Chapter 5

Variability and Trends in Spring Runoff in the Western United States

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Abstract

In the western United States, over half of the water supply is derived from mountain snowmelt, with the snow acting as a natural reservoir, delaying runoff and providing runoff in the spring and summer when it is needed most. Interannual variability of both the magnitude and timing of spring runoff is tremendous, and western states have developed extensive reservoir systems to store water from wet years in order to weather droughts. However, important changes in snowpacks and runoff timing have been noted in recent decades. The fraction of annual streamflow that runs off during late spring and summer has declined by 10 to 25%. Warmer winters and springs have led to earlier snowmelt and a higher percentage of precipitation falling as rain rather than snow. Snowmelt runoff timing has advanced so that it arrives approximately one to three weeks earlier in 73% of mountainous catchments across western North America. Even conservative climate-change projections suggest that California could lose one third of its present day spring snowpack by the middle part of the 21st century, and other western states will follow suit. Hydrologists have long known that snow is a key component of
the water budget, and yet most operational runoff models are based on historically
derived empirical relations, rather than on spatially and temporally distributed physical
processes. In a changing climate, the empirical relations may not remain valid, so more
observational attention to the high altitude snowfields that supply so much of the region’s
water supplies is needed to provide a basis for predicting the magnitude and timing of
snow melt, sublimation and runoff.

Introduction

In the western United States, over half of the water supply is derived from mountain
snowmelt, with the snow acting as a natural reservoir, delaying runoff and providing
water in the spring and summer when it is most needed. The total volume of spring
runoff can vary by an order of magnitude from one year to the next, and the timing of
spring runoff varies by ± several weeks. Western water managers have invested heavily
in reservoir and conveyance systems to store water from wet seasons and years for use in
subsequent drought seasons and years. However, while populations and water demands
in the region continue to grow, several recent studies have identified marked changes in
springtime runoff. The fraction of annual streamflow that runs off during late spring and
summer has declined by 10 to 25% since the 1950s (Roos 1991, Wahl 1992, Dettinger
and Cayan 1995). Warmer winters and springs have led to earlier snowmelt (Cayan et al
2001) and a higher percentage of precipitation falling as rain rather than snow (Knowles
et al 2006). Snowmelt runoff timing has advanced by approximately one to three weeks
earlier in the large majority of mountainous catchments across western North America
(Stewart et al. 2005, Regonda et al. 2004). These observed trends raise some urgent
questions: (1) Are they connected to global warming, or can the shifts be explained by natural variability? (2) What can we expect in the future? (3) How can we prepare for future changes? This paper summarizes observed changes in spring runoff in the Western U.S., plausible reasons for these changes, expected future changes, and current efforts to cope with expected change.

Observed trends in Spring Runoff

The examination of spring runoff timing began when Maury Roos noted that the fraction of total annual runoff occurring between April and July had declined in California since 1906 (Fig. 5.1, Roos 1987, 1991). The April through July (AMJJ) period encompasses most of the snowmelt season, and for most snowmelt-driven watersheds in the western United States, AMJJ flows are the most important contribution to the annual streamflow, comprising 50-80% of the annual total (Stewart et al. 2005). In contrast, mean AMJJ flow fractions are less than 30% at non-snowmelt-dominated gages, reflecting a lack of precipitation in much of the western United States during this period. Furthermore, AMJJ flows follow the main winter-storm season, during which many of the worst floods along the west coast occur. During the winter-storm season, reservoirs can only be partially filled, to allow space to mitigate floods. AMJJ flows are thus all the more important because this period is the most predictable portion of the annual runoff, allowing reservoirs and water supplies to be most confidently managed.

Following Roos’ work, subsequent studies revealed that the California trends were not isolated, and the AMJJ fraction of runoff was declining in rivers throughout the western United States (Wahl 1992, Aguado et al. 1992, Pupacko 1993, Dettinger and
Cayan 1995). The large regional coherence of these trends was further supported by studies finding that the snow-fed rivers of the Western U.S. rise and fall together each spring on time scales from days to year-to-year differences, along with the temperature patterns that control snowmelt (Peterson et al. 2000, Cayan et al. 1993).

Cayan et al. (2001) developed a second measure of spring runoff timing, the “spring pulse onset,” which is the date of the beginning of the spring snowmelt-derived streamflow for snowmelt-dominated rivers. Their algorithm identifies the day when the cumulative departure from that year’s mean flow is most negative, equivalent to finding the day after which most flows are greater than the season’s average (Fig. 5.2). In snowfed streams, the flows are typically an order of magnitude larger after the onset of snowmelt than before, so that this measure was as useful as it was simple. They found that these pulse dates are highly correlated with both regional temperatures and the dates of spring blooms of lilac (*Syringa vulgaris*) and honeysuckle (*Lonicera tatarica* and *L. korolkowii*), and that regionally blooms occurred 5-10 days earlier in the last half of the record (1976 to 2000) than in the first half (1950 to 1975).

Stewart et al. (2005) analyzed 302 stream gages in western North America for the period from 1948 to 2002 using both of the previous timing measures and a new measure, the timing of the center of mass of annual flow (CT) for each water year (Fig. 2), where

\[
CT = \frac{\sum (t_i q_i)}{\sum q_i}, \quad \text{where } t_i \text{ is time in days (or months) since the beginning of the water year (October 1) and } q_i \text{ is the corresponding streamflow in that month or day.}
\]

CT and the spring pulse onset date are strongly correlated \((r = 0.5 - 0.8)\) for most gages and significantly correlated at 74% of the gages. Regonda et al. (2004) used a similar measure to CT for a similar study period, calculating the date by which 50% of the water
year flow had passed a gage. CT is a robust measure of when most of the spring runoff arrived because, unlike the AMJJ fraction, it indicates whether the “missing” snowmelt-season water has shifted earlier or later, and unlike the spring pulse date, it is a measure at the seasonal/annual scale, rather than being controlled mainly by the events of a few brief weeks. Stewart et al. (2005) and Regonda et al. (2004) both reported statistically significant CT trends.

Stewart et al. (2005) found that AMJJ fractional flows have declined for 81% of the gages, even while 50% have measured increased overall annual flows. In the Pacific Northwest, annual flows decreased overall during the 1948-2002 period as well. Spring pulse onset dates shifted 10-30 days earlier, with the largest changes observed in the Pacific North-West and the Sierra Nevada. Of the 241 snowfed rivers, 66% (159) have experienced trends towards earlier spring pulse days by 3 days or more, and 54% (86) of these negative trends are significant at the 90% confidence level. Only 15-19% of the gages have shown no trends or trends towards later spring pulse days in later years. Results for CT were similar. Trends toward earlier centers of mass have been measured at 73% (214) of the gages, and 49% of these earlier trends, totaling 105 gages, are significantly different from zero at 90% confidence levels. By contrast, CTs for most non-snowmelt dominated gages, which usually are low-altitude coastal steams, have been trending in the opposite direction, towards later streamflow timing, with corresponding shifts of 5-25 days.

Most of these studies focused on the period of record after the mid-1940s because (1) this period has the most complete dataset available, and (2) the most significant contribution to trends took place during this period. Available data for the 1901-2002
period (not shown) reveals overall trends towards earlier streamflow timing for the coastal states and smaller or no trends for gages in the interior (Stewart et al. 2005). The 1900-1948 period did not have consistently trending streamflow timings (Dettinger and Cayan 1995), suggesting that the observed trends are a late 20th Century phenomenon.

Causes

The 1948-2002 timing changes were caused primarily by temperatures that have been observed to rise during the same period. This warming has been regional in extent (Cayan et al. 2001) and is believed to be part of global patterns of warming associated with increased greenhouse gases (Houghton et al. 2001, National Research Council 2001) in combination with warmth due to natural multidecadal variations of the climate system (e.g., Mantua et al. 1997). The timing of spring runoff is controlled by both temperature (Peterson et al. 2000) and precipitation (Aguado et al. 1992, Hamlet and Lettenmaier 1999, Stewart et al. 2005). Predicting future changes in Western U.S. runoff requires an understanding of how each of these processes has influenced spring runoff in the past.

Influence of temperature vs. precipitation

Observational studies have shown clear increasing trends in temperature and some less-certain increasing trends in precipitation over the West in the 20th century (McCabe and Wolock 2002, Mote et al. 2003, Mote 2003b, Shepard et al. 2002). While warmer temperatures cause snow to melt earlier, increased winter precipitation results in larger snowpacks, which are correlated with later spring melt (Hamlet and Lettenmaier 1999, Stewart et al. 2004). Additionally, the central timing of annual precipitation has tended to
come about a week later over the past 50 years (Stewart et al. 2005), a trend reflected in the timing of non-snowfed rivers (which shifted to later streamflow) but masked in snowmelt-dominated rivers by the effects of warming temperatures. That is, temperature and precipitation forcings on streamflow have had opposing influences in snowfed streams in recent decades. Hamlet et al. (2005) applied the Variable Infiltration Capacity (VIC) hydrologic model (Liang et al. 1994, Cherkauer and Lettenmaier 2003) over the western United States to isolate the effects of precipitation and temperature from the period of 1916-1997. They found that, although precipitation and temperature trends both contributed to runoff-timing changes, but that the combined effect of the temperature and precipitation trends on timing of peak snow accumulation and melt, and hence streamflow timing, was dominated by the temperature effects. Stewart et al. (2005) and Regonda et al. (2004) showed the same temperature dominance using statistical analyses of historical flow timing and climate records.

Natural variability vs. external forcings

Through shifts in regional patterns of temperature and precipitation, large-scale atmospheric circulation patterns have strong influences on spring runoff patterns. For example, circulation-driven trends toward warmer winters since the late 1940s can explain much of the western trend in runoff timing (Dettinger and Cayan 1995). Warmer winters and springs in recent decades were associated with deeper Aleutian lows and a southern displacement of the westerly wind fields over the central North Pacific, which brought warmer air masses over the west coast of the United States (Fig. 5.4).

During the period from 1951 to 2000, the PDO shifted from a regime that favors cool conditions in the West to the opposite regime, which favors warmth, in a major step change from a cool to warm PDO in 1976-77 (Mantua et al. 1997). The two multi-decade regimes of the PDO cover the period during which most records of streamflow timing were examined. Analysis of the period from 1924 to 1976, when the PDO shifted from its warm/dry phase to its cool/wet phase, revealed trends towards larger magnitude and later timing of maximum snow accumulation, concentrated in the Pacific Northwest (Hamlet et al. 2005). Thus, the sign of the PDO and the timing of spring runoff are negatively correlated ($r = -0.2$ to $-0.8$, Stewart et al. 2005), such that the cool phase of the PDO corresponds to later melt, and the warm phase of the PDO corresponds to earlier melt.

Because of this negative correlation between timing and the PDO, and the large PDO step towards warm PDO in the middle of the recent timing trends, the question of whether the timing trends were simply PDO responses naturally arises. Towards this end, Stewart et al. (2005) estimated the trend that the PDO would be expected to impose on CTs and temperatures from interannual variations of the PDO within the separate PDO
regimes from 1948-1976 and from 1977-1998. Applying the influences found before, and after, the 1976-77 PDO shift to the PDO history as a whole demonstrated that the long-term PDO variations explained less than half of the observed timing trends at most snowmelt stations in the western US (Fig. 5.5), with exceptions appearing in the Pacific Northwest where PDO climatic influences are generally strongest (Mantua et al. 1997). Furthermore, from 1977 to 1998, the PDO did not trend overall, and yet, streamflow timing did continue to trend towards earlier in the year. Other evidence that the shift in spring runoff timing cannot be entirely explained by the PDO includes: (1) Stewart et al. 2005 used linear regression to remove the PDO component of the CTs and then correlated the residual with temperature alone. CT variations correlated with temperature shifts without the PDO component. However, when the technique was reversed, with the temperature component removed, the residual was not correlated with the PDO (Fig. 5.6). (2) Hamlet et al. (2005) minimized the effects of decadal variability on streamflow timing by running the VIC model for back-to-back warm PDO epochs for the periods from 1924-1946 and 1977 to 1995. In the absence of a PDO shift, large scale warming still occurred, and the timing of snowmelt trended towards earlier in the spring. This result arose because the 1924-1946 “warm” PDO epoch was much cooler in the western US than was the more recent warm phase. Their model results suggest that, while decadal variability of the PDO probably does account for many trends in winter precipitation over shorter periods of the record, the temperature trends are greater than the PDO can account for. Because streamflow timing depends more on temperature than precipitation, these results suggest that the trends in spring streamflow are due to more than just the multidecadal natural variability of the PDO.
While warming temperatures affect the entire western United States in a similar manner, the effects of ENSO vary between the north and south, due largely to north-south precipitation contrasts (Redmond and Koch 1991, Dettinger et al. 1998). Stewart et al. (2005) found that El Niño conditions are associated with warmer spring temperatures and lower winter precipitation in the northern portion of the western United States, and therefore with earlier snowmelt-derived streamflow, leading to positive correlations of ENSO indices with CT, ranging from $r = 0.2$ to $r = 0.6$. In the southern portion of the study area, El Niño conditions are associated with higher-than-average winter precipitation, leading to a delay in snowmelt and negative correlations with CT ranging from $r = -0.2$ to $r = 0.6$. While interannual variations in streamflow timing can be at least partially explained by ENSO variations, the long-term trends towards earlier spring runoff have been observed throughout both the northern and southern portions of the western United States and cannot be explained by ENSO.

Regional Sensitivities
Changes in the timing of spring runoff vary primarily with elevation, such that areas with middle-elevation catchments, like the Pacific Northwest and the northern Sierra Nevada, have been the most sensitive to the observed warming trends (Mote 2003a, Regonda et al. 2004, Hamlet et al 2005, Stewart et al. 2005). While shifts of CT to 10-20 days earlier are common in basins below 2500 m, basins above 2500 m exhibit little or no change. The higher basins receive winter precipitation at the coldest temperatures and then remain below freezing for most of the spring (Regonda et al. 2004). Consequently, the high elevation areas in the Rockies and southern Sierra have been relatively less affected by
the statistically significant, but relatively modest, temperature trends observed to date (Hamlet et al. 2005). In contrast, the ratio of snow to winter rain is lower and more variable in the low-elevation Pacific Northwest (Serreze et al. 1999), and the shift of spring snows to spring rains enhances the trends towards earlier runoff. Long-term trends towards less snowfall and more rainfall, in response to warming, across most of the region have been documented recently by Knowles et al. (2006) and the rain-vs-snow changes were shown to be largest in the middle elevations. Because of these elevation effects, rivers in the warmer coastal states have been affected primarily by warming in the winter (Dettinger and Cayan, 1995), whereas areas with a more continental climate are more sensitive to precipitation trends during the winter and to warming in late spring (Cayan et al. 2001; Hamlet et al. 2005; Knowles et al. 2006).

To provide a global context for the regional trends discussed in this chapter, an analysis of CTs in rivers around the globe with long-term (20+ years) discharge records (described by Dettinger and Diaz 2000) is presented in Fig. 5.7. The results mapped there reveal that, while some of the largest trends toward earlier spring melt have occurred in western North America, similar trends have occurred in rivers worldwide during the period from 1945 to 1993 (Fig. 5.7, Dettinger et al. 2001). Significant trends, using nonparametric Kendall’s tau trend statistics, are found in rivers throughout eastern Europe and western Russia, across Canada, and, less robustly, in the southern hemisphere. As observed in the western United States, watersheds where cool-season temperatures are often at or above freezing have exhibited the largest changes in streamflow timing.
Climate Projections

The studies described above demonstrate that snowfed runoff timing has been trending towards earlier in the year in western North America, and that this trend is controlled primarily by temperature and secondarily by precipitation (Stewart et al., 2004). Thus, projections of future streamflow characteristics depend on projections of future temperature and precipitation.

Projections from many different climate models have recently become available, including projections from the U.S. PCM, Canadian CCCM, German ECHAM4, British HadCM3, Japanese NIES, and Australian CSIRO coupled ocean-atmosphere global climate models (as described by Dettinger, 2006). A growing number of climate-change responses to plausible future greenhouse-gas emission scenarios have been simulated with such models, e.g., in response to the A2, B2, and IS92a SRES emissions scenarios (Houghton et al. 2001), which represent projections of relatively rapid, moderate, and intermediate rates of 21st century emissions increases, respectively. Simulations by the six climate models listed above, each simulating responses to each of the three specified greenhouse-gas-plus-sulfate-aerosols emissions scenarios, indicate that temperatures in the western United States may increase between about +2.5 and +9 °C by 2100, and that precipitation may increase, decrease, or remain the same (e.g. Fig. 5.8). Dettinger (2006) resampled these projections to estimate probabilities for various levels of climate change in the western United States. Analyzed this way, the ensemble of projections shown in Fig. 5.8 suggest, at typical grid cells in the interior West, that (1) temperatures warm in all projections, much as in Fig 5.8, (2) precipitation is not projected to change as much in the future, and (3) projection uncertainties increase as we look further into the future,
more so for temperatures than for precipitation. A resampling of the temperature projections across the conterminous US (Fig. 2.9) illustrates that projected springtime warming in the interior West may be among the largest nationwide, and that uncertainties (as indicated by the growing widths of the 2050 distributions in Fig. 5.9) are largest in the West. Thus the historical streamflow-timing trends discussed thus far can only be expected to continue and, indeed, to enter the interior west even more forcefully in the 21st Century, unless current warming projections are askew.

Several efforts to model future streamflow scenarios have used one of the most conservative models of future warming, the NCAR Parallel Climate Model (PCM) (Washington et al. 2000), which gives ensemble estimates of +2°C to +3°C warming and ±10% precipitation change over the 21st century, assuming a business-as-usual emissions scenario. Even with this conservative projection, Knowles and Cayan (2002) used a physically based land-surface model to discover that California could lose one-third of its present-day spring snowpack by the middle part of the 21st century. Using linear regression models, Stewart et al. (2004) projected that, at many gages, CT could shift to 30 to 40 days earlier by the end of the 21st century, in good agreement with more detailed land-surface simulations, such as those by Wood et al. (2004) using the VIC model and by Dettinger et al. (2004) using the Precipitation-Runoff Modeling System (PRMS, Leavesley et al. 1983).

These simulations and others (Jeton et al. 1996) agree with the observational findings that (1) increases in temperature advance the timing of spring runoff, (2) changes in precipitation change the magnitude of both spring and annual runoff, and (3) the most pronounced differences between how basins respond depend on what elevations they
drain. Under a business-as-usual climate scenario, streamflow timing trends are projected to be continuations of the observed trends with no large-scale accelerations, unless emissions growth itself accelerates (Stewart et al. 2004, Dettinger et al. 2004). However, Lundquist and Flint (2006) recently have shown that, once streamflow timing trends shift timings by more than about a month, the seasonal cycle of daily insolation may intervene to slow subsequent timing changes somewhat in many watersheds.

Preparing for the future

Hydrologists have long known that snow is a key component of the water budget, and yet most operational runoff models are based on empirical relations between historical runoff and weather fluctuations in specific basins, rather than on invariant relations describing spatially and temporally distributed physical processes. In a changing climate, these empirical relations may become increasingly unreliable, and a historically-based empirical model may not accurately represent changing snowcover, snowmelt amounts and contributions, and snowmelt rates. At the same time, recent work by Jain et al. (2005) suggests that the year-to-year variability in streamflow volumes has increased since 1972, resulting in more large floods and extreme droughts. Thus, predicting next year’s flow timing may become even more difficult in the future.

Most of the observations and modeling efforts discussed above mention elevation as the primary factor differentiating between the magnitude of trends in spring runoff in various basins in the western United States. However, very few measurements exist at the highest elevations (e.g. Fig. 5.10, Lundquist et al. 2003), and surprisingly little is known about how temperature, precipitation, snowmelt, and runoff vary with elevation in
different regions and weather conditions. Thus, more monitoring at high altitudes is needed to understand and track the anticipated changes in temperature and precipitation with elevation, and to predict the magnitude and timing of critical processes like snow melt, sublimation and runoff. For example, recent increased hydroclimatic monitoring in Yosemite National Park has demonstrated that, in some years, like 2002, the onset of spring melt and runoff in California may be virtually independent of elevation (Fig. 5.11, Lundquist et al. 2004). This occurred when synoptic weather conditions resulted in over 13°C warming in less than a week, a temperature change large enough to overcome more typical springtime lapse rates that hold the highest altitudes below freezing well after the lowest altitudes have thawed. Understanding how surface- and free-air temperature lapse rates and precipitation lapse rates might change in a future climate are active areas of research with complex and often localized results (Aguado 1990, Barry 1990, Beniston et al. 1994, Beniston and Rebetez 1996, Dettinger et al. 2004b, Pepin and Losleben 2002, Singh 1991, Williams et al. 1996). As snow disappears from lower elevation areas, changes at high elevations will have a critical impact on western United States water supplies. Thus, further studies are needed (CIRMOUNT, 2006).

Conclusions
During the past 50 years, the climate of the western United States has warmed. Warmer temperatures have led to earlier runoff and to more precipitation falling as rain rather than snow. These effects are greatest where temperatures are close to 0 °C, such as at middle elevations, and particularly in the coastal ranges of the Pacific Northwest. Climate projections agree that temperatures will continue to warm in the future. Even the most
benign of these projected climate changes are sufficient to significantly alter the landscape, hydrology, and land and water resources of the western United States, and those alterations are likely to become significant within roughly the next 25 years (Barnett et al. 2005, Dettinger et al. 2004, van Rheenen et al. 2004). If even the modest observed warming in the last half of the 20th century was sufficient to cause the hydrological changes reported in the literature, then prospects for additional climate-change impacts in the western United States should be taken seriously, despite large remaining climate-change uncertainties.


Figure Captions:

Figure 5.1  AMJJ fractional runoff for the Sacramento River, in California. [updated from Roos 1987]

Figure 5.2  Daily streamflow discharges, Merced River at Happy Isles, 1983 and 1992. Both the timing and magnitude of snowfed rivers varies widely. Two methods of determining streamflow timing are the onset of spring melt and the center of mass of annual flow, both illustrated here.

Figure 5.3 CT trends for 302 snowfed streamflow gages in Western North America, 1948-2000. [From Stewart et al 2005]

Figure 5.4  Composites of spring 700-mb height anomalies for (a) low and (b) high values of the first principle component of western North American CTs from 1948-2000, associated with (upper panel) earlier and (lower panel) later timing of snowmelt-derived streamflow. Statistically significant correlations at the 95% confidence level are marked with small black circles. [From Stewart et al. 2005]

Figure 5.5  Fractions of observed (a) temperature (T1) and (b) CT trends that are explained by the PDO step-change “trend”, using coefficients from regressions of CT and TI with PDO; regression coefficients from the 1948-1976 and the 1977-1998 periods were averaged for use here; fractions are shown only for gages with significant observed CT trends. [From Stewart et al. 2005]

Figure 5.6 Correlation of (a) CT minus the portion of its variance attributable to the TI (CTwoTI), with PDO and (b) CT minus the portion of its variance attributable to the PDO index (CTwoPDO), with TI. The larger symbols indicate statistically significant correlations at the 95% confidence level. [From Stewart et al. 2005]
Figure 5.7 Global trends in the timing of streamflow, as measured by the CT, the flow-weighted average day of flows, in extratropical snowmelt-dominated rivers. Red filled circles denote retrogression of the annual hydrograph. [Source of data: Dettinger and Diaz 2000]

Figure 5.8 Ensembles of historical and future (a) temperature and (b) precipitation changes from seven coupled ocean-atmosphere general-circulation models of the global climate forced by various greenhouse-gas-plus-sulfate-aerosols emissions scenarios. Heavy black curves are 7-yr moving averages of simulated climates forced by a business-as-usual (IS92a) emissions scenario, which involves middle-of-the-road future emissions; red curves reflect SRES A2 emissions, which are moderately pessimistic (assuming moderately more emissions); blue curves reflect moderately optimistic B2 emissions; magenta curves reflect more pessimistic A1fi emissions; and green curves reflect more optimistic B1 emissions. Dotted black curves are unfiltered versions of all the scenarios. Models shown are the Canadian CCCM climate model (A2, BAU and B2 scenarios shown), Australian CSIRO model (A2, BAU, and B2 scenarios), European ECHAM4 model (A2, BAU and B2 scenarios), British HadCM2 model (BAU scenario), British HadCM3 model (A2, BAU, and B2, with A1fi, and B1 temperatures in panel a), Japanese NIES model (A2, BAU and B2 scenarios), and US PCM model (A2, BAU, B2, A1fi and B1 scenarios). In all, 23 projections are shown. [from Dettinger, 2006]

Figure 5.9 Probability density functions (pdfs) for projections of future temperature anomalies across the United States for (a) 2001 and (b) 2050, considering the ensemble of climate simulations shown in Fig. 5.8. All pdfs are scaled the same, and all reflect the
combination of natural temperature variability, uncertainties due to model differences, and uncertainties due to emission-scenario differences.

Figure 5.10  Most long-term climate stations in California, shown here, and throughout the west are located at low elevations. However, conditions at higher elevations are crucial for water resources and need to be more closely monitored. [from Lundquist et al. 2003]

Figure 5.11  In 2002, snow in the Sierra Nevada of California began melting simultaneously at all elevations [below about 9,000 ft] around the 88th day of the year (March 29th), and streams rose dramatically. The line and the right axis show discharge at the Merced River at Happy Isles, in Yosemite National Park, central Sierra Nevada. The circles show the date of maximum snow accumulation, i.e. the date when melt began, at 45 snow pillows in the central Sierra Nevada, California, plotted against elevation, left axis. [from Lundquist et al. 2004]
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