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USING SPRINGS TO STUDY GROUNDWATER FLOW AND ACTIVE GEOLOGIC PROCESSES

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■ **Abstract** Spring water provides a unique opportunity to study a range of subsurface processes in regions with few boreholes or wells. However, because springs integrate the signal of geological and hydrological processes over large spatial areas and long periods of time, they are an indirect source of information. This review illustrates a variety of techniques and approaches that are used to interpret measurements of isotopic tracers, water chemistry, discharge, and temperature. As an example, a set of springs in the Oregon Cascades is considered. By using tracers, temperature, and discharge measurements, it is possible to determine the mean-residence time of water, infer the spatial pattern and extent of groundwater flow, estimate basin-scale hydraulic properties, calculate the regional heat flow, and quantify the rate of magmatic intrusion beneath the volcanic arc.

INTRODUCTION

How fast is water moving? How much water is flowing? Where did the water originate and where is it going? These fundamental questions address both the temporal and spatial aspects of groundwater flow and water supply issues in hydrologic systems. As groundwater flows through an aquifer, its composition and temperature may change depending on the aquifer through which it flows. Thus, hydrologic investigations can also provide information about the subsurface geology of a region. But because such studies investigate processes that occur below the Earth's surface, obtaining the information necessary to answer these questions is not always easy. Springs, which discharge groundwater directly, afford a unique opportunity to study subsurface hydrogeological processes.

Drilling wells (often at great cost and difficulty) is the most common method used to obtain information about the subsurface. Because a well samples a small

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fraction of the region of interest, however, it is sometimes not feasible to make enough measurements, in both space and time, to adequately constrain geologic or hydrologic models. Spring water is therefore particularly useful in regions with few or no wells. However, because springs focus groundwater, the composition, temperature, and discharge variations of the spring water reflect processes that occur over the entire subsurface and history of transport. Thus, by using springs to understand the subsurface, rather than a set of point measurements, we need to interpret integrated information.

Interpreting both integrated and point measurements can be challenging because natural geological materials and settings are often complicated and heterogeneous. If characterizing the geological complexity of an aquifer system is not straightforward, developing a suitable mathematical representation of flow is not likely to be straightforward. Indeed, "heterogeneity has been one of the major areas of concern in ground water hydrology for the last 60 years" (Wood 2000). Thus, even though the questions listed at the beginning of this introduction may appear to be clearly defined, they involve stochastic issues (which, despite their importance, are not the focus of this review) as a result of the scale-dependence and the complexity of hydrogeologic properties and variables.

Rather than summarizing all the work done on springs and associated groundwater systems, in this review I outline a variety of measurement-based approaches that are often used to understand subsurface hydrological processes in spring systems. As a preliminary exercise, it is instructive to consider what can be learned from straightforward mass balance estimates. Consider an aquifer with surface area A, mean thickness $\overline{H}$, porosity ϕ, and mean discharge $\overline{q}$ (Figure 1). The mean residence time of water, hereafter called the mean age T_{age} is given by

$$T_{\text{age}} = A\overline{H}\phi/\overline{q} = \overline{H}\phi/\overline{N}, \tag{1}$$

where $\overline{N} = \overline{q}/A$ is the mean recharge rate. Equation 1 provides a connection between the mean residence time of water (T_{age} and the question "How fast is the water moving?"), the volume of circulating water ($\overline{q}$ and the question "How much

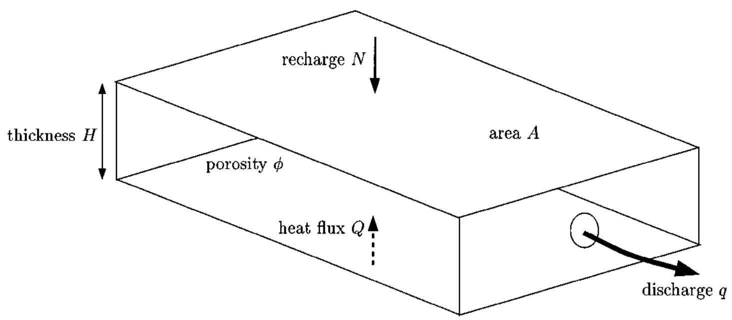

Figure 1 Schematic illustration of an aquifer and a spring.

water is flowing?"), and the source of the water (related to *A* and the question "Where did the water originate and where is it going?").

To illustrate some approaches that are used to interpret averaged information, I consider, as an example, a set of large volume springs that issue from basalt flows in the Oregon Cascades. Because these springs are located in an active volcanic arc, their waters provide an opportunity to learn about subsurface magmatic and geothermal processes. Photographs of two of the springs are shown in Figure 2, a map of their locations is shown in Figure 3, and some key

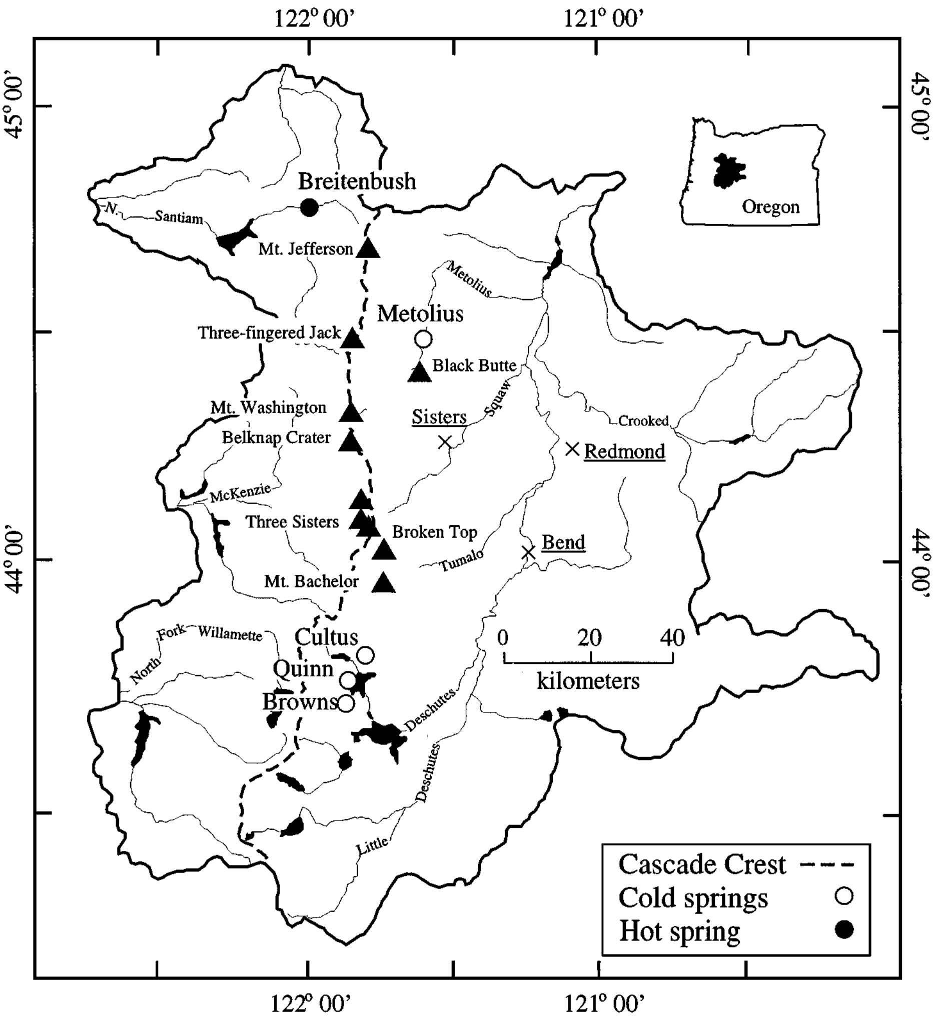

Figure 3 Location of springs discussed in this review.

measurements are listed in Table 1. These springs range from cold (a few °C) to hot (approaching the boiling point of water), and from small (liters per second) to large (several m^3/s). A more thorough discussion of the hydrogeologic setting of these particular springs can be found in Meinzer (1927), Ingebritsen et al (1992), and James et al (1999).

ORIGIN OF SPRINGS

A spring is a location where groundwater emerges from the Earth's subsurface in an amount large enough to form something resembling a stream. Natural springs may also discharge directly into lakes and oceans (Church 1996).

Rivers flowing out of the Earth have captured the imagination of scientists and philosophers for millenia and inspired some of the earliest ideas about the hydrologic cycle. Until the seventeenth century, the most widely held views in the Western world were that, given the large amount of water that can emerge at springs, springs must discharge water that "condenses" below the surface or water that is ultimately derived from the ocean and somehow flows uphill. Following watershed mass balance measurements and calculations done in the late 1600s, it became apparent that precipitation can supply more than enough water for rivers and springs. Adams (1938, Chapter XII) discusses the historical development of ideas on the formation of springs and related hydrogeological processes.

For many people, springs are the most obvious, interesting, and practical manifestation of the occurrence of groundwater. Springs have played a geographic role in human settlement, especially in arid environments. Spring waters, in particular those from mineral and hot springs, have long been purported to have medicinal or therapeutic value (Crook 1899, Waring 1915). Spring water is also associated

TABLE 1 Oregon springs discussed as examples

Name	Elevation (m)	Estimated Recharge Area (km^2)	Discharge (m^3/s)	Temperature °C
Quinn River (QR)	1354	33[a]	0.67[b]	3.4[c]
Brown's Creek (BC)	1332	60[a]	1.1[b]	3.8[c]
Cultus River (CR)	1356	44[a]	1.8[b]	3.4[c]
Metolius River (MH)	920	400[c]	3.1[d]	8.2[c]
Breitenbush Hot Spring[e]	682	—	0.013	8.4

[a]From Manga 1997.

[b]Average between 1939 and 1989.

[c]From James et al 2000. Temperatures change by less than 0.2°C over a period of several years.

[d]From Meinzer 1927.

[e]From Ingebritsen et al 1992.

in the public's mind with exceptional purity—witness the rapid growth in sales of bottled water in the United States, from 1.8% to 6.9% of bottled beverage sales between 1981 and 1997; in 1998 bottled water sales were $4.3 billion in the United States alone (all data from www.bottledwaterweb.com).

The existence of a spring requires that the subsurface is unable to transmit water as fast as it is supplied so that the potentiometric surface intersects the land surface. A range of geological structures and topographic features can thus bring water to the surface (Figure 4); Bryan (1919) provides a more comprehensive discussion. The discharge of large amounts of ground water requires some combination of a large recharge area, a high recharge rate, and a high permeability for large volumes of water to be concentrated at a single point. The presence of a large discrete spring, rather than diffuse seepage, is further evidence of heterogeneity of permeability in the subsurface.

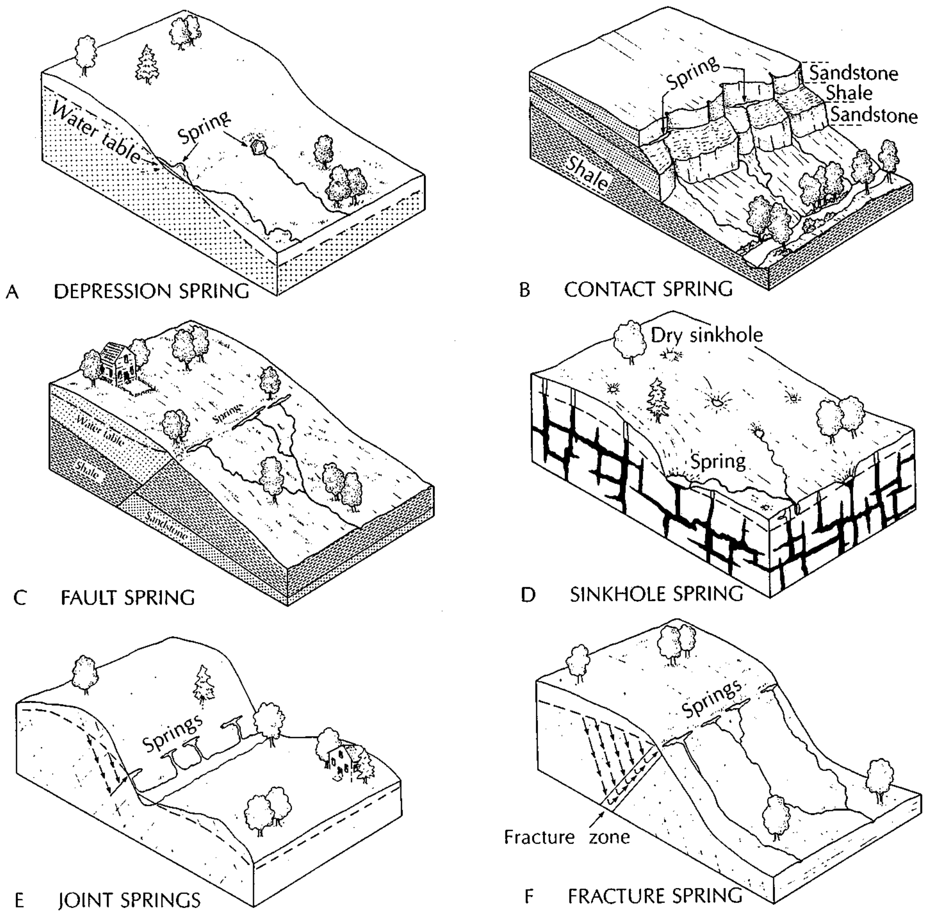

Figure 4 Formation of springs. From Fetter 1994, with permission.

ISOTOPE TRACERS

Isotopic tracer measurements and interpretations are widely used to determine how "old" groundwater is, and from where it came. Radiogenic isotopes (and tracers whose concentrations change in time) are particularly useful for estimating groundwater ages; stable isotopes are more useful for determining the original source of spring water. The study of a set of isotope tracers can thus provide information about groundwater sources and transport rates. This can be particularly valuable for the sometimes complicated geological settings of springs (for example, Nativ et al 1999, Davisson et al 1999). An analysis of tracer data in spring water, however, is complicated by the fact that recharge usually doesn't occur at a single location, but instead occurs over a large region, often over the entire length of the aquifer. The "age" of the water is thus a mean residence time for all the water that emerges at the spring. The unavoidable interaction of water and subsurface geology provides additional complications (Rose et al 1996), but also provides an opportunity to learn about subsurface geological processes (James et al 1999).

In this section I discuss some of the more commonly studied isotopes and focus on the key ideas involved in interpreting isotopic measurements. Fritz & Fontes (1980) and Clark & Fritz (1997) provide a more comprehensive summary and discussion of the use of isotopes in hydrogeologic studies, including a brief overview of recommended sampling and measurement techniques. The sampling and analytical methodology used for measurements discussed here are described in James et al (1999) and James (1999).

Besides the isotopic tracers discussed next, a variety of other tracers have been used to study spring flow, including alkalinity (Chandler & Bisogni 1999), suspended sediment (Mahler & Lynch 1999), temperature (Bundschuh 1993), and dye (Smart 1988).

How Old Is the Water?

The ideal tracer for estimating ground water ages changes its recharge concentration c_{in} over time, i.e., $c_{\text{in}} = c_{\text{in}}(t)$, and/or the tracer decays at a known rate λ ($= \ln 2/t_{1/2}$, where $t_{1/2}$ is the half-life). Ideally the tracer is also otherwise conservative and nonreactive so that chemical reactions and other sources or sinks of the tracer can be neglected. In the absence of mixing or dispersion the measured tracer concentration c_{out} can be used to obtain a water age by

$$T_{\text{age}} = \frac{t_{1/2}}{\ln 2} \ln\left(\frac{c_{\text{out}}}{c_{\text{in}}}\right). \qquad (2)$$

Here I focus on interpreting measurements of the two most widely used radiogenic tracers for estimating ages in hydrological problems: ^{3}H ($t_{1/2} = 12.3$ years) and its decay product ^{3}He, and ^{14}C ($t_{1/2} = 5730$ years). ^{3}H is thus useful for "young" waters (recharged within the last few decades) and ^{14}C is useful for "old" waters, $O\,(10^2{-}10^4)$ years. As discussed later, determining ages from ^{14}C measurements of dissolved inorganic carbon is notoriously difficult except in very simple systems, and many models and approaches have been developed to interpret measurements. Given the mass balance constraints of Equation 1, most large springs should discharge young water, though there are notable exceptions in arid climates where recharge rates are low and length scales are large (e.g. Winograd & Pearson 1976).

Other tracers that have been used for young waters include the anthropogenic radioisotope ^{85}Kr (Ekwurzel et al 1994) and anthropogenic chemicals such as chlorofluorocarbons introduced in the atmosphere (Busenberg & Plummer 1992). Interpretation of the measurements of these gases requires some care: The lag time for their transport through a thick unsaturated zone (>10 m) can lead to age overestimates as large as 10 years (Cook & Solomon 1995). ^{36}Cl ($t_{1/2} = 301{,}000$ years) has been used to study "ancient" groundwaters, such as those found in the Great Artesian Basin in Australia (Torgersen et al 1991). Dating "middle-aged" waters, with ages in the range of several decades to a few hundred years, is more challenging. The cosmogenic isotope ^{39}Ar has a suitable half-life of 269 years and can be applied to hydrogeological problems (Loosli 1983). However, sampling and measurement is not routine (Clark & Fritz 1997), and in situ subsurface production of ^{39}Ar must be accounted for (Andrews et al 1989). Solomon et al (1996) suggest that in some cases radiogenic ^{4}He, which is often used to trace ancient water (Heaton 1984), might be useful for ages in the range 10^1–10^3 years due to its diffusive loss from aquifer rocks.

Mixing Models

Owing to the mixing and dispersion of water recharged at different times and locations, models are often used to interpret tracer measurements. Integral-balance or lumped-parameter models representing the relationship between input and output can be described by a convolution integral,

$$c_{\text{out}}(t) = \int_{-\infty}^{t} c_{\text{in}}(\tau) g(t-\tau) e^{-\lambda(t-\tau)} d\tau. \tag{3}$$

Here, $g(t)$ is the impulse response function (or weighting function), c_{in} and c_{out} are the input and output concentrations, respectively, t is time, τ is an integration variable, and λ is the radioactive decay constant. Particular models are defined in terms of their response functions, $g(t)$.

Figure 5 illustrates two examples of mixing models. The simplest model is the so-called piston flow model in which water travels without mixing or dispersion. In this case the response function is a delta function, i.e. $g(t) = \delta(t - T_{\text{age}})$. The exponential model (often termed the well-mixed model) introduced by Eriksson (1958) to interpret tritium measurements, assumes an exponential distribution of transit times, $g(t) = T_{\text{age}}^{-1} \exp(-t/T_{\text{age}})$. The exponential model is typically used for distributed sources (Duffy & Lee 1992) and should be a good approximation for most springs (Dincer & Davis 1984). Both models shown in Figure 5 are characterized by a single parameter, the mean groundwater age T_{age}. It is possible

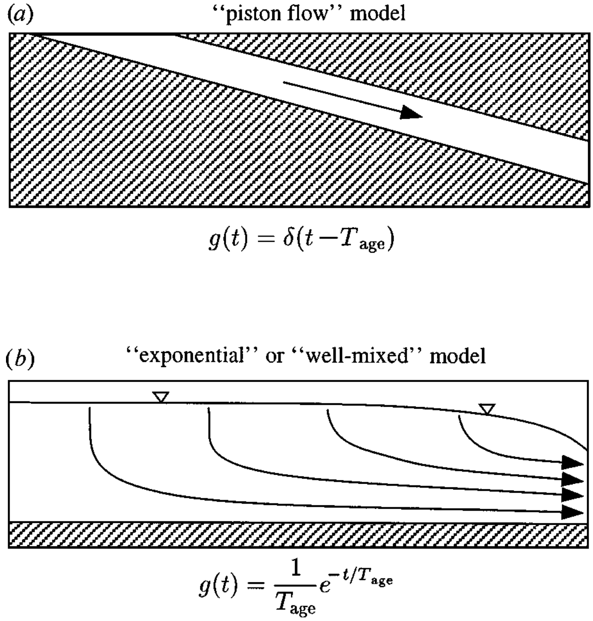

Figure 5 Schematic illustration of two commonly applied mixing models: (*a*) the "piston flow" model and (*b*) the "exponential" model. $g(t)$ is the weighting function for each model, and T_{age} is the mean residence time of water defined by Equation 1.

to develop more sophisticated models that will be characterized by additional parameters. For example, if dispersion is accounted for in the piston-flow model then $g(t)$ will depend on the ratio of flow velocity to dispersion rate, which is characterized by a Peclet number (Maloszewski & Zuber 1982). Mixing models can also account for geometric and transport properties in the system of interest. For example, in a dual porosity system, such as fracture-dominated flow through a porous matrix, apparent tracer ages will be affected by diffusive transfer of the tracer between the fractures and matrix (Sudicky & Frind 1982, Sanford 1997).

Given a single measurement of a tracer the inferred T_{age} is thus a model-dependent age. It is possible, in principle, to distinguish between mixing models in two ways. First, because $c_{out}(t)$ is a function of $g(t)$, time series of output-concentration measurements can be used to determine $g(t)$ or to distinguish between models (Amin & Campana 1996). It is then possible to estimate model parameters such as volume, travel time, and dispersivity (Duffy & Gelhar 1986). Second, because different tracers may have different diffusivities, the analysis of multiple tracers may be helpful (Maloszewski & Zuber 1990).

^{3}H and ^{3}He/^{3}H Ages In principle, the measurement of both ^{3}H and its decay product, ^{3}He, allows a "true" mean age (referred to hereafter as the ^{3}He/^{3}H age)

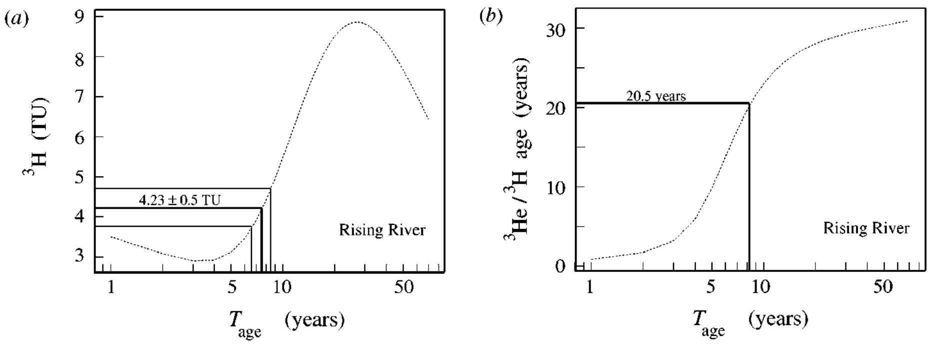

Figure 6 (*a*) ^{3}H concentration in spring water (measured in tritium units, TU), and (*b*) ^{3}He/^{3}H age as a function of groundwater age, T_{age}, for the exponential model (see Figure 5). The measured ^{3}H concentration at the Rising River springs is 4.23 ± 0.5 TU (Rose et al 1995), which implies a mean groundwater age of about 7–9 years. The measured ^{3}He/^{3}H age is 20.5 years (Rose et al 1995), which implies a groundwater age of about 8 years. From Manga 1999, with permission.

to be obtained without relying on the uncertain ^{3}H input function, i.e. the detailed values of $c_{in}(t)$ (Torgersen et al 1979):

$$^3\text{He}/^3\text{H age} = \frac{12.43}{\ln 2} \ln\left(1 + {}^3\text{He}_{\text{trit}}/^3\text{H}\right), \tag{4}$$

where the subscript trit indicates He produced by the decay of ^{3}H. Again, mixing will affect the ^{3}He/^{3}H age, and a model of the form of Equation 3 still needs to be applied for an aquifer with distributed recharge and measurements made at the spring. For example, in Figure 6, I show estimates of T_{age} for Rising River in northern California based on isotopic measurements reported by Rose et al (1995) and the exponential model. Both estimates are consistent with $T_{age} \approx$ 8 years. Notice in Figure 6*b* that for T_{age} less than about 10 years, the ^{3}He/^{3}H age overestimates T_{age}, whereas for T_{age} greater than about 30 years, the ^{3}He/^{3}H age will remain close to about 30 years. In both cases, the contribution of ^{3}He and ^{3}H from $\approx$30 year old water associated with the largest atmospheric nuclear testing peaks biases the age inferred from discharged ^{3}He and ^{3}H concentrations (Aeschbach-Hertig et al 1998, Manga 1999).

The analysis of He can be more complicated but also more geologically informative. He in groundwater can be derived from reservoirs in the atmosphere, the mantle, and the crust. Fortunately, the ratio of ^{3}He/^{4}He in each reservoir differs by up to several orders of magnitude so that in principle the contribution from each reservoir can be identified (Stute et al 1992). The contribution of atmospherically derived He can be determined by measuring the concentration of Ne and assuming an appropriate recharge temperature (Schlosser et al 1989). Three of the large-volume, cold springs listed in Table 1, the Quinn River, Brown's Creek, and Cultus River, contain only atmospherically derived ^{4}He (James et al 2000). Thus, excess ^{3}He is produced by the decay of ^{3}H and for these three springs ^{3}He/^{3}H ages are 2.1, 0.7, and 2.5 years, respectively (James et al 2000). A short residence time, for example, 2 years, combined with an estimated recharge rate of 1 m/year (Manga 1997) and a porosity of 10%, implies a mean aquifer thickness of 20 m (from Equation 1). The Metolius River, on the other hand, has nonatmospheric ^{3}He and ^{4}He requiring a contribution from a deep (mantle) source and, presumably, a longer residence time and a deeper flowpath. Magmatically derived He is often found in springs in regions that are, or have recently been, volcanically active (Kennedy et al 1985) and permits an estimate of magma degassing rates (Sorey et al 1998).

^{14}C Dating Figure 7 shows the relationship between ^{14}C and δ^{13}C for the springs listed in Table 1 (data from James et al 1999). The three end-members shown in Figure 7 represent dissolved inorganic carbon (DIC) equilibrated with soil CO_2, DIC equilibrated with the atmosphere, and typical magmatic CO_2 in a volcanic

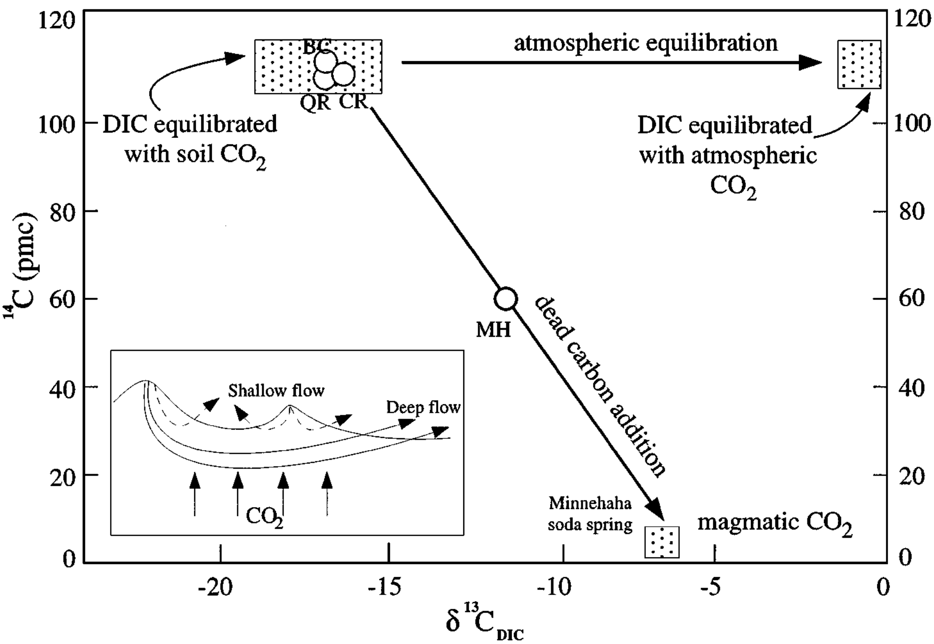

Figure 7 ^{14}C vs δ^{13}C for the 4 cold springs shown in Figure 3. Stippled boxes represent three endmember isotopic compositions of dissolved inorganic carbon (DIC) for pH between 6.5 and 8.5: DIC equilibrated with the atmosphere, DIC equilibrated with soil CO_2, and "dead" carbon of magmatic origin. The endmember for the latter is represented by Minnehaha soda spring in the southern Oregon Cascades (Rose & Davisson 1996). This spring is saturated in CO_2 (and thus bubbling, hence the name "soda" spring). The inset illustrates schematically why some springs obtain, or remain free of, magmatic carbon.

arc, represented by the isotopic composition of Minnehaha soda spring (Rose & Davisson 1996, James et al 1999). In the Oregon Cascades, carbonate minerals are generally absent, at least at shallow depths, so that sources of DIC other than those shown in Figure 7 can be neglected. Present day ^{14}C concentrations are greater than 100% modern carbon (pmc) due to ^{14}C produced by nuclear weapons tests. One interpretation of the low ^{14}C concentration in the Metolius River is that, rather than representing aging of the water (a few thousand years), the ^{14}C concentration reflects the addition of magmatically derived carbon ($^{14}C \approx 0$). An old age is precluded by Equation 1 because $\bar{q}$ is large. Measurements such as those shown in Figure 7 have been used in other regions to study magmatic CO_2 degassing (Allard et al 1997, Rose & Davisson 1996, Sorey et al 1998), though it is notable that there has not been any surface volcanic activity in central Oregon for more than 1300 years. James et al (1999) show that the rate of magmatic intrusion beneath the Cascades arc required to supply the magmatic CO_2 flux discharged at springs is consistent with the rate of intrusion inferred from heat flow studies (Ingebritsen et al 1989, Blackwell et al 1990).

The other three cold springs from Table 1 have no detectable magmatic C, consistent with the absence of magmatic He (see above section). The inset of Figure 7 illustrates schematically the different subsurface flow paths that the spring water may follow. Quinn River, Cultus River, and Brown's Creek discharge a shallow local-scale flow that does not dissolve magmatic CO_2, whereas the Metolius River discharges a more deeply circulating regional-scale flow. In the following section I demonstrate that these flow patterns are consistent with the temperature measurements made at the springs.

As illustrated in Figure 7, the most significant complication for ^{14}C dating is the addition of "dead" carbon, from a magmatic or metamorphic source as discussed here, or more commonly, from carbonate dissolution. Dead carbon added to the flow system biases inferred ages to older ages. There are a variety of approaches to account for this dead carbon addition. These range from isotopic and chemical mass balance calculations, e.g. using the NETPATH geochemical model (Plummer et al 1994), to measurement-based approaches. One example of the latter is to correct the "initial" ^{14}C activity for dissolution in the unsaturated zone based on ^{14}C measurements made in the recharge area (Bajjali et al 1997). Another example, similar to that illustrated in Figure 7, is to use $\delta^{13}C$ measurements to trace carbonate dissolution (Wigley et al 1978). Mook (1980) and Fontes (1992) provide reviews of ^{14}C dating of groundwaters.

Where Did the Water Originate?

One isotopic approach to answering this question is to examine the oxygen and hydrogen isotopes in the spring water—see Winograd & Friedman (1972), Christodoulou et al (1993), and Scholl et al (1996) for studies in a variety of geological and climatological settings. The usefulness of these isotopes stems from the fact that they fractionate (relative abundances change) in a predictable manner as water moves through the hydrologic cycle depending on the physical and chemical processes that operate (Criss 1999).

In mountainous regions, the isotopic composition of precipitation often varies systematically and in a quantifiable way with elevation. Analysis of snow core and small spring data in the central Oregon Cascades, for example, shows that δ^{18}O decreases by 0.18 per ml per 100 m rise in elevation (James 1999). This value is similar to those found farther north in British Columbia (−0.25 permil/100 m; Clark et al 1982) and farther south in Northern California (−0.23 permil/100 m; Rose et al 1996). The slightly lower rate of decrease in central Oregon may be caused by being in a rain shadow. The comparison of precipitation and spring discharge is appropriate here for two reasons. First, given the young ages found in the above section, discharge should be comparable to modern precipitation in terms of its O isotopic composition. Second, given the large volume of circulating water, there should be little isotopic effect from the possible addition of magmatic water (Giggenbach 1992). Nor should there be a "δ^{18} O-shift" produced by water-rock interactions, as is common in hot spring waters (Craig 1963). The isotopic composition of the spring water can thus be projected to the elevation at which precipitation is comparable, as shown in Figure 8, in order to infer a mean recharge elevation. From a topographic map and mass balance considerations it is then possible to estimate the recharge area for the springs.

If the mean residence time of water is sufficiently short, or if there is a component of rapid groundwater flow, there may be temporal variations in δ^{18} O at the spring. The isotopic composition of discharge and precipitation can then be used to perform

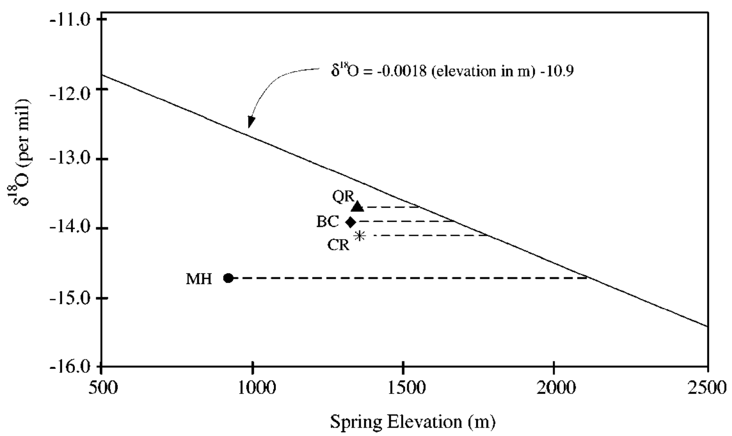

Figure 8 Elevation versus δ^{18}O in the central Oregon Cascades; line is a best-fit to data from snow cores and small springs (after James 1999), and symbols are data from large cold springs. The mean recharge elevation can be inferred by determining the elevation at which precipitation has a comparable isotopic composition. BC, Brown's Creek; CR, Cultus River; MH, Metolius River; QR, Quinn River.

a hydrograph separation to identify the volumes and rates associated with fast and slow flow paths (Lakey & Krothe 1996). For the springs studied here δ^{18} O is approximately constant over a period of several years (James 1999).

On the other hand, if the mean residence time of water is long and there is limited mixing of different water masses, the composition of the water reflects the conditions at the time of recharge. Thus, paleotemperature or atmospheric conditions (e.g. prevailing wind directions) can be inferred from δ^{18} O and δ D (Claassen 1986, Weyhenmeyer et al 2000).

TEMPERATURE MEASUREMENTS

Temperature is by far the easiest and least expensive property to measure at a spring. The temperature of groundwaters was first systematically studied in the 1700s (Davis 1999). These early studies demonstrated the key idea of the following discussion, namely that warming of groundwater reflects primarily the input of heat from within the Earth.

The use of temperature measurements to quantify the motion of groundwater has several advantages. Once a borehole has been drilled, it is inexpensive and straightforward to measure temperature accurately and obtain a high spatial resolution of measurements (Woodbury & Smith 1988). The temperature distribution satisfies an advective-diffusion equation in which the dispersivity tensor can be replaced (in most cases) by the thermal diffusivity of saturated rock because heat conduction through the rock dominates the macroscopic dispersion in interstices. Thus, variations in temperature can be related to the pattern and rate of groundwater flow rather than to the dispersive properties of the flow and porous material. The heterogeneity of porous materials that is so troublesome for problems of tracer and geochemical transport should therefore be less so for heat transport.

A wide range of solutions for the heat transport problem can be obtained for simple model geometries and compared with temperature records from single boreholes (Bredehoeft & Papadopulos 1965, Ziagos & Blackwell 1986, Ge 1998). At the regional scale, numerical groundwater flow models can be employed to study the relationship between groundwater circulation and the temperature distribution (Smith & Chapman 1983). Temperature measurements are only useful, however, when they deviate from the conductive solution. It is worth noting that temporal changes in surface temperature, uplift and erosion, topographic relief, and variations of thermal conductivity all produce nonconstant temperature gradients. These effects must be separated from those associated with groundwater flow. In many settings, a darcy velocity $>O(1)$ cm/year is sufficient to distort the conductive geothermal gradient. In the Oregon Cascades, recharge rates are $\approx$ 1 m/year and indeed borehole temperature measurements indicate that near-surface temperature gradients and the surface heat flux are approximately zero in the High Cascades (Ingebritsen et al 1989). In this case, it is possible to apply straightforward energy balance ideas to the entire groundwater system (Brott et al

1981), at least when spring temperatures are constant as they are here (Manga 1998).

Consider an aquifer with mean heat flux $\bar{Q}$ entering its base (see Figure 1). Assume that all this heat $H = \bar{Q}A$ is advected horizontally by groundwater flow in the aquifer and is then discharged at the springs. The total heat discharged at the spring is related to the temperature change $\Delta\theta$ by

$$H = \rho C \bar{q} \Delta\theta, \tag{5}$$

where ρ and C are the density and heat capacity of water, respectively, and again $\bar{q}$ is the mean spring discharge. In Northern California, there is a set of large springs that discharge about 40 m^3/s into the Fall River (Meinzer 1927). Spring temperatures are about 12°C, whereas small springs in the recharge area (on the flanks of the Medicine Lake shield volcano) are about 7°C (Davisson & Rose 1997). Thus, the "cold" Fall River springs discharge about 10^3 MW of geothermal heat, a value equivalent to a mean heat flux of about 0.56 W/m^2 (about 10 times the mean continental heat flux) over an estimated 2×10^3 km^2 recharge area. For comparison, the total amount of heat discharged by the large and very active hydrothermal system at Yellowstone is about 5×10^3 MW (Fournier 1989), and the total worldwide geothermal power being exploited in 1995 was 8.7×10^3 MW (Freeston 1996).

For the Metolius River, the estimated recharge area is about 400 km^2 (Table 1), and the background heat flux is about 120 mW/m^2 (Blackwell et al 1990). Assuming that all heat is removed advectively, I expect a temperature increase of about 4°C based on Equation 5. This value is consistent with Figure 9 if I make the reasonable assumption that the mean recharge temperature is close to the mean annual surface temperature (Taniguchi 1993, Perez 1997).

In contrast to the Metolius River, the other large springs shown in Figure 9 discharge water at temperatures that are nearly equal to the mean annual surface temperatures at the mean recharge elevations found in Figure 8. These springs thus discharge water that is colder (by about 1–3°C) than the mean surface temperature at the discharge elevation, an observation first noticed and explained by Humboldt in 1844 (see Davis 1999). The lack of geothermal warming combined with the near-zero near-surface heat flux implies that there is a larger-scale and more deeply circulating groundwater flow that advectively removes the geothermal heat. In the Oregon Cascades, much of this heat is discharged at lower elevations at hot springs (Ingebritsen et al 1989). Manga (1998) shows how the recharge rates of these two different systems (deep vs shallow) can be estimated from energy balance considerations.

It is sometimes assumed that the depth of groundwater circulation can be deduced by dividing the increase of the water's temperature by the geothermal gradient. Such a calculation implicitly makes two assumptions. First, groundwater flow does not affect the subsurface temperature distribution and thus flow rates must be low. Second, the spring water must rise sufficiently rapidly that it does not cool. Sanz & Yélamos (1998), for example, applied this argument to springs in Spain and inferred a reasonable circulation depth of 900 m. However, the total heat of

46 MW discharged by their thermal spring implies an unrealistically large geothermal heat flux over the 100 km^2 basin estimated by Sanz & Yélamos (1998), supporting the contention of other authors (Sánchez Navarro et al 2000) that the thermal water has a different origin. Thus, simple energy balance arguments can be useful for testing the reasonableness of conceptual hydrogeological models.

In summary, the temperature of spring water reflects the interaction between the advective and conductive transport of heat. The numerical calculations of Forster & Smith (1989) offer a particularly clear illustration of this relationship. In rocks with low permeabilities, velocities are low, conductive heat transport is dominant, and spring temperatures will be low. In rocks with high permeabilities, velocities will be high and advective transport of heat will dominate. High permeability also results in large volumes of circulating water, and spring temperatures will again remain low (as shown in Figure 9). The warmest springs occur for an intermediate range of permeabilities. Thus, spring temperatures will reflect a balance between

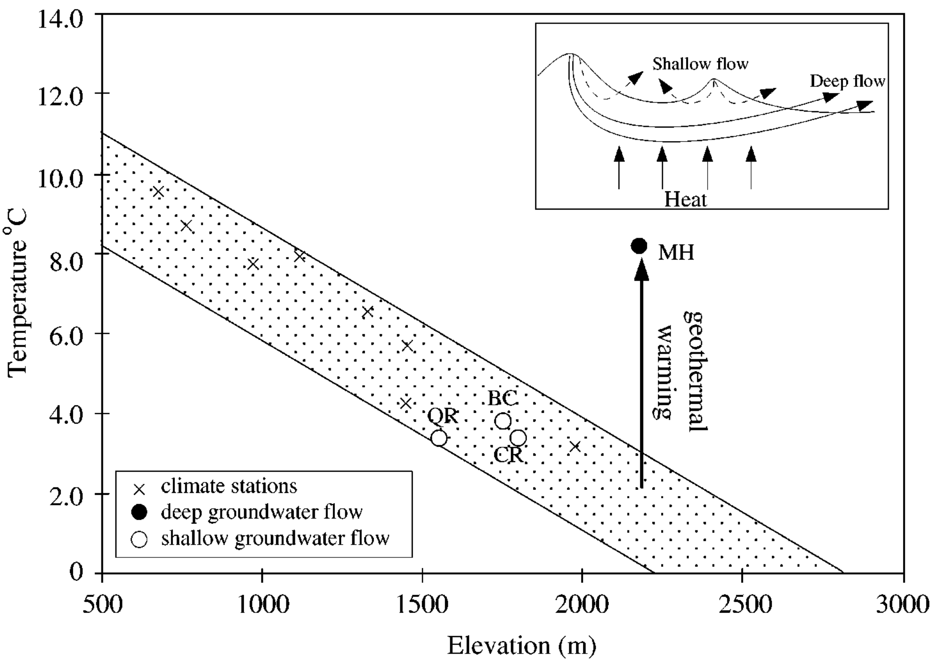

Figure 9 Relationship between elevation and temperature for climate stations in the Oregon Cascades (crosses). Spring temperature is shown as a function of the inferred mean recharge elevation (from Figure 8). The Metolius River water is warmed substantially by geothermal heat, whereas the other spring waters show little evidence of geothermal warming. The inset illustrates how deeply circulating groundwater can aquire geothermal heat, whereas water that remains at shallow depths will remain at temperatures close to the mean annual surface temperature. These inferences are consistent with those based on carbon isotopes (Figure 7).

the amount of heat transported advectively and the volume of water that must be warmed.

DISCHARGE MEASUREMENTS

Temporal variations in recharge must be reflected in changes in spring discharge. Figure 10 shows a one-year hydrograph for the Quinn River spring and nearby Deer Creek, a runoff-dominated stream. In the Oregon Cascades, where these rivers are located, most of the annual precipitation (and thus recharge) falls as snow during the winter, and the peak flow in Deer Creek represents snowmelt in the springtime. In contrast to the large variability of discharge in Deer Creek, discharge variations in the Quinn River are comparable to the mean annual discharge. In addition, peak discharge occurs in the summer, a few months after snowmelt.

Developing quantitative models to relate input (rainfall, snowmelt) and output (streamflow) has always been a major focus of hydrology (Jakeman & Hornberger 1993). Here, determining this relationship allows one to address the question posed in the introduction, "How much water is flowing?". Moreover, the magnitude and timing of the spring's response to input changes are related to, and thus allow one to determine, hydrogeologic properties of the aquifer(s) through which the water flows. Discharge variations can also be attributed in some cases to temporal changes in the hydrogeologic properties themselves. For example, postseismic changes in spring flow following the Loma Prieta earthquake (Rojstazcer et al 1995) and Kobe earthquake (Tokunaga 1999, Sato et al 2000) have been attributed to seismically-induced permeability changes.

Groundwater systems associated with springs are often structurally complex and poorly characterized. However, long time series of discharge measurements are often available. Thus, one of the more common approaches to studying spring

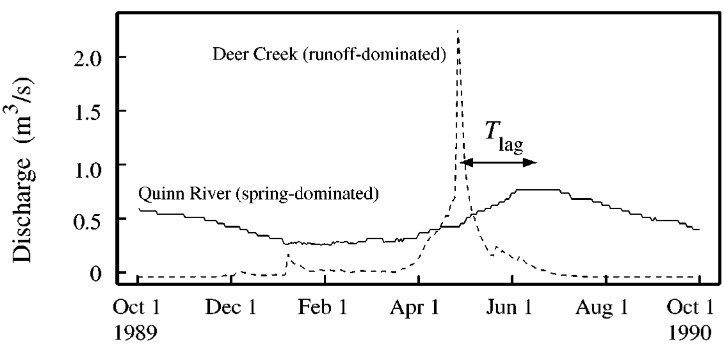

Figure 10 Discharge characteristics of a spring-fed stream. The time lag between recharge and discharge is typically long compared to runoff-dominated streams, and discharge variations are small. From Manga 1996, with permission.

discharge is to use time series and spectral analysis (Mangin 1984, Padilla & Pulido-Bosch 1995, Angelini 1997). Properties such as the cross-correlation of recharge and discharge (Figure 11) and autocorrelation of spring flow are useful for characterizing the nature of temporal discharge variations. More recent studies have applied the techniques of nonlinear time series analysis (Jian et al 1998) and artificial neural networks (Lambrakis et al 2000); both approaches offer improved predictive abilities relative to more traditional time series approaches.

Spectral analysis of spring flow is also informative. Consider an aquifer from which the outflow is a linear function of the average water level in the aquifer. The transfer function, $\eta(\omega)$, which describes the amplitude filtering characteristics of the system as a function of frequency, ω, is given by

$$|\eta(\omega)|^2 = \frac{S_{\text{out}}}{S_{\text{in}}} = \frac{1}{1+\omega^2 T_h^2}, \tag{6}$$

where S_{out} and S_{in} are the power spectra of discharge and recharge, respectively, and T_h is a characteristic hydraulic relaxation timescale. For an unconfined aquifer, with water level changes $\ll$ the mean water depth $\bar{H}$, $T_h = \phi L^2/3K\bar{H}$, where K is hydraulic conductivity and L is the aquifer length (Gelhar 1993). Figure 11 (right) shows that a transfer function with a form given by Equation 6 is compatible with measurements from the Quinn River. Other spectral properties, such as coherence and phase can be interpreted in a similar fashion (Duffy & Gelhar 1986).

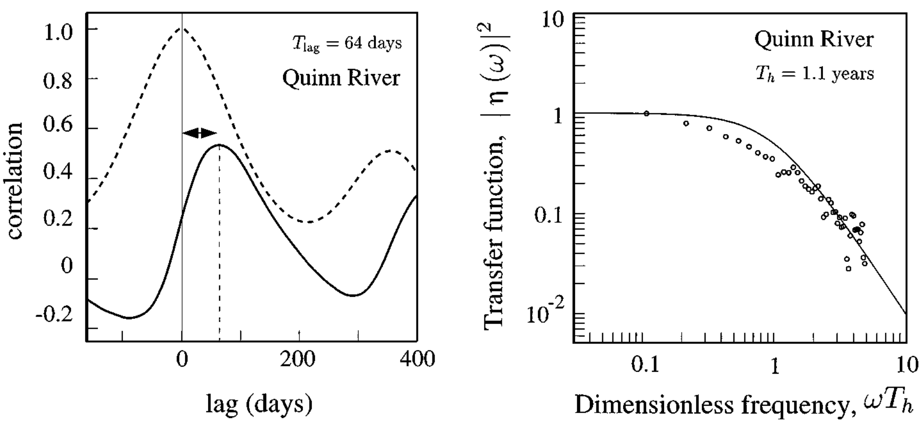

Figure 11 (Left) Cross-correlation between recharge (discharge in a runoff-dominated stream) and spring discharge for the Quinn River. The mean time lag is 64 days. The dashed curve shows the autocorrelation for the Quinn River. (Right) Ratio of output (spring discharge) and input (recharge) spectral densities as a function of frequency. Discharge in a runoff-dominated stream (Deer Creek in Figure 10) is used as a proxy for recharge, and is scaled so that the mean input equals the mean output. Solid curve is the prediction of the linear model (Equation 6). Circles are based on 50 years of discharge measurements with $T_h = \phi L^2/3K\bar{H} = 1.1$ years. Modified from Manga 1999, with permission.

It is also possible to simulate groundwater flow and the resulting spring discharge with models that describe the spatial distribution and values of hydrogeologic properties as well as boundary conditions for the flow. Gvirtzman et al (1997), Angelini & Dragoni (1997), and Eisenlohr et al (1997) provide examples in a range of hydrogeologic settings. Despite the large degree of heterogeneity implicitly associated with springs (see Figure 4), through calibration, such models permit the representation of known and hypothesized geometries and hydraulic properties (Larocque et al 1999). By introducing spatial dimensions to the problem, physically-based flow models are typically characterized by a large number of parameters, many of which may be interrelated. Thus, it is sometimes questioned whether discharge measurements have the ability to resolve more than just a few parameters (Jakeman & Hornberger 1993). Indeed, simplified models based on one-dimensional versions of groundwater flow equations, such as the Boussinesq equation, are often successful at explaining observed stream flow in a wide range of geological settings (Leonardi et al 1996, Manga 1997, Brutsaert & Lopez 1998).

Consider briefly one example of a simple one-dimensional model for flow to the Quinn River in central Oregon. Here the geology consists of stacked layers of Quaternary lava flows. Most of the horizontal groundwater flow occurs within the blocky top of each lava flow—a horizontally conductive layer—whereas in the relatively thick middle of the lava flow, groundwater flow is primarily in the vertical direction through cooling joints (Figure 12). By using Darcy's equation to

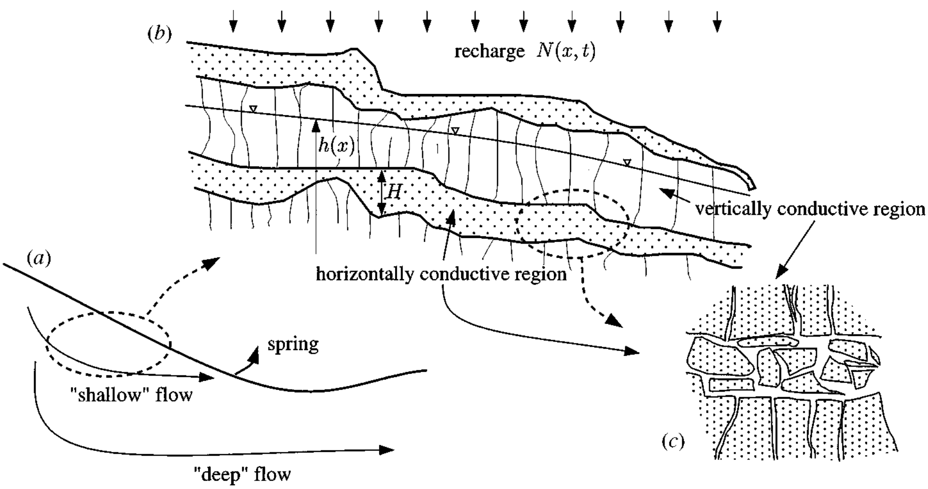

Figure 12 Groundwater flow geometry: (*a*) basin-scale regional flow system (see Figures 7 and 9); (*b*) flow at the scale of individual lava flow units; (*c*) origin of different hydraulic conductivities. The blocky tops of lava flows will have much higher horizontal permeabilities than the dense but jointed interior of flows.

describe flow (not necessarily valid in spring systems because of high flow rates) and conservation of mass, and assuming one-dimensional flow in the x direction, the evolution of head h in the horizontally conductive layer is described by

$$\frac{\partial h}{\partial t} = \frac{KH}{\phi}\frac{\partial^2 h}{\partial x^2} + \frac{N(x,t)}{\phi}, \tag{7}$$

where N describes the recharge to the system, ϕ is the appropriate storage property (here the specific yield or effective porosity) for the system, and K and H are assumed to be constant. Because Equation 7 is a linear diffusion equation, spring flow can be characterized by a diffusive timescale $T_{\text{diff}} = \phi L^2/K\bar{H}$, where L is the length of the aquifer (notice that $T_{\text{diff}} = 3T_h$). As a proxy for the time-dependence of recharge, I use the discharge measured in a nearby runoff-dominated stream and assume that recharge rates are proportional to precipitation (Manga 1997). Despite the gross oversimplification of geometry and the averaging of properties, Figure 13 shows that Equation 7 does a remarkably good job at reproducing the observed discharge.

Groundwater flow models also provide a theoretical basis for one of the oldest and most commonly used techniques of hydrograph analysis, recession flow analysis, which is a widely used technique to study springs (Bonacci 1993, Padilla et al 1994). Recession refers to the decrease in stream flow that follows some event, such as a rainstorm or snowmelt. It is convenient to express the rate of decrease of discharge in the form of a power law relationship,

$$\frac{dq}{dt} = -aq^b. \tag{8}$$

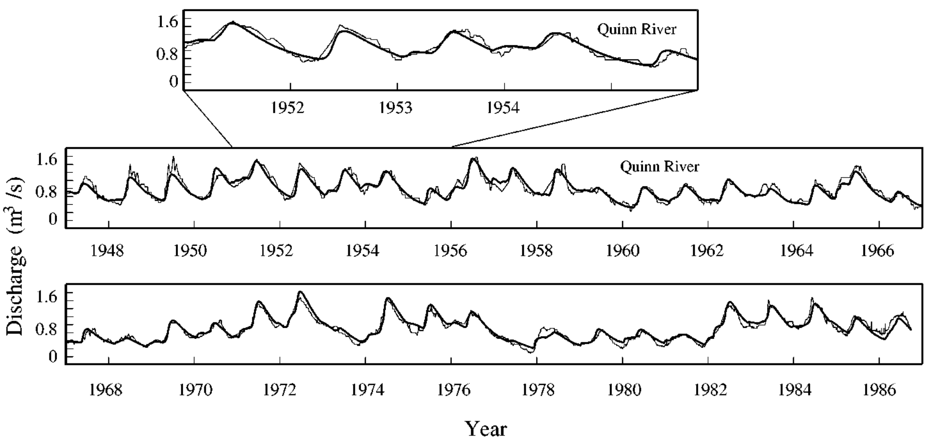

Figure 13 Modeled (bold curve) and measured (thin curve) discharge in the Quinn River. Model is illustrated in Figure 12 and described mathematically by Equation 7 with $T_{\text{diff}} = \phi L^2/K\bar{H} = 3.2$ years. Modified from Manga 1997, with permission.

The coefficients a and b can be related to hydraulic and geometric properties of the aquifer through a comparison with appropriate groundwater flow solutions (Szilagyi et al 1998). For an unconfined aquifer, the oldest and classic solution of Boussinesq (1903) has $b = 1$ (an exponential decrease in discharge) and describes the long-term behavior of the aquifer; by contrast, the short-term solution, representing the discharge of near-stream water, has $b = 3$ (Polubarinova-Kochina 1962).

One attractive feature of recession flow, time series, and spectral analyses is that the techniques involve a statistical analysis of data, which can then be interpreted within a framework provided by simple models. The value of such models is nicely summarized by Gelhar (1993):

> Models of this type are appropriate to treat situations where the time variation of aquifer conditions is the primary concern. Such an approach often will be appropriate when addressing the problems of overall policy and management decisions relating to the behavior of the aquifer over extended periods of time. This kind of model, obviously, cannot be used to address questions of the spatial distribution of the water level. Also, lumped-parameter models are often consistent with the kind of limited data that is available for analyzing such problems.

GEOCHEMICAL MEASUREMENTS

The geochemical composition of spring water is governed by water-mineral equilibria, reaction kinetics, the initial composition of the water and composition of the aquifer rocks, and the rate of groundwater transport. Interpreting the composition of groundwater can thus be complicated. Nevertheless, the composition of spring waters still contains information that can be useful. For example, in a classic study, Garrels & MacKenzie (1967) showed that it is possible to estimate the type and amount of weathering reactions from the chemical composition of spring waters. Geochemical time-series can also be interpreted with the same ideas used to interpret isotopic and discharge time-series, and can provide information about the sources and mixing of different waters (Mazor & Vuataz 1990).

Geochemical measurements at hot springs can further be used to study the conditions experienced by the water at depth. Because solubility, exchange reactions, and isotopic fractionation are temperature dependent, it is possible to estimate the temperature at which water and rock may have equilibrated at depth—the so-called reservoir temperature (θ_r) of the spring water. θ_r is usually greater than the water temperature at the surface owing to conductive heat loss as the water flows to the surface. A variety of geothermometers have thus been proposed (see Fournier 1981 for a review). The suitability and reliability of the different geothermometers depend on the rates of fluid flow at depth relative to the rate of equilibration of water with minerals at θ_r, and on the rate of ascent as the water moves. Additional factors are the nature of the rocks the water traverses and the mixing of deep, hot waters with shallow, cool waters. Figure 14 shows that it is also possible to compute

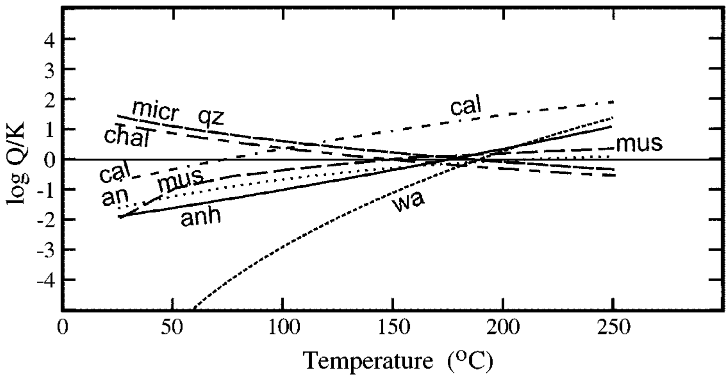

Figure 14 Saturation index $\log(Q/K)$ as a function of temperature for Breitenbush Hot Springs in the Oregon Cascades. Q is the activity product, and K is the equilibrium constant for the reaction: $Q/K < 1$ and $Q/K > 1$ correspond to water that is undersaturated and supersaturated, respectively. Curves are based on geochemical measurements reported by Mariner et al (1993) with sufficient CO_2 added to saturate the water with calcite at the discharge temperature. Convergence of Q/K curves for a number of minerals suggests a reservoir temperature of about 180°C. Abbreviation for minerals: an, analcime; anh, anhydrite; cal, calcite; chal, chalcedony; micr, microcline; mus, muscovite; qz, quartz; wa, wairakite. From Pang & Reed 1998, with permission.

multicomponent equilibria and determine a temperature at which a group of plausible minerals are in equilibrium with each other and the aqueous phase. Such equilibria calculations are sensitive to the often poorly known aluminum concentration (Pang & Reed 1998), and it is also sometimes necessary to account for degassing of CO_2 during ascent of the water to the surface (Reed & Spycher 1984). In Table 2, I list inferred reservoir temperatures for Breitenbush hot spring (see map in Figure 3) based on various geothermometers—a value of $\theta_r \approx 180$°C seems reasonable. Mariner et al (1993) suggest explanations for the range of reservoir temperatures listed in Table 2.

The energy budget for small hot springs can be reconsidered based on the estimate of a reservoir temperature. The Breitenbush hot spring, for example, with $\theta_r \approx 180$°C discharges about 13 L/s (Table 1). By assuming a mean recharge temperature of 5°C, Equation 5 implies that the spring discharges >10 MW of heat. The $>$ sign is used because the spring discharge is a lower bound on the total flux of hot fluid—the absence of ^{3}H in the water indicates that mixing with younger near-surface waters did not occur during ascent (Mariner et al 1993). Presumably much of this heat comes from the solidification and cooling of intrusives within the Cascades volcanic arc. The solidification of about 17 km^3 of magma/million years produces a heat flow of 1 MW (Ingebritsen et al 1989). Thus, a volume of at least 170 km^3 of magma/million years needs to be intruded to maintain this modest-size hot spring. This is a large volume of magma and implies that most hot springs are transient over time periods $>$O(10^5) years (Ingebritsen & Sanford 1998).

TABLE 2 Reservoir temperatures for Breitenbush hotspring (Figure 3) inferred from geothermometers. Surface temperature is 84°C

Geothermometer	Temperature[a] (°C)
Quartz[b]	166
SO_4-H_2O[c]	178
Anhydrite saturation[d]	176
K-Mg[d]	129
Na-K[d]	171
Na-K-Ca[d]	148
Na-Li[d]	202
fixAl[c]	~180

[a]All results are from Mariner et al 1993 except the fixAl result of Pang & Reed 1998.

[b]Based on mineral solubility.

[c]Based on temperature-dependent isotope fractionation.

[d]Based on empirical relation of various cation ratios to temperature.

[e]Based on multiple mineral equilibria with adjusted Al concentration.

CONCLUDING REMARKS

A variety of hydrological problems can be studied from measurements made at springs. These problems include determining from where the water came, the residence time and velocity of water, hydraulic properties of the aquifers, and the spatial scale and subsurface extent of the flow system. Because the water interacts with its geological environment, there is also the possibility of obtaining information about subsurface geological and geophysical processes. Examples discussed here include determining heat flow, quantifying magmatic degassing, and estimating geothermal reservoir temperatures. Finally, not only do springs provide access to water that has sampled distant regions, but in some cases they also discharge water from distant periods in time and preserve information about paleoclimate.

Signs posted at the headwaters of two of the springs I have discussed in this review illustrate some of the issues that can be addressed by studying springs. At the source of the Quinn River a sign states:

> The crystal clear water from this spring may have fallen several years ago as snow in the high Cascades. Each year, as the snow melts, most of the water seeps into the cracks of this "geologically young" country and travels through underground channels.

By using the stable isotopes of O and H, I confirm that recharge occurs along the crest of the Cascades. The analysis of radiogenic ^{3}H and its decay product ^{3}He suggests a mean age of about 1 year. The sign notes that the terrain is also "young." As water flows through these young volcanic rocks, it may dissolve and discharge gases released at depth from the solidification of magmas—measurements at springs provide a means of quantifying this degassing and inferring intrusion rates. At the headwaters of the Metolius River, a sign reads

> Down this path a full-sized river, the Metolius, flows ice cold from huge springs. The springs appear to originate from beneath Black Butte. However, geologists say this is misleading and believe the springs have their origin in the Cascade Mountains to the west.

Again, I use stable isotopes of O and H to confirm that the water originated as meteoric water along the crest of the Cascades. The sign also emphasizes the cold temperature of the water (8.3°C). In fact, despite the low temperature, because this spring is "huge" it carries virtually all the geothermal heat produced in the drainage basin (area about 400 km^2) from a region with a high background heat flux.

In summary, measurements made at springs integrate the signal of geological and hydrological processes over large spatial areas and possibly long periods of time—this can be either good, bad, or ugly, depending on the hydrogeologic information that is sought and the geological processes being studied. The challenge is that springs do not always directly measure the geological or hydraulic properties of interest. Thus, the interpretation of measurements requires the development of a model or mathematical framework in which a set of measurements can be related to processes.

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Figure 2 The Quinn River, Oregon (*left*) and Cultus River, Oregon (*right*) begin as full-fledged rivers that emerge from springs. A person, shown for scale, is indicated by an arrow in each photograph. (Adapted with permission from Manga 1999.)