fault zone of Figure 4b, can significantly affect the water table configuration despite the subtle difference in hydraulic conductivity between the fault zone and the surrounding rock mass.

Regional variations in climate influence the spatial variation of precipitation and available infiltration rates. For each flow system, a threshold infiltration rate can be defined above which the water table is everywhere at the bedrock surface. The threshold rate defines the maximum recharge rate which can be accepted by a particular system and depends upon surface topography and rock mass permeability. Low permeability systems in humid climates will have threshold rates which may be less than available infiltration rates. Thus, water tables are at or near the bedrock surface (Figure 5a). The converse is true for high permeability systems in arid environments where threshold rates may exceed available infiltration rates and produce water tables well below the bedrock surface (Figure 5b). When assigning arbitrary infiltration rates or water table positions, it is important to check that computed recharge rates are within the limits expected for the specified topographic and climatic conditions. The free-surface technique helps to reduce uncertainty in this regard by employing an internally consistent method to assign water table postions and ensure that available infiltration rates are not exceeded.

Figure 5a shows the fully saturated system occurring in a convex system with infiltration rates in excess of the threshold. The threshold rate for the convex system equals 29 percent of the reference hydraulic conductivity K_0. Threshold rates for the rectilinear and concave systems are equal to 56 percent and 72 percent of K_0, respectively. Therefore, the restricted discharge area produced by the convex topography of Figure 5a results in a significantly lower threshold infiltration rate.

Figures 5b and 3b show the lower water tables and modified flow patterns found for the convex system using sub-threshold infiltration rates equal to 10 percent of K_0 and 20 percent of K_0, respectively. Note that the total flux through a fully saturated system is the maximum which can be attained even if infiltration rates exceed the threshold rate. Using the computed flux through the fully saturated convex system as a reference, dimensionless fluxes of 0.96 and 0.52 correspond to the progressively lower water tables indicated in Figures 3b and 5b. Thus, water table configurations and total flux through the flow system are significantly affected by available infiltration rates. Similar results are found for other topographic types. Arbitrary definition of water tables in mountain regions may imply infiltration rates which do not adequately reflect the climatic environment.

Model results presented here indicate that water table configurations depend strongly on the magnitude and distribution of hydraulic conductivity and available infiltration rates. Arbitrary estimation of water table elevations with reference only to surface topography cannot easily account for the influence of these factors. Using the free-surface method, we employ assumed available infiltration rates and hydraulic characteristics of the flow system to calculate plausible water table configurations. It is recognized that infiltration rates depend upon a variety of complex factors which are not easily evaluated. Thus, there is a certain subjectivity involved in estimating infiltration rates, just as there is a subjective element in defining permeability distributions in mountain-scale systems. However, estimating these parameters by iterative comparison of field data and model results is expected to provide reasonable estimates of water table configurations and flow patterns in mountainous terrain.

SUMMARY

Isothermal free-surface modeling of groundwater flow in mountainous terrain illustrates the influence of topography, rock mass permeability and climate on water table configurations and groundwater flow patterns. Contoured plots of the velocity potential provide qualitative insights into the possible nature of thermal regimes in mountain systems. This isothermal model forms the basis for a fully coupled model of fluid and heat transfer in mountainous terrain.

The fully coupled model currently in development uses this method to solve the steady-state groundwater flow equation. The isothermal model is modified to include a thermal bouyancy term in the flow equation and to account for the temperature dependence of fluid viscosity and density. Steady-state thermal regimes can be simulated by solving the equation of heat transfer. Both thermal conduction and heat transfer by fluid flow are considered. Thermal boundary conditions include a basal heat flux and insulated vertical boundaries. On the top surface, temperatures in recharge areas can be fixed at an ambient air temperature which varies as a function of elevation. In discharge areas, surface temperatures can also be fixed at air temperature or computed using an iterative process. In the iterative procedure, surface temperatures are related to ambient air temperature through a simple linear cooling law. This method uses a highly simplified representation of the complex processes of heat transfer at the surface. However, this approach should provide useful insights into the nature of thermal regimes in mountainous terrain and aid in establishing relationships between thermal springs, thermal sources and possible reservoirs.

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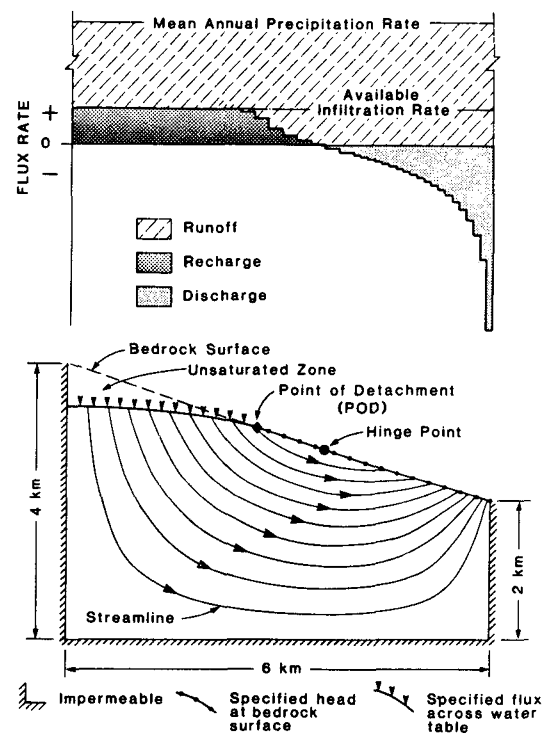

FIGURE 1. Schematic boundary value problem for isothermal groundwater flow in mountainous terrain.

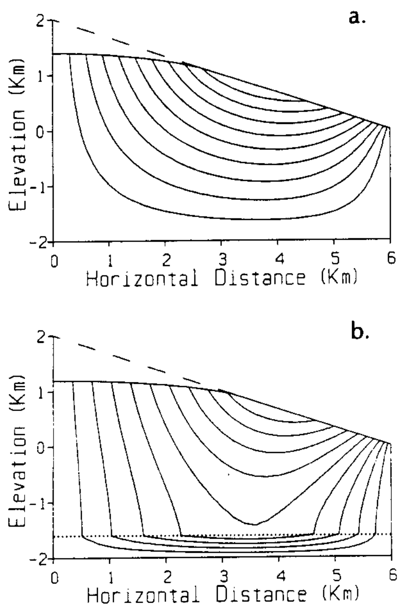

FIGURE 2. Influence of a reservoir at depth on patterns of flow in a rectlinear system. a) Homogeneous isotropic system with hydraulic conductivity K_0. b) Basal reservoir 400 meters thick with hydraulic conductivity 50 times K_0.

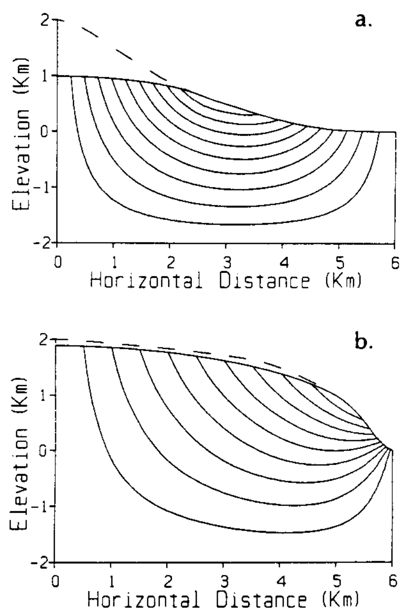

FIGURE 3. Influence of topography on patterns of flow in mountainous terrain. a) Concave volcano-form system. b) Convex glaciated system in crystalline terrain. Both systems have homogeneous isotropic conductivity K_0.

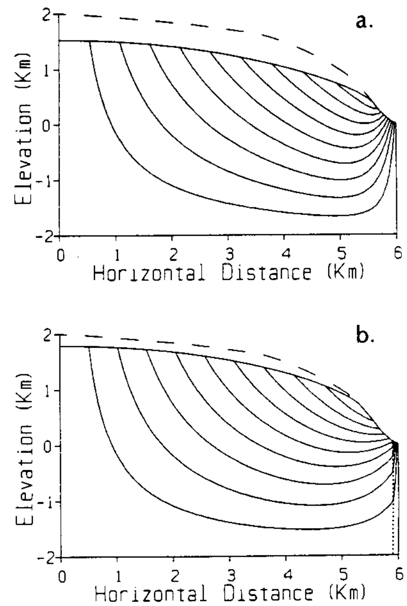

FIGURE 4. Influence of anisotropy and heterogeneity on patterns of flow in a convex system. a) Anisotropic system with vertical hydraulic conductivity 5 times K_0. b) Vertical fault zone 100 meters wide with hydraulic conductivity 5 times K_0.

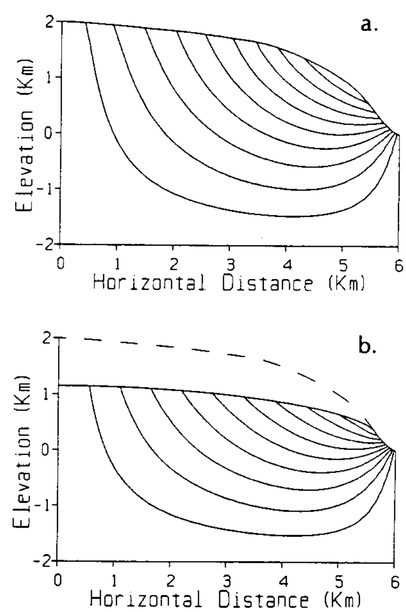

FIGURE 5. Influence of climate on patterns of flow in a convex system. a) Humid climate; infiltration rates exceed the threshold rate and the water table coincides with the ground surface. b) Arid climate; infiltration rates are less than the threshold rate and the water table falls below the bedrock surface. Both systems have homogeneous isotropic hydraulic conductivity K_0.