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Hydrology of spring-dominated streams in the Oregon Cascades

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Abstract. Spring-dominated streams in the Oregon Cascades are often characterized by nearly constant discharge and by peak flows that occur in late summer or fall, several months after the annual snowmelt. A model is presented that can account for the temporal variations of discharge and the delay between snowmelt and the period of peak streamflow. Springs are assumed to be fed by an unconfined aquifer that is recharged by the annual snowmelt. Model results depend primarily on the effective permeability and the dimensions of the aquifer. Four spring-fed streams in the Deschutes River basin in the Oregon Cascades are studied. The effective permeability of the young (<2 Ma) volcanic rocks that comprise the aquifers is inferred to be $O(10^{-11})$ m^2.

Introduction

Springs are points or small regions where groundwater emerges from an aquifer and forms a stream. Spring hydrographs often exhibit very little variation, even in regions with extended dry seasons or following a period of significant precipitation or snowmelt. In such systems, base flow (as opposed to direct runoff, which consists of both surface flow and interflow) accounts for most of the streamflow. Here we study certain aspects of the hydrology of spring-dominated streams in the upper Deschutes River basin in the Oregon Cascades, where precipitation falls primarily in the form of snow and the dominant source of water is the annual snowmelt [*Whiting and Stamm*, 1995]. The hydrographs of these spring-fed streams, as will be illustrated later, have two distinctive features: (1) nearly constant discharge and (2) peak discharge that typically occurs in the summer or fall, several months after the annual snowmelt. We present a one-dimensional model of flow in an unconfined aquifer recharged by the annual snowmelt that can account for both features. The relationship between precipitation and runoff, or snowmelt and streamflow, is sometimes used to infer hydrological properties. The results presented here suggest that the hydrographs of spring-dominated streams can also provide constraints on aquifer size and effective permeability.

Observations

We consider the daily discharge records of four spring-dominated streams and two runoff-dominated streams. All six streams are located within a few tens of kilometers km from each other, are fed by small drainage basins, and are at elevations of >1200 m. The near-surface geology consists of Quaternary (<2 Ma) volcanic rocks. The streams were monitored by the U.S. Geological Survey (USGS) from 1939 to 1991 and have no diversion or regulation upstream. The location, elevation, estimated size of the drainage basin, and mean annual flow rate are listed in Tables 1 and 2.

In Figure 1 we show hydrograph records for Quinn River and Deer Creek from water year 1990 (October 1, 1989, to September 30, 1990). The spring-dominated Quinn River has a nearly constant discharge, and the peak flow occurs in July. The hydrograph is approximately sinusoidal (but still with a steeper rise than recession), with a period of 1 year. The other three spring-dominated streams have similar hydrographs but exhibit smaller discharge variations. The runoff-dominated streams, such as Deer Creek, have by comparison relatively "flashy" hydrographs and very little base flow: the large discharge in April–May is due to direct runoff following snowmelt.

In order to characterize the spring-fed streams we measure two quantities, which describe (1) the annual variations of the discharge rate and (2) the lag between snowmelt and peak discharge. As illustrated in Figure 1, discharge variations are characterized by the magnitude of annual discharge fluctuations, ΔA, normalized by the mean annual discharge, A. The lag between peak spring flow and snowmelt is described by the time between the peak flow in the spring-dominated stream and the peak flow in the runoff-dominated streams, τ. The date of peak snowmelt is assumed to be the average date of peak flow in Deer Creek and Cultus Creek. For each of the four spring-fed streams we measured $\Delta A/A$ and τ for water years 1986–1991; the results are summarized in Table 3, and the uncertainties are standard deviations. Notice that $\Delta A/A$ decreases as τ increases.

Model

Since most of the precipitation occurs in the form of snow, groundwater is supplied by the spring snowmelt. This annual input of water is assumed to replenish an unconfined aquifer and to feed the springs. We further assume that water infiltrates and percolates relatively quickly from the surface to the top of the water table. The height of the water table will thus increase following the spring snowmelt, and variations in the height of this free surface will drive a flow, as illustrated schematically in Figure 2a.

The evolution of the water table is 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 dimension the height h of the free surface satisfies the nonlinear diffusion-like equation

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Table 1. Characteristics of Selected Spring-Dominated Streams in the Oregon Cascades

Name	Number	Elevation, m	Drainage Basin,[a] km^2	Mean Flow (1939–1991), m^3/s	Location
Cultus River	14050500	1356	43	1.77	43°49′06″N, 121°47′40″W
Quinn River[b]	14052500	1354	undetermined	0.67	43°47′03″N, 121°50′06″W
Brown Creek[c]	14054500	1332	54	1.08	43°42′57″N, 121°48′10″W
Fall River[d]	14057500	1286	117	4.16	43°47′48″N, 121°34′18″W

No regulation or diversions upstream.
[a]Estimated size based on surface topography; may differ from the subsurface drainage area contributing to the springs.
[b]Gage located 45 m downstream from spring.
[c]Geomorphology and hydrology discussed in detail by *Whiting and Stamm* [1995].
[d]Identified as spring-fed in *Water Resources Data, Oregon* [1992].

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

where k, ρ, μ, g, ϕ, t, and x are the permeability, water density, water viscosity, gravitational acceleration, effective porosity, time, and horizontal position, respectively, and N is 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 to be isotropic and homogeneous and to have a horizontal and relatively impermeable base. In many systems the discharge consists of both saturated and unsaturated flow; the neglect of unsaturated flow is a reasonable approximation for the aquifer sizes considered here (thickness $O(100)$ m and horizontal dimensions $O(10)$ km).

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 be neglected, (1) becomes

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

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

In order to provide some insight into the processes which relate $\Delta A/A$ and τ, consider a semi-infinite unconfined aquifer in which the height of the water table at one end varies periodically in time, $h_1 \propto \cos(\omega t)$, in order to simulate the annual snowmelt. The solution to this problem is well known [e.g., *Turcotte and Schubert*, 1982, pp. 389–390]: $h_1 \propto e^{-x/x_c}\cos(\omega t - x/x_c)$, where x_c is the characteristic "diffusive" length scale

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

Assuming discharge is proportional to h_1, we expect that as the time lag $\tau = x/\omega x_c$ increases, discharge variations will decrease, specifically, $\Delta A/A \propto e^{-\tau\omega}$, consistent with the observations in Table 3. A wide range of analytical solutions for $h(x,t)$ for related problems in infinite and semi-infinite aquifers are described by *Polubarinova-Kochina* [1962, chapters 13–15].

Results

Here we present results for simple models based on the linearized Boussinesq equation that illustrate the relationship between $\Delta A/A$ and τ. The aquifer is assumed to be one dimensional and has a finite length L, as illustrated in Figure 2b. At $x=L$ we apply the boundary condition $h=h_0$. Discharge is proportional to $\partial h/\partial x$ at $x=L$.

We set $t'=\omega t$, and $x'=x/x_c$, so that the linearized Boussinesq equation becomes

$$\frac{\partial h_1}{\partial t'}=\frac{1}{2}\frac{\partial^2 h_1}{\partial x'^2}+\frac{N(x',t')}{\omega\phi} \quad \text{with} \quad h=h_0 \quad \text{at} \quad d=L/x_c. \tag{4}$$

Snowmelt input is periodic with form

$$N(x',t')=N_0 e^{-[t'-\mathrm{nint}(t')]^2/\sigma^2} \quad \text{for} \quad 0<x'<L'/x_c, \tag{5}$$

where nint(t') means the nearest integer value of the time t' (i.e. a time in years), and a describes the period over which water enters the aquifer. The Boussinesq equation (4) is solved numerically using a Crank-Nicholson method for different values of d in (4), and σ and L' in (5).

Numerical results are presented in Figure 3 for $L'/L = 0.25$, 0.5, and 0.75 with $\sigma = 0.05$. The observations from Table 3 are shown for comparison. Each point along a curve is the result of a numerical simulation for different values of $d = L/x_c$ and L'/L. The results are largely insensitive to the value of σ, or even the particular functional form of the water input described by (5).

Numerical calculations run until the discharge versus time is "quasi-steady", i.e., until $\Delta A/A$ and τ do not change from one year to the next. Once this quasi-steady state is reached, $\Delta A/A$ and τ also lose sensitivity to details (volume and date of recharge) of previous years' water input; qualitatively, as is observed in Figure 3, $\Delta A/A \ll 1$ for $\tau > 1$ year so that it is

Table 2. Characteristics of Runoff-Dominated Streams Located Near the Spring-Dominated Streams Listed in Table 1

Name	Number	Elevation, m	Drainage Basin, km^2	Mean Flow (1939–1991), m^3/s	Location
Cultus Creek	14051000	1385	86	0.62	43°49′17″N, 121°49′22″W
Deer Creek	14052000	1378	56	0.21	43°48′48″N, 121°50′18″W

No regulation or diversions upstream.

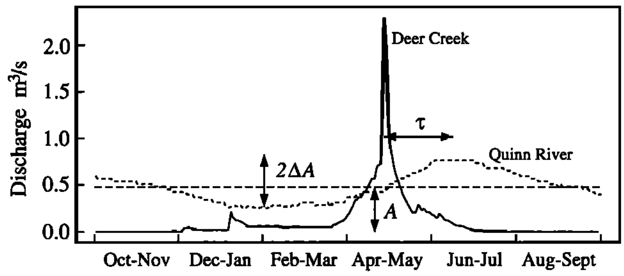

Figure 1. Hydrographs for water year 1990 from Deer Creek (runoff-dominated stream) and Quinn River (spring-dominated stream). Data gathered by the USGS. Spring-dominated stream hydrographs are characterized by the annual amplitude fluctuation $\Delta A/A$ and the time difference between the dates of peak flow in the spring- and runoff-dominated streams, τ. The mean annual flow in the Quinn River in 1990 is about twice as large as the long-term average (see Table 1).

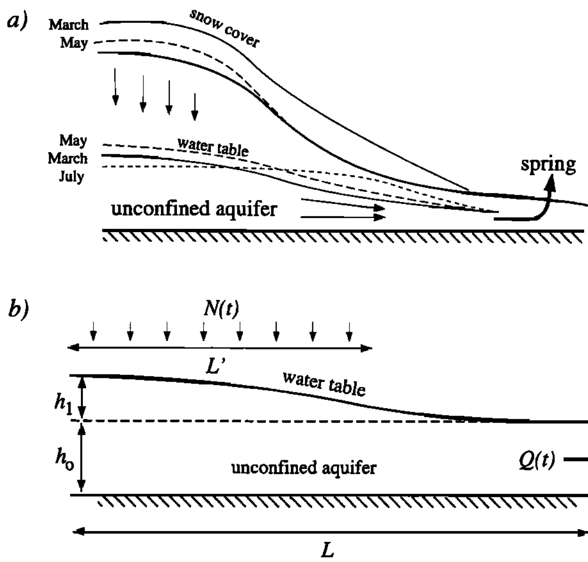

Figure 2. (a) Schematic illustration of the model for discharge into the spring-dominated streams. An unconfined aquifer is fed by the spring snowmelt. The time delay τ in Figure 1 is due to the time required for variations in the height of the water table to move along the aquifer. (b) Model geometry for the unconfined aquifer. Recharge occurs annually over the region $0 < x < L'$; the length of the aquifer is L. Flow is assumed to satisfy the linearized unsteady Boussinesq equation with accretion.

possible to correlate the peak flow in the summer with the preceding snowmelt.

In order to illustrate the model more clearly, in Figure 4 we show the predicted (bold curves) and observed (thin curves) hydrographs for Quinn River. For this calculation, we use the discharge in a runoff-dominated stream (Deer Creek) as a proxy for the time-dependent recharge of the unconfined aquifer. We make this assumption because recharge is related to the same processes of snowmelt and rainfall that result in streamflow in runoff-dominated streams. The two parameters $L'/L = 1.24$ and $d = L/x_c = 0.66$ are determined by minimizing the mean absolute error over the period of October 11, 1947, to October 11, 1951; the rest of the time series in Figure 4 is calculated based on the parameters estimated over the shorter time period. The discharge in the other runoff-dominated stream, Cultus Creek, produces a nearly identical predicted hydrograph. The Quinn River hydrograph is shown in Figure 4 because it has the largest discharge variations and thus the largest discrepancy between predictions and observations.

Other observations are consistent with the model presented here. For example, the mean annual discharge for the four spring-dominated streams exhibits less variation than that in the snowmelt-dominated streams, and streams with the largest τ also have the smallest mean annual discharge variations.

Discussion

The results described here, despite the simplifications and approximations, reproduce the observations at least qualitatively and thus support a model in which an unconfined aquifer is fed by the annual snowmelt. The governing equation resembles a diffusion equation, so that the delay of the peak discharge describes the time required for variations in the height of the water table due to the annual snowmelt to "diffuse" the length of the aquifer.

The "diffusion" process also results in small discharge variations, which in turn affects the geomorphology and riparian habitat associated with the spring-fed streams. *Whiting and Stamm* [1995] studied the geomorphology of spring-fed streams, including one of the streams considered here, and

Table 3. Discharge Characteristics of Spring-Dominated Streams

Name	$\Delta A/A$	τ, days
Quinn River	0.475 ± 0.124	55 ± 11
Cultus River	0.275 ± 0.037	113 ± 28
Brown Creek	0.187 ± 0.038	136 ± 18
Fall River	0.070 ± 0.019	184 ± 55

Average of water years 1986–1991. Parameters are defined in Figure 1 and in the observations section.

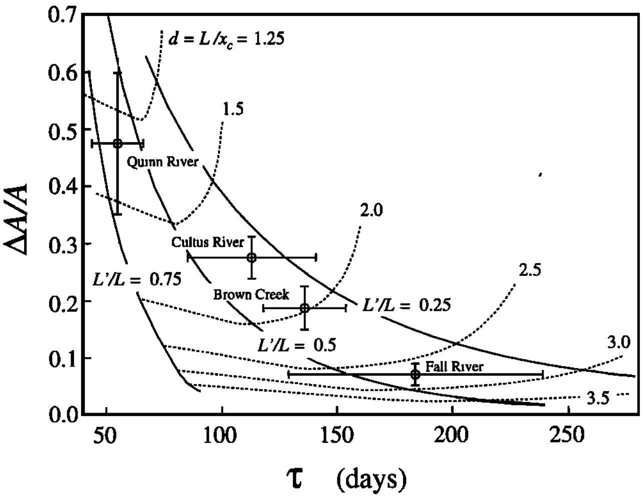

Figure 3. Relationship between $\Delta A/A$ and τ for the model illustrated in Figure 4 compared with observations for the four spring-dominated streams listed in Table 1. The solid curves are based on calculations for fixed L'/L along which d changes; the dashed curves are for fixed d and changing L'/L.

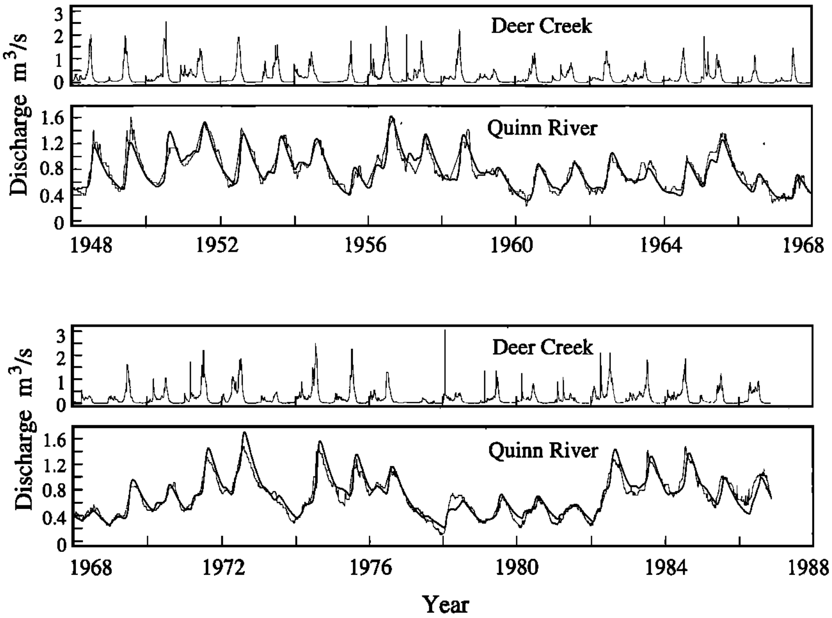

Figure 4. Predicted (bold curves) and observed (thin curves) hydrographs for Quinn River. The hydrograph for Deer Creek, a runoff-dominated stream, is used as the time-dependent recharge of the unconfined aquifer. The two parameters $L'/L = 1.24$ and $d = L/x_c = 0.66$ are determined by minimizing the mean absolute error over the period of October 11, 1947, to October 11, 1951.

concluded that the hydrology of spring-dominated streams results in a distinctive geomorphology. Specifically, as a result of the nearly constant discharge, spring-dominated channels have poorly developed bars, moist and organic-rich floodplains, and channel banks that typically rise steeply out of the water and terminate in a nearly horizontal surface along the channel.

The model depends primarily on two parameters: L'/L and $d = L/x_c$. For comparison, some rainfall-runoff models include more than 100 parameters, although two-component models with four parameters are sufficient to represent well the response of all catchments [*Jakeman and Hornberger*, 1993]. In contrast to some models for relating rainfall and runoff, the parameters here are directly related to the geometry and physical properties (permeability, effective porosity) of the aquifer. The parameter L'/L, which describes the horizontal extent over which water enters the aquifer, simulates a much more complicated source of water whose lateral extent and location varies during the year as snow melts and the snow cover recedes, i.e., $L' = L'(t)$. A value of $L'/L = 0.5$ would seem to be a reasonable average of $L'(t)/L$ and provides a good fit to the observations.

The value of d depends mainly on aquifer porosity and (most importantly) permeability. Comparing observations and model results (Figure 3), we find that the four spring-dominated streams, Cultus River, Quinn River, Brown Creek, and Fall River, are characterized by $d \approx 1.7$, 1.3, 2, and 2.7, respectively. The annual average output of these springs requires an input of $\approx$1 m/year over the entire drainage basin (see Table 1), a volume comparable to the annual precipitation in this region. Comparing the mean flow/basin size ratio for spring-fed streams (Table 1) and runoff-dominated streams (Table 2) suggests that a relatively small fraction of the annual precipitation is discharged as direct runoff and that a substantial fraction replenishes unconfined aquifers and is discharged in spring-dominated streams. Assuming the area of each basin is L^2 (recall we solved only the one-dimensional problem) and a recharge of 1 m/year, we estimate $L \approx 5.6$, 4.6, 5.9, and 11.6 km for the four springs listed above, and thus we obtain $x_c \approx 3.3$, 3.5, 3.0, and 4.4 km. The actual subsurface basin size for the spring-dominated streams, as noted in Table 1 and in the *Water Resources Data* books, is uncertain; however, the resulting uncertainty in L is less than or comparable to uncertainties in h_0, ϕ, and d inferred from Figure 3. Assuming $\phi = 0.01 - 0.15$ [*Ingebritsen et al.*, 1994] and $h_0 = 100$–500 m implies a permeability $k = O(10^{-11})$ m^2. The permeability calculated here is a basin-scale value and characterizes the relationship between τ, $\Delta A/A$, and x_c; studies of groundwater flow through volcanic rocks in mountainous regions indicate that details of precipitation input, aquifer geometry, and hydraulic properties can have a significant effect on spring and stream discharge, at least for flows over small ($O(100)$ m) length scales [e.g., *Ingraham et al.*, 1991; *Flerchinger et al.*, 1992].

The high calculated permeability is typical of young, unaltered volcanic rocks and may be due to a number of features including flow-top breccias, fractures, cooling joints, and porosity due to vesicularity. On the basis of a compilation of field measurements and numerical modeling experiments, *Ingebritsen and Scholl* [1993] found that the permeability of Hawaiian basalts decreases from $\leq 10^{-10}$ m^2 near the surface [*Williams and Soroos*, 1973] to $\leq 10^{-15}$ m^2 at a depth greater than about 500 m. Icelandic lavas have permeabilities of 10^{-9}–10^{-10} m^2 [*Arnorsson*, 1995]. For comparison, *Ingebritsen et al.* [1992] studied hot springs that lie at lower elevations, $\approx$500 m, in the Sisters region north of the springs considered here and found that the permeability of the upper $\approx$1 km of volcanic rocks is $O(10^{-14})$ m^2. The lower permeability calculated there is consistent with the sealing or partial sealing of cracks at higher temperatures and pressures. In fact, high recharge rates indicate a much higher near-surface (depths of <100 m) permeability in the Sisters region [*Ingebritsen et al.*, 1994].

Finally, we note that the annual discharge variations depend on the transport of variations in the height of the water table, which depends on the "diffusivity" of the aquifer, $k\rho g h_0/\mu\phi$, and thus on the thickness h_0 of the aquifer. However, the velocity of the water itself depends only on the horizontal gradient of h and is thus independent of h_0. The results discussed above allow us to estimate x_c which depends on k and h_0; correlating temperature variations and discharge variations [*Bundschuh*, 1993] or measuring the age of the groundwater using isotopic techniques may place additional constraints on h_0 and thus on the geometry of the aquifer.

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