Effects of Temperature and Precipitation Variability on Snowpack Trends in the Western United States

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ABSTRACT

Recent studies have shown substantial declines in snow water equivalent (SWE) over much of the western United States in the last half century, as well as trends toward earlier spring snowmelt and peak spring streamflows. These trends are influenced both by interannual and decadal-scale climate variability, and also by temperature trends at longer time scales that are generally consistent with observations of global warming over the twentieth century. In this study, the linear trends in 1 April SWE over the western United States are examined, as simulated by the Variable Infiltration Capacity hydrologic model implemented at 1/8° latitude-longitude spatial resolution, and driven by a carefully quality controlled gridded daily precipitation and temperature dataset for the period 1915–2003. The long simulations of snowpack are used as surrogates for observations and are the basis for an analysis of regional trends in snowpack over the western United States and southern British Columbia, Canada. By isolating the trends due to temperature and precipitation in separate simulations, the influence of temperature and precipitation variability on the overall trends in SWE is evaluated. Downward trends in 1 April SWE over the western United States from 1916 to 2003 and 1947 to 2003, and for a time series constructed using two warm Pacific decadal oscillation (PDO) epochs concatenated together, are shown to be primarily due to widespread warming. These temperature-related trends are not well explained by decadal climate variability associated with the PDO. Trends in SWE associated with precipitation trends, however, are very different in different time periods and are apparently largely controlled by decadal variability rather than longer-term trends in climate.

1. Introduction and background

Snowpack is crucial to the water resources of the western United States (Serreze et al. 1999). A dominant fraction (by some accounts, more than 70%) of streamflow in the western United States originates as melting montane snowpack. Over most of the western United States, precipitation in the mountains is strongly winter dominant and snow accumulation (along with storage in the soil and groundwater) is the primary physical mechanism by which winter precipitation is stored and transferred to the relatively dry summers. For river basins with substantial snow accumulation, scenarios of increasing winter and spring temperatures in climate change assessments typically result in increased winter runoff, reduced peak water equivalent stored as snow, earlier peak streamflows, and reduced summer streamflow volumes (e.g., Cayan 1996; Dettinger et al. 2004;
Observational studies have also shown increasing trends in both temperature and precipitation over the western United States in the twentieth century (McCabe and Wolock 2002; Mote et al. 2003, Mote 2003a; Sheppard et al. 2002) and widespread trends toward earlier runoff timing in the western United States associated with earlier spring snowmelt (Dettinger and Cayan 1995; Stewart et al. 2005; Regonda et al. 2005). Mote (2003b) examined trends in 1 April snow water equivalent (SWE) for the U.S. Pacific Northwest (PNW), and Mote et al. (2005) have extended the earlier study to examine snow course records over the entire western United States and southern British Columbia (BC), Canada. These studies both show strong downward trends in 1 April SWE over most of the domain from about 1950 onward. Mote et al. (2005) also corroborated observed SWE trends by comparing simulated and observed SWE trends from 1950 to 1997 and used correlations between 1 April SWE and winter temperatures to show that temperature trends were a major driver of these observed trends, particularly at moderate elevations with relatively warm winter temperatures. The role of temperature and precipitation in determining snowpack variations was also examined by Serreze et al. (1999).

Observed data provide direct empirical evidence of trends in SWE over the western United States, but model simulations are needed to corroborate and extend these observations in order to avoid problems with limited spatial coverage, coarse temporal resolution, and longevity of observed data sources (see also Mote et al. 2005). Observations from snow course measurements (i.e., direct measurements of average SWE over a small transect), for example, have had a reasonably consistent level of coverage over the western United States for the period from about 1950 to present, but the number of observations declines rapidly prior to 1950, and coverage is very uneven before about 1940 (Mote et al. 2005). These changes in station density prevent a consistent analysis for a period of record longer than about a half century. This is important because the period of record, 1950–present, has been shown to contain substantial (and regionally specific) trends in precipitation and temperature that are strongly influenced by climatic variability associated with the El Niño–Southern Oscillation (ENSO) and the Pacific decadal oscillation (PDO; Dettinger et al. 1998; Gershunov and Barnett 1998; Hamlet and Lettenmaier 1999a; Hidalgo and Dracup 2003; Mantua et al. 1997; Mote et al. 2003; Sheppard et al. 2002). In particular, the precipitation trends that accompany the well-known shift in the PDO from cool phase to warm phase in 1976–77 (Mantua et al. 1997) are a confounding element in the attempt to attribute observed losses of SWE from 1950 to present to long-term changes in climate.

Another important issue is that observations of SWE are representative of only a small subset of the spatial domain, and relatively few observations are available at very high elevations or in the transient snow zone where winter snowpacks are ephemeral. These areas that are not well observed are a dominant fraction of the land area and of the SWE contributing to streamflow in the mountainous western United States and, in the case of the transient snow zone, they represent one of the most sensitive areas to warming. Thus a comprehensive understanding of long-term changes in western U.S. snowpacks and associated water resources impacts requires attention to not only a period of record longer than the last 50 yr, but to parts of the domain not covered by the snow course observations, the locations of which were dictated by short-term water supply forecasting needs, rather than detection of the impacts of climate variations.

In addition, snow course observations lack temporal resolution (they are typically reported only on the first of the month) and are unavailable before midwinter at most stations. Automated observations of SWE via the Natural Resources Conservation Service Snowpack Telemetry (SNOTEL) network are available at daily time steps throughout the snow season, but the longest records available for these observations go back only slightly more than 20 yr in all but a few isolated cases. Thus observations alone cannot provide useful information about long-term trends in patterns of snow accumulation and melt (such as date of peak accumulation, date of 90% melt, etc.).

While the use of model simulations introduces uncertainties into the analysis that are not present in observational studies, these uncertainties can be minimized if the model forcing data are carefully quality controlled, and the use of model simulations avoids a number of important limitations in the observed datasets discussed above. Here we examine simulations of spring SWE on 1 March, 1 April, and 1 May, with particular attention to the critical 1 April date, which corresponds roughly to the date of maximum snow accumulation over much of the mountainous western United States. By isolating the effects of temperature and precipitation trends in the model’s forcing dataset in separate model runs, we also use the model to ex-
amine explicitly the effects of temperature and precipitation variability on SWE trends. The role of decadal-scale variability on SWE trends is also examined by selecting periods of record for analysis that reflect known shifts in decadal variability associated with the PDO (Mantua et al. 1997).

It is worth noting here that the trends in precipitation in the PNW appear to be consistently related to the PDO time series (i.e., dry to wet from 1925 to 1976 associated with warm to cool PDO; and wet to dry from 1947 to 2003 associated with cool to warm PDO; Hamlet and Lettenmaier 1999a; Mote et al. 2003; Mote 2003a), whereas a consistent relationship between precipitation trends and the PDO over time is not necessarily apparent in other parts of the domain, perhaps most notably in the Colorado River basin (Hidalgo and Dracup 2003).

The questions that motivate this paper are 1) To what extent are observed trends in western U.S. snowpack attributable to precipitation and temperature trends, and how do these effects vary with region, topography, climate, and the time period examined? 2) What areas of the western United States show the greatest trends in simulated snowpack, and what are the specific climatic regimes associated with these areas? 3) What is the role of decadal climate variability in determining the observed snowpack trends? 4) What changes in the timing of snow accumulation and melt are apparent and how do they vary with region, topography, and climate?

2. Hydrologic model and meteorological driving data

a. VIC hydrologic model

We use the Variable Infiltration Capacity (VIC) hydrologic simulation model (Liang et al. 1994; Cherkauser and Lettenmaier 2003) implemented at 1/8° (latitude–longitude) spatial resolution over the western US and southern BC. The 1/8° grid cells are roughly 12 km by 10 km, varying in size to some degree with latitude over the domain. Daily SWE for each grid cell (which represents the areal average SWE for the mosaic of elevation bands and vegetation types within the cell) is the primary output used in this study. SWE simulations from each elevation band were also used for comparison with point data from the snow course records, although these results are only briefly discussed here. We do not examine other variables from the simulations in this study, but the model also simulates a detailed water balance in each cell. Simulated daily runoff from the model produced by the runs in this study will be used in subsequent studies to examine changes in the water balance associated with changes in snow accumulation and melt. The VIC model has been used in numerous climate studies of large river basins around the world (e.g., Nijssen et al. 2001), and has been well validated with streamflow observations, particularly in the mountainous western U.S. (Christensen et al. 2004; Hamlet and Lettenmaier 2005, 1999b; Maurer et al. 2002; VanRheenen et al. 2004). The model has also been used extensively for streamflow forecasting applications (Hamlet and Lettenmaier 1999a; Hamlet et al. 2002; Wood et al. 2002) and for climate change assessments (Christensen et al. 2004; Hamlet and Lettenmaier 1999b; Payne et al. 2004; Snoover et al. 2003; VanRheenen et al. 2004; Lettenmaier et al. 1999). For this study we primarily make use of the detailed energy balance snow model incorporated in the VIC model (Cherkauser and Lettenmaier 2003). The snow model is well suited for simulating mountain snowpack and includes the effects of forest canopy on snow interception and the attenuation of wind and solar radiation, which are fundamental drivers of snow accumulation, sublimation, and melt processes (Storck 2000; Storck et al. 2002).

b. Meteorological driving data

For this study, a new climate dataset was developed following methods outlined by Hamlet and Lettenmaier (2005). This dataset was intended specifically for application to problems where careful quality control to avoid spurious trends in long records of daily temperature and precipitation is necessary. Daily values of maximum temperature, minimum temperature, and total precipitation were obtained for 1915–2003 at the 1/8° latitude–longitude spatial resolution by interpolating daily values from National Weather Service Cooperative Observer stations using the Symap algorithm (Shepard 1984) following methods described in Maurer et al. (2002). Temperatures were lapsed by 6.1°C km⁻¹ (i.e., the theoretical moist adiabatic lapse rate) during the interpolation process, and precipitation was scaled for topographic effects using monthly precipitation maps produced by the Precipitation Regression on Independent Slopes Method (PRISM) algorithm (Daly et al. 1994). Maurer et al. (2002) describe these data-processing steps in more detail. Temporal inhomogeneities (e.g., inconsistencies in time due to changing station groups, station moves, changes in instrumentation, urban heat island effects, etc.) in the regridded time series were adjusted for each calendar month using regridded and temporally smoothed data from the U.S. Historical Climatology Network and the monthly Historical Canadian Climate Database. Additional technical details are available in Hamlet and Lettenmaier (2005).
Despite the fact that the model forcing data are carefully quality controlled, there are inevitably some limitations associated with the approach. One is that the gridded (precipitation and temperature) forcing data are derived primarily from low or moderate elevation stations. Because the trends in the temperature and precipitation data are ultimately derived from the time series of these low elevation stations, there is no explicit information contained in the driving dataset regarding potentially different trends at very high elevations. In some parts of the domain, the primary driving data are also quite sparse, which may result in gridded data that may artificially suppress spatial variability, especially at short (e.g., daily) time steps. Topographic adjustments to the precipitation and temperature data (discussed above) likewise tend to suppress spatial variability in the gridded product. These limitations in the forcing data, while important for high-resolution modeling applications, are generally of secondary importance here. One major reason is that snow accumulation and ablation in mountainous regions is dominated by the characteristics of accumulated precipitation in snowpacks, and issues associated with suppression of spatial variability tend to be filtered out. In any event, the model is able to reproduce observed trends in spring snowpack quite well at a macroscale over the western United States (as discussed below); hence it meets the requirements for a study such as ours that focuses on macroscale aspects of the snow accumulation and ablation process.

c. Role of temperature and precipitation driving data in the snow simulations

The relationship between hourly snowfall and precipitation estimates in the model forcing data is fairly straightforward. The phase of the precipitation is determined by a simple partitioning scheme based on estimated hourly time step temperatures (derived from Tmax and Tmin) during the simulation. Temperatures below −0.5°C are assumed to result in precipitation that is 100% snow, those above 0.5°C are assumed to result in 100% rain, and a linear relationship is assumed between these two extremes.

Temperature driving data play a much more complex role in the simulations. Aside from the direct effects of temperature on convective heat transfer from the air to the snowpack (e.g., Storck 2000), temperature also affects snow accumulation and melt in the model due to meteorological variables that are derived from the temperature data. The difference between Tmax and Tmin determines the attenuation of solar radiation and variations in long-wave radiation due to cloudiness, for example, and Tmin is used in calculating the dewpoint and vapor pressure deficit which influence snow sublimation (Thornton and Running 1999). Thus trends in temperature are also indirectly related to trends in long- and shortwave radiation, humidity, and vapor pressure contributing to sublimation and snowmelt in the model. When we refer to “temperature-related trends” in later sections of the paper, we are referring to trends associated with these indirect effects as well.

d. Evaluation of simulated SWE trends

In corroborating trends in observations with the model results from this study, Mote et al. (2005) show remarkable broad-based agreement over the western United States between observed trends in 1 April SWE from 1950 to 1997 and trends derived from VIC simulations of SWE for the same period. Although some differences in the absolute value of relative trends are apparent, it is clear that the model successfully captures the large-scale characteristics of the spatiotemporal trends of snow accumulation and melt over the western United States. It is largely on the basis of this comparison and the longer streamflow comparisons discussed by Hamlet and Lettenmaier (2005) that we argue that the model results provide a reasonable surrogate for observations from 1916 to 2003 and are suitable for a trend analysis at these spatial scales.

It should be noted however that on a point-by-point basis, there are often substantial differences between the SWE observations analyzed by Mote et al. (2005) and the VIC simulations, both in absolute value and in relative trends. These differences between simulations and observations are primarily due to the fundamental differences in spatial scale between observations and simulations, and to frequent discrepancies between the actual precipitation and temperature at snow course sites (a point within the VIC cell) and the gridded meteorological driving data used in the VIC simulations. Snow courses are also generally located in open areas, which is inconsistent with the VIC simulations in parts of the domain with substantial forest canopy.

The issues associated with the model forcing data are to be distinguished from validation of the VIC snow model itself. When the model is driven by accurate temperature, precipitation, and vegetation characteristics (i.e., in locations where there is a nearby meteorological station and detailed information about the vegetation), the model very closely reproduces daily snow accumulation measurements recorded at these same locations (see, e.g., Storck 2000). Furthermore, additional analysis (not shown) found good general agreement between monthly VIC snow accumulation and melt statistics (e.g., date of peak snow accumulation, volume of accumulation and melt in each month, etc.) and those
extracted from daily time step observations from SNOTEL stations. These comparisons with the SNOTEL observations help to confirm that the VIC model captures the primary physical mechanisms associated with winter climate and topography that ultimately determine trends in mountain snowpack.

Mote et al. (2005) also examined the model’s ability to capture the relationship between winter temperature regimes and trends in 1 April SWE in the observations from 1950 to 2003. The model successfully reproduces this relationship over most of the western United States; however, in some specific parts of the domain the model does display some bias. There are, for example, some apparent systematic errors in the trends in high-elevation precipitation in the Sierra Nevada Mountains in the model forcing data that result in a general bias toward smaller simulated upward trends in the high-elevation areas in the southern part of the domain, and there are larger downward trends in simulated SWE at moderate elevations in California (CA) from 1950 to 1997 than are apparent in the observations. In the case of the lower-elevation sites in California, this problem might be partly attributable to sampling bias in the observations, or to inaccurate temperature lapse rates in spring when the weather is clear (i.e., the model forcing temperatures are probably biased toward warmer temperatures on clear days at high elevations because of the assumption of a fixed pseudoadiabatic lapse rate; see Hamlet and Lettenmaier 2005).

3. Methods and experimental design

The VIC model (version 4.0.5) was run from 1915 to 2003 at 1/8° resolution over the 16 526 grid cells that make up the continental United States west of the Continental Divide, plus the portion of the Columbia River basin that lies in southern British Columbia (see Fig. 1). In the following, we will refer to the PNW (containing the Columbia River basin and coastal drainages), California (CA), Colorado River basin (CRB), and Great Basin (GB) as subregions of this domain (see Fig. 1). The snow model time step was 1 h, which was required to capture the effects of the diurnal cycle of temperature and solar radiation on snow accumulation and
melt. In the absence of shorter time step observations, daily total precipitation was equally distributed throughout the 24 h of the day. To represent subgrid topographic variability, the model uses up to five equal-area elevation bands with a vertical interval of approximately 500 m. Simulated SWE values are reported as the area-weighted average SWE over all the elevation bands in each cell, each of which includes a vegetation mosaic comprised of up to 10 vegetation types. The vegetation characteristics of each cell were assumed to be stationary with time in the simulations. Note that the SWE values reported from the model represent the average snow water content over the entire grid cell, which can be quite different from the value that would be simulated for any particular point location within the cell. Elevation bands in each grid cell with July mean temperatures below 10°C were assumed to be above the tree line (Körner 1998), and vegetation classes with an overstory (if any) were removed from these bands during the simulation. The first nine months of the simulation (1 January 1915–30 September 1915) were used as model spinup and were discarded. This spinup is adequate for snow simulations, for which there is essentially no year-to-year carryover effect. Figure 1 shows the 1/8° digital elevation model for the simulation domain, average December–February (DJF) temperatures for each grid cell, and average November–March precipitation summarizing the model forcing data.

Three different model runs were made:

1) Base run: Unperturbed precipitation and temperature data from the meteorological input files (daily total precipitation and minimum and maximum temperature and wind speed) were used to drive the model. Trends in SWE are the result of both temperature and precipitation variations.

2) Fixed precipitation (fixed P) run: The precipitation forcing data were fixed (for each calendar month) at the climatological value for each grid cell, but daily temperature was allowed to vary as in the original time series. Trends in SWE are the result of trends in temperature and other temperature-related meteorological variables alone.

3) Fixed temperature (fixed T) run: The temperature forcing data were fixed (for each calendar month) at the climatological value for each grid cell, but daily precipitation varied as in the original time series. Trends in SWE are the result of precipitation trends alone.

For the fixed P and fixed T runs, the forcing data were perturbed as follows: For each grid cell, a monthly climatological value for either precipitation or temperature was calculated for each calendar month (i.e., a separate climatology for January, February, etc.). Then the daily time series of the variable to be held constant was forced to reproduce this climatological value in each month of the simulation—that is, the daily values were allowed to vary within the month as in the original time series, but the monthly totals (precipitation) or averages (maximum and minimum temperature) in each calendar month are identical in each year of the simulation. This method preserves the daily covariance between temperature, precipitation, solar radiation, etc., while removing the trends and monthly variation in the fixed variable from the simulation. Note that an average seasonal cycle remains in the fixed variable after this adjustment. Wind data were allowed to vary as in the original time series in each case. If precipitation and minimum and maximum temperature were all held fixed as described above, the mean of the resulting simulated SWE would be comparable (but not equal) to the long-term mean for each calendar date in the base simulation, and any trends in the simulated SWE would be small. It should be noted that the timing of precipitation within the month is unaffected by these adjustments. This has some important implications when evaluating the fixed P trends in the calendar date of 10% snow accumulation discussed in later sections, for example.

The 1 March, 1 April, and 1 May SWE values were extracted from the model output files for four time periods based on warm and cool PDO epochs defined by Mantua et al. (1997) as follows:

A) 1916–2003 (full period)
B) 1925–76 (warm PDO to cool PDO)
C) 1947–2003 (cool PDO to warm PDO)
D) 1925–46 concatenated with 1977–2003 (warm PDO to warm PDO)

Linear trends were calculated for the time series for each grid cell and were normalized by the long-term mean of the simulated 1 April SWE for each cell over the time period of analysis (i.e., the trends for scenario C were normalized by the mean SWE calculated from the 1947–2003 data). Cells included in the plots were required, on average, to have at least 50 mm of SWE on 1 April in order to avoid including spurious trends in the analysis.

For each grid cell, a time series of the date of peak SWE [day 1 = 1 September, day 365 (or 366 in leap years) = 31 August] was extracted from the daily time series of simulated SWE, as well as the day associated with 10% of peak accumulation and 90% melt. Linear trends were calculated for each of these variables from 1916 to 2003 for the base, fixed P, and fixed T simulations.
4. Results and discussion

a. Notes on presentation of results and interpretation of figures

The results in this section will be summarized primarily using spatial plots of the trends in SWE in each grid cell and scatterplots showing the relationship between average midwinter (DJF) temperatures in each cell and the trends in SWE. The dots in the scatterplots are color coded so that the different regions in the domain can be distinguished. By comparing the patterns in these figures for the base, fixed $P$, and fixed $T$ runs, qualitative conclusions about the effect of temperature and precipitation trends on the overall trends can be drawn. DJF average temperatures for each grid cell used in the scatterplots of trend versus winter temperature were calculated from the VIC driving data from 1916 to 2003 and are independent of the period of analysis used.

b. SWE trends for 1 March, 1 April, and 1 May for 1916–2003

To begin with, we show overall trends in simulated SWE for 1 March, 1 April, and 1 May for the base run from 1916 to 2003 (Fig. 2). As noted above, only grid cells with average SWE greater than 50 mm on 1 April are shown. The results show downward trends over many grid cells in the domain, and a relationship between midwinter temperatures and the relative trends in SWE characterized by an inverted “J” shape in the scatterplots. As we shall see in subsequent sections, this relationship is characteristic of downward trends in
snowpack associated with warming. Downward trends in SWE on 1 April and 1 May are more widespread than for 1 March. A substantial part of the domain shows upward trends in SWE, particularly over much of the Columbia River basin. As we shall see in subsequent sections, these upward trends in SWE are primarily due to upward trends in precipitation. The increased scatter for 1 May that is evident in Fig. 2 is probably due to the fact that many grid cells have relatively small amounts of snow remaining on 1 May.

c. Results for 1916–2003

This period is characterized by widespread, modestly upward trends in precipitation that result in upward trends in SWE in the fixed $T$ simulations (Fig. 3c) and strong downward trends in SWE due to upward temperature trends over essentially the entire domain in the fixed $P$ simulations (Fig. 3b). Thus the majority of the downward trends in SWE from 1916 to 2003 are attributable to large-scale warming, which overwhelms the effects of widespread increases in winter precipitation. Note, for example, the predominantly upward trends in grid cells in CA associated with precipitation (Fig. 3c), but predominantly downward trends for CA in warmer areas with DJF temperatures above about $-2.5^\circ C$ (Fig. 3a). Several distinct climatic regimes are also apparent in these results, which will be discussed separately below.

d. Results for 1924–76

Strong upward trends in precipitation throughout the region dominate the SWE trends in this period (Fig. 4c), and upward trends in SWE follow these precipitation trends over most of the domain (Fig. 4a). The overall trends in precipitation appear to reflect a shift from widespread drought from 1925 to 1946 to wetter conditions overall in the second half of the period. The fixed $P$ analysis (Fig. 4b) shows that there is not a consistent trend in SWE due to temperature alone, and this is the only period examined for which there were substantial upward trends in SWE associated with downward trends in temperature (and solar radiation) alone.

e. Results for 1947–2003

The 1947–2003 period of analysis corresponds most closely with the time periods covered by the snow course and streamflow records that were discussed in the introduction. This period is characterized by different precipitation trends in different regions (Fig. 5c) and by strong downward trends in SWE associated with temperature trends (Fig. 5b). The trends in precipitation in each region are broadly consistent with patterns of observed climate variability associated with the PDO since 1947 that were discussed in the introduction. The fixed $P$ analysis for this period (Fig. 5b) shows that without precipitation trends, essentially the entire domain would have experienced strong downward trends in 1 April SWE. Because the temperature-related trends are almost always downward for this period, areas with downward precipitation trends (e.g., most of the PNW cells with DJF temperatures above $-5^\circ C$) exhibit very strong downward trends in SWE. Some areas with upward precipitation trends (e.g., about half the cells in CA) show downward SWE trends due to temperature effects despite increases in precipitation. This effect is less evident in the colder areas of the domain (e.g., the CRB) but can still be seen in the base run (Fig. 5a) as a shift toward stronger downward trends in SWE when the effects of temperature trends are combined with the effects of precipitation trends. Trends in the base run analysis (Fig. 5a) are less than $-0.25\%$ per year (strong downward trends) over about 70% of the cells shown in the figures. This is a considerably larger fraction of the total area with downward SWE trends than for the base run for 1916–2003, in part because of widespread downward precipitation trends in the colder areas of the PNW from 1947 to 2003.

f. Results for 1924–46 concatenated with 1977–2003

The intent here is to minimize the effects of decadal variability associated with the PDO by combining two climatologically similar epochs. The fixed $P$ analysis for this period (Fig. 6b) shows somewhat more scatter than the fixed $P$ analysis for 1947–2003, but again the trends associated with temperature alone are overwhelmingly downward. The fixed $T$ analysis (Fig. 6c) shows little consistent trend in SWE associated with precipitation trends in comparison with the 1916–2003 or 1925–76 periods. As in the 1916–2003 and 1947–2003 periods, a major driver of the SWE trends in this period is a large-scale warming that affects most of the domain, and the analysis suggests that this warming cannot be readily explained by decadal-scale variability associated with the PDO. In particular, the earlier warm phase PDO epoch was clearly much cooler overall than the most recent warm phase epoch. Further evidence that decadal variability is a poor explanation for the temperature-related trends can be seen by comparing Figs. 3b, 5b, and 6b. It is apparent that any period paired with the 1977–2003 period shows dramatic downward trends in SWE due to temperature trends alone.

g. Effects of climatic regimes

Three distinct climatic regimes are apparent in the areas that experienced downward trends in 1 April
SWE from 1916 to 2003 (Fig. 7). Areas with large downward trends in 1 April SWE and DJF temperatures between $-1.0^\circ$ and $+2.5^\circ$C are shown to be coastal areas in the PNW and northern CA (Fig. 7, top right). Areas with large downward trends in 1 April SWE and DJF temperatures less than $-1.0^\circ$C are shown to be inland areas with a more continental climate (Fig. 7, middle right and lower right). These differences have to do with specific interactions between precipitation and temperature regimes that determine
the time scales of snow accumulation and melt. Coastal mountain ranges are warmer but are able to produce very large snowpacks in midwinter that can persist until 1 April because of high mean precipitation and cool, cloudy spring conditions. These areas are affected by temperature increases from midwinter through spring (which also explains why they are most sensitive to warming). Continental areas, by contrast, are typically dryer and colder, and snow accumulation that ultimately persists until 1 April takes place over a number of months from late fall to early spring. Analysis of winter precipitation and the seasonal cycle of snow ac-

Fig. 4. Same as in Fig. 3, but for 1925–76.
cumulation in these areas (not presented here) showed that precipitation trends over the entire winter and warmer temperatures during the late spring were the predominant drivers of the overall trends in SWE. For areas with cold midwinter temperatures (Fig. 7 lower right), the temperature sensitivity becomes relatively small and the large downward trends in SWE are associated primarily with strong downward trends in winter precipitation (cf. Fig. 7 with Fig. 3c). These climatic regimes are consistent with climatic characteristics observed by Serreze et al. (1999) in SNOTEL observations.

Fig. 5. Same as in Fig. 3, but for 1947–2003.
h. Trends in the timing of snow accumulation and melt

Figure 8 shows the trends in the date of simulated snow accumulation and melt. Trends in the date of simulated peak snow accumulation and the date of 90% melt are strongly related to DJF average temperatures. Almost all the cells with winter temperatures warmer than $-5^\circ C$, for example, show earlier dates of peak accumulation and 90% melt. In sensitive areas (such as the coastal areas in the PNW and CA shown in Fig. 7), peak accumulation occurs between 15 and 45 days earlier, and 90% melt occurs 15–40 days earlier. Comparison of the base run scatterplots with those from the
fixed $P$ and fixed $T$ runs (Fig. 8) shows that the changes in the timing of peak snow accumulation and 90% melt are a complex function of precipitation and temperature changes, but that the dominant effect is due to temperature trends. These changes in snow accumulation and melt are consistent with observed changes in streamflow timing discussed in the introduction. A similar examination of the scatterplots shows that trends in the date of 10% accumulation are predominantly due to upward trends in precipitation in the beginning of the snow accumulation season. Interestingly, the time between 10% accumulation and 90% melt has not changed very much overall. This somewhat counterintuitive result occurs because the 10% accumulation dates are most affected by precipitation trends (cf. base and fixed $T$ runs in Fig. 8), whereas the peak SWE and 90% melt dates are most affected by temperature trends (cf. base and fixed $P$ runs in Fig. 8). Despite different mechanisms at different times of the year, the overall trends in the date of snow accumulation and melt are characterized by shifts in the entire snow accumulation season earlier in time, but with a smaller amount of peak SWE.

The results for the four time periods are summarized for different portions of the domain and the different model runs in Tables 1 and 2. Table 1 shows the areal average of the relative trends calculated for each cell (i.e., the spatial average of the trends shown in Figs. 7 and 8).
Table 2 shows the trend in the areal average SWE (i.e., a time series of SWE averaged over each domain is extracted first, and the trend in this aggregate time series is reported).

The results for 1925–76 shown in Table 2 are somewhat confusing and require some additional explanation. For CA, for example, the fixed $P$ and fixed $T$ runs for this period show small downward trends in areal

Fig. 8. Cumulative changes (trend × number of years) in calendar date of (a) 10% SWE accumulation, (b) maximum SWE accumulation, and (c) 90% melt for 1916–2003 for three model runs. (top) Spatial plot for the base run, and (bottom three) scatterplots for the base (B), fixed $P$ (FP), and fixed $T$ (FT) runs, respectively.
average SWE, yet the trend in the base run has a strong upward trend. These counterintuitive effects are associated with the spatial distribution of the upward and downward trends associated with temperature and precipitation trends. For example, if specific areas in the domain that are getting wetter are also getting colder, then the overall trend may be strongly upward (as for CA for this period), whereas if areas that are getting wetter are also getting warmer there may be little overall trend. These complex effects are not seen in the other time periods because the temperature-related trends (and in many cases, the precipitation related trends as well) are much more spatially homogeneous.

5. Summary and conclusions

Widespread warming has occurred in the western United States from 1916 to 2003, resulting in downward trends in 1 April SWE over large areas of the domain. High-elevation areas with upward precipitation trends, however, are shown to produce upward trends in SWE over the same time period. The results show that almost all the upward trends in SWE from 1916 to 2003 are due to modest upward precipitation trends and that many of the downward trends in SWE are caused by widespread warming. In areas with relatively warm winter temperatures, such as coastal areas of the PNW and CA, the effects of warming frequently overwhelm the effects of increasing precipitation. Colder areas are predominantly driven by precipitation changes, and temperature effects on SWE trends, while apparent in the combined trends, are relatively small.

The period from 1925 to 1976 was characterized by widespread increases in precipitation, which dominated temperature trends to produce widespread upward trends in SWE. The period from 1925 to 1976 was also the only period in which widespread upward trends in SWE are observed due to temperature trends alone. From 1947 to 2003, regionally specific precipitation trends associated with decadal variability combine with strong downward trends in SWE due to widespread warming. The effects of warming are frequently large enough to overwhelm upward trends in SWE associated with precipitation trends alone in warmer areas of the domain.

The results for the back-to-back warm PDO epochs combined with the 1916–2003 and 1947–2003 analysis described above strongly suggest that widespread regional warming trends are not well explained by decadal-scale variability, but that the decadal variability probably does account for the trends in winter precipitation that have been apparent over shorter periods of the record. Although the precipitation trends from 1916 to 2003 are broadly consistent with many global-warming scenarios, it is not clear whether the modestly increasing trends in precipitation that have been observed over the western United States for this period are primarily an artifact of decadal variability and the time period examined, or are due to longer-term effects such as global warming.

Several distinct climatic regimes exist in areas that have experienced downward trends in 1 April SWE in the simulations. Coastal areas are strongly affected by warming throughout the winter and spring, whereas areas with a more continental climate are more sensitive to precipitation trends during the winter and to warming in late spring. Many high-elevation areas in the Rockies and southern Sierra Nevada are relatively insensitive to temperature trends, and downward trends in SWE are primarily due to downward trends in precipitation.
The dates of peak snow accumulation and 90% (of peak) melt have generally been occurring earlier in the year, and these trends are clearly sensitive to winter temperature regimes, with the greatest changes apparent in areas with warmer winter temperatures (e.g., near-coastal mountains in the PNW and CA). These effects are consistent with the observed trends toward earlier peak snowmelt runoff. The sensitivity analysis shows that the changes in the timing of peak accumulation and 90% melt are primarily a temperature-related effect. The date of 10% (of peak) accumulation has also trended earlier in the year but is shown to be related primarily to trends in fall precipitation rather than to trends in temperature.

The results of this study strongly suggest that the effects of global warming are apparent in the surrogate observations examined in this study. Furthermore, because the trends in SWE are most strongly affected by trends in temperature, we should expect, based on projections of continued warming, that these trends will continue. Because precipitation variability seems most strongly associated with decadal variability rather than long-term trends, the use of observed precipitation variability in conjunction with scenarios of warmer temperatures may currently be the best approach for understanding the overall effects of global warming on the hydrologic variability of the western United States. Such an approach implies, as well, that both “warm and wet” and “warm and dry” periods are likely to occur in the future at different times, and that water resources planning should consider both scenarios in testing alternative management plans.

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REFERENCES


