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INFLUENCE OF SNOW PROCESSES ON SEA ICE: A MODEL STUDY

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INTRODUCTION

Sea ice models have been developed for decades. The earlier models only simulated the one-dimensional vertical thermodynamics and the growth and melt of a uniform slab of ice. Complexity was then progressively brought to those models to represent sea ice more realistically. The dynamics of sea ice were added, as well as rheologies defining the sea ice behavior in response to mechanical constraints, comprehensive halo-thermodynamics and the concept of ice thickness distribution allowing the subgrid-scale representation of ice properties. The representation of snow in large-scale sea ice models, however, has been addressed in much less details and is still relatively basic. Even so, snow has long been known to be a leading player of the polar climate system and it has likewise been hypothesized that including snow mechanisms in sea ice models would turn out to be highly influential, if not beneficial, for current climate simulations. The need for snow-related model improvements, in order to validate this hypothesis, is now a consensus among sea ice modelers. The present thesis primarily aims at answering this need, providing the tools for assessing the importance of snow physics in large-scale sea ice simulations.

1.1 Snow and sea ice

1.1.1 Present-day sea ice state

WMO (1989) defines sea ice as ice ensuing from the freezing of seawater. Although it covers no more than about 4.5% of the Earth's surface on annual average and is hardly thicker than a few meters, sea ice is a crucial component of the polar climate system. Its relatively high albedo (i.e., the spectrally averaged fraction of solar radiation reflected by the sea ice surface) (Perovich et al., 2002), as compared to open water, considerably diminishes the absorption of solar radiation at the ocean surface and contributes to the positive albedo-temperature feedback loop, which amplifies any climate variation in polar regions (Curry et al., 1995). Besides, sea ice acts as a shield between ocean and atmosphere, that reduces the exchanges of heat (due to its low thermal conductivity; Pringle et al., 2007), mass and momentum. Thin sea ice grows faster in winter because its insulating power is accordingly weaker and enables larger conductive heat fluxes through the pack. This negative retroaction is known as the thickness-growth feedback (Bitz and Roe, 2004), that allows the sea ice volume to partly recover from summer losses. In addition to those feedbacks, sea ice affects the World Ocean circulation. As it forms and melts, sea ice releases salt and nearly fresh water, respectively. The outcome is a perturbation of the stratification of the upper ocean and convective fluxes of water that influence the thermohaline circulation (e.g., Goosse and Fichefet, 1999).

Sea ice is characterized by a strong seasonality, as shown in Figure 1.1. Sea ice extent¹ varies in the Arctic from a September minimum of $\sim 6 \times 10^6$ km² to a March maximum of $\sim 15 \times 10^6$ km² (calculated on average over 1979-2012; Chapter 4 of IPCC, 2013). The respective values for the Antarctic are 3 and 18×10^6 km² and are reached in February and September, respectively. Two main types of sea ice can be distinguished: seasonal ice (or first-year ice), that forms and melts entirely every year, and perennial ice (or multi-year ice), that is significantly thicker and survives the summer melt². It was concluded with high confidence in the 2013 report of the Intergovernmental Panel on Climate Change (IPCC) that both the total and perennial annual Arctic sea ice extent have decreased over 1979-2012 in response to the warming climate of

¹Sea ice extent is defined as the total area of model or satellite observation oceanic grid cells that are covered with more than 15% of ice.

²Note that, depending on sources, multi-year ice may also be defined as ice that survived at least two summers (IPCC, 2013), but for the sake of simplification in this thesis, perennial and multi-year ice will further be considered as the same types of ice.

the last half-century³. This suggests a progressive transition of the Arctic sea ice towards an exclusively seasonal cover in a near-to-medium-term future. According to the same source and in contrast, the Antarctic sea ice extent has very likely increased at a rate between 1.2 and 1.8% per decade over 1979-2012. The reasons for this increase are however not clearly established, especially given the geographical variability of the changes in Antarctic sea ice cover (Zwally et al., 2002).

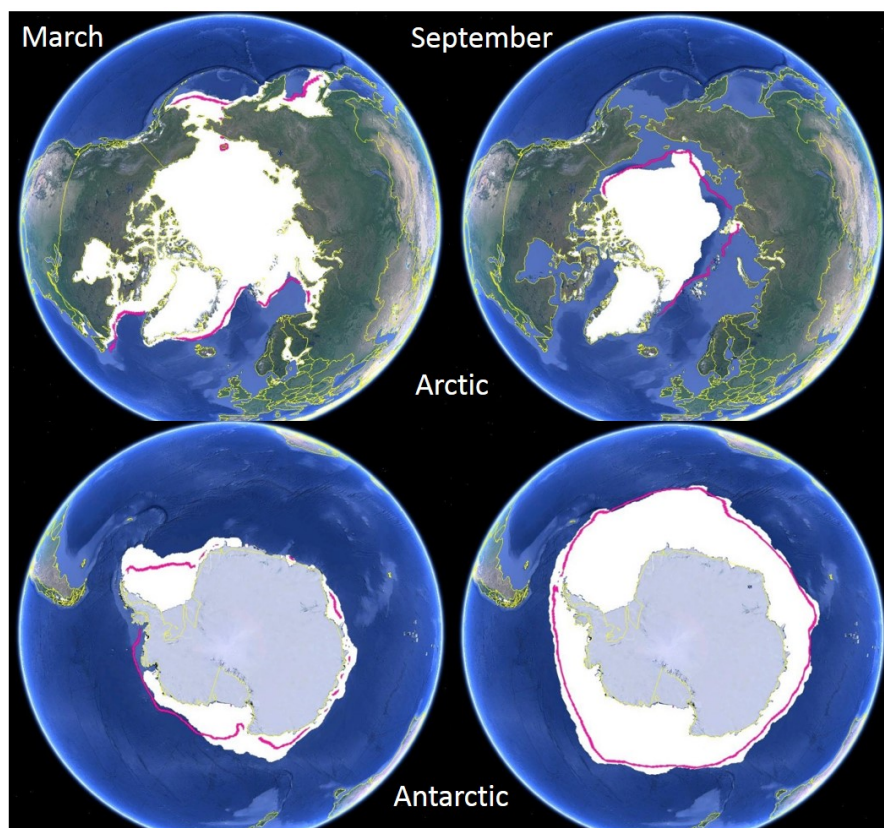


Figure 1.1: Google Earth view of the mean Arctic (top) and Antarctic (bottom) sea ice extents in March (left) and September (right) over 1979-2013, from the data archives of the National Snow and Ice Data Center (NSIDC; www.nsidc.org/data/seaice_index/archives.html). For the record, the pink contours delimit the monthly median sea ice extents over 1981-2010.

³The rate of decline in total annual mean Arctic sea ice extent lies very likely between 3.5 and 4.1% per decade over 1979-2012

1.1.2 Sea ice modeling

The current state of sea ice modeling is comprehensively reviewed in Hunke et al. (2010). The first of the basic components in a complete sea ice model (e.g., Salas y Mélia, 2002; Vancoppenolle et al., 2009; Hunke and Lipscomb, 2010) is the vertical thermodynamics and halodynamics, following for instance Bitz and Lipscomb (1999) and Vancoppenolle et al. (2006) and accounting for the vertical advection of brines in sea ice. The ice momentum equation is then solved by a dynamic component assuming a specific rheology for the sea ice deformational processes. Among those constitutive laws are for instance the non-linear viscous-plastic (VP) rheology of Hibler (1979), the elastic-viscous-plastic (EVP) method of Hunke and Dukowicz (1997) or the more recent anisotropic (Wilchinsky and Feltham, 2004) and elasto-brittle (Girard et al., 2011) rheologies. Lastly, the horizontal advection of sea ice state variables on the model grid is treated using various advection schemes (e.g., Prather, 1986) and in the framework of the so-called ice thickness distribution (ITD) theory (first introduced by Thorndike et al., 1975). Figure 1.2 gives insights of the typical heterogeneity of the sea ice cover. This heterogeneity emphasizes the gap existing between the usual size of a model grid cell (10-100 km) and the characteristic scales of ice thickness variations over an ice floe (1 m-1 km). So as to tackle this problem and reconcile those scales, the ITD theory was developed. The ice in a single grid cell is divided into several categories (usually 5 to 20), for which thickness and concentration (ice area fraction of grid cell) are tracked together with all prognostic variables associated with each category (e.g., snow thickness, snow and ice heat contents, etc.). In practice, this formalism, as opposed to a uniform slab of ice in a grid cell, enables to better resolve the growth and melt of thin ice. The model used specifically in this thesis is better described in the upcoming chapters, and in Vancoppenolle et al. (2009).

1.1.3 Physics of snow on sea ice

1.1.3.1 Major snow properties and processes with respect to the sea ice mass and energy balances

Snow at the sea ice surface accumulates from precipitation. Because sea ice is a peculiar, relatively thin and moving platform, the physics of its snow cover are different from those of land snow covers in various respects.

Radiative properties. Snow behaves almost as perfect black body in the thermal spectrum (longwave radiation, 5-40 μm). However, its reflectance increases in the shortwave spectral range (0.3-2.8 μm) and is especially very high over the visible spectrum (0.4-0.7 μm), which explains its white color. Even

though the snow albedo may be as low as 0.4 for snow containing high concentrations of soot (Warren and Wiscombe, 1980), it is most of the time much higher than the albedo of sea ice (with values usually ranging between 0.80 and 0.90 for clean snow; Wiscombe and Warren, 1980). It is therefore a property of major importance with respect to the sea ice summer decay, since it drastically reduces the surface absorption of solar energy that is either transmitted to the ice or the ocean below, subserving further bottom or lateral melt of sea ice. The albedo-temperature feedback mentioned earlier is thus influenced by the presence of snow on sea ice. Besides, as snow starts melting, the melt-water may accumulate on sea ice to form melt ponds, such as in Figure 1.2 (right picture). Although the occurrence of melt ponds is scarce in the Southern Ocean, it is very widespread on Arctic sea ice (e.g., Fetterer and Untersteiner, 1998; Polashenski et al., 2012). The effect of pond formation is a sharp drop in the surface albedo, since the albedo of water is much lower than the one of snow, or even the one of sea ice. As the snow volume on sea ice prior to the onset of melting is a controlling factor for the development of melt ponds (see Chapter 4), snow is at the origin of very fast and large-magnitude changes in sea ice surface albedo during the melt season.

The albedo and inner light scattering properties of snow mainly depend on the grain size and shape. Grain growth and rounding generally cause the albedo to decrease, because such grains reflect and refract incident light much less efficiently. These grain alterations may occur as a result of many different processes, from slow dry metamorphism due to natural snow ageing to very fast grain rounding due to water saturation in the snow pack, or even wind-driven metamorphism that fragments snowflakes and packs the snow (Sturm and Massom, 2009).

Thermal insulation properties. Due to its very low thermal conductivity, snow acts as a thermal shield insulating sea ice from the atmosphere and subsequently affects sea ice thermodynamics (e.g., Eicken et al., 1995; Fichet and Morales Maqueda, 1997; Sturm et al., 2002). In winter, the latter property curtails the upward conductive heat fluxes from the sea ice bottom (at a temperature close to its freezing point) to the cold snow surface (driven by the air temperature) and therefore limits the bottom growth rate of the ice. Snow thermal conductivity is mainly related to the snow crystalline structure, air content and liquid water content. It is a porous medium possibly including up to three phases: solid ice, liquid water and humid air (e.g., Jordan, 1991). Any variation of the fraction of these phases strongly affects the snow pack dynamics. Three distinct types of heat transfer mechanism through the snow pack can be distinguished. The first is the conduction of heat through the bonds between snow grains and through the interstitial air. The second is the transfer of latent heat by sublimation or condensation of water vapor on

snow grains. Icy particles of the warmer regions of a snow pack sublimate and are advected towards the cold regions by the water vapor flow between the grains to finally re-condensate on colder grains (depth hoar formation). The third is the free or wind-forced advection of humid air through the pack. The last two mechanisms are not conductive ones, but are sometimes implicitly considered in notions of “effective” or “apparent” thermal conductivity (see introduction of Chapter 3).

At macroscopic scales, the snow thermal conductivity is often parameterized as a function of density (see Chapters 2 and 3). Heat conduction through the snow grain connections is all the easier as snow is dense and grains are tightly bound. With regard to the sea ice snow cover, the wind during snow-fall and after deposition is the dominant factor driving the evolution of these properties (Sturm and Massom, 2009). Figure 1.3 illustrates the basic effects of winds on snow, involving both the local redistribution of snow (horizontal displacement of particles) and the formation of wind-packed layers within the snow pack. Depending on the meteorological conditions consecutive to its deposition, snow may undergo further metamorphism. As a result of this, the snow cover is characterized by a pronounced stratigraphy and complex density and thermal conductivity profiles, as demonstrated by Figure 1.4.

Snow depth. Contrary to the snow intrinsic characteristics presented previously, depth is a macroscale property of a snow pack. It is an influential factor for sea ice physics in various respects. For a given snow thermal conductivity, the deeper the snow pack, the more insulating, which means that the sea ice growth in winter is constrained by both the type and amount of accumulating snow. In summer, on the other hand, snow delays the onset of sea ice surface melts as it has to melt away first. The time-lag between snow and sea ice melt is thus naturally depending on snow depth. Additionally, snow may directly contribute to sea ice surface production via the formation of snow ice (Jeffries et al., 1998; Fichfet and Maqueda, 1999; Maksym and Jeffries, 2000). Snow may get water-soaked as a result of several processes. If the hydrostatic balance of the snow/ice column or ice deformation constraints are such that the snow-ice interface is depressed below the waterline, seawater may flood into snow. Lateral infiltration of seawater by waves breaking against ice floes, or vertical percolation of rain or meltwater may also contribute to this mechanism. The subsequent refreezing of the snow-seawater mix (often called slush) is called snow ice. Antarctic sea ice, mostly thin and seasonal, is particularly prone to sustain this mechanism (Massom et al., 2001). Last but not least, the horizontal distribution of snow thicknesses modifies the sea ice surface topography and its surface roughness with respect to the wind forcing. Hence, sea ice dynamics in response to winds is altered by the presence of snow.

If precipitation rates define the potential for snow accumulation on sea ice, the actual accumulation is mostly controlled by winds and the nature of the sea ice on which it falls (Sturm and Massom, 2009). Snow may accumulate only if the platform that sea ice constitutes is formed, which implies reduced accumulation on seasonal ice, as opposed to perennial ice. Whether the ice is level or deformed by dynamic processes is critical too. During the mechanism of snow-redistribution by the wind mentioned earlier, snow tends to be easily removed from flat surfaces, while it re-accumulates under the lee of pressure ridges (topographic features of sea ice that form when ice floes crush against each other during their convergent motion). This is clearly shown by the two pictures of Figure 1.2. Lastly, snow ice formation by definition is also a process directly inducing snow thinning, by transforming the lowermost snow layers into slush or solid ice.

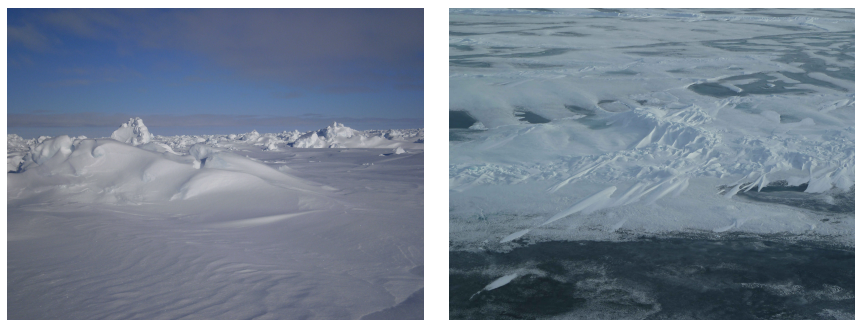


Figure 1.2: Pictures of snow-covered Antarctic (left; taken during the Sea Ice Physics and Ecosystem eXperiment –SIPEX-II field cruise, in East-Antarctic sea ice, Sept.-Nov. 2012) and Arctic (right; credit: Antoine Barthélemy, Université catholique de Louvain, during the Nansen and Amundsen Basins Observational System –NABOS cruise 2013) sea ice. Both pictures show the presence of relatively thin snow on flat sea ice areas, while the wind re-accumulated it around pressure ridges. On the left picture (Antarctic), heavily deformed sea ice with very deep patches of snow under the lee of ridges is clearly seen in the background. In the back of the Arctic picture, dark blue areas are refrozen melt ponds partially recovered with snow. Typical snow drift features are noticeable on the main pressure ridges of this picture.

1.1.3.2 The complex relationship between snow and sea ice : hemispheric differences and feedbacks

The previous section summarizes the most important snow-related processes on sea ice, at the scale of an ice floe. Not all those processes occur everywhere and on any type of sea ice. As explained, snow ice formation is more common in the Southern Ocean than in the Arctic Ocean due to the hemispheric differences in ice thickness distribution and snowfall. Furthermore, melt ponds,

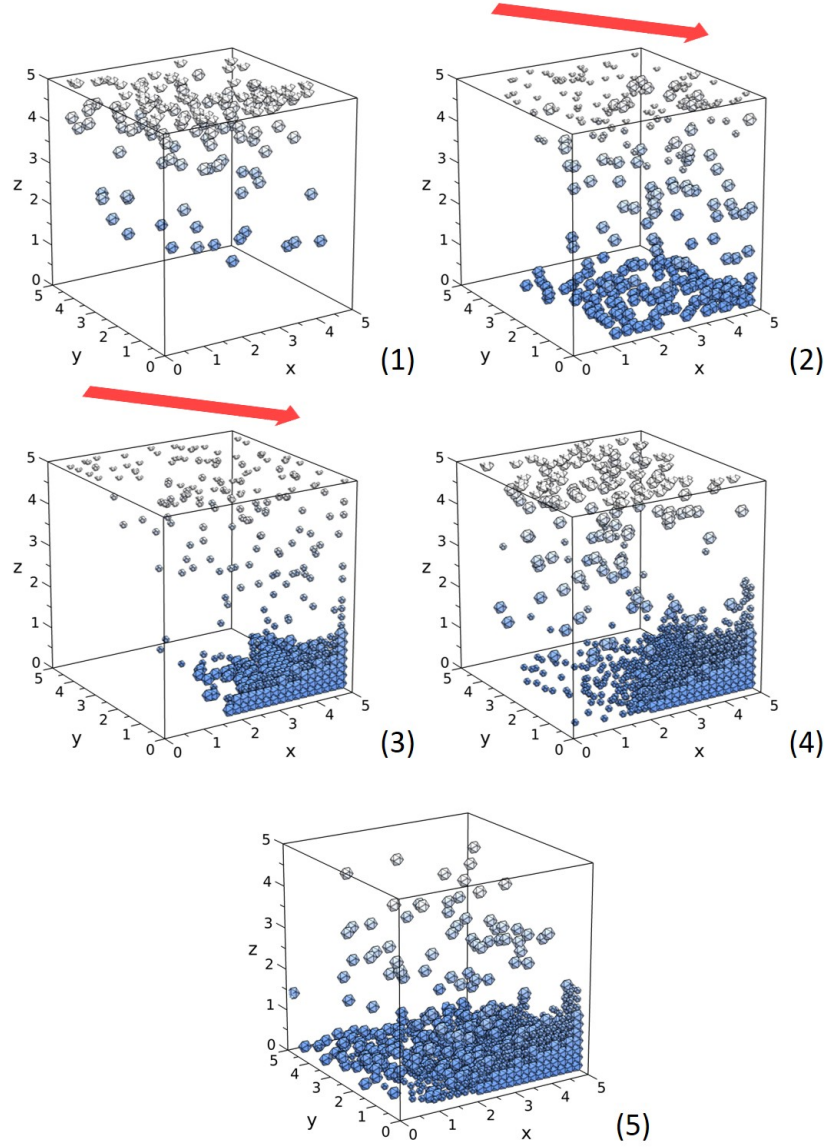


Figure 1.3: Illustration of the formation of dense wind-packed layers in a snow pack using a toy model developed during this thesis. At the initial stage (1), snow falls without wind. As the wind blows (represented by the red arrows in stages 2 and 3), snow particles (either snowflakes from precipitation or older grains formerly deposited and redistributed by the wind) are displaced horizontally until they are stopped by an obstacle. In addition, they are broken and fragmented into smaller grains, leading to denser layers when they are redeposited. In stages (4) and (5) accumulation restarts without wind, on top of the wind-crust formed during the wind gust. Within the snow pack, snow layers thus store the history of the previous blowing snow events.

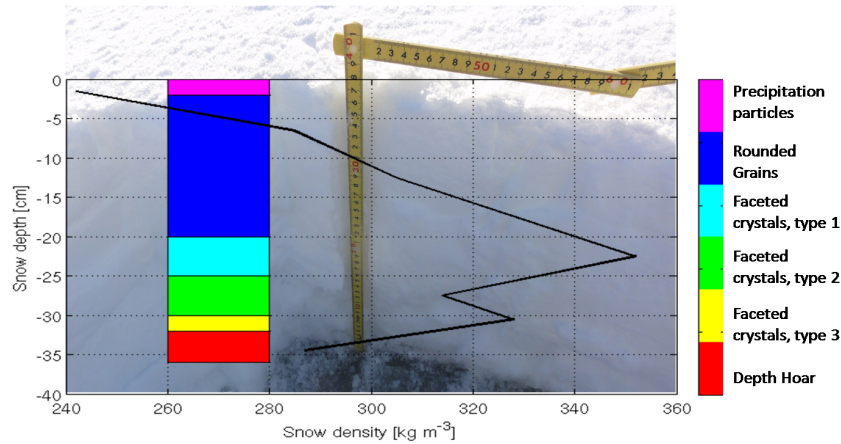


Figure 1.4: Snow pit taken during the Sea Ice Physics and Ecosystem eXperiment –SIPEX-II field cruise, in East-Antarctic sea ice, Sept.-Nov. 2012. The overlying graph shows the density profile and the various types of snow that were sampled in this snow pit.

that cover up to 50-60% of the Arctic sea ice area in summer, are virtually unseen on Antarctic sea ice. Although the reason for this feature is not clearly established, it seems related to atmospheric conditions that curtail snow surface melt in those regions (Andreas and Ackley, 1982). In any case, this means that the interactions between snow and sea ice strongly depend on external components: the ocean and the atmosphere. This statement, seemingly trivial, actually raises important issues in terms of modeling. It implies that a prerequisite for simulating those interactions properly is to have an a priori good representation of the ocean and atmosphere in both hemispheres in climate models, which is not necessarily achieved already and is bound to be a limitation. Second, the physical mechanisms described above sometimes have competing consequences. For instance, deeper snow may lead, on one hand, to decreased bottom ice congelation due to stronger insulating effect, but also to a larger production of snow ice (through increased flooding), in turn favoring the surface production of ice to the cost of snow. Such feedbacks are manifold and their large-scale implications are yet to be determined.

1.1.3.3 Observations of snow on sea ice

There is a number of sources of snow observations on sea ice. Here, we give some examples (some of them are actually used in this study), depending on the scale at which snow is observed, and present the major issues of those observations with regard to their use for large-scale model evaluation purposes.

Local - floe-scale in situ observations Many floe-scale snow observation datasets are available, both on Arctic and Antarctic sea ice. In the Arctic for instance, such observations are provided in the Sea Ice Mass Balance Probe datasets on landfast sea ice, collected within the SIZONET (Seasonal Ice Zone Observing Network) project ([url: http://seaice.alaska.edu/gi/data/barrow_massbalance](http://seaice.alaska.edu/gi/data/barrow_massbalance)) and include long time series of snow depth and temperature profiles in snow. On drifting ice, this kind of data is acquired using Ice Mass Balance (IMB) buoys such as those managed by the Cold Regions Research and Engineering Laboratory (CRREL) (Perovich et al., 2009). Yet another example is the data obtained during sea ice science cruises, such as SHEBA (Surface Heat Budget of the Arctic Ocean, Sturm et al., 2002). Those cruises provide generally shorter time series because they are acquired during temporary ice stations (icebreaker parked into the ice for a given period) but enable to perform broader measurements since scientists actually go on the ice to observe all kinds of snow structural properties via snow pit measurements, or to set up masts heavily equipped with optical sensors, measuring snow radiative properties. Such field campaigns have also been conducted in the Southern Ocean with, for instance, the ISPOL (Ice Station POLarstern) drift (e.g., Hellmer et al., 2008; Nicolaus et al., 2009), or the Sea Ice Physics and Ecosystem eXperiment 2007 ([url: http://www.sipex.aq/](http://www.sipex.aq/)) and 2012 ([url: http://www.acecrc.org.au/Research/SIPEX%202012](http://www.acecrc.org.au/Research/SIPEX%202012)). This kind of data is usually very reliable and, because meteorological variables are most of the time recorded simultaneously, allows to run and validate one-dimensional snow/sea ice models under the exact environmental conditions prevailing at the time observations were made. However, they are far more difficult to use with regard to large-scale models since, by nature, they are localized and not necessarily representative of snow at the hemispheric- or let alone global-scale. They probably could be used for such purposes if compiled and standardized into a single database covering basin-wide areas. Even so, since the measurements available in each dataset are performed with different methods, with distinct uncertainties, achieving the latter task would be no small endeavour.

Semi-global observations Since relatively recently, regional-scale snow observations are available. Examples of such data are the Operation IceBridge airborne surveys of snow depth in the Arctic (Kwok et al., 2011; Kurtz and Farrell, 2011) and the ASPeCt (Antarctic Sea Ice Processes and Climate) snow depth product (Worby et al., 2008a). Those product types have larger uncertainties and are not continuous in time but still provide data of good quality, which are very useful for estimating the major regional biases in snow depth simulated by models. Their drawback, is that their temporal and spatial cov-

erage is not really adequate for either local-scale snow process parameterizations or global-scale comprehensive evaluation of sea ice models.

Global - satellite observations Satellite observations of snow on sea ice have also been made available during the past two decades (e.g., Markus and Cavalieri, 1998; Comiso et al., 2003; Kwok and Cunningham, 2008; Worby et al., 2008b). Technically, these data are ideal for use in support to large-scale modeling of the sea ice. They are temporally nearly-continuous since 1979 and, with the ongoing fast progress in remote sensing sciences, their spatial resolution keeps increasing with years. Such data could be used for both model evaluation and improvement through the use of data assimilation techniques (Massonnet et al., 2014) if only they did not suffer from very large uncertainties. Indeed, the remote sensing of snow depth, snow water equivalent and other key snow properties is still problematic and is one of the reasons why the large-scale modeling of snow on sea ice is so challenging (Sturm and Massom, 2009).

1.2 Modeling snow on sea ice

Pre-existing land snow models. Very complex snow models have been long used on land (e.g., Brun et al., 1989; Jordan, 1991; Bartelt and Lehning, 2002; Punge et al., 2012), with applications going from the simulation of ice sheet snow covers under past and present-day climates to the operational forecasting of avalanches in Alpine environment. Those models are comprehensive and include very detailed representations of both the grain-scale snow properties (size, shape of snowflakes) and all metamorphism processes affecting them. Most frequently, those models are coupled to atmospheric circulation models, allowing the explicit simulation of the snow local transport by winds.

The challenge of modeling snow on sea ice : horizontal advection. The aforementioned land snow models were an inspiration for this work (in particular, the Soil Ice Snow Vegetation Atmosphere Transfer –SISVAT model; De Ridder and Gallée, 1998) but they present characteristics that are incompatible with current sea ice models. The scale is the first of these characteristics. Although this is bound to change in the future, most coupled ocean–sea ice models have so far been used in global configurations and at relatively low horizontal resolution ($1/4$ to 2-degree longitude-latitude). Therefore, they are not adapted for representing snow in every details and with high accuracy. More importantly, sea ice is also a moving platform, that requires advection schemes in models in order to track the horizontal transport of its state variables. Usually, those schemes are computationally expensive and including comprehensive snow schemes in such models would necessitate advecting a great number of physical variables for snow only, which cannot be afforded.

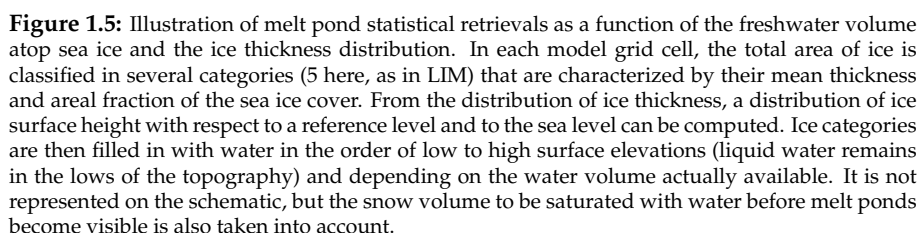
Going further, consistency must be kept between climate model components. Incorporating snow components as detailed as in land snow models into sea ice models that are not hardly as complex makes little sense. Finally, the physical processes of first importance in driving the land and sea ice snow covers at macroscopic scales are different. For instance, the density profiles of Alpine or glaciers snow covers are driven in large part by gravitatal settling, while this process is practically unseen on sea ice because the snow pack is too thin. Hence, although this is a possible solution, there seems to be no obvious reason to use those models as a starting point for improving the snow representation in sea ice models as a priority.

In part for those reasons, the representation of snow in large-scale sea ice models prior to the present thesis had been kept relatively simple. Snow was represented as a single layer, with constant density and thermal conductivity often equal to 330 kg m^{-3} and $0.31 \text{ W m}^{-1} \text{ K}^{-1}$, respectively. Efforts had already been made however to include snow ice formation in those models, following for instance the formalism of Fichefet and Morales Maqueda (1997), and the effect snow on the sea ice albedo. In LIM (Louvain-la-Neuve Sea Ice Model, version 3; Vancoppenolle et al., 2009) for instance, snow depth and distinct albedo values for dry and melting snow were taken into account in sea ice albedo calculations following Shine and Henderson-Sellers (1985). In CICE (Los Alamos Sea Ice Model, version 4; Hunke and Lipscomb, 2010), the more complex radiative scheme of Briegleb and Light (2007) accounts for the presence of snow on sea ice in albedo calculation as well. Yet another example is GELATO (sea ice model from the Centre National de Recherches Météorologiques, Toulouse, France; Salas y Mélia, 2002), in which a parameterization of the impact of snow ageing on the sea ice albedo had also been developed prior to this work. Notwithstanding, the other physical processes described in the previous section were still lacking in those models.

1.3 Identifying the most relevant snow processes with regard to the large-scale modeling of sea ice

1.3.1 Objectives of the thesis

Before and during the investigations of this thesis, several studies have shown the potentially large impacts that snow-related processes may have in global sea ice simulations (e.g., Eicken et al., 1995; Fichefet et al., 2000; Blazey, 2012). Therefore, at the starting point of this work stands the hypothesis that more advanced snow representations (as compared to former ones) in sea ice models are required to improve climate model simulations. **The first objective is thus to validate this hypothesis.** The second is to try to **give guidance to cli-**



1.3.2 Achievements

Given the stratified nature of the snow cover and the large number of snow layers included in existing land snow models, it seemed legitimate to extend the former representation of snow in LIM to a multi-layer structure. Even so, more layers to better define the vertical profiles of snow properties numeri-

cally means a greater number of variables to advect horizontally with sea ice, and thus considerably longer computation time. The issue of the minimal vertical resolution to be kept in a new one-dimensional snow scheme designed to be used in large-scale sea ice models is addressed in Chapter 2. Figure 2.3 and Table 2.1 of this chapter show how increasing the number of layers for the treatment of snow thermodynamics improves the simulation of temperature profiles in snow along the sea ice growth season. Despite this, threshold layer numbers above which the improvements in snow temperature profiles were no longer significant were found and provided the justification for using 3 layers (refined to 6 in thermodynamics only) in the snow scheme of the full three-dimensional version of LIM. The study also emphasizes the necessity of having both the temperature and thermal conductivity profiles in snow well represented in order to simulate the right conductive heat fluxes driving the wintertime ice growth rates. Several formulations of snow density and thermal conductivity are thus tested in Chapter 3, in the complete coupled ocean–sea ice model framework. In this chapter, the snow scheme previously developed was adapted with respect to the horizontal advection of snow variables. Several conclusions can be drawn from this analysis. First and foremost, the modeled Arctic mean sea ice thickness exhibits a large sensitivity to snow thermo-physical parameters, as demonstrated by Table 3.2 and Figure 3.4. The much weaker corresponding sensitivity in the Southern Ocean and the distinct sea ice mass balance in each hemisphere (see Figure 3.8, Chapter 3) then underline the different responses of the modeled sea ice state to changes in snow thermo-physical description. Hence, none of the usual parameterizations of snow thermal conductivity as function of density gives realistic results in both hemispheres simultaneously. In an attempt to solve this problem, a simple wind-packing parameterization linking the bulk snow density and thermal conductivity to the mean seasonal wind speed was developed and tested in the model. This parameterization, directly impacting on snow depth (see Figure 3.5, Chapter 3), leads to particularly good Arctic snow depth distributions and to a compromise in the model ability to reproduce both Arctic and Antarctic sea ice extents and thicknesses simultaneously.

Consequently, it seemed natural to implement a process that would further affect snow depth distributions on sea ice : blowing snow. As the melt pond model from Flocco and Feltham (2007) was incorporated in LIM concurrently to the latter process, the dual evaluation of the impacts of melt ponds and blowing snow on Arctic sea ice was made possible, and is presented in Chapter 4. The melt pond scheme uses the ice thickness distribution of the sea ice model to compute the fractional area and depth of the melt pond cover, as schematically illustrated by Figure 1.5. In addition, the latter coupling provided the opportunity to use a contribution to an other melt pond modeling

study (Hunke et al., 2013) to account for the infiltration of liquid water into snow in LIM. The final analysis of this chapter, based on Figures 4.12, 4.15 and 4.16, allows the identification of two separate melt pond regimes. On the simulated first-year ice, melt ponds tend to be shallow. In this situation, a deeper snow cover prior to the onset of summer surface melt likely results in larger pond fractions, since more freshwater coming from snow melt is available to make them grow, but the effect is modest. Indeed, owing to both its seasonal nature and its topography, snow and water accumulation on first-year ice are limited anyway. Conversely, ponds on multi-year ice may get deeper but cover a smaller fraction of the sea ice. In that case, persistent snow at the beginning of the melt season may hide forming melt ponds and impede their development, hence an opposite causal relationship is found: an increase in spring snow depths leads to a substantially reduced summer melt pond cover.

Those two typical model behaviors on first-year and multi-year ice ultimately raise the question of the uneven sensitivity of those two ice types in coupled ocean–sea ice models to the physical surface processes related to snow in a broader way. A short inter-comparison of simulations performed in the same conditions but changing various snow parameterizations in the model is done in Chapter 5 in order to tackle this problem. In this last chapter, Figures 5.1, 5.2 and Table 5.2 clearly show the weaker sensitivity of the simulated first-year ice (as compared to multi-year ice) to changes in snow representation or snowfall. In the context of global warming, it is therefore profitable to adequately represent snow in climate models in order to simulate the timing of perennial ice disappearance. Concurrently, the simulation of first-year ice covers, such as in the Southern Ocean or for future projections of the Arctic sea ice, will likely require to improve the oceanic and atmospheric components of those models, since they drive the essential part of the seasonal ice variability.

CHAPTER 2

A NEW SNOW THERMODYNAMIC SCHEME FOR LARGE-SCALE SEA-ICE MODELS

This Chapter is based on the following paper:

Lecomte, O., Fichefet, T., Vancoppenolle, M., Nicolaus, M., 2011. A new snow thermodynamic scheme for large-scale sea-ice models. *Annals of Glaciology* 52 (57), 337–346.

Abstract

This study assesses the skills of a new one-dimensional snow scheme developed for the thermodynamic component of the Louvain-la-Neuve sea-Ice Model (LIM). The model is evaluated at Point Barrow (Alaska) and at the Ice Station POLarstern (ISPOL) in the western Weddell Sea (Southern Ocean). The new snow thermodynamic scheme leads to better snow internal temperature profiles, with a setup-dependent increase in the correlation between simulated and observed temperature profiles. In average over all runs, these correlations are 27% better with the 6-layer configuration. The model's ability to reproduce observed temperatures improves with the number of snow layers, but stabilizes after a threshold layer number is reached. The lowest and highest values for this threshold are 3 (at

Point Barrow) and 6 (ISPOL), respectively. Overall, the improvement of the model skills in simulating the sea-ice thickness is not as significant as for snow temperature, which is probably due to the rather crude representation of the snow stratigraphy in the model.

2.1 Introduction

The Louvain-la-Neuve sea Ice Model (LIM) is a three-dimensional global model for sea-ice dynamics and thermodynamics that has been specifically designed for climate studies. Its latest version LIM3 (Vancoppenolle et al., 2009) is fully coupled with the oceanic general circulation model (GCM) OPA (Ocean PARallélisé) on the modeling platform NEMO (Nucleus for European Modelling of the Ocean). As in most coupled ice-ocean GCMs so far, the representation of snow is relatively simple in this model: it includes one snow layer with constant physical properties; snow depth increases when snow falls and decreases by snow-ice formation, sublimation and surface melt.

Yet snow plays a key role in the sea-ice physics and influences sea-ice heat/mass balance. First, the snow cover strongly curtails the heat exchanges between the ice and the atmosphere. Although snow nearly behaves like a black body in the longwave spectrum (Dozier and Warren, 1982; Warren, 1982), it reflects most shortwave radiation (e.g., Wiscombe and Warren, 1980; Perovich, 2001). A few cm thick snow layer can greatly lessen the input of solar radiation into the sea ice (e.g., Weller, 1968; Allison et al., 1993). In addition, due to its very low heat conductivity, snow has strong insulating abilities and absorbs large amounts of the surface temperature variability in its uppermost centimeters, meanwhile protecting the ice surface from sharp atmospheric state variations and limiting the thermodynamic ice growth (e.g., Maykut and Untersteiner, 1971; Maykut, 1986; Eicken et al., 1995; Fichefet et al., 2000). Finally, snow directly contributes to the sea-ice mass balance by both snow-ice formation (e.g., Eicken et al., 1994; Jeffries et al., 1997; Fichefet and Maqueda, 1999), when seawater infiltrates the snow/ice interface (through waves, negative freeboard, or permeability through ice), and superimposed ice formation (e.g., Jeffries et al., 1997; Kawamura et al., 1997; Haas et al., 2001) resulting from re-freezing of snow melt water and water vapor.

In view of those properties, it is of major importance to have a good representation of the snow physics in coupled ice-ocean models. This can be done only by having a good parameterisation of the heat fluxes through snow and ice, which themselves depend on snow physical properties. However, one constraint is that coupled ice-ocean models are computationally expensive. The challenge is therefore to improve snow models so as to get a better representation of the snow influence on sea ice while keeping rather simple parameterisations and reasonable computational costs in global-scale simulations.

Some studies involving one-dimensional modeling of sea ice using sub-models for snow have already been performed, like Huwald et al. (2005), Shirasawa et al. (2005), or Cheng et al. (2008). The snow module used in the

first study has constant thermal properties for snow. The last two ones are more similar to what is presented in this paper, except that the snow modules are slightly more advanced in terms of processes (e.g., they include refreezing of meltwater or gravitational settling of snow). However, our approach is different. We aim to find and evaluate the most simple and satisfying thermodynamic snow scheme to be easily incorporated into a large-scale sea-ice model with sea-ice thickness categories, and especially into the NEMO-LIM3 GCM. Physically, the model does not have to describe snow properties to last detail and accuracy but must represent well its large-scale features.

2.2 Model description

2.2.1 Sea ice model

The sea-ice model used here is a multi-layer halo-thermodynamic model of undeformed sea ice (Vancoppenolle et al., 2007, 2010). The thermodynamic component is based on the energy-conserving model of Bitz and Lipscomb (1999). Brine dynamics are represented using an advection-diffusion equation for brine salinity which represents brine convection (gravity drainage) and percolation (flushing).

2.2.2 Snow module

Given the high vertical heterogeneity of the snow cover above sea ice (Nicolaus et al., 2009), it is necessary to represent different types of snow depending on their characteristics. As a result, a multi-layer approach has been chosen for the scheme. We consider a horizontally uniform pack of snow on sea ice, with a thickness h_s . At each depth z within the snow, the thermodynamic state of the medium is characterized by temperature $T(z)$, density $\rho_s(z)$ and effective thermal conductivity $k_s(z)$. The horizontal variability of snow on sea ice (e.g., Massom et al., 2001; Sturm et al., 2002) will be treated later, in the global version of LIM. Vertical heat diffusion, surface and internal melt, snowfall and snow-ice formation are all included in the scheme. The details of the scheme characteristics are summarized in Figure 2.1.

2.2.3 Heat transport through the snow cover

The vertical snow temperature profile is governed by the one-dimensional heat diffusion equation:

$$c_s \frac{\partial(\rho_s T)}{\partial t} = \frac{\partial}{\partial z} \left(k_s \frac{\partial T}{\partial z} \right) - \frac{\partial I(z)}{\partial z} \quad (2.1)$$

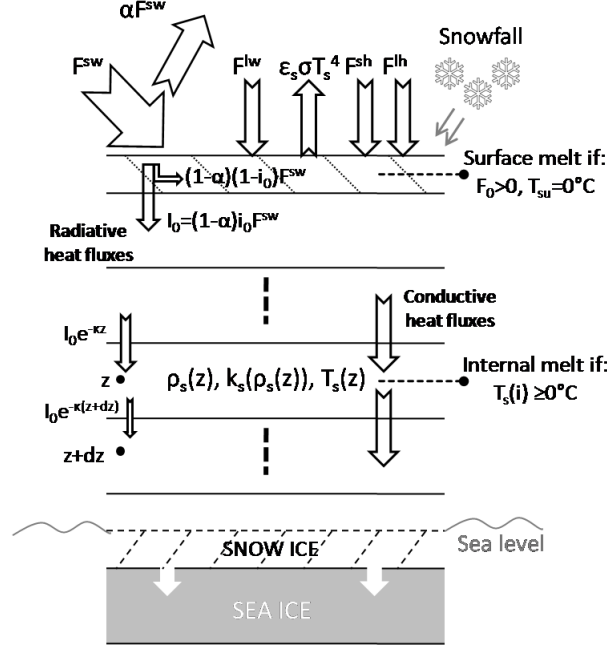


Figure 2.1: Schematic of LIM1D's new snow module.

where $c_s = 2100 \text{ J kg}^{-1} \text{ K}^{-1}$, ρ_s and k_s are the specific heat, density and thermal conductivity of snow, respectively.

$$I(z) = I_0 e^{-\kappa_s z} \quad (2.2)$$

is the solar radiation penetrating into snow at level z , following Beer's law. $\kappa_s = 16 \text{ m}^{-1}$ is the extinction coefficient and I_0 is the solar radiation penetrating under the snow surface:

$$I_0 = (1 - \alpha) i_0 F^{sw} \quad (2.3)$$

where α and $i_0 = 0.18$ are the albedo and fraction of solar radiation that penetrates the upper snow surface, respectively, and F^{sw} the incident shortwave radiation at the snow surface. For the sake of simplicity, we chose values of Grenfell and Maykut (1977) for i_0 and κ_s , but this solution for the computation of radiation transmission through snow may not be the best and work is currently done to improve it. This formulation is identical to the one for radiation transmission in sea ice, but due to a much bigger attenuation coefficient, snow attenuates much more solar radiation than ice. Similar methods are used in

the snow models of Gallée and Duynkerke (1997) and Loth et al. (1993), except that the extinction coefficient depends on snow grain properties.

Thermal conductivity of snow k_s is parameterised as a function of ρ_s and the three following parameterisations are tested in the model:

$$k_s = \begin{cases} k_i (\frac{\rho_s}{\rho_w})^{1.88} & \text{(a)} \\ k_i(T) (\frac{\rho_s}{\rho_w})^{1.88} & \text{(b)} \\ 0.138 - 0.00101\rho_s + 0.000003233\rho_s^2 & \text{(c)} \end{cases} \quad (2.4)$$

with ρ_w the liquid water density in kg m^{-3} , and k_s is in $\text{W K}^{-1} \text{m}^{-1}$. Formulation (4a) follows Yen (1981). (4b) is similar except that the ice thermal conductivity k_i is expressed as a function of temperature following Pringle et al. (2007). (4c) corresponds to the relationship of Sturm et al. (1997).

The surface energy balance provides a boundary condition at the top of the snow cover (fluxes are defined positive downwards):

$$F_0^{ct} = F_0 = (1 - \alpha)(1 - I_0)F^{sw} + F^{lw} - \epsilon_s \sigma T_s^4 + F^{sh} + F^{lh} \quad (2.5)$$

F_0^{ct} being the conductive heat flux in snow under the surface, F_0 the net energy flux from the atmosphere to the snow, F^{lw} the downward longwave radiation, $\epsilon_s = 0.97$ and T_s the surface emissivity and temperature, respectively, and F^{sh} and F^{lh} the turbulent fluxes of sensible and latent heat. The albedo is either a function of surface state, cloud cover, ice thickness and snow depth (Shine and Henderson-Sellers, 1985, see Appendix A), or taken from observations. The scheme also includes a parameterisation to account for the effect of a one-dimensional melt pond when snow has melted away and water is remaining on sea ice (Zuo and Oerlemans, 1996):

$$\alpha = \alpha_w - (\alpha_w - \alpha_i) e^{-\frac{p_{dpth}}{\omega}} \quad (2.6)$$

where α_i and $\alpha_w = 0.30$ are the ice albedo and the albedo of water in the pond, p_{dpth} is the pond depth in m and ω is a constant scale factor in the same units as p_{dpth} . In this particular case, α_i is taken to be constant and equal to 0.5 (Zuo and Oerlemans, 1996). The source for the water in the pond is nothing but the accumulated snow/ice melt.

The heat diffusion equation is solved in the snow - sea-ice system using 10 layers of sea ice and from 1 to 6 snow layers in this study. Once the new temperatures are computed, internal melt of snow may occur if the snow temperature reaches the melting point. The imbalance between external and inner heat fluxes at the interfaces are used to compute growth/melt of sea ice and surface melt of snow.

2.2.4 Sources and sinks of snow mass

For the snow density evolution, we did not choose to implement specific processes such as gravitational settling, wet snow metamorphism, destructive or constructive metamorphism (for a review of these processes, see Sommerfeld and LaChapelle, 1970; Colbeck, 1982; Armstrong and Brun, 2008) because we do not dispose of an accurate enough description of snow grain properties over sea ice. Indeed, these processes affect snow grain size, shape and constitution, and are very difficult to simulate without considering the small-scale features of a snow pack. As they indirectly affect snow density, we do not compute their potential impact on the snow characteristics of the model, but directly aim to get the best possible representation of snow density distribution based on observations. Therefore, the vertical density profile is initialized at the start of a run (with snow density profile data, if available) and is non-uniformly updated in a mass- and energy-conserving way at the end of each time step. More precisely, during each time step, new snow may come as snowfall with a density parameterised as a function of the surface wind speed. This is based on the assumption that the snow cover characteristics on sea ice mainly depend on the total amount of snowfall, the accumulation rate and the wind speed u at time of precipitation (Sturm and Massom, 2009). Indeed, when snow falls, the wind breaks and fragments snowflakes. The properties of the new snow layer are then settled in a few hours by means of metamorphism. We use the following formula for the density of fallen snow:

$$\rho_s = \max(150, 20 * u) \text{ (in kg m}^{-3}\text{)} \quad (2.7)$$

inferring an increase in density of 20 kg m^{-3} for each m s^{-1} increase in wind speed (Jordan et al., 1999). Densities are thresholded to avoid too small values for snow over sea ice. After new snow has fallen, the mass balance is computed:

$$\frac{\partial M_s}{\partial t} = R_S - R_M - R_{SI} \quad (2.8)$$

where R_S is the snowfall rate, R_M the melt rate, and R_{SI} the snow-ice formation rate. R_M accounts for both surface and internal melts. Snow mass per unit area M_s is the integral of snow thickness times density over all layers. The snow grid is then updated so as to keep a constant number of layers in the scheme, while minimizing numerical diffusion of density by merging adjacent layers with smallest density differences. If density differences are larger than 50 kg m^{-3} , the new layer from precipitation is merged with the former top layer and the rest of the snow pack is left untouched. Snow density thus evolves naturally through this remapping and the accumulation of snow due to precipitation.

2.2.4.1 Snowfall

R_s is given in input to the model as snow water equivalent. It is converted into snow depth with respect to its density, computed as described above. Snowfall data are taken from large-scale reanalyses, since no observational data are available. The scheme does not include liquid precipitation.

2.2.4.2 Melt

Surface and internal snow melt calculations are both based on enthalpy. Volumetric specific enthalpy of snow is expressed by :

$$q_s = \rho_s(c_s(T_0 - T) + L_0) \text{ (in } J m^{-3}) \quad (2.9)$$

in which the first term on the right hand side is the so called snow "cold content". T_0 is the melting point and L_0 is the latent heat of freezing ($334 \times 10^3 J kg^{-1}$). The enthalpy by surface unit of a layer of thickness h_s is then:

$$Q_s = q_s h_s \quad (2.10)$$

Internal melting may occur after diffusion, if the temperature of a layer comes to exceed the melting point temperature, which is made possible by the internal absorption of shortwave radiation by snow. In this instance, the layer temperature T is brought back to T_0 and h_s is reduced to ensure heat conservation. Otherwise, surface melt occurs whenever the surface energy balance is positive and energy available for melting overcomes the layer cold content, i.e. once T as reached T_0 ,

$$Q_s = \rho_s L_0 h_s \quad (2.11)$$

as snow state changes. If the top layer completely melts, and energy is still available for melting, the following layer starts to melt as well. Both kinds of melting are computed separately and can therefore be calculated quantitatively. It must be noted, however, that the physical impacts of meltwater percolation into the snow are not treated while, in reality, the snow pack rapidly becomes wet as it starts melting. Since water saturation in snow affects the snow effective thermal conductivity and the snow response to diurnal changes in temperature, this lacking process may induce errors that are not accounted for in the temperature profile and mass balance analyses of section 2.4.

2.2.4.3 Snow ice formation

When the snow pack is heavy enough to depress the snow-ice interface below sea level, the freezing of a mixture of snow and seawater results in snow-ice formation (e.g., Eicken et al., 1994, 1995). Ice core observations reported in

Massom et al. (2001) showed that snow ice can contribute up to 38% to the Antarctic sea-ice mass balance. In the model, the process is parameterised like in Fichefet and Maqueda (1999), assuming that the snow-ice column is in hydrostatic balance:

$$h_{SI} = \frac{\rho_s h_s - (\rho_w - \rho_i) h_i}{\rho_s + \rho_w - \rho_i} \quad (2.12)$$

where h_{SI} and ρ_w (1020 kg m^{-3}) are the thickness of snow ice that forms and the density of seawater, respectively. The snow-ice layer is then merged with the underlaying sea ice. Although surface flooding can actually lead to slush layers that do not refreeze for long periods (e.g., Haas et al., 2001), snow-ice formation is assumed to occur instantaneously in the model.

2.3 Observations and forcing

2.3.1 Observations

The model evaluation is done in one Arctic and one Antarctic site using four different in-situ data sets: Point Barrow 2007, 2008, 2009 and Ice Station Polarstern (ISPOL, Hellmer et al., 2008).

At Point Barrow, we use data gathered in 2007, 2008 and 2009 by the Floating Ice Group of University of Alaska Fairbanks (Hajo Eicken and Chris Petrich, personal communication, 2009). These data were collected on mass balance sites installed in the Chukchi Sea about 1000 meters offshore of Niksuiaraq, the hook at the end of the road out to Point Barrow. The covered period extends from mid-January to the beginning of June every year with intensive temperature profile measurements through the air - snow - ice - ocean column.

For the ISPOL configuration, we use snow/ice data from the interdisciplinary ISPOL project conducted in the western Weddell Sea (Southern Ocean) in austral spring and summer 2004/2005. During the experiment, the German icebreaker RV Polarstern was anchored to an originally 10 by 10-km large ice floe. The station drifted 35 days from 28 November 2004 to 2 January 2005. The floe was composed of both first-year (FYI) and second-year ice (SYI), and detailed measurements of snow and ice properties were performed on 4 sites. We use snow-pit observations from station S6, the most comprehensive data set (for a full description of these data, see Nicolaus et al., 2006, 2009). Regular measurements of snow temperature and density profiles over December 2004 are compared with model outputs.

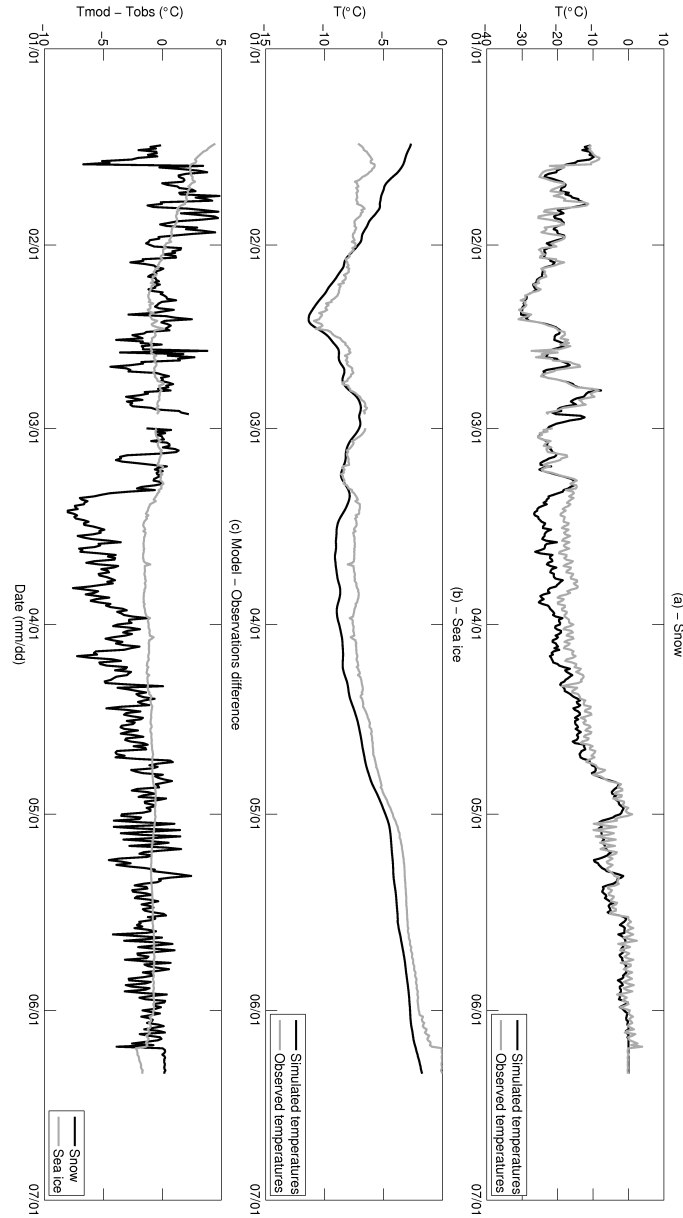


Figure 2.2: Temperature time series in snow at normalized height 0.5 (a) and sea ice at normalized depth -0.5 (b) for Point Barrow 2009 configuration. Plot (c) is the difference "Mod. minus Obs. temperature" at normalized levels 0.5 and -0.5 for snow and ice, respectively. Parameterisation (4a) is used for k_s , with 6 layers of snow in the model.

2.3.2 Forcing and model configuration

At Point Barrow, air temperature and relative humidity measurements are available in the data sets (at the same frequency as snow/ice data), but the other forcing variables, namely snowfall, wind speed and radiative fluxes (F^{sw} and F^{lw}) are retrieved from ECMWF ERA-Interim reanalyses and forecasts (Simmons et al., 2007) at a 1.5° spatial resolution. Data from the ocean/sea-ice grid point closest to the real location were taken and air temperature and relative humidity signals were checked for consistency with Point Barrow's signals. Temporal resolution of the runs is 6 hours and the modeling period is five months (period of observations). Snow and ice temperatures are initialized based on the observations, but snow density is initially set to 330 kg m^{-3} and the ice bulk salinity to 8 because no in-situ data are available.

At ISPOL, measurements of air temperature, relative humidity, wind speed, albedo and radiative flux measurements are used to run our model with a 1-hour time step during one month (December 2004). All snow and ice temperatures, snow density and ice salinity are initialized based on snow-pit and ice-core data.

The oceanic heat flux at the base of the ice slab is prescribed to 3 W m^{-2} in all Arctic configurations and to 17 W m^{-2} in the Southern Ocean. These values lie in the ranges proposed by Krishfield and Perovich (2005) and Heil et al. (1996), respectively, and were adjusted to improve the agreement between observed and simulated ice mass balances. The turbulent heat fluxes (F^{sh} and F^{lh}) are computed with bulk aerodynamic formulas as in Goosse (1997). The number of layers in the ice is set to 10 and several experiments with 1 to 6 six layers of snow are made.

2.4 Results and discussion

In order to assess the model's ability to simulate the internal conductive heat fluxes driving the sea-ice mass balance, both measurements of temperature and heat conductivity in snow are needed. Since we do not dispose of such data, we choose to evaluate temperature profiles instead. It is not fully equivalent of course, since a good agreement between observed and modeled conductive heat fluxes is bound to require both temperatures and thermal conductivities to be consistent with observations. However, the thermal conductivity tends to influence the magnitude of the conductive heat fluxes only, while the temperature gradients drive both their direction and intensity. We can therefore assume that comparing observed and modeled temperature gradients gives us a good insight on the model behaviour. In terms of correlations, agreement between conductive heat fluxes is mainly dependent on

temperature profile agreement, whereas bias or disagreement in heat conductivities in turn will mainly impact on the ice thickness maximum.

Hereafter, a simple comparison between observed and simulated temperature time series in snow and ice is done before showing the results of our more quantitative correlation analysis of temperature gradients in the snow pack. We finally look at the resulting mass balance in each model configuration.

2.4.1 Snow and sea ice temperatures

2.4.1.1 Method

To be able to compare observed and modeled snow/ice covers, we first normalize all vertical levels with respect to snow and ice thicknesses so that the height of any temperature point ranges between 0 and 1 for snow and 0 and -1 for ice (reference is taken at the snow/ice interface). Figure 2.2 shows temperature time series at normalized levels 0.5 and -0.5 in snow and ice, respectively. The plots are shown for Point Barrow 2009 setup, but are also representative of the other configurations.

2.4.1.2 Results

Snow and ice temperatures in the model tend to be lower than the observed ones, except at the beginning of the run. The maximum deviation in snow is about -8°C and the average deviation is -1.7°C . In the ice, the latter values are reduced to -4.5°C and -0.6°C , respectively. Large differences between modeled and observed ice temperatures are seen from the start of the run to the end of January and during March/April. The differences at the beginning of the run seem to be explained by the rough initialization of the ice bulk salinity profile (time series of ice salinity stabilize at the end of January too). March/April large errors may be due to the crude representation of the radiative transfer in the scheme.

2.4.2 Snow temperature gradients

2.4.2.1 Method

From the normalized snow and ice temperature grids, we first compute the linear correlation coefficient

$$r = \frac{\text{cov}(T_{\text{obs}}, T_{\text{mod}})}{\sigma_{T_{\text{obs}}} \sigma_{T_{\text{mod}}}} \quad (2.13)$$

between observed (T_{obs}) and simulated (T_{mod}) temperature profiles at times of available observations. $cov(x, y)$ is the covariance between x and y signals, and σ_x is the standard deviation of the x signal with respect to its mean. The number of correlation coefficients greater or equal to x relative to the total number of coefficients is then computed into r_x . Figure 2.3 shows this statistic at Point Barrow and ISPOL for different numbers of layers allowed in the snow scheme, and gives insight on the model performances in each setup. These configurations of the model are named LIM1D_ref for the reference run with the former thermodynamics (one layer of snow with constant physical properties), and LIM1D_i for a run with i layers of snow with varying density and thermal conductivity. Table 2.1 also presents the standard deviation, mean error and correlation between observed and simulated temperature profiles at Point Barrow in 2009.

To complement these statistics, we then look more quantitatively at snow temperature gradients using Taylor diagrams (Taylor, 2001). Taylor diagrams provide a way of graphically summarizing how closely a pattern (or a set of patterns) matches observations. The similarity between two patterns is quantified in terms of their correlation, their centred root-mean-square difference and the amplitude of their variations (represented by their standard deviations). These diagrams are often used for evaluating multiple aspects of complex models or in gauging the relative skill of many different models (e.g., Houghton et al., 2001). We use it here to compare the performances of our model in different configurations. Thus, time series of mean vertical temperature gradients in the model are retrieved in order to compute the correlation and root-mean-square difference with the observed ones. Taylor plots are finally made to summarize the information and shown in Figure 2.4 for both Point Barrow 2009 and ISPOL.

2.4.2.2 Results

Snow temperature gradients are sensitive to the number of layers in the model. This was expected, since the layer number determines the resolution and influences the heat conduction in snow. Indeed, correlation statistics and Taylor plots consistently show a significant increase of correlations between model and observations until a stabilization occurs from a certain configuration depending layer number. Hence, no run with more than six layers of snow, which was the stabilization threshold at ISPOL, is presented. The Point Barrow configurations resulted in thresholds of 4, 3, and 4 layers for years 2007, 2008, and 2009, respectively. The increase in the amount of good correlations (considered as correlations greater than 0.8 at Barrow and 0.4 at ISPOL) was different for each configuration. Statistics of Table 2.1 also exhibit a threshold;

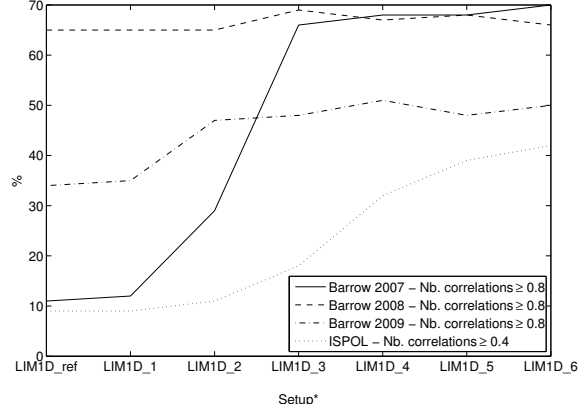


Figure 2.3: Fractional amount (in %) of correlations greater or equal to 0.8 (for Point Barrow) and 0.4 (for ISPOL) between observed and simulated snow temperature profiles, relative to the total number of observed profiles. The results are presented for each model configuration. "ref" index corresponds to a reference run with one layer of snow, constant density and thermal conductivity. Other indices correspond to the number of layers in the run. Parameterisation (4a) is used for k_s .

Table 2.1: Standard deviation, mean error (both in $^{\circ}\text{C}$) and correlation between observed and simulated temperature profiles in snow at Point Barrow 2009. Abbreviations: Obs. Std. and Mod. Std. mean standard deviation of observed and modeled temperature profiles over one run, Mean err. and Corr. mean error and mean correlation between observed and simulated profiles over one run. Parameterisation (4a) is used for k_s .

Setup ^a	Mod. Std.	Obs. Std. = 2.05	
		Mean err.	Corr.
LIM1D_ref	1.69	3.82	0.57
LIM1D_1	1.79	3.72	0.57
LIM1D_2	2.05	3.56	0.58
LIM1D_3	2.29	3.47	0.64
LIM1D_4	2.41	3.48	0.66
LIM1D_5	2.42	3.47	0.65
LIM1D_6	2.46	3.48	0.67

^a "ref" index corresponds to a reference run with one layer of snow, constant density and thermal conductivity. Other indices correspond to the number of layers in the run.

at Point Barrow 2009 the mean error (mean correlation) between observed and modeled temperature profiles decreases (increases) until the 4-layer configuration is reached, and then stabilizes. Initial correlation values (for reference or 1-layer run) were also very unlike in each run. We tend to attribute these results to the importance of representing well the real snow stratigraphy at the time and place of simulation. The scheme performs better with an increasing layer number, which is expected numerically, since the resolution allows a finer temperature profile characterization. However, the number of snow layers required to represent the different snow types in the real stratigraphy also comes into play. For one given layer number, the correlation between observed and modeled temperatures will be the better the closer the scheme describes observed snow properties. However, it must be noted that snow densities could only be initialized from observations at ISPOL and not at Point Barrow. This is in little contradiction with the rather weak ISPOL correlation values compared to Point Barrow ones, but somehow agrees with the uniform increase in the model's performances with the layer number at ISPOL. Indeed, the initialization file contains a snow pack with five layers of distinct densities, and the performances of the scheme are therefore enhanced when the layer number gets closer to the number of layers required to discriminate the snow types in the initial stratification.

This convergence in the models' skills with numerical resolution was also found by Cheng et al. (2008). When assessing the sensitivity of their thermodynamic snow/sea-ice model to its vertical resolution, they showed that the model produced better temperature profiles with greater layer numbers, with a stabilization of these improvements at around 20 layers (for the whole snow - sea-ice column). Increasing the vertical resolution in their model also enabled the subsurface melting of snow and ice. Our experiments lead to comparable results, except LIM1D exhibits a faster convergence (from 3 to 6 layers of snow, or 13 to 16 layers for the whole snow/sea-ice system) and allows the internal melting of snow to occur with 3 layers of snow or more (see ISPOL section below).

The frequency of observations in each data set and the differences in run's length may be the reasons for ISPOL's correlations being lower than Point Barrow's ones. Contrary to Point Barrow, observations at ISPOL are not available at each time step and are compared with the closest model iteration. Correlation's calculations in this case are therefore less robust because of this lack of regular data. Last but not least, ISPOL run covers only a short summertime period during which snow temperatures remain close to 0°C and vertical temperature gradients are likely to change or reverse rapidly. This consequently increase the chances of disagreement with observations compared to Point

Barrow wintertime conditions during which the downward vertical temperature gradient is well established.

Lastly, Table 2.1 exhibits large mean errors in snow temperature profiles. They could possibly be due to the missing impact of water saturation in snow (see section 2.2.4.2) as it starts to melt, or to an inadequate extinction coefficient in the computation of the radiative transfer in snow.

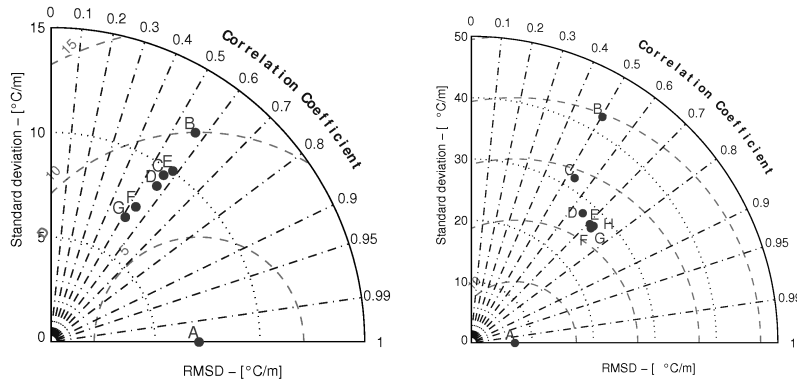


Figure 2.4: Taylor plots of snow temperature gradient time series at ISPOL station (left) and Point Barrow (right). Dotted circular lines are iso-lines of standard deviation. Dashed circular lines represent iso-lines of centred root-mean-square difference (RMSD) between simulated and observed temperature gradient time series. Cosine of the angle relative to the horizontal corresponds to the correlation between model times series and observations. A dot corresponds to observations, B is reference run and C to G are 1 to 6-layer configuration runs. For example, B dot on plot (a) means that the temperature gradient time series of reference run has a standard deviation of about $12.5^{\circ}\text{C m}^{-1}$, an RMSD of $10^{\circ}\text{C m}^{-1}$ and a correlation of 0.57 with the observed time series (A dot). Parameterisation (4a) is used for k_s .

2.4.3 Snow and ice thicknesses

2.4.3.1 Point Barrow

Figure 2.5 (a), (b) and Figure 2.6 (c) illustrate the snow and ice mass balances for 6-layer model configurations at Point Barrow for the three years. Over the entire observation period, the general snow depth evolution is consistent with observations, with an average error of 1 cm for 2007, -2.5 cm for 2008, and -3.4 cm for 2009. Maximum snow depth deviation is about -9 cm. Although these differences are significant, they lie within the expected range of natural short-term and small-scale variability. Indeed, since the geographical distribution

of snow depth is highly variable even at subkilometer scales, it is impossible to simulate the exact snow depth with the large-scale reanalyses of snow precipitation used to force our model. In addition, new snow accumulation also greatly depends on blowing snow effects that transport both falling and formerly-deposited snow from place to place. One solution to improve the match between simulated and observed snow thicknesses could have been to derive snowfall inversely from our snow depth data. However, this scheme being developed to be further included in a GCM, we chose to force it with forcings typically used in such models. Another reason for these discrepancies between observed and modeled snow depths from mid-May to the end of the run is that the mast on which snow pingers were installed at Point Barrow sometimes started to slip vertically through the ice pack at the beginning of the melt season. As we did not correct the data for this, they are questionable during the melt period. Still, simulated and observed melt periods seem to be consistently happening during the last 15 days of the run with relatively close melting rates (especially for 2007 melt period and May 2009).

Regarding the ice mass balance, average deviations in ice thickness (i.e. mean value of the "model minus observed thickness" difference) are -10.3 cm, -7 cm and -2.2 cm in 2007, 2008 and 2009, respectively. The corresponding errors for LIM1D_ref runs are -13.4 cm, -8.4 cm and +3.3 cm. The new model therefore produces a slightly better ice thickness, but 2007 and 2008 runs still exhibits large differences with observed ice thicknesses, as shown in Figure 2.5.

As expected, the maximum ice thickness is sensitive to the snow thermal conductivity. Parametrisation (4a) was initially used in every model run (e.g. run of Figure 2.6 (c)), but parameterisation (4b) sometimes proved to reduce the maximum ice thickness difference (used in runs of Figure 2.5). Error in maximum ice thickness is -1.8 cm with (4b) instead of -7 cm with (4a) at Point Barrow 2008. For both 2008 and 2009, parameterisation (4c) was also tested and seemed to cause a small overestimation in maximum ice thickness (about +4 cm). In the 2007 run however, the final shift between observed and modeled ice thicknesses (about -12 cm) could not be reduced significantly using any of (4a), (b) or (c) formulas for k_s without tuning them to values we are not able to justify. We can not fully explain why the difference between observed and modeled ice thicknesses is larger in 2007, but the two major uncertainties are the following. First, the oceanic heat flux, that controls ice growth at the bottom interface, is set constant, while it clearly has non-resolved temporal variations. The standard deviation of the oceanic heat flux in the Arctic has been shown up to 15 W m^{-2} (see, e.g., Krishfield and Perovich, 2005) over its seasonal variations. Second, snow physical properties could not be properly initialized at Point Barrow. Therefore, relatively large errors in snow density

and thermal conductivity may lead to significant discrepancies between the intensity of the real and simulated conductive heat fluxes from the ice to the snow.

2.4.3.2 ISPOL

Figure 2.6 (d) shows the simulated snow and ice thicknesses compared with ISPOL observations. Snow thickness variations are rather weak for two reasons. First, there were only two major snowfall events during the period of simulation. The first occurred between 28 November and 2 December, and increased the snow thickness by 6.8 cm. The second took place between 26 and 27 December, and added 1.3 cm of snow (Nicolaus et al., 2009). The run starts after the first snowfall episode. Second, melt conditions are not reached or only for short periods of time. Snow thinning by melt is therefore rather sporadic and weak, consistently with Nicolaus et al. (2009). More precisely, snow thinning events are all time-localized episodes of internal melting, that can be seen in Figure 2.6 (d) where 0°C closed-contour lines are drawn. The main internal melt events happen on December 4, 6 and 8, with respective melting rates of 0.18, 0.15 and 0.12 mm of water per hour and each time correspond to an approximate snow thinning of 1.5 cm. Surface melt conditions are never reached.

Snow depth and ice thickness average deviations are -1.6 cm and -1.2 cm, respectively, with a maximum deviation in ice thickness of -7.1 cm. The same comments as for Point Barrow configurations can be made to explain these errors, especially for the basal oceanic heat flux, of which the intensity depends very much on the feedbacks between the sea-ice growth rate and the oceanic convective activity. Indeed, Southern Ocean's stratification is weak and sensitive to perturbations in freshwater flux. A slower (faster) ice growth rate results in weaker (stronger) brine rejections and in a subsequent strengthening (weakening) of the ocean stratification, finally leading to a smaller (larger) oceanic heat flux at the ice bottom (Fichefet et al., 2000).

Observed and simulated ice ablation rates seem to be consistently constant. Thanks to the frequency of the meteorological observations available on the ISPOL floe, we were able to force the model at a higher temporal resolution (1 hour) than at Point Barrow. The diurnal cycle in snow temperatures is consequently much better resolved. However, if temperature contours in snow and ice show that the temperature of snow uppermost layers follows the diurnal cycle of air temperature and solar irradiance, ice hardly feels these variations. This is probably due to the fact that air temperature variations are not strong and long enough to overcome the insulating effect of snow, thus making the snow-ice interface temperature quite constant. Therefore, ice inner

temperature gradients being subsequently small and constant as well, added to a constant oceanic heat flux at the sea ice base, explains the almost constant ablation rate. At ISPOL, the snow and ice sensitivity to the snow heat conductivity is not as pronounced as at Point Barrow and the sea-ice bottom melting rate is mainly sensitive to the oceanic heat flux. However, to better assess this sensitivity, working with a coupled sea-ice - ocean model is suggested. Doing so would allow to account for the seasonal variability of the oceanic heat flux to the ice and for the associated feedbacks, related to the depth of the oceanic upper mixed layer as the ice grows or melts (Stössel et al., 1998; Winton, 1999).

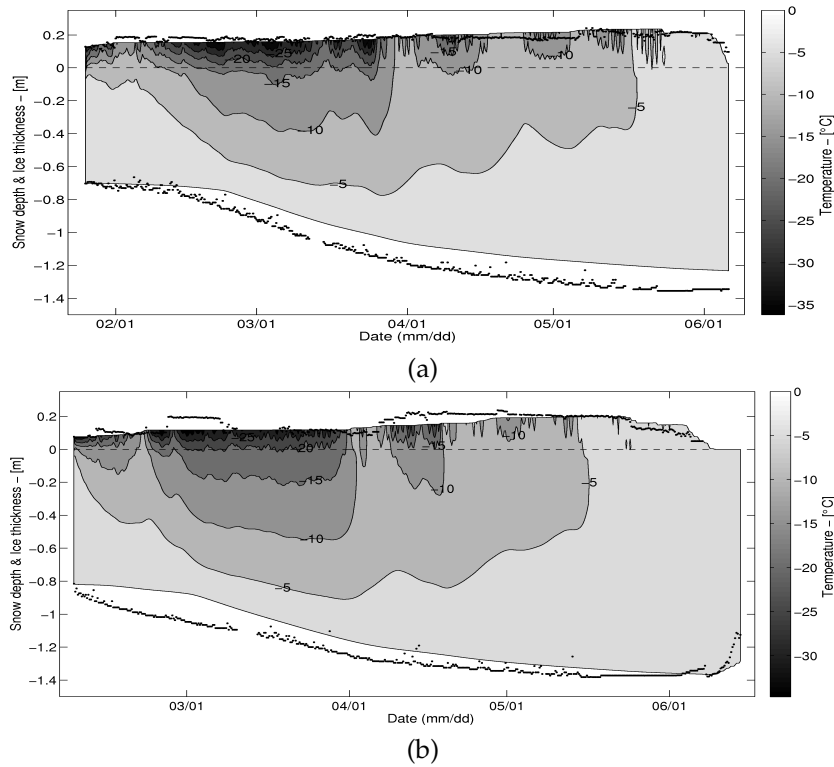


Figure 2.5: Mass balance and contours of temperature in the snow/ice system (part 1) as simulated by the model at Point Barrow 2007 (a) and 2008 (b). Dots refer to observations of snow height and ice depth. 0-level is the snow-ice interface. Parameterisation (4a) is used for k_s , with 6 layers of snow in the model.

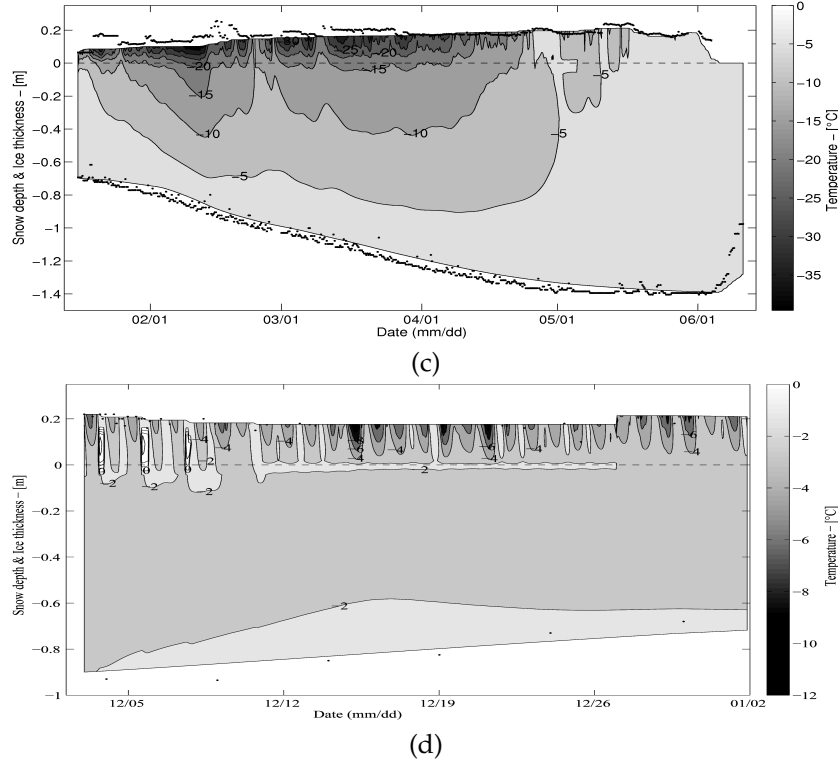


Figure 2.6: Mass balance and contours of temperature in the snow/ice system (part 2) as simulated by the model at Point Barrow 2009 (c) and ISPOL (d). Dots refer to observations of snow height and ice depth. 0-level is the snow-ice interface. Parameterisation (4a) is used for k_s , with 6 layers of snow in the model.

2.5 Conclusion

A new one-dimensional thermodynamic snow scheme with a view towards global-scale simulations was presented. The scheme includes several layers of snow with varying density and thermal conductivity, and particularly allows the surface and internal melting of snow, snow-ice production, and penetration of solar radiation into the snow pack. The evaluation of this new scheme was done by comparing snow temperature profiles and thickness of the snow/ice system in the model to data sampled at Point Barrow and on the ISPOL floe. The performances of the model were also compared to those of the former representation of snow in LIM, taken as a reference run. This representation includes only one layer of snow with constant physical properties.

Correlation values between observed and modeled temperature profiles increase with the number of snow layers allowed in the scheme. In average over all runs, these correlations are 27% better with the new scheme with six layers of snow, compared to the reference run. However, after a certain number of layers, correlation values stop improving. This number of layers is different for nearly each model setup (3 for Point Barrow 2008 setup, 4 for 2007 and 2009, 6 for ISPOL). Cheng et al. (2008) recommended that climate and weather prediction models should run with 3 to 15-20 layers (for snow and sea ice). Our results consistently suggest that 3 layers is a minimum for snow to produce reasonable vertical temperature profiles. Another common result between these two studies is the improvement of the model's skills with increasing number of snow layers. Large-scale sea-ice models would therefore benefit from running with higher vertical resolution, although this resolution is bound to be limited for computational cost reasons.

With the new snow scheme, sea-ice thickness is only slightly better represented, with an average error (over all runs) between observed and simulated thicknesses of 6.5 cm, against 8.4 cm for the reference runs. The latter improvement is weak compared to the improvement in the temperature profile reproduction. This is partially due to the fact that the real snow stratigraphy was not really well represented even in the new scheme. Indeed, both realistic temperature gradients and thermal conductivity profiles in snow are required to obtain accurate ice thickness estimates. These conclusions ultimately suggest that the representation of processes driving snow properties, lacking in global-scale models (e.g., metamorphic processes), must be compensated by at least a good representation of snow density distributions.

The modeled maximum ice thickness at the end of the growth period is, as expected, sensitive to the snow thermal conductivity parameterisation. Even so, we were not able to identify one specific formulation showing better results than the others in every case. Variations in the radiative parameters may also affect the results. These parameterisations will have to be tested again once the scheme is implemented in NEMO-LIM3.

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CHAPTER 3

SNOW THERMAL CONDUCTIVITY IN LARGE-SCALE SEA ICE MODELS

This Chapter is based on the following paper:

Lecomte, O., Fichefet, T., Vancoppenolle, M., Domine, F., Massonnet, F., Mathiot, P., Morin, S., Barriat, P., 2013. On the formulation of snow thermal conductivity in large-scale sea ice models. *Journal of Advances in Modeling Earth Systems* 5 (3), 542–557.

URL <http://dx.doi.org/10.1002/jame.20039>

Abstract

An assessment of the performance of a state-of-the-art large-scale coupled sea ice–ocean model, including a new snow multilayer thermodynamic scheme, is performed. Four 29-years-long simulations are compared against each other and against sea ice thickness and extent observations. Each simulation uses a separate parameterization for snow thermophysical properties. The first simulation uses a constant thermal conductivity and prescribed density profiles. The second and third parameterizations use typical power-law relationships linking thermal conductivity directly to density (prescribed as in the first simulation). The fourth

parameterization is newly developed and consists of a set of two linear equations relating the snow thermal conductivity and density to the mean seasonal wind speed. Results show that simulation 1 leads to a significant overestimation of the sea ice thickness due to overestimated thermal conductivity, particularly in the Northern Hemisphere. Parameterizations 2 and 4 lead to a realistic simulation of the Arctic sea ice mean state. Simulation 3 results in the underestimation of the sea ice basal growth in both hemispheres, but is partly compensated by lateral growth and snow ice formation in the Southern Hemisphere. Finally, parameterization 4 improves the simulated snow depth distributions by including snow packing by wind, and shows potential for being used in future works. The intercomparison of all simulations suggests that the sea ice model is more sensitive to the snow representation in the Arctic than it is in the Southern Ocean, where the sea ice thickness is not driven by temperature profiles in the snow.

3.1 Introduction

The thermo-physical properties of snow on sea ice play a key role in the physics of heat conduction through the snow - sea ice system. In particular, several studies have shown the importance of the snow thermal insulating effect on the sea ice growth and decay (e.g., Eicken et al., 1995; Fichefet and Morales Maqueda, 1997; Sturm et al., 2002). By strongly curtailing the conductive heat flux from the sea ice to the snow-atmosphere interface and therefore heat losses to the atmosphere, the low thermal conductivity of snow reduces the basal ice growth.

The process of heat transfer through the snow includes mainly three mechanisms : (1) heat conduction through the ice matrix formed by the interconnected grains, (2) heat conduction in the interstitial air and (3) the transfer of latent heat by water vapor that sublimates from warmer grains to condense on colder ones. With regard to this, a variable called *effective thermal conductivity* (k_{eff}) is typically used, accounting for all three processes. However, the term *effective thermal conductivity* is probably not the most appropriate, because it includes transfer by latent heat, which is not a conductive process. Following Calonne et al. (2011) we therefore refer to the variable regrouping the above three processes as the *apparent thermal conductivity*, k_{app} . Another potentially important process, heat advection by wind pumping of air over a rough snow surface, is not included in k_{app} and has to be simulated separately if represented in a model. The range of possible values for k_{app} extends from about 0.03 to $0.65 \text{ W m}^{-1} \text{ K}^{-1}$ (Sturm et al., 1997, 1998) and several regression curves relating snow thermal conductivity to snow density ρ_s were proposed (e.g., Yen, 1981; Sturm et al., 1997, among the most commonly known and used in the modeling community). Yet, most large-scale sea ice models (coupled to ocean and/or atmospheric general circulation models or not) have been using the $0.31 \text{ W m}^{-1} \text{ K}^{-1}$ constant value proposed by Maykut and Untersteiner (1971) and Semtner Jr (1976). Until now, the same constant has been used in the Louvain-la-Neuve Sea Ice model (LIM), which includes, like all those models, a relatively simple snow representation with only one snow layer and constant snow density and thermal conductivity. Even so, Wu et al. (1999) and Fichefet et al. (2000) showed that their simulated Antarctic sea ice exhibits a strong sensitivity to snow thermal conductivity, despite the rather simplistic snow representation in their models. In particular, they found the modeled average Antarctic sea ice to be on average 10% thinner when using a value reduced by half for k_{app} with respect to the usual value.

The present study aims at extending those works and investigates the sensitivity of a large-scale coupled sea ice-ocean model to the representation of snow thermal conductivity using a new, more elaborate snow thermody-

namic scheme. In particular, we show the impacts of a newly developed and simple parameterization of snow packing by wind on the simulated snow depth distributions. The second objective is to study the reproducibility of Lecomte et al. (2011)'s results, and assess the differences in response of the sea ice model to various thermal conductivity formulations between both hemispheres. Benefit is taken from the host coupled model including oceanic feedbacks to explain those dissimilarities. Lastly, we evaluate the ability of this model to simulate the sea ice extent and thickness in both hemispheres with respect to observations. The paper is laid out as follows: section 2 introduces the sea ice–ocean model and the snow scheme, section 3 details the experimental setup and section 4 describes the observations used to evaluate the model outputs. In sections 5–6, we respectively present and discuss the mean state and variability of the sea ice extent/thickness for each simulation. Findings are summarized in section 7.

3.2 Models description

For the purpose of this study, we use the global coupled ocean–sea ice model NEMO-LIM, in the same configuration as in Massonnet et al. (2011). Both the ocean–sea ice model and snow scheme we use are described in the next two sections.

3.2.1 Sea ice and ocean

LIM3 (Louvain-la-Neuve Sea Ice Model, version 3) is a state-of-the-art thermodynamic–dynamic sea ice model. Its comprehensive description is given in Vancoppenolle et al. (2009). This model includes an explicit representation of the subgrid-scale distributions of ice thickness, enthalpy, salinity and age (using 5 ice categories). The thermodynamic component is a multilayer scheme (5 layers of ice) based on the energy-conserving model of Bitz and Lipscomb (1999) and sea ice halodynamics are represented using an empirical parameterization of the effects of brine convection (gravity drainage) and percolation (flushing and snow ice formation) on the vertical salinity profile. Sea ice dynamics are solved using the elastic-viscous-plastic (EVP) rheology of Hunke and Dukowicz (1997) and the C-grid formulation of Bouillon et al. (2009).

LIM3 is fully coupled with the oceanic general circulation model (GCM) OPA (Ocean PARallelisé, version 9, see Madec (2008) for the full documentation) on the modeling platform NEMO (Nucleus for European Modeling of the Ocean) following the formulation of Goosse and Fichefet (1999). All simulations are performed on the model global tripolar ORCA1 grid (1 degree

resolution) with 42 vertical levels. A sea surface salinity restoring toward the Levitus (1998) climatological values is added to the freshwater budget equation to prevent spurious model drift.

3.2.2 Snow scheme

3.2.2.1 General description

Initially, the snow in LIM3 was, as in most ice-ocean GCMs, represented as one single layer with constant thermo-physical properties. From this starting point, two general lines of model improvement were considered. The first is the sophistication of the physical and numerical framework used for snow representation (e.g., increasing the number of layers, introducing varying density, etc.), which is a prerequisite for parameterizing important processes such as penetration of shortwave radiation, snow packing and drift. The second line of improvement, i.e., the representation of the latter processes themselves, is partially addressed in this study and will be further developed in future studies. Thus, a multilayer thermodynamic snow scheme of intermediate complexity was developed and tested in the one-dimensional (1-D) version of LIM (LIM1D) (Lecomte et al., 2011). This scheme was then modified and included into the full version of NEMO-LIM3 for the present work in order to add a slightly improved radiation scheme, a new snow thermal conductivity formulation, and make it consistent with the structure of a three-dimensional model with horizontal advection. The final scheme, described hereafter, therefore provides a better characterization of the snow layer properties (density and thermal conductivity) and of their impacts on heat conduction through the snow ice system, compared to the initial snow representation in LIM3.

For each sea ice thickness category in a given horizontal grid cell of the model, snow is considered as a horizontally uniform snow pack on sea ice, with a thickness h_s . At each depth z within the snow, the thermodynamic state of the medium is characterized by temperature $T(z)$, density $\rho_s(z)$ and effective thermal conductivity $k_{app}(z)$. The vertical snow temperature profile is governed by the one-dimensional heat diffusion equation:

$$c_s \rho_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(k_{app} \frac{\partial T}{\partial z} \right) - \frac{\partial I(z)}{\partial z} \quad (3.1)$$

where $c_s = 2100 \text{ J kg}^{-1} \text{ K}^{-1}$, ρ_s and k_{app} are the specific heat, density (handled as described in section 3.2.2.2.1.) and thermal conductivity of snow (parameterized as a function of either density or wind speed, as detailed in section 3.2.2.2.2.), respectively. $I(z)$ is the solar radiation penetrating into snow

at depth z . Practically, the downwelling radiation decreases exponentially through the snow, except in its highly scattering uppermost portion (Perovich, 2007). This layer is usually referred to as the “surface scattering layer”(SSL). In the new radiative scheme, we assume that the surface layer absorbs all near-infrared solar radiation (Dozier and Warren, 1982; Warren, 1982), so that the remaining radiation is in the visible and UV parts of the solar spectrum. Therefore, the snow column is optically divided into a highly scattering surface layer of thickness h_{SSL} and several deeper layers. The solar radiation penetrating under the SSL is:

$$I_0 = (1 - \alpha)i_0 F^{sw}, \quad (3.2)$$

where α is the surface albedo, F^{sw} the incident solar radiation at the surface, and $i_0 = 0.18$ the fraction of solar radiation penetrating under the SSL (value of Grenfell and Maykut, 1977). The albedo is computed as a function of the surface state, cloud cover, ice thickness and snow depth (Shine and Henderson-Sellers, 1985, see Appendix A). Radiation transmission through the deeper layers is then computed following Beer’s law, using a set of extinction coefficients κ_s for dry and wet snow from Järvinen and Lepparanta (2011):

$$I(z) = I_0 e^{-\kappa_s z}, \quad (3.3)$$

where $z > h_{SSL}$ is the snow depth. h_{SSL} is assigned to 3 cm (value of Perovich, 2007). Radiation absorption into the ice is also treated using Beer’s law, with an attenuation coefficient $\kappa_i = 1 \text{ m}^{-1}$. The crude representation of the radiative transfer in snow and sea ice should definitely be improved, for instance through the use of more sophisticated schemes (e.g., Briegleb and Light, 2007; Holland et al., 2012), which will be the subject of future work.

The surface energy balance provides the boundary condition at the top of the snow cover (fluxes are defined positive downwards):

$$F_0^{ct} = (1 - \alpha)(1 - I_0)F^{sw} + F^{lw} - \epsilon_s \sigma T_s^4 + F^{sh} + F^{lh} \quad (3.4)$$

F_0^{ct} being the conductive heat flux in snow just below the surface, F^{lw} the downward longwave radiation, $\epsilon_s = 0.97$ and T_s the surface emissivity and temperature, respectively, and F^{sh} and F^{lh} the turbulent fluxes of sensible and latent heat. In LIM3, the heat content of 3 snow physical layers and the total snow mass (per ice category) are advected horizontally with the sea ice, as suggested by Lecomte et al. (2011). This minimum number of layers was chosen to avoid an excessive increase in computing time, closely related to the number of state variables advected in the model. Considering that no optimization of the code was done at this stage, this configuration leads to

a lengthening of the actual computing time by $\sim 40\%$ compared to the standard version of NEMO-LIM3 with a single snow layer. Most of this cost is due to the advection of snow mass and temperature tracers (20 tracers) using the Prather scheme, which cost significantly increases with a large number of tracers. However, the snow vertical grid is refined in the thermodynamics so that the heat diffusion equation is solved within 6 layers instead. This is done using an energy-conserving method and enables the thermodynamic scheme to better simulate the temperature gradient within the snow cover.

Once the new temperatures are computed, the snow mass balance is calculated as:

$$\frac{\partial M_s}{\partial t} = R_S - R_{M/Su} - R_{SI} \quad (3.5)$$

where M_s is the snow mass per unit area (integral of snow thickness times density over all layers), R_S the snowfall rate, $R_{M/Su}$ the melt or sublimation rate and R_{SI} the snow ice formation rate. $R_{M/Su}$ accounts for both surface and internal melts, and surface sublimation whenever the snow surface specific humidity is larger than the air specific humidity. Basically, when the surface energy balance is positive for snow (taking into account all incoming and outgoing fluxes) and surface temperature reaches the melting point, snow melts. If the snow surface specific humidity (relative to the temperature of the surface layer) is larger than the air specific humidity and the air relative humidity is lower than 100%, snow sublimates instead of melting. Since the energy by unit mass required to sublimate snow is about ten times larger than the one for melting it, snow mass losses by surface ablation in case of sublimation are significantly reduced compared to regular melt conditions (Nicolaus et al., 2009).

R_{SI} is parameterized as in Fichfet and Morales Maqueda (1997). When the snow pack is heavy enough to depress the snow-ice interface below the sea level, a mixture of snow and sea water flooding the interface refreezes to form snow ice that is merged with the underlying sea ice. The thickness of the snow ice layer is computed assuming the hydrostatic equilibrium of the snow/ice column into the ocean.

The snowfall rate, R_S is initially derived from climatological data, as a surfacic water mass flux (in $\text{kg m}^{-2} \text{s}^{-1}$). Then, R_S is converted into snow accumulation Δh^{sfall} :

$$\Delta h^{sfall} = \frac{f^{sfall} f^{solid} R_S \Delta t}{\rho_s} \quad (3.6)$$

where f^{solid} , f^{sfall} and Δt are the fraction of precipitation that falls as snow, the fraction of solid precipitation falling on the ice-covered part of the grid cell and the model time step, respectively. f^{solid} is computed as a function of

surface air temperature following Ledley (1985), and f^{sfall} is defined as:

$$f^{sfall} = \frac{1 - (1 - a_i)^\beta}{a_i} \quad (3.7)$$

where a_i is the ice concentration (area fraction) and β a blowing snow parameter, accounting for the impact of the wind on the amount of snow blown into the ocean. If winds are assumed to have no impact on falling snow ($\beta = 1$), the same amount of snow falls onto ice and into open water. However, because of the winds, a fraction of the snow falls in leads, which is represented by $\beta < 1$. The unrealistic extreme case for which all the snow falls into open water is represented by $\beta = 0$. By default, its value is arbitrarily set to 0.6.

3.2.2.2 Parameterizations

3.2.2.2.1 Snow density profile The representation of snow metamorphism in a model, implying the characterization of the snow pack through the snow grain size and shape, requires a high level of sophistication that is not compatible with both the spatial scales resolved explicitly by GCMs and the computational costs of such models. Despite beneficial for the representation of some processes, such as the physics of depth hoar formation (see, e.g., Sturm and Massom, 2009), a comprehensive description of metamorphism like the one used in complex land snow models (e.g., CROCUS, Brun et al., 1989) was not included in LIM3. Besides, in contrast with the density of continental snow, the density of snow on sea ice is affected by sea ice dynamics and topography.

Nonetheless, the impacts of such mechanisms on snow are so poorly understood that prognosing the evolution of snow density in a sea ice model is hardly feasible at this stage. Instead, a density profile is prescribed as a function of the sea ice age and thickness, and only the snow mass (not volume) is advected on the horizontal grid of the model. This is done by assuming that the snow depth and stratigraphy are highly conditioned by the history of the underlying sea ice. Studies presenting specific snow stratigraphies on first-year ice (FYI) and multiyear ice (MYI) such as Nicolaus et al. (2009) support this assumption. Two main observational datasets were used to build the snow density profiles for FYI and MYI in the model. The first is from the interdisciplinary Ice Station POLarstern (ISPOL; Hellmer et al., 2008) project, Weddell Sea, Antarctica. The campaign was conducted in the western Weddell Sea, Southern Ocean in austral spring and summer 2004/05. Regular measurements (over December 2004) of snow density profiles were made in four sites including both FYI and MYI around the drifting station. A full description of these data can be found in Nicolaus et al. (2006) and Nicolaus et al. (2009). In the Arctic, snow density data were taken from the Canadian Arctic

Shelf Exchange Study (CASES) that took place in 2003. This dataset, described in Langlois et al. (2007), provides snow density profiles on first-year land-fast sea ice offshore in Franklin Bay, Northwest Territories, Canada.

Firstly, those snow pit data were processed to bring all densities and layer thicknesses back on a normalized vertical grid with depth $\hat{z}$ varying from 0 (surface) to 1 (bottom). The normalized density profiles were then averaged over snow samples originating from sea ice of similar type and thickness. From this, a uniform snow density profile of 290 kg.m^{-3} and a vertical profile $\rho_s(\hat{z}) = 291 + 89\hat{z} [\text{kg m}^{-3}]$ were chosen to be prescribed in the model for snow on FYI and MYI, respectively. Given the very high spatial (even at regional scale) and temporal variability of the snow cover on top of sea ice in both hemispheres (e.g., Massom et al., 2001; Sturm et al., 2002), this modeling choice might appear relatively simple. However, it is a first attempt to account for the impact of the variability of the snow stratigraphy on the large scale sea ice thermodynamics. In the context of the multiplication of snow in-situ data sources, this method could be refined to add more detailed and distinct profiles as a function of ice type in large-scale sea ice models.

3.2.2.2.2 Snow thermal conductivity The snow thermal conductivity is usually related to snow density through data regressions. Here, we use those of Yen (1981) (3.8) and Sturm et al. (1997) (3.9):

$$k_{app} = k_i \left(\frac{\rho_s}{\rho_w} \right)^{1.88} \quad (3.8)$$

$$k_{app} = 0.138 - 0.00101\rho_s + 0.000003233\rho_s^2 \quad (3.9)$$

with k_i and ρ_w being the thermal conductivity of pure ice ($2.01 \text{ W K}^{-1} \text{ m}^{-1}$) and the density of freshwater (1000 kg m^{-3}), respectively. Those relationships were chosen in Lecomte et al. (2011) because they are broadly used in the snow modeling community. However, due to the dissimilar meteorological conditions, sea ice types and thickness distributions in the two hemispheres, the Arctic and Antarctic sea ice snow cover properties (in particular density stratigraphy) are substantially different (Sturm and Massom, 2009). In light of the current state of snow representation in sea ice models, these differences cannot be well represented yet. The evaluation of snow thermal conductivity based on such relationships is therefore critical and sometimes leads to large errors in thermal conductivity (and subsequently in sea ice state estimates) according to whether they are used in the Northern or Southern Hemisphere. For these reasons, we have sought to develop simple parameterizations of snow density and thermal conductivity that would be independent from those

considerations and would directly relate the snow density and thermal conductivity to climate variables such as the surface wind speed, the surface air temperature or the temperature gradient in the snowpack. Here, we briefly present the correlations obtained in Domine et al. (2011a) between the snow density and the mean seasonal surface wind speed and between the snow thermal conductivity and the mean seasonal surface wind speed. The rationale is the observation, common and shared by virtually all field snow scientists, that the hardness and density of dry snow are largely determined by the wind intensity. This widespread observation has been published and used in continental snow models (e.g., Brun et al., 1997; Vionnet et al., 2012) but, to the best of our knowledge, never in sea ice models. Since both snow hardness and density are good predictors of the snow thermal conductivity (Domine et al., 2011b), relating this latter variable to the surface wind speed also appears logical.

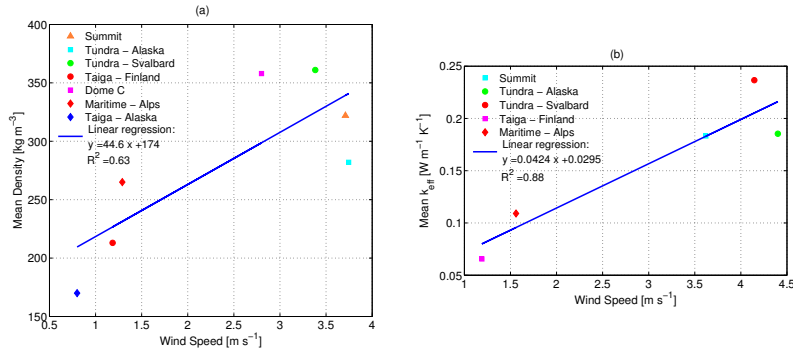


Figure 3.1: Correlation between the mean snow density (a) / thermal conductivity (b) and the mean seasonal surface wind speed at various locations (see section 2.2.3.1. for details).

We used data from field campaigns in Sodankylä (Northern Finland), Barrow (Alaska, Domine et al., 2012), the Col de Porte (French Alps), Ny Alesund (Svalbard, Domine et al., 2011b) and Summit (Greenland) for thermal conductivity. For density, in addition to the aforementioned locations, we used data from Fairbanks (Alaska, Taillandier et al., 2006) and Dome C (Antarctica). The data from places with references have been published, and most of the others have been obtained for a purpose totally different from that of the current study. Data from Col de Porte, Summit and Dome C have not been published yet. As detailed in Domine et al. (2012), the snow density was measured with density cutters, i.e., by extracting and weighting a known volume of snow. The snow thermal conductivity was measured with a TP02 heated

needle probe from Hukseflux. The principle is to measure the rate of dissipation of a 100 s heat pulse by monitoring the temperature rise of the heated needle (Sturm et al., 1997; Morin et al., 2010). This rate is related to snow thermal conductivity. Meteorological data was either measured during the campaigns or taken from permanent meteorological stations at the sites. The temperature gradient within the snow was determined from vertical strings of thermistors placed in the snow during the campaign. The site of Sodankylä is in the forest and wind speed is measured above the canopy. This is clearly not representative of wind speed at ground level. At this site, essentially no wind could be felt at ground level and there was never any blowing snow. These conditions were similar to those experienced in Fairbanks, so that the wind speed above the canopy was scaled to the ratio between the wind speeds above and below the canopy from Fairbanks to obtain the below canopy Sodankylä wind speed.

The correlation between the snow density and the surface wind speed is shown in Figure 3.1 (left panel). In this figure, the surface wind speed is the seasonal one. The snow density is the average of all density data over a season, as in the case of Fairbanks, or over the campaign duration, which ranges from 2.5 months (Summit) to 3 weeks (Finland). These data lead to the following relationship :

$$\rho_s = 44.6\bar{w} + 174 \text{ [kg m}^{-3}\text{]} \quad (3.10)$$

The associated correlation coefficient $R^2=0.63$, based on seven samples, is statistically significant ($p \leq 0.05$). It can be enhanced by adding a temperature gradient component. The rationale is that the temperature gradient tends to reduce the density in seasonal snowpacks, where the warmer basal layers sublimate and lose mass while not compacting (Sturm and Benson, 1997; Taillandier et al., 2006). However, computing the seasonal mean temperature gradient in the snowpack significantly increases the numerical memory requirements of the model, so only the wind speed correlation was actually retained. Practically in the model, at a given date, only the past winds during the season are considered, which may lead to relatively high wind velocities in early season (when the average is over a few days only). In those cases, a threshold value ensures snow density never becomes unrealistically high. The correlation between the snow thermal conductivity and the surface wind speed is illustrated in Figure 3.1 (right panel), and yielded the following regression:

$$k_{app} = 0.0424\bar{w} + 0.0295 \text{ [W m}^{-1} \text{ K}^{-1}\text{]} \quad (3.11)$$

Based on five points, $R^2=0.88$ ($p \leq 0.05$). It can be enhanced to 0.91 ($p \leq 0.05$) by adding a temperature gradient component but for the same reasons as above, this was not implemented in the model. Second order polynomial fits were also applied to these data but were not statistically significant for snow density

as a function of wind speed, so they were not retained. Note that, for all relationships, snow is not considered as saline while, on sea ice, it is quite frequently due to infiltrations of seawater at the snow/ice interface. Salinity may therefore alter those formulations.

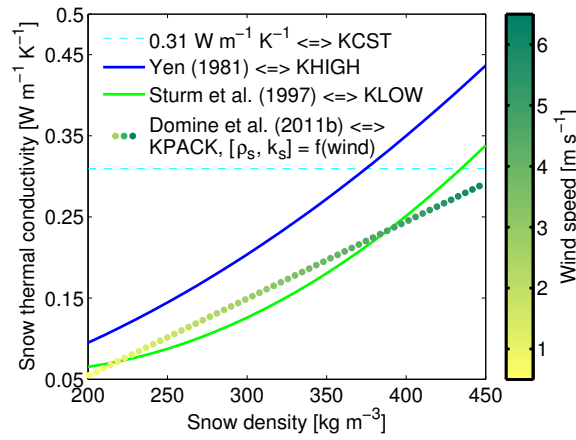


Figure 3.2: Snow thermal conductivity versus density for Yen (1981) and Sturm (1997) relationships. Colored dots of coordinates (ρ, k_s) represent snow thermal conductivity and density as functions of wind speed from Domine (2011).

3.2.2.2.3 Basic intercomparison The constant value of $0.31 \text{ W m}^{-1} \text{ K}^{-1}$ usually used for k_{app} in sea ice models was historically taken from the formulation of Abels (1892) for snow of density equal to 330 kg m^{-3} . As depicted in Figure 3.2, this value is greater than the thermal conductivities produced by all other relationships introduced above for low to medium densities, and especially for light snow. Equation (3.9) provides lower thermal conductivities than (3.8), because Sturm et al. (1997) studied depth hoar of subarctic taiga in large proportion. Since this snow type has the lowest k_{app} , this artificially decreased their values (Domine et al., 2011b). For this reason, the model simulations associated to formulations (3.8) and (3.9) in section 3.2. are named KHIGH and KLOW, respectively. The new regressions ((3.10) and (3.11)) lead to thermal conductivity values in between those of (3.8) and (3.9), for the usual range of snow densities in the model. Because regression 3.10 represents a simple parameterization of snow packing, the corresponding simulation is called KPACK in the following sections.

Simulation name	Parameterization used for snow density and thermal conductivity
KPACK	Density and effective thermal conductivity of Domine et al. (2011a), eq. (3.10) and (3.11).
KCST	Prescribed density (as detailed in section 2.2.2.1.). Constant thermal conductivity = $0.31 \text{ W m}^{-1} \text{ K}^{-1}$.
KHIGH	Prescribed density (as detailed in section 2.2.2.1.). Effective thermal conductivity of Yen (1981).
KLOW	Prescribed density (as detailed in section 2.2.2.1.). Effective thermal conductivity of Sturm et al. (1997).

Table 3.1: List of the simulations performed with the model, using the new snow scheme.

3.3 Model forcing and experimental design

3.3.1 Forcing

A combination of atmospheric reanalyses and climatologies are used to drive the sea ice–ocean model. We use the NCEP/NCAR daily reanalyses of 2 m air temperature and 10 m u – and v –wind components (Kalnay et al., 1996), and monthly climatologies of relative humidity (Trenberth et al., 1989), total cloudiness (Berliand and Strokina, 1980) and precipitation (Large and Yeager, 2004). River runoff rates are derived from Dai and Trenberth (2002).

All forcing fields are spatially interpolated from a $2^\circ \times 2^\circ$ latitude-longitude grid to the ORCA1 grid. Surface radiative and turbulent heat fluxes are calculated following Goosse (1997). The surface wind stress over sea ice is computed with a quadratic bulk formula assuming a drag coefficient $C_d = 1.40 \times 10^{-3}$.

3.3.2 Model setup and simulations

All simulations are run from 1948 to 2007, but the analysis and the comparison of model output with observations are made over the 1979–2007 period, for which satellite data are available. Initial ocean temperature and salinity fields are taken from Levitus (1998). Wherever the sea surface temperature is below 0°C , sea ice thicknesses (snow depths) of 3.5 m (0.3 m) and 1 m (0.1 m) are prescribed in the Arctic and the Southern Oceans, respectively. Ice concentration is initialized to 0.95 and 0.90 in the ice-covered regions of the Northern and Southern Hemispheres, respectively. Initial sea ice and snow temperatures are fixed to 270 K, and sea ice salinity to 6 PSU. The ocean model time step is $\Delta t_0 = 1$ hour and the sea ice model is called every 6 ocean time steps.

From the thermal conductivity and density formulations described in section 3.2.2.2.2., four simulations, hereafter referred to as KPACK, KCST, KHIGH and KLOW, were performed. Table 3.1 explicits which parameterization was used in each run. The first simulation (KPACK) was carried out using equations (3.10) and (3.11). KCST was run utilizing a constant snow thermal conductivity (equal to $0.31 \text{ W m}^{-1} \text{ K}^{-1}$) for each snow layer. KHIGH and KLOW were conducted using relationships (3.8) and (3.9), respectively, and prescribing the snow density profile as described in section 3.2.2.2.1.

3.4 Observations

In the next sections, we assess the skill of all model simulations in both hemispheres by comparing their sea ice thicknesses, concentrations and extent to observations coming from different sources.

Two sets of observational data are used to validate the modeled ice thickness. In the Northern Hemisphere, draft data (defined as the ice thickness below sea level) measured by Upward Looking Sonars (ULS) onboard submarines are available from 1979 to 2000. These data, described in Rothrock et al. (2008), are provided by the NSIDC (National Snow and Ice Data Center). They include mean drafts from more than 3000 50 km-long transects and about 30 cruises. According to Rothrock et al. (2008), the ULS observational errors have a standard deviation of 0.38 m with no significant bias. In the Southern Ocean, we use the ASPeCt (Antarctic Sea Ice Processes and Climate) data of Worby et al. (2008a), which consists of about 14,000 estimates of sea ice thickness over the 1983-2000 period. Sea ice thicknesses from this dataset are observed from ships when they break the ice and turn it sideways. Errors in the ASPeCt thickness range from $\pm 20\%$ to $\pm 30\%$ for undeformed and deformed ice, respectively (Worby et al., 2008a; Kurtz and Markus, 2012). Note that, here, PIOMAS reanalyses (Zhang and Rothrock, 2003) could have been used in order to evaluate our simulated sea ice volumes. However, we believe that in spite of their uncertainties, ASPeCt observations provide a baseline for comparison that is at least as accurate as PIOMAS model estimates, if not better.

For ice concentration, we use in both hemispheres the data from the Scanning Multichannel Microwave Radiometer (SMMR) and the Special Sensor Microwave/Imager (SSM/I) reprocessed by the EUMETSAT Ocean and Sea Ice Satellite Application Facility OSISAF (2010), from 1983 to 2007. The data were interpolated onto the model grid in order to perform the pointwise comparison with the simulated sea ice concentrations. Sea ice extent is then computed based on monthly fields of ice concentration from both model runs and

satellite data, as the total area of oceanic grid cells covered by more than 15% of sea ice.

Finally, we qualitatively compare the model snow depth distributions to those obtained by Kwok et al. (2011) and Kurtz and Farrell (2011), from large-scale airborne surveys over the Arctic basin, and by Worby et al. (2008a), who evaluated the EOS Aqua Advanced Microwave Scanning Radiometer (AMSR-E) snow depth product in the Southern Ocean.

3.5 Results

In the following, we discuss the model ability to simulate the mean state and variability of the Arctic and Antarctic sea ice covers over the 1979-2007 period. All deviations between observed and simulated quantities are expressed in terms of absolute value of the difference model - observation. Absolute values were chosen to avoid compensation of deviations with opposite signs, and resulting in misleading small values. A summary of these errors is given in table 3.2.

Table 3.2: Mean absolute value of the difference between observed and simulated variables. For draft and thickness errors, the average of the difference mod. - obs. (in absolute value) is computed over all observation locations. For sea ice extent, the mean annual cycle over 1979-2007 is computed for the model and the observations, and the annual mean of the difference between the two (in absolute value) is calculated. Observed and simulated monthly extent anomalies are retrieved with respect to the mean annual cycles of sea ice extent, in order to compute their trends and standard deviations over 1979-2007. The same method (absolute value of the difference mod. - obs.) is then applied to the latter quantities to get the errors. The best statistics for each diagnostic are enhanced using bold font.

		KPACK	KCST	KHIGH	KLOW
Mean thickness error [m]	NH (Draft)	0.63	1.08	0.64	1.55
	SH (Thickness)	0.36	0.36	0.35	0.34
Mean extent error [$\times 10^6 \text{ km}^2$]	NH	0.35	0.44	0.40	1.00
	SH	1.01	1.06	0.92	0.72
Ext. Ano. trend error [$\times 10^6 \text{ km}^2 / \text{year}$]	NH	0.027	0.032	0.030	0.013
	SH	0.012	0.013	0.008	0.005
Ext. Ano. std error [$\times 10^6 \text{ km}^2$]	NH	0.16	0.21	0.19	0.005
	SH	0.14	0.16	0.12	0.07

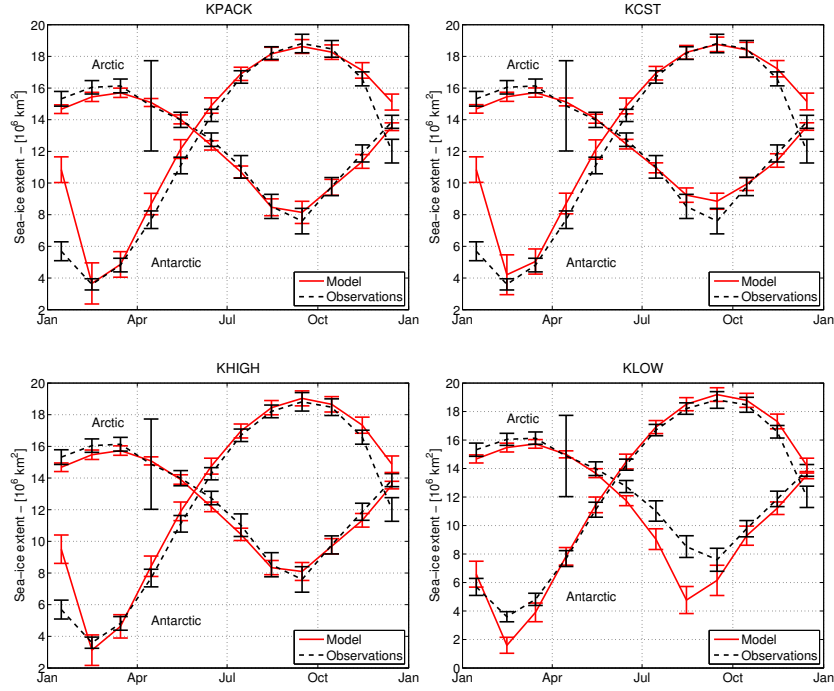


Figure 3.3: Simulated and observed (EUMETSAT OSISAF, 2010) mean seasonal cycle of sea ice extent over the period 1979-2007. Extents are calculated as the total area of oceanic grid cells with ice concentration larger than 15%. The error bars denote the standard deviation of monthly extents during the period of analysis.

3.5.1 Mean state

We invite the reader to refer to Figures 3.3 and 3.4 together with Table 3.2. Figure 3.3 shows the mean seasonal cycles of sea ice extent in both hemispheres, for all simulations and as observed. Figure 3.4 displays the differences between the simulated and observed sea ice drafts in the Northern Hemisphere at places where observations are available. Red (blue) areas show where the model overestimates (underestimates) the sea ice draft or thickness. The equivalent maps are not shown for the Southern Hemisphere for reasons that are given in section 3.5.2.2. The mean snow state is illustrated in Figure 3.5, which shows the April mean Snow Depth Distribution (SDD) averaged over the last decade of each run on both Arctic FYI and MYI (we use the category-averaged snow depths to compute them), while Figure 3.6 depicts the Antarctic annual mean snow depths over sea ice for each simulation.

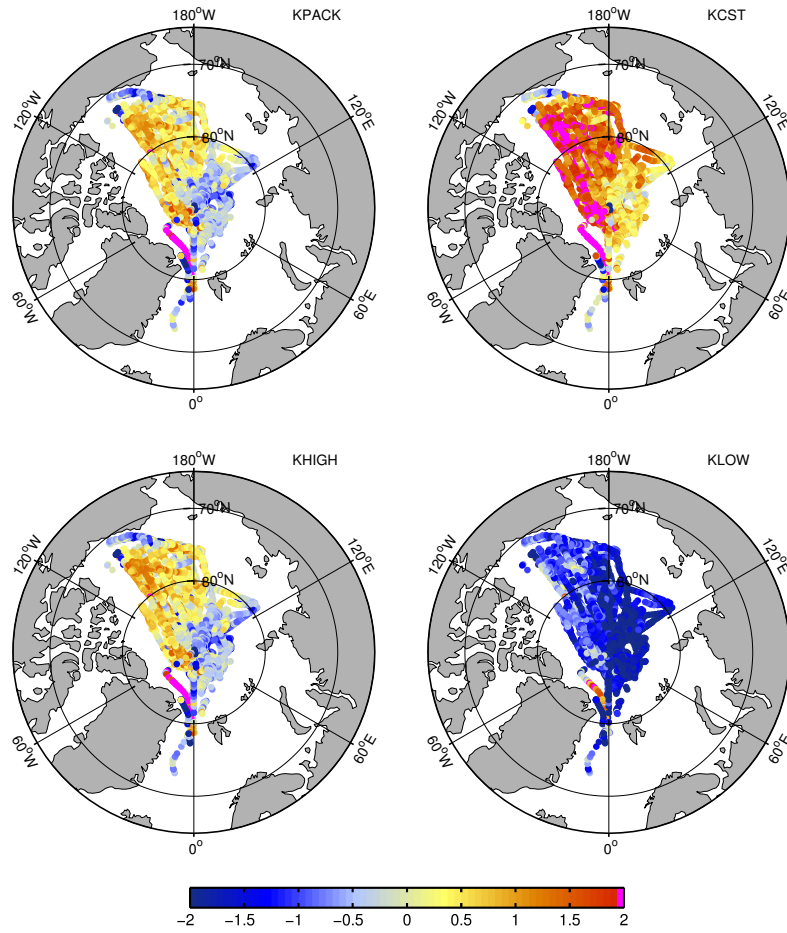


Figure 3.4: Differences (in meters) between simulated and ULS (National Snow and Ice Data Center, 1998, updated 2006) sea ice drafts in the Northern Hemisphere (mod. - obs.). Modeled drafts have been chosen according to the month and year of the observation. The corresponding grid cell has been chosen as the closest to the coordinates of the observation.

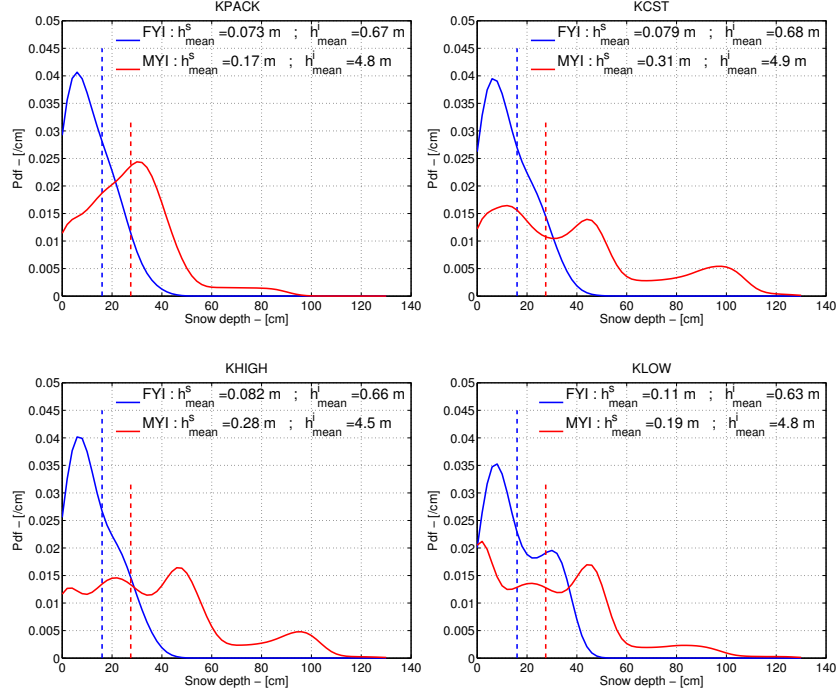


Figure 3.5: Arctic mean snow depth probability density functions (April 1997 - 2007) on FYI and MYI for all model simulations. Category-averaged snow depths are used to compute the distributions. Vertical dashed lines represent the modal snow depth from Kwok et. al. (2011), for FYI (blue) and MYI (red).

3.5.1.1 Northern Hemisphere (NH)

As shown by Figures 3.3 and 3.4, KPACK and KHIGH both provide a realistic mean sea ice state in the NH. For KPACK, the sea ice extent annual mean deviation and the draft mean absolute error with respect to observations are $0.35 \times 10^6 \text{ km}^2$ (2.8%) and 0.63 m, respectively. The mean draft in this simulation is overestimated by 2.3%. With such errors, typical for NEMO-LIM3 in the ORCA1 configuration (Massonnet et al., 2011), both globally and regionally, KPACK exhibits the best skill in simulating the Arctic sea ice extent and thickness among all simulations. KHIGH has slightly larger sea ice extent and draft mean errors ($0.4 \times 10^6 \text{ km}^2$ and 0.64 m). KCST and KLOW overestimate and underestimate the mean sea ice draft, with relative errors of +34.5% and -53.3%, respectively. KLOW presents by far the largest ice extent absolute differences with observations, with a mean value reaching $1.0 \times 10^6 \text{ km}^2$ ($\sim 10\%$).

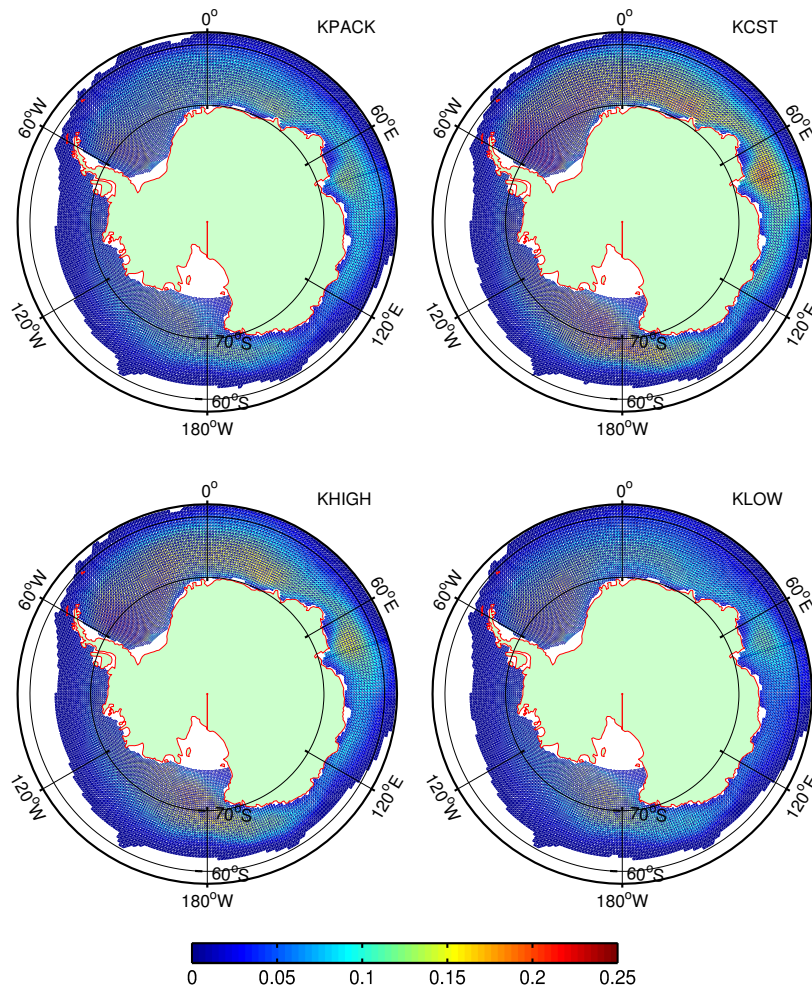


Figure 3.6: Antarctic annual mean snow depths (1997 - 2007) for all model simulations (in meters).

All model simulations do reasonably well in reproducing the observed SDD in the Arctic. In all cases, the snow depth on FYI ranges from 0 to 45 cm, with a peak at 10 cm (instead of 15 cm in Kwok et al. (2011)) in the SDD. On MYI, the maximum snow depth is about 120 cm for KCST, KHIGH and KLOW (peak in SDD around 45-50 cm), while it is only 100 cm for KPACK (peak at 30 cm). With mean snow depths (over 1997-2007) varying from 7 cm (15 cm) to 17 cm (31 cm) on FYI (MYI), the snow pack appears globally too thin but the shape and width of the SDDs are relatively well captured given the possible errors in the precipitation climatology used to force the model, especially in KPACK.

3.5.1.2 Southern Hemisphere (SH)

As revealed by Table 3.2, the model sea ice thickness is less sensitive to the snow thermal conductivity in the SH than in the NH. A similar mean error of about 35 cm is found for all runs, which explains why maps equivalent to those displayed in Figure 3.4 are not shown for the SH. In contrast, Figure 3.3 indicates that the SH sea ice extent is relatively sensitive to the snow thermal conductivity formulation. While KLOW is relatively poor in the Arctic, it performs better in the SH with respect to ice extent (error of 0.72×10^6 km², i.e., 10.5%) than the other runs ($\sim 1.0 \times 10^6$ km²). The model tends to underestimate the SH mean sea ice thickness by $\sim 30\%$, compared to the AS-PeCt ice thickness product. In Fichefet et al. (2000), a negative ice-ocean feedback was detected when reducing the snow thermal conductivity, with less ice growth and therefore less brine rejection, in turn resulting in weaker oceanic heat fluxes to the ice due to the stronger ocean stratification. In their study, the latter process triggered an enhanced sea ice thermodynamic production competing with, and sometimes compensating, the initial drop in sea ice growth rate. Such a feedback is not clearly observed here, probably because some of the model components are substantially different. In particular, NEMO-LIM3 includes the sea ice thickness distributions, a prognostic salinity, a treatment of the vertical mixing in the ocean different from the one in the ocean model of Fichefet et al. (2000), and was used at a higher resolution. In addition, our snow scheme is multi-layered, which makes it more responsive to the surface thermodynamic forcing compared to its former version. However, it may also make the whole sea ice/snow system less sensitive to variations in conductive heat fluxes through the snow, driven by air-temperatures, since this variability is more efficiently contained by the multilayer snow scheme. This might be an other reason for obtaining results that are in part different from those of Fichefet et al. (2000).

A common problem of all model runs is their inability to reproduce the fast decrease in extent in Nov.-Dec. (see Figure 3.3). This typical feature for the NEMO-LIM3 ORCA1 configuration of the model was already observed in Massonnet et al. (2011) and is still unexplained.

As in the NH, although the observed geographical distribution of snow depth is reasonably well reproduced, in particular with regards to the ASPeCt snow data, the model tends to underestimate the snow depths compared to the Advanced Microwave Scanning Radiometer (AMSR-E) snow depth product (Worby et al., 2008b) (not shown). The maximum snow depth is found in the Weddell Sea and reaches 25 cm in KCST, against 19 cm only in KPACK.

3.5.2 Variability

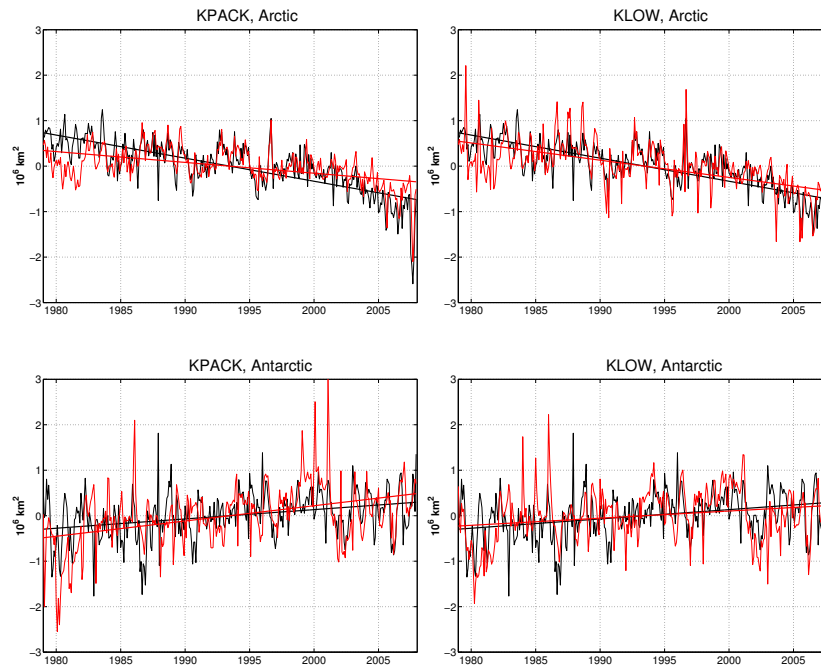


Figure 3.7: Simulated (red) and observed (black, EUMETSAT OSISAF, 2010) monthly anomalies in sea ice extent (relative to the mean seasonal cycle of sea ice extent) in the NH (top panels) and SH (bottom panels) over the period 1979-2007 for the KPACK and KLOW simulations. The plain lines indicate the trends computed from linear regression over the same period.

Surprisingly, the simulations yielding the best estimates for mean state are not necessarily those providing the ice extent anomalies in best agreement with observations. Figure 3.7 shows the time series and trends of the monthly anomalies in sea ice extent in both hemispheres through 1979-2007 for KPACK and KLOW. Because they are qualitatively similar, we only show these plots for KPACK, and compiled the statistics for the other runs in Table 3.2. Both Figure 3.7 and Table 3.2 reveal that KLOW systematically outperforms all other runs in reproducing the observed sea ice extent anomalies, with relative errors of -25.3% (-23.1%) for the trend in extent anomaly and +0.8% (+13.3%) for the standard deviation in extent anomaly in the NH (SH).

3.6 Discussion

3.6.1 Mean state

As shown in section 5.1, KCST overestimates the Arctic sea ice volume. As explained in section 2.2.2.3., the snow thermal conductivity in KCST seems to be overestimated compared with other formulations, particularly for snow on FYI that has smaller densities in the model. The result is a strongly enhanced FYI growth rate in winter. This feature is also observed in the Southern Ocean, although less significantly. As clearly shown by Figure 3.8, presenting the seasonal cycle of the sea ice mass balance spatially integrated terms, the surface energy budget in the Southern Ocean is such that it rarely leads to surface melt. Hence, the sea ice mass balance in the South is rather driven by snow ice formation, and formation and melting at the base. Therefore, the relative importance of atmospheric processes is likely smaller than in the Arctic. Compared to KCST or KPACK, KLOW produces more insulating snow on top of sea ice that curtails the heat loss from the ice to the atmosphere and thus limits the basal ice growth in winter (see Figure 3.8). FYI therefore covers a larger part of the Arctic Ocean during wintertime, in turn leading to a significantly reduced sea ice summer extent. Among all simulations, KHIGH leads to the most realistic sea ice thicknesses with a snow thermal conductivity relationship naturally leading to values widely varying between those produced by KLOW and KCST.

The Arctic sea ice draft simulated by the KPACK and KLOW runs differ substantially from each other, as depicted in Figure 3.4. Yet the snow thermal conductivities for these two parameterizations look similar (Figure 3.2). To explain this apparent contradiction, we put forward two reasons. First, the snow thermal conductivity in KPACK is a function of the wind speed only, and the winds from the atmospheric forcing are large ($\sim 5\text{-}10 \text{ m s}^{-1}$). For this simulation, k_{app} reaches values at the upper tail of the distribution, say ~ 0.3

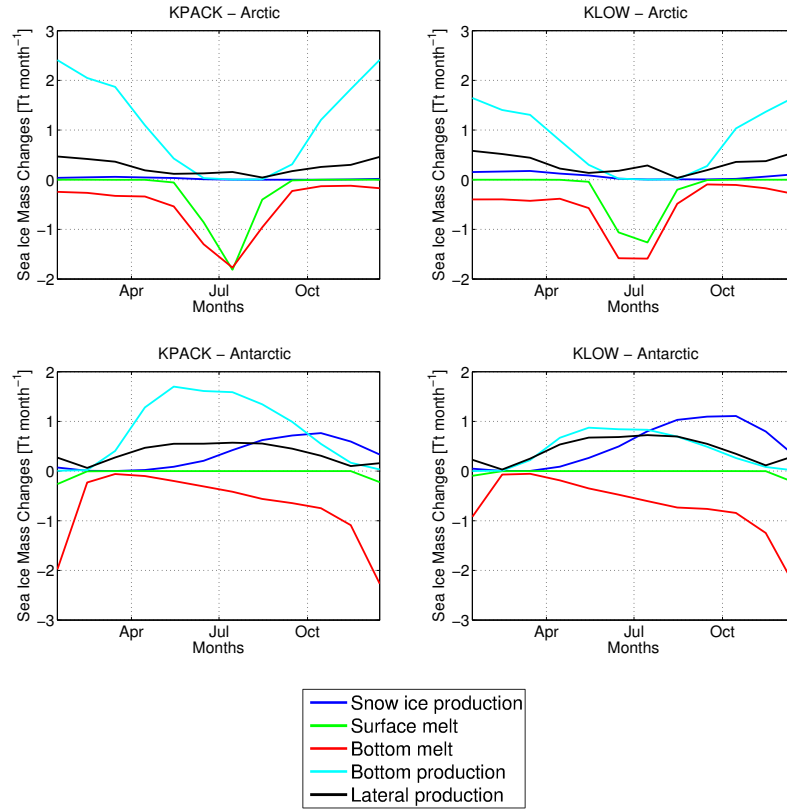


Figure 3.8: Spatially integrated terms from the simulated mean sea ice mass balance (over 1979-2007), for KPAC and KLOW. Positive (negative) diagnostics represent sea ice mass source (sink) terms. Snow ice formation corresponds to the surface production of sea ice whenever seawater floods the snow - sea ice interface and refreezes. Lateral production depicts the mass of sea ice forming in leads. Bottom production/melt and surface melt are the integrated diagnostics for the basal thermodynamic accretion/ablation of sea ice and the surface melt due to positive imbalance in the surface energy budget, respectively. There is no explicit lateral melting in the model.

$\text{W m}^{-1} \text{K}^{-1}$. In the KLOW run, k_{app} is a function of snow density only ($\sim 300 \text{ kg m}^{-3}$ in the model), thus reaching lower values and limiting ice growth in winter. Second, snow density is also a function of wind speed in the KPACK simulation, and for the same reasons it is higher on average than the KLOW density. For the same snow mass, the snow cover will accordingly be thinner in KPACK (see also top left panel of Figure 3.5), thus enhancing heat transfers compared to KLOW. These two effects add up to produce a thicker ice in KPACK than in KLOW, and explain the significant difference in sea ice bottom production between the two simulations (Figure 3.8).

In all simulations and in both hemispheres, the sea ice thickness and extent are also considerably sensitive to the SDDs. Although qualitatively those distributions seem to be rather well reproduced, quantitatively the observed SDDs of Kurtz and Farrell (2011) or Kwok et al. (2011) as well as the geographical distribution of mean snow depth in the Southern Ocean are not captured. Errors can be attributed to the forcing or to the model itself. Regarding the forcing, using a climatology for precipitation involves the loss of interannual variability. Besides, the atmospheric reanalyses and climatologies used to force the model have known biases (e.g., Bromwich et al., 2007; Vancoppenolle et al., 2011). In particular, a bias in snow precipitation is likely to affect the simulated SDDs by shifting them towards thicker or thinner snow, thus improving or worsening the quantitative comparison with observed SDDs, respectively. Interestingly, KPACK provides the best distribution shapes compared to the observational studies mentioned previously. This simulation uses regression (3.8), which emphasizes the importance of accounting for the direct influence of wind on snow depth, and ultimately suggests that developing a representation of wind-forced snow redistribution in the model would significantly impact on the results and may be a lever for improving the simulated SDDs. We therefore recommend this kind of parameterization for snow density and thermal conductivity in sea ice models designed for climate studies.

The mean snow depths on Arctic MYI (Figure 3.5) also indicate that runs with the smallest (largest) mean snow thermal conductivities produce the smallest (largest) snow depths. The reason is that a smaller snow thermal conductivity induces less heat carried into the inner snow - sea ice pack, therefore leading to an earlier onset of surface melt at the end of spring and a more intense thinning all along the snow melt period, and conversely. This also explains the early ice extent minimum of KLOW, which is shifted from September to August.

Similar observations can be made in the SH, although for different reasons since surface melt conditions are rarely reached there. As sea ice is thinner in KLOW (due to the limited bottom growth during winter), more snow ice production occurs due to more frequent negative freeboards and seawater flood-

ing the snow-ice interface. This process, virtually non-existent in the NH (see Figure 3.8 top panels), enhances the snow thinning but counterbalances the sea ice thermodynamic thinning. The latter mechanism partly explains why, in the SH, the differences in ice thickness between all runs are still small (as suggested by Table 3.2) relatively to the differences in mean snow thermal conductivity through the various formulations that are used. Furthermore, even though KLOW exhibits the smallest absolute deviations with regard to sea ice extent observations, it is the simulation with the smallest mean sea ice thickness and volume (not shown). The reason is that with a smaller volume and minimum sea ice extent (see Figure 3.3), the larger area fraction of leads in late summer enables a more intense lateral growth of sea ice all along the following winter (Figure 3.8), hence compensating for the initial underestimation in the seasonal sea ice extent. Finally, the significantly less pronounced sensitivity of the SH sea ice thickness to the snow thermal conductivity representation is consistent with the findings of Massonnet et al. (2011). They showed in particular that the NEMO-LIM3 SH ice concentration, extent and thickness are less sensitive to the model physics compared to their NH counterparts, in the ORCA1 configuration. This result is due to the crucial importance of the oceanic heat supply in the Antarctic sea ice mass balance. The average value of this flux is larger in the SH, with typical values around $\sim 15\text{--}20 \text{ W m}^{-2}$ (Heil et al., 1996), against $\sim 3 \text{ W m}^{-2}$ in the NH (Krishfield and Perovich, 2005). Such values are also observed in the model. Because of this, sea ice responds more significantly to oceanic anomalies than to atmospheric ones, as already mentioned by Maykut and Untersteiner (1971).

3.6.2 Variability

As shown by the model statistics, KLOW provides the smallest errors for the monthly sea ice extent anomalies with respect to observations in both hemispheres. However, only the 2-m air temperature and 10-m u-v winds bear an interannual component in our setup, while the other forcings (humidity, cloudiness and precipitation) are climatological. Therefore, the full potential impacts of interannual variability and trends of the actual atmospheric state on the sea ice cover is only partially accounted for in our simulations. As an example, a significant trend in the Southern Ocean's precipitation in the context of a warming climate, as in Zhang (2007), would likely affect the sea ice extent anomalies and trends over the simulation period. Additionally, the restoring of the sea surface salinities towards the Levitus (1998) climatology further strengthens this problem. This in turn suggests limitations in the conclusions that can be drawn from our analysis of statistics in Table 3.2.

KLOW simulation yields a smaller sea ice volume and a MYI areal coverage accordingly smaller, resulting in two consequences. First, the ice has a lower mass and responds faster to the dynamic forcing from the ocean and the atmosphere. Second, larger direct inputs of solar radiation are allowed into the ocean during the summertime drop in FYI concentration, affecting the heat budget in leads and the oceanic turbulent heat flux at the ice base. A positive ice-ocean feedback is then triggered as the bottom melt of the remaining ice is enhanced, subserving further decreases in ice concentration. Being aware of the aforementioned issues, the seemingly better ice extent variability in KLOW may therefore be attributed to larger oceanic interactions occurring during the simulation, while the impacts of the climatological forcing (clearly lacking of atmospheric feedbacks) are lessened. Nonetheless, this is to the detriment of the modeled sea ice mean volume, which seems to be remarkably underestimated, especially in the NH.

3.7 Conclusion

A new snow thermodynamic scheme was embedded into the sea ice model LIM3 and the sensitivity of the global coupled ice-ocean model NEMO-LIM3 to the representation of snow thermal conductivity was assessed through four runs (at a 1 degree horizontal resolution) using different snow thermal conductivity formulations. These include a constant thermal conductivity (simulation KCST), relationships of Yen (1981) (KHIGH), Sturm et al. (1997) (KLOW) and a new one, based on Domine et al. (2011a) (KPACK), which is specifically designed for climate studies. The first result of this study is that the $0.31 \text{ W m}^{-1} \text{ K}^{-1}$ value for snow thermal conductivity, which is commonly used in large-scale sea ice models with a single-layer representation of snow, cannot be used in a multilayer snow scheme with varying density. This value, which seems too large, except for very dense snow, leads to an overestimation of the Arctic sea ice thickness, particularly in a multilayer scheme that is naturally more sensitive to the thermodynamic surface forcing than a single layer with constant thermo-physical properties. None of these simulations provides good results in terms of all snow and ice variables and in both hemispheres simultaneously, which stresses the specific response of each hemisphere to varying snow physics. In the Arctic, the KPACK and KHIGH simulations give a reasonable estimate of the mean sea ice extent (absolute error between model and observations of $0.35 \times 10^6 \text{ km}^2$ for KPACK, i.e. 2.8%) and thickness (absolute error of 0.63 m, KPACK) although the observed monthly sea ice extent anomalies over 1979-2007 are better reproduced with KLOW. In the Southern Ocean, KLOW leads to the minimum absolute errors, with mean values of $0.72 \times 10^6 \text{ km}^2$ (10.5%) and 0.34 m ($\sim 30\%$) for ice extent and thickness, respec-

tively. However, these better statistics may be due to compensating errors. Like in the NH, the sea ice basal thermodynamic growth is clearly curtailed by the low snow thermal conductivity values in the simulation, but it is partly compensated by both lateral production and snow ice formation maintaining the hydrostatic equilibrium of the snow - ice column. In the Arctic, low snow-fall and thicker ice virtually prevent snow ice formation. KLOW also seems to feature a better sea ice variability compared to all other simulations, potentially allowing the oceanic feedbacks to play a more prominent role in driving the geographical distributions and temporal evolution of the ice concentration through the oceanic heat flux at the ice base. The new snow thermal conductivity formulation used in KPACK seems the only one reasonably well suited for both hemispheres even though the mean errors with respect to observed sea ice extent ($1.01 \cdot 10^6 \text{ km}^2$ (13.0%)) and thickness (0.37 m) are still quite large in the Southern Ocean, which could admittedly be due to other sources of errors in the simulation. The most important peculiarity of KPACK is its ability to reproduce the observed shape of the snow depth distributions, thanks to relationships simply translating the effects the snow packing by wind, and indirectly influencing the snow depth. The importance of the snow depth distributions in the model physics ultimately suggests that accounting for snow redistribution by the wind would presumably impact positively on the model performance. Finally, because of the lack of atmospheric feedbacks and the intrinsic limitations of the forcing, due to potential biases in the reanalyses and climatologies, those findings will have to be investigated using fully coupled configurations of the model, once available. This contribution stresses the need for further understanding of snow processes on sea ice and of efforts towards better representation in large-scale models.

Acknowledgments

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WIND-BLOWN SNOW REDISTRIBUTION AND MELT PONDS

This Chapter is based on the following paper:

Lecomte, O., Fichefet, T., Flocco, D., Schroeder, D., Vancoppenolle, M., 2014a. Impacts of wind-blown snow redistribution on melt ponds in a coupled ocean–sea ice model. Submitted. Ocean Modelling.

Abstract

Snow depth conditions the shape of melt ponds in their early stages of formation on Arctic sea ice. So as to study this mechanism, an explicit melt pond scheme is introduced into a coupled ocean–sea ice model including new snow thermophysics and a wind-blown snow redistribution parameterization. The model comparison against observations demonstrates realistic melt pond statistics on average over Arctic sea ice, but a clear underestimation of the pond coverage on the multi-year ice (MYI) of the western Arctic Ocean is noticed. The latter shortcoming originates from the concealing effect of persistent snow on forming ponds, impeding their growth. Analyzing the influence of intensified snow drift in simulations enables the identification of two distinct modes of sensitivity in

the melt pond formation process. First, the larger proportion of wind-transported snow that is lost in leads directly curtails the late spring snow volume on sea ice and facilitates the early development of melt ponds on MYI. In contrast, a combination of higher air temperatures and thinner snow prior to the onset of melting sometimes make the snow cover switch to a regime where it melts entirely and rapidly. In the latter situation, seemingly more frequent on first-year ice (FYI), a smaller snow volume directly relates to a reduced melt pond cover. Notwithstanding, changes in snow and water accumulation on seasonal sea ice is naturally limited, which lessens the impacts of wind-blown snow redistribution on FYI, as compared to those on MYI. At the basin scale, the overall increased melt pond cover results in decreased ice volume via the ice-albedo feedback in summer, which is experienced almost exclusively by MYI.

4.1 Introduction

Soon after the initiation of the summer surface melt on Arctic sea ice, melt-water starts accumulating in pools called melt ponds that usually cover up to 50-60% of the sea ice area. The processes driving the formation and evolution of those melt ponds are well documented (e.g., Fetterer and Untersteiner, 1998; Perovich et al., 2002; Polashenski et al., 2012). The most important consequence of the pond formation with respect to the sea ice energy and mass balance is the critical drop in surface albedo wherever ponds form, triggering further ice surface and basal melt through the ice-albedo feedback. The sea ice albedo may afterwards quickly come back to high values as the melt pond water drains through permeable ice (Perovich et al., 2002) or starts refreezing at the surface. A partly refrozen melt pond then constitutes a suitable platform for the snow to reaccumulate, ultimately affecting the albedo and the temperature profile of the ice underneath (Flocco et al., 2014). The crucial role of melt ponds in controlling the evolution of the sea ice albedo thus make them a key component of the polar climate system.

Serious efforts have already been invested in the representation of melt ponds in large-scale sea ice models (e.g., Pedersen et al., 2009; Flocco et al., 2010, 2012; Holland et al., 2012; Hunke et al., 2013). Those studies evidence for the models sensitivity to the representation of melt ponds and show that actually accounting for their influence on the ice-albedo feedback leads to consequential sea ice volume reductions. Melt ponds have also been shown to influence the trends of the current Arctic sea ice decline significantly (Roegner et al., 2012). Even so, those models are not comprehensive yet and still lack a few of the processes driving the formation and evolution of melt ponds. Among them are snow-related processes. The late-winter sea ice snow cover persisting into the early part of the melt season is one of the controlling factors for melt pond distributions (Petrich et al., 2012), for two main reasons. First, the snow volume on sea ice (in snow water equivalent) defines in part the fresh water volume directly available to feed the ponds when snow melts and, second, new ponds in formation may remain hidden by the snow as long as it has not melted entirely. Hence, the spatial distribution of snow alters the initial distribution of melt ponds by affecting where and when they become visible first. Subsequently, the low albedo of ponds as compared to snow or bare ice causes them to preferentially absorb solar radiation, which further affects the surface ice melt locally. Snow depths on sea ice thus have both short term and longer term indirect impacts on the evolution of melt ponds. This is all the more important as the depth of the sea ice snow cover is extremely heterogeneous horizontally (e.g., Sturm et al., 2002), mostly due to blowing snow effects and to a frequently uneven sea ice surface topography. Depend-

ing on the nature of an ice floe, whether it is multi-year or seasonal, level or deformed, the wind tends to draw manifold snow drift features at its surface, such as dunes, sastrugi or accumulation patterns under the lee of sea ice pressure ridges (Sturm and Massom, 2009). In light of those elements, the question of the indirect influence of blowing snow on melt pond formation through the reshaping of the snow cover becomes legitimate, and may lead to different answers for different ice types.

In the present paper, we therefore aim at (1) understanding the large-scale implications of snow accumulation and redistribution by winds on the formation and evolution of melt ponds and (2) assess the similarity or dissimilarity of those impacts on first-year ice (FYI) and multi-year ice (MYI). Both because melt ponds are relatively uncommon and because their extensive observation is nonexistent in the Southern Ocean, our study focusses on the Arctic. So as to achieve this work, the explicit melt pond formalism of Flocco and Feltham (2007) was incorporated into the Louvain-la-Neuve Sea Ice Model (LIM), which is fully coupled with the ocean general circulation model NEMO-OPA (Nucleus for European Modelling of the Ocean - Ocean PARallelisé). In particular, LIM includes a snow scheme of intermediate complexity (Lecomte et al., 2013) and a newly developed parameterization of blowing snow (or snow drift) effects. The model is presented in the following section, before we proceed to the assessment and intercomparison of two forced-atmosphere configuration simulations with regard to observations (forcing and observations are detailed in sections 4.3 and 4.4) and two other melt pond modeling studies (sections 4.5 and 4.6). Section 4.7 finally summarizes the results.

4.2 Model description

4.2.1 Ocean and sea ice

NEMO-LIM (Nucleus for European Modeling of the Ocean - Louvain-la-Neuve Sea Ice Model) is a state-of-the-art global coupled ocean-sea ice model. Its ocean component is the general circulation model OPA (Ocean PARallelisé, version 9) and is fully documented in Madec (2008). The thermodynamic-dynamic sea ice model, on the other hand, is LIM3 (LIM, version 3) and is coupled to the ocean component following Goosse and Fichefet (1999). This model, comprehensively described in Vancoppenolle et al. (2009), includes an explicit representation of the subgrid-scale distributions of ice thickness, enthalpy, salinity and age. Sea ice thermodynamics are computed for each sea ice thickness category following Bitz and Lipscomb (1999) and halodynamics use empirical parameterizations for gravity drainage and percolation of brines. The elastic-viscous-plastic (EVP) rheology of Hunke and Dukowicz

(1997) in the C-grid formulation of Bouillon et al. (2009) is used to solve the sea ice dynamics. We run the model in the same configuration as in Lecomte et al. (2013), i.e., on the global tripolar ORCA1 grid of the ocean model (1 degree resolution), with 5 ice thickness categories (each of them being divided into 5 layers for sea ice halo-thermodynamics) and 42 vertical levels in the ocean. Spurious model drift in salinity is prevented by a sea surface restoring term (toward climatological values of Levitus, 1998) in the freshwater budget.

4.2.2 Snow

4.2.2.1 Thermodynamics

The thermodynamic snow scheme we use was developed in previous works (for its comprehensive description, see Lecomte et al., 2013). It is multilayer, with time and space dependent snow thermo-physical properties (i.e., density and thermal conductivity) and includes heat conduction through the snow, penetration and absorption of solar radiation in the uppermost snow layers (following Beer's law and extinction coefficient from Järvinen and Lepparanta, 2011), surface melt or sublimation based on the imbalance of the surface heat budget and a representation of snow ice formation based on Fichefet and Morales Maqueda (1997).

Two important additions were made to the snow thermodynamics as compared to Lecomte et al. (2013). They are related to the inclusion of melt ponds in the model. The first is the albedo calculation (see section 4.2.3.2) and the second is the treatment of the physical impacts of liquid water presence in snow. Indeed, benefit was taken from the new water storage variables, required for the coupling to an explicit melt pond model (see section 4.2.3), to account for wet snow properties and superimposed ice formation into the snow. Because snow gets saturated with water very quickly (Jordan et al., 2008; Sturm and Massom, 2009), we assume that the snow cover is wet whenever the liquid water reservoir is not empty (for a given ice category, in a single grid cell). As in Hunke et al. (2013), water infiltration in the snow is computed based on the amount of water available and the volumes being possibly occupied in snow, depending on its density. The mass fraction of liquid water relative to the total mass of water and snow in a layer is then used to alter the snow thermal conductivity, calculated as a weighted mean of snow and freshwater thermal conductivities. Finally, freshwater that may refreeze in case of divergence of the conductive heat fluxes going in and out of a saturated snow layer is given by:

$$-L_f \frac{dm_{r fw}}{dt} = f c_s^{in} - f c_s^{out} \quad (4.1)$$

where L_f is the latent heat of fusion, m_{rfw} the mass of refrozen freshwater, and $f c_s^{in}$ and $f c_s^{out}$ the conductive heat fluxes at the interfaces of the snow layer, respectively. The adopted formalism is similar to Cheng et al. (2006), except that the refrozen water is not treated exactly as ice. Instead, it is incorporated to the mass of snow, affecting its density and slackening the early stages of the snow melt. This process acts as a sink for the melt pond water, as long as snow remains present and water volumes are small. Note that, on the contrary, in case of convergent conductive heat fluxes in a snow layer at the melting point, snow melts internally and becomes a source for the freshwater reservoir.

4.2.2.2 Blowing snow

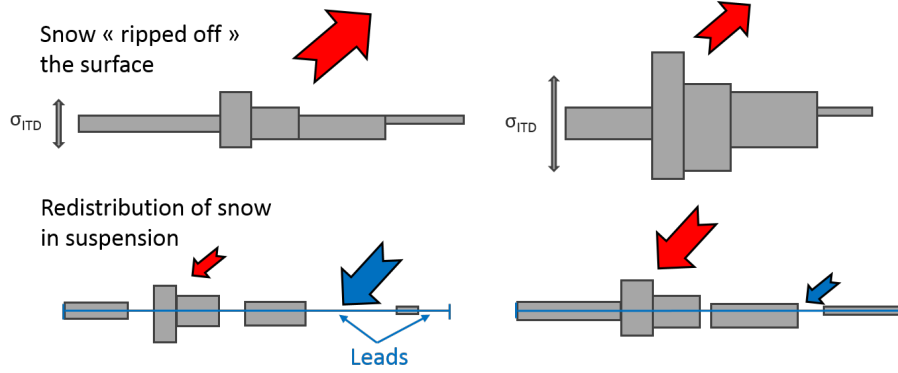


Figure 4.1: Schematic representation of the snow redistribution by winds on the subgrid-scale distribution of ice thicknesses in the model. σ_{ITD} stands for the standard deviation of the ice thicknesses (from categories) within a grid cell. Red arrows correspond to the snow mass fluxes from and to the ice during blowing snow transport and redeposition, respectively. The blue arrows represent the snow mass fluxes to the ocean through the cracks between the ice floes during redeposition (see section 4.2.2.2). Basically, flat ice (translating in small σ_{ITD}) enables stronger erosion rates, hence larger mass fluxes of snow removed from the ice when snow drift is triggered, and conversely for deformed ice. During redeposition, the fraction of the total transported snow mass that is lost in leads is proportional to the area of open water in the grid cell.

An extended version of this section, including further technical details about the blowing snow parameterization in the model, is provided in Appendix B.

Blowing snow (or snow drift) on sea ice tends to redistribute the snow mass from thin level ice to thicker and deformed ice. The process smoothes out small and high frequency changes in sea ice topography while enhancing larger topographic features such as pressure ridges (Sturm et al., 2002). Usually, the “blowing snow” term includes several sub-processes such as snow

erosion and motion by saltation or suspension (snow transport by winds). Snow drift is also a sublimation-enhancing process, facilitating the sublimation of grains thrown in suspension (D  ry and Yau, 2002). The impacts of such mechanisms are well documented (e.g., D  ry and Tremblay, 2004; King et al., 2008; Sturm and Massom, 2009; Leonard and Maksym, 2011). Most importantly, snow that is ripped off the pack and transported by winds may be lost into leads, or redeposited on the ice, altering the spatial distribution of snow depths. As the wind direction shifts or the floe rotates, snow undergoes further erosion that may affect the surface drag coefficient and the dynamic response of sea ice to the wind forcing. Given the difficulty of establishing a new parameterization due to the lack of observations available for this specific process, we only thought to represent the zero-order impact of blowing snow on the subgrid-scale snow depth distribution. To our knowledge, it is the first time such developments are made in a large-scale coupled ocean–sea ice coupled model.

The usual conservation equation for a global state variable such as the total snow mass per unit area m_s on a single ice thickness category writes:

$$\frac{\partial m_s}{\partial t} = -\nabla(m_s \mathbf{u}) + \Psi^{m_s} + \Theta^{m_s} + \Phi^{bs} \quad (4.2)$$

where $-\nabla(m_s \mathbf{u})$ represents the horizontal advection of snow due to sea ice motion, Ψ^{m_s} the effects of the mechanical redistribution of sea ice on snow, Θ^{m_s} the changes in snow mass due to the thermodynamic processes described in the previous two sections and Φ^{bs} the wind-redistribution of snow. The later term may be written as:

$$\Phi^{bs} = \Phi_{local}^{bs} + \Phi_{nonlocal}^{bs} \quad (4.3)$$

the local term accounting for redistribution of the snow mass over the sea ice thickness categories, and the non local one representing the snow transport from grid cell to grid cell. For the sake of simplification, we assume that, at the synoptic scale of a storm, the snow mass being transported out of a grid cell matches the amount of snow that is transported in and we therefore neglect the non local effect of blowing snow. The validity of this assumption may depend on the horizontal resolution of the model grid, but would require the coupling with an atmospheric circulation model in order to be properly tested. Φ^{bs} may be separated into two terms:

$$\Phi^{bs} = \Phi_T^{bs} + \Phi_R^{bs} \quad (4.4)$$

where Φ_T^{bs} and Φ_R^{bs} are the ripping out and re-deposition rates, respectively. Based on the thought, commonly shared by snow and sea ice field scientists,

that snow drift mainly depends on wind speed and sea ice surface topography, we formulate those rates in a simple way, as illustrated by Figure 4.1. We assume that the mass flux of snow removed from the pack and transported by winds is (1) negatively proportional to the snow density (packed snow is less erodable than light snow), (2) proportional to the wind speed, as long as it is greater than a threshold velocity V^* and (3) inversely proportional to the standard deviation of the ice thickness distribution σ_{ITD} (i.e., over the thickness categories) in a grid cell. V^* is computed as:

$$V^* = \frac{\rho_s - \beta}{\alpha} \quad (4.5)$$

where ρ_s is the snow density and α and β are coefficients that were retrieved from a linear regression of observations relating the seasonal mean wind speed to the actual snow density (Lecomte et al., 2013). Physically, V^* represents the seasonal mean wind velocity that is required to make the snow pack as dense as ρ_s , and it is assumed that blowing snow will start only as soon as the instantaneous wind speed exceeds this value. σ_{ITD} on the other hand is considered as a roughness length limiting the snow drift in case of large disparities in ice thicknesses (represented by a large σ_{ITD}). Given the relatively poor ice deformation diagnostics in current sea ice models, we chose this simple solution instead. A mass flux coefficient γ is used to tune the relationship between Φ_T^{bs} and the wind speed. For a same wind velocity, a higher γ induces a larger snow mass removed from the ice (and potential losses into leads), and conversely. Since this parameter is currently impossible to calibrate based on observations, it was adjusted to provide comparable results as those obtained in Lecomte et al. (2013) and set to the standard value of $10^{-5} \text{ kg m}^{-2}$. In our results below, we assess the sensitivity of our model to this new parameterization and its influence on melt pond distributions. A fraction of the total mass of snow transported by winds is then lost in leads and the rest is redistributed among ice thickness categories via Φ_R^{bs} , proportionally to their relative coverage of the total ice area. Again, this was done on the simple assumption that wind-blown snow redistribution on sea ice is largely dependent on the size of open water areas between ice floes (Leonard and Maksym, 2011).

4.2.3 Melt ponds

4.2.3.1 The model

The melt pond formulation we use is the one of Flocco and Feltham (2007) and the way it was incorporated into our sea ice model is very similar to the works of Flocco et al. (2010, 2012). The real topography of sea ice is not represented in LIM3, but from the ice thickness distribution formalism and assuming the

hydrostatic equilibrium of each ice thickness category, the melt pond model retrieves a discretized distribution of ice surface height and basal depth with respect to the sea level position. After calculating the changes in water volume retained on the ice due to the manifold source and sink processes described hereafter, the depth and fractional coverage of melt ponds on the ice can be computed by distributing the water among ice thickness categories, depending on their surface height and the volume of their snow cover.

In order to couple the sea ice and melt pond models, three new tracers were added: the melt pond volume, area and ice lid volume in case they refreeze (all three tracers are per unit area of ice). Melt pond volume is fed by snow/sea ice surface melt, snow internal melt and rain. At each time step, a fraction of this volume (depending on the total ice concentration, Holland et al., 2012) is lost by lateral run-off through cracks or ice floe edges. If the pond surface is above the sea level, the vertical drainage of meltwater through permeable sea ice into the ocean is computed following Darcy's Law, as a function of the pressure head relative to the sea level and the ice permeability. When the surface heat budget gets negative, melt ponds start refreezing and the growth rate of the ice lid forming on top of the melt pond is calculated using Stefan's law. The ice lid may undergo further growth or melt depending on the surface heat balance.

As in Flocco et al. (2012), the fraction of ice covered by ponds in each category is limited by a prescribed value that is function of its thickness. Once the melt pond area fraction, depth and ice lid thickness are calculated, they are used to retrieve the effective albedo of the pond covered ice.

4.2.3.2 Specific details

A few specific details related to the coupling of the melt pond scheme to LIM3 have to be noted. In addition to the sink processes described above, the meltwater volume in the model may also be reduced by physical infiltration and refreezing of water into the snow, as presented in section 4.2.2.1. The pond volume is also affected by sea ice advection and deformation. Fractions for melt pond and ice lid volume lost during sea ice ridging or rafting were added to the model. A few test cases showed that our model is not very sensitive to those parameters. At this stage, it is not clear whether this weak sensitivity was to be expected. We therefore prescribed the value of those fractions to 50%, as it was already the case for snow in LIM3. For the sake of clarity, Figure 4.2 summarizes the fresh water balance during a model time step.

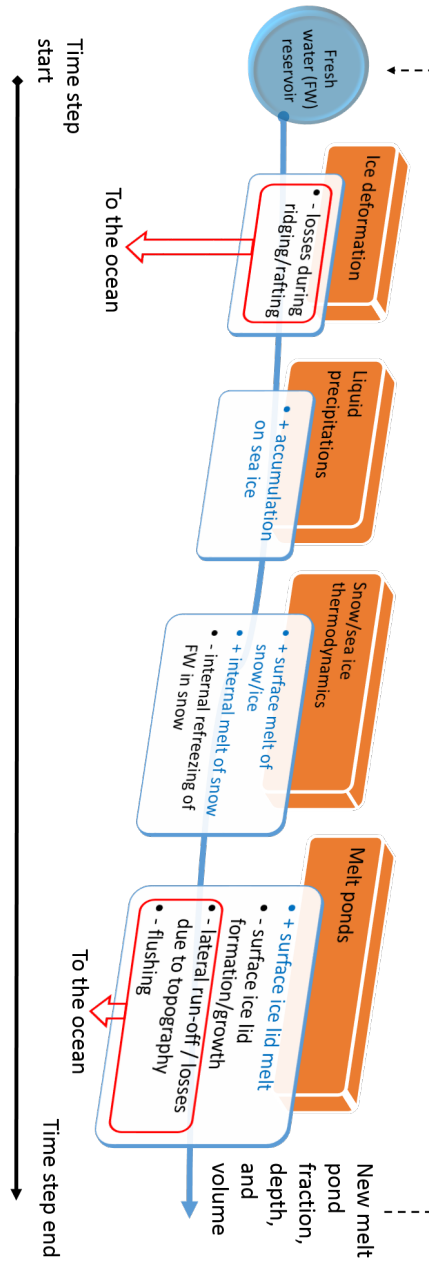


Figure 4.2: Evolution of the fresh water reservoir in a grid cell during a model time step. At the time step start, this reservoir is equal to the melt pond volume. All source and sink processes listed in this diagram are described in section 4.2.3.

In order to account for melt pond tracers in the albedo calculations, the melt pond albedo is first computed as a function of its depth, as in Lecomte et al. (2011):

$$\alpha_p = \begin{cases} \alpha_w - (\alpha_w - \alpha_i)e^{-\frac{h_p}{\omega}} & \text{if no active ice lid and active melt pond} \\ \alpha_{bi} & \text{if active ice lid} \end{cases} \quad (4.6)$$

where α_i and $\alpha_w = 0.30$ are the snow-covered ice and deep pond albedo, h_p is the pond depth in m and ω is a constant scale factor in the same units as h_p . α_i is computed as a function of the surface wetness (melting conditions), cloud cover, snow depth and ice thickness (Shine and Henderson-Sellers, 1985, see Appendix A), as in the former version of the model without melt ponds. α_{bi} is the albedo of bare ice, equal to 0.58. Note that, like in Flocco et al. (2012), melt ponds/ice lids are optically active and affect the sea ice albedo only when they are deeper/thicker than a critical value of 1 cm. Then, the mean albedo of the various surface types weighed by their relative coverage of the ice cover is calculated:

$$\alpha = \alpha_i(1 - a_p) + \alpha_p a_p \quad (4.7)$$

where a_p is the melt pond fraction of ice area.

4.3 Forcing and setup

Except for precipitation, the model is driven by the same forcing fields as in Lecomte et al. (2013), all spatially interpolated onto the ORCA1 grid. NCEP / NCAR daily reanalyses of 2 m air temperature and 10 m u - and v - wind components of Kalnay et al. (1996) are used, together with monthly climatologies of total cloudiness and relative humidity of Berliand and Strokina (1980) and Trenberth et al. (1989), respectively. River runoff rates are taken from Dai and Trenberth (2002). Usually, we use the climatology of Large and Yeager (2004) as precipitation input to our model. Here, in order to get a more realistic regional variability of the snowfall, we added the precipitation anomalies from DFS5.2 (DRAKKAR Forcing Set, version 5, Dussin and Barnier, 2013) to the climatology. The making of DFS5 follows the same method as in Brodeau et al. (2010), applied on the ERA-interim reanalysis product (Simmons et al., 2007; Dee et al., 2011). The reason why the full DRAKKAR precipitation forcing set was not used was to avoid introducing a large bias in model results due to a bias in the mean precipitation rates (as compared to the former climatology), which would have required further tuning of the model. Surface heat fluxes (radiative and turbulent) are derived from Goosse (1997), and the

wind stress over sea ice is calculated with a quadratic bulk formula and a drag coefficient $C_d = 1.40 \times 10^{-3}$.

Since the DFS5 precipitation data we use are available from 1979, model simulations are performed from 1979 to 2011. In sections 4.5 and 4.6, the first three years of the simulations are not considered in order to let a short model spin-up before we start the analysis. The initial state for all sea ice covered regions of the Arctic Basin (defined by the locations where the sea surface temperature is below 0°C) is set to 3.5 m (0.3 m) for ice (snow) thickness, 0.95 for ice concentration, 270 K for snow and ice temperatures and 6 PSU for sea ice salinity. The snow scheme is used with 3 layers advected horizontally, but the vertical resolution is refined to 6 layers in the thermodynamics. Ocean temperatures and salinities are initialized from Levitus (1998). The ocean model runs hourly and calls to the sea ice model are made every 6 time steps.

4.4 Observations

In the result sections below, we compare our simulations to several observational datasets. We use Advanced Microwave Scanning Radiometer (AMSR-E), Special Scanning Microwave Imager (SSM/I) and Scanning Multichannel Microwave Radiometer (SMMR) sea ice extent data, processed by the NASA Team and Bootstrap algorithms and distributed by the NSIDC (National Snow and Ice Data Center) (Comiso and Nishio, 2008). In addition, airborne surveys of snow depth and ice thickness from Operation IceBridge (Kurtz et al., 2012, updated 2013) are used to evaluate our modeled snow depth and ice thickness distributions on Arctic MYI (defined as the ice remaining at the time of the minimum extent). In order to properly compare the model outputs with those observations, the data were interpolated on the ORCA1 grid and used to generate a mask for model fields. Therefore, whenever the model results are compared to IceBridge observations in the following, it is done only where observations at matching locations are available. MODIS (Moderate Resolution Imaging Spectroradiometer) melt pond fraction satellite data over 2000-2011 (ICDC MODIS, 2013; Rösel et al., 2012) were also used for comparison with simulated melt pond spatial distributions. The dataset, initially on the $12.5 \text{ km} \times 12.5 \text{ km}$ polar stereographic grid of the NSIDC, was interpolated on the ORCA1 grid for comparison with model outputs.

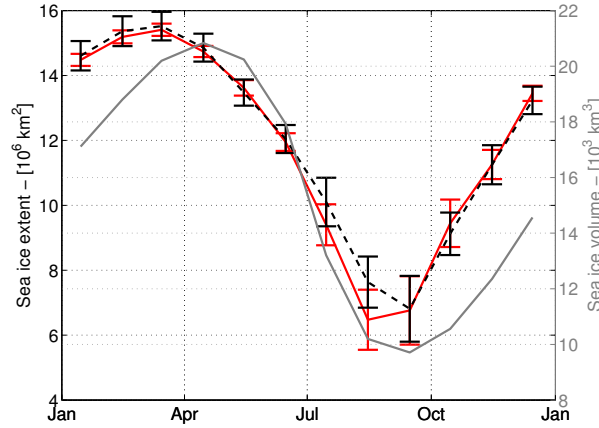


Figure 4.3: Simulated (red) and observed (NASA data, Comiso and Nishio, 2008, dashed black) mean annual cycles of Arctic sea ice extent over 1982-2011 and simulated mean annual cycle of Arctic sea ice volume (grey) over the same period. Error bars represent the standard deviation of the monthly extents, over the simulation period.

4.5 Results

4.5.1 Control run (CTL) versus observations

Here, we first analyse a control run (hereafter referred to as CTL) performed through 1982-2011, using the model as described in the previous sections. We compare CTL against the set of observations mentioned above to appraise the model skills in simulating the sea ice extent, volume, snow and melt pond cover.

4.5.1.1 Sea Ice

Figure 4.3 shows the simulated and observed mean seasonal cycle of sea ice extent through 1982-2011, and the simulated mean cycle of sea ice volume over the same period for CTL. Sea ice extents produced by the model match exactly the observed ones, except for the summer minimum, which is underestimated by half a million square kilometers and shifted from September to August. Although sea ice extent seems to be well reproduced in the simulation, sea ice volume appears too low, with a maximum and a minimum of $20.8 \times 10^3 \text{ km}^3$ and $9.7 \times 10^3 \text{ km}^3$, respectively. Comparing, for instance, these values with PIOMAS estimates (Zhang and Rothrock, 2003) would lead to a mean underestimation of the sea ice volume of about 20%. In comparison to the reference simulation of Lecomte et al. (2013), without the new processes

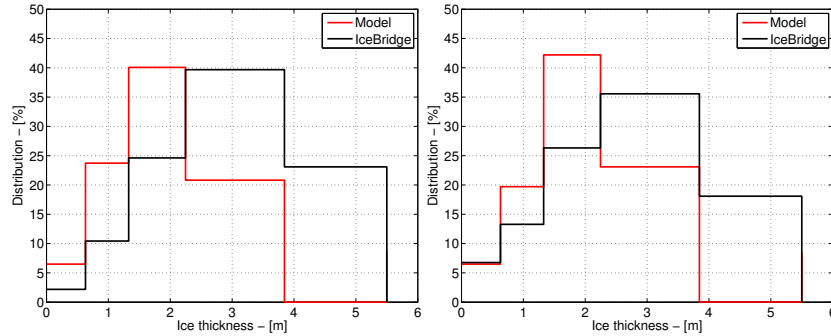


Figure 4.4: Ice thickness distributions (ITDs) as simulated by the model (red) and observed during IceBridge flights (Kurtz et al., 2012, updated 2013) (black), in March (left panel) and April (right panel). Simulated ITDs are averaged over the analysis period (1982-2011). Bounds in sea ice thickness correspond to those from the ice thickness distribution in the model.

(i.e., melt pond representation and blowing snow), the Arctic mean sea ice extent is similar, but the volume is reduced by about 40% in winter and 50% in summer (reasons are discussed in section 4.6.1).

The simulated ice thickness distributions (ITDs) (Figure 4.4) consistently show an underestimation of IceBridge ice thicknesses (Kurtz et al., 2012, updated 2013), mainly including MYI data. In March, the coverage of ice in the first three categories of ice thickness in the model is overestimated by up to 15% (for category 2 and 3), while the amount of ice in the last two categories is largely underestimated. In particular, all the ice thicker than 4 m observed during IceBridge flights is missing in the simulation. In April, the observed ITD indicates that ice thinner than 2 m has kept forming since March and the relative coverage of 0-2.3 m thick ice has increased to the cost of 2.3-5.5 m thick ice. Consequently, the biases between CTL and IceBridge ITDs are reduced, but the simulated sea ice thicknesses remain thinner than the observed ones in general, with an underestimation of 2.3-3.8 m thick ice of $\sim 13\%$ and 4-5.5 m thick ice that is still completely missing.

4.5.1.2 Snow

Like in Lecomte et al. (2013), snow depths are relatively well simulated by the model overall, as suggested by Figure 4.5 which shows the snow depth distributions (SDD) in CTL versus IceBridge data. As for ice thickness, the simulations exhibits a bias towards thinner snow, with a large proportion of 0-5 cm deep snow (12% in March, 9% in April, against 3% and 5% for IceBridge, respectively) and an underestimation of 30-45 cm deep snow by $\sim 3\%$. Note

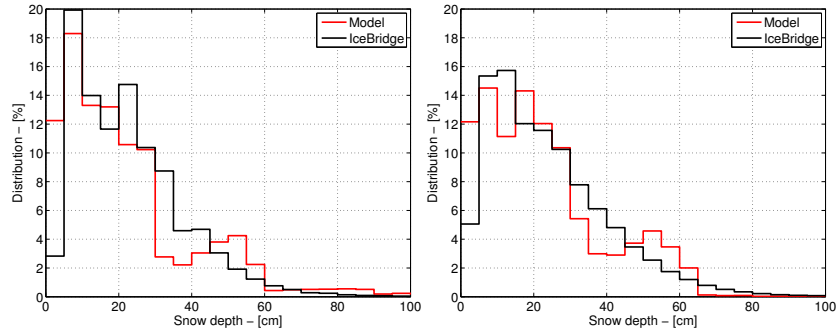


Figure 4.5: Snow depth distributions (SDDs) as simulated by the model (red) and observed during IceBridge flights (Kurtz et al., 2012, updated 2013) (black), in March (left panel) and April (right panel). Simulated SDDs are averaged over the analysis period (1982-2011).

that the simulation features a second mode in snow depth around 45-60 cm deep snow, which is not apparent in the observations.

4.5.1.3 Melt ponds

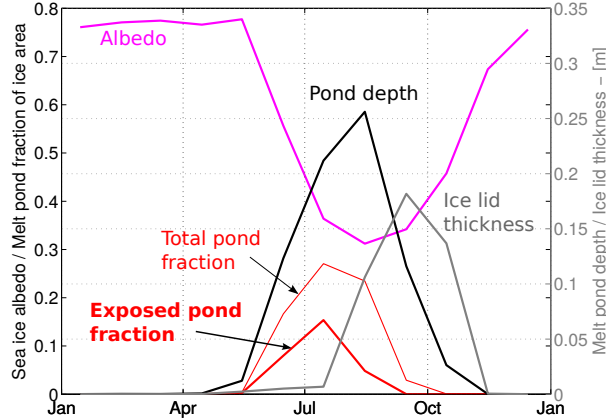


Figure 4.6: Mean annual cycles of melt pond-related diagnostics (through 1982-2011). The purple curve represent the ponded-ice clear sky broadband albedo. The bold and light red curves depict the exposed (unfrozen ponds) and total melt pond fraction of ice area, respectively. Black and grey curves read on the right hand y-axis, and picture the pond depth and ice lid thickness, respectively.

Figure 4.6 depicts the mean summer cycles of the main melt pond diagnostics in the model, namely the area fraction of ice covered in ponds, melt pond

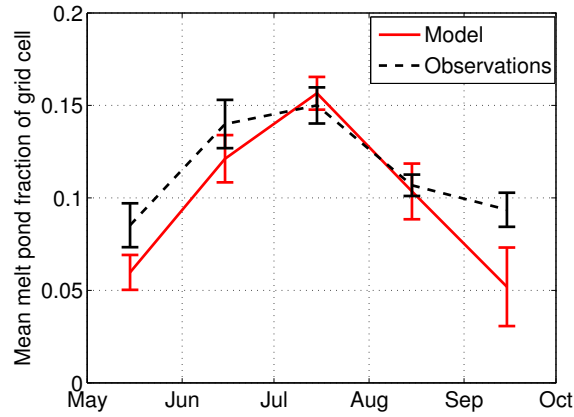


Figure 4.7: Mean seasonal cycles of melt pond fraction (over 2000-2011), expressed as a fraction of grid cell, from the model (red) and from MODIS observations (Rösel et al., 2012, dashed black). Error bars represent the standard deviation of the monthly melt pond fractions over 2000-2011.

depth, ice lid thickness (refreezing atop the pond) and the surface broadband albedo of pond covered ice. Melt ponds start forming in May, and reach their maximum coverage of 0.15 for exposed-ponds (i.e., still uncovered by a refreezing ice lid) and 0.28 for all-ponds in mid-July. The maximum depth of melt ponds (~ 0.25 m) is reached a month later, in August, before ponds shrink rapidly in September due to surface refreezing. The albedo consistently follows the evolution of the sea ice surface state, dropping from ~ 0.8 to ~ 0.3 in mid-August as melt ponds develop and raising again as soon as their refreezing stage begins. Specifically looking at the agreement between simulated exposed-ponds and MODIS observations of melt ponds (ICDC MODIS, 2013; Rösel et al., 2012) yields a good timing of the melt pond development in the model, and a consistent maximum value for the mean melt pond fraction over the Arctic Basin (Figure 4.7), although it is underestimated in early season. As stated in Rösel et al. (2012), MODIS melt pond fractions are overestimated in late season since their algorithm misclassifies areas of thin ice as ponded ice. Therefore, our simulated melt pond fractions may be realistic during this period. The generally low pond fractions in the Central Arctic Ocean is emphasized by Figures 4.8, 4.9 and 4.10, which display the mean geographical distributions of melt pond fraction, depth and difference between CTL and MODIS satellite melt pond fractions through June-September for 2000-2011 (period with available observations). The maximum melt pond fractions (~ 0.4) are found in July in the Canadian Archipelago and in the transitional region be-

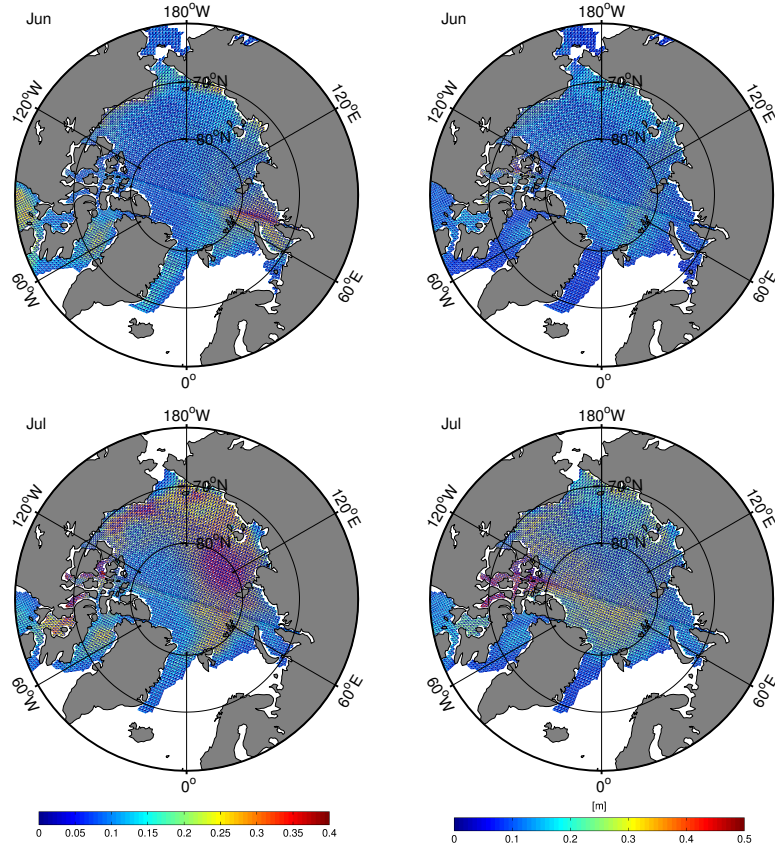


Figure 4.8: Average spatial distributions of melt pond fraction of ice (left) and depth (right, in meters) through June-July over Arctic sea ice, for the 1982-2011 period (part 1).

tween the Laptev Sea and the Arctic Ocean, whereas very deep melt ponds (0.5 m) are found in the Canadian Archipelago only. Overall, melt ponds tend to remain shallow in the marginal sea ice zone of the eastern basin (with fractional coverages of up to 0.35), while they get deeper (up to ~ 0.35 m) on the MYI of the western and central basin. This anti-correlation pattern is particularly noticeable in July, and to a lesser extent in August. Figure 4.10 shows a general model tendency to underestimate the melt pond relative coverage, especially in the central basin, while some coastal regions are significantly over-covered (overestimation by 0.15-0.25 in July).

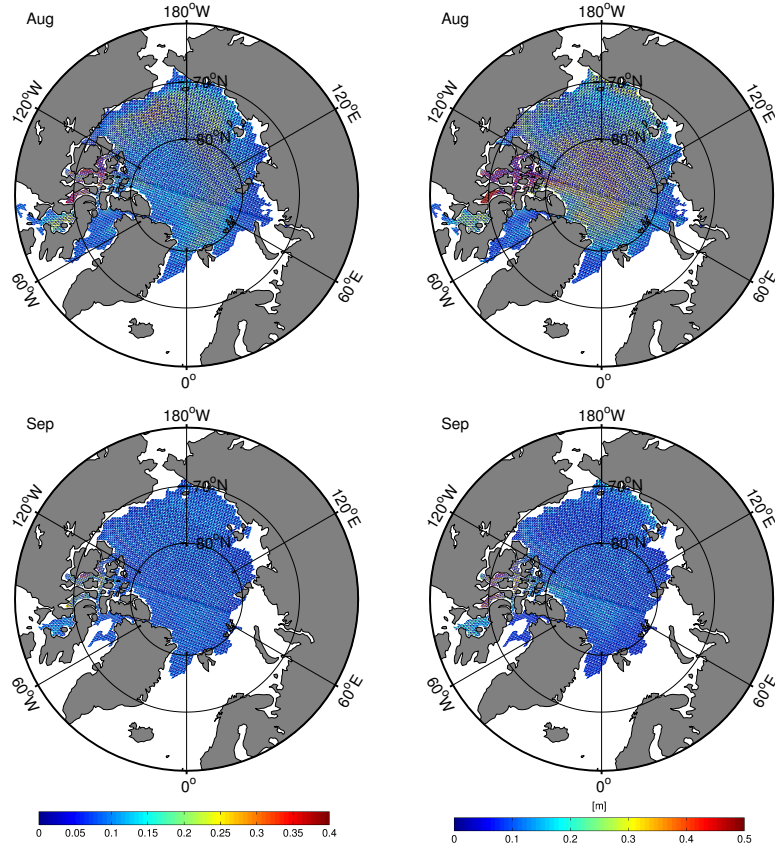


Figure 4.9: Average spatial distributions of melt pond fraction of ice (left) and depth (right, in meters) through August-September over Arctic sea ice, for the 1982-2011 period (part 2).

4.5.2 Sensitivity study : impacts of blowing snow on sea ice and melt ponds

In this section, we take advantage from having melt ponds and blowing snow both represented in our sea ice model to study the impacts of enhanced wind-transport of snow on melt pond distributions. So as to do this, CTL is compared to a second simulation called BSE (blowing snow enhanced simulation), in which the blowing snow tuning parameter γ was increased to $10^{-4} \text{ kg m}^{-2}$. Raising the value of this parameter practically increase the mass flux of snow removed from the pack when winds are strong enough for the snow drift process to occur. As mentioned previously, this parameter is underconstrained owing to lacking observations of wind-blown snow redistribution on sea ice,

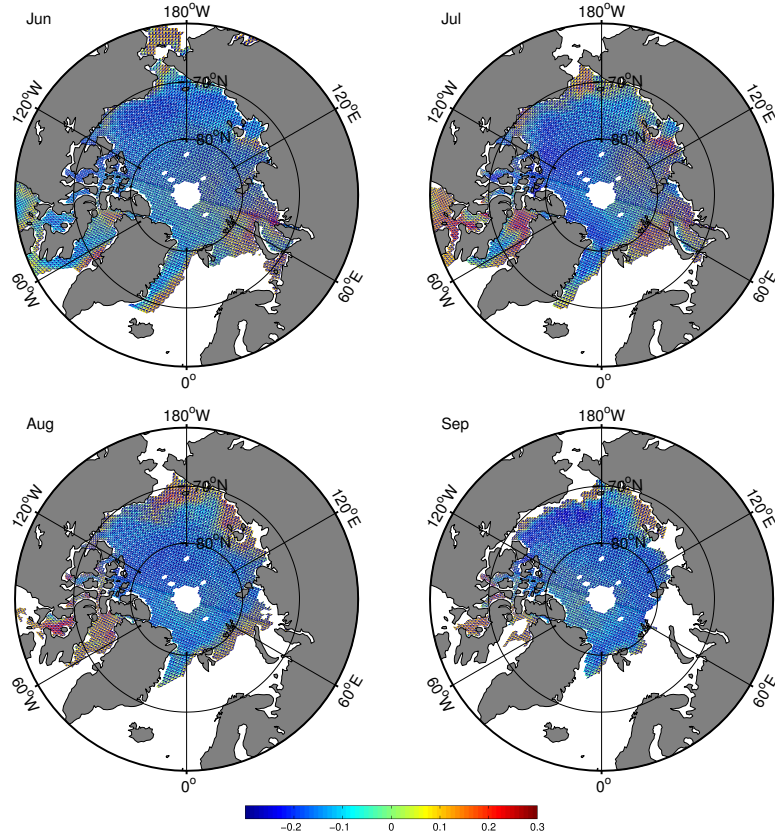


Figure 4.10: Average spatial distributions of the difference between modeled and observed pond fractions (of grid cell) through June-September 2000-2011.

so the choice of its value for BSE was simply made keeping a compromise between getting noticeable differences in model outputs (BSE vs. CTL) and avoiding unrealistically strong transport rates. In part of the analysis, we distinguish the impacts on MYI from those on FYI.

4.5.2.1 Impact on sea ice volume

The mean annual cycles of sea ice extent and volume from BSE simulation, are compared to those from CTL in Figure 4.11. Only small differences are observed between the two cycles of sea ice extent, with a summer minimum slightly lower in BSE than CTL (6.1 against 6.5×10^6 km²). By contrast, the sea ice volume in BSE is systematically lower than in CTL throughout the

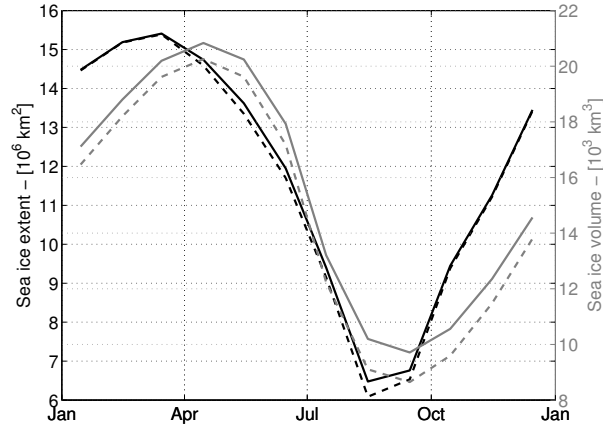


Figure 4.11: Mean annual cycles of sea ice extent (in black) and volume (grey) over 1982-2011 for CTL (solid lines) and BSE (dashed lines) simulations

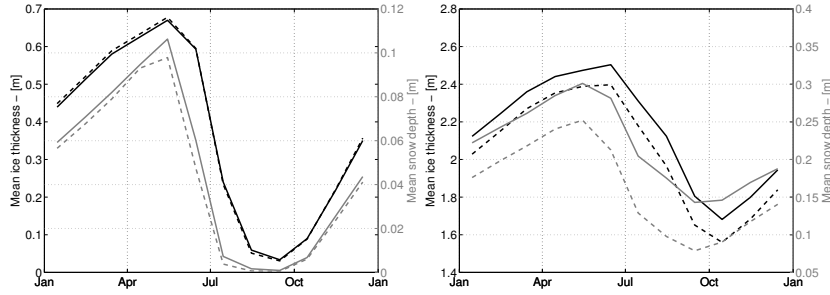


Figure 4.12: Mean annual cycles of ice thickness (black) and snow depth (grey) for CTL (solid) and BSE (dashed), computed over FYI (left) and MYI (right), through 1982-2011.

year, with a winter maximum and a summer minimum decreased by 0.7 and $1.0 \times 10^3 \text{ km}^3$, respectively. In summer, the latter difference translates into a volume loss of about 10% with respect to CTL. Figure 4.12 shows that this volume loss is exclusively experienced by MYI, with a 10-15 cm lower mean ice thickness in BSE, while the annual cycle of FYI thickness is almost identical in both simulations.

4.5.2.2 Snow depth changes

As illustrated by Figures 4.12 and 4.13, the primary effect of enhanced wind-transport of snow in the model is to reduce the snow depth on sea ice by removing it from the pack and transferring it into leads during the redistribu-

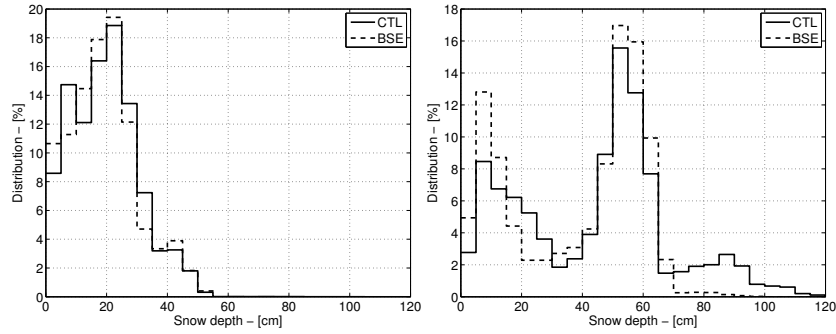


Figure 4.13: May snow depth distributions for CTL (solid lines) and BSE (dashed lines), averaged over the FYI (left) and MYI (right) area, over 1982-2011.

tion process. Therefore, the MYI snow depth distributions of BSE (in May, at the onset of melt pond formation) are clearly shifted towards thinner snow. In BSE, snow deeper than 70 cm, that represents $\sim 13\%$ of the snow pack in CTL, has virtually disappeared, whereas the proportion of 0-20 cm deep snow has increased by 10%. This snow thinning effect is much less noticeable on FYI. An explicit bimodal SDD is observable on MYI in both simulations, with peaks at 10 and 55 cm of snow that are all the more distinct in case of stronger erosion (BSE).

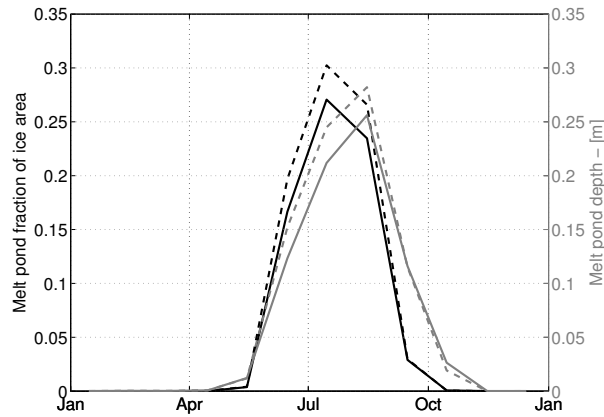


Figure 4.14: 1982-2011 mean seasonal cycles of melt pond fraction (black) and depth (grey) for CTL (solid lines) and BSE (dashed lines) simulations.

4.5.2.3 Impacts on melt pond distributions

The SDDs at the onset of the melting season impact on the melt pond development in the simulations. Figure 4.14 shows that the main effect of the overall reduced snow depths on Arctic sea ice is a small increase in average melt pond fraction ($\sim 3\%$) and depth (2.5 cm). Nonetheless, those global diagnostics hide interesting changes on MYI, as suggested by Figure 4.15, which depicts the changes in melt pond fractional coverage and depth between CTL and BSE for each ice type (i.e., FYI and MYI). The maximum melt pond area and depth on FYI seems to decrease in BSE, as compared to CTL, but those differences are very small (-1% and -0.5 cm, respectively). In contrast, the MYI melt pond cover in BSE exhibits a faster development, which leads to more considerable changes in both fractional coverage ($+ 5\%$ coverage) and depth ($+ \sim 4$ cm). The observed magnitude of the changes in melt pond distribution between the two simulations are consistent with those in SDD. Interestingly, although the changes on FYI are weak, they suggest a melt pond regime that differs from the one on MYI, with a melt pond volume seemingly decreasing in BSE (further discussed in section 4.6.2).

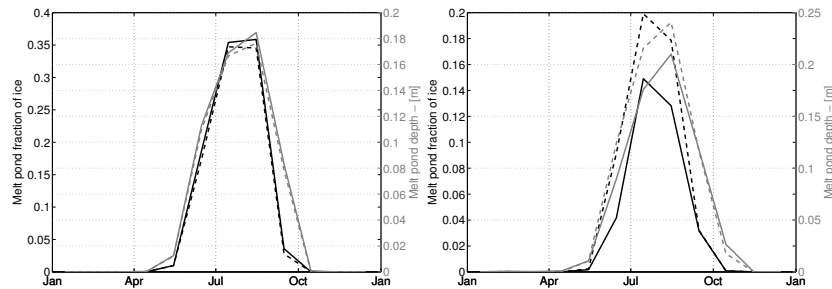


Figure 4.15: 1982-2011 mean seasonal cycles of melt pond fraction (black) and depth (grey) for CTL (solid lines) and BSE (dashed lines) simulations, and for each ice type (FYI on the left; MYI on the right).

4.6 Discussions

4.6.1 Control run versus observations

The sea ice volume loss observed in CTL compared to a no-melt pond run such as in Lecomte et al. (2013) is huge (up to 50% volume loss in summer, not shown) but consistent with Flocco et al. (2012) (using the same melt pond scheme in a similar multi-category sea ice model), in which they noticed a $\sim 40\%$ volume loss between pond and no-pond configurations. The early oc-

currence of the minimum ice extent on the other hand is not realistic with regard to present-day observations, but typical of a low ice volume simulation (Lecomte et al., 2013), where the ice has a weaker thermodynamical inertia and melts faster in summer. In addition, in the simulation, the minimum ice extent consistently occurs at the time of the maximum unfrozen melt pond volume.

The slight underestimation of snow depths compared to IceBridge observations is primarily explained by the underestimation of the ice thickness itself, suggesting younger ice overall that is not formed yet in the early accumulation season and therefore misses some snowfall events (Sturm and Massom, 2009). The shape of the SDDs, in Figure 4.13 especially, depict two maxima at 20-30 cm and 50-60 cm, while snow of moderate depth (~ 30 cm) seems to be lacking. This is not inconsistent with existing observations. Such bimodal SDDs on MYI were observed in some individual data sets of the IceBridge product (Kurtz and Farrell, 2011; Kwok et al., 2011) and in other snow observation works (Gerland et al., 2014).

Although the mean annual melt pond fraction cycle over 1982-2011 is in good agreement with MODIS observations, it seems lower than in existing studies. In Flocco et al. (2012) (same melt pond scheme), their total and exposed melt pond fractions were reaching a maximum of 0.33 and 0.2 in average, against 0.28 and 0.15 in our control run. In the control simulation of Hunke et al. (2013) (different melt pond scheme), the maximum pond fraction of ice area was ~ 0.4 and the maximum pond fraction of grid cell was 0.19 (against ~ 0.15 in our case). The reason for this underestimation of the mean annual cycle of pond fractions is the lack of melt ponds in the western Arctic Basin, as compared to both MODIS observations (see Figures 4.8, 4.9 and 4.10) and those studies (especially Flocco et al., 2012). Apart from this shortcoming, melt ponds in CTL are in relatively good agreement with the latter studies, especially in terms of spatial variability. Both Flocco et al. (2012) and Hunke et al. (2013) had large melt pond fractions in July in the eastern basin (fractions of ~ 0.3 - 0.35) and in the Canadian Archipelago, like in CTL. The geographical distribution of melt pond depths in our control run (in July) is also very consistent with those of the two mentioned studies, with the deepest ponds (0.3 to 0.5 m) along the north coasts of Greenland and Canada. The regional variability of melt ponds in the simulation highlights an anti-correlation pattern between pond depth and fractional coverage, that is characteristic of the melt pond scheme behavior. A shallow but spread out melt pond regime is observed on FYI, while concentrated and deeper melt ponds form atop MYI with a pronounced topography (also confirmed by Figure 4.15). In addition, Flocco and Feltham (2007) showed that their model is very sensitive to the snow volume on top of sea ice, that can partially or even totally conceal the

ponds. This provides the physical reason for the lack of melt ponds on the MYI of the western and central Arctic Basin in CTL, where some snow persists throughout the summer (Figure 4.12).

4.6.2 Sensitivity study : impacts of blowing snow on sea ice and melt ponds

Figures 4.14 and 4.15 suggest that, on average over the whole pond-covered ice and on MYI specifically, increased snow drift leads to increased pond volume. On the other hand, the FYI melt pond cover seems to respond differently. The first physical reason for these differences in response lies in the primary effect of the blowing snow process: sending a fraction of the sea ice snow cover into leads. The resulting snow thinning is illustrated in Figures 4.12 and 4.13. As mentioned in the previous section, snow does not entirely melt away in MYI regions and the blowing snow process influences the snow depth distributions both during winter and after the onset of the melt season. In case of a greater wind-transport of snow, such as in BSE, less snow is available on sea ice to hide melt ponds and impede their development (as shown in Figure 4.15). The overall increased melt pond fraction in this simulation in turn results in increased ice melt through the ice-albedo feedback and explains the lower MYI volume, as compared to CTL (see Figures 4.11 and 4.12, right panel).

The latter mechanism was not necessarily to be expected since, for instance, the opposite consequence was observed in Flocco et al. (2012) in which a simulation with increased precipitation led to more snow available to melt and feed the ponds. Nonetheless, those two behaviors are not inconsistent and are due to the specific influence of our blowing snow parameterization on FYI and MYI. Less (more) snow leading to smaller (larger) melt pond area is to be expected when the total snow cover is melted whatever its depth, which is the case on FYI of both simulations (see Figure 4.12). FYI however, once it starts forming, quickly reaches high concentrations and covers most of a model grid cell, meaning that when snow is blown by winds, little is lost into the ocean and most of it is redeposited on the ice it was removed from. Figure 4.13 left panel clearly underlines the reduced effect of blowing snow redistribution on FYI. Eventually, when the open water fraction is large enough in summer to capture snow, most of it is already melted and filling in melt ponds. In addition, independently of the snow cover on flat FYI, the liquid meltwater volume transportable on such ice is naturally limited by its topography. As a consequence, FYI in the model might naturally tend to fill with meltwater at full capacity anyway and be less sensitive to the overlying snow volume prior to the onset of the melting season. Thus, strengthened snow

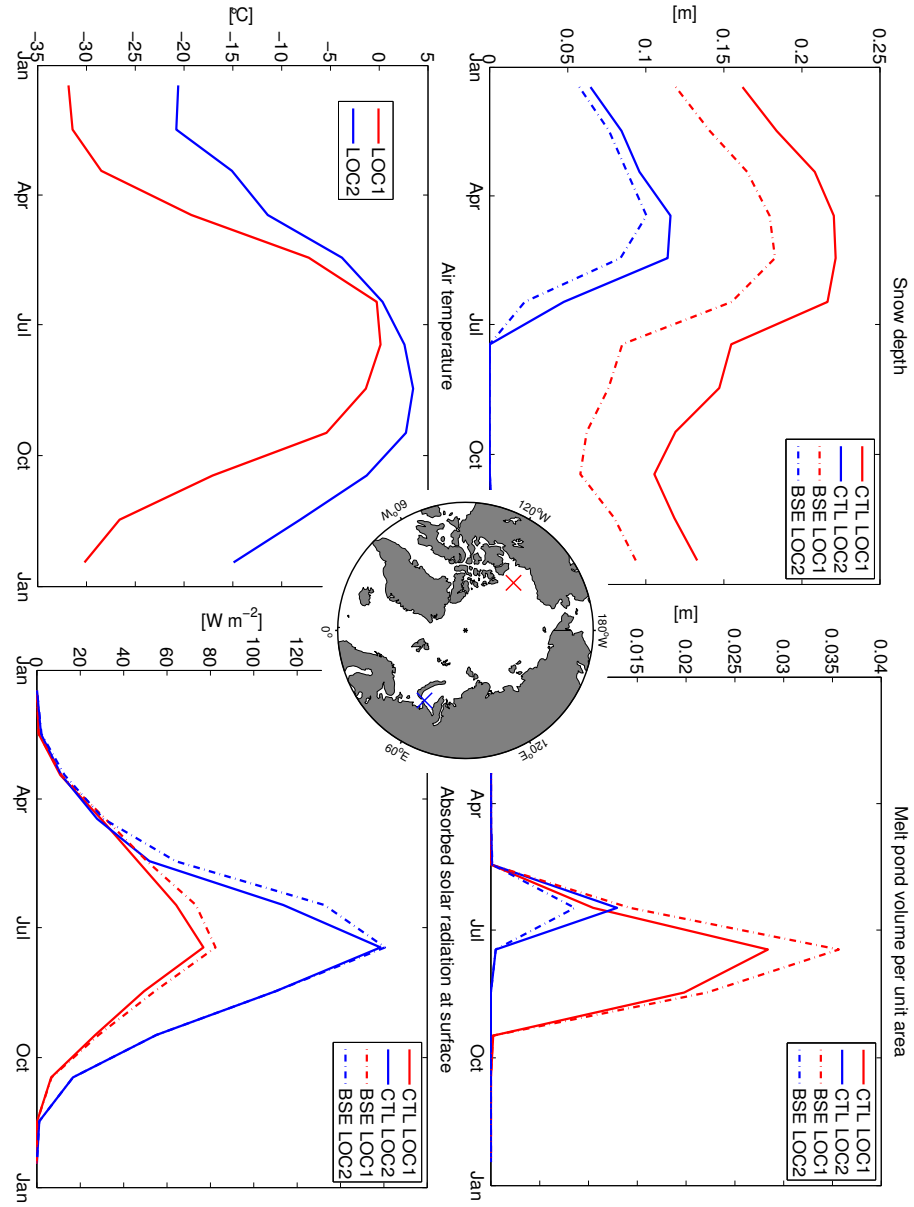


Figure 4.16: 1982-2011 climatologies of (a) snow depth, (b) melt pond volume, (c) air temperature and (d) absorbed solar radiation at the snow/ice surface, at two locations of the Arctic Basin. Those locations are shown on the central map. The first (LOC1, in red) is on MYI, and the second (LOC2, in blue) on FYI. Solid and dashed lines depict the climatologies for CTL and BSE simulations, respectively.

drift has little influence on melt pond formation on FYI (Figure 4.15), which also explains the rather weak impact at the basin scale (Figure 4.14).

Figure 4.15 (left panel) does not only suggest a reduced sensitivity of FYI to wind-blown snow redistribution. It also seems to reveal that the impact of enhanced snow drift on FYI melt pond formation tends to be the opposite of its counterpart on MYI ponds. This is in line with the findings of Flocco et al. (2012). One question then remains unanswered: what are the physical reasons for one model behavior to occur instead of the other? In order to address this issue, we chose two specific locations in the Arctic where the same effect of increased snow drift on snow depth leads to opposite changes in melt pond volume. The climatologies (1982-2011) of snow depth, melt pond volume, air temperature and absorbed solar radiation at the sea ice/snow surface for these two places are displayed in Figure 4.16. For both locations, increased wind-transport of snow leads to a snow thinning (panel a). The amplitude of those changes is expectedly different, since those two ice points are part of different ice covers (i.e., MYI and FYI). However, because snow is thinner and air temperature higher at location 2 (panel c), snow melts and disappears much faster in summer than at location 1. Since less snow is available to melt in BSE (as compared to CTL), the resulting maximum melt pond volume is smaller too (panel b). As soon as the ice is bare of snow, the drop in albedo triggers a sharp increase in absorbed solar radiation at the surface (panel d). Naturally, melt ponds at location 2 are ephemeral, and disappear as the ice melts entirely in July. Hence, two distinct causal connections can be highlighted between snow and melt pond formation, depending on air temperature conditions and snow depth prior to the melt onset. The first, in presence of a resilient snow cover that curtails the surface melting rates and hides the liquid water at the ice surface, relates thicker snow to reduced melt pond development. The second, probably typical of seasonal sea ice carrying a thinner snow cover, is the converse relationship. In the latter configuration, the amount of snow directly predetermines the melt pond volume. Finally, the small differences observed between CTL and BSE in Figure 4.15 (FYI, left panel) and Figure 4.16 for location 2 might be lessened by the averaging effect over years, in relation to year-to-year signals.

4.7 Conclusion

In order to study the influence of the sea ice late-spring snow cover on the development of melt ponds and the large-scale implications for sea ice, the explicit melt pond model of Flocco and Feltham (2007) was incorporated into a coupled ocean-sea ice model including an advanced snow representation and a blowing snow parameterization. The comparison of a control simulation

with MODIS satellite observations over the 2000-2011 period demonstrates a relatively good model ability to reproduce the spatial variability of melt pond depth over the Arctic Basin and a proper mean annual cycle of melt pond coverage. However, even keeping in mind the degree of uncertainty of MODIS melt pond fractions, the model seems to fail in simulating satisfactory pond fractions on MYI, where the concealing effect of persistent snow curtails the horizontal development of melt ponds. This results in an underestimation of pond fraction by up to 0.2 in the western basin, along the northern coasts of Greenland and Canada, while flat FYI regions of the peripheral basin are sometimes over-covered by ponds. Overall, the control simulation exhibits a realistic melt pond cover, despite a bias of the maximum mean melt pond fraction of about 5 to 10% toward smaller pond fractions, compared to two other studies (Flocco et al., 2012; Hunke et al., 2013). As in those studies, the sea ice volume in our control run compared to a no-pond configuration (Lecomte et al., 2013) is much lower (up to 50% volume loss) due to the ice-albedo feedback. Consistently, the simulated late-spring ITD compared to the IceBridge dataset exhibits an excessive proportion of 0.7-2.3 m thick ice, while it lacks ice in the thickest two categories of the model (2.3-5.5 m). The comparison of corresponding SDDs versus IceBridge observations therefore shows a slight bias toward thinner snow as well, which is characterized by an overestimation of very thin snow (0-5 cm). In spite of this, the tail of the SDDs remains in good agreement with observations.

The main effect of the blowing snow parameterization is to reduce snow depth on sea ice by sending a fraction of the transported snow into the leads. Increasing the intensity of snow redistribution by winds therefore subserves further snow thinning in the associated simulation. From there, two melt pond regimes and causal relationships are clearly identified. On MYI, where some snow persists through the summer, the smaller snow volume ensuing from enhanced snow drift has a weaker concealing effect on melt ponds, and facilitates their development. In an other situation, apparently more typical of FYI, a combined effect of higher air temperatures and relatively thin snow makes it a snow cover that entirely melts in summer. Hence, less snow leads to smaller pond fractions because the amount of freshwater available to fill them in is decreased, but the effect remains limited by the intrinsic nature of seasonal ice that bears restrained snow and water accumulation. The magnitude of such changes is thus smaller than those observed on MYI. The enhanced melt pond formation on MYI in case of increased wind-blown snow redistribution is thus substantially lessened if observed in average over the whole Arctic sea ice cover. Even so, it is influential enough to cause a 10% summer ice volume loss, with regard to our control simulation. It should finally be stated that our results might be influenced by the number of ice thickness

categories in the model given potential sensitivity of both our blowing snow parameterization and the melt pond scheme (Flocco and Feltham, 2007; Flocco et al., 2010) to that parameter.

Ultimately, the particular behavior of the blowing snow parameterization and the specific melt pond regimes on FYI as opposed to MYI raises two issues. The first concerns the distinct or shared sensitivity of all ice types to snow-related processes and properties in sea ice models, in general. Going further, the second regards the actual benefits of including detailed snow parameterizations into sea ice–ocean models running under current climate conditions, i.e., sustaining a progressive transition of sea ice towards a FYI exclusive cover. Those questions still require to be addressed.

Acknowledgements

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CHAPTER 5

BENEFITS FROM REPRESENTING SNOW IN COUPLED OCEAN–SEA ICE MODELS

This Chapter is based on the following paper:

Lecomte, O., Fichefet, T., Massonnet, F., Vancoppenolle, M., 2014b. Benefits from representing snow properties and related processes in coupled ocean–sea ice models. Submitted. Ocean Modelling.

Abstract

Several large-scale sea ice simulations are performed over the last three decades using a coupled ocean–sea ice model under the same experimental setup but partly modifying the representation of snow physics in the model. The inter-simulation spread yields the unreported result that the simulated multi-year ice is sensitive to such changes while the seasonal sea ice, is rather dominantly driven by the external oceanic and atmospheric forcings. In the context of a thinning Arctic sea ice cover, those findings suggest that including snow processes in large-scale sea ice models is beneficial, if not necessary, to predict the timing of the Arctic multi-year ice disappearance, whereas the operational forecasting of first-year

ice extent using fully coupled models will likely require to improve the oceanic and atmospheric components themselves.

5.1 Introduction

Snow on sea ice is a key component of the polar climate system. A review of the snow-related properties and processes of major relevance with respect to the sea ice energy and mass balances was established by Sturm and Massom (2009). Among them, the high albedo of snow as compared to the ocean or bare ice is probably the most important, drastically lessening the shortwave radiation input into the ice pack. Second, snow is a highly efficient insulator that reduces ice-atmosphere heat exchange, which smoothes temperature changes into snow and ice compared to the atmosphere and moderates the bottom ice growth rate. Snow directly contributes to the sea ice mass balance through snow ice formation, widespread in the Southern Ocean. The presence of snow generally delays the surface melting of the ice; however once snow starts melting, melt ponds start increasing in size, which lowers the albedo and enhances surface sea ice melting.

Snow has therefore long been expected to be an important component in sea ice models, with several studies supporting this idea (e.g., Eicken et al., 1995; Fichefet and Maqueda, 1999; Wu et al., 1999; Fichefet et al., 2000; Blazey, 2012; Blazey et al., 2013). Even so, its actual representation has so far been somewhat disregarded and kept relatively crude. Only recently, a new snow scheme was proposed by Lecomte et al. (2014a) for use in large-scale sea ice models, including a melt pond formalism as well (Flocco and Feltham, 2007). The latter study evidenced in particular an increased sensitivity of the Arctic multi-year ice (MYI) volume and of its summer melt pond cover to the effect of blowing snow on the late spring snow depth distributions on sea ice, in comparison with first-year ice (FYI). Extrapolating from those findings, the question of the distinct sensitivities of FYI and MYI to the physics of snow in present-day climate simulations may be asked in a broader way. This paper aims at addressing this issue in both the Arctic and the Antarctic, by analyzing the spread (simple statistical dispersion) between various simulations of a global ocean–sea ice model for which only snow parameterizations vary. The next two sections therefore give a brief overview of the models we use and the experiments we carried out, before results and their implications for future snow developments in sea ice models are discussed in the last two sections.

5.2 Model description

The coupled ocean–sea ice model we use here is NEMO-LIM (Nucleus for European Modeling of the Ocean - Louvain-la-Neuve Sea Ice Model), described in Madec (2008) for the ocean general circulation model OPA (Ocean PARallelisé, version 9) and Vancoppenolle et al. (2009) for the sea ice model LIM3

(LIM, version 3). LIM3 is a so-called multi-category, dynamic-thermodynamic sea ice model providing an explicit representation of the subgrid-scale ice thickness distribution. In order to complete this study, a physically based melt pond scheme and a new snow physics scheme were recently incorporated in the model. The melt pond formulation of Flocco and Feltham (2007) is utilized here. It retrieves the pond depth and fractional coverage of the sea ice based on the ice thickness distribution and the fresh water volume available to fill in the ponds. The snow representation includes a multi-layer snow thermodynamic scheme, accounting for the vertical heat transfer through snow layers with varying density and thermal conductivity, surface melt and sublimation, internal melting and refreezing of fresh water into the snow, snow ice formation subsequent to flooding and a simple formulation of the effect of snow packing by winds on the snow density profile. The model also includes intuitive parameterizations of the snow redistribution by winds on the ice categories within a grid cell and of snow losses in leads when winds blow snow away on top of an open sea ice pack. The comprehensive physical design of the snow scheme is available in Lecomte et al. (2013, 2014a).

5.3 Simulations

In this section, we provide a succinct description of the main characteristics of each model run. Except for the last simulation described hereafter, the experimental setup and the atmospheric fields used to force the model are the same as in Lecomte et al. (2014a). NCEP/NCAR daily reanalyses are used for 2 m air temperature and 10 m u – and v – wind components (Kalnay et al., 1996). Climatologies of Berliand and Strokina (1980) and Trenberth et al. (1989) are utilized for total cloudiness and relative humidity, respectively. Surface heat fluxes are computed following Goosse (1997). For the snowfall specifically, we use the precipitation anomalies from DFS5.2 (DRAKKAR Forcing Set, version 5, Dussin and Barnier, 2013) added to the climatology of Serreze and Hurst (2000).

The first simulation was performed enabling all snow and melt pond processes available in the model. This run is described and evaluated against observations in Lecomte et al. (2014a). In the second simulation, melt ponds were disabled, which resulted in a 40% (winter) to 50% (summer) higher mean Arctic sea ice volume as compared to the first one, due to the higher albedo and weaker sea ice surface melting in summer. Experiment 3 is the same as 2, except that the snow thermal conductivity was set equal to $0.31 \text{ W m}^{-1} \text{ K}^{-1}$, a commonly used value in ocean–sea ice coupled models. This led to even higher sea ice volumes in both hemispheres as a result of increased winter ice growth rates (Lecomte et al., 2013). The melt pond scheme was kept active in

Simulation number	Simulation specificity
1.	Simulation with all new components enabled (snow and melt ponds), as in Lecomte et al. (2014a).
2.	1. with melt ponds disabled.
3.	2. with constant snow thermal conductivity equal to $0.31 \text{ W m}^{-1} \text{ K}^{-1}$, as in Lecomte et al. (2013).
4.	1. with decreased deep pond albedo from 0.3 to 0.2.
5.	1. with internal melting/refreezing of fresh water into snow disabled.
6.	1. with increased erosion rates in blowing snow parameterization, as in Lecomte et al. (2014a).
7.	1. with climatology of (Serreze and Hurst, 2000) only for snowfall forcing (no variability).

Table 5.1: List of all performed simulations.

the fourth simulation, but the albedo of deep melt ponds was lowered from 0.3 to 0.2. Owing to a slightly enhanced ice-albedo feedback in this run, the mean Arctic sea ice volume decreased, but in small proportions. The reason for this is that ponds in the model are probably too shallow for the pond albedo to reach the deep-pond value. No major impact was observed on Antarctic sea ice, as melt ponds formation is very limited in the southern hemisphere. Run 5 was performed turning off the internal melting and refreezing of the fresh water into the snow. Again, the large-scale impacts on the simulated sea ice was not significant overall, because fresh water and cold snow are not necessarily present simultaneously, at least on FYI. The sixth simulation is also described in Lecomte et al. (2014a) and was achieved increasing the intensity of the blowing snow process in the model. Although the effect on the Arctic sea ice melt pond cover is small on average, it is clearly noticeable on MYI specifically, and influential enough to cause a $\sim 10\%$ loss in seasonal mean sea ice volume over the whole basin. The seventh and last simulation is identical to the first except that the snowfall forcing was changed back to a single climatology (Serreze and Hurst, 2000), mainly inducing changes in the geographical distribution of snow depths. For the sake of clarity, simulations are outlined in Table 5.1.

In the following, all simulations are analyzed over 1982-2011. The aim is not to proceed to the detailed analysis of each simulation, but rather to generally determine the extent to which changing the physical representation of snow in the model affects the main sea ice state variables, namely the total area and volume.

5.4 Results and discussion

Figures 5.1 and 5.2 depict the time series of the spatially integrated multi-year, first-year and total sea ice volumes in each hemisphere through 1982–2011, and the relative spread between the various simulations. This relative spread in ice volume at a given date is defined as the standard deviation normalized by the mean value over all simulations, for each ice type. The trends of all time series and the mean relative spread in ice volume between simulations over the period of analysis were computed and are reported in Table 5.2, together with the same statistics for sea ice area and snow volume in both hemispheres.

The analysis indicates negative and positive trends in both the area and volume of the Arctic MYI and FYI, respectively, which is in agreement with the current status of Arctic sea ice studies (e.g., Maslanik et al., 2007; Kwok and Untersteiner, 2011), showing a progressive transition towards a seasonal Arctic sea ice cover. In the Antarctic however, those trends are both positive and corroborate the findings of Comiso and Nishio (2008) and Zhang (2014). Note that the signs of the trends in snow volume are consistent with those in ice volume. This is to be expected since thicker ice, potentially older, is predisposed to a larger snow accumulation. Second and most importantly, Figures 5.1, 5.2 and Table 5.2 highlight systematically larger spreads between simulations for MYI than for FYI, except for Arctic sea ice area. The latter exception is discussed thereafter. The smaller spread on FYI can be explained by both the shorter lifetime of this type of ice and the competing processes it may undergo. Indeed, snow may accumulate on young ice only after it starts freezing, as opposed to the accumulation on older ice that survived the melt season. This natural limitation is critical, since the maximum of snowfall is in September, which reduces the impact of changes in daily snowfall rates on the FYI mass balance in a simulation. In addition, an increase in snowfall rate triggers a series of competing processes that tend either to increase or decrease the ice thickness. First, in the cold season, more snow induces more thermal insulation, a smaller oceanic heat loss to the atmosphere and, in turn, less basal ice growth. Second, if sufficient snowfall pushes the snow–ice interface below the sea level, snow ice starts to form, contributing to thicker ice. Finally, the summer melting of a thicker snow cover may ultimately lead to larger melt pond fractions on the sea ice and hence larger ice melt through the decrease in surface albedo. Thus, those competing processes moderate the impacts of an initial change in snow depth, due to altered snow physics or precipitation in the model, on the growth and melt of FYI. In contrast, MYI, which is thicker, enables both larger accumulation of snow throughout the year and larger water retention for melt pond formation in summer due to its uneven topography. A significantly thicker snow cover may persist through all or part

of the summer, therefore protecting the ice surface from intense melting and impeding the development of melt ponds. On a different course, the increased insulation effect of such a deeper snow pack combined with its total summer melting would result in both larger sea ice melting due to a larger melt pond cover and reduced wintertime ice growth. The latter processes, subserving a progressive decrease in ice thickness, would then contribute to the replacement of MYI by FYI that would in turn accumulate less snow. In this sense, FYI is less sensitive than MYI to changes in snow physics or snowfall forcing in the model, since the snow-related processes and the snow-ice feedbacks that take place on FYI are limited by its seasonality. Besides, FYI has by definition no memory, whereas MYI may cumulate changes over the years. Hence, the larger sensitivity of MYI volume to changing physics in the model might even be a more general characteristic, with respect to FYI. As already mentioned, the Arctic sea ice area makes exception to the previous comments, with similar spreads between simulations for both MYI and FYI ($\sim 10\%$). The reason is twofold. First, snow processes in the model do not influence sea ice area directly. Second, contrarily to the Antarctic MYI that is relatively thin and may switch to a FYI regime more easily, a significant fraction of Arctic MYI is thick and resilient enough for not shifting regime and suffering from losses in coverage. Notwithstanding, this might be expected to change in a longer simulation where the MYI fraction in the Arctic would be further reduced.

An other interesting fact from Table 5.2 is that, although the relative spread between simulations is smaller for FYI than MYI in the Southern Ocean, it is still considerably larger than for Arctic FYI. Antarctic sea ice in simulations features only little surface melt, as compared to its Northern counterpart (Lecomte et al., 2013), and its mass balance is mainly governed by basal growth and melt, and snow ice formation. As explained earlier, those processes tend to compete against each other in terms of impact on the ice growth, but snow ice formation also mitigates any change in snow depth since it is a sink mechanism for snow. The snow and ice column being assumed in hydrostatic equilibrium in the process, a larger snow load on the sea ice induces larger snow ice production, all the while decreasing the snow depth, and conversely. In consequence, alterations of the snow insulation effect by means of snow depth variations are mitigated likewise. The widespread occurrence of snow ice formation on Antarctic seasonal sea ice therefore gives the snow a greater influence through this process.

The temporal evolution of the inter-simulation spread in Figures 5.1 and 5.2 is also naturally explained by the aforementioned remarks. In the Arctic, as MYI is gradually replaced by FYI, the model sensitivity to the snow physics is reduced, as suggested by the decreasing spread in all-ice volume toward the end of the experiments. The Antarctic sea ice pack however, mainly seasonal,

yields a weaker sensitivity to changes in snow parameterizations or accumulation in the model all along the period of analysis, which is unexpected. This ultimately suggests that, in both hemispheres, the interannual variability of FYI is mainly driven by external forcing such as the heat fluxes from the ocean and the atmosphere, while both the variability and trend of the total Arctic MYI volume are responsive to both the forcing and the representation of snow in the model.

Table 5.2: Mean relative inter-simulation spreads and trends of the total sea ice volume, sea ice area and snow volume over 1982–2011, for both hemispheres. The mean relative inter-simulation spread of a variable is calculated as the ratio between its standard deviation and its mean value over all simulations (hence in %), averaged over the whole period of analysis. Trends are all significant at the 95%-confidence level.

	FYI		MYI	
	Mean spread	Trend	Mean spread	Trend
Arctic	Units*/decade		Units*/decade	
Sea ice volume [$\times 10^3 km^3$]	0.06	0.36	0.51	-4.5
Sea ice area [$\times 10^6 km^2$]	0.089	0.45	0.10	-0.93
Snow volume [$\times 10^3 km^3$]	0.14	0.041	0.20	-0.66
Antarctic				
Sea ice volume [$\times 10^3 km^3$]	0.17	-0.031	0.49	0.13
Sea ice area [$\times 10^6 km^2$]	0.16	0.05	0.46	0.17
Snow volume [$\times 10^3 km^3$]	0.21	-0.025	0.43	0.0097

*Corresponding units from the leftmost column.

5.5 Conclusion

This brief study reports on the sensitivity of the ocean–sea ice coupled model NEMO-LIM to the representation of snow processes and properties. A series of experiments using different snow parameterizations or snowfall forcings is analyzed over the same period (1982–2011). Consistently with the current literature on the sea ice extent and volume trends over the last decades, model simulations show a decline in Arctic MYI and its progressive replacement by FYI. Likewise, the recently observed increase in Antarctic sea ice extent is qualitatively captured by model simulations. The distinct analysis of

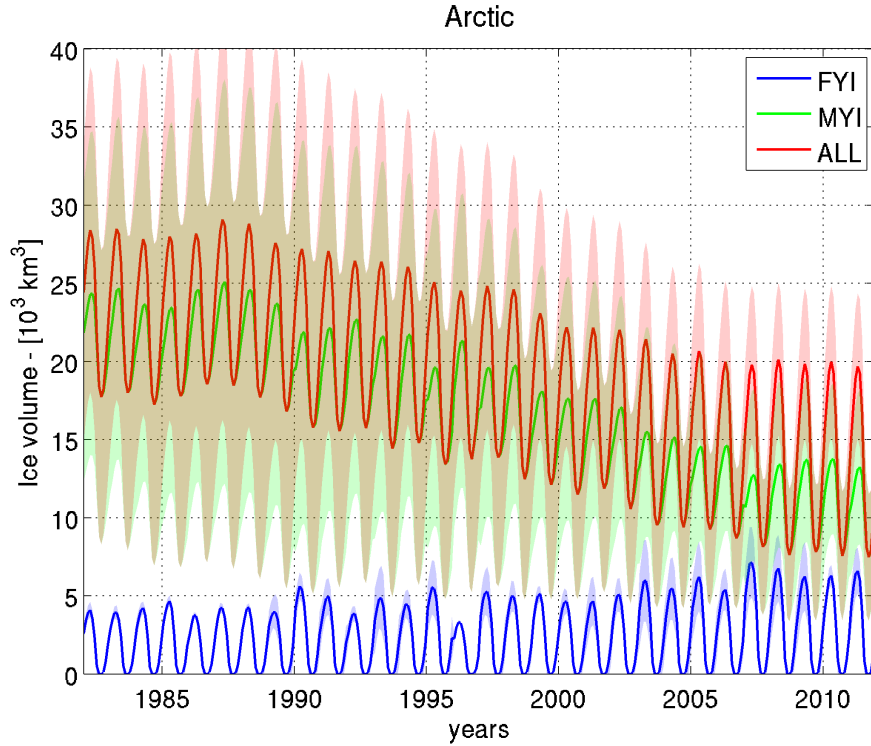


Figure 5.1: Time series of spatially integrated multi-year (green), first-year (blue) and total (red) sea ice volume in the Arctic over 1982-2011. For each ice type, the main solid line represents the mean sea ice volume over all experiments. The shaded area of the same color below and above each solid line corresponds to the spread between simulations, calculated as plus or minus one standard deviation of the sea ice volume over the experiments.

the inter-simulation spread for MYI and FYI yields the expected but yet unreported finding that the simulated MYI state and its snow cover are affected by the snow processes in larger proportions than FYI is. Those conclusions therefore explain the reduced sensitivity of NEMO-LIM to snow physics in the Southern Ocean, as compared to the Arctic's, and the decreasing dispersion between model runs in the northern hemisphere as the sea ice cover shifts from MYI to FYI toward the end of the simulations. Furthermore, they have several implications in terms of future model developments. As the growth and decay of seasonal sea ice seems primarily driven by external forcing, due to its smaller thermodynamical inertia, the priority in the framework of the operational predictability of sea ice using fully coupled models (that, in particular, currently tend to fail in simulating the increase in Antarctic sea ice

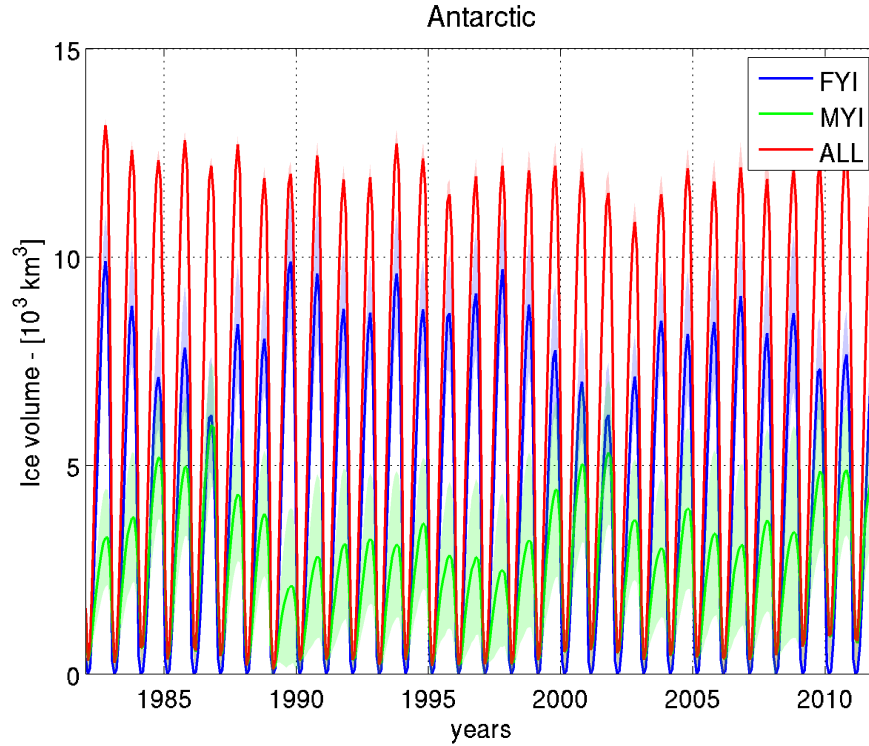


Figure 5.2: Time series of spatially integrated multi-year (green), first-year (blue) and total (red) sea ice volume in the Antarctic over 1982-2011. For each ice type, the main solid line represents the mean sea ice volume over all experiments. The shaded area of the same color below and above each solid line corresponds to the spread between simulations, calculated as plus or minus one standard deviation of the sea ice volume over the experiments.

extent over the past decades; Arzel et al., 2006; Zunz et al., 2013) will likely be to get the oceanic and atmospheric mean states and variability well simulated. This task is complex and in turn prerequisites well resolved subcomponents, such as atmospheric chemistry or meltwater inflow from ice sheets and glaciers. Including and improving snow processes in sea ice models will nonetheless be a necessary condition to predict the timing of MYI disappearance. In our model, MYI production proved to be most sensitive to the effect of the heat transfer through the snow layers on the ice bottom growth and to the melt pond scheme. However the other parameterizations, in particular blowing snow, are still insufficiently constrained and might happen to be just as important as the latter processes in a fine-calibrated fully coupled model. Extensive observations of such mechanisms are thus crucially needed in order

to address properly the problem of snow parameterization improvement in sea ice models. For instance, measuring the mass fluxes of transported snow on sea ice during a storm, as well as snow depth distributions before and after the event, may allow to constrain our parameterization and improve it so as to represent the widespread occurrence of partially bare ice, suggested by existing observations.

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CHAPTER 6

CONCLUSIONS

6.1 Contributions of the thesis

What we wanted to investigate. In Chapter 1, a literature-based series of the major snow processes on sea ice was introduced. Although the actual representation of snow in large-scale ocean–sea ice models had been kept relatively simple prior to this work, several studies (e.g., Eicken et al., 1995; Fichefet et al., 2000) had shown that accounting for the latter processes in such models would potentially bring improvements to the sea ice simulations over the last decades. This thesis essentially serves the purpose of (1) validating this hypothesis and (2) providing the sea ice modeling community with directions for the most relevant snow-related implementations to be made in its models, depending on the simulations it wishes to perform. In the following, we attempt to summarize how those scientific objectives were tackled.

What we learned. First and foremost, this work led to the implementations of new snow physical components in our sea ice model. Although this does not answer any scientific question, it constitutes a technical achievement that allowed us to test the interactions presented in Chapter 1 and will probably be a valuable modeling toolbox for quantitative evaluations of the impacts of those snow processes in future works. The numerical developments made to the ocean–sea ice model NEMO-LIM3 (Nucleus for European Modelling

of the Ocean - Louvain-la-Neuve Sea Ice Model, version 3) could have led to significant increases in computational cost. Therefore, our first concern was to determine the minimal vertical resolution to be kept for solving snow thermodynamics in the model. Chapter 2 addressed this problem and showed improving model skills in reproducing observed snow/sea ice temperature profiles with increasing resolution, in the context of 1D sea ice simulations. In spite of this, threshold snow layer numbers above which the statistics stopped improving were found. Those thresholds were comprised between 3 and 6 snow layers depending on the locations where the model was used, and gave the justification for finally using 3 layers of snow in NEMO-LIM3 with regard to the horizontal advection, refined to 6 in the treatment of vertical thermodynamics. Chapter 3 comes as an extension to both the latter study and the older works of Wu et al. (1999) and Fichefet et al. (2000), which assessed the sensitivity of coupled atmosphere–sea ice and ocean–sea ice models to snow thermal conductivity using mono-layer snow schemes. In our investigations, we found that none of the parameterizations of snow thermal conductivity regularly used in large-scale models yields satisfying model results with respect to sea ice thickness and extent in both hemispheres simultaneously. It was demonstrated that this originated from different sea ice mass balances in the Arctic and Antarctic. So as to engage this problem, a new formulation linking the bulk snow thermal conductivity and density to the mean seasonal wind speed was developed based on observations and tested in the model. The advantage of such a formulation is to have snow thermo-physical properties related to a space-dependent meteorological parameter that may be used in both hemispheres equivalently. This parameterization indeed led to a compromise in the goodness of Arctic and Antarctic sea ice simulations with respect to observations and, in particular, to Arctic snow depth distributions close to those observed during recent airborne observation campaigns (Ice-Bridge; Kurtz et al., 2012, updated 2013).

Even though Chapter 4 deals with altogether different processes and focuses on the Arctic, as opposed to the global analysis made in Chapter 3, those two chapters tightly connect to each other through one of their main conclusions. In Chapter 4, a melt pond component and the blowing snow mechanism were introduced in NEMO-LIM3. In the resulting analysis, a reduced impact of wind-blown snow redistribution on melt pond formation on Arctic FYI, as compared to MYI, was found. Consistently, findings of Chapter 3 had previously indicated a weaker sensitivity of the simulated Antarctic sea ice state, mostly seasonal, to changes in snow thermal conductivity and density parameterizations. Those complementary conclusions thus motivated the broader discussion of Chapter 5, about the distinct sensitivity of modeled FYI and MYI to snow processes and properties in general. In addition, Chapter

4 brings clarifications about a process-oriented physical issue. In Flocco et al. (2012), a sensitivity experiment in which the authors increased the amount of snow precipitation resulted in an increase in the Arctic melt pond cover, due to larger water input coming from snow melt. In our study, an increase in snow depth prior to the onset of summer surface melt (through a reduced effect of the wind-redistribution process) results in smaller pond fractions overall, as the snow hides the ponds in the early stages of their formation and impedes their developments. By distinguishing FYI from MYI in the analysis of our simulations, it was evidenced that both mechanisms actually happen to take place. Two melt pond regimes were identified depending on snow depth and air temperature combined conditions. It appears that, in the presence of a thin snow cover and air temperatures favoring the early and complete melting of snow, more snow prior to summer melt leads to larger pond fractions. This situation seems likely to occur on FYI, but the effect is moderate in terms of magnitude owing to the naturally limited accumulation of snow and water on seasonal ice. In the case of a thicker and hence more resilient snow cover, on MYI typically, the second of the aforementioned mechanisms prevails and deeper snow curtails the development of melt ponds more efficiently.

Lastly, Chapter 5 supports the conclusions inferred by the previous two chapters and confirms the larger sensitivity of MYI state variables to the representation of snow in our model. Quantitatively, sea ice and snow volumes among the simulations considered in this study appear to be at least twice as responsive, on MYI, to changes in model snow physics, as it is on FYI (Table 5.2).

Those conclusions ultimately give parts of the answer for the scientific questions of the present thesis. With respect to (1), our study validates and emphasizes the importance of representing snow processes and properties in large-scale coupled ocean–sea ice models, in particular for two types of simulations. The first regards simulations of Arctic sea ice under present-day climate conditions, in which determining the rate of decline of the MYI cover is critical. The second include all kinds of simulations of colder or past climates during which proxy-based observations of the perennial sea ice cover are existing but scarce (e.g., Crespin et al., 2013; Klein et al., 2013; Stranne et al., 2013). As for (2), including specific snow processes in models may be necessary depending on the simulations one wants to perform. In the Arctic, our model proved to be very sensitive to the thermo-physical description of snow (solely changing snow thermal conductivity and density formulations led to a ~ 2 m difference in mean Arctic sea ice thickness; Figure 3.4, Table 3.2) and to melt ponds interactive with snow (Chapter 4). The wind-blown snow redistribution process by itself is also important through its effect on snow depth distributions, but will need further calibrations for future applications

(see discussion below). In consequence, the incorporation of those processes in models is advised. However, the priority for simulations of the Antarctic sea ice cover in the framework of fully coupled models will more likely be to improve the oceanic and atmospheric components, which seem to drive most of FYI variability.

6.2 Remaining technical issues

6.2.1 Forcing uncertainties

The forcing uncertainties have been little discussed in the present work. Yet, the atmospheric reanalyses and climatologies used to force our model simulations have known biases (e.g., Bromwich et al., 2007; Vancoppenolle et al., 2011) that could explain part of the disagreement between model outputs and observations. In particular, the snowfall forcing is critical and affects, although in different ways, sea ice and its snow cover in both hemispheres. Antarctic sea ice features very widespread snow ice formation, due to relatively thin ice and high precipitation rates throughout the year. The freeboard of Antarctic FYI is, when not negative, typically near-zero (Jeffries et al., 1998), implying a specific linear relationship of snow depth as a function of ice thickness (from the hydrostatic balance of the snow and ice column), with the slope of this relationship depending on snow and ice densities. Since winds drive snow densities in the model, a bias in wind velocities from the forcing would change the latter relationship on one hand, and any bias in snowfall would affect the total amount of snow ice production. This is important for snow depth, but also for sea ice, as snow ice formation during austral summer is sometime attributed a role in the retention of snow-covered sea ice in some regions (Massom et al., 2006). In the Arctic, where the snow depth distributions are more likely determined by the age of snow, whether it is lying on FYI or MYI (at least in the Central Basin, Iacozza and Barber, 1999), biases in solid precipitation rates may have significant impacts on both the mean and the shape of these distributions.

Forcing uncertainties may also have special implications with respect to one of the main conclusions of this thesis: FYI in the model seems less sensitive than MYI to the representation of snow and is mostly responsive to the oceanic and atmospheric forcings at its interfaces. Trying other forcing datasets to drive our model or running it in a fully coupled framework with an interactive atmosphere would then probably result in quantitative changes in FYI mean state and interannual variability.

6.2.2 Computational costs

At the time this thesis is written, our ocean–sea ice model NEMO-LIM exhibits significantly increased computational costs ($\sim 50\%$), when running with the new snow and melt pond components as compared to its former version. However, we are very confident that there is relatively large room for optimization in order to reduce those additional costs. Besides, the horizontal advection of snow and sea ice properties is computed using the scheme of Prather (1986), which conserves the second-order moments of the spatial distribution of global state variables and preserves their positivity after transport. This scheme is practically non-diffusive but has the drawback of being computationally expensive. Hence, after optimization of the code, the extra computational costs induced by our new snow model will mainly depend on the number of new snow state variables to advect. A solution could then be to switch toward an alternative, numerically less demanding, advection scheme (e.g., Dukowicz and Baumgardner, 2000).

6.3 Perspectives

While this whole thesis revolves around representing snow processes and their impacts on sea ice, the perspectives rising from this work all relate to the better quantification of those impacts. In order to proceed to such an effort, four tasks may be undertaken.

(1) Use the effective tools. As already explained, this work led to the implementation of numerous mechanisms in NEMO-LIM3. This modeling toolbox provides the possibility of testing many configurations of the model and the ensemble of all possibilities has not been fully explored. For instance, sensitivity tests related to snow thermal insulating properties were performed in configurations where melt pond and blowing snow processes were disabled, since they were unavailable in the model at that precise time. Redoing those experiments in a configuration with all new processes activated would bring valuable information about the relative importance of ice thickness-growth and albedo-temperature feedback mechanisms, both on seasonal and perennial ice. Since experiments with melt ponds resulted in 40 to 50% reductions in Arctic sea ice volume (see Chapter 4), the smaller MYI cover in those cases implies a reduced sensitivity of the simulated Arctic sea ice to the thermal properties of its snow cover. This tentative assumption has been made but requires to be verified.

(2) Better quantification of impacts through process-oriented improvements. Because observational data are lacking at this stage to better constrain the new snow parameterizations implemented in our model, our conclusions

are often qualitative. However, improving those parameterizations is a necessary condition for better understanding the snow - sea ice interactions. Better quantifying the impacts of snow processes would enable to simplify the conceptualization of the snow-sea ice interactive system. In other words, putting numbers (using appropriate metrics) on those mechanisms would allow to prioritize them in order of top to least importance with respect to the large-scale sea ice mass and energy balances and ultimately simplify this schematic. Floe-scale observations of processes, such as wind-blown snow redistribution on sea ice, are imperatively needed before any further model developments associated to those processes are made.

(3) Better quantification of impacts through data assimilation. An other way to tackle the latter issue is data assimilation. More reliable large-scale satellite observations of snow start to become available (Kwok and Cunningham, 2008; Maaß et al., 2013). Assimilating those data in our model may improve future simulations and especially allow to optimally estimate uncertain snow parameters. Those methods have already proven to be successful for the calibration of sea ice parameters (Massonnet et al., 2014). Although the uncertainties of this kind of data remain large, they could be used for such purposes, at least as long as process-oriented In Situ observations are lacking.

(4) Include other processes. In our study, relatively little attention was paid to the radiative properties of snow. This was done voluntarily. Choosing or developing a snow radiative scheme adapted to sea ice modeling applications requires a big effort in itself. At the very beginning of this work, choosing to investigate the role of other processes such as blowing snow offered greater novelty, since the radiative aspects were concurrently being investigated by other groups (e.g., Holland et al., 2012), hence the path we followed. Nonetheless, better representing the albedo and optical characteristics of snow in our model is bound to be a mandatory task to go through, as the radiative transfer in snow may be one of the main reasons for the disagreements between the model and observations in Chapters 2 and 3. Lastly, between the beginning and end of the present work the continual advances in computer sciences have lessened the constraints of running models with increasing computational costs. Thus, secondary processes, such as wind-forced heat transfers through the advection of air in snow or depth hoar formation, may be tested and happen to have non-negligible influences on the vertical snow-sea ice thermodynamics.

This thesis makes NEMO-LIM3 the first large-scale coupled ocean-sea ice model including a snow representation of intermediate complexity. Whether this complexity is required or not to improve the model results is still an open question, but the better snow physics, now available in the model, will defi-

nitely allow to quantify the impacts of snow processes on sea ice with more accuracy. This work also highlights the necessity of a proper snow representation in climate models so as to suitably reproduce or forecast the evolution of multi-year ice in simulations of the past, present and future climates. Although sea ice models have been developed for decades, snow in these models has until now remained an under-addressed issue. This thesis paves the way in the investigation of these research questions. In the context of a projected global warming, possibly accompanied by increases in snowfall in polar regions, a lot is still to be done in order to understand the intimate relationship between snow and sea ice.

APPENDIX

SNOW AND SEA ICE ALBEDO PARAMETERIZATION IN THE LOUVAIN-LA-NEUVE SEA ICE MODEL (LIM)

This supplementary material was written based on LIM's users guide (Vancoppenolle et al., 2012) and Shine and Henderson-Sellers (1985).

A.1 Surface albedo

The surface albedo α determines the fraction of incoming solar radiation reflected by the surface. The clear-sky sea ice albedo α_{cs}^i is parameterized following Shine and Henderson-Sellers (1985), as a function of snow depth, ice thickness and surface temperature (see table A.1 and A.2). As these variables are specific for each ice thickness category, each category ends up with its own surface albedo. The parameterization interpolates albedo between several pivotal values, which are hardcoded in the routine computing the albedo (see table A.1). Depending on the forcing used, $\bar{\alpha}_m^i$ can be tuned within 10%, which has a moderate effect on summer ice characteristics.

Snow and sea ice albedo parameterization in the Louvain-la-Neuve Sea Ice Model (LIM)

Snow / sea ice class	Symbol	Value
Thick dry snow	$\bar{\alpha}_f^s$	0.8
Thick melting snow	$\bar{\alpha}_m^s$	0.65
Thick frozen bare ice	$\bar{\alpha}_f^i$	0.72
Thick melting bare ice	$\bar{\alpha}_m^i$	0.53

Table A.1: Key albedo parameters in the parameterization of Shine and Henderson-Sellers (1985)

Frozen bare ice	α_f^i	$= \bar{\alpha}_f^i$ if $h_i > 1.5$ $= 0.472 + 2(\bar{\alpha}_f^i - 0.472)(h^i - 1.0)$ if $1.0 \leq h^i \leq 1.5$ $= 0.2467 + 0.7049h^i - 0.8608h^{i2} + 0.3812h^{i3}$ if $0.05 \leq h^i \leq 1.0$ $= 0.1 + 3.6h^i$ if $0 \leq h^i \leq 0.05$
Melting bare ice	α_m^i	$= \bar{\alpha}_m^i$ if $h^i > 1.5$ $= 0.472 + 2(\bar{\alpha}_m^i - 0.472)(h^i - 1.0)$ if $1.0 \leq h^i \leq 1.5$ $= 0.2467 + 0.7049h^i - 0.8608h^{i2} + 0.3812h^{i3}$ if $0.05 \leq h^i \leq 1.0$ $= 0.1 + 3.6h^i$ if $0 \leq h^i \leq 0.05$
Dry snow	α_f^s	$= \bar{\alpha}_f^s$ if $h^s \geq 0.05$ $= \bar{\alpha}_f^i + h^s(0.8 - \alpha_f^i)/0.05$ if $h^s \leq 0.05$
Melting snow	α_m^s	$= \bar{\alpha}_m^s$ if $h^s \geq 0.1$ $= \bar{\alpha}_m^i + (\bar{\alpha}_m^s - \bar{\alpha}_m^i)h^s/0.1$ if $h^s \leq 0.1$

Table A.2: Surface albedo parameterization of Shine and Henderson-Sellers (1985). h^s and h^i are the snow depth and ice thickness, respectively.

A.2 Clear-sky versus overcast-sky albedo

An overcast-sky value is also computed $\alpha_{os}^i = \alpha_{cs}^i + 0.06$. The broadband shortwave sea ice albedo increases in nature because clouds change the spectral distribution of incoming light (Grenfell and Perovich, 1984). Following this, the absorbed shortwave flux at the snow/sea ice surface is formulated as a linear combination of clear sky and overcast sky contributions, weighted by cloud fraction.

APPENDIX B

WIND-BLOWN SNOW REDISTRIBUTION IN LIM: EXTENDED DESCRIPTION

This supplementary material is an extended version of section 4.2.2.2 (Chapter 4), describing the blowing snow parameterization in more technical details.

Blowing snow (or snow drift) on sea ice tends to redistribute the snow mass from thin level ice to thicker and deformed ice. The process smoothes out small and high frequency changes in sea ice topography meanwhile enhancing larger topographic features such as pressure ridges (Sturm et al., 2002). Usually, the “blowing snow” term includes several sub-processes such as snow erosion and motion by saltation or suspension (snow transport by wind). Snow drift is also a sublimation-enhancing process, facilitating the sublimation of grains thrown in suspension (Déry and Yau, 2002). The impacts of such mechanisms are well documented (e.g., Déry and Tremblay, 2004; King et al., 2008; Sturm and Massom, 2009; Leonard and Maksym, 2011). Most importantly, snow that is ripped off the pack and transported by the wind may be lost into leads, or redeposited on the ice, altering the spatial distribution of snow depths. As the wind direction shifts or the floe rotates, snow undergoes further erosion that may affect the surface drag coefficient and the dynamic response of sea ice to the wind forcing. Given the difficulty of establishing a

new parameterization due to the lack of observations available for this specific process, we only thought to represent the zero-order impact of blowing snow on the subgrid scale snow depth distributions in our model.

The usual conservation equation for a global state variable such as the total snow mass m_s on a single ice thickness category writes:

$$\frac{\partial m_s}{\partial t} = -\nabla(m_s \mathbf{u}) + \Psi^{m_s} + \Theta^{m_s} + \Phi^{bs} \quad (\text{B.1})$$

where $-\nabla(m_s \mathbf{u})$ represents the horizontal advection of snow due to sea ice motion, Ψ^{m_s} the effects of the mechanical redistribution of sea ice on snow, Θ^{m_s} the changes in snow mass due to the thermodynamic processes described in the previous two sections and Φ^{bs} the snow redistribution by the wind. The later term may be written as:

$$\Phi^{bs} = \Phi_{local}^{bs} + \Phi_{nonlocal}^{bs} \quad (\text{B.2})$$

the local term accounting for redistribution of the snow mass over the sea ice thickness categories, and the non local one representing the snow transport from grid cell to grid cell. For the sake of simplification, we assume that, at the synoptic scale of a storm, the snow mass being transported out of a grid cell matches the amount of snow that is transported in and we therefore neglect the non local effect of blowing snow. The validity of this assumption may depend on the horizontal resolution of the model grid, but would require the coupling with an atmospheric circulation model in order to be properly tested. Φ^{bs} may be separated into two terms:

$$\Phi^{bs} = \Phi_E^{bs} + \Phi_R^{bs} \quad (\text{B.3})$$

where Φ_E^{bs} and Φ_R^{bs} are the erosion and re-deposition rates, respectively. Based on the thought, commonly shared by snow and sea ice observationalists, that snow drift mainly depends on wind speed and sea ice surface topography, we formulate those rates as follows:

$$\Phi_E^{bs} = \frac{\partial m_s}{\partial t}_{erosion} = -\frac{\gamma}{\sigma_{ITD}}(V - V^*) \frac{\rho_{sMAX} - \rho_s}{\rho_{sMAX}} \quad (\text{B.4})$$

$$\Phi_R^{bs} = \frac{\partial m_s}{\partial t}_{redemption} = g(h_i)(1 - f_l)m^{bs} \quad (\text{B.5})$$

$$f_l = (1 - A_i)e^{-\frac{\sigma_{ITD}}{\sigma_{REF}}} \quad (\text{B.6})$$

ρ_s and ρ_{sMAX} are the density and maximum density of snow in the model, respectively. The ratio involving those two parameters in equation B.4 accounts

for the increased (decreased) resistance of snow to erosion as it is denser (lighter). V is the wind speed and V^* the threshold wind velocity required to trigger erosion. V^* is computed as:

$$V^* = \frac{\rho_s - \beta}{\alpha} \quad (\text{B.7})$$

where α and β are coefficients retrieved from a linear regression of observations relating the seasonal mean wind speed and the actual snow density (Lecomte et al., 2013). Physically, V^* represents the seasonal mean wind velocity that is required to make the snow pack as dense as ρ_s , and it is assumed that erosion will start only as soon as the instantaneous wind speed exceeds this value. σ_{ITD} is the standard deviation of the ice thickness distribution (i.e., over the thicknesses of the ice categories in the grid cell), considered as a roughness length and limiting the erosion in case of large disparity of ice thicknesses (represented by a large σ_{ITD}). Given the relatively poor ice deformation diagnostics in current sea ice models, we chose this simple solution instead. Finally, γ is a mass flux tuning coefficient, which standard value was set to $10^{-5} \text{ kg m}^{-2}$. This value is currently impossible to calibrate based on observations, so it was adjusted to provide comparable results as those obtained in Lecomte et al. (2013) and we aim at assessing the sensitivity of our model to this new parameterization and its influence on melt pond distributions. Integrating equation B.4 over time and sea ice thickness (i.e., summing numerically over ice thickness categories), the total mass of snow in suspension in a grid cell m^{bs} is obtained. A fraction f_l of this mass, defined as a function of the total ice concentration A_i and σ_{ITD} , is lost into leads. Intuitively, the larger A_i is, the smaller f_l becomes. In contrast, larger σ_{ITD} values retain snow on sea ice more efficiently. σ_{REF} (set to 1 m) is a scale factor for σ_{ITD} . The fraction of snow kept on the ice is then redistributed to the various thickness categories proportionally to their relative coverage of the total ice area, hence the use of the ice thickness distribution function $g(h_i)$ as originally defined by Thorndike et al. (1975). This formalism is the most simple but ensures mass conservation, and the redistribution of snow enthalpies is done accordingly. A more complex solution using an other redistribution function f under the constraint

$$\int_0^\infty f(h_i) dh = 1$$

could certainly be used, but would definitely require observations in order to be formulated realistically.

REFERENCES

- Abels, G., 1892. Measurement of the snow density at Ekaterinburg during the winter of 1890-1891. *Academia Nauk, Memoirs* 69. *Cited page 50*
- Allison, I., Brandt, R., Warren, S., 1993. East Antarctic sea ice: Albedo, thickness distribution, and snow cover. *Journal of Geophysical Research* 98 (C7), 12,417–12,429. *Cited page 19*
- Andreas, E., Ackley, S., 1982. On the differences in ablation seasons of Arctic and Antarctic sea ice. *Journal of the Atmospheric Sciences* 39 (2), 440–447. *Cited page 9*
- Armstrong, R., Brun, E., 2008. *Snow and climate: physical processes, surface energy exchange and modeling*. Cambridge University Press. *Cited page 23*
- Arzel, O., Fichefet, T., Goosse, H., 2006. Sea ice evolution over the 20th and 21st centuries as simulated by current AOGCMs. *Ocean Modelling* 12 (3), 401–415. *Cited page 104*
- Bartelt, P., Lehning, M., 2002. A physical SNOWPACK model for the Swiss avalanche warning: Part I: numerical model. *Cold Regions Science and Technology* 35 (3), 123–145. *Cited page 11*
- Berliand, M., Strokina, T., 1980. Global distribution of the total amount of clouds. *Hydrometeorological, Leningrad* 71. *Cited pages 51, 77, and 98*
- Bitz, C., Lipscomb, W., 1999. An energy-conserving thermodynamic model of sea ice. *Journal of Geophysical Research* 104 (C7), 15,669–15,677. *Cited pages 4, 20, 42, and 70*
- Bitz, C., Roe, G., 2004. A mechanism for the high rate of sea ice thinning in the Arctic Ocean. *Journal of Climate* 17 (18). *Cited page 2*
- Blazey, B., 2012. *Snow Cover on the Arctic Sea Ice: Model Validation, Sensitivity, and 21st Century Projections*. *Cited pages 12 and 97*

- Blazey, B., Holland, M., Hunke, E., 2013. Arctic Ocean sea ice snow depth evaluation and bias sensitivity in CCSM. *The Cryosphere* 7 (6). *Cited page 97*
- Bouillon, S., Morales Maqueda, M., Legat, V., Fichefet, T., 2009. An elastic-viscous-plastic sea ice model formulated on Arakawa B and C grids. *Ocean Modelling* 27 (3), 174–184. *Cited pages 42 and 71*
- Briegleb, B., Light, B., 2007. A Delta-Eddington multiple scattering parameterization for solar radiation in the sea ice component of the Community Climate System Model. NCAR Technical Note. Tech. rep., Note NCAR/TN-472+ STR, National Center for Atmospheric Research. *Cited pages 12 and 44*
- Brodeau, L., Barnier, B., Treguier, A., Penduff, T., Gulev, S., 2010. An ERA40-based atmospheric forcing for global ocean circulation models. *Ocean Modelling* 31 (3), 88–104. *Cited page 77*
- Bromwich, D., Fogt, R., Hodges, K., Walsh, J., et al., 2007. A tropospheric assessment of the ERA-40, NCEP, and JRA-25 global reanalyses in the polar regions. *Journal of Geophysical Research -All series-* 112 (D10), 10111. *Cited pages 62 and 110*
- Brun, E., Martin, E., Simon, V., Gendre, C., Coleou, C., 1989. An energy and mass model of snow cover suitable for operational avalanche forecasting. *Journal of Glaciology* 35 (121), 333–342. *Cited pages 11 and 46*
- Brun, E., Martin, I., Spiridonov, I., 1997. Coupling a multi-layered snow model with a GCM. *Annals of Glaciology* 25, 66–72. *Cited page 48*
- Calonne, N., Flin, F., Morin, S., Lesaffre, B., du Roscoat, S., Geindreau, C., St Martin d'Heres, F., Grenoble, F., Grenoble, I., 2011. Numerical and experimental investigations of the effective thermal conductivity of snow. *Geophysical Research Letters* 38 (23), L23501. *Cited page 41*
- Cheng, B., Vihma, T., Pirazzini, R., Granskog, M., 2006. Modelling of superimposed ice formation during the spring snowmelt period in the Baltic Sea. *Annals of Glaciology* 44 (1), 139–146. *Cited page 72*
- Cheng, B., Zhang, Z., Vihma, T., Johansson, M., Bian, L., Li, Z., Wu, H., 2008. Model experiments on snow and ice thermodynamics in the Arctic Ocean with CHINARE 2003 data. *Journal of Geophysical Research* 113 (C9), C09020. *Cited pages 19, 31, and 37*
- Colbeck, S., 1982. An overview of seasonal snow metamorphism. *Reviews of Geophysics* 20 (1), 45–61. *Cited page 23*

- Comiso, J., Cavalieri, D., Markus, T., 2003. Sea ice concentration, ice temperature, and snow depth using AMSR-E data. *Geoscience and Remote Sensing, IEEE Transactions on* 41 (2), 243–252. *Cited page 11*
- Comiso, J., Nishio, F., 2008. Trends in the sea ice cover using enhanced and compatible AMSR-E, SSM/I, and SMMR data. *Journal of Geophysical Research* 113 (C2), C02S07. *Cited pages 78, 79, and 100*
- Crespin, E., Goose, H., Fichefet, T., Mairesse, A., Sallaz-Damaz, Y., 2013. Arctic climate over the past millennium: Annual and seasonal responses to external forcings. *The Holocene* 23 (3), 321–329. *Cited page 109*
- Curry, J., Schramm, J., Ebert, E., 1995. Sea ice-albedo climate feedback mechanism. *Journal of Climate* 8 (2), 240–247. *Cited page 2*
- Dai, A., Trenberth, K., 2002. Estimates of freshwater discharge from continents: Latitudinal and seasonal variations. *Journal of hydrometeorology* 3 (6), 660–687. *Cited pages 51 and 77*
- De Ridder, K., Gallée, H., 1998. Land surface-induced regional climate change in southern Israel. *Journal of applied meteorology* 37 (11), 1470–1485. *Cited page 11*
- Dee, D., Uppala, S., Simmons, A., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M., Balsamo, G., Bauer, P., et al., 2011. The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society* 137 (656), 553–597. *Cited page 77*
- Déry, S., Tremblay, L., 2004. Modeling the Effects of Wind Redistribution on the Snow Mass Budget of Polar Sea Ice*. *Journal of Physical Oceanography* 34 (1), 258–271. *Cited pages 73 and 117*
- Déry, S., Yau, M., 2002. Large-scale mass balance effects of blowing snow and surface sublimation. *Journal of Geophysical Research: Atmospheres* (1984–2012) 107 (D23), ACL–8. *Cited pages 73 and 117*
- Domine, F., Arnaud, L., Bock, J., Carmagnola, C., Champollion, N., Gallet, J., Lesaffre, B., Morin, S., Picard, G., 2011a. Vertical profiles of specific surface area, thermal conductivity and density of mid-latitude, Arctic and Antarctic snow: relationships between snow physics and climate. *American Geophysical Union Fall Meeting, San Francisco, Calif., 5-9 Dec.* *Cited pages 48, 51, and 64*

- Domine, F., Bock, J., Morin, S., Giraud, G., 2011b. Linking the effective thermal conductivity of snow to its shear strength and density. *Journal of Geophysical Research* 116, F04027. *Cited pages 48 and 50*
- Domine, F., Gallet, J., Bock, J., Morin, S., 2012. Structure, specific surface area and thermal conductivity of the snowpack around Barrow, Alaska. *Journal of Geophysical Research* 117, D00R14. *Cited page 48*
- Dozier, J., Warren, S., 1982. Effect of viewing angle on the infrared brightness temperature of snow. *Water Resources Research* 18 (5), 1424–1434. *Cited pages 19 and 44*
- Dukowicz, J., Baumgardner, J., 2000. Incremental remapping as a transport/advection algorithm. *Journal of Computational Physics* 160 (1), 318–335. *Cited page 111*
- Dussin, R., Barnier, B., 2013. The Making of DFS 5.1. personal communication, <http://www.drakkar-ocean.eu/forcing-the-ocean>. *Cited pages 77 and 98*
- Eicken, H., Fischer, H., Lemke, P., 1995. Effects of the snow cover on Antarctic sea ice and potential modulation of its response to climate change. *Annals of Glaciology*, volume 21, pp., 369–376. *Cited pages 5, 12, 19, 24, 41, 97, and 107*
- Eicken, H., Lange, M., Wadhams, P., 1994. Characteristics and distribution patterns of snow and meteoric ice in the Weddell Sea and their contribution to the mass balance of sea ice. *Annales Geophysicae* 12 (1), 80–93. *Cited pages 19 and 24*
- Fetterer, F., Untersteiner, N., 1998. Observations of melt ponds on Arctic sea ice. *Journal of Geophysical Research: Oceans* (1978–2012) 103 (C11), 24821–24835. *Cited pages 5 and 69*
- Fichefet, T., Maqueda, M., 1999. Modelling the influence of snow accumulation and snow-ice formation on the seasonal cycle of the Antarctic sea-ice cover. *Climate Dynamics* 15 (4), 251–268. *Cited pages 6, 19, 25, and 97*
- Fichefet, T., Morales Maqueda, M., 1997. Sensitivity of a global sea ice model to the treatment of ice thermodynamics and dynamics. *Journal of Geophysical Research-Oceans* 102 (C6). *Cited pages 5, 12, 41, 45, and 71*
- Fichefet, T., Tartinville, B., Goosse, H., 2000. Sensitivity of the Antarctic sea ice to the thermal conductivity of snow. *Geophysical Research Letters* 27 (3), 401–404. *Cited pages 12, 19, 34, 41, 58, 97, 107, and 108*

- Flocco, D., Feltham, D., 2007. A continuum model of melt pond evolution on Arctic sea ice. *Journal of Geophysical Research* 112 (C8), C08016.
Cited pages 14, 70, 74, 89, 92, 94, 97, and 98
- Flocco, D., Feltham, D., Bailey, E., Schroeder, D., 2014. The impact of refreezing of melt ponds on Arctic sea-ice thinning, Abstract presented at 2014 IGS International Symposium on Sea Ice in a Changing Environment, Hobart, Tas., 10-14 Mar. 69A572. *Cited page 69*
- Flocco, D., Feltham, D., Turner, A., 2010. Incorporation of a physically based melt pond scheme into the sea ice component of a climate model. *Journal of Geophysical Research* 115 (C8), C08012. *Cited pages 69, 74, and 94*
- Flocco, D., Schroeder, D., Feltham, D., Hunke, E., 2012. Impact of melt ponds on Arctic sea ice simulations from 1990 to 2007. *Journal of Geophysical Research: Oceans* (1978–2012) 117 (C9).
Cited pages 69, 74, 75, 77, 88, 89, 90, 92, 93, and 109
- Gallée, H., Duynkerke, P., 1997. Air-snow interactions and the surface energy and mass balance over the melting zone of west Greenland during the Greenland Ice Margin Experiment. *Journal of Geophysical Research-Atmospheres* 102 (D12), 13,813–13,824. *Cited page 22*
- Gerland, S., Haas, C., Wang, C., Granskog, M., Haapala, J., Makshtas, A., 2014. Snow thickness distributions over sea ice in the vicinity of the North Pole, 2005-13, Abstract presented at 2014 IGS International Symposium on Sea Ice in a Changing Environment, Hobart, Tas., 10-14 Mar. 69A668. *Cited page 89*
- Girard, L., Bouillon, S., Weiss, J., Amitrano, D., Fichet, T., Legat, V., 2011. A new modeling framework for sea-ice mechanics based on elasto-brittle rheology. *Annals of Glaciology* 52 (57), 123–132. *Cited page 4*
- Goosse, H., 1997. Modeling the large-scale behavior of the coupled ocean-sea ice system. Ph.D. thesis, 231 pp., Univ. Catholique de Louvain, Louvain-la-Neuve, Belgium. *Cited pages 27, 51, 77, and 98*
- Goosse, H., Fichet, T., 1999. Importance of ice-ocean interactions for the global ocean circulation: A model study. *Journal of Geophysical Research* 104 (C10), 23337–23355. *Cited pages 2, 42, and 70*
- Grenfell, T., Maykut, G., 1977. The optical properties of ice and snow in the Arctic Basin. *Journal of Glaciology* 18 (80), 445–463. *Cited pages 21 and 44*
- Grenfell, T., Perovich, D., 1984. Spectral albedos of sea ice and incident solar irradiance in the southern Beaufort Sea. *Journal of Geophysical Research: Oceans* (1978–2012) 89 (C3), 3573–3580. *Cited page 116*

- Haas, C., Thomas, D., Bareiss, J., 2001. Surface properties and processes of perennial Antarctic sea ice in summer. *Journal of Glaciology* 47 (159), 613–625. *Cited pages 19 and 25*
- Heil, P., Allison, I., Lytle, V., 1996. Seasonal and interannual variations of the oceanic heat flux under a landfast Antarctic sea ice cover. *Journal of Geophysical Research-Oceans* 101 (C11), 25,741–25,752. *Cited pages 27 and 63*
- Hellmer, H., Schröder, M., Haas, C., Dieckmann, G., Spindler, M., 2008. The ISPOL drift experiment. *Deep Sea Research Part II: Topical Studies in Oceanography* 55 (8-9), 913–917, ice Station POLarstern (ISPOL): Results of interdisciplinary studies on a drifting ice floe in the western Weddell Sea.
URL <http://www.sciencedirect.com/science/article/B6VGC-4S38BVJ-1/2/01b91ee36132c09bf1a259639eee0435>
Cited pages 10, 25, and 46
- Hibler, W., 1979. A dynamic thermodynamic sea ice model. *Journal of Physical Oceanography* 9 (4), 815–846. *Cited page 4*
- Holland, M., Bailey, D., Briegleb, B., Light, B., Hunke, E., 2012. Improved sea ice shortwave radiation physics in CCSM4: the impact of melt ponds and aerosols on Arctic sea ice*. *Journal of Climate* 25 (5), 1413–1430. *Cited pages 44, 69, 75, and 112*
- Houghton, J., Ding, Y., Griggs, D., Noguer, M., Van der Linden, P., Dai, X., Maskell, K., Johnson, C., 2001. IPCC, 2001: Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge, United Kingdom, New York, USA, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 881 pp. *Cited page 29*
- Hunke, E., Dukowicz, J., 1997. An elastic-viscous-plastic model for sea ice dynamics. *Journal of Physical Oceanography* 27 (9), 1849–1867. *Cited pages 4, 42, and 70*
- Hunke, E., Hebert, D., Lecomte, O., 2013. Level-ice melt ponds in the Los Alamos sea ice model, {CICE}. *Ocean Modelling* 71 (0), 26–42.
URL <http://www.sciencedirect.com/science/article/pii/S1463500312001680>
Cited pages 15, 69, 71, 89, and 93
- Hunke, E., Lipscomb, W., 2010. CICE: the Los Alamos sea ice model, documentation and software user's manual, Version 4.1. *Cited pages 4 and 12*

- Hunke, E., Lipscomb, W., Turner, A., 2010. Sea-ice models for climate study: retrospective and new directions. *Journal of Glaciology* 56 (200), 1162–1172.
Cited page 4
- Huwald, H., Tremblay, L., Blatter, H., 2005. A multilayer sigma-coordinate thermodynamic sea ice model: Validation against Surface Heat Budget of the Arctic Ocean (SHEBA)/Sea Ice Model Intercomparison Project Part 2 (SIMIP2) data. *Journal of Geophysical Research* 110, C05010. *Cited page 19*
- Iacozza, J., Barber, D., 1999. An examination of the distribution of snow on sea-ice. *Atmosphere-Ocean* 37 (1), 21–51. *Cited page 110*
- ICDC MODIS, 2013. Arctic melt pond cover fractions were obtained for 2000-2011 from the Integrated Climate Data Center (ICDC, <http://icdc.zmaw.de/>), University of Hamburg, Hamburg, Germany, September 2013. *Cited pages 78 and 82*
- IPCC, 2013. Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T. F., D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P. M. Midgley (eds.)]. in press, cambridge University Press, Cambridge, United Kingdom and New York, NY, USA. *Cited page 2*
- Järvinen, O., Lepparanta, M., 2011. Transmission of solar radiation through the snow cover on floating ice. *Journal of Glaciology* 57 (205), 861–870. *Cited pages 44 and 71*
- Jeffries, M., Li, S., Jana, R., Krouse, H., Hurst-Cushing, B., 1998. Late winter first-year ice floe thickness variability, seawater flooding and snow ice formation in the Amundsen and Ross Seas. *Antarctic Research Series* 74, 69–87. *Cited pages 6 and 110*
- Jeffries, M., Worby, A., Morris, K., Weeks, W., 1997. Seasonal variations in the properties and structural composition of sea ice and snow cover in the Bellingshausen and Amundsen Seas, Antarctica. *Journal of Glaciology* 43 (143), 138–151. *Cited page 19*
- Jordan, R., 1991. A one-dimensional temperature model for a snow cover. Tech. rep., U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Special Report 91-16. *Cited pages 5 and 11*
- Jordan, R., Albert, M., Brun, E., 2008. Physical processes within the snow cover and their parameterization, in *Snow and Climate: Physical Processes, Sur-*

- face Energy Exchange and Modeling, by Armstrong, R. and Brun, E., Cambridge University Press, 12–69.
Cited page 71
- Jordan, R., Andreas, E., Makshtas, A., 1999. Heat budget of snow-covered sea ice at North Pole 4. *Journal of Geophysical Research-Oceans* 104 (C4), 7785–7806.
Cited page 23
- Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Sana, S., White, G., Woollen, J., et al., 1996. The NCEP/NCAR 40-Year Reanalysis Project. *Bulletin-American Meteorological Society* 77, 437–471.
Cited pages 51, 77, and 98
- Kawamura, T., Ohshima, K., Takizawa, T., Ushio, S., 1997. Physical, structural, and isotopic characteristics and growth processes of fast sea ice in Lützow-Holm Bay, Antarctica. *Journal of Geophysical Research* 102 (C2), 3345–3355.
Cited page 19
- King, J., Pomeroy, J., Gray, D., Fierz, C., Föhn, P., Harding, R., Jordan, R., Martin, E., Plüss, C., 2008. Snow-Atmosphere energy and mass balance, in *Snow and Climate: Physical Processes, Surface Energy Exchange and Modeling*, by Armstrong, R. and Brun, E., Cambridge University Press, 70–124.
Cited pages 73 and 117
- Klein, F., Goosse, H., Mairesse, A., de Vernal, A., 2013. Model-data comparison and data assimilation of mid-Holocene Arctic sea-ice concentration. *Climate of the Past Discussions* 9 (6).
Cited page 109
- Krishfield, R., Perovich, D., 2005. Spatial and temporal variability of oceanic heat flux to the Arctic ice pack. *Journal of Geophysical Research* 110 (C07021).
Cited pages 27, 33, and 63
- Kurtz, N., Farrell, S., 2011. Large-scale surveys of snow depth on Arctic sea ice from Operation IceBridge. *Geophysical Research Letters* 38 (20), L20505.
Cited pages 10, 53, 62, and 89
- Kurtz, N., Markus, T., 2012. Satellite observations of Antarctic sea ice thickness and volume. *Journal of Geophysical Research* 117 (C8), C08025.
Cited page 52
- Kurtz, N., Studinger, M. S., Harbeck, J., Onana, V., Farrell, S., 2012, updated 2013. IceBridge Sea Ice Freeboard, Snow Depth, and Thickness. Boulder, Colorado USA: NASA DAAC at the National Snow and Ice Data Center. <http://nsidc.org/data/idsi2.html>.
Cited pages 78, 80, 81, and 108

- Kwok, R., Cunningham, G., 2008. ICESat over Arctic sea ice: Estimation of snow depth and ice thickness. *Journal of Geophysical Research: Oceans* (1978–2012) 113 (C8). *Cited pages 11 and 112*
- Kwok, R., Panzer, B., Leuschen, C., Pang, S., Markus, T., Holt, B., Gogineni, S., 2011. Airborne surveys of snow depth over Arctic sea ice. *Journal of Geophysical Research* 116 (C11), C11018. *Cited pages 10, 53, 58, 62, and 89*
- Kwok, R., Untersteiner, N., 2011. The thinning of Arctic sea ice. *Physics Today* 64 (4), 36–41. *Cited page 100*
- Langlois, A., Mundy, C., Barber, D., 2007. On the winter evolution of snow thermophysical properties over land-fast first-year sea ice. *Hydrological processes* 21 (6), 705–716. *Cited page 47*
- Large, W., Yeager, S., 2004. Diurnal to decadal global forcing for ocean and sea-ice models: The data sets and flux climatologies. NCAR Tech. Note TN-460+STR. *Cited pages 51 and 77*
- Lecomte, O., Fichefet, T., Flocco, D., Schroeder, D., Vancoppenolle, M., 2014a. Impacts of wind-blown snow redistribution on melt ponds in a coupled ocean–sea ice model. Submitted. *Ocean Modelling*. *Cited pages 97, 98, and 99*
- Lecomte, O., Fichefet, T., Vancoppenolle, M., Domine, F., Massonnet, F., Mathiot, P., Morin, S., Barriat, P., 2013. On the formulation of snow thermal conductivity in large-scale sea ice models. *Journal of Advances in Modeling Earth Systems* 5 (3), 542–557.
URL <http://dx.doi.org/10.1002/jame.20039>
Cited pages 70, 71, 74, 77, 79, 80, 88, 89, 93, 98, 99, 101, and 119
- Lecomte, O., Fichefet, T., Vancoppenolle, M., Nicolaus, M., 2011. A new snow thermodynamic scheme for large-scale sea-ice models. *Annals of Glaciology* 52 (57), 337–346. *Cited pages 42, 43, 44, 47, and 77*
- Ledley, T., 1985. Sensitivity of a thermodynamic sea ice model with leads to time step size. *Journal of Geophysical Research* 90 (D1), 2251–2260. *Cited page 46*
- Leonard, K., Maksym, T., 2011. The importance of wind-blown snow redistribution to snow accumulation on Bellingshausen Sea ice. *Annals of Glaciology* 52 (57), 271–278. *Cited pages 73, 74, and 117*
- Levitus, S., 1998. NODC World Ocean Atlas, <http://www.esrl.noaa.gov/psd/>. *Cited pages 43, 51, 63, 71, and 78*

- Loth, B., Graf, H., Oberhuber, J., 1993. Snow cover model for global climate simulations. *Journal of Geophysical Research-Atmospheres* 98 (D6), 10,451–10,464. *Cited page 22*
- Maaß, N., Kaleschke, L., Tian-Kunze, X., Drusch, M., 2013. Snow thickness retrieval over thick Arctic sea ice using SMOS satellite data. *The Cryosphere Discussions* 7 (4), 3627–3674. *Cited page 112*
- Madec, G., 2008. NEMO ocean engine: Note du Pole de Modelisation, Institut Pierre-Simon Laplace.
URL <http://www.nemo-ocean.eu/About-NEMO/Reference-manuals> *Cited pages 42, 70, and 97*
- Maksym, T., Jeffries, M., 2000. A one-dimensional percolation model of flooding and snow ice formation on Antarctic sea ice. *Journal of Geophysical Research: Oceans* (1978–2012) 105 (C11), 26313–26331. *Cited page 6*
- Markus, T., Cavalieri, D., 1998. Snow depth distribution over sea ice in the Southern Ocean from satellite passive microwave data. *Antarctic Research Series* 74, 19–39. *Cited page 11*
- Maslanik, J., Fowler, C., Stroeve, J., Drobot, S., Zwally, J., Yi, D., Emery, W., 2007. A younger, thinner Arctic ice cover: Increased potential for rapid, extensive sea-ice loss. *Geophysical Research Letters* 34 (24). *Cited page 100*
- Massom, R., Eicken, H., Hass, C., Jeffries, M., Drinkwater, M., Sturm, M., Worby, A., Wu, X., Lytle, V., Ushio, S., et al., 2001. Snow on Antarctic sea ice. *Reviews of Geophysics* 39 (3), 413–445. *Cited pages 6, 20, 25, and 47*
- Massom, R., Stammerjohn, S., Smith, R., Pook, M., Iannuzzi, R., Adams, N., Martinson, D., Vernet, M., Fraser, W., Quetin, L., et al., 2006. Extreme anomalous atmospheric circulation in the West Antarctic Peninsula region in austral spring and summer 2001/02, and its profound impact on sea ice and biota. *Journal of Climate* 19 (15). *Cited page 110*
- Massonnet, F., Fichefet, T., Goosse, H., Vancoppenolle, M., Mathiot, P., Beatty, C., 2011. On the influence of model physics on simulations of Arctic and Antarctic sea ice. *The Cryosphere* 5 (3), 687–699. *Cited pages 42, 56, 59, and 63*
- Massonnet, F., Goosse, H., Fichefet, T., Counillon, F., 2014. Calibration of sea ice dynamic parameters in an ocean–sea ice model using an ensemble Kalman filter. In revision. *Cited pages 11 and 112*
- Maykut, G., 1986. The surface heat and mass balance. *Geophysics of sea ice*, edited by N. Untersteiner, NATO ASI Ser., New York, 395–463. *Cited page 19*

- Maykut, G., Untersteiner, N., 1971. Some results from a time-dependent thermodynamic model of sea ice. *Journal of Geophysical Research-Oceans* 76 (6), 1550–1575. *Cited pages 19, 41, and 63*
- Morin, S., Domine, F., Arnaud, L., Picard, G., 2010. In-situ monitoring of the time evolution of the effective thermal conductivity of snow. *Cold Regions Science and Technology* 64 (2), 73–80. *Cited page 49*
- Nicolaus, M., Haas, C., Bareiss, J., Willmes, S., 2006. A model study of differences of snow thinning on Arctic and Antarctic first-year sea ice during spring and summer. *Annals of glaciology* 44 (1), 147–153. *Cited pages 25 and 46*
- Nicolaus, M., Haas, C., Willmes, S., 2009. Evolution of first-year and second-year snow properties on sea ice in the Weddell Sea during spring-summer transition. *Journal of Geophysical Research* 114 (D17109). *Cited pages 10, 20, 25, 34, 45, and 46*
- OSISAF, E., 2010. Global sea ice concentration reprocessing dataset 1978–2007 (v1), <http://osisaf.met.no>. *Cited page 52*
- Pedersen, C., Roeckner, E., Lüthje, M., Winther, J., 2009. A new sea ice albedo scheme including melt ponds for ECHAM5 general circulation model. *Journal of Geophysical Research: Atmospheres* (1984–2012) 114 (D8). *Cited page 69*
- Perovich, D., 2001. Ultraviolet radiation and optical properties of sea ice and snow. *UV Radiation and Arctic Ecosystems* (Ed. D. Hessen) 153, 73–89. *Cited page 19*
- Perovich, D., 2007. Light reflection and transmission by a temperate snow cover. *Journal of Glaciology* 53 (181), 201–210. *Cited page 44*
- Perovich, D., Grenfell, T., Light, B., Hobbs, P., 2002. Seasonal evolution of the albedo of multiyear Arctic sea ice. *Journal of Geophysical Research: Oceans* (1978–2012) 107 (C10), SHE–20. *Cited pages 2 and 69*
- Perovich, D., Richter-Menge, J., Elder, B., Claffey, K., Polashenski, C., 2009. Observing and understanding climate change: Monitoring the mass balance, motion, and thickness of Arctic sea ice. URL <http://imb.crrel.usace.army.mil/> *Cited page 10*
- Petrich, C., Eicken, H., Polashenski, C., Sturm, M., Harbeck, J., Perovich, D., Finnegan, D., 2012. Snow dunes: A controlling factor of melt pond distribution on Arctic sea ice. *Journal of Geophysical Research: Oceans* (1978–2012) 117 (C9). *Cited page 69*

- Polashenski, C., Perovich, D., Courville, Z., 2012. The mechanisms of sea ice melt pond formation and evolution. *Journal of Geophysical Research: Oceans* (1978–2012) 117 (C1).
Cited pages 5 and 69
- Prather, M., 1986. Numerical advection by conservation of second-order moments. *Journal of Geophysical Research: Atmospheres* (1984–2012) 91 (D6), 6671–6681.
Cited pages 4 and 111
- Pringle, D., Eicken, H., Trodahl, H., Backstrom, L., 2007. Thermal conductivity of landfast Antarctic and Arctic sea ice. *Journal of Geophysical Research* 112 (C04017).
Cited pages 2 and 22
- Punge, H., Gallée, H., Kageyama, M., Krinner, G., 2012. Modelling snow accumulation on Greenland in Eemian, glacial inception and modern climates in a GCM. *Climate of the Past Discussions* 8 (2).
Cited page 11
- Roeckner, E., Mauritsen, T., Esch, M., Brokopf, R., 2012. Impact of melt ponds on Arctic sea ice in past and future climates as simulated by MPI-ESM. *Journal of Advances in Modeling Earth Systems* 4 (3).
Cited page 69
- Rösel, A., Kaleschke, L., Birnbaum, G., 2012. Melt ponds on Arctic sea ice determined from MODIS satellite data using an artificial neural network. *The cryosphere* 6 (2), 431–446.
Cited pages 78 and 82
- Rothrock, D., Percival, D., Wensnahan, M., 2008. The decline in arctic sea-ice thickness: Separating the spatial, annual, and interannual variability in a quarter century of submarine data. *Journal of Geophysical Research* 113, C05003.
Cited page 52
- Salas y Mélia, D., 2002. A global coupled sea ice–ocean model. *Ocean Modelling* 4 (2), 137–172.
Cited pages 4 and 12
- Semtner Jr, A., 1976. A model for the thermodynamic growth of sea ice in numerical investigations of climate. *Journal of Physical Oceanography* 6 (3), 379–389.
Cited page 41
- Serreze, M., Hurst, C., 2000. Representation of Mean Arctic Precipitation from NCEP–NCAR and ERA Reanalyses. *Journal of Climate* 13 (1).
Cited pages 98 and 99
- Shine, K., Henderson-Sellers, A., 1985. The sensitivity of a thermodynamic sea ice model to changes in surface albedo parameterization. *Journal of Geophysical Research-Atmospheres* 90 (D1), 2243–2250.
Cited pages 12, 22, 44, 77, 115, and 116

- Shirasawa, K., Leppäranta, M., Saloranta, T., Kawamura, T., Polomoshnov, A., Surkov, G., 2005. The thickness of coastal fast ice in the Sea of Okhotsk. *Cold Regions Science and Technology* 42 (1), 25–40. *Cited page 19*
- Simmons, A., Uppala, S., Dee, D., Kobayashi, S., 2007. ERA-Interim: New ECMWF reanalysis products from 1989 onwards. *ECMWF newsletter* 110 (110), 25–35. *Cited pages 27 and 77*
- Sommerfeld, R., LaChapelle, E., 1970. The classification of snow metamorphism. *Journal of Glaciology* 9 (55), 3–17. *Cited page 23*
- Stössel, A., Kim, S., Drijfhout, S., 1998. The impact of Southern Ocean sea ice in a global ocean model. *Journal of physical oceanography* 28 (10), 1999–2018. *Cited page 35*
- Stranne, C., Jakobsson, M., Björk, G., 2013. Arctic Ocean perennial sea ice breakdown during the Early Holocene Insolation Maximum. *Quaternary Science Reviews*. *Cited page 109*
- Sturm, M., Benson, C., 1997. Vapor transport, grain growth and depth-hoar development in the subarctic snow. *Journal of Glaciology* 43 (143), 42–59. *Cited page 49*
- Sturm, M., Holmgren, J., König, M., Morris, K., 1997. The thermal conductivity of seasonal snow. *Journal of Glaciology* 43 (143), 26–41. *Cited pages 22, 41, 47, 49, 50, 51, and 64*
- Sturm, M., Holmgren, J., Perovich, D., 2002. Winter snow cover on the sea ice of the Arctic Ocean at the Surface Heat Budget of the Arctic Ocean (SHEBA): Temporal evolution and spatial variability. *Journal of Geophysical Research* 107 (C10). *Cited pages 5, 10, 20, 41, 47, 69, 72, and 117*
- Sturm, M., Massom, R., 2009. Snow and Sea Ice, in *Sea ice*, 2nd Ed., by Thomas, D.N. and Dieckmann, G., Wiley-Blackwell, Chapter 5, 153–204. *Cited pages 5, 6, 7, 11, 23, 46, 47, 70, 71, 73, 89, 97, and 117*
- Sturm, M., Morris, K., Massom, R., 1998. The winter snow cover of the West Antarctic pack ice: its spatial and temporal variability. In: *Antarctic Sea Ice Physical Processes, Interactions and Variability*, Ed. M.O. Jeffries, 1–18, American Geophysical Union, Washington, DC. *Cited page 41*
- Taillandier, A., Domine, F., Simpson, W., Sturm, M., Douglas, T., Severin, K., 2006. Evolution of the snow area index of the subarctic snowpack in central Alaska over a whole season. Consequences for the air to snow transfer of pollutants. *Environmental science and technology* 40 (24), 7521–7527. *Cited pages 48 and 49*

- Taylor, K., 2001. Summarizing multiple aspects of model performance in a single diagram. *Journal of Geophysical Research* 106 (D7), 7183–7192.
Cited page 29
- Thorndike, A., Rothrock, D., Maykut, G., Colony, R., 1975. The thickness distribution of sea ice. *Journal of Geophysical Research* 80 (33), 4501–4513.
Cited pages 4 and 119
- Trenberth, K., Large, W., European Center for Medium Range Weather Forecasts, 1989. A global ocean wind stress climatology based on the ECMWF analyses.
URL <http://dss.ucar.edu/datasets/ds110.1/>
Cited pages 51, 77, and 98
- Vancoppenolle, M., Bitz, C., Fichefet, T., 2007. Summer landfast sea ice desalination at Point Barrow, Alaska: Modeling and observations. *Journal of Geophysical Research-Oceans* 112 (C4), C04022.
Cited page 20
- Vancoppenolle, M., Bouillon, S., Fichefet, T., Goosse, H., Lecomte, O., Maqueda, M. M. M., Madec, G., 2012. The Louvain-la-Neuve sea ice model. Tech. rep., Tech. rep., Georges Lemaître Centre for Earth and Climate Research, Proudman Oceanographic Laboratory and Laboratoire d’Océanographie et du Climat.
Cited page 115
- Vancoppenolle, M., Fichefet, T., Bitz, C., 2006. Modeling the salinity profile of undeformed Arctic sea ice. *Geophysical Research Letters* 33 (21), L21501.
Cited page 4
- Vancoppenolle, M., Fichefet, T., Goosse, H., Bouillon, S., Madec, G., Maqueda, M., 2009. Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 1. Model description and validation. *Ocean Modelling* 27 (1-2), 33–53.
Cited pages 4, 12, 19, 42, 70, and 97
- Vancoppenolle, M., Goosse, H., de Montety, A., Fichefet, T., Tremblay, B., Tison, J., 2010. Modeling brine and nutrient dynamics in Antarctic sea ice: The case of dissolved silica. *Journal of Geophysical Research* 115 (C2), C02005.
Cited page 20
- Vancoppenolle, M., Timmermann, R., Ackley, S., Fichefet, T., Goosse, H., Heil, P., Leonard, K., Lieser, J., Nicolaus, M., Papakyriakou, T., et al., 2011. Assessment of radiation forcing data sets for large-scale sea ice models in the Southern Ocean. *Deep Sea Research Part II: Topical Studies in Oceanography* 58 (9), 1237–1249.
Cited pages 62 and 110

- Vionnet, V., Brun, E., Morin, S., Boone, A., Faroux, S., Le Moigne, P., Martin, E., Willemet, J., 2012. The detailed snowpack scheme Crocus and its implementation in SURFEX v7.2. *Geoscientific Model Development* 5, 773–791.
Cited page 48
- Warren, S., 1982. Optical properties of snow. *Reviews of Geophysics and Space Physics* 20 (1), 67–89.
Cited pages 19 and 44
- Warren, S., Wiscombe, W., 1980. A model for the spectral albedo of snow. II: Snow containing atmospheric aerosols. *Journal of Atmospheric Sciences* 37 (12), 2712–2733.
Cited page 5
- Weller, G., 1968. The heat budget and heat transfer processes in Antarctic plateau ice and sea ice.
Cited page 19
- Wilchinsky, A., Feltham, D., 2004. A continuum anisotropic model of sea-ice dynamics. *Proceedings of the Royal Society of London. Series A: Mathematical, Physical and Engineering Sciences* 460 (2047), 2105–2140. *Cited page 4*
- Winton, M., 1999. Polar water column stability. *Journal of physical oceanography* 29 (6), 1368–1371.
Cited page 35
- Wiscombe, W., Warren, S., 1980. A model for the spectral albedo of snow. I: Pure snow. *Journal of Atmospheric Sciences* 37 (12), 2712–2733.
Cited pages 5 and 19
- WMO, 1989. WMO sea-ice nomenclature. Vol. No. 259, Supplement No. 5. World Meteorological Organisation.
Cited page 2
- Worby, A., Geiger, C., Paget, M., Van Woert, M., Ackley, S., DeLiberty, T., 2008a. Thickness distribution of Antarctic sea ice. *Journal of Geophysical Research* 113 (C5), C05.
Cited pages 10, 52, and 53
- Worby, A., Markus, T., Steer, A., Lytle, V., Massom, R., 2008b. Evaluation of AMSR-E snow depth product over East Antarctic sea ice using in situ measurements and aerial photography. *Journal Geophysical Research* 113 (C5), C05.
Cited pages 11 and 59
- Wu, X., Budd, W., Lytle, V., Massom, R., 1999. The effect of snow on Antarctic sea ice simulations in a coupled atmosphere-sea ice model. *Climate Dynamics* 15 (2), 127–143.
Cited pages 41, 97, and 108
- Yen, Y., 1981. Review of thermal properties of snow, ice and sea ice. US Army Cold Regions Research and Engineering Laboratory (CRREL) Report 81-10. Hanover, NH, USA.
Cited pages 22, 41, 47, 51, and 64

- Zhang, J., 2007. Increasing Antarctic sea ice under warming atmospheric and oceanic conditions. *Journal of Climate* 20 (11), 2515–2529. *Cited page 63*
- Zhang, J., 2014. Modeling the Impact of Wind Intensification on Antarctic Sea Ice Volume. *Journal of Climate* 27 (1), 202–214, doi: 10.1175/JCLI-D-12-00139.1. *Cited page 100*
- Zhang, J., Rothrock, D., 2003. Modeling global sea ice with a thickness and enthalpy distribution model in generalized curvilinear coordinates. *Monthly weather review* 131 (5), 845–861. *Cited pages 52 and 79*
- Zunz, V., Goosse, H., Massonnet, F., 2013. How does internal variability influence the ability of CMIP5 models to reproduce the recent trend in Southern Ocean sea ice extent? *The Cryosphere* 7 (2). *Cited page 104*
- Zuo, Z., Oerlemans, J., 1996. Modelling albedo and specific balance of the Greenland ice sheet: calculations for the Sondre Stromfjord transect. *Journal of Glaciology* 42 (141), 305–317. *Cited page 22*
- Zwally, H., Comiso, J., Parkinson, C., Cavalieri, D., Gloersen, P., 2002. Variability of Antarctic sea ice 1979–1998. *Journal of Geophysical Research: Oceans* (1978–2012) 107 (C5), 9–1. *Cited page 3*