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**Key Points:**
- Computed temperature change at springs depends on changes to land surface temperature and to groundwater recharge temperature
- A new analytic solution can be used to estimate the timing of thermal response to climate change
- For the sample system examined, thermal response time is of the same order as climate change (i.e., tens to hundreds of years)

**Supporting Information:**
- Supporting Information S1

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# Thermal effect of climate change on groundwater-fed ecosystems

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**Abstract** Groundwater temperature changes will lag surface temperature changes from a changing climate. Steady state solutions of the heat-transport equations are used to identify key processes that control the long-term thermal response of springs and other groundwater discharge to climate change, in particular changes in (1) groundwater recharge rate and temperature and (2) land-surface temperature transmitted through the vadose zone. Transient solutions are developed to estimate the time required for new thermal signals to arrive at ecosystems. The solution is applied to the volcanic Medicine Lake highlands, California, USA, and associated springs complexes that host groundwater-dependent ecosystems. In this system, upper basin groundwater temperatures are strongly affected only by recharge conditions. However, as the vadose zone thins away from the highlands, changes in the average annual land-surface temperature also influence groundwater temperatures. Transient response to temperature change depends on both the conductive time scale and the rate at which recharge delivers heat. Most of the thermal response of groundwater at high elevations will occur within 20 years of a shift in recharge temperatures, but the large lower elevation springs will respond more slowly, with about half of the conductive response occurring within the first 20 years and about half of the advective response to higher recharge temperatures occurring in approximately 60 years.

**Plain Language Summary** Tools are developed (and demonstrated) that allow prediction of the effect of climate change on groundwater temperature at springs and seeps that support critical habitat.

## 1. Introduction

The sensitivity of stream temperatures to projected changes in atmospheric conditions (air temperature and precipitation) has been the subject of much recent investigation [*Arismendi et al.*, 2012; *Luce et al.*, 2014; *Snyder et al.*, 2015; *Isaak et al.*, 2016]. Collectively, these statistical studies paint a picture of strong spatial variability in sensitivity and increasing likelihood of nonstationarity in time [*Milly et al.*, 2008; *Hirsch*, 2011]. All of these studies highlight the need for a more mechanistic understanding of relationships between air and stream temperature, particularly with respect to the influence of hyporheic and regional groundwater flow.

Downstream of springs, thermal and stream-discharge regimes influenced by groundwater inputs support species commonly encountered and valued by humans, such as salmon and trout [*Boulton and Hancock*, 2006; *Dunham et al.*, 2008; *Jonsson and Jonsson*, 2009; *Nichols et al.*, 2014; *Penaluna et al.*, 2016]. The distribution of these cold-water species at the southern margins of their geographic ranges is often tied to the presence of strong groundwater influences in streams [e.g., *Koizumi and Maekawa*, 2004; *Shepard et al.*, 2016; *Benjamin et al.*, 2016]. The effects of warming groundwater may lead to shifts in volume and temporal availability of cold-water refuges for salmon and trout, resulting in poorer growth or survival [*Torgersen et al.*, 1999; *Ebersole et al.*, 2001]. A host of lesser-known and unique species occur exclusively within freshwater springs, due to their need for specific ambient conditions, and these species are often represented in lists of species of conservation concern [*Cantonati et al.*, 2012].

Recent advances in quantification of regional groundwater temperatures, developed to better understand geothermal processes [*Burns et al*., 2016], can be used to provide insights into the climate-response of groundwater temperature, an important factor in understanding future habitat of sensitive biota [cf., *Nichols et al*., 2014]. Temperature has been used as a tracer to understand hydrogeologic and geothermal processes [cf., *Smith and Chapman*, 1983; *Forster and Smith*, 1988a,b,; *Manga*, 2001; *Anderson*, 2005; *Saar*, 2011]. Groundwater gains or loses heat by conduction, chemical reactions, and radiogenic decay of isotopes in minerals, and gains heat from friction as mechanical energy is viscously dissipated along the groundwater-flow path [*Manga and Kirchner*, 2004; *Burns et al*., 2016]. Steady state solutions to the energy balance provide a means of estimating the potential effects of climate change [e.g., *Burns et al*., 2016]. When evaluating the time scale over which groundwater temperature will change in natural systems, thermal response can be divided into two components: (1) land-surface temperature effects and (2) recharge effects. Land-surface temperature effects will occur on the same time scale as conduction through the vadose zone. Recharge effects occur on a time scale governed by how rapidly groundwater heats (or cools) the surrounding rock as it flows along groundwater-flow paths.

In this paper, we demonstrate the use of recently developed steady state analytic solutions [*Burns et al*., 2016] and a new transient analytic solution (supporting information to this manuscript) to evaluate climate-driven changes to the temperature of groundwater discharging from regional flow paths. First, the analytic solution methods are described and associated limitations are summarized. Second, we apply the methods to an illustrative example problem and discuss the results. The example is a demonstration of the methods and is not intended to be an exhaustive analysis of the system. In fact, the site-specific study is in its infancy, and application of the analytic methods encapsulates much of our early effort to quantitatively understand heat and groundwater flow in this study area. This exercise demonstrates the use of new analytic methods early in the process of scientific study in order to focus future data collection on important parameters and observations. Finally, we discuss broader implications and conclusions.

## 2. Methods

The steady state groundwater temperature distribution depends on two climate-driven boundary conditions that are variably important as groundwater flows from recharge to discharge areas: (1) land-surface temperature and (2) groundwater recharge rate and temperature. The importance of each boundary condition can be evaluated using the steady state solution of *Burns et al*. [2016], but this solution cannot predict the timing of thermal response. To estimate the time scale of thermal response to each boundary condition, the temperature changes owing to conduction (from land surface) and advection (from recharge areas) are summarized below. The new solution for advection is derived in the supporting information.

### 2.1. Steady State Analysis of Climate-Driven Change

The steady state analytic solution of *Burns et al*. [2016] applies to transport of heat through an aquifer system and is a one-dimensional (1-D) solution that assumes piston flow (i.e., no dispersion or conduction of heat in the direction of groundwater flow) and that the aquifer is well mixed (i.e., negligible vertical gradients in temperature within the aquifer). The physical processes represented are heat conduction through the vadose zone (assumed to be 1-D orthogonal to groundwater flow), prescribed heat flux into the bottom of the aquifer, heat generation due to viscous heating, and advection of heat by flowing groundwater. A Python script allows aquifer system geometry and boundary conditions to vary piecewise (constant or linearly varying across each segment) to represent some 2-D and 3-D complexities of natural aquifer systems [*Burns et al*., 2016]. Properties that can be varied piecewise are recharge rate and temperature, discharge rate (e.g., springs), vadose zone thickness and thermal conductivity, land-surface temperature, heat flux into the bottom of the aquifer, aquifer width (which controls the surface area for heat conduction), and hydraulic head loss (i.e., energy loss to viscous heating).

#### 2.1.1. Limitations of These Steady State Methods

The usual limitations of steady state solutions apply, and *Burns et al*. [2016] discussed the conditions under which a steady state solution for heat transport by groundwater is applicable, but, for the example herein, the steady state solution is used only to identify the important physical mechanisms controlling temperature change and to estimate the magnitude of that change if the system is allowed to reequilibrate.

An additional limitation is the thermally well-mixed aquifer assumption. For the physics considered in the analytic solution, the only mechanism that explicitly depends upon temperature is conduction of heat through the vadose zone, which depends upon temperature at the top of the aquifer. In practice, it is sufficient that mean temperature gradient within the aquifer is small compared with most vadose zone and basal thermal gradients. This requirement may be violated for thick, slow-flowing aquifers, but estimates made with the analytic solution can still be instructive regarding physical processes controlling heat flow.

We do not assert that the 1-D solution is sufficient to capture all of the complex 3-D interactions between groundwater and heat flow for many systems, but *Burns et al*. [2016] demonstrated that the method can efficiently identify the important physical processes within a complex regional aquifer system (the Eastern Snake River Plain aquifer), permitting quantitative estimates, facilitating sensitivity analyses (allowing improved experimental design), and guiding selection of more complex 2-D and 3-D tools for subsequent analyses. Piecewise application of the analytic solution is performed using an open-source Python script that is included in *Burns et al*. [2016, supporting information], and the example problem herein is archived in accordance with U.S. Geological Survey policy and is freely available upon request.

### 2.2. Transient Response to Climate-Driven Change

Whereas the steady state solution of *Burns et al*. [2016] can be used to assess the importance of different hydrologic and thermal controls (including boundary conditions), this tool cannot be used to estimate when the effects of changes will be detectable at different points within a hydrologic system. To estimate time of response, we use a classical solution to the 1-D heat conduction problem and develop a new solution to the heat advection problem.

#### 2.2.1. Heat Conduction Through the Vadose Zone

The propagation of land-surface temperature changes will occur on the time scale of conduction through the vadose zone. The time scale of response resulting from a step change in temperature at land surface ($\Delta T$) is estimated by solving the heat conduction problem for a semi-infinite medium with initially uniform temperature $T_0$:

$$T(z,\ t)=T_0+\Delta T\mathrm{erfc}\left(\frac{z\sqrt{\Gamma_{vz}}}{2\sqrt{\sigma_{vz}t}}\right) \tag{1}$$

where $z$ is depth, $t$ is time, $\sigma_{vz}$ is bulk thermal conductivity of the vadose zone, and $\Gamma_{vz}$ is the heat storage coefficient of the vadose zone, which is given for a three-phase porous media as

$$\Gamma=\rho_{solid}(1-\varphi)c_{solid}+\theta\rho_{water}\varphi c_{water}+(1-\theta)\rho_{air}\varphi c_{air} \tag{2}$$

where $\theta$ is saturation of the pore spaces with water (ranges from 0 to 100%), $\varphi$ is porosity, and $\rho$ and $c$ are the density and specific heat capacity of each phase, respectively. Temperature (as a function of time) at the top of the aquifer is given by equation (1) when $z$ equals vadose zone thickness.

#### 2.2.2. Advection of Heat Along the Groundwater-Flow Path

To estimate how a change in recharge temperature propagates through an aquifer, we must account for heat exchange with the aquifer skeleton and with the overlying vadose zone and the underlying confining unit. Due to this heat exchange, the thermal front will lag the advective front. Local thermal equilibrium is assumed within the aquifer, and the rate of heat exchange with the vadose zone and with the material underlying the aquifer is controlled by the rate of heat conduction. In addition to assuming that the aquifer is well mixed (i.e., no vertical temperature gradients within the aquifer), we assume that conduction and dispersion in the direction of flow is negligible (i.e., piston flow). Because thermal properties can differ between the vadose zone, aquifer, and underlying material, the method of *Zhu et al*. [2016] was used to develop a new analytic solution for how a temperature perturbation ($\Delta T$) at recharge location will move through a groundwater system. The full derivation, which assumes a constant groundwater-flow rate and a step change in temperature at a single recharge location, is provided in the supporting information, and the solution for how temperature evolves within the aquifer as a function of distance along flow path ($s$) and time ($t$) is

$$T_{aqfr}(s,t)=T_0+\Delta T\mathrm{erfc}\left[\frac{\left(\sqrt{\Gamma_{vz}\sigma_{vz}}+\sqrt{\Gamma_{base}\sigma_{base}}\right)s}{2b\sqrt{\rho_{water}q_{water}c_{water}}\sqrt{\rho_{water}q_{water}c_{water}t-\Gamma_{aqfr}s}}\right] \tag{3}$$

where equation (3) is equation (A21a) from the supporting information, converted by recognizing that $v_{water}\varphi_{aqfr}=q_{water}$ is the Darcy flux, $b$ is the thickness of the aquifer, and the subscripts *vz*, *aqfr*, and *base*

indicate properties of the vadose zone, aquifer, and the geologic unit beneath the aquifer (confining unit), respectively.

#### 2.2.3. Limitations of the Transient Solutions

It is not a simple matter of mathematical superposition to use equations (1–3) to estimate the arrival of thermal signals at a groundwater discharge location, so these equations can only estimate thermal arrivals. In real hydrologic systems, vadose zone thickness is variable, and conductive signals will arrive at the aquifer at different times (governed by equation (1)), then be transmitted within the aquifer (governed by equation (3)). Further, recharge is distributed across the landscape, so the advective thermal signal associated with recharge will not begin at a single location as assumed in the derivation. The last significant complication when estimating advective arrival of heat is the lack of accounting for dispersion along flow path (e.g., some water travels via fast flow paths and some via slow flow paths), but this complication is mitigated by the fact that heat conduction is relatively large (compared to chemical diffusion), resulting in more significant equilibration of temperatures across the aquifer thickness compared to other transported constituents. The net result of these limitations is that estimates obtained using equations (1–3) are instructive, but not definitive. Ultimately, 2-D or 3-D models accounting for complex geometry and dispersion may be necessary to refine estimates and assess uncertainty for real hydrologic systems.

### 2.3. Workflow for Analysis of Climate-Driven Change

We use the following workflow to examine the influence of climate-driven change on the temperature of groundwater exiting regional flow systems:

1. Steady state solution to estimate the magnitude of climate-driven temperature changes associated with changes in recharge and land-surface temperature.
2. If temperature changes are insignificant, stop. Otherwise,
3. evaluate the timing of each significant mechanism.

In this way, the principal mechanisms controlling temperature change are identified and future research can be targeted on vulnerable locations and on the most important measurements. The proposed workflow is demonstrated for the example problem.

## 3. Example: Medicine Lake Highlands Hydrologic System

We develop a general model that is applied to the Medicine Lake highlands, California, and an associated large ($\sim$34 $m^3/s$) springs complex that flows into the Pit River via the Fall River [*Davisson and Rose*, 2014]. Cumulatively, springs and groundwater seeps contribute approximately half of the annual average flow ($\sim$140 $m^3/s$) to the Pit River, which is the largest tributary to Shasta Lake, the largest reservoir in California. The tributary groundwater-fed springs provide a relatively cool and drought-resistant source of water for fisheries and agricultural, municipal, and industrial use.

The volcanic terrane of the northwestern U.S. host more than half of the U.S. high-volume ($>$3 $m^3/s$) springs [*Meinzer*, 1927]. This region is home to a variety of biota that depend on groundwater-fed ecosystems, including many threatened and endangered species [cf., *Dunham et al*., 2008; *Nichols et al*., 2014; *Isaak et al*., 2016; *Penaluna et al*., 2016; *Benjamin et al*., 2016]. While the methods developed and illustrated here can be applied to any groundwater-fed system, large-volume springs provide an important and clear example.

### 3.1. Description of the Study Area

The Medicine Lake highlands hydrologic system (Figure 1) is conceptualized (Figure 2) as consisting of an active volcanic center surrounded by heterogeneous radially distributed volcanic deposits that slope away from the center. Even though the volcano is roughly radially symmetric, water budget and isotopic data indicate that almost all of the recharge in the Medicine Lake highlands occurs at high elevation and drains to the south through young lava flows, mainly the Giant Crater Lava [*Davisson and Rose*, 2014]. Precipitation decreases from $\sim$100 cm/yr in the west to $\sim$60 cm/yr in the east. There are few surface water features and no surface water drainages on the highlands, indicating that evapotranspiration and groundwater flow are the only hydrologic outlets. Rubbly, coarse lava flows in the highlands allow a very large fraction of precipitation to recharge groundwater, much of this recharge occurring during snow melt. Vadose zone temperatures at the volcano are highly variable as a result of hydrothermal heating of groundwater, but the lowest deep vadose zone temperatures represent groundwater recharge temperature prior to heating

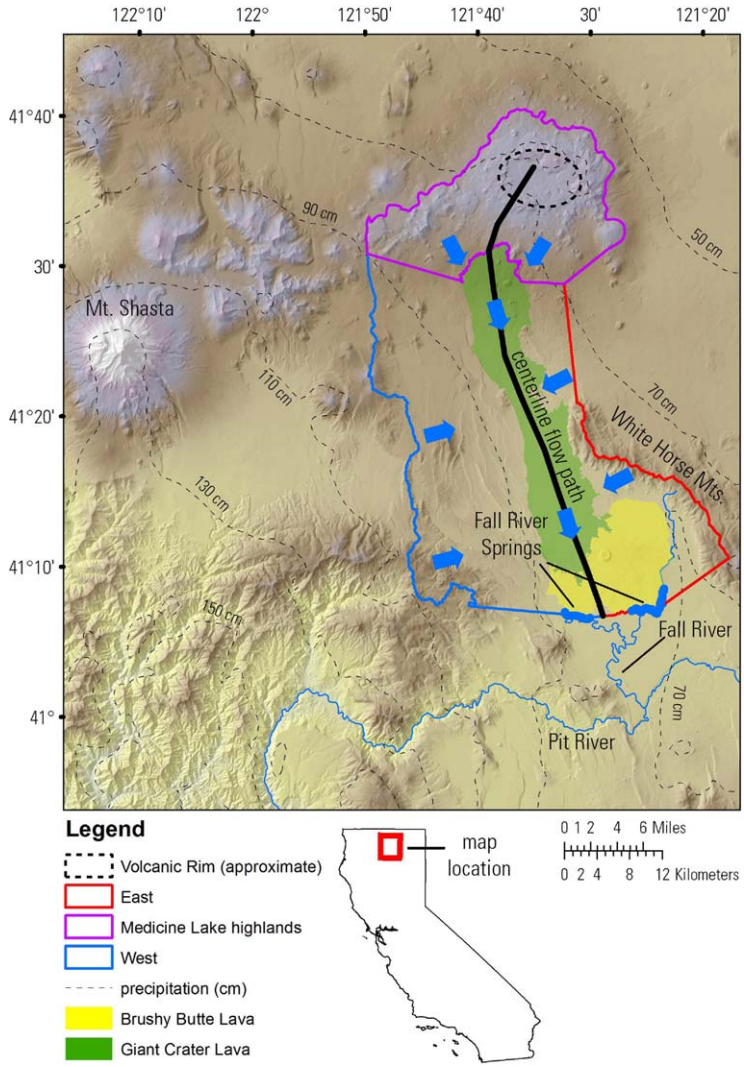


**Figure 1.** Groundwater accumulates and is rapidly transmitted to the Fall River Springs system through the permeable Giant Crater and Brushy Butte Lavas. Precipitation contours are annual average values (1980–2013) estimated using the 1 km gridded DayMet data set [*Thornton et al.*, 1997, 2016]. One-dimensional simulations, presented in this manuscript, are for the centerline flow path. Medicine Lake highlands recharge is simulated as directly contributing groundwater along the centerline flow path, and one-dimensional models for the east and west sides were used to estimate the temperature of groundwater entering the centerline flow path from the surrounding uplands. Groundwater contributions per unit length of the centerline flow path were estimated as proportional to precipitation in each of the three regions, so that total flow = 34 $m^3/s$ [*Davisson and Rose*, 2014].

(~3.0 ± 0.1°C, logged by USGS October 1998). This temperature agrees well with the temperatures measured at the coldest springs (~2.6°C) within the volcanic crater [*Schneider and McFarland*, 1996]. Spring flows in the vicinity of Fall River Springs (Figure 1) average ~34 $m^3/s$ of ~12°C water [*Waring*, 1915; *Davisson and Rose*, 2014]. The uppermost isothermal zones measured in geothermal boreholes represent flow to the water table, so may be used to infer that the vadose zone thickness near the volcano is hundreds of meters (assumed for our analyses to be ~200 m). Away from the volcano, groundwater is likely transmitted through the thin permeable base of the Giant Crater Lava and Brushy Butte Lava flows, implying a vadose zone thickness similar to lava flow thickness. In addition to contributions from the Medicine Lake highlands, these young lavas accumulate groundwater flow from the adjacent eastern and western highlands (Figure 1). While the lava flow thickness is variable along flow (thinning at the distal margin) and across flow (thickest near the center where the lava flow followed topography), it is assumed here that representative vadose thickness is ~20 m, based on estimated lava thickness in the range 10–30 m [*Donnelly-Nolan*, 1991]. Geothermal heat flow into the base of the aquifer is assumed to be higher near the volcano and lower away from the volcano, and these heat flows are varied during the analysis to match typical measured groundwater temperatures.

The analysis of the Medicine Lake system assumes three precipitation zones (East, West, and Medicine Lake highlands; Figure 1) that accumulate water along the centerline flow path (Figure 2). The total recharge of 34 $m^3/s$ was divided into the three zones, proportional to 1980–2013 average precipitation for each zone, estimated from the DayMet 1 km gridded daily precipitation [*Thornton et al.*, 1997, 2016]. The centerline flow path (starting in the Medicine Lake highland zone) model geometry is shown schematically in Figure 2, and the East and West geometries (not shown) are modeled as having uniform conditions. Output from the East and West simulations were then used as input to the centerline model after being adjusted for an

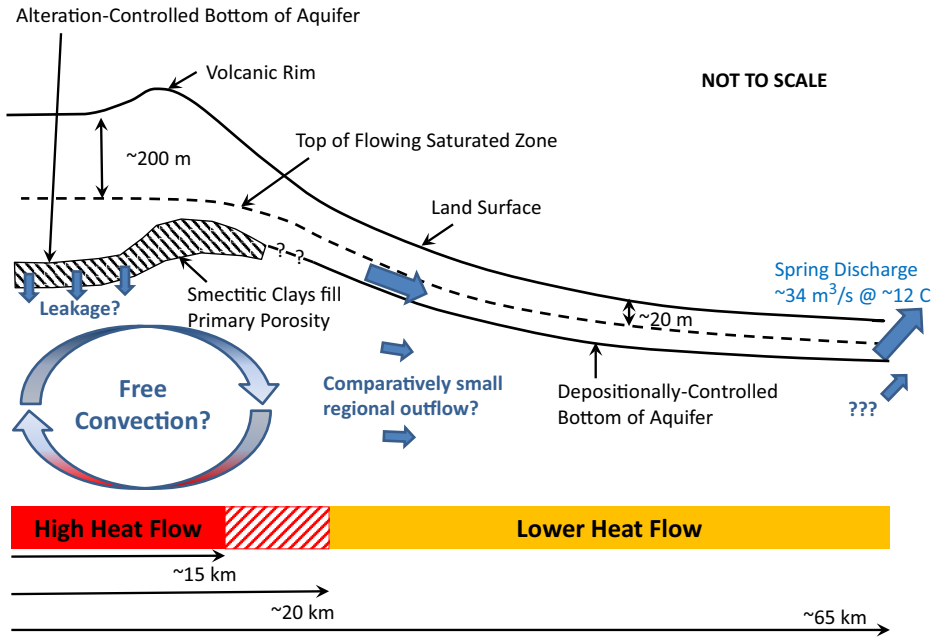


**Figure 2.** The Medicine Lake hydrologic system along the centerline flow path (Figure 1) can be conceptualized as consisting of two regions: (1) much of the recharge occurs in the vicinity of the high-elevation volcano, where it flows through a thick unsaturated zone and is subjected to high heat flow from the volcano and (2) groundwater leaving the uplands, from the volcano and the highlands to the west, is transmitted by very permeable, relatively thin young lava flows. Along the length of these lava flows, groundwater accumulates from the adjacent eastern and western uplands (Figure 1). The bottom of the aquifer is shown schematically to indicate that it generally follows the topography, but it is more complex than shown, especially in the vicinity of the volcano. Knowledge of aquifer bottom position is not necessary for the analytic solutions used and developed here. Question marks show uncertainty in deep flow paths for advective transport of heat (groundwater flow is much smaller than in the aquifer), which translates into uncertainty in deep heat flow pattern below the aquifer.

adiabatic gradient along the centerline flow path. For the Medicine Lake highlands, recharge of 7 $m^3$/s at 3°C (~4–5°C lower than average annual temperature, reflecting significant snowmelt) was distributed evenly across the first 15 km (i.e., the volcanic crater and rim). Eastern recharge of 10 $m^3$/s and western recharge of 17 $m^3$/s was assumed to enter uniformly along the ridgeline. Because no midelevation recharge temperatures were available, the mean temperature of upper basin springs (7°C) [*Manga and Kirchner*, 2004] was assumed for the east and west cross-sectional models, consistent with lower snow fraction at these lower elevations. Simulated temperature from these cross-sectional models was used as input to the centerline flow path model. The geometry of the centerline flow path model is shown in Figure 2. The eastern flow path was assumed to be 8 km long from ridge to centerline on average, extending along the last 40 km of the centerline flow path, and the western flow path was assumed to be 15 km long, extending along the last 45 km of the centerline flow path. Both eastern and western models assumed a uniform vadose zone thickness of 30 m. Vadose zone thermal conductivity varies as a function of air and water content, but a uniform value of 1.6 W/m/°K is assumed for all flow paths, consistent with estimates for other volcanic terrains in the northwestern U.S. [e.g., *Lachenbruch and Sass*, 1977; *Brott et al.*, 1981; *Burns et al.*, 2015]. Land-surface temperatures were estimated by averaging the DayMet gridded daily minimum and maximum temperatures for the period 1980–2013 [*Thornton et al.*, 1997, 2016], and to account for the influence of increasing mean air temperature (in the DayMet data) along the length of Giant Crater Lava (~0.1°C/km); the temperatures of groundwater contributions from the east and west were increased at the same rate.

The head distribution (used to compute heat generation from viscous losses) was assumed to follow topography, so was estimated using a digital elevation map. An exception is the head gradient within the young lava flow aquifer (distance 25–65 km) because, assuming that transmissivity of the lava flow is approximately uniform along its length, steady state conditions require that the gradient steepen as flow accumulates from the west and east (Figure 1). Assuming that flow linearly increases with distance over the last 40 km and that transmissivity and lava width are constant, hydraulic head can be shown to be

$$h(x)=h_0+\frac{(h_f-h_0)\left(\frac{\Delta Q(x-x_0)^2}{2(x_f-x_0)}+Q_0(x-x_0)\right)}{\left(\frac{\Delta Q(x_f-x_0)^2}{2(x_f-x_0)}+Q_0(x_f-x_0)\right)} \tag{4}$$

where $x$ is the distance from $x_0$ (=25 km for the example), $h_0$ is the estimated head at $x_0$, $h_f$ is the head at Fall River Springs ($x_f$ = 65 km), $Q_0$ is the groundwater flow at $x_0$ (7 $m^3$/s), and $\Delta Q$ is the total flow added uniformly by the eastern and western flow paths (27 $m^3$/s).

The remaining input for the analysis is geothermal heat flow into the base of the aquifers. There are no thermal gradient or heat flow estimates along the groundwater-flow path shown in Figure 1 except for 24 thermal profiles from geothermal exploration holes drilled within 10 km of the center of the volcano (highly variable heat flow, ranging from 0.1 to nearly 2.0 W/$m^2$, based on available USGS temperature logs (not shown)). Average simulated geothermal heat flow was adjusted to match observed groundwater temperatures.

### 3.2. Results

Figure 3 shows the equilibrium effect of a +1.5°C rise in average annual land-surface temperature and a +3.0°C rise in recharge temperature on the Medicine Lake highlands and +2.0°C rise in recharge temperature elsewhere (solution in *Burns et al*. [2016]). The value of +1.5°C rise in land-surface temperature corresponds to the +1.5°C rise in average annual air temperature projected near the Medicine Lake highlands. While high-emission scenarios predict eventual temperature rises >+2.5°C, +1.5°C is expected by the mid-21st century under relatively low emissions scenarios [*Kunkel et al*., 2013; *Rupp et al*., 2017]. Further, many of these western U.S. climate scenarios indicate that while total precipitation will not change substantially (<10%), more precipitation will occur as rain rather than snow. For demonstration purposes, we assume that net recharge does not change, and that the shift from snowmelt-dominated recharge on the highlands, coupled with changes in air temperature, will result in a +3.0°C rise in recharge temperature. Recharge temperature is affected by a variety of processes, including snow insulation, latent heat of melting, snow-affected land-surface albedo, and timing of precipitation and, as a result, a priori estimation of recharge temperature change is challenging. Because of the controlling importance of recharge temperature on groundwater discharge temperatures, further research is needed. The +3.0°C temperature rise selected herein still leaves recharge temperature less than average annual temperature, consistent with the winter

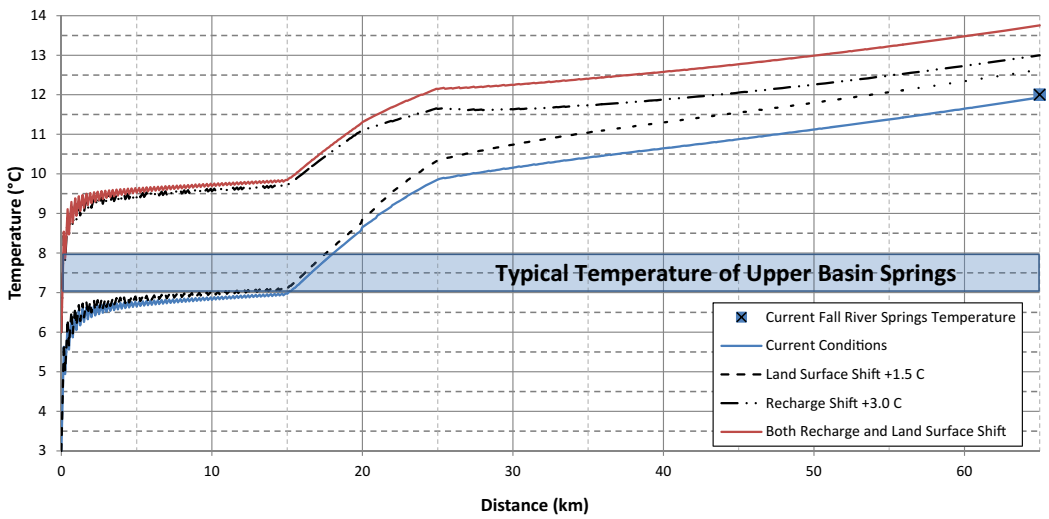


**Figure 3.** Simulated steady state current and future groundwater temperatures along the shallow-groundwater-flow path shown in Figure 2. Lines show current conditions (blue), conditions that would result from a +1.5°C shift in average annual land-surface temperature (dash), conditions that would result from a +3.0°C shift in recharge temperature on Medicine Lake highlands coupled with a +2.0°C shift in recharge temperature at all other locations (dash-dot), and conditions that would result from both a shift both in land-surface and recharge temperatures (red). While there are no upper basin springs on the Medicine Lake highlands, adjacent volcanoes have springs at similar elevations and positions on the slopes. These upper basin springs will be strongly affected only by changes in recharge temperature, but temperatures at lower springs, such as Fall River Springs, will be affected by changes in both land-surface and recharge temperatures.

and spring recharge pattern of the Pacific Northwest. Because there is less snow on the lower elevation ridges, the rise in recharge temperature there was limited to +2.0°C.

The baseline simulation (blue line in Figure 3) was constructed using conditions summarized above and adjusting heat flow in the upper and lower zones until measured temperatures were matched. Final values of simulated geothermal heat flow of 0.3 W/m$^2$ at distances <15 km (i.e., within the volcanic rim), 0.25 W/m$^2$ for distances of 15–20 km, and 0.2 W/m$^2$ for all other regions (including the east and west cross sections) provided a good match (Figure 3, current conditions) to Fall River Springs temperature (12°C) and to typical upper basin spring temperatures across the region (7–8°C) [cf., *Davisson and Rose*, 2014]. Examination of available USGS temperature logs (not shown) indicates that ~0.5 W/m$^2$ is a reasonable estimate within the rim of the volcano, but the highlands capture a significant amount of water outside the volcano (Figure 1), making our estimate of 0.3 W/m$^2$ at distances <15 km reasonable. No data are available to directly support the more distal estimate of 0.2 W/m$^2$, but published estimates of heat flow for the western U.S. frequently show heat flow >0.1 W/m$^2$ in the vicinity of active volcanoes [*Williams and DeAngelo*, 2008, 2011]. These heat flow estimates are of interest for geothermal studies and are similar to regional heat flows elsewhere in the Cascades [e.g., *Ingebritsen et al.*, 1989; *Blackwell et al.*, 1990] and lower than heat flow estimates for large, vigorous volcanic centers such as Yellowstone [*Morgan et al.*, 1977] and Taupo [*Hochstein*, 1995]. While understanding heat and groundwater flow in Medicine Lake volcanic area is an area of ongoing research, we conclude that this conceptualization of the system is reasonable, so may be used to evaluate system response to climate change. In order to evaluate the timing and response of climate-driven boundary conditions, it is only necessary that the estimates are reasonable.

Upper basin groundwater temperatures are predicted to increase ~3.0°C, almost entirely as a result of changes in recharge conditions (Figure 3), indicating that groundwater discharging from upper basin spring complexes will respond strongly to changes in recharge temperature. Changes in average annual land-surface temperature will have little effect on upper basin springs, so can likely be ignored when considering the timing of system response. Most temperature changes for the upper basin are predicted to occur within the first 20 years (Figure 4). Figure 4 shows that the conductive time scale for the highlands (vadose zone thickness = 200 m) is very long compared with other time scales of interest.

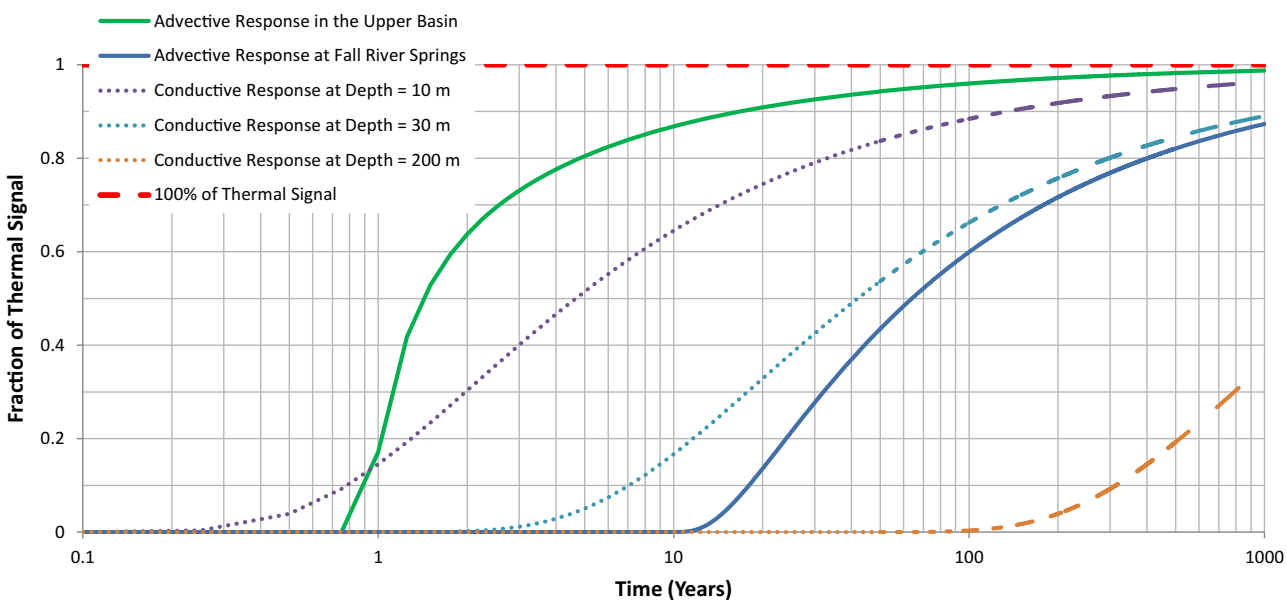


**Figure 4.** Thermal signal arrival, estimated using equations (1)–(3), can be used to evaluate the time scale of response for different processes shown in Figure 2, with the magnitude of final response due to each mechanism shown in Figure 3. Advective response in the upper basin is the signal that will be seen ~5 km along the groundwater-flow path away from the Medicine Lake highlands, so is an estimate of how long it would take for changes in recharge temperature to be seen at the typical upper basin springs in the northern California Cascades examined by *Davisson and Rose* [2014]. Fall River Springs has a longer time scale of response to changes in recharge temperature. Conductive response to land-surface temperature changes depend on the thickness of the vadose zone. Conductive arrival times are shown for the estimated variation in Giant Crater Lava Flow (Figure 1) thickness (10–30 m) [*Donnelly-Nolan*, 1991] and, for comparison, an estimated vadose thickness (~200 m) near the Medicine Lake highlands. At Fall River Springs, conductive signals through the relatively thin vadose zone will precede the advective signals associated with changes in recharge on the highlands.

The temperature of the Fall River Springs is predicted to increase by ~1.75°C and is also strongly affected by changes in recharge conditions, but with changes in land-surface temperature having a similarly large effect at this distance (Figure 3). About half of the recharge-driven temperature change (on the order of 1.0°C) will occur in the first 60 years, with little change predicted during the first 15 years. Because the thickness of the vadose zone at distances >20 km is variable and uncertain, the conductive time scale is shown for both 10 and 30 m thickness. Comparison of the conductive time scale with the advective time scale shows that most of the land-surface temperature changes will be transmitted to the aquifer within 50 years, after which time these thermal signals (on the order of 0.75°C) still need to be carried to the springs advectively.

Care should be taken when interpreting our results and when applying the transient solutions (equations (1–3)), because these solutions assume piston flow and neglect hydrodynamic dispersion (e.g., fast and slow groundwater-flow paths associated with geologic heterogeneity). However, because thermal conductivity is sufficiently large to result in substantial thermal equilibration between fast and slow flow paths on the time scale of heat conduction out of the aquifer, the limitations of the assumptions are not overly restrictive, making travel time estimates sufficiently robust for planning purposes (e.g., thermal response monitoring or land management).

## 4. Discussion

We have developed and demonstrated methods to translate climate-science results into practical estimates that will help ecologists and land managers to identify vulnerable groundwater-dependent ecosystems, develop monitoring programs, and plan responses to changes in available habitat. The methods identify distance from cold-water recharge areas as an important control on whether changes in recharge rate and temperature or changes in land-surface temperature will dominate groundwater temperature changes. Subsequent uncertainty analysis would have two benefits: the robustness of the estimates could be assessed subject to the limitations in available data, and effort could be devoted to narrowing the uncertainty of parameters that have the largest effect on predicted discharge temperatures. Because the purpose of this manuscript is to illustrate new methods, and not to advocate for a specific method of assessing uncertainty, no uncertainty analysis was performed for the example problem.

While the system being examined is special in that it involves very high flow rates through a volcanic aquifer, the methods used here can be used to examine a variety of systems. At a minimum, methods are applicable to high flow rate systems in volcanic and karst terrains, i.e., systems that have high-volume springs [e.g., *Meinzer*, 1927]. The major limitation to applying the tool to other systems is the requirement for a thermally well-mixed aquifer characterized by small temperature gradients relative to the gradients above and below. However, this does not preclude the use of the model to get a basic understanding of the distance over which recharge temperature strongly influences spring temperatures (see *Burns et al*. [2016] for an expanded discussion).

Though absolute temperatures within the example system do not exceed those required by many salmon and trout in the region, projected future increases in temperature may have adverse consequences for native cold-water taxa [*Benjamin et al*., 2016], including increased probability of mortality associated with pathogens [e.g., *Ray et al*., 2012]. Such effects are likely magnified in a downstream direction as the distance from cold groundwater inputs increases and surface water temperatures increase. From an energy-balance perspective, cool groundwater creates a volume of cool habitat, and even if groundwater-flow rate does not change, an increase in temperature reduces the size of the habitat, because that volume is a flow-weighted averaging of the temperature.

We have demonstrated that spring temperatures are buffered from rapid responses to changing surface conditions because of the length of groundwater-flow paths, because the vadose zone is a thermal insulator, and because of the heat capacity of the aquifer and surrounding geology. Although groundwater temperature changes slowly relative to the time scales typical of most ecology and biology studies cited herein [cf., *Isaak et al*., 2016; *Kovach et al*., 2016], our analysis demonstrates that, on time scales of decades, climate change may alter conditions that sensitive species currently rely upon. In order to aid land managers, future research needs to account for anticipated changes in groundwater conditions, particularly in areas where biota are dependent upon groundwater inputs.

For the example problem herein, recharge rate was not varied, consistent with many climate projections for this locality. However, the methods employed in the analysis remain valid for a change in recharge rate. Because recharge rate is an input to the analytic solutions, the steady state analysis allows variation of the recharge rate and estimation of the difference in temperature that would ultimately occur. The method of estimating time scale of response is still valid, provided that the new recharge rate is used in the estimate, i.e., equation (3).

Our analysis and corresponding assumptions highlight some future research needs, including the following:

1. Trends in precipitation under future climate scenarios were used to infer trends in recharge. However, changes in timing and type of precipitation (e.g., transition from snow to rain) could change the relationship between precipitation and recharge.
2. Similarly, a transition from snowmelt-dominated to rainfall-dominated recharge could cause a shift in recharge temperature that is not directly correlated to the shift in average annual temperature. Recharge temperature is affected by a variety of processes, including snow insulation, latent heat of melting, snow-affected land-surface albedo, and timing of precipitation, so that estimation of recharge temperature changes is challenging.
3. Better knowledge of vadose zone thickness would improve estimates of the magnitude and timing of thermal response, as would constraints on regional heat flow.

## 5. Access to Data

In accordance with American Geophysical Union data policy, all data used for the analysis contained herein are publicly available from the cited source. The model used to generate Figure 3 is archived and maintained in accordance with USGS policy, is publicly available, and can be obtained by contacting the USGS Oregon Water Science Center. All other data are documented within this manuscript.

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