THE EFFECT OF BASIN SCALE ON DIURNAL STREAMFLOW TIMING

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ABSTRACT

Hourly streamflow timing, as revealed by diurnal fluctuations in discharge in snowfed watersheds, provides a new tool for understanding transport times and processes in river basins. Travel time delays at different basin scales were measured in nested subbasins (6 to 775 km^2) of the Tuolumne River in Yosemite National Park throughout the spring 2002 and 2003 melt seasons. The travel time increases with longer percolation times through deeper snowpacks, increases with longer travel times overland and along longer stream channels, and increases with slower in-stream flow velocities. In basins smaller than 30 km^2, snow properties that determine the travel times through the snowpack dominate streamflow timing. In particular, daily peak flows shift to earlier in the day as the snowpack thins and mean discharge increases. In basins larger than 150 km^2, snowpack heterogeneity and mixing cause the hour of peak flow to be remarkably consistent, with little or no variation due to snowpack properties. Basins with areas in between 30 and 150 km^2 exhibit different characteristics in different years, illustrating the transition between small and large-scale basin characteristics. Increasing channel travel times as the snowline retreats to higher elevations are not enough to offset the observed decrease in mean snowpack travel times. However, the observed patterns can be reproduced by a model that couples porous-medium flow through evolving snowpacks and free-surface flow in stream channels, provided that the model incorporates snowpack heterogeneity.

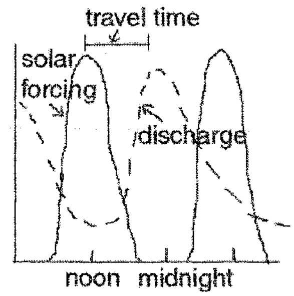

Figure 1. Travel time can be measured as the difference between the time of peak solar forcing (solid line, peak at noon) and the time of peak discharge (dashed line).

INTRODUCTION

Diurnal fluctuations in river discharge in snowfed watersheds, as revealed by hourly streamflow measurements, provide a new tool for understanding basin hydrology. The diurnal cycle of radiative forcing yields major changes in snowmelt and streamflow over the course of each day, and the difference between the time of peak forcing and the time of peak runoff provides a measure of average runoff travel times through the river basin (Fig 1). Timing provides information on snowpack properties, channel velocities, and distances traveled. An understanding of travel times is useful for flood forecasting, reservoir and hydropower operations, and characterizing and predicting chemical processes.

Most previous studies of short-term streamflow timing have focused on small basins and local processes (Bengtsson, 1981; Caine, 1992; Colbeck, 1972; Dunne et al., 1976; Jordan, 1983; Woo and Slaymaker, 1975). Braun and Slaymaker (1981) examined changes in diurnal runoff at different spatial scales, but their largest basin, Miller Creek, encompassed only 23 km^2. Kobayshi and Motoyama (1985) examined streamflow timing in four adjacent watersheds from 0.4 to 108 km^2, and all were completely snow covered throughout the period of study. All of these studies report that travel times decrease as the snowpack thins and matures, reflecting shorter travel times for melt-water to pass from the snow surface to the base of the snowpack.

Basin-scale runoff travel times cease to be dominated by snowpack properties as the basins considered become larger. Grover and Harrington (1966) state that, below the snowfield, peaks and troughs of the diurnal

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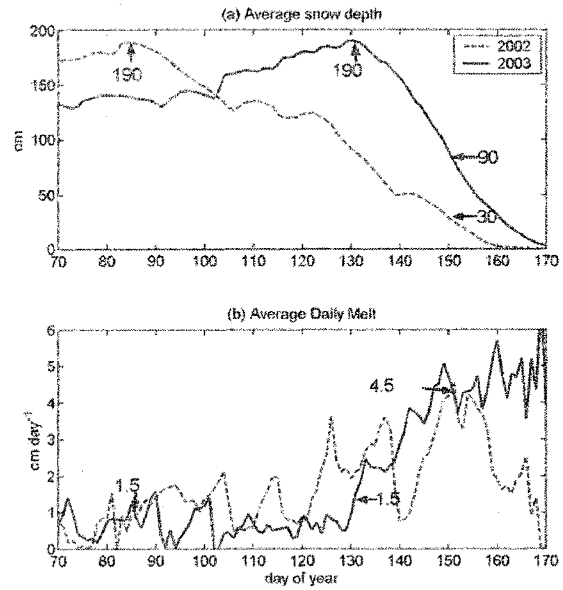

Figure 2. (a) Average snow depth vs. day of the year for 2002 and 2003. Arrows indicate average depth at the start of the melt season (time of peak snow accumulation) and on the day of peak discharge (30 May, day 150, shown in Figure 3) (b) Average daily melt vs. time for the two years. Arrows indicate melt rates at the start and peak of the melt season.

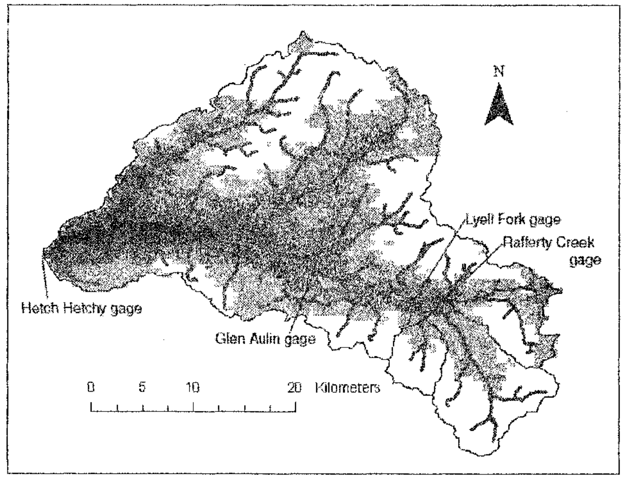

Figure 3. Map of retreating snow cover for the Tuolumne River basin above Hetch Hetchy. Darkest shade represents area that was snow free from 9 May 2003 on, next darkest from 17 May, next from 25 May, and lightest gray becomes snow free by 2 June. White areas are still snow-covered on 2 June, 2003.

cycle occur later than at the edge of a snowfield, with a delay that depends on the in-stream distance from the snowfield and on the stream's velocity. Lundquist and Cayan (2002) show that, during the peak melt season, most USGS-gauged rivers in the western United States have clear diurnal cycles but little or no consistency in the season evolutions of the hour of peak flow on month-long time-scales. Lundquist and Dettinger (submitted) show that heterogeneity in snowpack properties can lead to nearly constant daily streamflow timing, with no change in the time of peak flow over the first half of the melt season.

The interaction of small catchments (<30 km^2) that sum to form larger (>150 km^2) basins, however, has not previously been described. This paper presents an observational study of hourly streamflow timing within nested basins in the Tuolumne River above Hetch Hetchy, Yosemite National Park, California. The study shows how changing channel distance and stream velocities change travel times at different basin scales (as suggested by Grover and Harrington, 1966) and tests the relative importance of heterogeneous snow properties and channel distances in determining streamflow timing (as suggested by Lundquist and Dettinger, submitted). Section 2 describes snow and streamflow observations for 2002 and 2003. Section 3 uses the observations to estimate travel times within the basin throughout the snowmelt season. Section 4 employs a simple numerical model to show that changing in-channel travel velocities and distances are not sufficient to explain the observed timing changes. Rather, heterogeneity in snowpack properties is required. Section 5 summarizes the results and discusses applications.

YOSEMITE INTENSIVE STUDY: OBSERVED PATTERNS IN NESTED SUBBASINS

Overview

Over half of California's water supply comes from the high-elevation snowpacks of the Sierra Nevada. California has a Mediterranean climate, and most precipitation falls between November and March. Thus, the snow provides a natural reservoir, releasing winter precipitation in the summer when the population needs it most. Climate variability in the region is high, and annual precipitation and runoff fluctuate from under 50% to over 200% of climatological averages. Natural climate fluctuations, global warming, and the growing needs of water consumers demand intelligent management of this water resource. Unfortunately, because of complex terrains and limited access, relatively few high-altitude measurements exist. In order to improve understanding of hydrologic processes at high altitudes, a network of meteorological and hydrological sensors has been deployed in Yosemite National Park from Summer 2001 to present.

Snow Observations

Snow observations identify where, when, and how fast snow melts in the region, and these factors play a large role in streamflow timing. Measurements of daily snow water equivalent (SWE) were obtained at 47 snow pillows throughout the central Sierra Nevada (described and mapped in Lundquist et al., 2004). Based on average measurements from 1 April snow course samples in the area, density is assumed to be a constant 400 kg m^{-3}, and snow depth at each station is calculated from SWE by dividing by 0.40. Daily melt rates are calculated from the difference in SWE at each station from one day to the next.

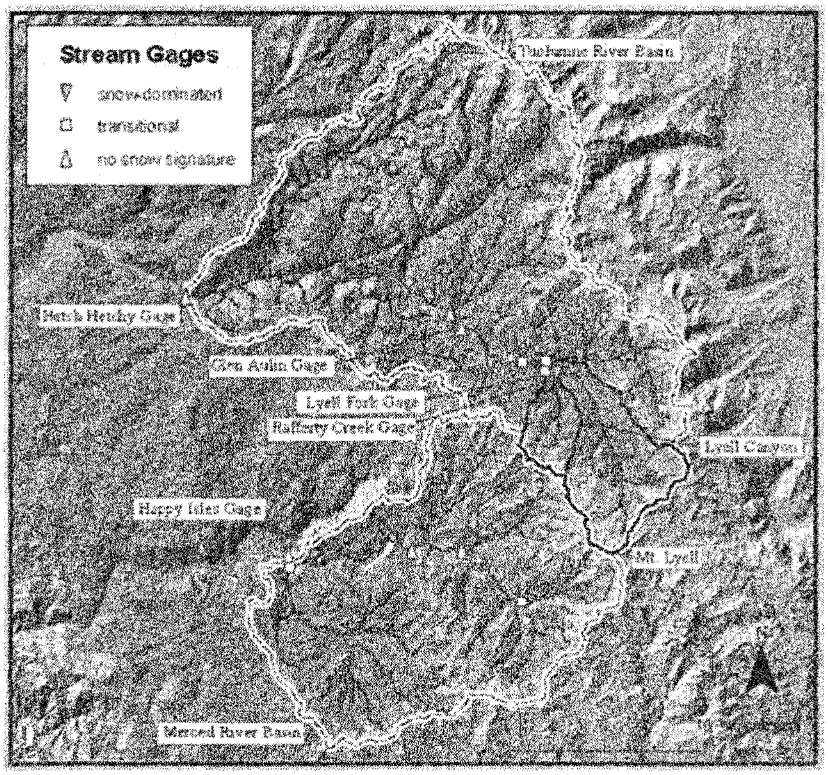

Figure 4. Map of stream gauges in intensive study area in Yosemite National Park. Gages marked with ▽ (a downward pointing triangle) indicate sites where snow depth strongly influences streamflow timing, and travel times decrease as the snowpack thins each year. Gages marked with □ (a box) indicates sites where snow depth appeared to influence streamflow timing in 2002 but not 2003. Gages marked with Δ (an upward pointing triangle) indicate sites where streamflow timing varies less than $\pm$ 1 hour during the first half of the melt season, showing no influence from changing snow depths.

In 2002, spring snowmelt began simultaneously at all elevations on 30 March (Lundquist et al., 2004). The initial melt pulse was followed by several spring snowfalls, which brought snow to areas where it had previously been depleted. The average rate of decrease in snow depth from melt initiation (30 March, day 89) to peak melt (30 May, day 150) was 2.6 cm day^{-1}. In 2003, extensive April snowstorms delayed melt until 10 May. At that point, melt occurred very rapidly, with no interrupting snowstorms, and average snow depth decreased about 100 cm in 20 days, for an average rate of 5 cm day^{-1} (Fig. 2).

Maps of snow-covered area for the 2002 and 2003 melt seasons were obtained from the MOD10A2 data product with a 500-m spatial resolution and 8-day temporal resolution (Hall et al., 2000) and were projected over the DEM (Fig. 3). On 18 June 2003, most of the snow in the Northern areas of the Tuolumne watershed was gone, so most melt originates from the headwaters of the Lyell Fork of the Tuolumne River. By 26 June 2003, only Mt. Lyell appeared snow-covered in the satellite image.

Yosemite Stream Network

The dynamics of streamflow timing at different spatial scales are well illustrated by nested subbasins. For the 2002 and 2003 melt seasons, numerous subbasins (Fig. 4) of the Tuolumne and Merced Rivers were monitored with pressure sensors (Solinst Leveloggers), which provide river stage measurements at half-hour intervals. Basins were delineated with a 30-m resolution digital elevation model (DEM). Basin areas range from 6 to 775 km^2, and elevations range from 1,200 to 3,700 m.

During the first half of the melt season, from the day flows start rising up until the day of peak flow (day 150, for both 2002 and 2003), different sized basins exhibit different trends in streamflow timing (as marked in Fig. 4). Tuolumne River gages monitoring basins larger than 200 km^2, illustrated by Glen Aulin (251 km^2) and Hetch Hetchy (775 km^2, Fig. 5), report remarkably constant streamflow timing, similar to the consistency observed in most USGS-gauged basins throughout the Western U. S. (Lundquist and Cayan, 2002, Lundquist and Dettinger, submitted).

In basins smaller than 30 km^2, illustrated by Rafferty Creek (25 km^2, Fig. 3), travel time decreases 5-6 hours during this period in both years, similar to patterns observed in other small basin studies (Bengtsson, 1981; Caine, 1992; Colbeck, 1972; Dunne et al., 1976; Jordan, 1983). In basins larger than 30 but smaller than 200 km^2, travel time patterns vary between years. At the Lyell Fork of the Tuolumne River (109 km^2, Fig. 5) in 2002, travel time increases 2 hours during the first half of the melt season, yielding travel times similar to those of larger basins. In 2003, streamflow timing at Lyell Fork decreases 3 hours over the same period, in a pattern similar to that observed at smaller basins.

During the latter half of the melt season, following the day of peak spring discharge, travel times increase in all basins as flows decline. During this period, all snowmelt originates from the highest peaks, and streamflow velocities decrease as water levels fall. Travel times increase most dramatically at the largest basins, where long travel distances magnify the time delays caused by declining velocities.

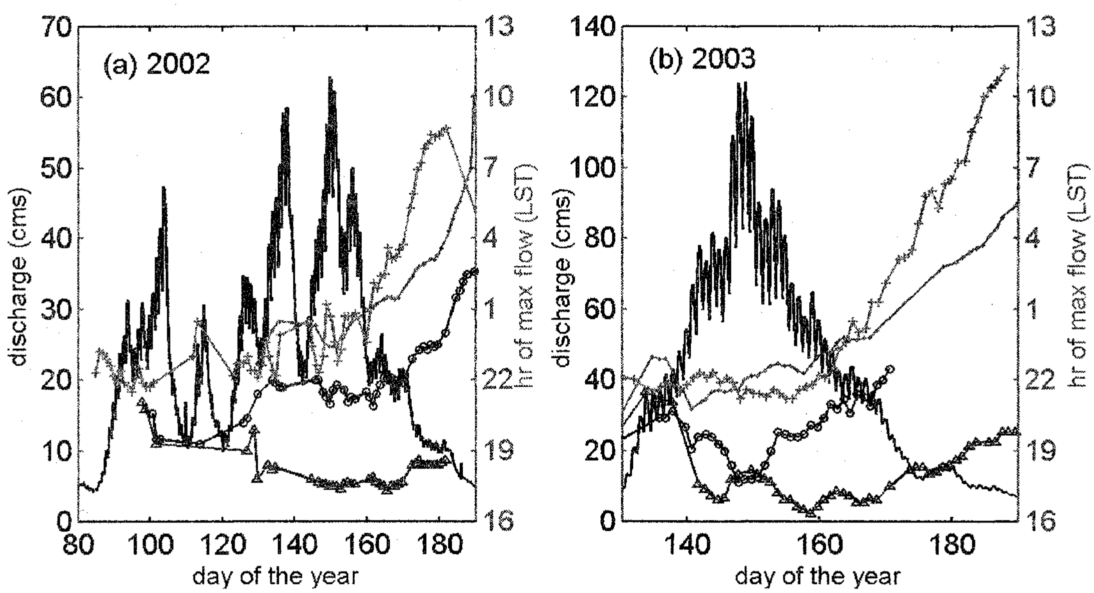

Figure 5. Observations of hour of peak streamflow (right axis) for Rafferty Creek (triangles), Lyell Fork (circles), Glen Aulin (dots) and Hetch Hetchy (pluses) along the Tuolumne River in (a) 2002 and (b) 2003. Merced River at Happy Isles discharge (highly correlated with discharge at all gages) is plotted in black, left axis, to reference the relative discharge magnitude at each time. Note: The focus of the present study is the trend before peak flow on day 150.

CHANGING CHANNEL LENGTH AND VELOCITY: OBSERVATIONS AND TRAVEL TIME ESTIMATES

Travel time in a basin equals distance traveled divided by travel velocity, summed along segments of flow paths from melting to measurement. The distance traveled depends on 1) the distance through the snow, 2) the distance traveled from the base of the snowpack to the channel, and 3) the distance through the river channel network. The travel velocity in both the snowpack and the river channel increases with increasing discharge. To interpret the observed patterns, we partition the travel time between travel in the snowpack and travel along the hillslope and stream channel. Adding a snowpack component to D'Odorico and Rigon (2003)'s estimated mean travel time, $E[t]$, yields

$$E[t] = \frac{< x_c >}{u_c} + \frac{< x_h >}{u_h} + \frac{< x_s >}{u_s} \tag{1}$$

where $<x_{c/h/s}>$ is the average travel distance and $u_{c/h/s}$ is the average velocity of the channel/hillslope/snowpack.

Hillslope length, $<x_h>$, is calculated as 1/2D (Bras, 1990), where D is the drainage density, measured from the 30-m resolution DEM for the watershed (Fig. 2). For the Tuolumne River drainage, $<x_h> = 250$ m. Estimates of u_h range from 14 m hr^{-1} for travel along the saturated base of a snowpack (Dunne et al., 1976) to 10.8 to 540 m hr^{-1} for overland flow (Dunne, 1978). Thus, the time spent in hillslope transport could vary from less than 1 hour to just under a day. Because time spent in hillslope transport is believed to be independent of basin scale (D'Odorico and Rigon, 2003), hillslope transport should not be responsible for the observed scale-dependent variations in timing.

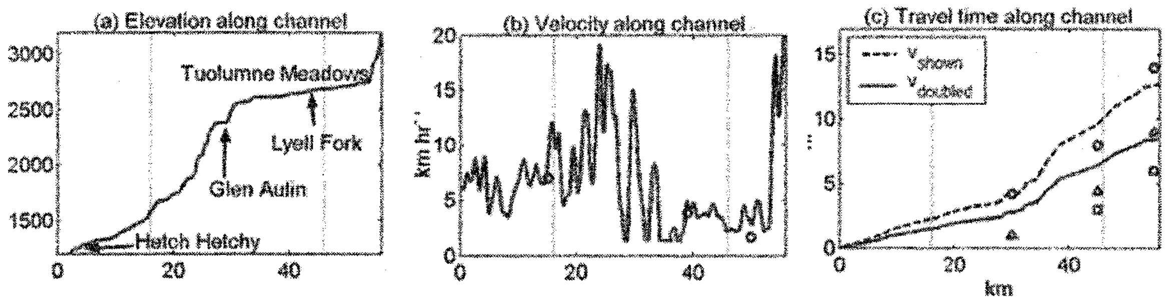

Figure 6. (a) Elevation vs. distance along the main channel of the Tuolumne River from Hetch Hetchy to the top of Mt. Lyell. The channel slope greatly decreases through Tuolumne Meadows (the section near the Lyell Fork gage). (b) Estimated velocity vs. distance along the same channel, assuming $v = 42 \sqrt{s}$ (line). Dots represent average velocities for each section measured from hourly streamflow records on 29 June 2002, when discharge in the Merced River at Happy Isles was 10 cms. (c) Travel time from a point along the channel to the Hetch Hetchy gage as a function of distance. Travel times measured from hourly records of streamflow peaks (Fig. 3) are represented by circles (29 June 2002), triangles (14 June 2002), and squares (14 June 2003). Dashed line shows the integrated travel time using the velocity curve shown in b. Note that the lines fall below the observed delays from the headwaters of Lyell Canyon and fall above the observed delay from the Lyell Fork gage.

Flow through the snowpack can be modeled as the vertical propagation of water through an unsaturated porous medium (Colbeck, 1972). In California, the snowpack is typically isothermal at 0°C at all depths (Smith, 1974) before spring melt begins. Based on snow pillow measurements from 2002 and 2003, typical measured rates of snowmelt in the central Sierra range from 1.5 cm day^{-1} early in the season to 4.5 cm day^{-1} during peak melt (Fig. 2b). These daily melt rates are assumed to follow the diurnal cycle in radiation, which is modeled as half a sinusoid with a peak at noon. Assuming that the integrated melt over half a sinusoid equals the measured daily melt rates at the snow pillows, we calculate average daytime melt rates of 0.125 to 0.375 cm hr^{-1}, and maximum melt rates of 0.196 to 0.589 cm hr^{-1}. From a snowmelt propagation model (Dunne et al., 1976),

$$u_s = Km^{2/3} \tag{2}$$

where u_s is the vertical velocity through the snowpack, m is the melt rate at the surface, and K is a constant based on snowpack properties, $K = 99$ cm$^{1/3}$ hr$^{-1/3}$ (1.4 m$^{1/3}$ s$^{-1/3}$, from Lundquist and Dettinger, submitted). Combining this equation with the average hourly melt rates above, typical velocities range from 25 to 70 cm hr^{-1}. These velocity values are consistent with those found in field experiments: 25 cm hr^{-1} (Dunne et al., 1976), 36 cm hr^{-1} (Colbeck and Anderson, 1982), and 50 cm hr^{-1} (Kobayashi and Motoyama, 1985).

Based on the initial average snow depth and melt rate at the snow pillows (Fig. 4), $\langle x_s \rangle = 190$ cm and $u_s = 25$ cm hr^{-1} in both 2002 and 2003, yielding an initial snowpack travel time estimate of 7.7 hours. The snowmelt season was much more compressed in 2003, due to the late start of melt, but in both years peak flows occurred around 30 May (day 150), when average melt rates reached 4.5 cm day^{-1} (Fig. 2) and $u_s = 50$ cm hr^{-1}. At this time, $\langle x_s \rangle = 30$ cm in 2002 and 90 cm in 2003, for average snowpack travel times of 0.6 hr and 1.7 hr, respectively. The combined decrease in travel distance, $\langle x_s \rangle$, and increase in velocity, u_s, result in a decrease in travel time of 6 to 7 hours. This shift is observed in subbasins with area less than 30 km^2, illustrated by Rafferty Creek (Fig. 5), in both years.

Estimating the travel time spent in the stream channels requires knowledge of channel length and time-integrated velocity. Satellite images show how channel lengths increase as the snowline retreats to higher elevations (Fig. 3). If most discharge originates from the edge of the snowline along the main channel of the Hetch Hetchy basin $\langle x_c \rangle = 16$ km on 09 May 2003, when melt begins, and $\langle x_c \rangle = 46$ km on 02 June 2003, when peak flow occurs. If all snow-covered areas contribute equally to basin run-off, then $\langle x_c \rangle = 33$ km on 09 May 2003, and $\langle x_c \rangle = 40$ km on 02 June 2003.

Channel velocities depend strongly on channel geometry and roughness and are difficult to parameterize in mountain areas (Bathurst 2002). Velocity increases with discharge (Rickenmann, 1994) and slope (Bras, 1990), but the precise relationship varies between river reaches. Following the form of the Manning and Chezy equations, velocity can be calculated as

$$v = C(Q)\sqrt{s} \tag{3}$$

where u_c is velocity, s is channel slope, and $C(Q)$ is an unknown parameterized function of channel shape and roughness that varies with discharge level, Q.

Late in the season, the propagation of the diurnal cycle down the channel can be used to estimate average velocities (Fig. 5). By mid-June of 2002 and 2003, only the highest elevations had snow, and most of the water originated from the glaciers and snowfields surrounding Mt. Lyell, the highest peak in Yosemite. The time delay measured from one channel gage to the next can be combined with measurements of slope and discharge to estimate $C(Q)$ and model the velocity down the channel. During the latter half of the melt season, peak timing varies less than one hour between all monitored small basins in the region, so we assume that the peak time observed at Rafferty Creek is about the same as the peak time at the first-order streams at the head of Lyell Canyon. With this assumption, we measure the difference in hours of peak flow between gages (Fig. 5) and combine that with distance measurements from the DEM (Fig. 3) to estimate mean velocities for different channel sections during different levels of discharge (Fig. 6).

During periods of higher discharge, the Tuolumne River at Hetch Hetchy gains considerable water from closer regions in the north of the basin (Fig. 3), so that discharge peaks there at close to the same time as discharge peaks near Glen Aulin (Fig. 5). The channel slope near the Lyell Fork gage is considerably less than the slope downstream of Glen Aulin (Fig. 6a). Hence, the delay introduced by the relatively flat Tuolumne Meadows is much greater than the delay in the steeper channel sections downstream of Glen Aulin. Modeling velocity as a function of slope (Fig. 6b), with $C(Q{=}10\ cms) = 38$, illustrates the dramatically lower velocities in the low-gradient Tuolumne Meadows section. Combining velocity with distance, cumulative travel times can be estimated along the channel (Fig. 6c). The low-gradient section introduces a considerable increase in travel time over a relatively short distance. Travel times with a doubled velocity represent conditions more likely during higher flows.

The change in travel time caused by discharge-level-induced velocity changes is much greater for flows originating upstream of Tuolumne Meadows than those originating downstream. Thus, the most dramatic increase in streamflow travel time occurs when the location of snowmelt changes relative to the low-gradient channel section. Early in the season most melt originates downstream of the flat section of Tuolumne Meadows, so all channel travel times are fast. Later, most melt originates upstream of the flat section, and the travel delay through the meadows can be measured at gages downstream. Once most flow originates upstream of the slow velocity section, channel velocities become more important.

Table 1. Effect of changing snow line and water velocity on channel travel times at different scales.

2003	$\langle x \rangle_{initial}$ 09 May (km)	$\langle x \rangle_{final}$ 02 June (km)	$u_{initial}$ (km hr^{-1})	u_{final} (km hr^{-1})	$t_{initial}$ (hr)	t_{final} (hr)	Δt (hr)
Rafferty Cr	0	2	6	8	0	0.25	+0.25
Lyell Frk	0	9	1.7	3.3	0	2.7	+2.7
Hetch Hetchy	16	46	7	15	2.3	3.1	+0.8

Table 2. Estimated change in time of peak flow (from mean parameters) versus observed change in time of peak flow for the first month of melt in 2003 for three Tuolumne River subbasins.

	2003 Δt_{est}	2003 Δt_{obs} (Fig 5b)
Rafferty Cr.	-6+0.25= -5.75	-6
Lyell Fork	-6+2.7= -3.3	-3
Hetch Hetchy	-6+0.8= -5.2	0 ($\pm$1)

Table 1 summarizes the change in channel travel time caused by the combined influences of increased travel distance and increased velocity on basins of different scales. The increased travel distance introduces the largest delay at the Lyell Fork gage, which is immediately downstream of the low-gradient, low-velocity section of Tuolumne Meadows. Overall, increasing travel velocities, which decrease travel times, offset the delays caused by increasing travel distances. The estimated net change in channel travel time from the snowfield to Hetch Hetchy is less than 1 hour.

For all of the observed basins, the time delay increase in the stream channel is not enough to offset the 6-hr decrease in travel time observed in the snowpack. Table 2 lists the estimated (from mean measured parameters) and observed changes in hour of peak flow in three subbasins of the Tuolumne River for 2003. The slow-velocity section immediately upstream of the Lyell fork gage introduces delays immediately following the snowline retreat, so channel properties can explain the diminished change in timing. However, changing channel lengths alone are not enough to erase the influence of decreasing snow depth on diurnal cycle timing before the water reaches Hetch Hetchy.

VARIABLE VELOCITY MODEL

The observations presented above suggest that the channel geometry and changing snowline in the Hetch Hetchy basin cannot fully explain the observed patterns in diurnal streamflow timing. This section applies the variable velocity model (VVM) developed by Lundquist and Dettinger (submitted) to the Tuolumne River Basin to test the hypothesis that basin snowpack heterogeneity must also be considered in order to accurately simulate hourly streamflow timing.

The VVM distributes the melt rates measured from the snow pillows as half-sinusoids with peaks at solar noon. This hourly melt is input at the top of a snowpack and propagates vertically according to (2), following Dunne et al. (1976). Once the melt water reaches the base of the snowpack, the model separates flow into surface and subsurface components. Surface flow is routed using the Manning equation for a triangular cross-section, and subsurface flow is routed using a simple linear reservoir (Bras 1990). Flow is routed from upstream to downstream channel reaches, which have lengths, slopes, and orientations as determined from the DEM (Fig 7). Hillslope processes are not modeled explicitly. See Lundquist and Dettinger (submitted) for a complete model description.

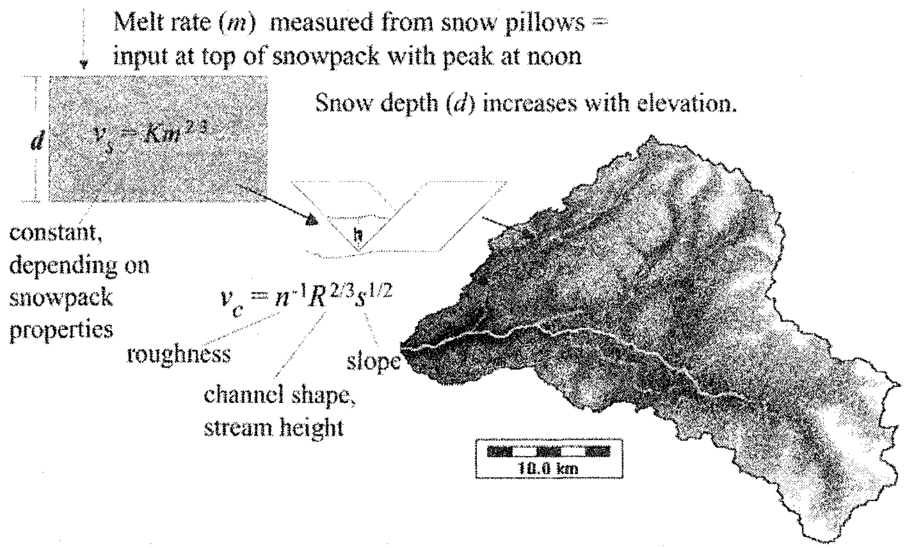

Figure 7. Elements of the Variable Velocity Model (VVM) applied to the Tuolumne River basin.

Input to the VVM consists of time series of mean melt rates and snow depths for each channel link. These are obtained from 47 Sierra Nevada snow pillows (discussed above). Elevation provides the primary control on snow accumulation in the Sierra Nevada (McGurk et al. 1993 and Aguado1990), and the observed snowline is generally co-located with contours of elevation.

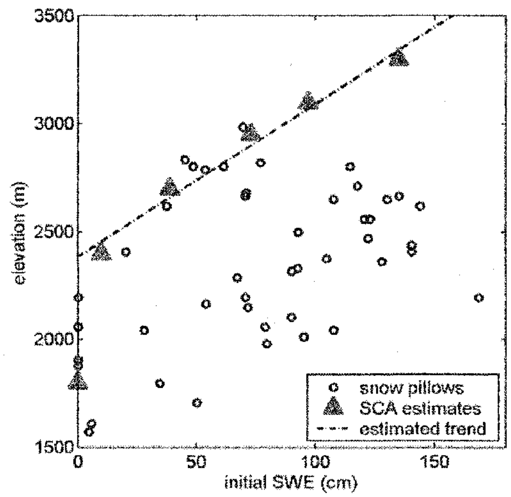

Figure 8. Elevation trends in SWE for the start of the 2003 melt season, measured from snow pillow (circles) and satellite images of snow-covered area (SCA, triangles). The line represents the best linear fit to the SCA estimates, excluding the outlier with 0 cm initial SWE. Using this linear estimate, SWE=-0.14*elevation-336. Units for the equation are the same as those shown in the graph.

Mean melt rates and maps of snow-covered area were combined to model mean snow depths at each elevation. Spatially, melt rates vary much less than snow depth in alpine basins (Anderton et al. 2002, Dunn and Colohan 1999, Luce et al. 1998, Hartman et al. 1999). Thus, for simplicity, we assume that melt rates are not functions of elevation or location. The initial snow depths were specified according to elevation such that, when the mean melt rates during the season were applied, snow cover disappeared at the times observed in satellite images (Fig 8). This slope falls along the line of snow pillows with relatively low snow accumulation compared to other snow pillows at similar elevations (Fig 8). Because snow pillows are generally installed where large amounts of snow accumulate, they can over-represent the snow depth for a given elevation. This over-representation is greater at lower elevations where snow-cover is very patchy.

To represent variations in snow depths and melt rates within each subbasin, the VVM runs in a Monte-Carlo mode (Lundquist and Dettinger, submitted), where the mean snow depths and melt rates for contributing areas are sampled repeated from among normally- distributed random variates with means and standard deviations as measured at snow pillows. The perturbed depths and melt rates are used in the complete model, and the results from 100 samplings are averaged. This technique represents the combined contributions of daily melt pulses originating from the wide variety of snow patches of differing depths and receiving differing solar forcing within each sub-basin.

HETCH HETCHY MODEL RESULTS

When the model is run with snow depth increasing with the mean elevation for each channel link (Fig. 8) but without including snow depth heterogeneity within each elevation, the simulated times of peak daily flows shift earlier in the day in all basins at all scales (Fig. 9a). Daily peak flows at Rafferty Creek arrive about 8 hours earlier by 30 May (day 150, the seasonal peak) than they do at the start of the melt season. This shift is about 35% larger than the observed shift of 6 hours (Table 2). At the Lyell Fork of the Tuolumne River, daily travel time shifts about 4 hours earlier between the start and peak of the melt season, compared to an observed and estimated shift of 3 hours (Table 2). At the Hetch Hetchy gage, the simulated timing shift is about 3 hours, compared to an estimate of 5 hours and an observation of 0 hours. The simulated shift of daily peak flows to earlier in the day is interrupted at Hetch Hetchy and Lyell Fork after the season peak discharge. By this time, the snow-covered area has decreased to the point where most of the discharge originates upstream of Tuolumne Meadows, the very flat-gradient and slow-velocity river reach shown in Fig 6.

When the effects of heterogeneity and elevation are both included in VVM (Fig. 9b), simulations for 2003 show results similar to observations. Simulates times of daily peak flows at Rafferty Creek shift about 6 hours earlier, as both estimated and observed. Simulated peak flows at the Lyell Fork of the Tuolumne River arrive 3 hours earlier, and simulated times of daily peak flow at Hetch Hetchy remain relatively constant. Both results match observations of changes in travel times (Table 2). The simulated flows at Hetch Hetchy peak consistently at 1900 local standard time (LST), 3 hours earlier than the observed peak time of 2200 LST. Simulated peaking times at Rafferty Creek and the Lyell Fork are also 1-2 hours earlier than observed times. The offset may be due to an underestimate of channel roughness, resulting in overestimates of channel velocities. The too-early simulations also may be due to hillslope travel times, which were not included in the model.

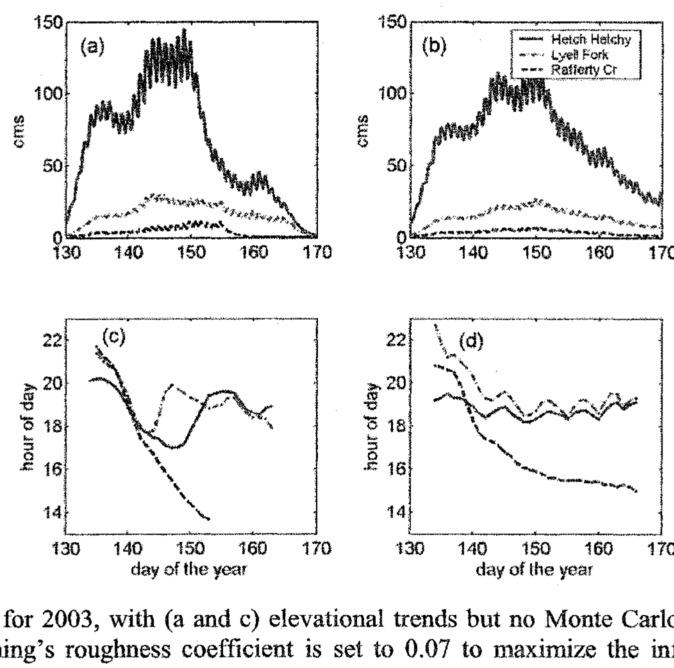

Figure 9. Model results for 2003, with (a and c) elevational trends but no Monte Carlo variations, averaged over 100 simulations. Manning's roughness coefficient is set to 0.07 to maximize the influence of channel delays. Decreasing the roughness changes the timing patterns in c but not in d.

Modeled results for 2002 (not shown) yield similar results. Without heterogeneity, all basins evolve unrealistically towards earlier daily peak flows. When heterogeneity in snowpack properties is included, model results agree with observations. Peak flows arrive earlier each day at the Rafferty Creek gage and remain relatively constant at the Lyell Fork and Hetch Hetchy gages as the season progresses. These results suggest that heterogeneity in snow properties is a necessary component for accurately simulating streamflow travel times in large basins.

SUMMARY, DISCUSSION, AND APPLICATIONS

Prior studies of nested basins (Braun and Slaymaker, 1981; Kobayshi and Motoyama, 1985) have suggested that the diurnal cycle in streamflow might yield information about snow characteristics in basins of all scales. This study examines nested basins across a wider range of areas and elevations and demonstrates that diurnal streamflow timing is dominated by different characteristics at different basin scales. Basin-mean snow depths dominate diurnal streamflow timing in small basins with limited elevation range and thus, limited variations in snow depth and melt rate. In basins with areas less than 30 km^2, the hour of daily peak flow generally begins the season near midnight and then shifts progressively earlier in the day.

Snow heterogeneity and in-channel travel distances from where snow is melting have greater influences on the timing of diurnal cycles in streamflow in basins that span large ranges of elevations, with a variety of snow depths, melt rates, and in-channel travel distances. During the first half of the melt season, streamflow from these basins tends to peak consistently near midnight. As the melt season progresses, snowlines retreat to higher elevations, and higher elevations correlate with longer travel distances. However, during the first half of the melt season, increases in travel time due to increased distance are often mostly offset by decreases in travel time due to increased velocity. Thus, both simple travel-time estimates and model simulations indicate that the retreat of the snowline is not sufficient to explain the observed near-constant time of peak flow observed at Hetch Hetchy from the start of the melt season until the day of peak discharge. Heterogeneity in snow depth and melt rates (Lundquist and Dettinger, submitted) must be taken into account in order to properly simulate observations.

Pressure sensors recording stream stage, such as those used in the Yosemite study, are small, inexpensive, and easy to deploy in remote mountain areas. Developing rating curves for sites in remote areas is labor-intensive and time-consuming, but precise information about diurnal cycle timing can be gleaned from stage information alone, without rating curves. Hence, the ideas presented here provide practical ways to learn about snow and basin properties in remote, previously unmonitored basins. Diurnal timing measured at different locations along a single channel gives direct measurements of travel time along channel segments. This information can be combined with distance and standard flow equations to estimate changing velocities and roughness coefficients for these segments at various levels of discharge (Fig. 6).

Diurnal timing in small streams, typically with areas less than 30 km^2, provides information about snow water reserves throughout the melt season and might be used to forecast water supply. Timing in larger river basins reflects the extent of heterogeneity within the basin. Timing in larger rivers is also remarkably consistent during the first half of the melt season each year, and this information can be used by hydroelectric power plants to predict the best hours of operation, by river rafters to decide the best time to run a river, and by hikers to decide the best time to cross a stream. The diurnal cycle also provides information during spring storms when clouds obscure the basin from satellites but snow-covered area is changing rapidly.

Mountain channel morphology is characterized by pronounced and abrupt variations in gradient (Wohl, 2000), and almost all mountain basins in the western U.S. have sections of very flat topography (generally lakes and meadows). The largest effect of changing travel distances occurs when the location of most snowmelt shifts from downstream to upstream of a low-gradient channel section with slow velocities, such as at Tuolumne Meadows. Where low-gradient sections are long enough, they can introduce pronounced delays in diurnal cycle timing. When these sections are located at elevations upstream of the gage, diurnal cycle timing can be used to determine what proportion of melt is originating upstream or downstream of this low-gradient section. A full examination of this effect is beyond the scope of the present study, but we are currently investigating a method of combining channel slope information (easily obtainable from DEMs) with diurnal cycle information (already measured at most gages) to identify locations in the basin where snowmelt originates throughout the melt season.

Current hourly discharge/stage information is already available at 83% of USGS gages, and this is available for operators and forecasters. Using the context presented here, scientists can start applying the diurnal cycle to better understand snowpack properties, basin heterogeneity, the location of snowmelt, or stream channel velocities.

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