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A model for discharge in spring-dominated streams and implications for the transmissivity and recharge of quaternary volcanics in the Oregon Cascades

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Abstract. A model for discharge in spring-fed streams is described and applied to streams in the Oregon Cascades. These streams are assumed to be fed by an unconfined aquifer composed predominantly of quaternary basalts and basaltic andesites. The model is based on the unsteady Boussinesq equation and is characterized by a single parameter, a normalized length scale of the aquifer. Flow in a runoff-dominated stream is used as a proxy for the time-dependent groundwater recharge. Four spring-dominated streams with 50 years of daily discharge records are studied, and modeling efficiencies are in the range of 0.76–0.89. The effective transmissivity is found to be approximately proportional to the length scale of the aquifers. Groundwater recharge rates are in the range of 66–127 cm/yr and 40–73% of the mean annual precipitation.

1. Introduction

A central problem in hydrology is understanding and characterizing the relationship between the response of hydrologic systems (such as streamflow), various inputs (such as rainfall), and the physical attributes of the system. Determining such hydrologic relationships at the regional scale involves developing models and fitting parameters, for example, developing models that relate rainfall/snowmelt and runoff. Owing to the geologic and hydrologic complexity of most natural systems, models that are physically based are typically characterized by a large number of physical and geometric parameters [e.g., *Beven*, 1989]. Alternatively, conceptual models can be developed and can often successfully account for measured streamflow even when characterized by a small number of parameters [e.g., *Jakeman and Hornberger*, 1993]. Here we show that a certain class of stream hydrographs, namely those of spring-fed streams, can sometimes be explained by physically based models that are characterized by a small number of parameters.

Unlike "typical" runoff-dominated streams, those that are spring-dominated often have simple discharge characteristics (terminology after *Whiting and Stamm* [1995]). The hydrographs of spring-dominated streams often exhibit very little variation, even in regions with extended dry seasons or following a period of significant precipitation or snowmelt. In such systems, subsurface flow originating from an aquifer (as opposed to direct runoff, which consists of both surface flow and interflow) accounts for most of the streamflow. Previous studies have shown that it is possible to explain many of the characteristics of such spring-fed streams with simplified physically based models [e.g. *Manga*, 1996; *Leonardi et al.*, 1996].

Here we describe a one-parameter model for discharge in spring-dominated streams that is based on the Boussinesq equation for unsteady flow in an unconfined aquifer and apply the model to six streams and rivers near the headwaters of the Deschutes River in the High Oregon Cascades. A more detailed and complete hydrologic model for the entire Upper Deschutes River Basin is currently being developed by the U.S. Geological Survey (USGS) [*Gannett et al.*, 1996]. Our model is physically based in that it follows from Darcy's equation and conservation of mass, but it greatly simplifies basin geometry and groundwater recharge patterns. Despite these approximations, the model allows us to determine an effective transmissivity (see work by *Sanchez-Vila et al.* [1996] for a definition) of the aquifers that feed the springs and provides constraints on groundwater recharge and evapotranspiration rates.

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2. Geologic and Hydrologic Setting

We focus on a set of four spring-dominated streams in the Upper Deschutes River basin in the High Oregon Cascades: Quinn River, Cultus River, Browns Creek, and Fall River (Table 1 and Figure 1). Daily discharge measurements were made by the USGS from 1939 to 1991. Discharge in these spring-dominated streams typically varies by less than a factor of 2 over the course of a year (hydrographs are shown later in Figure 5). Peak discharge typically occurs in the summer or fall, even though most of the precipitation falls as snow, and peak snowmelt, as observed in runoff-dominated streams, occurs in the spring [*Whiting and Stamm*, 1995]. By contrast, runoff-dominated streams, for example, Cultus Creek, Deer Creek, and Charlton Creek (Table 1 and Figure 1), have relatively flashy hydrographs with little or no base flow for much of the year (e.g., the hydrograph for Deer Creek, shown later in Figure 5).

The local geology consists primarily of Quaternary basalts and basaltic andesites. The Mount Bachelor volcanic chain, which comprises about half the recharge area for the Fall River (shown later in Figure 10), consists of basalt and basaltic andesites that erupted from cinder cones and shield volcanos from about 18 to 8 ka [*Scott and Gardner*, 1992]. The recharge area for the other streams is underlain primarily by basaltic andesites with well preserved surface features; a K-Ar date of this unit gives an age of 0.73 Ma [*Sherrod*, 1991]. The entire region was covered at 7 ka by Mazama ash; the thickness of the ash layer varies from about 0.5 m in the southern parts shown

Table 1. Stream Locations

Name	Number	Elevation, m	Latitude, Longitude
	Spring-Dominated Streams		
Cultus River	14050500	1356	43°49′06″, 121°47′40″
Quinn River	14052500	1354	43°47′03″, 121°50′06″
Browns Creek	14054500	1332	43°42′57″, 121°48′10″
Fall River	14057500	1286	43°47′48″, 121°34′18″
	Runoff-Dominated Streams		
Cultus Creek	14051000	1385	43°49′17″, 121°49′22″
Deer Creek	14052000	1378	43°48′48″, 121°50′18″
Charlton Creek	14053000	1359	43°46′51″, 121°50′06″
	Other Streams Studied Here		
Deschutes River	14051000	1385	43°49′17″, 121°49′22″
Snow Creek	14074900	1378	44°06′59″, 121°39′34″

in Figure 1 to less than 10 cm in the northern parts of the study area shown in Figure 1 [*MacLeod and Sherrod*, 1992].

Annual precipitation in the (assumed) recharge area varies from about 20 inches (50 cm) in the east to more than 100 inches (250 cm) near the top of the South Sister and Broken Top mountains (see inset of Figure 1). The recharge area is forested by Ponderosa and Lodgepole pines at lower elevations. At higher elevations, and along the crest of the Cascades, forests are dominated by Hemlock and mixed conifers. As we will show later, groundwater recharge rates are very high, presumably a result of the high permeability of the young lava flows and the layer of Mazama ash.

The four spring-dominated streams discharge meteoric water. In Figure 2 we plot the measured $\delta^{18}O$ and δD values (in parts per thousand relative to standard mean ocean water) for

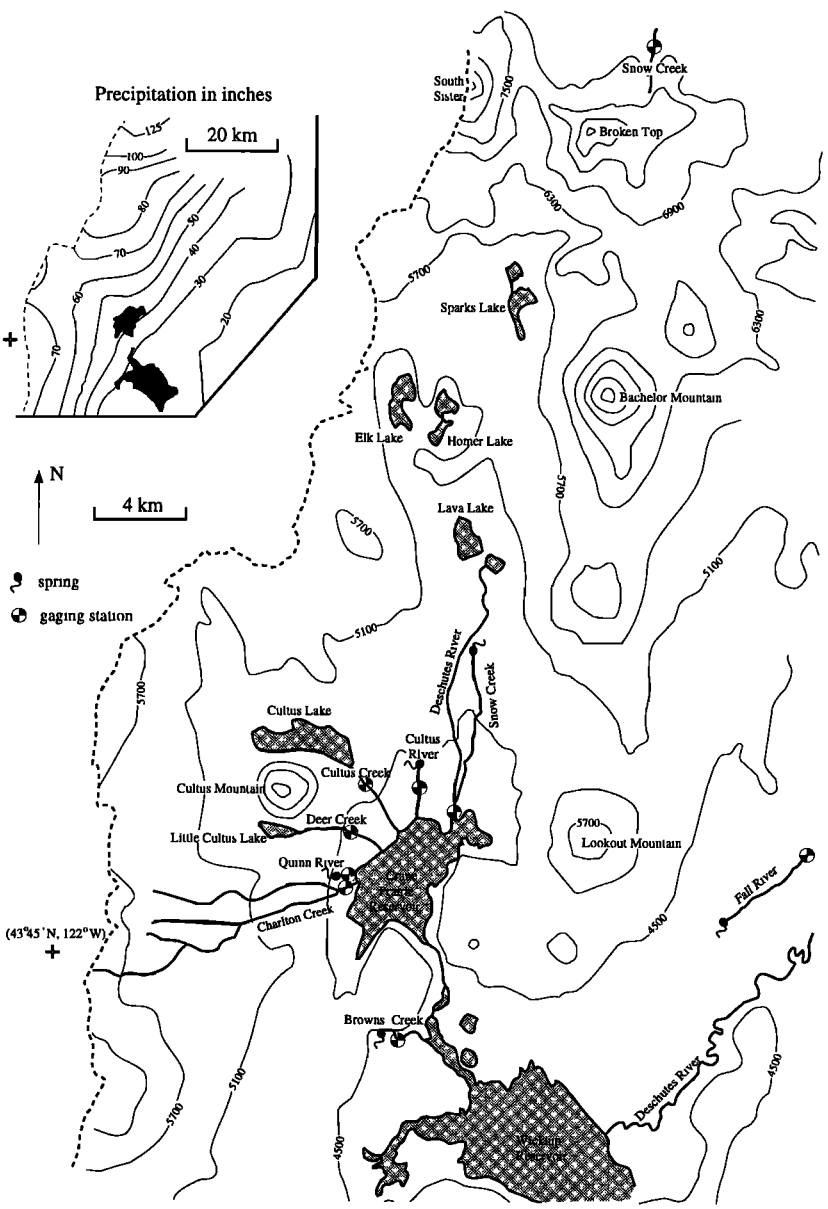

Figure 1. Map showing the location of streams, gaging stations, and elevation (in feet). The dashed curve shows the crest of the Cascades. Only streams studied in this paper are shown. Inset shows the isohyetal map of mean annual precipitation in inches per year of rainfall equivalent (based on records from 1961 to 1991 compiled by G. Taylor, state climatologist, Oregon Climate Service).

waters sampled from the four spring-dominated streams on September 28, 1996 (solid disks). For comparison we also plot the global meteoric water line, $\delta D = 8\delta^{18}O + 10$. The close agreement between the meteoric water line and the measurements is probably the result of integrating the compositions of rainfall and snowmelt from at least several decades of precipitation. For comparison with the (large) springs studied here, in Figure 2 we show $\delta^{18}O$ and δD values (from *Ingebritsen et al.* [1994]) for other springs on the east and west side of the Cascades. Deviations from the meteoric water line of these other springs are attributed to local weather pattern variability and varying amounts of evaporation [*Ingebritsen et al.*, 1994].

3. Model

We assume that the spring-dominated streams are fed by unconfined aquifers that are replenished by rainfall and snowmelt (because meteoric water is being discharged). Here we describe a one-parameter model, illustrated in Figure 3, for subsurface flow and discharge that can account for the two distinctive features of the hydrographs of spring-dominated streams: (1) peak discharge occurs in the summer or fall even though snowmelt typically occurs in April or May, and (2) discharge varies by less than a factor of about 2 over the 50-year time period for which we have stream gaging records. Below we describe the governing equations, boundary conditions, and the numerical methods employed to determine the numerical value of the one parameter which characterizes the model.

3.1. Boussinesq Equation

The evolution of a water table in an unconfined aquifer, with height h, can be described by the Boussinesq equation for unsteady flow with accretion, which follows from conservation of mass and Darcy's equation [e.g., *Bear*, 1969, section 8.2]. In one horizontal dimension, the Boussinesq equation is given by

$$\frac{\partial h}{\partial t} = \frac{k\rho g}{\mu S_y}\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right) + \frac{N(x,t)}{S_y} \tag{1}$$

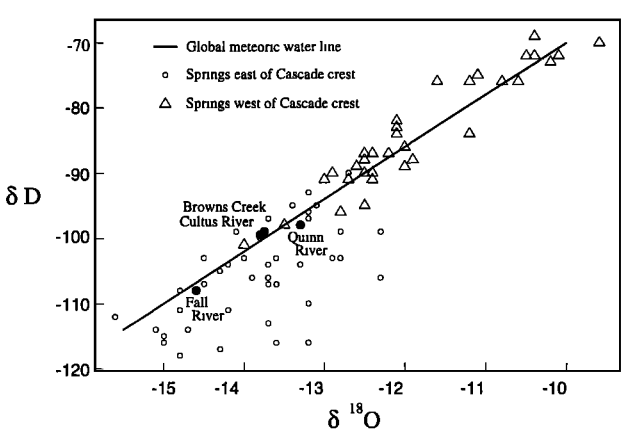

Figure 2. Relationship between $\delta^{18}O$ and δD values (in parts per thousand relative to standard mean ocean water) for the four spring-dominated streams studied here (solid circles). For comparison, values for other springs on the east and west side of the crest of the cascades are shown with open circles and open triangles, respectively [from *Ingebritsen et al.*, 1994]. The solid line is the global meteoric water line $\delta D = 8\delta^{18}O + 10$.

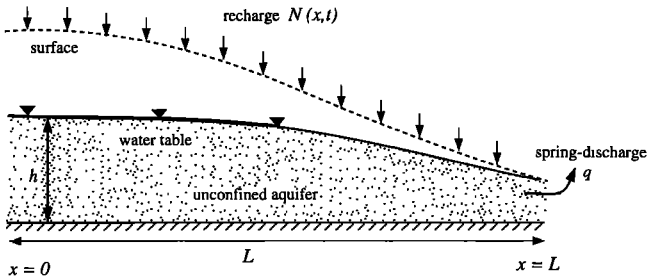

Figure 3. Geometry of the one-dimensional model.

where k, ρ, μ, g, S_y, t, and x are permeability, water density, water viscosity, gravitational acceleration, effective porosity or specific yield [*Freeze and Cherry*, 1979, p. 61], time, and horizontal position, respectively, and N describes the time-dependent water input. Equation (1) for saturated flow is subject to the standard Dupuit approximations, ignores the capillary fringe and any seepage faces, and assumes the aquifer is isotropic and homogeneous and that the aquifer has a horizontal and relatively impermeable base.

The Boussinesq equation is frequently linearized by setting $h(x, t) = h_0 + h_1(x, t)$, where h_0 is a constant; assuming h_1/h_0 is small so that quadratic terms in h_1 can neglected, (1) becomes

$$\frac{\partial h_1}{\partial t} = \frac{k\rho g h_0}{\mu S_y}\frac{\partial^2 h_1}{\partial x^2} + \frac{N(x,t)}{S_y}. \tag{2}$$

The coefficient $k\rho g h_0/\mu S_y$ can be interpreted as a diffusivity that describes the rate at which variations in the height of the water table decay.

Manga [1996] showed that a model based on the Boussinesq equation can quantitatively explain the observed time lag between snowmelt and peak discharge at the springs, as well as the magnitude of annual discharge variations. *Leonardi et al.* [1996] successfully applied a similar model based on the diffusion equation to springs in Armenia which discharge from basaltic aquifers.

3.2. Recharge

Because runoff is related to the same events that result in groundwater recharge, namely, rainfall and snowmelt, we use flow in a runoff-dominated stream as a proxy for the time-dependent recharge, $N(t)$. Groundwater recharge may also vary in space. In this case we expect $N(x)$ to increase with increasing distance from the spring because the mean annual precipitation increases upslope, which, for all the streams studied here, corresponds to increasing distance from the mouth of the springs (see Figure 10, shown later). Here we assume N increases linearly with increasing distance from the spring.

The model described here is similar to that of *Manga* [1996] except for the assumed spatial distribution of recharge: *Manga* [1996] assumed that recharge occurred at a uniform rate over the region $0 < x < L'$ and was zero over the region $L' < x < L$.

3.3. Boundary Conditions and Geometry

We assume that flow is primarily in one direction so that we can apply the one-dimensional model, (2). This approximation is not unreasonable because the recharge areas (see Figure 10) are long and narrow. In addition, estimated water table heights in this region are also compatible with nearly unidirectional flow beneath the recharge area towards the springs [*Gannett et*

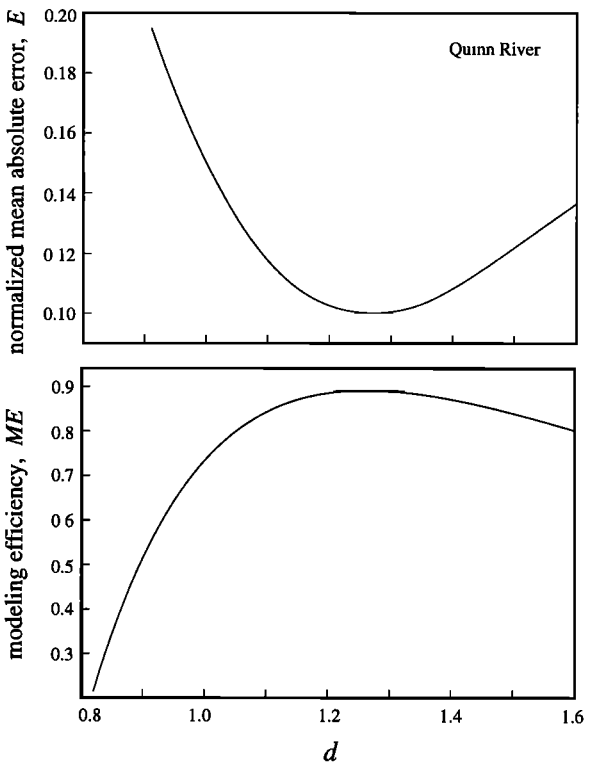

Figure 4. Normalized mean absolute error (top), equation (6), and modeling efficiency (bottom), equation (7), as a function of the diffusion length scale d, equation (5), for the Quinn River. The best model parameter is $d = 1.27$.

al., 1996]. We let L denote the length of the aquifer, and at $x = L$ we apply the boundary condition $h = h_0$. Discharge, q, is proportional to $\partial h/\partial x$ at $x = L$.

3.4. Modeling Procedure

In order to nondimensionalize the linearized Boussinesq equation, we choose ω to be the characteristic recharge frequency (1 year) and the characteristic length scale x_c to be the "diffusive" length scale (see work by *Manga* [1996] for a detailed discussion),

$$x_c = \left(\frac{2k\rho g h_0}{\omega \mu S_y}\right)^{1/2}. \tag{3}$$

Thus setting $t' = \omega t$ and $x' = x/x_c$, the linearized Boussinesq equation becomes

$$\frac{\partial h_1}{\partial t'} = \frac{1}{2}\frac{\partial^2 h_1}{\partial x'^2} + \frac{N(x', t')}{\omega S_y} \qquad h_1 = 0 \text{ at } x' = L/x_c = d. \tag{4}$$

We solve the linearized Boussinesq equation numerically using a Crank-Nicholson finite-difference method.

Our model, (4), is characterized by a single parameter, the normalized length of the aquifer,

$$d = L/x_c. \tag{5}$$

The average annual recharge is determined by requiring that the mean stream discharge equals the mean groundwater recharge.

In order to determine the best value of d we minimize the mean-absolute error

$$E = \frac{1}{N}\sum_{i=1}^{N} |q_i^M - q_i^O|, \tag{6}$$

and maximize the modeling efficiency

$$\mathrm{ME} = 1 - \frac{\sum_{i=1}^{N} (q_i^M - q_i^O)^2}{\sum_{i=1}^{N} (q_i^M - \bar{q}^M)^2}, \tag{7}$$

where q is discharge, the superscript M denotes the model prediction, the superscript O denotes the observed values, and $\bar{q}$ is the mean of q. In all cases the choice of d that minimizes E is similar to the value that maximizes ME. An example of E and ME for different values of d is shown in Figure 4 for the Quinn River.

4. Results

Here we consider four spring-dominated streams for which reliable daily discharge records exist over a period of 50 years: Quinn River, Cultus River, Browns Creek, and Fall River (Figure 1 and Table 1). As discussed in the previous section, we use discharge in a runoff-dominated stream as a proxy for recharge. For the results presented here we use flow in Deer Creek. We have also carried out the same analysis using the hydrographs of two other nearby runoff-dominated streams (Cultus Creek and Charlton Creek; see Figure 1 and Table 1) and obtained nearly identical results.

Before presenting and discussing the results, it is important to highlight certain key model approximations and assumptions so that results are not misinterpreted: the model is a one-dimensional approximation of a two-dimensional problem; the parameter d describes an effective (aquifer-scale) transmissivity; the aquifer is assumed to be homogeneous and isotropic; and recharge in the entire region is assumed to be proportional to discharge in the runoff-dominated streams. We also note that there is some uncertainty and error associated with the streamflow measurements (determined by relating measured water stage to discharge). In particular, the highest peak flows and the low base flows in Deer Creek, which we use as a proxy for recharge, are noted as being estimated values in the *Water Resources Data, Oregon* [1992] books.

For all four spring-dominated streams there is a well-defined minimum E and maximum ME (see Figure 4 for the Quinn River). Our best values for d and the corresponding modeling efficiencies (based on the entire time series) are listed in Table 2. Modeling efficiencies are in the range of 0.76–0.89 and are comparable to those of the IHACRES lumped conceptual rainfall-runoff model [e.g., *Schreider et al.*, 1996]. In Figure 5 we show observed hydrographs (thin curves) and model hydrographs (bold curves) for all four streams over the time period 1947–1987. In general, the model describes both the annual and decadal discharge variations.

Table 2. Model Parameters and Modelling Efficiencies for Spring-Dominated Streams

Name	d	ME
Quinn River	1.27	0.89
Cultus River	1.54	0.76
Brown Creek	1.78	0.80
Fall River	2.50	0.78

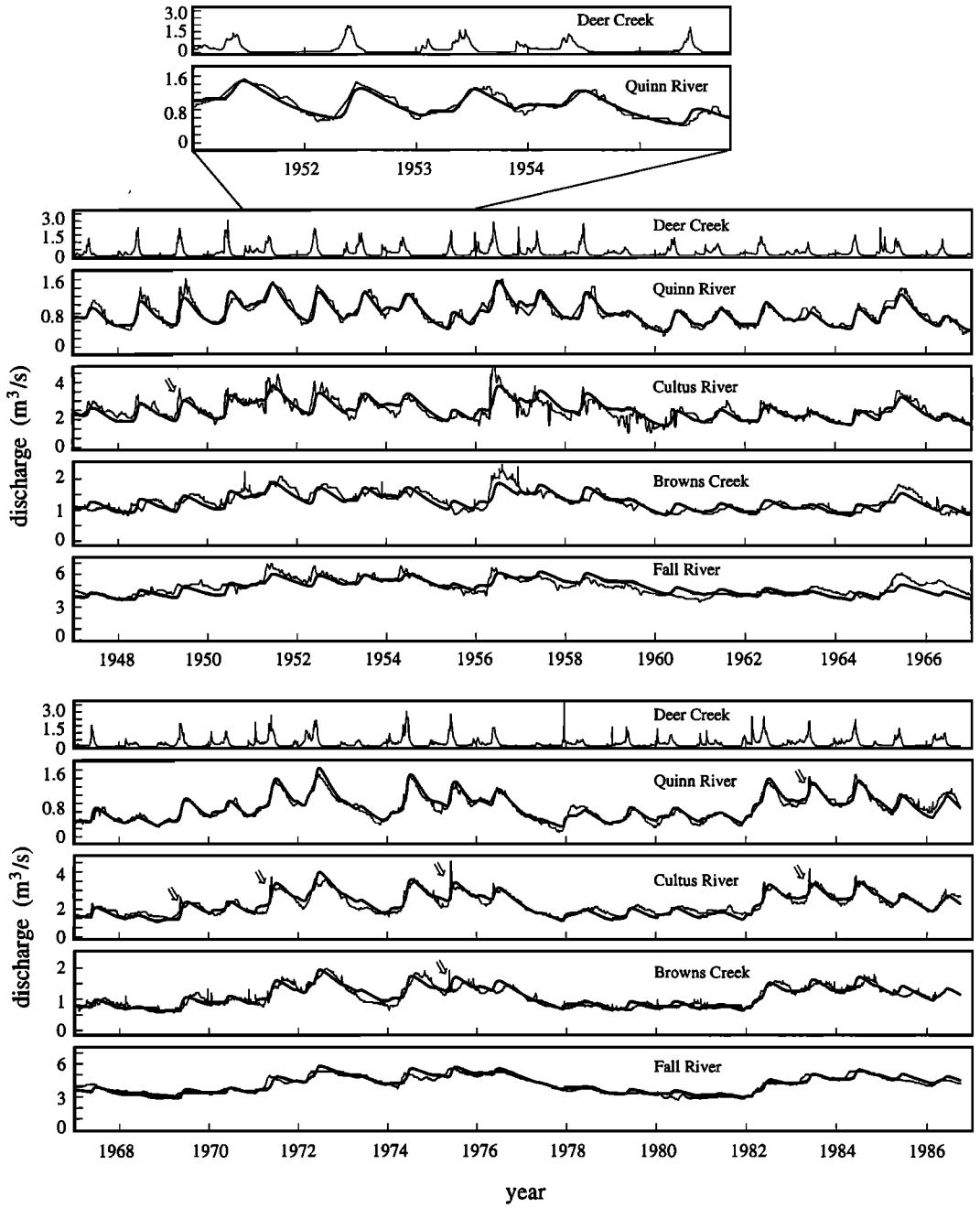

Figure 5. Discharge in spring-dominated streams for Quinn River, Browns Creek, Cultus River, and Fall River for the period 1947–1986 (time in calendar years). Observed discharge is shown with a thin curve, and the model discharge is shown with a bold curve. Arrows indicate possible surface runoff contributions.

The range of d for these four streams is 1.27–2.50. For the Quinn River ($d = 1.27$) the annual discharge variations are large. At the other extreme, for the Fall River ($d = 2.50$) the annual discharge variations are small, and the largest streamflow variations occur on the timescale of several years and are related to drought years. In particular, water years (October–September) 1973 and 1977 had almost no precipitation (see Deer Creek hydrograph, Figure 5), and all four springs have no new pulse of discharge in the summer or fall.

We note that the model successfully reproduces the slopes of both the rising and falling limbs of the hydrograph (see inset for the Quinn River) and in general reproduces the amplitude of discharge variations. In Figure 6 we plot the difference between observed and modeled discharge for the Quinn and Fall Rivers, the streams with the smallest and largest value of d, respectively. The largest differences are generally at the beginning of the rise of the hydrograph. For both streams these differences appear as negative spikes and indicate that the model predicts a slightly earlier onset of increasing streamflow. In particular, the small response time difference between the model and observation is most clear for the Fall River. This discrepancy could be due to a slight delay between recharge of Fall River's aquifer relative to that inferred from Deer Creek; such variations might be associated with microclimate variability.

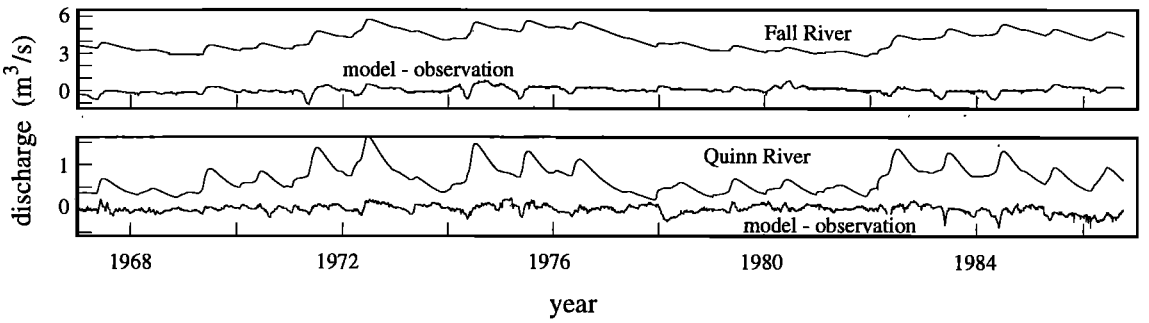

Figure 6. Model prediction and difference between model prediction and observation for Quinn River (top) and Fall River (bottom), streams with the smallest and largest values of d, respectively.

In order to determine d for the results shown in Figure 5 and Table 2, we used the entire time series. In Figure 7 we show values of d obtained using shorter time series of 1 year (open circles) and 7 years (solid circles) centered around the symbol (i.e., the value of d for the 7-year time interval 1947–1953 is shown by a solid circle at 1950). The greater variability of inferred d for Fall River based on 1 year of data is due to the relatively small signal which remains after subtracting the mean. For example, between 1978 and 1982 the hydrograph is nearly flat so that determining d is problematic. For the Quinn River, annual discharge variations are large enough that 1 year of flow measurements is sufficient provide reasonable estimates of d. In fact, using discharge measurements at 1-month intervals over a time period of 1 year is sufficient to determine d for the Quinn River.

For the Cultus River there are a few years with a spike in the springtime (indicated with arrows in Figure 5). Between the spring source and gaging station there is a small runoff-dominated stream (indicated with a blue-dashed curve on 7.5-min USGS topographic maps) which suggests that these spikes are due to a runoff component. Indeed, these spikes are correlated in time with peaks in the runoff-dominated streams.

5. Application to Other Streams

Here we apply the model to two other stream systems for which daily discharge records are available: Snow Creek (north of Broken Top Mountain) and the Deschutes River above Crane Prairie Reservoir (Figure 1 and Table 1). Both streams have no diversion or regulation upstream, and both have discharge characteristics similar to the spring-dominated streams studied in the previous section.

5.1. Snow Creek

The model described above was applied to Snow Creek, and the resulting observed and predicted hydrographs are shown in

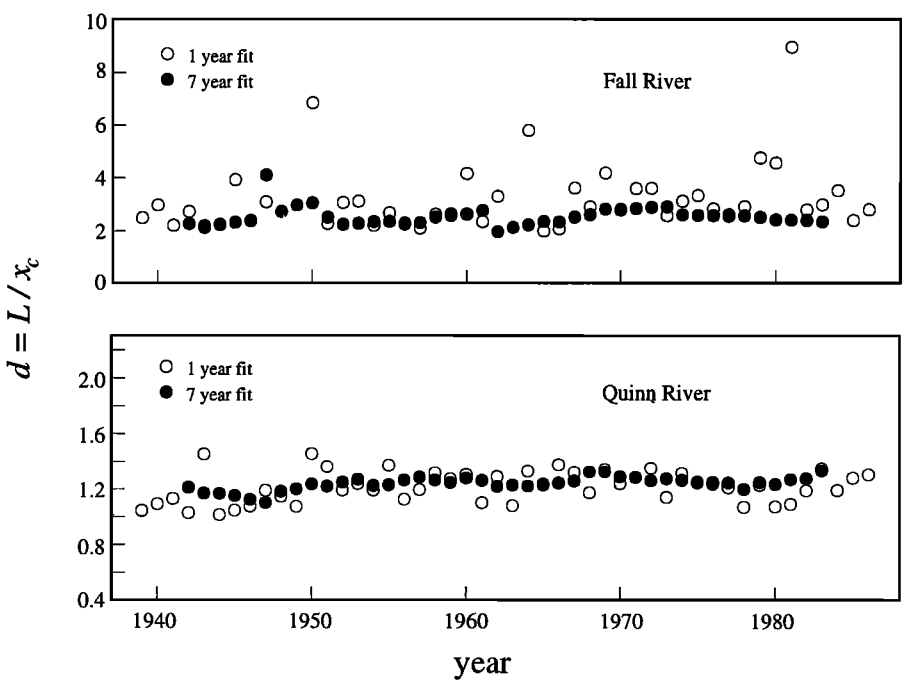

Figure 7. Model values of the diffusion length scale, d, using only selected lengths of time series for Fall River (top) and Quinn River (bottom). The open circles are values of d based on 1 year of data corresponding to the date on the x axis. The solid circles are values of d based on 7 years of data centered around the date on the x axis.

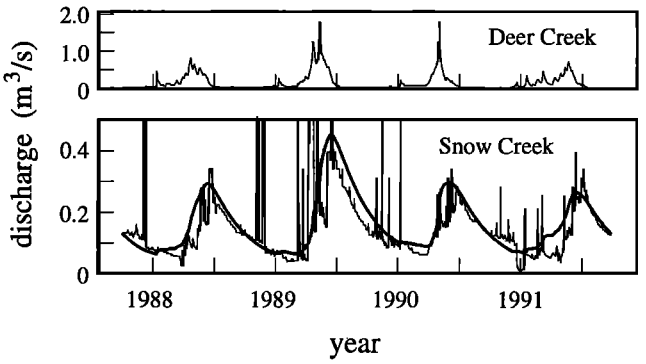

Figure 8. Model discharge (bold curve) and observed discharge (thin curve) for Snow Creek.

Figure 8, again using the Deer Creek hydrograph as a proxy for recharge; the best fit occurs for $d = 0.90$. The Snow Creek hydrograph clearly contains two components. The first is a short-period and large-amplitude component (big spikes in Figure 8) associated with runoff and interflow from snowmelt in the springtime and rainfall. The model successfully describes a second component which has a long period and small amplitude, and in our model corresponds to flow associated with groundwater flow. The modeling efficiency of the best model is 0.057 (a small value because the model does not reproduce the large and frequent runoff-dominated discharges).

5.2. Deschutes River

The hydrograph of the Deschutes River, just above the Crane Prairie Reservoir, also has features similar to the spring-dominated streams studied in section 4. The Deschutes River drains the spring-fed Lava Lake (Figure 1), and the gaging station record for the Deschutes River also contains a smaller contribution from spring-dominated Snow Creek (a different Snow Creek from the one studied in section 5.1).

The model described here results in ME = 0.69 and underestimates peak discharges by several cubic meters per second. Because this modeling efficiency is low, we also used the model studied by *Manga* [1996]. Briefly, rather than assuming recharge was proportional to distance from the spring, we assume that the recharge occurs uniformly in the region $0 < x < L'$ (and that the aquifer extends from $0 < x < L$); thus two parameters need to be estimated, d and L'/L. For the Deschutes River we find $d = 1.07$ and $L'/L = 0.24$, resulting in ME = 0.82. For comparison, the four spring-dominated streams studied in the previous section are best fit by $L'/L \approx 0.5$ for this second model. Model and observed hydrographs are shown in Figure 9. Although this second model results in a better modeling efficiency, it still underestimates peak discharge, at least for years with peak discharges greater than about 10 m³/s.

The ability of this second model to account better for discharge variations may provide information about the subsurface flow pathways for the Deschutes River's aquifer. The model suggests that most of the recharge occurs over a small area at some distance from the spring because L'/L is so small. The basin drained by the Deschutes River is a large basin (area ≈340 km²) surrounded by the crest of the Cascades to the west, South Sister and Broken Top mountains to the north, and Mount Bachelor to the east. A number of springs and runoff-dominated streams discharge into Sparks, Homer, and Elk lakes from which there is no outlet. The small values of L'/L and d (compared with the Fall River) suggest that much of recharge for the aquifer which feeds the Lava Lakes and Deschutes River occurs near Sparks, Elk, and Homer lakes.

6. Discussion

The modeling results indicate that the discharge characteristics of the spring-dominated streams in any given year are governed primarily by the previous winter's snowmelt and rainfall. For example, in the 1972–1973 and 1976–1977 winters there was relatively little precipitation (see Figure 5), and the hydrographs for all the streams show a smooth decrease over a 2-year period. However, as discussed in section 6.2, the "age" of the groundwater being discharged is probably in the range of decades to centuries. Clearly there are two timescales operating in the unconfined aquifer: the first is the timescale over which variations in the height of the water table propagate (a diffusive timescale), and the second is related to the actual velocity of water in pore spaces, which moves much more slowly.

The time lag between peak runoff and peak flow in spring-dominated streams is related to the time required for a "pulse" of groundwater to propagate and diffuse the length of the aquifer. As the pulse moves, its amplitude decreases. Thus, as the time lag increases, discharge variations decrease. Such a relationship is characteristic of diffusive processes and is consistent with our model, (4), which is of the same form as a diffusion equation.

6.1. Length Scale of the Aquifers

In Figure 10 we show estimated recharge areas for the spring-dominated streams Quinn River, Browns Creek, Cultus

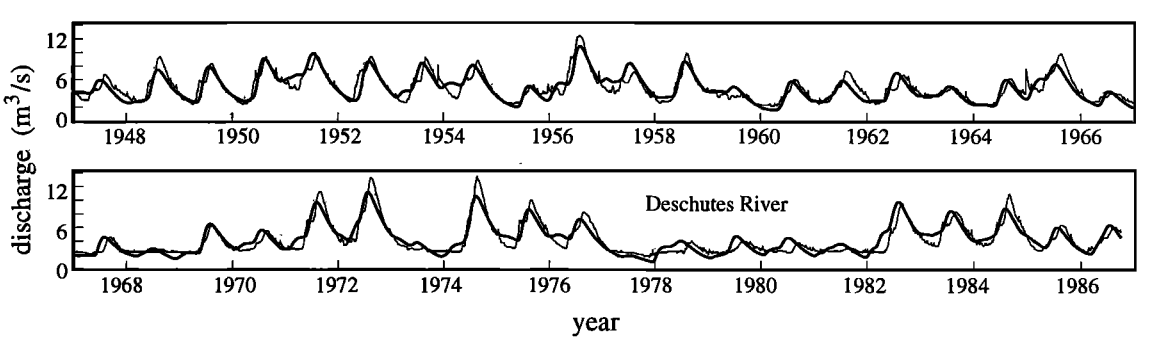

Figure 9. Model discharge (bold curve) and observed discharge (thin curve) for the Deschutes River above Crane Prairie Reservoir.

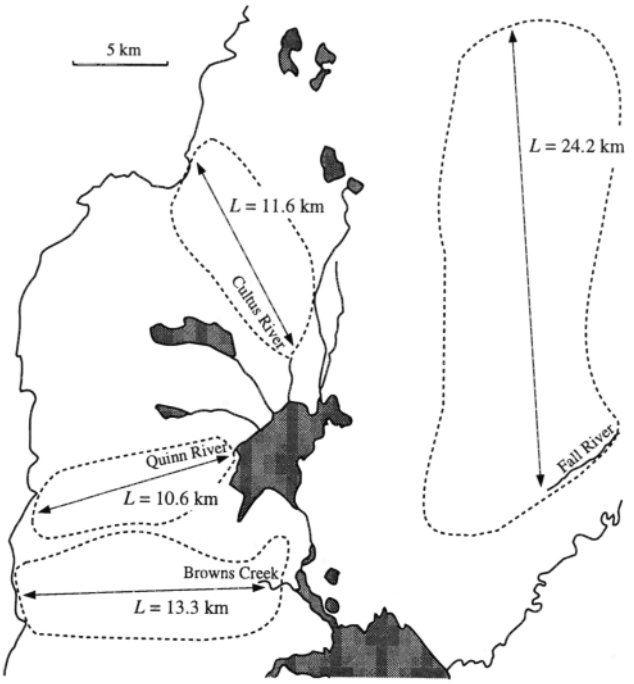

Figure 10. Estimated regions (areas bounded by the dashed curves) of groundwater recharge for Quinn River, Browns Creek, Cultus River, and Fall River. L is the characteristic length scale for groundwater flow.

River, and Fall River, as well as the characteristic length scale L of the aquifers (10.6, 13.3, 11.6, and 24.2 km, respectively). The boundaries of these regions are based on surface topography. However, as noted in the USGS *Water Resources Data, Oregon* [1992] publications, the actual recharge area is uncertain because of interbasin exchange. We assume hereafter an uncertainty of 15% in our estimated length scale L.

6.2. Determining Groundwater Age From Temperature

In order to estimate the age of the groundwater being discharged at the springs, we recorded water temperatures at the mouth of the springs. Measurements were made once per month for 5 months, and the temperature was found to be constant to within the accuracy of the thermometer (±0.2°C). For Quinn River, Browns Creek, Cultus River, and Fall River we measured 3.4, 3.6, 3.6 and 5.9°C, respectively. Temperatures are shown in Figure 11 as a function of the length scale L of the aquifer (see Figure 10).

We can relate spring temperature to groundwater age as follows. Because most of the water enters the ground as snowmelt, it should be close to freezing (y intercept on the graph). As the water flows towards the spring, it is slowly warmed by input of geothermal heat. Now consider a vertical column of water with height h. Assuming the direction of fluid flow is perpendicular to the temperature gradient (an assumption consistent with the Dupuit approximation used in the model), the rate of temperature increase is given by

$$\frac{dT}{dt} = \frac{Q}{\rho C h S_y} \tag{8}$$

where Q is the heat flux into the bottom of the aquifer and C is the specific heat. Choosing $Q = 100$ mW/m^2 [e.g., *Blackwell and Priest*, 1996], $h = 500$ m, and $S_y = 0.2$ [*Ingebritsen et al.*, 1994], we find that the groundwater warms up at the rate of about 1°C per 30 years. We note that borehole temperature measurements in the High Cascades show that temperature in many areas is nearly constant to depths of hundreds of meters [e.g., *Blackwell et al.*, 1990], an observation consistent with the large volumes of circulating groundwater suggested by our model [*Ingebritsen et al.*, 1996]. On the basis of (8), the age of the water discharged at Quinn River, Cultus River, and Browns Creek should be about 50–100 years, and that discharged at the Fall River should be about 100–200 years. These numbers are compatible with age estimates from residence time arguments [e.g. *Rose et al.*, 1996]: assuming a recharge rate of 1 m/year, an aquifer thickness of 500 m, and $S_y = 0.2$ leads to a residence time of 100 years.

The tritium concentration of water samples collected on September 28, 1996, are 8.1, 12.8, 6.2, and 9.9 tritium units (TU) for the Quinn River, Browns Creek, Cultus River, and Fall River, respectively, with an uncertainty of ±2.7 TU on each measurement. These concentrations cannot be unambiguously related to an apparent groundwater age. If the groundwater being discharged consisted entirely of pre-1952 water, we would expect tritium concentrations lower than about 3 TU. However, mixing of water of all ages occurs along the length of the aquifer so that tritium concentrations in the range of 6–12 TU may indicate that some post-1952 water has been mixed with older water.

6.3. Transmissivity

Our model allows us to determine the effective transmissivity (and equivalent permeability if we know h_0) of the young basalts and basaltic andesites which comprise the aquifers. The effective transmissivity is related to d by

$$T = \frac{S_y \omega L^2}{2d^2}. \tag{9}$$

Shown in Figure 12 are estimates of T based on the length scales L shown in Figure 10; for Quinn River, Browns Creek, Cultus River, and Fall River aquifers we find $T = 6.9$, 5.6, 5.6, and 9.4 m^2/s, respectively. In order to extend the range of L we also plot the transmissivity obtained for Snow Creek (see section 5.1).

We did not include a point for the Deschutes River because,

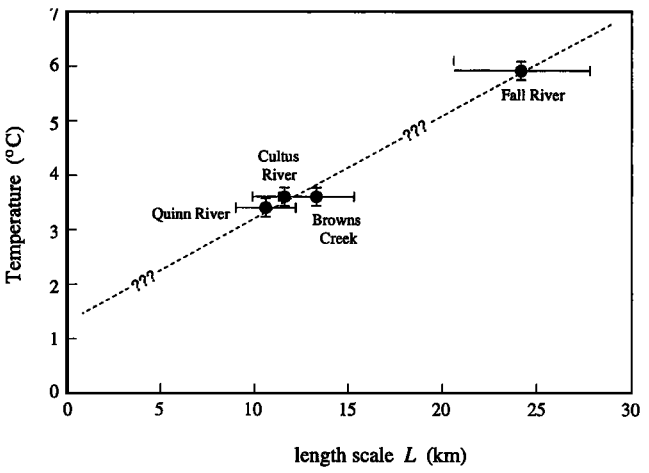

Figure 11. Spring temperature as a function of the length scale of the aquifer, L (see Figure 10).

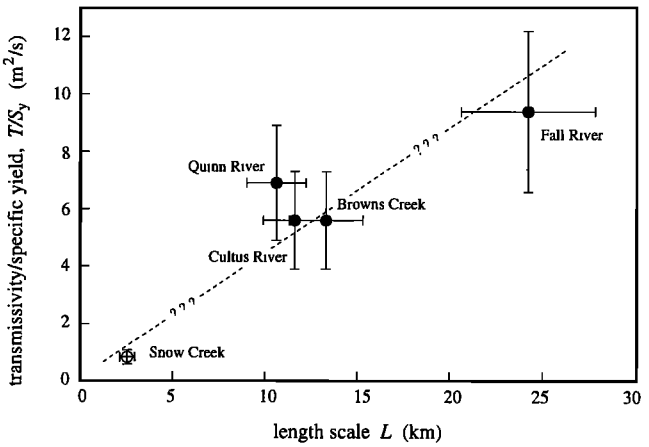

Figure 12. Transmissivity divided by specific yield as a function of the length scale of the aquifer, L (see Figure 10).

as discussed in section 5.2, the hydrology of the system feeding the Deschutes River is relatively complicated and required modifications to the model. However, if we choose L to be the distance between the Lava Lakes and Sparks Lake (see Figure 1), we find $L \approx 10$ km and $T \approx 4.4$ m²/s, and thus would plot close to the points for the Quinn River, Cultus River, and Browns Creek.

Despite the large uncertainties, there is evidence for a scale-dependent effective transmissivity, with T increasing by about a factor of 10 as L increases by a factor of 10. The scale dependence of transmissivity is a well-known phenomenon due to heterogeneity (e.g., review by *Sanchez-Vila et al.* [1996]), and the results shown in Figure 12 may indicate that the scale-dependence of transmissivity persists at the regional scale.

6.4. Groundwater Recharge, Runoff, and Evapotranspiration

The water budget in a watershed can be expressed as

$$P = \mathrm{ET} + R + G + \Delta S \tag{10}$$

where P is precipitation, ET is evapotranspiration, R is surface runoff plus interflow, G is groundwater recharge, and ΔS is the change in storage. Over the 50-year time period considered here, ΔS will make a relatively small contribution to the total water budget. As noted by many authors, ET is the most difficult component to measure directly. For the streams studied here, R can be determined from discharge in runoff-dominated streams, G can be estimated from the discharge in the spring-dominated streams (assuming that the springs are fed by groundwater), and ET can be calculated by $\mathrm{ET} = P - G - R$. We are implicitly assuming that all the groundwater is discharged at springs, which is clearly not the case because at least a small amount of groundwater circulates to greater depths and is discharged at lower elevations [*Ingebritsen et al.*, 1992]. However, the effective permeability of the upper 1–2 km of the Cascades found by *Ingebritsen et al.* [1992] is $O(10^2 - 10^3)$ times smaller than the near-surface permeability [*Manga*, 1996] so that only a small fraction of groundwater circulates to greater depths. Surface runoff can be calculated only for the Quinn River recharge area where surface runoff is measured in Charlton Creek (see Figure 1). However, for the Fall River there are no runoff-dominated streams in the recharge area. For Browns Creek and the Cultus River, runoff-dominated streams enter the spring-dominated streams upstream of the gaging station, and the absence of large and clear runoff signals in the hydrographs shown in Figure 5 indicates that direct runoff makes a small contribution to the overall water budget.

In Table 3 we summarize estimated values for P, R, G, and ET on the basis of basin sizes shown in Figure 10 and the mean annual precipitation map shown in the inset of Figure 1. Groundwater recharge rates are in the range of 66–127 cm/yr (between 40 and 73% of mean annual precipitation), and ET is in the range of 43–98 cm/year (27–60% of mean annual precipitation). These high rates of groundwater recharge are similar to those estimated by *Ingebritsen et al.* [1994] for the Oak Grove Fork Basin, which lies on the western side of the crest of the Cascades, and is composed of <7 Ma andesites. The rate of ET in Table 3 is within the range of ET in other watersheds in the western United States [*Claassen and Halm*, 1996].

7. Concluding Remarks

The one-parameter model presented in this paper is physically based in that it follows from the Boussinesq equation, which is derived from conservation of mass and Darcy's equation. The single parameter can be related to the effective transmissivity of the aquifers that supply groundwater to the springs and can be determined by model calibration.

The model can quantitatively explain several features of spring-dominated streams. For example, the time lag between peak runoff and peak discharge in spring-dominated streams is due to the time required for variations in the height of the water table to propagate through the aquifer, and discharge

Table 3. Estimates of Groundwater Recharge, Runoff, and Evapotranspiration

Name	Recharge Area,[a] km²	Precipitation,[b] cm	Groundwater Recharge, cm	Runoff, cm	Evapotranspiration, cm
Quinn River[c]	33	166	66	4[d]	98
Cultus River[c]	44	175	127	···	48
Browns Creek	60	149	60	···	89
Fall River[e]	202	108	65	···	43

For Cultus River and Browns Creek all runoff-dominated streams enter the spring-dominated stream upstream of the gaging station. As is evident in Figure 4, the runoff contribution makes a small contribution to the total water budget.

[a]Uncertain due to interbasin exchange.
[b]Based on precipitation map shown in Figure 1 and recharge areas shown in Figure 10.
[c]Recharge area similar to that estimated by the USGS.
[d]Based on Charlton Creek (Figure 1).
[e]Recharge area is twice that estimated by the USGS.

variations decrease as the time lag increases (see work by *Manga* [1996] for a detailed discussion). For the streams studied here, annual discharge variations are governed primarily by the preceding winter's precipitation. However, the actual age of the groundwater being discharged is probably in the range of 50–200 years and is governed by the porous flow.

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