Complexity of Snow Schemes in a Climate Model and Its Impact on Surface Energy and Hydrology

EMANUEL DUTRA*
Centro de Geofísica da Universidade de Lisboa, Instituto Dom Luiz, University of Lisbon, Lisbon, Portugal, and Institute for Atmospheric and Climate Science, ETH, Zurich, Switzerland

PEDRO VITERBO
Institute of Meteorology, and Centro de Geofísica da Universidade de Lisboa, Instituto Dom Luiz, University of Lisbon, Lisbon, Portugal

PEDRO M. A. MIRANDA
Centro de Geofísica da Universidade de Lisboa, Instituto Dom Luiz, University of Lisbon, Lisbon, Portugal

GIANPAOLO BALSAMO
European Centre for Medium-Range Weather Forecasts, Reading, United Kingdom

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ABSTRACT

Three different complexity snow schemes implemented in the ECMWF land surface scheme Hydrology Tiled ECMWF Scheme of Surface Exchanges over Land (HTESSEL) are evaluated within the EC-EARTH climate model. The snow schemes are (i) the original HTESSEL single-bulk-layer snow scheme, (ii) a new snow scheme in operations at ECMWF since September 2009, and (iii) a multilayer version of the previous. In offline site simulations, the multilayer scheme outperforms the single-layer schemes in deep snowpack conditions through its ability to simulate sporadic melting events thanks to the lower thermal inertial of the uppermost layer. Coupled atmosphere–land/snow simulations performed by the EC-EARTH climate model are validated against remote sensed snow cover and surface albedo. The original snow scheme has a systematic early melting linked to an underestimation of surface albedo during spring that was partially reduced with the new snow schemes. A key process to improve the realism of the near-surface atmospheric temperature and at the same time the soil freezing is the thermal insulation of the snowpack (tightly coupled with the accuracy of snow mass and density simulations). The multilayer snow scheme outperforms the single-layer schemes in open deep snowpack (such as prairies or tundra in northern latitudes) and is instead comparable in shallow snowpack conditions. However, the representation of orography in current climate models implies limitations for accurately simulating the snowpack, particularly over complex terrain regions such as the Rockies and the Himalayas.

1. Introduction

Accurate simulations of the snow cover strongly impact on the quality of weather and climate prediction as the absorption of solar radiation at the land–atmosphere interface is modified by a factor up to 4 in response to the presence of snow (albedo effect) (Viterbo and Betts 1999). Changes in snow cover modify the surface albedo, which then feedbacks to surface temperature (Groisman et al. 1994). In northern latitudes and mountainous regions, snow also acts as an important energy and water reservoir (Immerzeel et al. 2010). Therefore, a correct representation of snow mass and density is crucial for predicting the snow thermal insulation, especially in climate change conditions, with direct consequences for soil hydrology and ground temperature (Bartlett et al. 2011).
2. Snow schemes

The three snow schemes coupled to HTESSEL are compared in Table 1 with respect to the number of snow layers, liquid water representation, snow density, and snow albedo parameterizations. CTR (ECMWF 2011) was the snow scheme used at ECMWF from the late ‘90s until 2009, including the 40-yr ECMWF Re-Analysis (ERA-40) (Uppala et al. 2005) and ERA-Interim (ERAI) (Dee et al. 2011) reanalyses. In 2009, a new snow scheme, OPER, was introduced in the integrated forecast system at ECMWF, and its description, validation, and comparison against CTR in offline mode is presented by DU10. A multilayer version of OPER (ML3) was developed for this study and its main characteristics are presented in the appendix. The three snow schemes are identically coupled with the overlying atmosphere and underlying soil within HTESSEL.

The snow scheme represents one (CTR and OPER) or up to three (ML3) additional layers on top of the upper soil layer, with independent prognostics thermal and mass contents. The net energy flux at the top of the snowpack is the residual of the skin energy balance from
3. Site simulations

Data from the micrometeorological site in Col de Porte (CdP) was used to force, validate, and intercompare the three snow schemes. CdP is operated by the Center of Snow Studies in Grenoble, France, and is located at 45°N, 6°E at an altitude of 1320 m. Data from this site has been widely exploited in model validation (e.g., Boone and Etchevers 2001) and was one of the sites that participated in the first Snow Model Intercomparison Project (SnowMIP; Etchevers et al. 2004). Offline simulations of snow mass (SWE), density, and runoff for the winter seasons of 1996/97 and 1997/98 were validated against CdP observations (Fig. 1). From November to mid-January, in both years, all the simulations properly represent the snow accumulation. From February to May both CTR and OPER tend to overestimate SWE, whereas ML3 is closer to the observations. The final ablation is delayed in all simulations with a stronger overestimation by CTR. The mean absolute error (MAE) of SWE was 98, 82, and 42 kg m\(^{-2}\) during the 1996/97 winter in CTR, OPER, and ML3, respectively, and 128, 84, and 40 kg m\(^{-2}\) during the 1997/98 winter. On average for the two winter seasons, the MAE of SWE reduced 25% from CTR to OPER and 64% from CTR to ML3. Snow density evolution is correctly captured by OPER and ML3. CTR shows a fast increase and saturation at 300 kg m\(^{-3}\), similar to what was found by DU10, and by Bartlett et al. (2006) in other site locations, when using the CTR snow density parameterization. Pronounced density changes due to snowfall events are evident in the observations. In mid-April 1998, the sharp increase of SWE is accompanied by a significant decrease of snow density that is poorly simulated by OPER but correctly captured by ML3.

The accumulated runoff (Figs. 1c,f) is defined as the liquid precipitation and snowmelt that is in excess of the snow-cover holding capacity. Its temporal evolution gives an integrated view of the snowpack water balance. The difficulties of the three snow schemes in simulating short-duration melting events during late winter and spring are evident. These melting episodes are mainly driven by synoptic conditions, and are poorly represented by CTR and OPER. On the other hand, ML3 partially reproduces those episodes with melting in the first layer. These results clearly show one of the main drawbacks of a single snow layer associated with the large thermal inertia of thick snowpacks (Marks et al. 2008). The multilayer scheme has an enhanced accuracy in simulating midseason sporadic melting events, since the top snow layer has a reduced thermal inertia.

Surface albedo and surface temperature simulation are compared against observations for the 1997/98 winter season at CdP in Fig. 2. The overestimation of SWE and delayed melting in CTR is explained by its overestimation of snow albedo and limitations in representing the albedo decay between snowfall events. On the other hand, both OPER and ML3 (which share the same snow albedo parameterization) are able to represent the albedo decay between snowfall events, especially during

<table>
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<tr>
<th>Snow scheme</th>
<th>Number of layers</th>
<th>Liquid water</th>
<th>Snow density</th>
<th>Snow albedo</th>
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<tbody>
<tr>
<td>CTR (ECMWF 2011)</td>
<td>1</td>
<td>Not represented (dry snow)</td>
<td>Exponential increase with age</td>
<td>Exponential decay with age; 0.15 for shaded snow</td>
</tr>
<tr>
<td>OPER (DU10)</td>
<td>1</td>
<td>Diagnosed from temperature, mass, and density of the snowpack</td>
<td>Overburden and thermal metamorphisms, meltwater retention</td>
<td>Exponential decay with age; vegetation type dependent for shaded snow</td>
</tr>
<tr>
<td>ML3 (appendix)</td>
<td>Up to 3</td>
<td>Bucket-type prognostic in each layer</td>
<td>As OPER but solved for each layer</td>
<td>As OPER</td>
</tr>
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spring. However, in the beginning of April 1998, there was a sharp decrease of surface albedo that both OPER and ML3 did not simulate. This was a critical point for an accurate final ablation, which would have triggered melting because of the high available energy. The simulations of surface temperature (skin temperature) are in good agreement with the observations, except during late January and early February where all the simulations underestimate surface temperature, particularly ML3.

Other site locations were tested (from SnowMIP2; Rutter et al. 2009) and showed similar results with significant improvements from CTR to OPER (as presented by DU10). Figure 3 compares the root-mean-square error of SWE simulations in open and forest plots with different configurations of the model. In open sites there is an improvement from CTR to OPER and to ML3, while in forest sites ML3 is similar to OPER. In open sites, there is also an improvement when increasing the number of snow layers from two (ML2) to three (ML3), but with five layers (ML5) the results are similar to ML3. Changing the minimum snow layer depth (see appendix) from 0.2 to 0.1 m with three or five snow layers (ML3* and ML5*) worsens the results because of the time-implicit coupling with the atmosphere. The reduction of the minimum snow layer depth leads to a reduction of the thermal capacity of the uppermost snow layer and its associated time scale in response to the energy forcing. Further analysis is needed to fully understand this impact and its relation with the algorithmic details of the
implicit coupling to the atmosphere. From OPER to ML3 the improvements are mainly restricted to deep snowpack locations, while for shallow snowpacks and highly forested locations ML3 returns very similar results to OPER. These findings suggest that over forested regions other processes, such as radiation interaction with the canopy and snow interception by the canopy, should be also addressed in the model formulation.

4. Atmospheric simulations

The atmospheric model of EC-EARTH version 2 (Hazeleger et al. 2010) is based on ECMWF’s Integrated Forecasting System (IFS), corresponding to the current seasonal forecast system of ECMWF (system 3). The model runs at TL159 horizontal spectral resolution (about 125 km on the gridpoint space) with 62 vertical levels that extend up to 5 hPa. Some aspects of newer IFS cycles have been implemented, particularly the new land surface scheme HTESSEL. EC-EARTH also has ocean and sea ice model components, but in this study atmosphere-only runs were performed. The boundary conditions of sea surface temperature and sea ice cover are prescribed as time-varying monthly fields derived from the ECMWF reanalysis ERA-40 up to December 1988 and ERA-Interim onward. This setup was adopted over coupled atmosphere–ocean simulations to avoid the long-term spinup necessary for the ocean, and to confine the impacts of the different snow schemes to the atmospheric evolution and variability.

Both CTR and OPER snow schemes are available in EC-EARTH—the latter being the default—and ML3 was implemented. A three-member 30-yr-long ensemble (January 1979–December 2008) was performed for each of the snow schemes. Each ensemble member differs in its initial conditions: 1 November 1977, 1 December 1977, and 1 January 1978, taken from ERA-40. The first months of the simulation (until December 1978) were discarded from the analysis. The results will be presented as the ensemble mean unless otherwise stated. Throughout the discussion the statistical significance of the differences between the mean of the simulations (monthly or seasonal means) is assessed using a two-sample two-tailed Student’s t test assuming equal variances (Wilks 2006). In the following sections, results are analyzed over the Northern Hemisphere landmasses.
poleward of 40°N, excluding Greenland (hereafter NH40). This region was selected as representative of the area where seasonal snow cover plays an important role in controlling the surface water and energy fluxes.

a. Snow cover and surface albedo

Simulated snow cover is compared with the weekly snow-cover data from the National Snow and Ice Data Center (NSIDC) (Armstrong and Brodzik 2007). The weekly files consist of gridded binary counts of snow or no snow at 25-km resolution available from 1979 to 2006. This dataset has been used to validate climate model simulations (e.g., Roesch 2006) and to analyze the spatial and temporal patterns of the snow season length during the last three decades of the twentieth century (Choi et al. 2010). The weekly data were averaged into monthly means and aggregated to the model grid.

The simulated Northern Hemisphere snow-covered area (SCA) is compared against NSIDC in Fig. 4. The bias was calculated for each month (synchronous in time) and normalized by the mean NSIDC SCA. CTR underestimates SCA throughout the cold season, with particular problems during spring because of early melting. OPER significantly reduces the biases, especially during spring. ML3 shows similar results during early spring, but in May and June is outperformed by OPER. The mean NSIDC snow cover and the model biases during spring are represented in Fig. 5. The early melting in CTR is marked in central Canada, eastern Europe, Scandinavia, and north of the Himalayan range. Over those regions, both OPER and ML3 significantly reduce the early melting bias of CTR.

The Moderate Resolution Imaging Spectroradiometer (MODIS) albedo product MCD43C3 provides data describing both directional hemispheric reflectance (black-sky albedo) and bihemispherical reflectance (white-sky albedo) in seven different bands and aggregated bands. Data from the Terra and Aqua platforms are merged in the generation of the product that is produced every 8 days, with 16 days of acquisition, and available on a 0.05° global grid. The accuracy and quality of the product has been studied by several authors in different locations (e.g., Roman et al. 2009; Salomon et al. 2006; Stroeve et al. 2005). The MODIS product has already served as reference for model validation (e.g., DU10; Wang and Zeng 2010; Zhou et al. 2003). As for the NSIDC snow cover, the MODIS albedo was averaged to monthly fields and aggregated to the simulations grid.

In this study the mean annual cycle of MODIS white-sky broadband shortwave albedo (2001–09) is compared against the model simulations (1979–2008) and is represented in Fig. 6 for the NH40 region. OPER and ML3 have an almost coincident mean annual cycle on the hemispheric scale. This is expected since OPER and ML3 share the same snow albedo parameterization, and their evolution of snow cover is also similar (see Fig. 4). From July to November all simulations overestimate surface albedo, whereas CTR is closest to MODIS from October to December. This overestimation of OPER and ML3 was also identified by DU10 in OPER and is mainly related with the snow albedo parameterization, since such a signal is not visible on snow cover. From March to June, OPER and ML3 are similar to MODIS while CTR shows an underestimation that is associated with the early melting discussed before. The mean spring surface albedo maps from MODIS and simulation differences are represented in Fig. 7. The strongest underestimation of spring surface albedo in CTR is mainly localized over snow-covered regions poleward of

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**Fig. 3.** Boxplot summaries describing the normalized RMSE of SWE for different model configurations, combined at all SnowMIP2 locations at (a) open sites and (b) forest sites (nine winter simulations each; see Rutter et al. 2009 for a list and description of the sites). The boxes have horizontal lines at the lower quartile, median, and upper quartile and the whiskers extend from min to max. The configurations ML2 and ML5 are similar to ML3, but with max of snow layers set to two and five, respectively. The configurations ML3* and ML5* are similar to ML3 and ML5 but change minimum snow layer depth from 0.2 to 0.1 m (see appendix).
60°N. Both OPER and ML3 reduce those biases. However, over the Himalayan range and Rockies, OPER and ML3 reinforce the overestimation of surface albedo that was already visible in CTR. This overestimation of surface albedo in complex terrain regions is associated with an overestimation of snow cover (see Fig. 5). These results highlight a current deficiency of the model in representing snow cover over complex terrain regions. This is not surprising, since none of the snow schemes include parameterizations for subgrid-scale processes. For such processes, the multilayer representation of the snowpack does not improve the simulations, indicating that further improving the model might need to address the representation of other important processes such as subgrid-scale variability.

The early spring snowmelt and underestimation of snow albedo in CTR are closely related, but it is not possible to clearly identify a cause–effect: low snow albedo leading to excessive energy absorbed on the snowpack resulting in an early snow ablation, or the early snow ablation exposing the land surface (with an albedo lower than a snow surface) leading to the underestimation of surface albedo. It is most likely to result from the conjunction of the two factors associated with the SAF.

b. Surface water balance

The mean annual cycle of the surface water fluxes averaged over NH40 are presented in Fig. 8a for the CTR simulation. Figures 8b–e illustrate the differences
The sign of the fluxes follow the model convention with positive for incoming water (or energy) to the surface and negative for outgoing water (or energy) from the surface. The runoff and terrestrial water storage variation (TWSV) exhibit a distinct behavior in spring and summer in both OPER and ML3 when compared with CTR. There is a significant reduction of runoff during spring (peak in April) matching an increase of soil water storage that is also evident in the mean soil moisture (see Fig. 9b). During summer there is the opposite signal with higher soil moisture depletion in OPER and ML3 than in CTR owing to higher runoff rates.

OPER and ML3 also show a higher snow mass peak in winter and delayed snowmelt (Fig. 9a) that contribute to the reduction of runoff. The delayed snowmelt was discussed in the previous section in terms of snow cover. The delayed depletion of snow mass in OPER and ML3 is also associated with the interception of rainfall by the snow schemes (Table 1). The increased soil moisture availability during summer leads to higher evapotranspiration rates in OPER and ML3. Following the increased moisture supply to the atmosphere, there is also an increase of the precipitation rates due to local–regional recycling and to the soil–precipitation feedback (Schar et al. 1999). Compared with OPER, ML3 has a slightly higher spring and summer soil moisture content, which explains the higher evapotranspiration rates and, consequently, precipitation changes in respect to CTR.

Figure 10 compares the mean annual cycle of precipitation of the coupled simulations against ERA-Interim and several observational-based global datasets: Climate Research Unit (CRU version 3.1 updated by Mitchell and Jones 2005), the Global Precipitation Climate Centre (GPCC version 4.0; Schneider et al. 2010; see http://gpcc.dwd.de/), the Global Precipitation Climatology Project [GPCP version 2.1 (Huffman et al. 2009) and version 2.2] and the Climate Prediction Centre (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997). The mean annual cycles were evaluated for the period 1979–2006 in all the datasets and for different regions: the Himalayas, Rocky Mountains, and NH40. In the NH40 region (Fig. 10c) the uncertainty between the different datasets reaches about 25% of the amplitude of the annual cycle. In this region, the coupled simulations...
are in good agreement with the datasets, apart from an overestimation during spring. It is also possible to identify the increase of precipitation in OPER and ML3 (discussed before in Fig. 8). In both the Himalayan and Rocky Mountains regions, the coupled simulations tend to overestimate precipitation during winter and underestimate it during summer. However, these results highlight the large uncertainty among these global datasets; its comparison with the coupled simulations is not trivial and is beyond the scope of the present work. Moreover, this comparison does not evaluate the partitioning between solid and liquid phases, which is of paramount importance in snow-cover simulations (Dutra et al. 2011a). Nevertheless, the coherent signal of overestimation of precipitation over the Himalayan and Rockies regions partially explains the overestimation of snow cover (see Fig. 5) and surface albedo (see Fig. 7) during spring.

c. Surface energy balance

Following the analysis of the water balance components in the previous section, the mean annual cycles of the surface energy fluxes, averaged over NH40, are represented in Fig. 11. There is a significant reduction of surface solar radiation (SSR—downward minus upward) from March to May in both OPER and ML3. This reduction is mainly associated with the increased surface albedo and delayed snowmelt (see Figs. 6 and 9a). There is a reduction of both downward and upward thermal radiation resulting from the cooling of the atmosphere and surface from CTR to OPER and ML3 (see Fig. 12). This results in an increase of surface thermal radiation (STR—downward minus upward), which is more pronounced during winter. The decrease (positive differences) of the surface sensible heat flux (SSHF) in OPER...
and ML3 is also associated with the cooling of the atmosphere and with the reduction of SSR. The surface latent heat flux (SLHF) shows a significant increase (negative differences) in summer due to increased evapotranspiration rates (see Fig. 8c). The surface net radiation reflects the combination of the above-mentioned changes, with a warming of the soil and cooling of the atmosphere during winter in OPER and ML3 when compared with CTR.

The mean annual cycle of 2 m (T2M, or air temperature), skin, and soil temperature (surface to 7-cm depth) in Fig. 12 are consistent with the changes in surface fluxes presented in Fig. 11. The changes in surface fluxes, as well as temperature, are larger in ML3 than in OPER when compared with CTR. These changes are mainly due to the increased snow thermal insulation of the atmosphere from the underlying soil, which is much stronger in ML3. In CTR, heat exchanges between the atmosphere and the surface during winter led to colder soil temperature and warmer air temperature. OPER simulates lower snow densities (see DU10 for details), leading to a stronger insulation of the atmosphere from the soil. This effect is stronger in ML3 since the multilayer snow scheme also accounts for the thermal insulation within the snowpack (i.e., the upper layer thermally insulates the lower layers).

The simulated T2M is compared against the CRU Time Series version 3.0 (TS3.0) global dataset of monthly temperature (updated from Mitchell and Jones 2005) for the period 1979–2006. Over NH40 (see Fig. 12a) the T2M mean bias during winter was reduced from +2.2 K in CTR to 0.9 K in OPER and −0.1 K in ML3. The mean winter differences between CTR and CRU T2M (Fig. 13a) show a pronounced warm bias, reaching 8 K over snow-covered regions. Both OPER and ML3 lead to a significant cooling over snow-covered regions. Over the Himalaya and Rockies, CTR presents a negative T2M bias that was intensified in OPER and ML3. Part of the cold bias can be associated with an overestimation of snow cover and surface albedo (see Figs. 6 and 7) that is related with an overestimation of precipitation during winter in those regions (see Fig. 10).

d. Snow–albedo feedback

The SAF can be quantified by the variation in net incoming shortwave radiation (Q) with surface air temperature (Tair) because of changes in surface albedo (αs) (Cess and Potter 1988; HQ06):

\[
\left( \frac{\partial Q}{\partial T_{air}} \right)_{SAF} = -I_i \frac{\partial \alpha_s}{\partial \alpha_s} \Delta \alpha_s, \tag{1}
\]

where \( \alpha_p \) is the planetary albedo, \( I_i \) the incoming solar radiation at the top of the atmosphere, and the subscript SAF denotes that the partial derivative refers only to changes in Q with Tair due to changes in surface albedo (excluding all the other factors affecting solar radiation). The SAF is decomposed as the product of two terms in Eq. (1): (i) \( (\partial \alpha_s/\partial \alpha_s) \) accounts for the attenuation effects of the atmosphere on anomalies in \( \alpha_s \) (due to solar absorbers in the atmosphere, including clouds) and (ii) \( (\Delta \alpha_s/\Delta T_{air}) \) represents the changes of surface albedo induced by changes in surface air temperature resulting from surface processes.

HQ06 proposed the calculation of \( (\Delta \alpha_s/\Delta T_{air}) \) in the seasonal cycle by taking climatological changes in the Northern Hemisphere from one month to another. They also estimated an observed \( (\Delta \alpha_s/\Delta T_{air}) \) in the seasonal cycle context, taking changes from April to May values from the satellite-based International Satellite Cloud Climatology Project (ISCCP) and T2M from reanalysis data. In 17 climate models used for Fourth Assessment Report (AR4) climate change experiments, the ratio of \( (\Delta \alpha_s/\Delta T_{air}) \) evaluated as changes from present to future were very similar to the changes from April to May in present climate. The models showed a strong variability of the ratio \( (\Delta \alpha_s/\Delta T_{air}) \) with differences up to a factor...
HQ06 also found that the intermodel variation of \( \frac{\partial \alpha_p}{\partial \alpha_s} \) was small because of the convergence of the models in representing the atmosphere’s interaction with upwelling solar photons (Qu and Hall 2006).

The \( \frac{\Delta \alpha_p}{\Delta T_{air}} \) SAF component estimated from the EC-EARTH simulations and the observational estimates from HQ06 are presented in Table 2. CTR had a reduced sensitivity of surface albedo to surface temperature changes—almost 50% less than the estimates of HQ06. Both OPER and ML3 have similar SAF sensitivity values that are closer to the HQ06 estimates. The increased values of OPER and ML3 when compared with CTR were mainly due to an increased change in surface albedo from April to May. This is due to the later spring melting and to a stronger sensitivity of surface albedo to surface temperature changes in both OPER and ML3 when compared with CTR.

e. Upper air

The changes in upper air temperature and specific humidity of OPER and ML3 with respect to CTR are presented in Fig. 14 as the mean annual cycle of time–pressure cross-section average of NH40. Both OPER and ML3 show a cooling of the lower troposphere up to 700 hPa during winter. During spring the cooling extends further in altitude, reaching 300 hPa. Although the surface cooling in OPER and ML3 has their strongest signal during winter, the vertical propagation is constrained by the strong stability of the planetary boundary layer. As for the surface temperature, the cooling in ML3 is stronger than in OPER (cf. Figs. 14a,b).

The air temperature cooling has a direct impact on the reduction of the atmosphere water content at saturation. Both OPER and ML3 show a reduction of specific humidity in the lower troposphere (see Figs. 14c,d). In addition, the near-surface cooling in OPER and ML3 (in comparison with CTR) implies a reduction of moist convection, which in turn would reduce the moistening of the atmospheric column. However, during summer there is an increased water availability in the lower troposphere from CTR to OPER and ML3—higher in the latter. This is coherent with the increased evapotranspiration (see Fig. 8c) that ultimately led to an increase of precipitation. The changes in upper air temperature and specific humidity in ML3...
are stronger, both in spatial distribution and intensity, than in OPER when compared with CTR. This is consistent with the previous analysis to the surface water and energy fluxes, showing that the snow thermal insulation plays an important role in controlling the land–atmosphere coupling.

There is a reduction of the geopotential height in OPER and ML3 as a result of the atmospheric cooling during winter and spring (not shown). The mean sea level pressure presents a positive anomaly (mainly over Siberia; not shown) in OPER and ML3 when compared with CTR that is associated with the cooling of the atmosphere. However, the changes in mass distribution are difficult to analyze because of their reduced statistical significance. Further analysis of changes in local and/or hemispheric circulation patterns would require an increase of the ensemble size to properly disentangle the changes due to the different snow schemes from those linked with atmospheric internal variability, which is out of the scope of the present study.

5. Conclusions

The offline simulations of the three snow schemes at Col de Porte and SNOWMIP2 sites were evaluated as a validation of the ML3 snow scheme. ML3, the multi-layer version of OPER, improved the snowpack simulations mainly over deep snowpack conditions. This was expected, since a bulk snowpack has a large thermal inertia, with difficulties in representing sporadic melting events during winter and early spring. On shallow snowpacks under high vegetation, ML3 had similar results to OPER. These results show that developments associated with the increased complexity of the snowpack
vertical structure should be also associated with a revision of other processes such as subgrid-scale variability, radiative effects of the canopy, and coupling with the atmosphere via the turbulent heat fluxes, among others. However, ML3 has a very simple vertical structure (only three layers with a minimum depth of 0.2 m in each layer). Although sharing a similar construction to other multilayer snow schemes used within climate models (e.g., Boone and Etchevers 2001; Sun et al. 1999) but with a different vertical grid, there are more complex formulations for the vertical structure and physical process of the snowpack available in the literature (e.g., Anderson 1976; Brun et al. 1989). Nevertheless, the implementation and validation of such a complex snow scheme within EC-EARTH is beyond the scope of this study.

EC-EARTH climate simulations of snow cover and surface albedo were compared against remote sensing products. ML3 simulations of snow cover and surface albedo were analogous to OPER. This was foreseen, since both OPER and ML3 share the same parameterizations for snow-cover fraction and snow albedo. The early spring snowmelt in CTR associated with an underestimation of surface albedo was partially reduced with the OPER snow scheme. This underestimation of surface albedo can be due to an early melting (exposing surfaces with lower albedos) and/or an underestimation of snow albedo in the model. The early melting could be also related with an underestimation of snowfall in the coupled model. Dutra et al. (2011a) found that in offline simulations the quality of atmospheric forcing (both total precipitation and the partitioning between liquid and solid phases) is very important for an accurate snow cover, while the model parameterizations were more important for simulating surface albedo. While it was possible to clearly identify the problems in the CdP simulations (overestimation of snow albedo in critical time periods), the general underestimation of surface albedo in the coupled simulations in northern latitudes during spring is associated with a series of interconnected processes. The improved performance of OPER over CTR was also found in offline simulations by DU10. The resemblance between the offline (DU10) and atmosphere-coupled simulations highlight the reliability of the offline strategy to validate land surface models and the importance of the model physics in simulating snow cover.

Over the Himalayan range and Rockies, OPER and ML3 overestimate snow cover and surface albedo during winter and spring (see Figs. 5 and 7). This bias was already present in CTR, but was intensified in OPER, and a similar signal was found in offline simulations with HTESSEL (Dutra et al. 2011a). Over those areas there is a cold bias of T2M (see Fig. 13). Part of the bias can be directly associated with the excess snow cover and surface albedo that is reflecting part of the incoming solar radiation. This explanation is supported by the overestimation of precipitation in those regions during winter (see Fig. 10). However, other atmospheric process such as cloud cover can also be responsible for the cold bias. Furthermore, the subgrid-scale distribution of snow cover over complex terrain is not taken into account in EC-EARTH, with a significant impact on snow-cover evolution and associated energy and moisture fluxes (Liston 2004). These results suggest that the impact of snow schemes over high–complex orography deserves dedicated studies that involve the capacity of resolving fine resolution scales (Bernier et al. 2011). This is also associated with the uncertainty of precipitation simulations in such regions. For example, Minder et al. (2011)
found that the choice of microphysical parameterizations is an important source of uncertainty in simulating mountain snowfall with a mesoscale numerical model.

The increased snow thermal insulation from CTR to OPER and ML3 lead to warmer soil temperatures during winter and spring. This reduced soil freezing, allowing snowmelt and rainfall to infiltrate the soil instead of generating surface runoff. The increased soil moisture content during late spring and summer is then responsible for higher evapotranspiration rates. This, in turn, resulted in an increase of summer precipitation due to the increased moisture in the lower troposphere and an intensification of the soil–precipitation feedback. The sensitivity of surface albedo to surface temperature changes from April to May ($\Delta \alpha_s / \Delta T_{air}$) averaged over the Northern Hemisphere landmasses poleward of 30°N for the three simulations and the observational-based estimates from HQ06. The values of surface albedo were weighted by April incoming radiation prior to averaging. The values between brackets represent the std dev of the mean for CTR, OPER, and ML3 and the error estimate for HQ06.

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<th>((\Delta \alpha_s / \Delta T_{air})) (% K(^{-1}))</th>
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<tbody>
<tr>
<td>CTR</td>
<td>-0.56 (±0.05)</td>
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<tr>
<td>OPER</td>
<td>-0.92 (±0.07)</td>
</tr>
<tr>
<td>ML3</td>
<td>-0.93 (±0.08)</td>
</tr>
<tr>
<td>HQ06</td>
<td>-1.07 (±0.07)</td>
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</table>

Furthermore, the increased snow thermal insulation in OPER and ML3 reduced the T2M warm bias over snow-covered regions during winter and spring. Hazeleger et al. (2010) found similar improvements in coupled climate models on both surface water and energy budgets.
ocean–atmosphere simulations when running EC-EARTH with CTR and OPER snow schemes. Cook et al. (2008) reported similar results when investigating the sensitivity of surface climate to the treatment of snow thermal conductivity. In their experiments, high versus low insulation cases led to soil cooling of up to 20 K in winter and T2M warming of 6 K. Compared with this study, CTR would be the low-insulation case, OPER an intermediate case, and ML3 a high insulation. While Cook et al. (2008) demonstrated the importance of snow thermal conductivity by prescribing its value in a set of experiments, in this study we achieved similar results by using different snow schemes.

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APPENDIX

Multilayer Snow Scheme

The ML3 snow scheme is a multilayer version of OPER for the energy and mass balance within the snowpack. ML3 has the same snow-cover fraction, surface albedo, and snow density parameterizations (see DU10 for further details)—the latter evaluated for each snow layer separately. This multilayer scheme follows closely the formulation of the Interactions between Soil, Biosphere, and Atmosphere–explicit snow (ISBA-ES) snow scheme (Boone and Etchevers 2001) with some adaptations required for HITESSEL, and with a different snow-layering procedure.

The mass balance of the entire snowpack reads as

$$\frac{\partial S}{\partial t} = F + c_{\text{sn}}(F_I - E_{\text{sn}}) - R_{\text{sn},N},$$

(A1)
where $S$ (kg m$^{-2}$) is the total mass in the snowpack (solid and liquid phases), $c_{sn}$ is the gridbox snow-cover fraction, and $F_i$, $F_t$, $E_{sn}$, and $R_{sn,N}$ are the mass fluxes of snowfall, rainfall, snow sublimation, and runoff from the $N$th snow layer ($N$ is the number of snow layers) at the interface with the soil (kg m$^{-2}$ s$^{-1}$).

The liquid water content in each layer is a prognostic, differing from OPER, which uses a diagnostic approach, and the radiation flux follows Boone and following a bucket-type formulation:

$$\frac{dS_{n,j}}{dt} = R_{n,j-1} - R_{n,j} + Q_{n,j}/L_f,$$  \hspace{1cm} (A2)

where $S_{n,j}$ is the liquid water content of the $n$th layer (kg m$^{-2}$), $L_f$ the latent heat of fusion (J kg$^{-1}$), $R_{n,j}$ (kg m$^{-2}$ s$^{-1}$) the flux of water at the bottom of layer $i$, and $Q_{n}$ (W m$^{-2}$) the energy released or absorbed because of water–ice phase changes. The top boundary condition, $R_{sn,0}$, is the rainfall interception ($c_{sn}F_i$).

Runoff is generated when the liquid water exceeds the snow liquid water capacity of the layer following Anderson (1976) (see also DU10 for definition).

The snow layer’s energy balance is given by

$$(\rho C)_{sn,j} D_{sn,j} \frac{dT_{sn,j}}{dt} = G_{sn,j-1} - G_{sn,j} - Q_{sn,j},$$ \hspace{1cm} (A3)

where $(\rho C)_{sn,j}$ is the snow layer volumetric heat capacity (J m$^{-3}$ K$^{-1}$), $D_{sn,j}$ the snow layer depth (m), $T_{sn,j}$ the snow layer temperature (K), and $G_{sn}$ the heat flux (W m$^{-2}$), representing the sum of the heat diffusion and turbulent heat fluxes. The bottom boundary condition, $G_{sn,N}$, is calculated as the basal heat flux in OPER.

The snow layering follows depth coordinates with an increased number of snow layers with increased depth up to a maximum of $N$ layers given by

$$D_{sn,1} = \begin{cases} \frac{D_{sn}}{2}, & D_{sn} < 2D_{sn}^{MIN} \\ \frac{D_{sn}}{2}, & D_{sn} \geq 2D_{sn}^{MIN} \end{cases}$$

$$D_{sn,j}^{MIN} = \begin{cases} 0, & D_{sn} < iD_{sn}^{MIN} \\ \min \left( \frac{D_{sn} - D_{sn,1}}{N-1}, 3^{i-1}D_{sn}^{MIN} \right), & D_{sn} \geq iD_{sn}^{MIN} \end{cases}$$

$$D_{sn,N} = D_{sn} - \sum_{i=1}^{i=N-1} D_{sn,j},$$ \hspace{1cm} (A4)

where $D_{sn}$ is the snowpack depth (m), and $D_{sn}^{MIN}$ is the minimum snow layer depth. In the present study, $N = 3$ and $D_{sn}^{MIN} = 0.2$ m were selected. This configuration was achieved after testing different configurations up to $N = 5$ and $D_{sn}^{MIN} = 0.1$ m in site simulations for CdP and SNOWMIP2. Three snow layers is a common number in the literature for snow schemes within numerical weather prediction or climate models (e.g., Boone and Etchevers 2001; Sun et al. 1999). The minimum snow layer depth was set to 0.2 m primarily to avoid numerical instabilities since both HTESSEL offline and EC-EARTH run with a time step of 1 h, but also because the energy balance in HTESSEL is solved implicitly for the skin temperature and is not explicitly coupled to the snowpack energy balance. With the present configuration there is one snow layer up to $D_{sn} < 0.4$ m, two layers for $0.4 \leq D_{sn} < 0.6$ m, and the three layers are active for $D_{sn} \geq 0.6$ m. The first layer is kept constant at 0.2 for $D_{sn} \geq 0.4$ m and the second layer is limited to a maximum depth of 0.6 m. For 1-m depth the layering is as follows: 0.2, 0.4, and 0.4 m, while for 1.5-m depth: 0.2, 0.6, and 0.7 m.

The adopted time step numerical procedure is the following:

1) skin energy balance is solved to return the residual energy to the snowpack (HTESSEL component);
2) updated snow mass with snowfall and snow sublimation (first snow layer);
3) the energy balance for each layer is evaluated by solving the linearized system of equations simultaneously;
4) water–ice phase changes are evaluated from the updated temperature profile. In the case of melting in the first layer, $T_{sn,1}$ is set to the freezing point, and point 3 is repeated again;
5) phase changes and layer runoff are evaluated to update the profiles of $T_{sn,i}$, $S_i$, and $S_{n,j}$;
6) snow density (for each layer) and surface snow albedo are updated; and
7) reset of vertical grid according to Eq. (A4) for the depth of each layer, conserving mass and energy to obtain new profiles of $T_{sn,i}$, $S_i$, $S_{n,j}$, and $\rho_{sn,i}$. This last point leads, in some degree, to a vertical mixing of properties, especially during snowmelt conditions.

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