We introduce a physically-based, multilayer snow radiative transfer into the NCAR Community Land Model 3, improving representation of the vertical distribution of shortwave (SW) heating within snowpacks. We show that 37-70% of the SW absorption by deep snowpacks occurs beneath the top 2 cm layer where it is prescribed to occur in the present model, leading to snow warming relative to the current model. Prediction of snow depth and 2-meter air temperature over the Tibetan Plateau (TP), where incident SW flux is relatively intense, is greatly altered and generally improved relative to observation without worsening predictions elsewhere. Snow cover on the TP can affect the strength and timing of the South Asian Monsoon, and snowmelt from this region feeds into rivers that supply freshwater to half of the world’s population.

1. Introduction

The Blanford Hypothesis [Blanford, 1884] states that the strength of the South Asian Monsoon is directly related to the springtime quantity of snow on the TP (30°-40°N, 80°-100°E). Less residual snow in late spring and summer enhances the land-ocean temperature contrast on the Indian subcontinent, driving stronger monsoonal flow in summertime. This hypothesis has undergone many supportive and critical [e.g., Robock et al., 2003] evaluations. Fasullo [2004] recently showed that in non-ENSO (El Nino Southern Oscillation), strong-monsoon years there is a significant negative correlation between snow cover fraction over Tibet and the Himalayas during December-February (DJF) and June-September rainfall over all of India, lending conditional support to the hypothesis. Tibetan winter snow cover has also been linked to summer rainfall over the Yangtze River Valley [Wu and Qian, 2003], and snow cover that lingers into the late spring acts as a constraint on dust emissions from this region [Zender]. Finally, snow melt and glacial meltwater on the TP helps feed ten of Asia’s largest rivers, bringing freshwater to roughly half of the world’s population. For these reasons, accurate representations of snow processes on the TP by global climate models is crucial.

GCMs represent snow radiative processes in simplified manners, generally accounting for snow reflectivity with empirical functions and not accounting for realistic vertical SW heating. For example, the Community Land Model 3 (CLM3) [Oleson et al., 2004], the land component of the NCAR Community Atmosphere Model (CAM3) [Collins et al., 2004], prescribes all SW absorption by the snowpack to occur in the top 2 cm layer. We show this to be unrealistic, especially in snowy regions of intense incident solar radiation such as the TP. Un-
realistic representations of snow radiative processes may be responsible for the nearly ubiquitous overprediction of snow depth on the TP that Foster et al. [1996] found in an analysis of seven GCMs, and for the 5-fold overprediction by CAM3. Sophisticated snow and ice radiation models have been developed [e.g. Wiscombe and Warren, 1980; Jordan, 1991; Grenfell, 1991], but not incorporated into GCMs. In this paper we assess some of the climate implications of applying realistic snow radiative and heating processes in the TP region.

2. Methods

We employ a multi-layer two-stream radiative transfer model which computes fluxes at each layer interface and depends on snow grain radius, aerosol concentrations, solar-zenith angle, layer thickness and density, ratio of diffuse to direct incident flux, and reflectance of the underlying surface. We use Mie Theory offline to compute single-scattering albedo ($\omega_\lambda$), asymmetry parameter, and extinction coefficient for collections of ice spheres of different mean radii at fine spectral resolution from 0.3-5.0 $\mu$m. Broadband Mie parameters matching the two spectral bands used by CLM (0.3-0.7 $\mu$m and 0.7-5.0 $\mu$m) are computed by weighting the fine-resolution parameters with top-of-atmosphere solar flux ($F_\lambda$). Furthermore, because of the dominant role of $\omega$ in determining bulk snowpack albedo, and the non-linear relationship between $\omega$ and bulk albedo, $\omega$ is also weighted by the semi-infinite, diffuse-light snowpack albedo ($A_\lambda$), computed using the method from Wiscombe and Warren [1980]. Hence, the broadband single-scattering albedo ($\overline{\omega}$) is computed as follows:

$$\overline{\omega} = \frac{\sum_\lambda^{\lambda_1} R_\lambda \cdot A_\lambda \cdot \omega_\lambda}{\sum_\lambda^{\lambda_1} R_\lambda \cdot A_\lambda}$$ (1)

This approach yields much better agreement between broadband snowpack albedo computed from single-band Mie parameters and the exact, spectrally-integrated albedo, especially in the near-IR. Because these calculations are computationally expensive, they are done offline and a lookup table is created for use in climate simulations.

To represent bulk radiative transfer, Mie parameters are transformed using the Delta-Eddington Approximation [Joseph et al., 1976] in the visible, and Hemispheric-Mean Approximation [Toon et al., 1989] in the near-IR. We use the Hemispheric-Mean Approximation in the near-IR because the Delta-Eddington Approximation can predict negative albedos for diffuse incident light in some highly-absorptive regions of the spectrum. Both approximations are good for strongly forward-scattering mediums like snow [Toon et al., 1989]. Furthermore, use of the Hemispheric-Mean Approximation allows replication of satellite-observed albedo over Greenland [Zhou et al., 2003] in both the visible and near-IR bands using a single snow grain radius. Finally, we apply the multi-layer radiative transfer solution presented by Toon et al. [1989] which utilizes a tridiagonal matrix solution to compute flux at each layer interface.

CLM represents snow columns with up to five layers, allowing for heterogeneity in density (from snow compaction), thermal diffusion, melting and freezing processes, and solid vs. liquid phase state [Oleson et al., 2004]. We assimilated SNICAR into CLM, replacing the original snow reflectance model, matching radiative lay-
ers with snow layers already defined by the model. We performed two sets of experiments. The first set was with CLM in offline mode, forced with 1990’s data from the National Center for Environmental Prediction (NCEP) reanalysis project [E. Kalnay, 1996]. The second set was with the CLM coupled to the atmosphere model CAM3, forced with climatologic mean sea surface temperatures.

Table 1 summarizes the simulations that we conducted. The 'A' simulations were controls with the original snow physics of CLM. Bulk surface albedo is predicted with SNICAR in the 'B' simulations, but we artificially prescribe all SW absorption to occur in the top snow layer. We chose 100 µm as the mean snow radius because it yields comparable visible and near-IR albedos observed by satellite [Zhou et al., 2003] [Nolin?]. The 'C' simulations predict the same snow reflectance as 'B', but realistic SW absorption occurs in subsurface layers.

Finally, we allow for snow aging in simulation CAM-D. Snow grain metamorphism causes snow albedo to decrease with time. The rate of grain growth (albedo reduction) increases with increasing temperature. CLM accounts for snow aging with an exponential decay function that depends on time and temperature [Oleson et al., 2004]. In the absence of available observational data of snow albedo vs time and temperature, we preserved the aging characteristics of CLM snow albedo in the near-IR by developing a linear relationship between CLM’s artificial snow age, \( \tau \), and snow grain radius, \( r \):

\[
r = r_0 + \alpha \cdot \tau
\]

The slope, \( \alpha \), of this function was found as follows: 1) Find the snow grain size which yields CLM near-IR diffuse snow albedo over the domain: \( 230 \text{ K} \leq T_{\text{snow}} \leq 273 \text{ K} \), and \( 0 \leq \text{SnowAge} \leq 10 \text{ days} \). 2) Find the slope of the linear function relating \( \tau \) to \( r \) which minimizes the error between snow grain size from (1) and snow grain size from equation 2, over this domain. This was accomplished with a multi-dimensional regression technique utilizing the Nelder-Mead Simplex (function minimization) method [Lagarias et al., 1998]. The advantage of this method is that it can easily be applied to a new set of data describing the evolution of snow albedo in the time and temperature domains. For now, however, we use essentially the same aging relationship as CLM, but described with snow grain size. Finally, because subsurface layers are older and contain larger-grained snow [Aoki et al., 2000], we linearly ramp up the snow radius in descending subsurface layers to the maximum snow radius of 1000 µm.

The offline simulations were run for 10 years, and the coupled simulations for 15 years. 15 years was ample because in our most-altered experiment (D), the rate of change in global annual mean snow depth dropped to less than 1%/yr within 6 years and net top-of-atmosphere radiative flux showed no significant trend, averaging +0.79 W m\(^{-2}\). The resolution of all runs was T42, 2.8° \( \times \) 2.8° resolution.

3. Results and Discussion

3.1. Model Sensitivity to Subsurface Snow Heating

We first analyze how SNICAR predicts SW radiative absorption in a single snow column compared with how CLM prescribes it. Figure 1 shows the SW absorption in a 20 cm snow column, assuming the TP climatologic mean incident surface radiation of 135 and 137 W m\(^{-2}\)
in the visible and near-IR bands, respectively. CLM visible and near-IR albedos are 0.95 and 0.65, respectively; those of fresh snow with diffuse incident radiation. With SNICAR we assume that snow density increases linearly from 190 kg m$^{-3}$ at the top to 270 kg m$^{-3}$ at 20 cm depth, a rate derived from mean snow densities predicted by CLM. As illustrated by the figure, significant absorption occurs below 2 cm for all snow grain sizes. 37%, 60%, 69%, and 73% of the net column absorption occurs below 2 cm with 100, 250, 500, and 1000 μm grain radius snowpacks, respectively. Radiation penetrates deeper into larger-grained snowpacks, resulting in greater heating rates with depth. Because our choice of 100 μm grain radius for the subsurface heating experiment is smaller than what is typically measured in the field, results that we ascribe to subsurface heating may be underestimated.

To isolate the climatologic effects of subsurface heating, we compare Simulations B and C. 'B' is the control and 'C' the experiment in this case. Figure 2 depicts the climatologic-mean global changes in column-integrated snow temperature and 2-meter air temperature for the offline and coupled experiments. Hatching is where there is a statistically-significant difference at the 1% confidence level. Table 2 summarizes the globally- and TP-averaged changes for this experiment. Snow temperature is averaged spatially and temporally only over gridcells that have snow.

While the surface snow layer cools because of less SW absorption, all subsurface layers warm because of increased absorption, leading to column-integrated snow warming in both cases. Snow warming over Antarctica is the greatest, topping 5°C in both experiments. Northern Hemisphere snow warming ranges from 0 to +4°C in the offline experiment, and from -1 to +2°C in the coupled experiment. Snow warming is slightly greater over the TP than most other land because the incident SW flux is relatively large at its latitude, resulting in greater subsurface absorption. However, snow temperatures in the TP tend to be near the melting point in the controls, constraining the possible warming.

While the global 2-meter air warming in both experiments is negligible, warming over the TP is +2 °C in the coupled experiment, and the TP is the only region of statistically significant warming in the offline experiment. Warming is greatest over the TP because subsurface melting within the snowpack occurs, reducing snow depth and therefore snow fraction, significantly lowering the areal albedo. Surface air warming is much greater in the online experiment than in the offline experiment because of atmospheric response. Temperature from the lowest atmospheric level (992mb) is prescribed in the offline runs, placing a strong constraint on surface temperature feedbacks. As the snow column warms and melts in the more realistic online experiment, bare ground becomes exposed, allowing greater surface warming and SW absorption because of lowered gridcell-albedo. The warm surface air aids in melting more snow, completing a feedback that is not present in the offline experiment. Air warming does not occur over Antarctica in spite of snow column warming because the surface snow layer temperature change is negligible and subsurface warming is not great enough to induce melting, so there is no change in snow fraction.

### 3.2. Model Comparison with Observation

We have compared the annual cycle of online model-
predicted snow depth over the TP with three observational datasets. The global snow depth climatology assembled by the US Air Force Environmental Technical Applications Center (USAF/ETAC) [Foster and Davy, 1988] is the only global snow-depth dataset based on ground measurements covering an annual cycle. The second observational dataset is snow-depth derived from the Nimbus-7 Scanning Multi-channel Microwave Radiometer (SMMR) from November 1978 to August 1987 [Chang et al., 1992; Chang, 1995], retrieved on a $0.5^\circ \times 0.5^\circ$ grid. Third, we use global snow water equivalent derived from the Advanced Microwave Scanning Radiometer (AMSR-E) for 2003 [Chang and Rango, 2004]. For the sake of comparison, we converted snow water equivalent to snow depth by using the mean TP snow density predicted by CLM. All three datasets were averaged monthly over the TP.

Figure 3 compares snow depth from the three observational datasets, our online control run, and CAM-D, which we deem the most realistic experiment. While the AMSR time-series duration only covers one year, it closely matches the 9-year SMMR average annual cycle, differing by no more than 4 cm in any month. Both satellite-derived datasets show greater snow depth over the winter months than USAF/ETAC, which shows average snow depth less than 2 cm all year. Microwave sensors predict snow depth poorly over mountainous terrain, and often cannot detect shallow, dry snow less than 5 cm thick [SMMR]. Furthermore, Gao [2003], who compared SSM/I data with ground measurements over the TP, hypothesizes that microwave sensors overpredict snow on the TP because the permafrost produces a similar microwave signature as snow. Based on this data, we assume that average December-May snow depth on the TP is less than 10 cm.

The CAM-A control simulation predicts average TP Dec-May snow depth of 53 cm, while CAM-D predicts 5.5 cm, agreeing very well with the observational data. CAM-D snow depth closely resembles satellite data from Jun-Feb, but predicts snow depth closer to that of the USAF/ETAC data during spring. Globally, the change in snow depth between CAM-A and CAM-D was -5.1 cm, much less than over the TP. A visual inspection of global CAM-A and CAM-D differences with USAF/ETAC snow depth shows that CAM-D generally agrees slightly better with observation, especially over the western U.S., Canada, Alaska, and eastern Siberia (could present figure...).

Finally, we compare the seasonal cycle of 2-meter surface air temperature over the TP from CAM-A and CAM-D with the Willmott-Matsuura mean 1990s observational temperature [Willmott and Matsuura, 2000], shown in Figure 4. Globally, 2-meter temperature was only 0.18°C warmer in CAM-D relative to CAM-A, but over the TP, it was 2.43°C warmer. Depicted in Figure 4, CAM-A has a mean MAM cold bias of nearly 5°C, likely due to its gross overprediction of snow depth. CAM-D fairs better, predicting temperatures within 1°C of observation from Mar-Oct, but slightly overpredicting during Nov-Feb. The annual mean bias of CAM-A is -1.93°C, and that of CAM-D is +0.48°C.

4. Conclusion

We have shown that accounting for a realistic vertical distribution of SW heating within a snowpack can have a large influence on GCM-prediction of snow temperature,
snow depth, and surface air temperature. This is especially true of low- and mid- latitude snowy regions such as the Tibetan Plateau, where mean incident radiation is relatively intense. With a more realistic snow model, prediction of snow depth and surface air temperature over the TP is improved in a coupled land-atmosphere model. Much of this improvement is likely due to the change in albedo associated with change in snow depth in the model. More research is needed to assess the realism of predicted surface albedo by the new snow model and to evaluate new relationships relating snow cover fraction to snow depth.

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diometer (MODIS) and Common Land Model, *J. Geophys. Res.*, 108(D15), doi:10.1029/2002JD003326, 2003.
## Table 1. Summary of Simulations

| Offline Simulation | Coupled Simulation | Snow Model | Grain Size | Shortwave Absorption |
|--------------------|--------------------|------------|------------|----------------------|
| CLM-A              | CAM-A              | Original CLM | -          | Top Layer            |
| CLM-B              | CAM-B              | SNICAR     | 100 µm     | Top Layer            |
| CLM-C              | CAM-C              | SNICAR     | 100 µm     | All Layers           |
| -                  | CAM-D              | SNICAR     | 100-1000 µm | All Layers           |

**Figure 1.** Comparison of shortwave absorption with depth in a snow column as prescribed in the Community Land Model and predicted with a physically-based radiative transfer model using different sizes of snow grains.
Figure 2. Global differences in snow temperature and 2-meter air temperature between simulations with and without subsurface shortwave absorption in snowpacks. Top-left: Difference in snow temperature with offline models. Top-right: Difference in snow temperature with coupled models. Bottom-left: Difference in 2-meter air temperature with offline models. Bottom-right: Difference in 2-meter air temperature with coupled models.

Table 2. Snow temperature and 2-meter air temperature over the globe and Tibetan Plateau

|            | Snow Temp | 2-meter Air Temp |
|------------|-----------|------------------|
|            | Global    | Global           |
| CLM-C - CLM-B | +1.31     | +1.86            |
| CAM-C - CAM-B | +0.74     | +1.28            |
|            | TP        | TP               |
| CLM-C - CLM-B | +0.03     | +0.25            |
| CAM-C - CAM-B | +0.07     | +1.99            |
Figure 3. Seasonal cycle of observed and modeled snow depth over the Tibetan Plateau. Observations are from in-situ measurements and microwave observations from satellite. Simulated snow depths are from models with and without subsurface shortwave absorption.

Figure 4. Observed and simulated 2-meter air temperature over the Tibetan Plateau.