Advective heat transport by low-temperature discharge in the Oregon Cascades

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ABSTRACT
In the central Oregon Cascades there is a large region of near-zero near-surface heat flux due to downward and lateral flow of groundwater. Within this region, there are numerous large cold springs. Despite the low temperatures of the springs, about 3.5 °C, the heat discharged at the springs is equivalent to a mean heat flux of 56 mW/m² and represents approximately half the background heat flux. The rest of the background geothermal heat is advected laterally and discharged at lower elevations.

INTRODUCTION
The high permeability of near-surface rocks in Quaternary volcanic arcs permits high recharge rates and rapid groundwater flow. As a result, interpretations of borehole temperature data are challenging and the resulting estimates of heat flux are uncertain. The interpretation of temperature records from single boreholes is often based on analytical models (e.g., Bredehoeft and Papadopulos, 1965; Sorey, 1971; Ziagos and Blackwell, 1986). At the regional scale, numerical groundwater flow models can be employed to study the effects of groundwater circulation on the temperature distribution and heat transport in volcanic arcs (e.g., Smith and Chapman, 1983; Forster and Smith, 1989; Ingebritsen et al., 1992).

In the central Oregon Cascades, several investigators attempted to determine regional variations of the background heat flux (Blackwell et al., 1982, 1990a; Ingebritsen et al., 1992, 1994). The magnitude and spatial distribution of heat flux are relevant not only for evaluating potential geothermal resources, but also for inferring the deep structure of volcanic arcs, in particular the nature and size of heat sources (such as magma or hot intrusive rocks) at depth. The inability to quantitatively account for the effects of groundwater flow is ultimately responsible for differing views of both the hydrological and thermal system in the Oregon Cascades (Blackwell and Priest, 1996a, 1996b; Ingebritsen et al., 1996a, 1996b).

Here, I determine the amount of heat that is removed advectively by large cold springs that discharge shallow groundwater from highly permeable near-surface aquifers (Manga, 1996). I adopt a one-dimensional heat budget approach (e.g., Brot et al., 1981) that permits determination of the spatially averaged heat flux. Despite temperatures of only 3.5–6.0 °C, the cold springs discharge a substantial amount of geothermal heat and thus are a significant component of the regional heat budget.

GEOLOGICAL AND HYDROLOGICAL SETTING
I focus on a section of the central Oregon Cascades to the east of the crest of the Cascades. This region, a large part of the High Cascades, consists of Quaternary volcanic rocks (primarily basaltic andesites) overlain in some areas by glacial deposits (MacLeod and Sherrod, 1992). The Bachelor volcanic chain, located to the east of the crest of the Cascades, was emplaced between about 18 and 7 ka (Scott and Gardner, 1992), and is probably underlain by glacial, alluvial, and lacustrian deposits.

A highly schematic illustration of a cross section across the eastern Cascades is shown in Figure 1, along with a sketch of possible groundwater flow paths. Almost all the springs in this region are fed by large streams (Meinzer, 1927). The springs represent local flow and reflect topographic control of groundwater circulation. A larger scale regional flow certainly exists (Gannett et al., 1996) and is responsible for high-temperature discharge at lower elevations (Ingebritsen et al., 1989). The specific locations of the cold springs appear to be governed by permeability contrasts: Almost all the large cold springs are located near the surface...
contact of permeable basalts and less permeable sedimentary deposits.

I consider four streams for which there are more than 60 years of U.S. Geological Survey daily discharge measurements, the Cultus River, Quinn River, Fall River, and Browns Creek. The recharge area for each spring, estimated from surface topography, is shown in Figure 2. As noted in U.S. Geological Survey Water Resources Reports (e.g., Hubbard et al., 1992), drainage areas are uncertain owing to interbasin exchange. The areas shown in Figure 2 are consistent with values reported in the U.S. Geological Survey reports, with the exception of the Fall River recharge basin. I assume that the recharge area for the Fall River is bounded on the west by Bachelor, Lookout, and Sheridan Mountains and on the east by Edison, Kuamaksi, Klak, Lolo, and Wake Buttes. The rate of groundwater recharge can be estimated by dividing the discharge at the cold springs by the (estimated) recharge area (Manga, 1997).

Temperatures of the spring waters were measured at irregular intervals, typically 1 to 3 mo, over a period of 2 yr via a calibrated thermometer with a precision of 0.1 °C. Measured temperatures varied by less than 0.2 °C at each spring. For example, identical temperatures were measured in December 1996, September 1997, and May 1998. Thus, the residence time of groundwater is sufficiently long to dampen annual periodic temperature variations (Bundschuh, 1993). Table 1 lists the recharge area, groundwater recharge, and temperature for each spring.

![Figure 2. Recharge areas for Fall River, Quinn River, Browns Creek, and Cultus River. Numbers on contour lines are elevations in feet (1 ft = 0.3048 m). Black circles show locations of large cold springs.](image)

**TABLE 1. PROPERTIES OF LARGE COLD SPRINGS**

<table>
<thead>
<tr>
<th>Spring</th>
<th>Recharge Area (km²)</th>
<th>Recharge* (m/yr)</th>
<th>Temperature (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quinn River</td>
<td>33</td>
<td>0.66</td>
<td>3.5</td>
</tr>
<tr>
<td>Browns Creek</td>
<td>60</td>
<td>0.60</td>
<td>3.6</td>
</tr>
<tr>
<td>Cultus River</td>
<td>44</td>
<td>1.27</td>
<td>3.6</td>
</tr>
<tr>
<td>Fall River†</td>
<td>202</td>
<td>0.65</td>
<td>6.0</td>
</tr>
</tbody>
</table>

* From Manga (1997).
† Recharge area is twice that estimated by the U.S. Geological Survey: see text for details.

**GEOTHERMAL HEAT DISCHARGED BY COLD SPRINGS**

Consider an aquifer with thickness $h$, effective porosity $\phi$, and heat flux $Q$ entering its base. Assume that all this heat is advected horizontally by groundwater flow in the aquifer and is then discharged at the springs. This approximation is consistent with the near-zero heat flux and temperature gradients observed near the surface in the central Cascades (e.g., Blackwell et al., 1982; Ingebritsen et al., 1989).

The rate at which the mean temperature, $T$, of water in the aquifer increases is

$$\frac{dT}{dt} = \frac{Q}{\rho C h \phi},$$

where $\rho$ and $C$ are the density and the heat capacity of water, respectively, and $t$ is time. The total change in temperature, $\Delta T$, will be equation 1 integrated over the mean residence time of water in the aquifer (hereafter referred to as the "age" of the water):

$$\Delta T = \int \frac{dT}{\text{age} dt} = \text{age} \left( \frac{dT}{dt} \right).$$

From conservation of mass, we know that

$$\text{age} = \frac{h \phi}{R},$$

where $R$ is the recharge rate of the aquifer. Thus,

$$Q = \rho C R \Delta T,$$

and we are able to determine the mean heat flux $Q$ given $\Delta T$ and $R$. The right-hand side of equation 4 is the total heat discharged at the springs per unit time divided by the recharge area,

$$Q = \rho C \frac{\Delta T}{\text{recharge area}},$$

assuming that all the groundwater recharge is discharged at the springs.

In order to determine $\Delta T$ we need to know the mean recharge temperature of the water discharged at the springs. I estimate this temperature by assuming that (1) the local recharge temperature is the same as the local mean annual surface temperature, and (2) recharge rates are proportional to precipitation. The first assumption is commonly made in groundwater flow studies (e.g., Forster and Smith, 1989) and does not account for microclimatic variations and near surface hydrological processes that may affect subsurface temperatures. Even though the recharge is snowmelt, the heat capacity of the annual recharge is much less than that of the rock and soils through which this water infiltrates; at a sufficiently great depth, annual temperature variations will thus be reduced, and the temperature will approach the mean annual surface temperature (Taniguchi, 1993).

One may calculate the precipitation-weighted mean recharge elevation for the recharge areas shown in Figure 2. Precipitation is based on records...
from 1961 to 1991 compiled by G. Taylor, State climatologist, Oregon Climate Service (see inset of Fig. 1 in Manga, 1997). This calculation was done manually because the resolution of existing PRISM (parameter-elevation regressions on independent slopes model) data (e.g., Daly et al., 1994) for precipitation and temperature is far too coarse (a 4 km grid) to be useful for the present calculations. The relationship between elevation and mean annual temperature is based on data from five climate stations. For the three springs along the crest of the Cascades, I use data from Santiam Pass (1448 m, 4.29 °C) and Sisters (969 m, 7.64 °C); for the Fall River recharge area that lies further east, I use data from Bend (1116 m, 7.94 °C), Wickiup (1329 m, 6.55 °C), and Chemult (1451 m, 5.67 °C).

Figure 3 is a plot of the relationship between temperature and elevation for these two regions, and shows the relationship between the temperature of the springs, the elevation of the springs, and the mean-recharge elevation. The spring temperatures are between 1 and 2 °C colder than the mean annual temperature at the discharge elevation, whereas they are slightly warmer than the temperature at the mean-recharge elevation. ΔT in equation 4 is the temperature difference between the mean recharge elevations and the springs.

Using the results in Figure 3 and recharge values from Table 1, I obtain heat fluxes, Q, of 42–70 mW/m² along the crest of the cascades and 111 mW/m² for the Fall River recharge area (Table 2). These heat fluxes are not the total background heat fluxes, but only the heat flux that results in geothermal warming of the water discharged at the cold springs. The uncertainty in Q, between 16 and 33 mW/m², is based on an estimated uncertainty of 0.2 °C for ΔT. Because the two assumptions described above are only approximations, it is possible that the error in Q is even greater.

I have only considered large springs for which I have long gauging records because the estimate of Q requires values for discharge (in order to obtain R). The temperature of North Davis Creek (Fig. 2) is 3.5 °C. Assuming a recharge rate R similar to that of the other recharge areas implies a similar value of Q.

HEAT BUDGET

The heat flux for the three springs along the crest of the Cascades, weighted by their respective recharge areas, averages 56 mW/m², about half the mean background flux in the Western Cascades (e.g., Ingebritsen et al., 1989; Blackwell et al., 1990a, 1990b) and some other volcanic areas (e.g. Furukawa et al., 1998). By contrast, the temperature increase for the Fall River springs suggests a heat flux of about 1.1 × 10² mW/m². This difference may reflect differences in the underlying stratigraphy and geology. Almost all of the groundwater that recharges the Fall River aquifer probably emerges at the springs owing to the low permeability of sediments that underlie the young volcanic rocks of the Bachelor chain (Fig. 1). However, along the crest of the Cascades, there is a deeper regional circulation, in addition to the local flow that emerges at the cold springs: This deeper flow reduces the geothermal gradient and thus the heat flux supplied to the cold springs.

One can estimate the spatially averaged volume of water that participates in large-scale regional circulation (dashed arrows in Fig. 1) relative to the total recharge. For the deeper component of groundwater flow in the recharge area, I adopt a one-dimensional model for heat transport in the vertical direction, z, that consists of a layer of thickness L and a background flux Q_total entering the base of this layer. The heat flux out of the top of the layer, Q, is that determined by equation 4. Assume for simplicity that the downward component of groundwater velocity decreases linearly from U (a Darcy velocity) at the top of the layer, to 0 at the bottom of the layer (Phillips, 1991).

The solution to the one-dimensional advective diffusion equation

\[ \frac{U_T}{L} \frac{dT}{dz} = \kappa \frac{d^2T}{dz^2} \]

with these boundary conditions is

\[ U = \frac{2\kappa}{L} \ln \frac{Q}{Q_{total}}. \]  

Here κ is the effective thermal diffusivity, i.e., the thermal conductivity of the saturated rock divided by the density and heat capacity of water. The conditions under which equations 6 and 7 are good approximations for the temperature distribution are discussed in detail by Phillips (1991, p. 191–194). Assuming the depth of the regional flow, L, is about 1 km, U must be \( \approx 4 \) cm/yr in order to obtain \( Q/Q_{total} = 1/2 \). U is thus much less than the recharge of the surface aquifers, 60–127 cm/yr, used in the preceding section.

Ingebritsen et al. (1989, 1994) presented a heat budget for the central region of the Cascades north of the area considered here. For an 85 km length of the volcanic arc, they estimated that the region of near-zero, near-surface heat flux in young volcanic rocks (younger than 7 Ma) has a heat

![Figure 3. Relationship between temperature (T) and elevation. Solid lines—mean annual surface temperatures for Fall River (top curve), and Cultus River, Quinn River, and Browns Creek (bottom curve) recharge areas. White circles show elevations and temperatures of springs. Black circles show mean-recharge elevations and spring temperatures.](image-url)
TABLE 3. HEAT BUDGET

<table>
<thead>
<tr>
<th>Component</th>
<th>Heat flow east of arc (MW/km of arc)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heat deficit in Quaternary rocks*†</td>
<td>2.4</td>
</tr>
<tr>
<td>Heat deficit in 2 to 7 Ma rocks*</td>
<td>0.61</td>
</tr>
<tr>
<td>Anomalously heat discharge in &gt;7 Ma rocks*</td>
<td>1.5</td>
</tr>
<tr>
<td>Heat discharged by cold springs§</td>
<td>0.61 ± 0.24</td>
</tr>
</tbody>
</table>

* After Ingebritsen et al. (1989, 1994).
† Heat deficit on both sides of the Cascades divided by two.
§ Based on the 12.5 km of arc represented by the discharge in the Quinn River, Cultus River, and Browns Creek.

deficit of 470 MW. The excess advective and conductive heat loss in the older rocks on the sides of the Cascades is 360 MW and represents heat carried by circulating ground water from the region overlain by younger volcanic rocks. The difference between the heat deficit in the younger rocks and excess heat in older rocks “may occur as lower temperature advective discharge” (Ingebritsen et al., 1989, p. 1461). The heat fluxes determined here quantify this lower temperature groundwater flow.

Table 3 lists the various components of the heat budget, expressed as heat flow per kilometer of arc, so that these results can be compared with those of Ingebritsen et al. (1989, 1994). A direct comparison of the relative magnitude of the various components of the heat budget is not straightforward because I am considering a different segment of the Cascade arc, and I do not account for along-arc and spatial variations of groundwater flow. Nevertheless, my estimates indicate the importance of low-temperature advective discharge in the heat budget of volcanic arcs.

CONCLUDING REMARKS

The one-dimensional model considered here does not describe the spatial distribution of temperature and heat flux. Nevertheless, it is useful for inferring spatially averaged and integrated quantities and should be independent of the detailed pattern and rate of groundwater flow. Although the temperatures of the cold springs are low, they typically discharge about half of the geothermal heat flux from their respective recharge areas. These results highlight the potential importance of accounting for low-temperature discharge in heat budgets.

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REFERENCES CITED


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