A framework for the study of seasonal snow hydrology and its interannual variability in the alpine regions of the Southwest

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Abstract. A framework is developed to study variations in winter season precipitation and snow processes and spring season runoff in the alpine basins of the southwestern United States. The framework exploits available high-elevation precipitation and snow water equivalent (SWE) data, and a regional climate model (RegCM) to study two contrasting winter/spring seasons (1994 and 1995). This work examines the influences of large-scale atmospheric conditions on cold season hydrology in the Southwest, assesses the capability of the regional climate model for simulating snowpack and runoff in two major alpine basins and evaluates methodologies for employing high-elevation observations on regional spatial scales and seasonal time-scales. High elevation data and streamflow measurements from the winter/spring of 1995 show relatively wet conditions, larger snow accumulations and more substantial springtime streamflows in the Rocky Mountains and Sierra Nevada than observations from 1994. Analysis of the large scale atmosphere and surface station data during these periods reaffirms that the wintertime quasi-stationary wave across North America is an important modulator of precipitation, snowpack, and streamflows in the west. The RegCM simulates the large-scale circulations in these 2 years well. Errors are apparent in the simulated precipitation fields and are greater when comparing modeled and observed snowpack and runoff. While the simulated snowpack/streamflow errors are controlled by precipitation deficiencies in the Rocky Mountain region, in the Sierra Nevada region the errors are determined by elevation and associated temperature errors. Despite an underprediction of snowpack and runoff, the model is able to reproduce the overall patterns of precipitation and snowcover and to some extent the year to year hydrologic variations. Continued time series of the data employed in this study (high-elevation precipitation and SWE and streamflow measurements) are crucial for seasonal and interannual studies of water resources in the west.

1. Introduction

The American Southwest, with a fast growing population and arid climate, depends on seasonal alpine snowpacks for a significant fraction of its annual water needs. Snowmelt runoff provides much of the 45% of annual wa-

ter supplies that are from surface flows. It also provides recharge of groundwater aquifers which supply ≈55% of the annual water needs [Maddock and Hines, 1995]. The largest surface flows in the Southwest are the Colorado River, the Rio Grande, and several river basins in California; all have their origins in high elevations in the Rocky Mountains or the Sierra Nevada. Water management in the Southwest depends on skillful forecasts of year-to-year variations in large-scale basin runoff and the high-elevation snowpack that supplies these major rivers.

The primary precipitation peak is in the winter months throughout the west, and the annual cycle of streamflow
in the major rivers shows a pronounced peak in spring and summer [Cayan and Peterson, 1989]. The time lag between peak precipitation and peak streamflow results largely from water stored as snow at higher elevations. Several observational studies have related seasonal anomalies in streamflow and snowpack to large-scale circulation anomalies over the North Pacific and the western United States. Cayan and Peterson [1989] find that west coast streamflows are influenced by local sea level pressure (SLP) anomalies in the North Pacific, while streamflows in the interior west are affected by more remote SLP anomalies in the western North Pacific. Redmond and Koch [1991] relate regional temperature, precipitation, and streamflow anomalies to the Pacific North American (PNA) index and the Southern Oscillation Index (SOI). Cayan [1996] shows that a smaller western snowpack is associated with a negative phase of the PNA index. Another study by Cayan et al. [1993] isolates streamflow dependence on temperature and precipitation and shows that winter precipitation has the greatest influence on springtime flows. They find that temperature is important in determining streamflow at higher elevations only in springtime and that both precipitation and temperature are controlled by atmospheric circulation patterns which occur on a much larger spatial scale than individual basins. These atmospheric circulation anomalies may be predictable with some skill on seasonal time-scales.

In order to examine the potential of climate models for simulation of winter season precipitation, topography that is more highly resolved than commonly used for global climate models (≈200 km resolution) is needed. Hence it is appropriate for this purpose to use regional climate models (RegCMs). These models employ resolutions from 40 to 200 km, are run for specified limited area domains, and are driven by time-dependent large-scale meteorological fields in a buffer area adjacent to the domain’s lateral boundaries. The driving fields can be provided either by analyses of observations or by output from general circulation model (GCM) simulations. Lee and Chan [1995] have employed a limited area model on a small regional scale for the northwest to study precipitation and snow processes in mountainous areas. Their subgrid parameterization for precipitation incorporates simple airflow and thermodynamic models, and subgrid calculations are performed for a distribution of elevation classes.

Recent results suggest that larger domains are needed to evaluate how well RegCMs can simulate land hydrological processes. Seth and Giorgi [1998] have shown that choice of regional model domain can have a significant impact on simulation results: a small domain constrains the solution and prohibits feedback to the larger-scale atmosphere. A regional model domain should consider the processes being studied and be large enough to reduce imbalances between the internal simulation and the external forcing. Seasonal forecasts will require the use of lateral boundary forcing derived from ensembles of simulations with a global model. Global model simulations can at best describe large-scale atmospheric anomalies and generally do not perform as well on small regional scales. Thus boundary conditions provided to a small regional model domain by a GCM are less likely to be skillful than boundary conditions provided to a large regional model domain. In this study the regional model simulates the continental-scale climate and is influenced by topography, land use, and soil moisture on spatial scales not represented in the GCM. Given this choice of domain size, horizontal resolution is limited by computational costs. The National Center for Atmospheric Research (NCAR) RegCM domain encompasses the United States; and a horizontal resolution of 60 km is employed. Snow hydrology and runoff in the RegCM are represented using the Biosphere-Atmosphere Transfer Scheme (BATS) [Dickinson et al., 1993], which simulates surface physics at each model grid cell. The boundaries of the limited area domain are driven by analyses for the period March 1, 1993, through September 30, 1996. The simulation can thus be verified against observed atmospheric and surface data sets.

This paper builds a framework to study cold season hydrology (precipitation and snow processes) and its variations in the mountain west and springtime runoff in the two major basins that supply surface flow to the Southwest. Because circulation anomalies are larger in spatial scale than individual basins, a continental-scale perspective is taken. A downscaling approach is employed, using a regional climate model combined with data from global analyses, continental-scale networks of station temperature and precipitation, high-elevation snow courses, airborne and satellite derived maps of western snow cover, and stream gauges. Our goals are (1) to examine the relationship between streamflow anomalies in the Southwest and large-scale circulation anomalies, (2) to evaluate the NCAR RegCM, in its current state, for seasonal prediction of winter/spring hydrology in the mountain west, and (3) to investigate the use of high-elevation data for validation of the RegCM.

2. Methods and Data

The methodology employed in this study is as follows: Two contrasting winter seasons (January-March 1994 and January-March 1995) are studied using observations and a RegCM simulation for the observed period March 1, 1993 - September 30, 1996. Large-scale circulation and moisture transport differences between these 2 years provide the starting point for the analysis. Seasonal precipitation totals are next examined using high-elevation station data. Modeled and observed snow water equivalent (SWE) are presented descriptively on a continental scale for the two years. Our analysis then narrows to focus on two high-elevation basins which supply water to the Southwest, the upper Colorado River basin (hereafter referred to as the Rocky Mountains region) and a combined basin consisting of 10 smaller river
basins in the Sierra Nevada (hereafter referred to as the Sierra Nevada region). Time series of basin-averaged precipitation, temperature, and snow water equivalent are examined to determine their influence on the time series of basin-scale runoff.

Several datasets are used in this study. Large-scale analyzed atmospheric fields for the period of the simulation were taken from the European Centre for Medium Range Weather Forecasting (ECMWF), as processed by Trenberth [1992]. The atmospheric fields are available at 6-hour intervals and are employed both as driving fields for the boundaries of the regional climate model and as verification data to compare with the regional model generated fields.

Monthly precipitation and surface air temperature data are taken from the National Climate Data Center (NCDC) U.S. Historical Climatology Network (USHCN) [Karl et al., 1990], which includes 1000 stations across the continental United States. For comparison with the model results the station data are objectively analyzed onto a 1/4 degree grid and then bilinearly interpolated onto the 60-km model grid. An elevation correction is made to the surface temperature data using the standard lapse rate (−6.5 °C/km) and the difference between mean station elevation and the mean elevation of each 1/4 degree grid cell.

In addition to the NCDC precipitation data, high-elevation daily precipitation is obtained from the USDA Natural Resources Conservation Service, Snow Survey through the Western Regional Climate Center. We employ 580 SNOTEL sites (remote stations which employ automated sensors and radio telemetry of data) providing accumulated daily precipitation data throughout the western United States. Monthly values are computed for each station for January-February-March (JFM) 1994 and 1995. Combining the SNOTEL and NCDC precipitation data is not an entirely straightforward process. Because in the west there are more SNOTEL stations than NCDC station, if the two data sets are simply combined in the objective analysis onto the 1/4 degree grid, the resulting observed precipitation will be skewed toward the high-elevation data. In order to combine the data in a physically meaningful way the total precipitation \( P_t \) in a 1/4 degree grid cell is a weighted sum of the NCDC and SNOTEL data after each has been objectively analyzed onto the 1/4 degree grid, separately. For each 1/4 degree grid cell this is given by

\[
P_t = F_e P_{\text{SNOTEL}} + (1 - F_e) P_{\text{NCDC}},
\]

where the weight \( F_e \) is the fraction of the grid cell above a critical elevation \( H_c \) as determined by a high resolution (30 s) elevation data set. \( P_t \) is sensitive to the choice of \( H_c \). Station elevations vary between 1700 m at higher latitudes and 3100 m at lower latitudes. We assume that the SNOTEL precipitation data are representative at elevations down to 500 m below the mean station elevation \( H_s \) within a 1/4 degree grid cell, so that \( H_c = H_s - 500 \). The sensitivity of resulting precipitation to this determination of \( H_c \) is discussed in section 4. Thus the total precipitation for a 1/4 degree grid cell is a weighted sum of the SNOTEL and NCDC data. This combined product is then bilinearly interpolated to the 60-km regional model grid.

Since 1990 the National Operational Hydrologic Remote Sensing Center (NOHRSC) has been generating snow water equivalent (SWE) maps for the western United States. The SWE product is a 2-5 day composited field derived from station observations (including SNOTEL sites), which report SWE and snow depth, and satellite imagery (advanced very high resolution radiometer (AVHRR) and Geostationary Operational Environmental Satellite (GOES)) which give information on extent of snow cover. These maps are used for visual comparison with the model-generated SWE, and regional means are computed for the analysis of JFM time series.

Monthly streamflow data are from two sources. For the upper Colorado River basin, monthly data from two U.S. Geological Survey gauges, located on the Green River and the Colorado River just upstream of their confluence, are used. These two gauges effectively integrate the upper part of the basin and are upstream of the major regulations on these rivers. Ten river basins are used to integrate the Sierra Nevada. The California Department of Water Resources (DWR) provides monthly reconstructed (or full) natural flows (FNF) for each of the 10 basins. The total monthly runoff is computed by combining the flows for the basins in the Sierra Nevada and in the Colorado River basin.

The regional model domain includes the continental United States (see Figure 1, which shows the domain and contours of model topography) and has a horizontal grid resolution of 60 km. The 60-km resolution provides more realistic topography than most current GCMs (which use ≈200-km resolution). Figures 2a-2d show elevation distributions for the Rocky Mountains (Figures 2c and 2d) and the Sierra Nevada (Figures 2a and 2b) from a 30-s data set compared with model elevations. The model captures the distribution of the Rocky Mountains reasonably well. The mean elevation is 2290 m and the model mean is 2332 m. However, the model shows very little area above 2800 m. The Sierra Nevada is more difficult to represent due to the narrowness of the range. The mean elevation for the Sierras is 1906 m and the model mean is 1524 m. The model does not represent elevations above 2000 m at the 60-km resolution employed. The effects of model topography on simulation errors are addressed in section 5.

3. Model Description

The RegCM [Giorgi et al., 1993a, b] is a limited area model whose dynamical component is essentially that of the Pennsylvania State/NCAR mesoscale model version 4.0 (MM4). The model is hydrostatic, compress-
Figure 1. RegCM domain covering the continental United States and showing contours of model topography (m). The domain is 3720x5940 km or 62x99 grid points at 60-km horizontal resolution. The solid boxes depict the Sierra Nevada region and the Rocky Mountain region.

visible, and based on primitive equations and employs a terrain-following $\sigma$-vertical coordinate. Sigma is defined $\sigma = p - p_{top}/p_s - p_{top}$, where $p$ is pressure, $p_{top}$ is pressure at the uppermost model level, and $p_s$ is surface pressure. The model includes parameterizations of surface, boundary layer, and moist processes which account for the physical exchanges between the land surface, boundary layer, and free atmosphere. The vertical resolution is 14 levels with about five levels within the planetary boundary layer (below 800 mbar). This model top is defined at 80 mbar. A brief description of the surface, boundary layer, and convective parameterizations is presented here, which is sufficient for our analysis and later discussion. Further details are available in the references.

The bottom boundary is composed of land and ocean in the regional model domain. Surface-atmosphere exchange fluxes are computed using the BioSphere Atmosphere Transfer Scheme (BATS) [Dickinson et al., 1993]. BATS is a quasi-one-dimensional surface physics/soil hydrology model which consists of a vegetation layer, three soil layers for moisture, and two soil layers for temperature. Each atmospheric model grid point is specified with characteristics for a single vegetation and soil class. Air temperature, humidity, pressure, winds, radiation, and precipitation are provided by the atmospheric model to BATS at each domain grid point. For grid points designated as land, BATS computes surface radiative, sensible and latent heat, momentum fluxes, and surface temperature based on the assigned vegetation and soil parameters. The snow model employed is relatively simple. Only the snow surface processes are modeled explicitly. Water on the snow surface is put directly into the soil; that is, melt water or rainwater does not percolate through the snow pack or refreeze. Also ground melt is neglected unless this heat reaches the snow surface. Given these assumptions, snow cover is given by

$$\frac{\delta S_{cv}}{\delta t} = P_s - P_e - S_m,$$

(2)

where $S_{cv}$ is the snow cover amount measured in terms of liquid water content, $P_s$ is the snow precipitation rate, and $P_e$ is the rate of sublimation. Snow melt $S_m$ is assessed from the energy required for the surface heat balance. Precipitation is assumed to fall as snow if for the lowest model layer $T \leq T_c$, where $T_c = 273.15 + 2.2$ K.

The runoff calculation is guided by the criterion that there should be small surface runoff at the soil moisture of field capacity and complete surface runoff at saturated soil,

$$R_s = \begin{cases} (\rho_s/\rho_{w\text{nat}})^4G & \text{for } T_{ul} \geq 0^\circ C \\ (\rho_s/\rho_{w\text{nat}})G & \text{for } T_{ul} < 0^\circ C \end{cases}$$

(3)

Here $\rho_{w\text{nat}}$ is the saturated soil water density, $\rho_s$ is the soil water density weighted toward the top layer, and $G$ is the net water applied to the surface. The exchange between atmosphere and ocean is also computed by BATS, with sea surface temperatures specified from a data set of monthly mean values [Reynolds, 1988].

Resolvable scale precipitation is treated with a simplified version of the explicit moisture scheme of Hsieh et al. [1984], as described by Giorgi and Marinucci [1996].
This scheme includes only an equation for cloud water, which is formed when supersaturation is attained at a given model level. It is then advected and can be reevaporated or converted into rainwater. Rainwater is immediately precipitated out. The cloud properties provided by this scheme (cloud water and fractional cover) are used for cloud radiation computations. For convective precipitation the scheme developed by Grell [1993] is employed. In this scheme, clouds are defined as two steady state circulations consisting of an updraft and a downdraft with no mixing between cloudy air and environmental air except at the cloud top and base. The scheme employs a stability based closure similar to that adopted by Arakawa and Schubert [1974], which states that cumulus clouds stabilize the environment as fast as the large-scale and surface fluxes destabilize it.

The RegCM2 requires initialization of model prognostic variables (u, v, T, q, p, and T*), at each horizontal model grid point and, excluding surface pressure ($p_s$) and surface temperature ($T_s$), at each model level. In
addition, because the domain is of a limited area (not global), the lateral boundaries of the domain require periodic forcing of the model prognostic variables by applied external conditions. The external conditions are prescribed using 6-hour ECMWF global analyses, interpolated spatially to the model grid and temporally to the model time step. A weighting function, equal to one at the boundaries, decreases exponentially inward. The number of grid points affected depends on the size of the domain, and in this study the buffer zone consists of 10 grid points.

4. Results

4.1. Mean Seasonal Circulation Differences 1995-1994

ECMWF analyses of the 1994 Northern Hemisphere JFM circulation show that the Aleutian low, a quasi-stationary feature in the North Pacific in winter time, was anomalously weak, while in the North Atlantic the Icelandic low was shifted southward and intensified relative to the 1979-1988 mean. The large amplitude was seen in the stationary wave pattern with a significant
ridge and elevated 500-mbar heights (Figure 3a) over the western United States and a significant trough over the east. This ridge over the western United States resulted in weak upper-level westerly winds in the west during JFM 1994, while the cyclonic flow in the eastern half of the United States and Canada was intensified. The northward displacement of the Pacific jet carried moisture (Figure 3b) to the northwest and British Columbia, but the weakened westerlies transport little moisture from the Pacific to the interior west and southwest, our regions of interest. In the eastern United States the intensified flow carries moisture from the Gulf of Mexico in the southeasterly flow along the eastern seaboard. This moisture transport resulted in record snowfall in the east.

In JFM 1995, analyses show lower than normal 500-mbar heights in the midlatitude central and eastern Pacific and above normal heights in much of the east. The differences between JFM 1995 and JFM 1994 500-mbar heights are shown in Figure 3c. These differences result in a significantly reduced amplitude of the stationary wave in the 500-mbar height field across North America. The deeper Pacific low was associated with enhanced upper level westerlies across the eastern Pacific between 25 and 35°N and with reduced westerlies at higher latitudes (40-50°N). The conditions reflected a southward shift and eastward extension of the exit region of the Pacific jet stream, which resulted in a more westerly flow of the North American jet across the continent. Moisture transport differences between JFM 1995 and JFM 1994
(Figure 3d) show the enhanced mid-latitude westerlies provided substantial moisture from the Pacific basin to the entire west coast and significantly more to the interior west than was seen in JFM 1994. The southeasterly flow from the Gulf of Mexico, though not as strong as in 1994, still provides significant moisture to the Atlantic seaboard in JFM 1995.

The difference in stationary wave patterns simulated by the RegCM for JFM 1995 minus 1994 (Figure 4a) closely resembles that from ECMWF in Figures 3c. The model-generated vertically integrated moisture transport differences are shown in Figure 4b. The enhanced transport along the Pacific seaboard and into the interior west in 1995 are simulated well by the RegCM. However, the model simulates moisture transport larger than the analyses in eastern Utah, Colorado, and northern New Mexico.

4.2. Precipitation

The circulation and moisture transport differences between the winter seasons of 1994 and 1995 have a substantial impact on precipitation in the west. This analysis employs monthly precipitation data from 1000 stations across the United States from NCDC, most of which are located in populated areas that are not representative of high elevations in alpine regions.
served precipitation from NCDC stations is presented first, then a combined analysis of NCDC and SNOTEL station data, which adds information at higher elevations is discussed. The uncertainties in the analysis method are provided, and we evaluate RegCM simulated seasonal precipitation against these observations for JFM 1994 and 1995.

Figures 5a and 6a show the JFM observed total precipitation computed from NCDC data for 1994 and 1995, respectively. The ridge over the west in 1994 resulted in abundant precipitation for the northwest, and little precipitation in the interior west, while the deep trough in the east brought record precipitation to the Appalachian range and the eastern seaboard. The NCDC stations show substantial precipitation amounts across the west coast due to the increased westerly flow in 1995. The interior west also shows more precipitation in these observations. Most notable, however, is
the lack of information in the mountain regions of the interior west in both years.

Precipitation data collected at high-elevation SNOTEL stations are combined with the NCDC data in Figures 5b and 6b using the section 2 analysis method described in section 2. The contrast between the NCDC only precipitation and the combined NCDC/SNOTEL precipitation is striking. The incorporation of SNOTEL data adds substantial information to the observed NCDC precipitation in the interior west. Precipitation in the Rocky Mountains from New Mexico north through through Idaho, as well as in the mountains of Utah, Nevada, and Arizona, is revealed by the SNOTEL data. That little change is seen in the coastal ranges is
due to steep gradients and smaller spatial scale of these mountains. Recall that we apply the high-elevation precipitation data using a critical elevation $H_c$. The coastal ranges have relatively little area above the critical elevation, hence the SNOTEL data make a negligible impact along the west coast. It is important to note that combined precipitation is sensitive to the choice of critical elevation. If $H_c = H_0$, and the SNOTEL data is applied only above the mean station elevation in a grid cell, the precipitation in the Rocky Mountains is reduced by 30%. The choice of $H_c$ in this study approximates the high end of precipitation supported by observations and provides a maximum estimate of model error.

The circulation differences between 1995 and 1994 resulted in enhanced precipitation in 1995 along the west coast that was related to the local SLP anomaly in the midlatitude Pacific and the stronger moisture transport associated with the enhanced westerlies. These observations support the results of Cayan and Peterson [1989], which show positive streamflow anomalies along the
west coast during climatologically similar conditions.

Figures 5c and 6c show the RegCM simulated total precipitation for JFM 1994 and 1995, respectively. The model simulates abundant precipitation along the west coast in 1995 and relatively smaller precipitation amounts in the Southwest and southern California in 1994. Precipitation is underpredicted in the Colorado Rockies in both years but simulated quite well in the Sierra Nevada. The model captures the pattern and approximate amplitude of precipitation in the eastern United States in both 1994 and 1995, although there is a small underprediction in the maximum in the Appalachian region in 1994 and in the gulf states in 1995. A close inspection also shows that the model tends to precipitate on the windward (nominally) western slope of the higher ranges, leaving less moisture for the eastern slope, as seen throughout the Rockies. From this visual comparison the model precipitation errors appear to be largest in the interior west, where the errors in moisture transport occur.

4.3. Snow Water Equivalent and Time Series

How the variations in seasonal precipitation and temperature translate into snowpack are described in this section. The NOHRSC combined SNOTEL and satellite SWE product for the western United States for 5-day composited periods at the end of January, February, and March are presented for 1994 and 1995 and are used to evaluate qualitatively the model-simulated SWE through the winter seasons of these 2 years. In order to remain concise the observed SWE product and simulated SWE are presented for the end of March 1994 and 1995 only. These values effectively integrate the winter season precipitation and snow processes. Plates 1a and 1b show the observed SWE in the western United States at the end of March 1994 and 1995, respectively. Note first the difference in color scales between the 2 years. In 1995 the blue-violet represents greater than 80 inches (203 cm) SWE while in 1994 the same color is for greater than 20 inches (51 cm) only.

At the end of March 1994 the western snowpack shows 14-20 inches (36-51 cm) SWE at the highest elevations of the Sierra Nevada and the central Rockies and more than 20 inches (51 cm) SWE in the northwest range of Washington. The northern Rockies (Columbia River Basin) has generally <8 inches (20 cm) SWE at this time. By contrast, the end of March 1995 shows >64 inches (163 cm) SWE in much of the Sierra Nevada and the north west. The interior west, e.g., Colorado Rockies, the Teton range, the Salmon River Mountains in Idaho, and the northern Rockies, all show >28 inches (71 cm) SWE with regions of >46 inches (117 cm).

The observed alpine snowpacks in the coastal ranges in 1995 were 3 times those in 1994. Throughout the Rockies the 1995 snowpacks were twice those in 1994. This is consistent with the differences in observed precipitation for the 2 years. In order to simulate the seasonal snow processes that produce high-elevation snowpack the RegCM employs BATS, which accumulates, evaporates and melts snow in its account of SWE at each 60-km model grid cell. Snapshots of the model-generated SWE are given for the end of each period shown in the observations. The color scales have been defined to match as closely as possible those in the observations.

The RegCM simulated SWE is given in Plates 2a and 2b for the end of March 1994 and 1995. These maps show the entire United States, but our discussion will focus on the west. At the end of March 1994 the snow extent had retreated throughout the west but was still present in the northern and central Rockies. Less than 4 inches (10 cm) was present in the Sierra Nevada. The snow accumulation in the model is ~20-25% of that observed for the Sierra Nevada and 25-30% of observed in the central Rockies. In contrast to the 1994 simulation the model has accumulated more snow throughout the west at the end of March 1995. Two grid points in the Sierra Nevada show SWE in excess of 10 inches (25 cm). The central Rockies show larger areas that have greater than 10 inches SWE, and the Southwest shows regions having greater than 24 inches. Thus the interior west in the model is getting approximately one half to one third of the observed snowpack, while the simulated snowpack in the coastal ranges is perhaps 20% of the observed.

As with the simulated precipitation, the model snowpack is preferentially located on the western slopes of the highest ranges. Thus most of the precipitation occurs during the resolvable orographic uplift, which occurs on the windward side of the model mountains, leaving less moisture for precipitation on the leeward side.

In order to understand how precipitation and temperature combine to produce the observed and simulated SWE, time series of area and monthly average precipitation, maximum daily temperature, minimum daily temperature, and SWE are computed for the Rocky Mountain region and for the Sierra Nevada region. These time series are provided in Figures 7 and 8, respectively, and Table 1 gives mean and root-mean-squared (rms) errors for each field in each region.

Figure 7 shows that in 1994 the model predicts one-half to one-third of the observed precipitation in the Rocky Mountains in each month. Surface temperatures reveal that the model is cold relative to station data during daytime (TMAX) and warmer than observed at night (TMIN). This error in the diurnal temperature range results from an overprediction of clouds in the model (E. Small, personal communication, 1998). The resulting simulated SWE is significantly smaller than the observed, 7 inches (17.8 cm) less than the mean for the Rockies by end of March 1994. In 1995, precipitation amounts are much larger due to the enhanced westerlies and moisture transport. In this winter, daytime maximum temperatures are again colder in the model relative to observations, and the SWE errors are quite large by end of March (20 inches, 51 cm). Figure 8 confirms
Plate 1. Snow water equivalent maps (inches, 1 inch = 25.4 mm) for (a) March 22-29, 1994, and (b) March 22-28, 1995, from NOHRSC.
Plate 2. Snow water equivalent maps (inches) for (a) March 29, 1994, and (b) March 28, 1995 from RegCM.
that model simulated precipitation is much better in the Sierra Nevada than in the Rocky Mountains, with mean errors <30 mm/month (except January 1995). Daytime maximum temperatures do not show as systematic an error, and nighttime temperatures are warmer than observed. In this region the daytime temperatures are above $T_c$, which determines if the precipitation will fall as rain or snow. The season-integrated SWE errors are 6 inches (15 cm) in March 1994 and 9 inches (23 cm) in 1995.

The possible sources of these very large errors in simulated snowpack include the model’s circulation and transport of moisture through the regions, the precipitation, the surface temperatures, and inadequacies in the model topography. The circulation and moisture transport errors appear to be small. The precipitation errors are substantial in the Rocky Mountains and relatively small in the Sierra Nevada. Yet the SWE errors are large in both regions. Surface temperatures must have an important effect in the model-simulated snowpack.

Figure 7. JFM 1994 and 1995 time series of monthly precipitation (mm), maximum daily temperature (K), minimum daily temperature (K), and SWE (inches) from observations (solid) and the RegCM simulation (shaded) averaged spatially for the Rocky Mountain region.

Figure 8. Same as Figure 7, but for the Sierra Nevada region.
Table 1. Mean and RMS Errors Between RegCM-Simulated Fields and Observations for the Rocky Mountain Region and the Sierra Nevada Region for January-March 1994 and 1995

<table>
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<th>Jan. Mean</th>
<th>Jan. RMS</th>
<th>Feb. Mean</th>
<th>Feb. RMS</th>
<th>March Mean</th>
<th>March RMS</th>
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</tbody>
</table>

While the error in diurnal temperature range is similar in both the Rocky Mountains and the Sierras, the Sierras are warmer than the Rockies. From the results of Cayan [1993] we know that temperature has more influence on streamflow in lower elevation basins than in higher elevation basins during winter. Because the daytime temperatures in the Sierra Nevada are above the critical temperature for snowfall, much less of the precipitation falls as snow. This can be seen in a diurnal time series of precipitation and runoff from the model (not shown), where precipitation and runoff occur often coincidentally in the Sierra Nevada. In contrast the surface temperatures in the Rocky Mountains are below freezing until March, and diurnal time series show very little runoff until temperatures warm in March. Thus it is proposed that in the RegCM the SWE errors seen in the Rocky Mountains result primarily from errors in precipitation until March when temperature begins to have an effect. While in the Sierra Nevada, where precipitation errors are relatively small, the errors in snowpack result primarily from the lack of higher elevations in the model topography. There is insufficient area at high, cold elevations to develop the observed snowpack in this region. These mechanisms are consistent with the model errors in topography, which are larger in the Sierra Nevada than in the Rocky Mountains.

4.4. Monthly Streamflow

The RegCM computes runoff from each model grid cell diagnostically and monthly and spatial means are calculated for the Rocky Mountain and Sierra Nevada regions. The regions are chosen within the RegCM domain to approximate the upper Colorado River basin and the 10 smaller river basins of the Sierra Nevada, in order to enable comparison of model generated runoff with observed streamflow measurements. Recall that the stream gauges employed effectively integrate the upper Colorado River basin and separately the basins of the Sierra Nevada. It is important to note also that diagnosed model runoff is not physically the same as the observed streamflow against which it is compared. The model incorporates no horizontal flow and thus no time lag between melting of snow and peak stream flow.

Figure 9 shows monthly times series of observed streamflow (solid line) and model generated runoff (dashed line) for water years 1994 and 1995 for the Rocky Mountain region (Figure 9a) (upper Colorado River basin) and the Sierra Nevada (Figure 9b) (ten basins). Both major basins show the observed 1995 peak flows to be 3-4 times those in 1994. Spring snowmelt begins earlier in the Sierra Nevada than in the Rocky Mountains due to the relatively warmer temperatures in the former. In 1995 the observed runoff in the Sierra Nevada shows three peaks, with the minimum in February due to the dry period experienced during that month.

The model-simulated runoff also shows an increase between the spring of 1994 and 1995 in both basins, but the interannual difference is not as large as observed. The peak model runoff in the upper Colorado River basin precedes the observed runoff by approximately one month. The observed springtime peak runoff from the Rockies is in May/June, while the simulated runoff peaks in April/May. In addition, the model runoff ends
by June, while observed flows continue through July (and August in 1995). We have seen that the model is cold enough to generate a snowpack in the model Rocky Mountains. In fact, model daytime temperatures show a slight cold bias when compared with analyzed station temperatures. Yet the daytime temperatures in February and March do exceed the critical temperature for snowmelt. In the RegCM when the critical temperature is exceeded snowmelt and runoff occur simultaneously as seen in 6-hour time series which were discussed earlier but not shown. (The model warm bias in minimum daily temperature is unlikely to affect runoff because the minimum temperatures are well below the critical temperature for snow melt to occur.) Thus the early peak in runoff appears to be related to daytime temperatures and to the fact that model runoff is instantaneous. The errors in late season model runoff are due to the lack of snow accumulation in the model Rocky Mountains. Simulated precipitation is deficient in this region, and the highest terrain that actually accumulates snow is warmer than observed because of the elevation bias, resulting in a significantly smaller than observed snowpack. Recall that the model lacks elevations above 2800 m in the Rocky mountains where larger snow amounts would accumulate.

In the Sierra Nevada, which has a lower elevation and thus greater fraction or rain versus snow as compared to the Rocky Mountains, the model simulates the timing of the early spring runoff quite well. This early spring runoff is a combination of rain and snowmelt. However, the model accumulated insufficient snow to simulate the late spring runoff peak, which is comprised largely of runoff from higher-elevation snowmelt. Though the snowmelt model in BATS is relatively simple, on a seasonal basis it does simulate alpine snowmelt reasonably well [Jin et al., 1999]. Thus the problem with late spring runoff is again lack of snow accumulation associated with lack of high-elevation regions in the model.

5. Discussion

While the RegCM can capture the broadest changes from year to year in the annual runoff cycle of the large basins in the Rocky Mountains and Sierra Nevada, the model shows significant errors in both the timing and the amplitude of the springtime peak flow. These errors result from errors in simulated precipitation and also from errors in how the precipitation translates to snow in the model. Both sources of error are fundamentally, though not exclusively, tied to the model topography. Precipitation errors in the Rocky Mountains and the interior west are larger than those near the moisture source of the Pacific. Thus errors in the Rocky Mountains are likely due to the lack of higher elevations in the model topography and insufficient moisture transport to the interior West. Although lack of higher elevations is a problem in the Sierra Nevada as well, there is enough moisture transported into the region that the model precipitates it. Still, the sources of precipitation error in the RegCM require further investigation. The error may result in part from our analysis of precipitation observations. Analysis of additional winter seasons will show if errors in moisture transport are important. Higher resolution simulations are necessary to isolate the effects of elevation errors on orographic precipitation.

The mechanisms which result in large errors in simulated snowpack are quite different in the Rocky Mountains and in the Sierra Nevada. Precipitation is the
primary source of error in the snowpack of the Rocky Mountains, though the model topography and moisture transport may both play a role in the precipitation error. In the Sierra Nevada, precipitation errors are generally small; however, temperatures (observed and simulated) are well above the critical temperature for snowmelt. Much of the precipitation goes directly into runoff and little accumulates as snow due to the lack of higher colder elevations in the model.

Improved model topography would permit more mountain precipitation and more snowfall (i.e., a larger snowpack), particularly in the Sierra Nevada. It would also improve the timing of snowmelt. Improvements in model topography can be achieved by a number of methods. The most straightforward method is a high-resolution atmospheric model. Resolutions higher than 60 km (used in this study) are feasible with regional models but become quite expensive computationally over large domains due to the smaller time steps required. Alternate methods have been suggested for incorporating the subgrid scale elevation effects on the amount of precipitation that falls as snow, thus improving the depth and distribution of the snowpack. These include statistical distributions of elevation within a model grid cell [Avisar, 1992; Giorgi, 1997], explicit break down of a model grid cell into subgrid cells each represented by an elevation [Seth et al., 1994], and employment of a subgrid airflow model within a number of elevation classes [Leung and Gahn, 1995].

6. Conclusions

Our results indicate that winter precipitation, SWE, and springtime runoff variations in the west are organized by atmospheric circulation patterns which occur on large-scales relative to individual basins. Global analyses from JFM 1994 and 1995 show different large scale circulations across the North Pacific and North America in these two winter seasons. JFM 1994 was dominated by a negative PNA pattern associated with a lower than normal Aleutian low in the North Pacific, a strong ridge over western North America, and a deeper Icelandic low in the east. The displacement of the upper level westerlies northward into Canada resulted in dry conditions, smaller snowpack, and less springtime streamflow in the interior west and southwest. In JFM 1995 there was an anomalous low pressure in the mid-latitude eastern Pacific. This resulted in stronger upper level westerlies and large moisture flux from the Pacific into the west, which led to larger snowpack and springtime streamflows in the southwest. These results are consistent with those of Cayan [1996] and Cayan and Peterson [1989] who studied climatological influences of large-scale atmospheric circulation on SWE and streamflows in the west.

The NCAR RegCM simulates the large-scale circulations in the two studied winter seasons quite well. Errors in the regional climate model are apparent in the precipitation fields and are greater in comparison with observed snowpack and runoff. Regionally in the Rocky Mountains and the Sierra Nevada, the model errors accumulate rather different from two basins. In the Rocky Mountains the model is deficient and leads to inadequate snowpack which is too early and too little. In the Sierra Nevada, however, the precipitation is quite close to observed yet errors in snowpack are quite large. Errors in model topography are large in the northern 60-km resolution does not represent up to 2000 m, and because temperatures are below the melting point of snow, relatively little snow accumulates in the model Sierras. Thus, while the snowpack/streamflow errors are controlled better in the Rocky Mountains, the errors are elevation/temperature in the Sierra Nevada. RegCM runoff simulates a factor of 2 increase between 1995, where the observations show a factor of 2 in the Colorado River basin and a factor of 3 in the Sierra Nevada. Despite these problems, it is able to reproduce the overall patterns of snowmelt and snowcover and approximately half-year variations.

The SNOTEL precipitation data provides an essential test for evaluating the model simulated precipitation at high elevations. However, there is a sensitivity to the method used for combining elevation station data. Such methods show how to determine their accuracy. The NOHRSC provide a valuable resource for qualitative analysis of the model snow processes, but quantities that require a gridded SWE data set (not is needed but as yet not available. More data are essential for this type of study.

We have discussed possible improvements to the RegCM within the context of the proposed model. These include the need for subgrid elevation classes to improve the snowpack. For the initial study we have employed an analysis scale forcing to drive the boundaries of the domain. Seasonal prediction studies will also benefit from GCM ensemble forecasts, the skill of which at high latitudes remains to be determined. Gregs at large-scale seasonal anomalies from a GCM could improve models may be effectively used for seasonal and interannual water resource planning.

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