Physically Based Global Downscaling: Climate Change Projections for a Full Century

STEVEN J. GHAN AND TIMOTHY SHIPTERT

Pacific Northwest National Laboratory, Richland, Washington

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ABSTRACT

A global atmosphere–land model with an embedded subgrid orography scheme is used to simulate the period 1977–2100 using ocean surface conditions and radiative constituent concentrations for a climate change scenario. Climate variables simulated for multiple elevation classes are mapped according to a high-resolution elevation dataset in 10 regions with complex terrain. Analysis of changes in the simulated climate leads to the following conclusions. Changes in surface air temperature and precipitation differ from region to region in a manner similar to simulations without the subgrid scheme. Subgrid elevation contributes little to spatial variability of the change in temperature and the relative change in precipitation. In some regions somewhat greater warming occurs at higher elevations because of the same tendency in the free troposphere, but in others greater warming occurs near the melting level where snow albedo feedback amplifies the warming. Changes in snow water are highly dependent on altitude because of its nonlinear dependence on changes in the melting level. Absolute changes usually increase with altitude because more snow is currently available for depletion, but for extremely cold conditions the simulated warming is insufficient to increase melting. Relative changes in snow water always decrease with altitude as the likelihood that a warming will enhance melting or change the phase of precipitation decreases with decreasing temperature at higher altitudes. In places where snow accumulates, an artificial upper bound on snow water (which is required in any climate model that does not treat lateral snow transport) limits the sensitivity of snow water to climate change considerably. The simulated impact of climate change on regional mean snow water varies widely, with little impact in regions in which the upper bound on snow water is the dominant snow-water sink, moderate impact in regions with a mixture of seasonal and permanent snow, and profound relative impacts on regions with little permanent snow.

1. Introduction

The climatic signature of global warming is both local and global. The forcing by increasing greenhouse gases is global, so there is clearly a global component to the climatic signature. But there is also a local climatic signature because of the interaction of local topography with nonlinear processes such as precipitation and melting (Paul et al. 2004). A uniform rise in the freezing level, for example, can produce a strong orographic signature in the distribution of changes in snowfall, snow water, and runoff on scales less than 10 km in regions with complex terrain.

Projecting the impact of global climate change on water resources therefore requires the ability to represent climate processes on a variety of spatial scales, from global (40 000 km) down to local (5 km). Current computing resources are insufficient to permit global climate simulations explicitly resolving such a wide range of spatial scales for more than a few months. Century climate simulations spanning the full range of spatial scales will not be feasible for a decade or more.

Recognizing this, the climate impacts community has developed a variety of methods for downscaling global model projections of climate change. These methods include regional climate modeling, high-resolution time slice, stretched grid, statistical downscaling, and physically based downscaling.

Regional climate models (RCMs) provide high-resolution climate simulations for a century for selected regions (Giorgi 1990; Leung et al. 2004; Wang et al. 2004). RCMs use treatments of climate processes that are comparable (in some cases identical) to those used in global climate models. They are driven by lateral boundary (winds, temperature, and humidity) and ocean surface (temperature and ice cover) conditions from a global climate simulation, but use finer-resolu-
tion land surface conditions (elevation and soil and vegetation type). Although the realism of the climate simulated by RCMs is limited to some extent by the realism of the lateral boundary conditions from the global model, the climate simulated by RCMs is usually more realistic than that simulated by the parent model, particularly at spatial and temporal scales not resolved by the global model. The primary limitation of the RCM downscaling method is of course the limited region of the climate simulation.

A second downscaling method is running the global model at higher resolution for selected time slices, again using ocean surface conditions from a global climate simulation (Cubasch et al. 1995; Ohmura et al. 1996a,b; Timbal et al. 1997; May and Roecker 2001; Wild et al. 2003). The time slice is typically a decade, which is long enough to distinguish climate signals from noise but short enough to permit much finer resolution than is used in global climate simulations for a century. Given current computational resources, 60–100-km grid sizes are most common (Govindasamy et al. 2003; Iorio et al. 2004). The primary limitation of this method is the resolution, which is marginal for many impact studies.

A third method is the stretched grid, which is applied to one or more regions in a few global climate models (Paegle 1989; Fox-Rabinovitz et al. 2001). This provides finer resolution in selected regions without the high cost of fine resolution for the whole globe. This promising method combines the advantages of regional climate modeling and high-resolution time slices. Its primary limitation is doubt about the applicability of subgrid physics parameterizations to the range of grid sizes used in the model.

A fourth method is statistical downscaling, which develops relationships between local and regional meteorology using historical observations, and applies them to global climate model projections of future climate change (Wilby and Wigley 1997; Murphy 1999; Widmann et al. 2003; Wood et al. 2004; Benestad 2004; Maurer and Duffy 2005). This method can be used to rapidly account for the complex influence of land surface heterogeneities and to correct for biases in global climate model simulations. Its primary disadvantage is its limited applicability to climates significantly different from the historical record. It also requires an adequate density and quality of observations.

A fifth method is physically based downscaling, in which a physically based treatment of the climatic influence of subgrid land surface conditions is applied either following or preferably during the global climate model simulation. One example of this approach is the subgrid orography scheme developed by Leung and Ghan (1995, 1998). The full column atmospheric and land surface physics of the climate model are applied to each of a modest set of elevation classes within each grid cell, with the orographic forcing of temperature and humidity determined from conservation of energy and moisture applied to an estimated vertical displacement of air parcels given the height of the elevation class and the stability of the atmosphere. The history is written for each elevation class, which can then be distributed in postprocessing according to the high-resolution distribution of surface elevation. The computational cost of this method is about a factor of 2, which is far less than the cost of the high-resolution time slice. This method provides a physically based link between global and local scales that can be applied to century climate simulations. The primary disadvantage of it is the limited ability of the current scheme to represent rain shadows. The scheme typically produces more precipitation at higher elevations, but similar precipitation at the same elevation on the windward and lee side of mountain ranges.

Leung and Ghan (1999) used the method in an RCM to simulate the climate response to double CO$_2$. Ghan et al. (2002, hereafter G2002) applied the method to a prototype of the National Center for Atmospheric Research (NCAR) Community Atmosphere Model (CAM2) and showed that its simulation of temperature, precipitation, and particularly snow water is clearly superior to simulations without the subgrid scheme, even at much finer explicit resolution. Here we present the results of the application of the G2002 physically based downscaling method to century climate change simulations by a global climate model, the NCAR Community Climate System Model (CCSM3; Collins et al. 2006a) for one Intergovernmental Panel on Climate Change (IPCC) scenario. A companion paper (Ghan et al. 2006, hereafter G2006) describes the downscaling method and evaluates the simulation of the period 1980–2000 by this model. Section 2 describes the design of the experiment. Section 3 presents results from the climate change downscaling simulation. Conclusions are summarized in section 4.

2. Experiment design

As explained by G2006, the downscaling method has been applied to the NCAR CAM3 (Collins et al. 2006b; Boville et al. 2006; Hack et al. 2006) and Community Land Model (CLM3; Dickinson et al. 2006). CAM3 and CLM3 are run in an offline mode, that is, driven by ocean surface conditions taken from a CCSM3 climate simulation (Meehl et al. 2006), for the period 1977–2100. Although this strategy does not permit feedback of the slightly different atmosphere simulated by
CAM3 with the downscaling method, G2002 have shown that differences between CAM3 simulations with and without the downscaling method are much smaller than model biases.

The simulated monthly mean ocean surface temperature and sea ice cover are mapped from the CCSM3 ocean model grid to the grid of CAM3 and CLM3 using the Spherical Coordinate Remapping and Interpolation Package (SCRIP; Jones 1999; see also http://climate.lanl.gov/Software/SCRIP/index.htm). Area means are conserved by the mapping. The monthly mean values are then converted to midmonth values (Taylor et al. 2000) such that the monthly mean of values interpolated between the midmonth values will match the monthly mean values mapped from the ocean model grid.

To correct for biases in the ocean simulation, the ocean surface temperature bias for the period 1979–96 is subtracted from the simulated ocean temperature,

\[ T^* = T + (T_{\text{obs}} - T), \]

where \( T \) is the simulated ocean surface temperature, \( T^* \) is the corrected temperature, and \( T_{\text{obs}} \) and \( T \) are, respectively, the observed and simulated climatological mean temperature for the period 1979–96. The climatological mean of the corrected temperature for the period 1979–96 is therefore identical to that observed.

This type of correction will not work for sea ice, because for points where sea ice is observed but none is simulated the bias correction will not permit sea ice to decrease as the climate warms. We therefore use the following treatment:

\[ c^* = c_{\text{obs}} + \frac{dc}{dT}(T - T), \]

where \( c_{\text{obs}} \) is the observed climatological mean sea ice fractional coverage for the period 1979–96 and \( dc/dT \) is the slope of the regression of the simulated sea ice on ocean surface temperature for all months and grids cells in which sea ice fraction is between 0.01 and 0.99. This treatment yields the observed climatological mean sea ice distribution during the period 1979–96 and allows sea ice to decrease in a quasi-realistic manner as the climate warms. It does not produce any discontinuities in the sea ice distribution.

For consistency with the CCSM3 simulation the same concentrations of the greenhouse gases \( \text{CO}_2, \text{CH}_4, \text{N}_2\text{O}, \text{CFC}-11, \) and \( \text{CFC}-12 \) used in the CCSM3 simulation are also used in the downscaling simulation.

The CCSM climate simulation is also forced by up to 10 aerosol species. The concentrations of all but one of the aerosol types are prescribed from monthly means simulated by an offline model (Collins et al. 2001). The concentration of sulfate aerosol is simulated by CCSM3 using the historical emissions and IPCC projections of emissions of sulfur dioxide. We were neither permitted to use the sulfate simulation capability of CAM nor its emissions, so we cannot ensure absolute consistency between the sulfate mass simulated by CCSM3 and the downscaling simulations. But we have obtained the monthly mean sulfate concentrations from the CCSM3 simulation and have used the decadal mean sulfate for each month, converted to a midmonth value for temporal interpolation, in our downscaling simulation. Such an approximation is unlikely to significantly influence the climate signal because ocean surface conditions are prescribed. For the other aerosol types the same monthly climatology is used.

How large are the ocean surface temperature and sea ice cover adjustments? Figure 1 shows the spatial distribution of the annual mean ocean surface temperature and sea ice cover bias from the CCSM3 simulation for the years 1980–96. The ocean surface temperature is more than 2°C too cold in two wide swaths across the
North Atlantic, from Labrador to Spain and from the Caribbean Sea to Africa. Other regions in the Pacific Ocean and in the Antarctic Circumpolar Current are also 1°–4°C too cold in the CCSM simulation. These biases are similar to the difference between the observed surface temperature and that simulated by CCSM3 (Collins et al. 2006a). Consistent with the biases, the sea ice cover is 20% too large in the Labrador Sea and along the Antarctic ice shelf.

Such large biases are likely to impact the climate simulated by CAM3. To quantify this, we have run CAM3 with the adjusted and unadjusted ocean surface temperature and sea ice datasets for the period 1977–99. Figure 2 shows the global distribution of the ratio of the annual mean and grid cell mean precipitation simulated with unadjusted/adjusted ocean surface conditions for the years 1980–99.

On scales larger than the grid cell the distribution of the change for 2080–99 is very similar to that from the coupled CCSM simulation without the subgrid scheme (Meehl et al. 2006). During the period 2080–99, in both the downscaling simulation and the CCSM simulation the subtropical east Pacific, western United States, southern Europe, and midlatitude Andes are drier than during the period 1980–99, while the intertropical convergence zone, Alaska, Canada, the eastern United States, Russia, and the tropical Andes are wetter. This suggests modest influence of the subgrid scheme on the circulation, so that the discussion of the change by Meehl et al. (2006) applies to this simulation as well.

3. Results

Our analysis will slice the climate change simulation dataset two different ways. First, we will focus on two 20-yr periods: 2030–49 and 2080–99. For temperature and precipitation we will examine the global distribution of the change in the annual mean with respect to the period 1980–2000. For snow water we will focus on the change within the 10 regions evaluated by G2006: the western United States, Alaska/western Canada, Greenland, Scandinavia, the European Alps, the Himalaya Range, New Zealand, the Andes Range, Mexico, and Mt. Kilimanjaro. Second, we will look at the time dependence of the spatial mean response for the full period 1977–2100, focusing on snow water because of its implications for water resources.

a. Precipitation

Figures 3 and 4 show the global distribution of the change in annual mean precipitation (expressed as a difference and a ratio) with respect to 1980–2000, for the periods 2030–49 and 2080–99, respectively. The distributions have been plotted at a 10-min resolution. To relate the precipitation distribution to orography, Fig. 5 shows the global distribution of surface elevation at the same resolution.

On scales larger than the grid cell the distribution of the change for 2080–99 is very similar to that from the coupled CCSM simulation without the subgrid scheme (Meehl et al. 2006). During the period 2080–99, in both the downscaling simulation and the CCSM simulation the subtropical east Pacific, western United States, southern Europe, and midlatitude Andes are drier than during the period 1980–99, while the intertropical convergence zone, Alaska, Canada, the eastern United States, Russia, and the tropical Andes are wetter. This suggests modest influence of the subgrid scheme on the circulation, so that the discussion of the change by Meehl et al. (2006) applies to this simulation as well.
Closer examination of Figs. 3 and 4 reveals a variety of patterns of precipitation change at subgrid scales. In some regions (the western United States and New Zealand) the orographic signature is small. In others (Alaska, western Canada, and the Andes) the absolute change in precipitation has a strong orographic signature (increasing with altitude), but the relative change is insensitive to elevation. For some regions (Norway) precipitation increases on the windward side of mountains but changes little on the lee side. In others (central Asia, Chili, Peru, and Mt. Kilimanjaro) even the sign of the precipitation change depends on elevation. In most regions the absolute precipitation change increases with altitude, which is consistent with other simulations that resolve orographic effects (Leung and Ghan 1999), but in some regions (Greenland) the precipitation change decreases with altitude. These patterns are similar for both analysis periods, but of course are stronger for 2080–99. Relative changes are typically 10%–50% for 2080–99, with much weaker spatial variability than on scales larger than the grid cell.

b. Surface air temperature

Figure 6 shows the annual mean surface air temperature change with respect to 1980–99 for the two periods, plotted at a 10-min resolution. As for precipitation, on scales larger than the grid cell the distribution of the change for 2080–99 is very similar to that from the coupled CCSM simulation without the subgrid scheme.
(Meehl et al. 2006). The discussion of the change by Meehl et al. (2006) applies to this simulation as well.

At the subgrid scale, the surface air temperature change is insensitive to elevation in most regions, including the western United States, Alaska, western Canada, Greenland, Norway, central Asia, Mexico, and Mt. Kilimanjaro. In some regions (New Zealand and the Andes) the warming increases with altitude, reflecting the well-known influence of the same structure in the free troposphere. In southern Europe the distribution of temperature change is complex, being relatively low (1°–2°C for 2030–49 and 2°–3°C for 2080–99) in the valleys and on the highest mountains, but larger (2°–3°C for 2030–49 and 3°–4°C for 2080–99) at intermediate elevations; this response may reflect the snow albedo feedback, which is strongest at intermediate elevations near the snow line.

c. Snow-water equivalent

Although the change in precipitation and surface air temperature exhibit weak or inconsistent orographic
signatures, the change in snow water has a strong regional dependence on surface elevation. To see this, Figs. 7–15 show the absolute and relative change in annual mean snow water for the western United States, Alaska and western Canada, Greenland, Scandinavia, the European Alps, the Himalayas, New Zealand, the Andes, and Mexico, with respect to 1980–99, for the periods 2030–49 and 2080–99. To better see the fine-scale structure of the response, the changes have been plotted at 2.5-min resolution. Also shown is the surface elevation and the time dependence of the regional and annual mean snow water for the period 1980–2100.

In most regions the change in snow water has a strong orographic signature, with absolute reductions increasing with altitude within grid cells and relative reductions decreasing with altitude. More snow is depleted at elevations where more is simulated in the control climate, which in most regions is greater at higher elevations. But in extremely cold regions or elevations (e.g., central Greenland, the Himalayas, and the Rocky Mountains of the western United States), where a warming of a few degrees does not influence melting or the phase of precipitation, the absolute reduction in snow water is generally smaller than in more temperate regions or elevations. For the same reasons, relative reductions range from 100% at elevations near the present snow line to less than 10% on the highest mountains. Leung and Ghan (1999), using a regional climate model with the same subgrid scheme, showed that changes in the distribution of snowmelt as well as snowfall have a strong elevation dependence and hence also contribute to the elevation dependence of the snow.
Fig. 7. (top) Simulated difference and (bottom) ratio of annual mean snow-water equivalent (m) for the periods (left) 2030–49 and (middle) 2080–99 with respect to the period 1980–99, mapped at 2.5-min resolution for the western United States. (top right) Surface elevation (m) plotted for the region at the same resolution. (bottom right) Annual mean snow water for the period 1980–2100 averaged between latitudes 36°–49°N and longitudes 125°–115°W.

Fig. 8. Same as in Fig. 7, but for Alaska and western Canada between 50°–70°N and 160°–120°W.
Fig. 9. Same as in Fig. 7, but for Greenland between 60°–85°N and 60°–20°W.

Fig. 10. Same as in Fig. 7, but for Scandinavia between 55°–70°N and 5°–30°E.
Fig. 11. Same as in Fig. 7, but for the European Alps between 42°–48°N and 5°–15°W.

Fig. 12. Same as in Fig. 7, but for Tibet and the Himalayas between 25°–40°N and 70°–100°W.
water change. Mote et al. (2005) find a similar signature in snow-water data for the period 1950–97. Clearly the nonlinear dependence of snowfall and melting on temperature are responsible for the much stronger elevation dependence of the snow change compared with changes in temperature or precipitation. Simulations without the subgrid scheme, as illustrated by Leung et al. (2004), cannot provide this level of information; even the grid cell means are vastly different.

Although the spatial mean snow water varies widely from year to year in many regions, the downward trend is obvious for all regions. The relative decrease varies widely from region to region, with little impact in some regions and profound impacts in others. As might be expected, the relative impact is greatest in regions with little permanent snow (New Zealand, Mexico, and the Andes). Moderate relative reductions are simulated for regions with a mixture of seasonal and permanent snow (the western United States, Alaska, Scandinavia, the Alps, and the Himalayas). The smallest reductions are simulated for Greenland and Mt. Kilimanjaro, which require further explanation.

Although part of the insensitivity for Greenland is due to its cold temperature, another factor plays a major role there and for Kilimanjaro as well. In Greenland the strong orographic signature on the baseline temperature (G2006) produces an extremely strong orographic signature on the change in snow water. Snow water decreases only near the freezing level, which is along the edges of the Greenland ice dome. The snow-water loss there is limited to 1 m because the model does not permit snow water to exceed 1 m, with the excess snow discarded as runoff. Thus, it is likely that the snow-water loss on Greenland is underestimated.

To see this more clearly, we have saved each term in the snow water budget, which can be expressed as

\[
\text{storage} = \text{snowfall} + \text{rain_on_snow} - \text{sublimation} - \text{runoff} + \text{transport} - \text{adjustment}
\]

Snow water in the CLM3 consists of both liquid and ice phases, so snowmelt has no direct impact on the snow water (snowmelt increases the runoff by increasing the liquid water content of the snow). Transport is neglected in the CLM3. The last term is an adjustment applied to ensure that the snow water never exceeds 1 m. This ad hoc treatment is needed because otherwise snow water accumulates indefinitely in places where neglected terms such as lateral transport are important. G2006 evaluate several of these terms. The annual and area mean of each term for Greenland is shown in Fig. 16. The largest terms in the balance are snowfall and

![Fig. 13. Same as in Fig. 7, but for New Zealand between 47°–34°S and 165°W–180°.](image)
adjustment, which nearly cancel each other out. Snowfall increases with time, but the adjustment increases too. Rain on snow also increases, but more slowly. Table 1 lists values of each term averaged over the periods 1980–2000 and 2080–2100. The storage term becomes increasingly negative with time, increasing from 0.004 mm day$^{-1}$ for 1980–2000 to 0.034 mm day$^{-1}$ for 2080–2100. It is unlikely that this is driven by the adjustment, which serves only to mitigate the increase due to snow and rain. Sublimation changes little, but runoff increases slightly (from 0.063 mm day$^{-1}$ for 1980–2000 to 0.079 mm day$^{-1}$ for 2080–2100), presumably because of increased snowmelt. The disparity between the change in the storage and the change in the runoff can only be explained by the adjustment term, which is counterintuitive. Regardless of the sign of the change in the adjustment term, it is clear that if the adjustment term is active throughout the simulation pe-

Fig. 14, Same as in Fig. 7, but for the Andes between 60°S–0° and 90°–50°W.
period the snow will remain near its upper limit and hence will change little. Clearly the sensitivity of the snow water to the warming is inhibited by the application of the upper bound. Thus, the only other estimate of the impact of climate warming on the Greenland ice mass balance (Ohmura et al. 1996a,b) used methods that did not require treatment of lateral transport or accurate simulation of precipitation. If the mass balance method for simulation snow water (which is used by most current climate models) is to be used to simulate changes in the ice sheets, the precipitation must be simulated.

**Table 1. Greenland snow balance (mm day<sup>−1</sup>).**

| Source                  | 1980–99 | 2080–99 |
|-------------------------|---------|---------|
| Storage                 | −0.004  | −0.034  |
| Snowfall                | 1.254   | 1.419   |
| Rain on snow            | 0.168   | 0.247   |
| Precipitation           | 1.422   | 1.666   |
| Adjustment              | 1.275   | 1.529   |
| Sublimation             | 0.088   | 0.092   |
| Runoff                  | 0.063   | 0.079   |
| Ablation (sublimation + runoff) | 0.151   | 0.171   |

**Fig. 15.** Same as in Fig. 7, but for Mt. Kilimanjaro between 3.5°–2.5°S and 37°–38°E.

**Fig. 16.** Components of the annual mean snow balance for Greenland averaged over latitudes 60°–85°N and longitudes 60°–20°W for the period 1980–2099.
more accurately; and lateral and transport of snow by slides, creep, and glacial flow will be required.

On Mt. Kilimanjaro the absolute change in snow water is less than 0.02 m at the summit, even though the snow water during 1980–2000 exceeds 0.4 m. The snow in the highest elevation class does not melt away, even though the elevation of the highest class is more than 600 m below the actual summit elevation (5893 m). The insensitivity near the summit reflects the colder temperatures there of course, but other factors are at work as well. First, G2006 find that the upper bound on snow water is frequently encountered there during the simulation for 1980–2000. The adjustment associated with the upper bound therefore acts as a snow sink and essentially controls the snow water in the highest elevation class. The snow water in the highest elevation class continues to brush against the 1-m upper bound throughout the simulation, even though snowmelt increases tenfold by 2080–2100. Clearly the upper bound is limiting the sensitivity of the snow water to climate change. The upper bound plays a significant role because, as noted by G2006, the simulated precipitation is excessive on the mountain. Second, as the climate warms, precipitation increases enough that, even though the climate warms, snowfall on the summit changes little. The precipitation increase with climate change may therefore contribute to the stability of the snow. Thus, both the excessive precipitation for 1980–2000 and the simulated increase in precipitation contribute to the stability of the snow water at the summit. The temporal variability in the regional mean snow water is due to variability in the snow water at intermediate elevations. As the climate warms, snow water decreases markedly at those elevations, but not at the summit. Clearly this insensitivity of the snow water on Kilimanjaro is inconsistent with current trends and predictions that its snow would disappear completely this century (Thompson et al. 2002).

Other downscaling studies have found similar sensitivity of snow water to warming. Leung and Ghan (1999) found a 50% reduction in snow water in the Cascade range simulated by an RCM with a 90-km grid size and the same subgrid orography scheme. Leung et al. (2004) found a 70% snow-water reduction in the Cascade and Sierra ranges simulated by an RCM with a 40-km grid size and no subgrid scheme. Govindasamy et al. (2003) found that global high-resolution simulations at T170 (about 0.7° latitude and longitude) were unable to simulate snow water in the Cascade and Sierra ranges realistically enough to determine the impact of climate change on snow water there. Wood et al. (2004) concluded that biases in the temperature and precipitation from the Leung et al. (2004) simulation had to be corrected before they could be used to provide snow water accurate enough for watershed hydrology simulations.

4. Conclusions

We have examined the simulated climate change for the IPCC A1B scenario using a climate model run for the period 1977–2100 with an embedded subgrid orography scheme. We have focused our analysis on three fields: precipitation, surface air temperature, and snow-water equivalent. The mapping of the simulated fields according to a high-resolution distribution of surface elevation provides an unprecedented picture of a climate change scenario. We have also focused our analysis on a set of 10 regions, each with complex terrain but with differences in other attributes. Our analysis leads us to the following general conclusions:

- Changes in precipitation differ from region to region, with precipitation increasing more with increasing altitude in some regions, decreasing more with altitude in others, and changing little in still others. In some regions the sign of the precipitation change depends on surface elevation. Relative changes in precipitation are largely controlled by large-scale processes.
- Changes in surface air temperature are rather uniform on subgrid scales, with at most a twofold difference between the largest and smallest changes within a region. In most cases the warming increases with altitude, reflecting the same trend in the free troposphere.
- Changes in snow water are highly dependent on altitude and hence exhibit strong subgrid variability in regions with complex terrain. Absolute changes usually increase with altitude because more snow is currently available for depletion, but for sufficiently cold conditions snow changes little with global warming. Relative changes always decrease with altitude as the likelihood that a warming will enhance melting or change the phase of precipitation decreases with decreasing temperature at higher altitudes.
- In places where snow accumulates, an artificial upper bound on snow water limits the sensitivity of snow water to climate change considerably. Any climate model that simulates snow water using the mass balance method must apply a similar upper bound unless it treats the physics that limits snow accumulation, namely downward transport by slides, creep, and glacial flow. This upper bound is necessary in regions where glaciers form but is also needed for places where snow accumulates but glaciers do not exist. Most of these spurious glaciers arise because of ex-
cessive precipitation simulated for the 1980–2000 period. In these places the precipitation biases are typically larger than the precipitation change, so the simulated precipitation change has little impact on the snow water. The adjustment to the upper bound continues to control the snow water there. This limits the interpretation of the sensitivity of snow water to warming on Mt. Kilimanjaro and most notably on Greenland but is not a concern for most other regions that are dominated by seasonal snow.

Table 2 summarizes the relative change of the area mean snow water for the periods 2030–50 and 2080–2100. The simulated impact of climate change on the regional mean snow water varies widely, with little impact in regions in which the upper bound on snow water is the dominant snow-water sink (Greenland and Kilimanjaro), moderate impact in regions with a mixture of seasonal and permanent snow (the western United States, Alaska, Scandinavia, the Alps, and the Himalayas), and profound impacts on regions with little permanent snow (New Zealand, Mexico, and the Andes).

These conclusions must be tempered with the recognition of significant limitations in the simulated climate. These limitations and potential solutions are discussed in depth by G2006. Thus, although this work represents a significant improvement over studies without the subgrid scheme (which rarely even address changes in snow water, except for one particular region if a regional model is used), further work is required to improve the climate simulation and reduce uncertainty in projections of the impacts of climate change on snow water and water resources in regions with complex terrain.

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