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Influence of wet conditions on snow temperature diurnal variations: An East Antarctic sea–ice case study

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ABSTRACT

A one-dimensional snow–sea–ice model is used to simulate the evolution of temperature profiles in dry and wet snow over a diurnal cycle, at locations where associated observations collected during the Sea Ice Physics and Ecosystem eXperiment (SIPEX-II) are available. The model is used at two sites, corresponding to two of the field campaign's sea–ice stations (2 and 6), and under two configurations: dry and wet snow conditions. In the wet snow model setups, liquid water may refreeze internally into the snow. At station 6, this releases latent heat to the snow and results in temperature changes at the base of the snow pack of a magnitude comparing to the model-observation difference (1–2 °C). As the temperature gradient across the snow is in turn weakened, the associated conductive heat flux through snow decreases. At station 2, internal refreezing also occurs but colder air temperatures and the competing process of strengthened heat conduction in snow concurrent to snow densification maintain a steady temperature profile. However, both situations share a common feature and show that the conductive heat flux through the snow may significantly be affected (by 10–20% in our simulations) as a result of the liquid water refreezing in snow, either through thermal conductivity enhancement or direct temperature gradient alteration. This ultimately gives motivation for further investigating the impacts of these processes on the sea–ice mass balance in the framework of global scale model simulations.

1. Introduction

Snow on sea ice is a thermal barrier that strongly curtails the ice–atmosphere heat exchanges. The physical origin for the insulation power of dry snow is primarily its very low thermal conductivity, due to its large air content (e.g., Eicken et al., 1995; Maykut and Untersteiner, 1971). However, the sea–ice snow cover is often wet, which influences the snow thermal conductivity and triggers a variety of heat transport processes well apart from regular conduction. The reasons for the occurrence of wet snow packs on sea ice are manifold. Liquid water in snow may directly come from rain, surface melt or internal melt following absorption of solar radiation in the pack. Freshwater may then percolate down into the snow to potentially refreeze and form superimposed ice (Haas et al., 2001; Jeffries et al., 1997). Although snow thinning processes on Antarctic sea ice are believed to be associated to melting to a much lower proportion than in the Arctic (Nicolaus et al., 2006), and more to evaporation, the occurrence of internal-diurnal-freeze-thaw cycles within snow has been reported in this region (Willmes et al., 2006). Salty water from the ocean may also infiltrate the basal snow layers, by the process of wave overwashing ice floes in the outer pack or the marginal ice zone (Ackley and Sullivan, 1994; Massom et al., 1997, 1998), flooding whenever the snow load is large enough on thin ice to suppress the snow/sea–ice interface below the sea level (Eicken et al., 1994; Sturm, 1998) or even brine capillary wicking through permeable ice (Massom et al., 2001). Water-saturated and salty snow layers or slush may in turn refreeze to form snow ice, as it is very widespread in the Southern Ocean (Jeffries et al., 1998; Maksym and Jeffries, 2000; Massom et al., 2001). Those freezing mechanisms, thus possibly occurring at all vertical levels of the snow, initiate latent heat transfers between liquid and solid phases within the snow, ultimately affecting the vertical snow temperature gradients and hence the associated conductive heat fluxes.

In most large-scale sea–ice models, even those newly including advanced snow representations (Castro-Morales et al., 2014; Lecomte et al., 2015a; Saenz and Arrigo, 2014), these alternative heat transport processes are ignored or their impacts are unevaluated. In Lecomte et al. (2015a), the internal refreezing of freshwater in snow was incorporated but the model showed basically no sensitivity to this process, likely because the temporal

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resolution of the simulations was not high enough to resolve the diurnal cycle of atmospheric surface parameters.

Here, we propose to assess the impacts of the previously described refreezing process on diurnal variations of temperature and conductive heat fluxes in snow. To do so, day-long simulations of a one-dimensional sea–ice model, including the snow internal refreezing scheme of Lecomte et al. (2015a), are performed and intercompared. The analysis of dry snow simulations with respect to observations, collected during the SIPEX-II field experiment, is presented first and gives a baseline for the model evaluation. Snow temperature changes associated with wet snow conditions are then examined before providing some concluding remarks.

2. Model description

2.1. Sea–ice model

In this study, we use the one-dimensional, multi-layer, halo-thermodynamic model of undeformed sea ice proposed in Vancoppenolle et al. (2007, 2010). The energy-conserving formalism of Bitz and Lipscomb (1999) is used for the thermodynamic component of the model and the advection–diffusion equation of brine salinity is used to account for brine convection (gravity drainage) and percolation (flushing) in the model.

2.2. Snow model

Very comprehensive snow models (e.g., SNTHERM, Jordan, 1991), treating all three phases of water, exist and have already been used to study the differences in snow thinning on Arctic and Antarctic sea ice (Nicolaus et al., 2006). However, such models are very sophisticated and for combined reasons of scale (of snow representation), consistency of model complexity between climate components and computational costs, they are not necessarily well adapted for use in large-scale models. It is on the contrary the case of the snow scheme used here (Lecomte et al., 2011), which was specifically designed to represent snow thermo-physics at an intermediate level of complexity and intended to be used large-scale ocean–sea–ice models. By using this kind of snow scheme instead of a more complex one, processes are reproduced with a lower level of accuracy at the local scale, but robust information about their importance and physical impacts to be expected in global models is easily obtainable. For the purpose of the present paper, a few modifications were made to the model and are described in the next section.

The model treats the sea–ice snow cover as a multi-layer, horizontally uniform slab of snow. Each snow layer at depth z is characterized by its temperature $T_s(z)$, density $\rho_s(z)$ and thermal conductivity $k_s(z)$, which may all vary in time. In terms of the snow mass balance, the model accounts for snow accumulation by snowfall, surface and internal melt, and snow losses by snow ice formation when the snow/sea–ice interface is flooded by sea water. Note that in the simulations presented thereafter, the aforementioned processes do not have the time to alter neither the total snow mass nor its stratigraphy substantially, both because the simulations are very short and because melting conditions are never reached. This is consistent with observations (cf. Section 3), showing no melt feature at any of the sampled sites. For this reason, we invite the reader needing further details on those processes or the handling of the density profile in the model to refer to the Lecomte et al. (2011) study. In this 1D vertical model, snow depth changes due to horizontal redistribution by the wind are not treated. This mechanism, while important for snow on sea ice in general, is not critical in this particular case because there was no major snowfall or stormy weather event at the time and location of the simulations. The vertical heat transport through the snow cover, on the other hand, is very active and drives the evolution of snow temperature profiles during the simulations. It is represented in the model using the regular one-dimensional heat diffusion equation:

$$c_s \rho_s \frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z}\left(k_s \frac{\partial T_s}{\partial z}\right) - \frac{\partial I(z)}{\partial z} \tag{1}$$

where $c_s = 2100\,\mathrm{J\,kg^{-1}\,K^{-1}}$ is the snow specific heat and $I(z)$ the amount of solar radiation penetrating at depth z. $I(z)$ is calculated assuming that a fraction of the net solar radiation (i.e., incoming minus reflected shortwave radiation) is absorbed by a surface scattering layer and using an exponentially decaying law for the evolution of penetrating solar radiation with depth (Beer's law). All parameters, namely the thickness of the surface scattering layer, the fraction of net shortwave energy absorbed by this surface layer and the extinction coefficients used in the exponential formula for $I(z)$ are derived from observations. However, this formalism is the one major shortcoming of the present snow scheme. With respect to light propagation, snow is known to be a diffusing- and not an absorbing-medium (Wiscombe and Warren, 1980). Beer's law is thus inappropriate for use in this context. As explained in Section 4, this rather crude radiation scheme in the snow is one of the main sources of disagreement between the model and observations. Solutions to address this issue exist, such as the Delta–Eddington radiation model approach (Briegleb and Light, 2007) or the more recent works of Petrich (2012) and Katlein et al. (2014), but implementing (and evaluating) them is out of the scope of the present study. In the model, the sea–ice/snow albedo may either be prescribed from observations or computed following Shine and Henderson-Sellers (1985). In lack of local observations in the present case, the second option is used.

In the simulations presented hereafter, the model runs with 10 layers in both snow and ice. Studies like Cheng et al. (2008) or Lecomte et al. (2011) have shown that there is a threshold layer number in the snow–ice column below which high frequency temperature changes in snow are better captured with increasing resolution, but above which the model skills in reproducing those changes no longer improve. The highest value for this threshold layer number was 20 (Cheng et al., 2008), for the whole snow–ice column, and 3 snow layers appeared to be sufficient in Lecomte et al. (2011). The choice of 10 snow layers in the present model configuration therefore seems reasonable.

2.3. New implementations

Two modifications were made with respect to the initial Lecomte et al. (2011) model. The first is the use of a new relationship for snow thermal conductivity. In the snow scheme, $k_s(z)$ is parameterized as a function of $\rho_s(z)$ and the formulation of Calonne et al. (2011) was used here. This relationship actually provides snow thermal conductivity values close to those of Yen (1981)'s formula, but was established based on more recent observational datasets.

The second model modification is the incorporation of the internal refreezing of liquid water within the snow, as in Lecomte et al. (2015a). Freshwater, coming from snow melt or rain, may accumulate in a dedicated snow water content model variable. This liquid water content, in m, is a single-bulk-variable independent of the vertical coordinates and has no associated enthalpy. It is assumed to remain at the freezing point at all times, and it may refreeze (in part or entirely) whenever the conductive heat fluxes going in and out of a saturated snow layer are divergent. The refreezing can occur at any level in the snow, since the whole snow pack is assumed to be wet as soon as the snow water content is non-zero. In other words, no specific equation or

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