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Modelling the mass balance and salinity of Arctic and Antarctic sea ice

Doctoral dissertation presented by
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Tout a été dit cent fois
Et beaucoup mieux que par moi
Aussi, si j'écris des vers
C'est que ça m'amuse
C'est que ça m'amuse
C'est que ça m'amuse...

B. Vian (et R. Binamé)

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Contents

1	Modelling the mass balance and salinity of Arctic and Antarctic sea ice: global summary	1
1.1	Sea ice in a changing climate	2
1.2	Sea ice models	5
1.3	Downscaling issues in sea ice modelling	8
1.4	Models developed and used in this thesis	10
1.5	Objectives of this thesis	11
1.6	Sea ice mass balance	13
1.6.1	Mechanisms of ice growth and melt	13
1.6.2	Ice thickness distribution theory	14
1.6.3	Simulation of the local sea ice mass balance by LIM1D	15
1.6.4	Discussion	16
1.6.5	Simulation of the recent large-scale sea ice mass balance with LIM3	17
1.6.6	Discussion	25
1.7	Salinity variations in sea ice	27
1.7.1	Ice desalination and brine drainage mechanisms.	27
1.7.2	Vertical variations in ice salinity simulated by LIM1D	28
1.7.3	Large-scale variations in ice salinity simulated by LIM3	29
1.7.4	Discussion	30
1.8	Impact of salinity variations on sea ice	31
1.8.1	Local impact of salinity studied with LIM1D	32
1.8.2	Large-scale impact of salinity variations studied with LIM3	33
2	Sea ice halo-thermodynamics in LIM1D	35
2.1	Heat diffusion in the snow-sea ice system	35
2.2	Phase equilibrium in sea ice	36
2.3	Sea ice thermal properties	37

2.4	Thermodynamic transformations of saline ice	39
2.5	Sea ice desalination	40
3	On the sensitivity of undeformed Arctic sea ice to its vertical salinity profile (VFB05)	43
3.1	Introduction	44
3.2	Model description	45
3.3	Model Results	46
3.3.1	Impact of bulk salinity	46
3.3.2	Impact of the shape of the salinity profile	47
3.3.3	Salt/freshwater flux	49
3.3.4	Discussion on application	50
3.4	Conclusions	50
4	Summer landfast sea ice desalination at Point Barrow, Alaska: modeling and observations (VBF07)	53
4.1	Introduction	54
4.1.1	Importance of sea ice and its salinity	54
4.1.2	Observations of ice salinity	55
4.1.3	Modeling heat transfer and salinity variations in sea ice . .	56
4.1.4	Goals of this study	58
4.2	Observation of fast ice desalination at Point Barrow, AK	59
4.2.1	Mass balance	61
4.2.2	Salinity profiles	61
4.3	The model	68
4.3.1	Thermodynamic component	68
4.3.2	Halodynamic component	71
4.4	Model results	77
4.4.1	Mass balance	77
4.4.2	Winter desalination	78
4.4.3	Summer halo-thermodynamics	79
4.4.4	Sensitivity of summer halo-thermodynamics to the representation of the snow in the model.	83
4.4.5	Sensitivity of summer desalination to melt water flushing parameters (i_0 , ϕ , e_T).	86
4.5	Summer melt water budget	88
4.6	Discussion and conclusions	89
4.6.1	Observations of early summer desalination	89
4.6.2	Modeling the winter sea ice salinity profile	90

4.6.3	Modeling the summer Arctic bare ice mass balance and salinity profile	91
4.6.4	Flushing and superimposed ice formation	92
4.6.5	Meltwater balance	93
4.7	Perspectives	93
5	Modeling the salinity profile of undeformed Arctic sea ice (VFB06)	97
5.1	Introduction	98
5.2	The model	99
5.3	Results	101
5.3.1	FY ice salinity variations	101
5.3.2	The MY ice salinity profile and its sensitivity to several model features	103
5.3.3	Impact of a prognostic salinity on ice thickness	104
5.3.4	Salt flux to the ocean	104
5.4	Discussion and conclusions	106
6	Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 1. Model description and validation (V08-1)	109
6.1	Introduction	110
6.2	Model description	112
6.2.1	Ocean general circulation model	113
6.2.2	Ice dynamics	113
6.2.3	Ice state variables	115
6.2.4	Transport	116
6.2.5	Mechanical redistribution	117
6.2.6	Ice thermodynamics	120
6.2.7	Ice age.	125
6.2.8	Ice-ocean coupling	125
6.3	Forcing and experimental design	127
6.3.1	Model grid	127
6.3.2	Simulation setup	127
6.3.3	Model forcing	127
6.4	Results	128
6.4.1	General view of the ice physical state	128
6.4.2	Ice areal coverage	130
6.4.3	Ice thickness and volume	138
6.4.4	Ice drift	144
6.4.5	Snow volume and depth	146

6.4.6	Sea ice age	147
6.4.7	Ice salinity	151
6.5	Discussion and conclusions	152
6.6	Acknowledgements	153
7	Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 2. Sensitivity to salinity processes (V08-2)	155
7.1	Introduction	156
7.2	Ice thickness and salinity distribution theory	158
7.3	One-dimensional fomulation of the thermodynamic changes in ice salinity	159
7.3.1	Thermodynamic component	160
7.3.2	Reference salinity model	160
7.3.3	Simple parameterization of brine entrapment and drainage	161
7.3.4	Experimental setup	161
7.3.5	Results	163
7.3.6	Summary	164
7.4	Representation of the salinity distribution in OPA9-LIM3	165
7.4.1	LIM3 overview	165
7.4.2	Forcing and experimental conditions	167
7.5	Results of the control simulation	167
7.5.1	Mean seasonal cycle of bulk sea ice salinity averaged over the ice thickness categories	167
7.5.2	Ice salinity distribution in thickness space	171
7.5.3	Geographical variations of ice salinity	172
7.6	Sensitivity experiments	172
7.6.1	Northern Hemisphere	174
7.6.2	Southern Hemisphere	183
7.7	Discussion, conclusions and perspectives	190
7.8	Acknowledgements	192
8	Conclusion	193
A	Acronyms	197
	Bibliography	198

Chapter 1

Modelling the mass balance and salinity of Arctic and Antarctic sea ice: global summary



- Ca caille hein ti!
- Tu l'as dit fieu, on aurait mieux fait de rester à la maison et de jouer à l'ordinateur!

Hergé, Destination New York, Collection Jo et Zette, éditions Casterman,
Tournai, 1952.

Sea ice, which refers to any form of ice found at sea which has originated from the freezing of seawater (WMO, 1970) covers about 7% of the global ocean and 4.5% of the Earth surface, on annual average. Due to its important insulating power on the atmosphere-ocean exchanges in polar regions and because it is much thinner than wide, sea ice is considered not only as an important actor in but also as a sensitive indicator of climate change. In addition, sea ice formation and melt are associated with respectively salt and freshwater releases that affect the World Ocean circulation at a global scale. Over the past thirty years, a sharp decrease in Arctic sea ice coverage and volume has been observed from sensors onboard satellites and from upward-looking sonars onboard submarines. The ongoing and future sea ice retreat projected by climate models could also affect life on Earth, as sea ice has an important role on marine ecosystems. For all these reasons, the study of sea ice and its modelling, which is the subject of the present thesis, are increasingly important research areas in the world of geosciences.

1.1 Sea ice in a changing climate

The climate of the Earth is on a track to serious and unprecedented change (Bernstein *et al.*, 2007) characterized by an increase in surface temperatures, a shift in atmospheric and oceanic patterns, an accelerated sea level rise and wide ranging socio-demographic-economic consequences. The warming signal is particularly strong in the Arctic¹ (Serreze *et al.*, 2000), which notably includes sharp changes in the sea ice.

Sea ice is an important actor of the climate system. Sea ice extent² varies spectacularly through the year (Parkinson *et al.*, 1999). In the Northern Hemisphere (NH), sea ice extent is maximum in March ($15 \times 10^6 \text{ km}^2$) and minimum in September ($5 \times 10^6 \text{ km}^2$). In the Southern Hemisphere (SH) (Zwally *et al.*, 2002), it is minimum in February ($4.5 \times 10^6 \text{ km}^2$) and maximum in September ($20 \times 10^6 \text{ km}^2$). In polar regions, the presence of a sea ice cover largely affects the interactions between the atmosphere and the ocean. In particular, due to its high albedo³ (Perovich *et al.*, 2002; Eicken *et al.*, 2004; Brandt *et al.*, 2005) and low thermal conductivity (Pringle *et al.*, 2007), sea ice largely reduces the

¹In the Arctic, the changes observed since the beginning of this century include a decrease in sea ice extent and thickness, a deepening of the permafrost active layer, mountain glacier melting and treeline shifts. They are larger than elsewhere (ACIA, 2004) and are expected to remain large until the end of the present century.

²Sea ice extent is defined as the geographical area covered by at least 15% of sea ice.

³The albedo is defined as the fraction of the incident solar radiation that is reflected back to space.

solar radiation absorbed by the ocean and affects the surface energy budget. In addition, the rejection of salty water during sea ice growth and the release of meltwater during sea ice melt induce sea surface salinity variations that affect the oceanic stratification, which is critical to the circulation of the high-latitude oceans, the formation of deep water masses (Goosse *et al.*, 1997; Goosse and Fichefet, 1999; Stössel *et al.*, 2002) and, in turn, the thermohaline circulation (Aagaard and Carmack, 1989).

Changes in the sea ice involve changes in marine ecosystems and in the major biogeochemical cycles. Sea ice is a unique habitat for a group of bacteria and microscopic plants and animals that are encased in the ice matrix at low temperatures and light levels, with the only liquid being contained in the brine pockets⁴ (Thomas and Dieckmann, 2002). Sea ice and its dynamic biogeochemistry modulate the exchanges of gases (e.g., CO_2) and tracers between atmosphere and ocean. Hence, sea ice has been identified as an important component of several biogeochemical cycles for example carbon (Delille, 2006) and iron (Lannuzel *et al.*, 2007).

Sea ice is considered by the IPCC⁵ as a key indicator of short-term climate change. The large sensitivity of sea ice is due, first, to feedbacks between atmosphere, ice and ocean and, second, to the fact that ice is much thinner (1-10 m) than horizontally wide (>1000 km). The main sea ice-related feedback mechanism is the ice-albedo feedback (see, e.g., Ebert and Curry (1993)). A small positive perturbation in the surface energy budget triggers a decrease in ice coverage. This enhances the absorption of solar radiation in the upper ocean, which further increases melting at the base of the ice cover. Therefore small climate warming perturbation are amplified by sea ice. The understanding of the sea ice mass balance (i.e., ice extent, thickness and velocity) and interannual variations is therefore of crucial importance for the assessment of ongoing climate changes in polar regions and at a global scale (Bamber and Payne, 2003).

Ice coverage is observed for more than 35 years by passive microwave⁶ (PMW) sensors onboard satellites. PMW sea ice cover retrievals show that the Arctic sea ice extent has been decreasing since 1979 by about 3% per decade, on annual average (Cavalieri *et al.*, 2003). The trend is higher in summer and accelerated in the 2000s (-9.1% per decade over 1979-2006, see Stroeve *et al.* (2007)). A downward winter trend is also observed for a few years, now reaching -2.9% per

⁴Brine pockets are the liquid inclusions of water saturated with salt (i.e., brine) that are encased inside the ice crystalline lattice.

⁵IPCC stands for Intergovernmental Panel on Climate Change.

⁶PMW are characterized by a moderate resolution but are practically cloud insensitive and have a wide swath and therefore have an almost global coverage.

decade (Comiso, 2007). In September 2007, the summer Arctic sea ice extent plummeted to the lowest levels since satellite measurements began and shattered the previous record (2005) by more than twenty percent (Comiso *et al.*, 2008). The Arctic ice retreat has been related to changes in longwave radiative fluxes, atmospheric circulation and ocean heat flux (Rigor *et al.*, 2002; Lindsay and Zhang, 2005; Francis *et al.*, 2005; Francis and Hunter, 2007; Polyakov *et al.*, 2005). In the SH, no retreat but instead an insignificant increase in ice extent for the last 30 years is reported, except near the Antarctic Peninsula, where significant warming has occurred during the last half of the past century (Zwally *et al.*, 2002). In 2007, the SH ice extent had a historical maximum.

In the NH, information on ice thickness comes from upward-looking sonars (ULS) onboard submarines (Rothrock *et al.*, 1999; Wadhams and Davis, 2000; Rothrock *et al.*, 2003) and from in-situ electromagnetic sounding (Haas, 2004). Ice freeboard is now measured by laser and radar altimetry (Kwok *et al.*, 2007; Laxon *et al.*, 2003). Nevertheless, ice thickness cannot yet reliably be estimated because of the lack of reliable remote snow depth observations. In the Arctic, the radiative footprint of snow, very close to multi-year (MY) ice, currently prevents to retrieve snow depth, and in turn, to reliably convert freeboard into ice thickness (Kwok *et al.*, 2007). In order to fill in this gap, algorithms of snow depth retrieval are in their early stages of development (e.g., Langlois *et al.* (2007)). Available NH ice thickness data consistently point to a significant decrease in ice draft⁷ between 22.5 % and 32 % in several regions of the Arctic basin in the 1990s (Rothrock *et al.*, 2003; Haas, 2004). The recent ice volume downward trend and its significance are not firmly established because of high interannual and interdecadal variability indicated by altimeter data (Laxon *et al.*, 2003) and historical landfast ice⁸ measurements (Polyakov *et al.*, 2003). Large-scale ice-ocean models running over the last 30 years (Fichefet *et al.*, 2003; Bitz and Roe, 2004; Lindsay and Zhang, 2005; Rothrock and Zhang, 2005) all simulate a decrease in ice volume consistent with observations.

In the SH, knowledge on ice thickness is limited. Information comes from a few moored ULS in the Weddell Sea (Strass and Fahrbach, 1998) as well as from the Antarctic Sea ice Processes and Climate (ASPeCt) climatology. The latter dataset derives from ship-based visual observations and has large-scale coverage (Worby *et al.*, 2008). Nevertheless, the lack of SH ice thickness data will presumably be resolved within a few years. In the Southern Ocean, remote snow depth data do exist (Markus and Cavalieri, 2006) and the first large-scale dataset of freeboard-derived ice thickness has been published recently, for the Weddell Sea

⁷Sea ice draft is the thickness of the part of the ice that is submerged under the sea.

⁸Landfast sea ice is sea ice which remains fast along the coasts, where it is attached to the shore (WMO, 1970).

(Zwally *et al.*, 2008). Contrary to the Arctic, snow depth is observable from space (Markus and Cavalieri, 1998, 2006) and therefore ice thickness can be computed from freeboard observations, deriving either from ICESat (laser) or ERS (radar) altimeters (Zwally *et al.*, 2008; Giles *et al.*, 2008).

The high variability in the sea ice characteristics, the relative shortness (30 years) of the satellite sea ice records, as well as the sparsity of historical data and proxies raises the question of the significance of the recently observed changes in sea ice. Among the historical records, the Divine and Dick (2006) reconstruction of ice edge position in the Nordic Seas from whaling boats indicates persistent ice retreat from 1850, with significant superimposed interdecadal variability (with periods of 20–30 y and 60–80 y). This supports the hypothesis of a climate sensitive sea ice but also raises the variability issue. Longer-term sediment-based proxies from both hemispheres (Crosta *et al.*, 1998; Sarthein *et al.*, 2003; de Vernal *et al.*, 2005) indicate large-amplitude sea ice variations associated to past climate variations.

1.2 Sea ice models

Understanding the past and projecting the future variations of the Arctic and Antarctic sea ice covers is crucial to perform reliable projections of future climate. Sea ice models are useful tools to simulate and understand mechanisms regulating past sea ice variations. Included in climate models, they enable projections of the future ice conditions. Numerical sea ice models are either one-dimensional (1D) or three-dimensional (3D).

Because heat transfer preferentially occurs along the vertical direction, the problem of heat transfer and storage in sea ice can be formulated one-dimensionally. **1D sea ice models** are designed to resolve this problem at local scales. Also called thermodynamic models, they usually neglect dynamical and oceanic processes. Under prescribed atmospheric and oceanic forcings, the vertical temperature field is computed by solving the heat diffusion equation in the ice-snow system. Boundary conditions state how to convert imbalances between internal (conductive) and external (atmospheric and oceanic) heat fluxes into ice growth or melt. Integrating the model over time, the evolution of ice thickness, snow depth and other ice characteristics is derived.

Maykut and Untersteiner (1971) developed the first thermodynamic sea ice model, which was then simplified in the so-called 3-layer and 0-layer models (Semtner, 1976). The Maykut and Untersteiner model then underwent several improve-

ments, as rigorous heat conservation (Bitz and Lipscomb, 1999), inclusion of a melt pond scheme (Taylor and Feltham, 2004), as well as an enthalpy-based formulation (Huwald *et al.*, 2005a; Notz and Worster, 2006). In parallel, Cox and Weeks (1988) introduced an empirical formulation of the winter sea ice desalination in the 0-layer model.

1D models have been used to simulate the landfast ice growth and decay (Flato and Brown, 1996), to study the sensitivity of sea ice to internal parameters and external forcing (Maykut and Untersteiner, 1971; Eisenman *et al.*, 2007) and to identify several ice-related feedbacks (Ebert and Curry, 1993).

3D sea ice models⁹ are more complex and computationally expensive than 1D models but give more interesting results. 3D models are used to simulate the geographical variations of sea ice characteristics and understand large-scale sea ice dynamics. Under time-varying atmospheric and oceanic forcings, they simulate ice coverage, volume and drift.

Present 3D sea ice models are based on the same dynamic-thermodynamic structure, discretized using finite differences on structured grids. Typically, the model thermodynamic component is based on the Semtner (1976) 0- or 3-layer model, or on the more sophisticated model of Bitz and Lipscomb (1999). The dynamic component solves the ice momentum equation assuming that sea ice is a two-dimensional continuum with a viscous-plastic (VP) rheology, using either the original VP numerical method (Hibler, 1979), the elastic-viscous-plastic (EVP) method (Hunke and Dukowicz, 1997) or a semi-implicit alternate method (Zhang and Hibler, 1997). An advection scheme transports ice state variables on the model grid (e.g., Prather (1986)). In addition, some models include the subgrid-scale ice thickness distribution (ITD) physics (Thorndike *et al.*, 1975) with various physical (Flato and Hibler, 1995; Zhang and Rothrock, 2001) and numerical (Lipscomb, 2001) improvements. 3D models can be used either in a regional (see, e.g., Maslowski and Lipscomb (2003)) or a global configuration, which was the one adopted in Louvain-la-Neuve (see, e.g., Timmermann *et al.* (2005)). The size of model grid cells range from ten kilometers in regional models to several hundreds of kilometers in large-scale models. 3D model development is limited by the current understanding of ice physics, by the lack of validation data and by computational power.

⁹Large-scale sea models ice are actually not fully three-dimensional. They actually combine a dynamic (two-dimensional) and thermodynamic (one-dimensional) components. Nevertheless, we will use the word "three-dimensional" throughout this text in order to refer to them.

3D models applications include sea ice hindcasts, forecasts and climate projections. Hindcasts are simulations of the past sea ice conditions forced by available atmospheric information. Forecasts are short-term previsions of the future sea ice state, with timescales from weeks to months. They are the equivalent of meteorological forecasts in atmospheric sciences. Climate projections are assessments of the mean and variability of future climate, which include future sea ice conditions, given assumptions on emissions of greenhouse gases. Since there sea ice forecasts are only in early stages of development, only hindcasts and projections will be developed hereafter.

Hindcasts are performed either with 3D sea ice models in isolation (e.g., Flato and Hibler (1995)) or coupled with oceanic models. Sea ice hindcasts are forced with atmospheric reanalyses (e.g., NCEP-NCAR¹⁰ reanalyses, see Kalnay et al., 1996) and give reconstructions of past ice conditions, typically for the last 50 years. Hindcasts are useful to validate the ice-ocean models against satellite data (Timmermann *et al.*, 2005). Furthermore, they generate information on ice volume otherwise unknown. For example, all ice-ocean models show a decrease in ice volume in the 1990s (Rothrock *et al.*, 2003). They also can give information on the role of atmospheric forcing on interannual and interdecadal variability (Köberle and Gerdes, 2003; Fichefet *et al.*, 2003; Lefebvre *et al.*, 2004). Nevertheless, hindcasts are tainted by errors in the model and forcing.

Climate projections are simulations of future climate performed with a series of global, coupled climate models, including at least atmosphere, land, sea ice and ocean components, in the framework of the IPCC assessments. In the "Climate of the twentieth century experiment", models are forced by observed greenhouse gases and aerosol concentrations as well as by variations in solar and volcanic activity. To simulate future climate, they are forced by several emission scenarios. Emission scenarios describe how greenhouse gases emissions could evolve between 2000 and 2100, depending on various socio-economical hypotheses.

The large-scale sea ice models developed in the last decades (e.g., Fichefet and Maqueda (1997); Bitz *et al.* (2001); Hunke and Lipscomb (2004)) are included in the coupled models used to perform the climate projections (see, e.g., Arzel *et al.* (2006); Zhang and Walsh (2006)). In the Arctic, only a few models reasonably simulate the geographical distribution of ice concentration for the past 30 years and all of them underestimate the observed NH downward trend in ice extent (Stroeve *et al.*, 2007). In the SH, all the models have strong deficiencies (Lefebvre and Goosse, 2008). The large remaining uncertainties in sea ice simulations for

¹⁰NCEP (National Center for Environmental Prediction) and NCAR (National Center for Atmospheric Research) are both located in Boulder, Colorado, USA.

the 20th and 21st centuries stress the need for further model development and validation.

1.3 Downscaling issues in sea ice modelling

The evolution of sea ice is driven by physical processes, whose governing equations are so complex that they can only be solved through the use of numerical models. These processes occur over a wide range of scales, from the microscopical to the geophysical scale, as summarized in Figure 1.1. Current computer power constrains the model grid cells to be larger than the smallest pieces of ice (called "floes"). Consequently, the smallest-scale processes have to be parameterized. In the present work, two downscaling issues were studied with particular attention. The first one is related to subfloe-scale variations in ice thickness. The other one is devoted to the representation of the thermal effect of brine pockets.

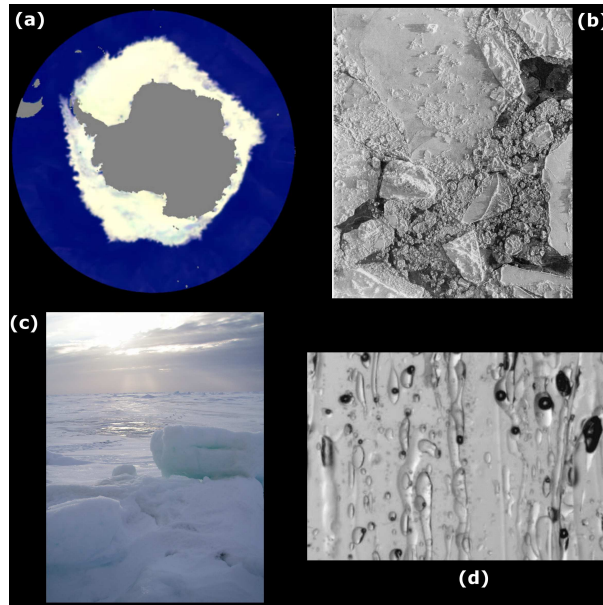


Figure 1.1: Illustration of the various scales of sea ice. (a) Large-scale image of Southern Ocean sea ice cover derived from PMW recordings (Source: www.nsidc.org). (b) Aerial picture of the fracturing of Arctic sea ice by Steffen (1986), 160 m x 200 m. (c) Old pressure ridge in the Laptev Sea, September 2005 (picture by M.V.). (d) Picture of the microstructure of artificial sea ice by Light *et al.* (2003) at -15 °C, using a cold-specialized microscope. Width: 5 mm.

Ice thickness may vary from zero to about twenty meters, even at subfloe scales (see Figure 1.2, panel a). Accounting for ice thickness variations is important because, first, the ice growth rate depends nonlinearly on ice thickness particularly for thin ice (see Figure 1.2, panel b). Second, the rapid disparition of thin ice in early summer drives a significant reduction of the surface albedo. Subgrid-scale variations in ice thickness are represented in models through the use of an **ice thickness distribution** (ITD) function, which describes the relative surface coverage of sea ice of different thicknesses. Changes in the ITD are due to thermodynamic (growth and melt) and dynamic (opening and deformation) effects. Models including an interactive ITD improve the representation of ice growth and of the summer ice-albedo feedback, which in turn significantly improves the simulation of the large-scale sea ice characteristics (Holland *et al.*, 2006b). This suggests that a representation of ITD physics is required in the sea ice models used for global climate simulations.

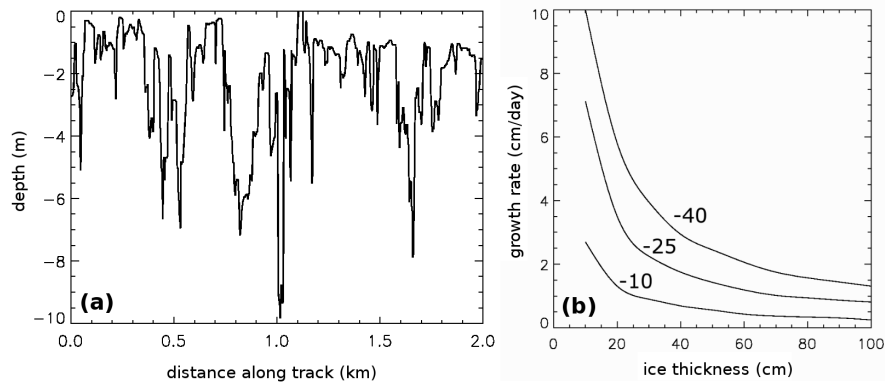


Figure 1.2: (a) Ice draft profile from under-ice submarine sonar ice thickness sounding, Beaufort Sea, October 2000, data from Wensnahan *et al.* (2007). (b) Growth rate of young sea ice as a function of ice thickness and air temperature, for air temperatures of -10, -25 and -40°C. Redrawn from Anderson (1961) and Maykut (1986).

The second downscaling issue is linked to the **brine inclusions** (see Figure 1.1, panel d). Sea ice is a saline medium. A fraction of oceanic salt is trapped in the ice during formation, while the remainder is rejected into the ocean. The salt in sea ice is dissolved within liquid inclusions of brine. Due to brine entrapment and drainage, the ice salinity varies in time (Cox and Weeks, 1974; Weeks and Ackley, 1986).

The ice salinity and the associated brine inclusions affect the sea ice mass balance and the ice-ocean salt / freshwater exchanges. The relative volume of brine inclusions depends on ice salinity and temperature. Consequently, there are, inside sea ice, permanent variations in brine volume, which consume or release large amounts of heat. Hence, the sea ice thermal properties strongly depend on salinity and temperature. In turn, this influences the seasonal cycle of ice growth and decay. The sea ice salinity also affects salt and fresh water fluxes associated to ice growth and melt as well as to brine drainage.

Presently, brine drainage mechanisms and salinity variations are not included in sea ice models: the salinity is either ignored or, at best, prescribed. The role of ice salinity on the ice-ocean system through changes in the sea ice thermal properties and in the ice-ocean salt / freshwater flux has never been assessed. Consequently, the errors due to the poor representation of the sea ice salinity in models are presently unknown.

1.4 Models developed and used in this thesis

*In the framework of this thesis, two sea ice models were developed, based on earlier work done in Louvain-la-Neuve. The first one, **LIM1D**¹¹, is a 1D sea ice model that has a complete representation of brine physics. The second one, **LIM3**, is a large-scale 3D dynamic-thermodynamic sea ice model with subgrid-scale distribution of its main state variables.*

LIM1D is a 1D halo-thermodynamic sea ice model. The halodynamic module computes the vertical salinity profile based on empirical formulations of brine entrapment during growth, brine expulsion and gravity drainage (Cox and Weeks, 1988) and a physical formulation of flushing (newly developed). The thermodynamic module is based on Bitz and Lipscomb (1999). In order to represent the salinity effect on heat transfer and storage through its impact on the sea ice relative brine volume, the sea ice thermal properties (i.e., the specific heat c , the thermal conductivity k and the energy of melting q , defined as the energy required to melt a unit volume of sea ice) depend on S and T .

LIM is originally a large-scale 3D dynamic-thermodynamic sea ice model developed by Fichefet and Morales Maqueda (1997). The thermodynamic component is the Semtner (1976) 3-layer model and the dynamic component is the Hibler

¹¹LIM (Louvain-la-Neuve sea Ice Model) is the sea ice model developed at the Institut d'Astronomie et de Géophysique G. Lemaître.

(1979) viscous-plastic (VP) rheology. **LIM** includes various parameterizations (open water, snow, snow ice formation, increased heat conduction through thin ice). The basis of the coupling with ocean models was introduced by Goosse and Fichefet (1999). **LIM** was then coupled to the ocean model OPA¹² v8.2, leading to the ice-ocean model configuration ORCA2-LIM¹³ (Timmermann *et al.*, 2005). The code was then rewritten formally without significant physical changes (work done by Christian Ethé and Gurvan Madec), which resulted in **LIM2**.

Based on **LIM2**, the new sea ice model developed in the framework of this thesis has logically taken the name of **LIM3**. **LIM3** is a 3D large-scale sea ice dynamic-thermodynamic model with subgrid-scale distributions of its main state variables (snow volume and enthalpy, ice area, volume, enthalpy, age content and salt content, all expressed per unit area). **LIM3** is coupled to OPA v9 and is part of the NEMO¹⁴ ocean modelling system. **LIM3** includes three major new developments. First, the C-grid elastic-viscous-plastic (EVP) rheology (S. Bouillon *et al.*, The Elastic-Viscous-Plastic rheology on different grid types, manuscript in preparation, hereafter referred to as Bouillon *et al.*, 2008) replaces the former VP formulation of Hibler (1979). EVP gives a better solution of the ice momentum equation, allows easier parallelization and the C-grid allows the dynamical coupling to OPA (also in C-grid) in a natural way. Second, in order to account for unresolved variations in ice thickness, several thickness categories have been included into the model. Ice volume is redistributed among categories due to thermodynamic and dynamic processes. Finally, a multi-layer halo-thermodynamic component replaces the former 3-layer model. This includes an explicit representation of brine entrapment and drainage, as well as the impact of brine on sea ice growth and decay. **LIM3** includes also several other parameterizations of processes of minor importance that are not described here.

1.5 Objectives of this thesis

The present thesis is in the line of the sea ice model development effort reviewed above, with a climate simulation perspective. More specifically, I tried to improve the understanding and the formulation in sea ice models of the interactions between the sea ice mass balance and the salinity variations. My contribution addressed three questions :

¹²OPA (Océan Parallélisé) was developed at LOCEAN, Université de Paris VI, Paris, France.

¹³ORCA2-LIM is a configuration of OPA on a $2^\circ \times 2^\circ$ tripolar grid.

¹⁴NEMO stands for Nucleus for European Modelling of the Ocean.

- What are the main features of the present-day sea ice mass balance and the ability of 1D and 3D sea ice models to simulate them ?
- How does the sea ice salinity evolve in space and time. How to simulate salinity variations ?
- What is the impact of the sea ice salinity variations on the sea ice mass balance?

In order to address these questions, a significant part of the present work was to develop **LIM1D** and **LIM3**. The results obtained throughout this thesis have been grouped in the next three sections of the present chapter. Each of the sections specifically covers one of the scientific questions. Insight gained with **LIM1D** and **LIM3** is used to address each question at local and large scales, respectively. In Chapter 2, the theoretical framework of the halo-thermodynamic formulation of **LIM1D** is described.

Chapters 3–7 cover the details of the scientific topics, corresponding to a series of scientific articles¹⁵. The first three papers describe the 1D studies. In "*On the sensitivity of undeformed Arctic sea ice to its vertical salinity profile*" (Vancoppenolle *et al.*, 2005) (hereafter **VFB05**), the importance of the vertical salinity profile on ice growth and decay is assessed. Then, in "*Summer landfast sea ice desalination at Point Barrow, Alaska: Modelling and observations*" (Vancoppenolle *et al.*, 2007) (**VBF07**), the **LIM1D** halodynamic¹⁶ module is described and validated against observed landfast sea ice characteristics over a seasonal cycle. In "*Modeling the salinity profile of undeformed Arctic sea ice*" (Vancoppenolle *et al.*, 2006) (**VFB06**), we study, first, the variations of the salinity profile simulated by **LIM1D** over a 50-year equilibrium simulation and, second, the sensitivity to salinity variations of the simulated ice thickness and ice-ocean salt fluxes.

The two other papers document the 3D studies and compose the two parts of a single contribution: "*Simulating the mass balance and salinity of Arctic and Antarctic sea ice*". In the first one, "*Model description and validation*" (Vancoppenolle *et al.*, 2008c) (**V08-1**), OPA9-**LIM3** is described and its results over a 1979-2006 hindcast are thoroughly compared to available in-situ and satellite data (snow depth and ice concentration, thickness, salinity and drift) and to outputs from other models. In the second part, "*Sensitivity to salinity processes*" (Vancoppenolle *et al.*, 2008a) (**V08-2**), the 3D parameterization of salinity variations

¹⁵The references, section and figure numbers in bold characters refer to the publications and manuscripts that constitute the chapters of the present thesis.

¹⁶The halodynamic module is the section of the code in which salinity variations are computed.

used in **LIM3** is validated against ice core data. In addition, the sensitivity of the simulated sea ice mass balance to the representation of the salinity of sea ice is investigated. Chapter 8 is dedicated to the conclusions of the present thesis.

1.6 Sea ice mass balance

*In this section, we first review the sources and sinks of ice mass and outline how they are represented in the thickness distribution theory. Then, we describe how **LIM1D** and **LIM3** simulate the sea ice mass balance at local and geophysical scales, respectively.*

1.6.1 Mechanisms of ice growth and melt

Changes in sea ice mass are controlled by thermodynamic and dynamic processes. By convention, thermodynamic processes include all the process by which sea ice grows and melts. Initially, when seawater has reached its freezing point, small frazil crystals start to grow. In calm waters, the frazil crystals can coalesce and form a thin ice layer. In wavy, turbulent waters, the frazil mixes with slush and forms rounded pieces of ice (called pancakes). These pancakes will form a consolidated layer only later, once the wave influence on ice will be small enough not to break the bonds between them (Lange *et al.*, 1989; Shen *et al.*, 2004).

Once an ice layer has formed, the ice grows by congelation at the ice base if the heat conduction flux is directed upwards and is higher than the oceanic heat flux (Maykut and Untersteiner, 1971). In winter, in open water areas, frazil ice keeps on forming and coalesces at the edges of the floes (Biggs *et al.*, 2000). If the snow is deep enough to depress the snow-ice interface under the sea level, seawater infiltrates the snow, and if the heat loss is sufficient, seawater freezes into the snow, which forms the highly saline *snow ice* (Markus and Maksym, 2008). In summer, snow melts on top of the ice. The resulting meltwater percolates through snow and may refreeze within the snow and, if the ice is impervious, on top of the saline ice surface (which is colder than 0°C, the freezing temperature of pure water). This forms the fresh *superimposed ice* (Gerland *et al.*, 1999; Haas *et al.*, 2001). Ice melts at the ice surface or at its base under a convergence of heat (Maykut and Untersteiner, 1971; Perovich *et al.*, 2003). Lateral melting was also reported in several instances (Steele, 1992; Perovich *et al.*, 2003).

Dynamic processes include opening, ridging and rafting. Under a local divergence of the ice velocity field, the ice pack opens, which creates leads (Kwok and Cunningham, 2002). Along the coasts and islands, wind blowing from the coasts creates coastal (latent heat) polynyas¹⁷. Offshore, sensible polynyas, thermodynamically driven, appear as a result of oceanic sensible heat entering the area of polynya formation in amounts large enough to melt any pre-existing ice and prevent the growth of new ice (Maqueda *et al.*, 2004). Under a local convergence of the ice velocity field, if the ice is thin, the ice plates override on top of each other (rafting), which approximately doubles the ice thickness. If the ice is thick, or more irregular, the initial ice cover breaks into small pieces, which pile together into pressure ridges. The necessary conditions for the ice to ridge or raft (Tuhkuri and Lensu, 2002) are not yet fully understood.

1.6.2 Ice thickness distribution theory

In order to account for unresolvable variations in ice thickness (h) in large-scale sea ice models, Thorndike *et al.* (1975) proposed to introduce a distribution function $g(h)$ for ice thickness in a given region defined by :

$$\int_{h_1}^{h_2} g(h)dh = \frac{A(h_1, h_2)}{R}, \quad (1.1)$$

where R is the total area of the region of interest, $A(h_1, h_2)$ is the area in R covered by ice with thickness lying between h_1 and h_2 . Invoking continuity (i.e., assuming that $g(h)$ is valid for any subregion of R), Thorndike *et al.* (1975) argued that the conservation of ice mass can be written in the following way :

$$\frac{\partial g}{\partial t} = -\nabla \cdot (g\mathbf{u}) - \frac{\partial}{\partial h}(fh) + \psi^g, \quad (1.2)$$

where $\mathbf{u}$ is the 2-D horizontal ice velocity vector, f is the thermodynamic vertical growth rate (m/s) and ψ^g is the so-called mechanical redistribution function. Functional of h and $g(h)$, it determines the changes in $g(h)$ during a deformation event, which redistributes ice area between the different thickness categories. Equation (1.2) is forced with $\mathbf{u}$, f and the strain rate tensor. The terms on the right-hand side of (1.2) refer to advection by the velocity field, advection in thickness space by the thermodynamic processes and mechanical redistribution by dynamic processes, respectively.

¹⁷A polynya is an open water feature inside an ice pack with a nonlinear shape.

1.6.3 Simulation of the local sea ice mass balance by LIM1D

In **LIM1D**, equation (1.2) is not solved. f – which accounts for basal growth, surface and basal melt, as well as snow ice formation – is integrated over time to obtain ice thickness, given initial conditions and prescribed time series of atmospheric and oceanic heat fluxes. There is no account for superimposed ice formation in **LIM1D**. Several results that are of direct interest to the study of the sea ice mass balance were obtained with **LIM1D**.

Landfast Arctic first-year (FY) ice was studied at Point Barrow, on the northern Alaskan coast (see **VBF07**, **Fig 4.1**, for a map). The simulated seasonal cycle of ice thickness and vertical temperature profile are in good agreement with observations. The simulation is very realistic if the snow cover is thin. The model reproduces the observed winter basal ice growth and the summer surface and bottom melt. The simulations suggest that sea ice stores 25% of the meltwater produced at the surface during the first 20 days of summer melt. This is coherent with the observed freshwater input in the upper ocean.

Though **LIM1D** does not explicitly include it, the model behaviour and sensitivity experiments suggest that the formation of superimposed ice is conditioned by the snow depth and by the permeability structure of the upper ice layers (see **VBF07**, **sect. 4.6.4**). First, a deeper snow cover induces more potential fresh meltwater production able to refreeze at the ice surface. Second, the early summer period in the Arctic is characterized by relatively high (low) atmospheric (oceanic) heat input. In these conditions, deep snow thermally insulates the upper ice layers, which keeps them cold and impervious. This favours storage and refreezing of the percolating snow meltwater within the snow and at the ice surface.

In addition, idealized Arctic MY ice was simulated by applying the same climatological forcing typical of central Arctic conditions (Fletcher, 1965) over 50 years (see **VFB05** and **VFB06**). **LIM1D** was found to stabilize after 50 years to an annual mean thickness of 3.3 m, which is realistic. Other case studies performed with **LIM1D** include the use of in-situ data for Arctic MY and Antarctic FY ice¹⁸.

In conclusion, **LIM1D** manages to simulate the observed seasonal cycle of sea ice thickness (within around 10 cm) and the vertical temperature profile (within

¹⁸In additional experiments which are not part of the present work, **LIM1D** was used to simulate (1) the seasonal cycle of Arctic MY ice growth and decay (Beaufort Sea), using SHEBA data (Vancoppenolle and Fichefet, 2005), and (2) the decay of thick FY ice in the Weddell Sea, using ISPOL data. ISPOL (Ice Station POLarstern) was a drift station in the Weddell Sea in December 2004 (Hellmer *et al.*, 2006; Tison *et al.*, 2008).

around 1°C) over a wide range of conditions. These 1D simulations increase confidence in **LIM1D** which can therefore be used in a 3D model.

1.6.4 Discussion

Several problems remain in the understanding of the sea ice mass balance at local scales and the associated 1D models. The first problem is the presence of spatial variability in the validation data, the second one is the snow cover and the third one is associated to the longwave (LW) and oceanic forcings.

Spatial variability. Sea ice characteristics may vary substantially up to subfloe scales. The instruments (or sampling methods) used to derive the different variables (meteorological mast, ice thickness gauges, thermistor strings, ice cores) cannot, for technical reasons, be set up (or taken) at the same location. Therefore, despite a lot of precautions, the different components of the 1D datasets used for model validation are tainted with spatial variability (see, e.g. Huwald *et al.* (2005b)). One possibility to overcome spatial variability is to average data from several locations, taken at the same time, in order to simulate a "mean" sea ice column, typical of a given floe. Nevertheless, such a concept is not realistic. For example, it is not possible to find the right way to compute the mean vertical temperature profiles over ice of different thicknesses. In addition, the small-scale horizontal variations in ice thickness and snow depth result in large variations in ice temperatures and ice growth. In summary, it is not possible to reach a perfect 1D model-observation agreement. To overcome this problem, I suggest two solutions. First, instead of the single-column setup, running a series of simulations covering the 2D ice-snow thickness space, which allows to resolve the spatial variability in the simulations, provides an improved framework for model-observation comparison (see, e.g., Lindsay (2003)). Second, as they allow to collect time series on a single location, non-destructive observation methods reduce the inherent spatial variability in the validation datasets (see, e.g., Notz *et al.* (2005)).

Snow. The representation of snow in **LIM1D** is approximate and deteriorates the sea ice simulation. In **winter**, snow depth has to be forced. This is because, first, snow gauges that are used to perform snowfall measurements underestimate snowfall significantly, by wind undercatch, wetting and evaporation losses (Goodison *et al.*, 1998). In addition, the link between snowfall and snow depth is not understood (Massom *et al.*, 2001). Finally, because of the effect of the wind on snow resuspension and deposition, snow is fundamentally 3D and cannot be simulated in a 1D model. In **summer**, basic snow physics are missing in **LIM1D** (meltwater percolation, superimposed ice formation, metamorphism and radiative

transfer, see **VBF07, sect. 4.4.4**). In turn, this overestimates snow melt, anticipates the sea ice melt onset and overestimates surface melt. Finally, the range of the values for the thermal conductivity of the snow found in the literature is very large¹⁹ (0.14-0.50 W/m/K) and models are very sensitive to it (Fichefet *et al.*, 2000).

Forcing. There are two main problems in the fields used to force 1D models: the incoming LW radiation and the oceanic heat flux. Observations of radiative fluxes, including incoming LW, are not always available. Shortwave (SW) and LW fluxes are extremely sensitive to cloud cover (Gorodetskaya and Tremblay, 2008). SW flux can be reasonably estimated using astronomical formula (Zillmann, 1972) and a cloud cover climatology. But, though the LW formula we used could be optimized, biases remain, as LW radiation depends on cloudiness, cloud type and depth (Key *et al.*, 1996; Guest, 1998). In **LIM1D**, incoming LW radiation has to be tuned for the model to match the observed winter ice growth rate and date of melt onset.

The second major source of error in the forcing is the oceanic heat flux. First, the spatio-temporal variations of the oceanic heat flux are poorly known, except a study by Krishfield and Perovich (2005). In addition, simulation at the SHEBA site has shown that using in-situ observations of the oceanic heat flux²⁰ leads to substantial errors in the simulated basal growth and melt. Therefore, in 1D simulations, the oceanic heat flux has either to be prescribed or to be reconstructed from basal growth and temperature profile data. These reconstructions (Perovich and B. Elder, 2002; Huwald *et al.*, 2005b) make assumptions on the sea ice properties (latent heat, ice thermal conductivity) that, if coherent with model parameters, warrant model-observation agreement but do not exclude errors in ocean-ice heat transfer.

1.6.5 Simulation of the recent large-scale sea ice mass balance with LIM3

In **LIM3**, equation 1.2 is solved. f accounts for new (frazil) ice growth in open water, congelation growth at the ice base, snow ice formation, surface melt and bottom melt (see **V08-1, sect. 6.2.6.3**). There is no account for the pancake

¹⁹See Huwald *et al.* (2005b), section 3.3, for a detailed discussion on the topic.

²⁰Oceanic heat flux is measured by applying the eddy covariance method to oceanic velocity measurements (see., e.g., McPhee (1992)).

cycle, superimposed ice formation and lateral melting. ψ^g accounts for opening, porous ridging and rafting (see **V08-1, sect. 6.2.5**)²¹.

In **V08-1, sect. 6.4**, a simulation with OPA9-LIM3 of the Arctic and Antarctic sea ice mass balance over the period 1970-2006 is presented. Thorough comparisons of the simulation results to available observations of ice areal coverage, thickness, texture, drift, age and snow depth as well as to other models are performed.

In the Arctic, the model provides a very good simulation of the large-scale sea ice characteristics. The sources and sinks of ice mass (see Fig. 1.3 and **V08-1, sect. 6.4.3.3**) are in agreement with information from textural analysis of ice cores (see, e.g., Tucker *et al.* (1987); Perovich *et al.* (2003)). In LIM3, the ice thickness increases mostly by basal congelation growth and melts in comparable amounts from top and bottom interfaces. New ice growth in open water contributes in the seasonal ice zones (SIZ)²², while snow ice formation is confined to the ice edge because this region combines thin ice and large oceanic heat flux and snowfall.

In the NH, using LIM3 considerably improves the simulation of the sea ice characteristics compared to LIM2 (see Vancoppenolle *et al.* (2008b)). OPA9-LIM3 produces mean sea ice coverage (see Figure 1.4) and thickness fields that compare significantly better to available observations than OPA9-LIM2.

In LIM3, the mean seasonal cycle and interannual variations of ice area, as well as the mean geographical distribution of ice concentration, are rather realistic compared to observations derived from satellite recordings (Comiso, 2007). The difference between simulated and observed Arctic sea ice extents averaged over 1979-2006 is $-0.51 \times (0.71)10^6$ km² with LIM3 (LIM2). In winter, both models simulate well the geographical distribution of ice concentration, though LIM2 overestimates ice concentration near the ice edge. In summer, LIM2 overestimates the ice coverage in the seasonal ice regions, in particular along the east coast of Greenland and on the Siberian shelf. In contrast, the LIM3 simulation is much more realistic. In LIM3, the ice-albedo feedback is governed by the rapid melt of thin ice and can be characterized as follows. In early summer, thin ice in the marginal ice zones disappears quickly, which significantly reduces the ice concentration compared to LIM2 and promotes higher absorption of shortwave

²¹Rafting and porous ridging are newly developed features included in the model. Porous ridging is found of minor importance, while rafting is important in the SH. See **V08-1, sect. 6.2.5.5** for details on their role.

²²The seasonal ice zones are the regions which are covered seasonally, i.e., only in winter and from which ice retreats in summer.

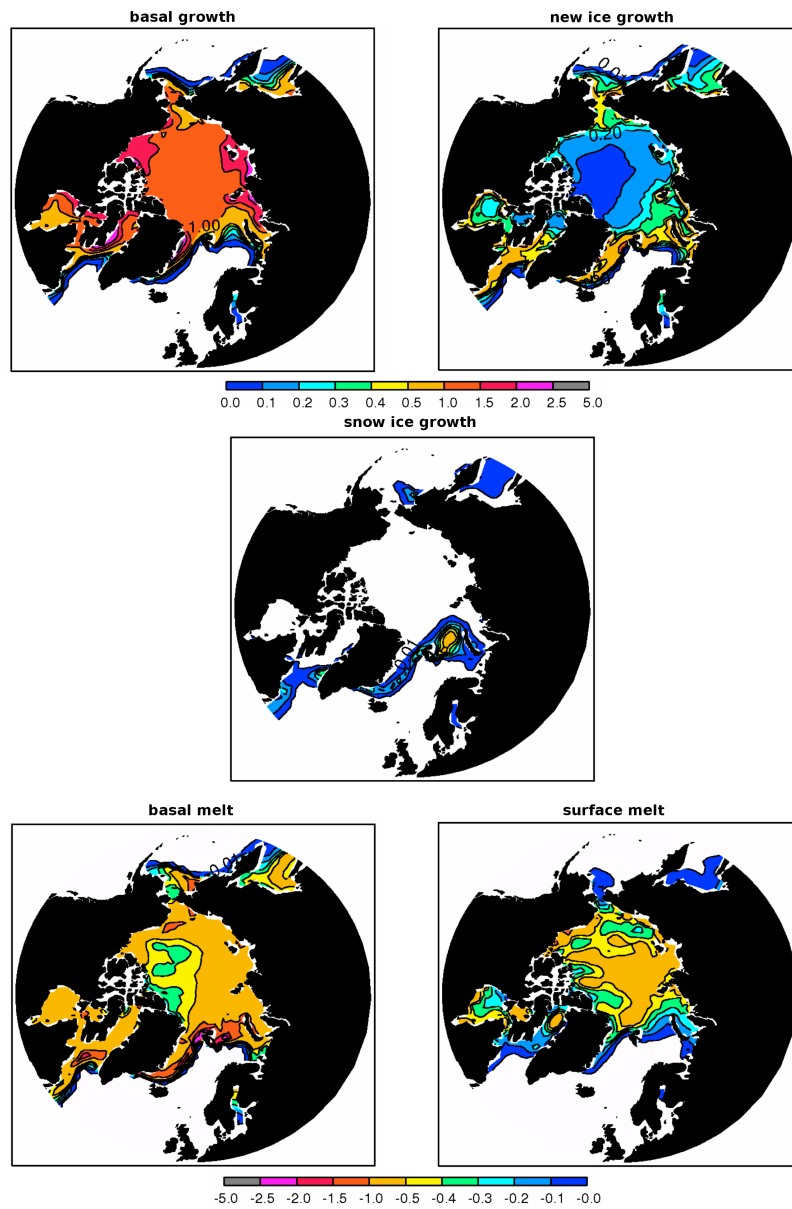


Figure 1.3: Annual mean tendencies of mean ice thickness (m/y) in the NH simulated by LIM3, over the period 1979-2006.

radiation in the ocean. This enhances the oceanic heat flux at the ice base and hence bottom melt, leading to a decrease in ice thickness.

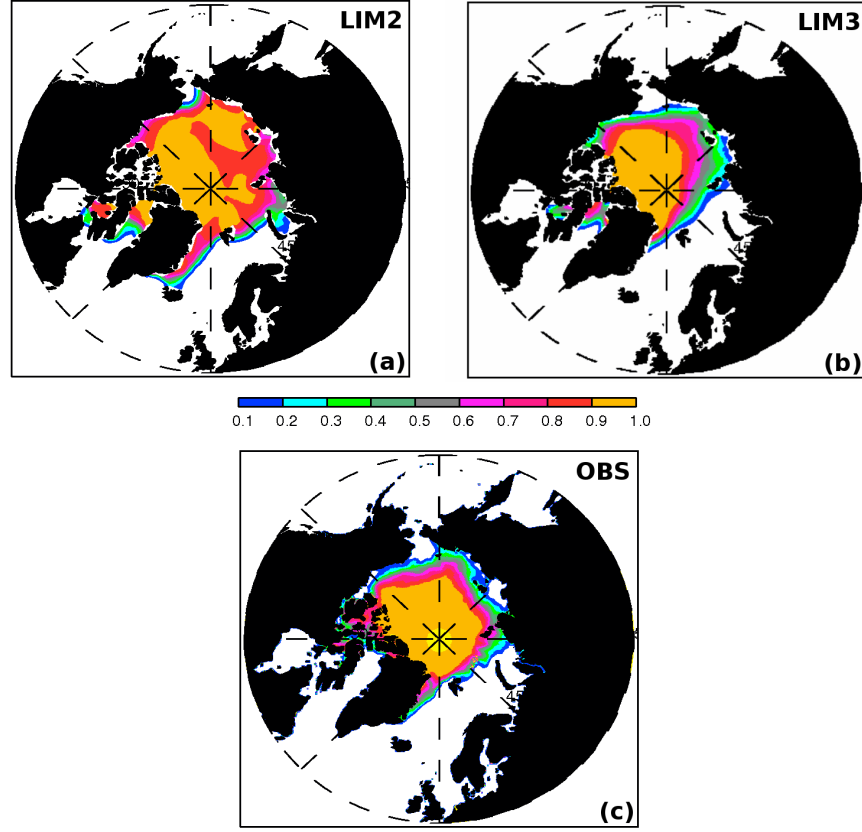


Figure 1.4: Simulated September geographical distribution of ice concentration in the NH simulated by **LIM2** (a), as derived from PMW observations (Comiso, 2007) (b) and simulated by **LIM3** (c). Reproduced from Vancoppenolle *et al.* (2008b).

Ice volume is well described in terms of ice thickness. Comparison to data from upward-looking sonars (ULS) onboard submarines (Wensnahan *et al.* (2007), see Figure 1.5) reveals a mean 1976-2000 **LIM3** (**LIM2**) – data difference of -0.55 ± 1.04 m (1.88 ± 0.97 m). The most significant changes in the simulation are due to the new dynamical component, to the introduction of prognostic salinity and of an explicit ITD. The latter results in a more realistic ice-albedo feedback which is governed by the fraction of thin ice, rather than by the summer ice albedo value. Because of the strong feedback of the stronger thermohaline circulation²³

²³The thermohaline circulation (THC) is the global density-driven circulation of the oceans. Density of seawater is function of salinity, temperature and pressure.

(THC) on the simulated sea ice volume, it is difficult to point a particular physical parameterization as responsible for the better simulation. Significant ice thickness biases remain in **LIM3**. Ice is too thick in the Beaufort Gyre and too thin in the Atlantic sector of the Arctic Ocean. Comparison with several other models showed that this pattern is typical of multicategory ice models with a VP/EVP rheology, has been suggested to be due to an underestimated shear strength and can notably be reduced by including sliding friction in the mechanical redistribution scheme (Wilchinsky *et al.*, 2006).

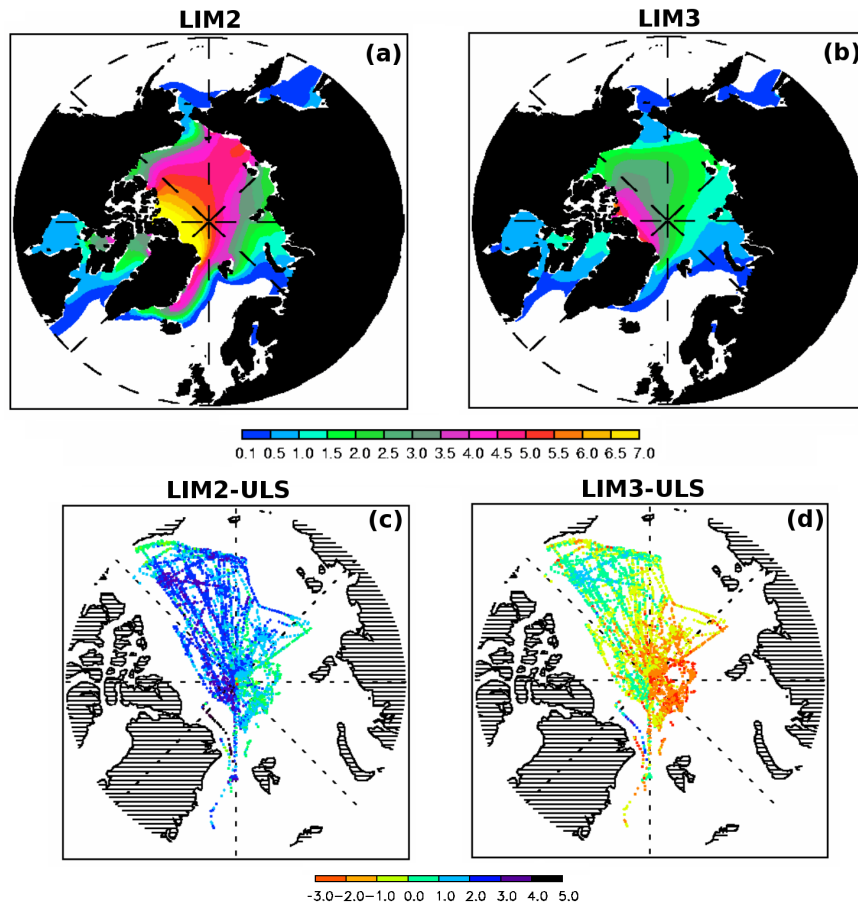


Figure 1.5: Geographical distribution of January Arctic ice thickness (m): (a) difference **LIM2** – ULS data, (b) **LIM2**, (c) **LIM3** and (d) difference **LIM3** – ULS data. ULS data come from Wensnahan *et al.* (2007). Reproduced from Vancoppenolle *et al.* (2008b).

Associated with the better mean seasonal cycle, the simulation of interannual variability is also improved. In the Arctic, the correlation between the simulated

and observed monthly mean anomalies of ice area is 0.52 in **LIM2** and 0.74 in **LIM3**. **LIM3**'s better representation of the ice-albedo feedback makes it more efficient to simulate the September minimum of extent, the interannual variations and the long-term trend. In addition, the spatial distribution of anomalies is also much better captured with **LIM3**, as illustrated by the simulation of the recent 2007 minimum of extent, which reached a record value of 4.28×10^6 km² (Comiso *et al.*, 2008). **LIM3** simulates a pattern close to observations and slightly underestimates the summer 2007 ice extent (3.66×10^6 km²). **LIM2** does not manage to melt enough ice and significantly overestimates the ice extent (5.85×10^6 km²).

The simulated large-scale ice circulation is in agreement with satellite observations of ice motion, as the areal export of ice through Fram Strait (Kwok *et al.*, 2004), including seasonal and interannual variations. Associated to underestimated ice thickness near the North Pole, the export of ice volume through Fram Strait is smaller than observed. Information on snow depth in the Arctic is too limited to draw conclusions regarding the model ability to simulate it.

Finally, the improvement in the simulated sea ice using **LIM3** is associated to large improvement of the simulated features of the ocean. In particular, the simulated THC is much more realistic using **LIM3** instead of **LIM2**, characterized by a strength of the meridional overturning stream function of 18 Sv²⁴, which is 10 Sv higher than in **LIM2**, averaged over 1990-1999. The THC is extremely sensitive to deep convection in the North Atlantic. In the North Atlantic, the role of sea ice inflow dominates the buoyancy forcing, which controls deep convection. Associated to the smaller ice thickness all over the Arctic Ocean (see Figure 1.6), the ice volume export and freshwater transport through Fram Strait are more realistic in **LIM3**. Consequently, the surface waters are 0 – 0.5 ‰ more saline. In turn, the distribution of the convection sites in the North Atlantic is much better simulated than in **LIM2**, with differences in mixed layer depth between the two ice models up to 2000 m in the Labrador sea.

Around Antarctica, basal congelation growth, snow ice formation and new ice growth in open water contribute significantly to ice growth. Ice melts mostly at the bottom interface (see Fig. 1.7 and **V08-1, sect. 6.4.3.3**). This is in agreement with textural analyses of ice cores, though the amount of new (frazil) ice formation in open water seems underestimated by the model (see, e.g., Jeffries *et al.* (1997); Worby *et al.* (1998)). Nevertheless, associated to the low bias in ice thickness on the east side of the Antarctic Peninsula, due to the wrong direction of winds in the NCEP/NCAR forcing (R. Timmermann, personal communication),

²⁴1 Sverdrup (Sv) is equal to 10^6 m³/s.

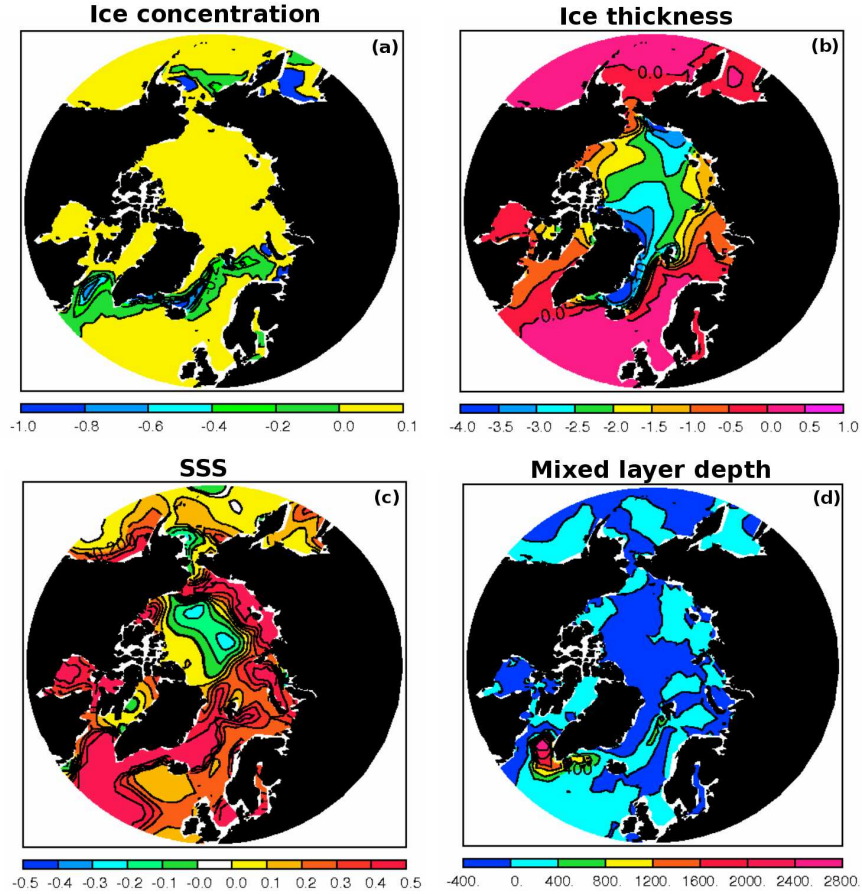


Figure 1.6: 1979-2006 **LIM3**–**LIM2** differences, in March, in terms of (a) ice concentration, (b) ice thickness (m), (c) sea surface salinity (‰) and (d) mixed layer depth (m).

in this location, there are spurious maxima of new ice growth, bottom melt and snow-ice formation.

The seasonal variations in ice extent are rather well represented compared to satellite observations (Comiso, 2007), though the winter ice area and concentrations are overestimated. In summer, the simulated ice retreats slightly too far southward. The simulated ice thickness resembles the ASPeCT climatology (Worby *et al.*, 2008), with a tendency towards underestimation, in particular in coastal seas. Compared to satellite observations (Markus and Cavalieri, 1998, 2006), snow depth is quite realistic in the model. In addition, the general features of the large-scale ice motion are well reproduced by **LIM3**, though the coastal currents

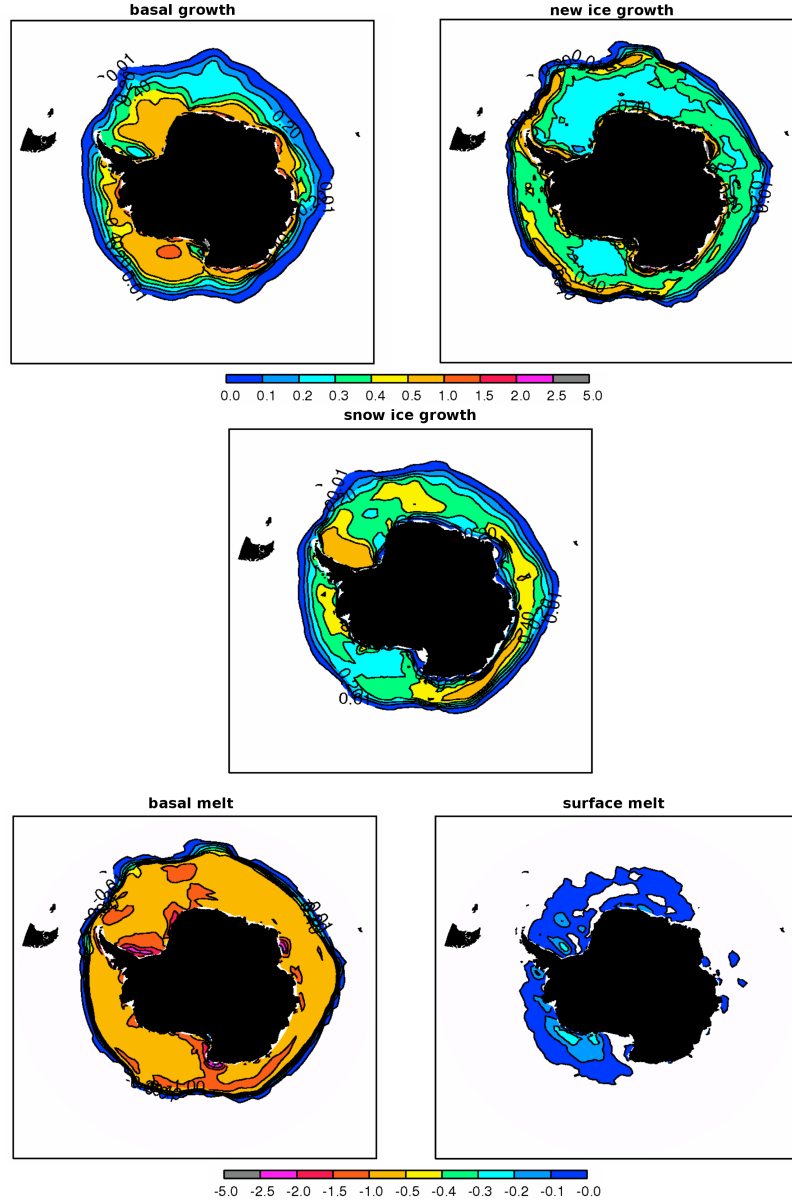


Figure 1.7: Annual mean tendencies of mean ice thickness (m/y) in the NH simulated by LIM3, over the period 1979-2006.

are significantly overestimated²⁵. Compared to **LIM2**, the simulation of Antarctic

²⁵Simulations with LIM2, using VP on a B-grid, similarly overestimate the coastal currents (see, e.g., Timmermann *et al.* (2005); Bouillon *et al.*, 2008). Therefore, it is argued here that this is due to the rheology.

sea ice is neither improved nor deteriorated. It is argued that there are significant missing processes for the simulation of Antarctic sea ice which are responsible for this behaviour.

1.6.6 Discussion

There are a series of remaining problems that should be addressed to improve 3D sea ice simulations. The first problem is the missing model physics, the second one is the forcing and the third one is the lack of validation data.

Model physics. There are several missing physical parameterizations in the model. First, the ice deformations simulated by 3D sea ice models using VP or EVP rheologies disagree with observed deformations derived from synthetic aperture radar (SAR) data (Lindsay *et al.*, 2003). The hypotheses underlying VP-like models, examined in the light of SAR data, prove to be inadequate (Coon *et al.*, 2007). Newly developed in-situ stress sensors (Richter-Menge *et al.*, 2002) and SAR deformations (Kwok and Cunningham, 2002) offer a direct and accurate validation background for sea ice rheological models, enabling new progresses in sea ice rheology (see, e.g., Schreyer *et al.* (2006); Sulsky *et al.* (2007)). Newly developed rheologies are of strong interest for sea ice mass balance through the impact of ice dynamics on transport and mechanical redistribution by rafting and ridging.

In addition, a better representation of snow, meltwater and flooding is required. Models of snow thermodynamics and metamorphism exist and could be included (see, e.g., Nicolaus (2006)). Nevertheless, there are two majors obstacles to the development of snow models : (1) in polar regions, not much is known either on precipitation or on wind effects on snow resuspension and deposition on sea ice rough surfaces; and (2) in the Arctic Ocean, there is no remote-sensing method to derive large-scale snow depth data. Meltwater parameterizations, including over- and under-ice melt ponds (Flocco and Feltham, 2007) and superimposed ice formation (Cheng *et al.*, 2006) have recently been developed and should be included in models²⁶. Better representation of flooding, slush, and gap layers²⁷

²⁶Melt ponds would require significant model development work to be included.

²⁷Flooding occurs when the snow load is high enough to depress the snow-ice interface under sea level. The subsequent penetration of seawater through the snow creates slush (see, e.g., Jeffries *et al.* (1998), which, if the snow-ice interface gains heat, remains unfrozen and may turn into a gap layer Haas *et al.* (2001)). If the snow-ice interface loses heat, slush refreezes into snow ice (Jeffries *et al.*, 1997).

should also be taken into consideration. Surface flooding is frequent in Antarctic sea ice (see., e.g., Perovich *et al.* (2004)).

Furthermore, though pancake ice is estimated to contribute to about one third of the ice-covered Southern Ocean in winter (Wadhams, 2000) , the dynamical growth of thin ice and the impacts of oceanic waves on sea ice is not well understood and therefore not included in the models.

The question of lateral melting at the edges of the floes is far from being resolved. Lateral melting is controlled by floe size (Steele, 1992; Perovich *et al.*, 2003) and only contributes significantly for floes that have a diameter smaller than about 100 m. The ice floe size distribution is poorly known, which renders estimates of the contribution of lateral melting difficult. SHEBA data has indicated a variable contribution of lateral melting to the total energy consumed by sea ice melt: nearly zero in July and around 30% of in early August (Perovich *et al.*, 2003). Nevertheless, conceptually, it is difficult to distinguish between the total removal of thin ice and lateral melting. Bitz *et al.* (2001) proposed a parameterization of lateral melting²⁸ in the marginal ice zones²⁹. In models including ITD, there is some lateral melting accounted for by the melt of thin ice. In summary, conceptual and observational evidence of lateral melting is needed.

Finally, due to a lack of ice thickness data, the mechanical redistribution schemes are not validated in isolation. Amundrud *et al.* (2004) have shown that mechanical redistribution algorithms in common use generate much more deep ridged ice than is observed. They concluded that the existence of ridges and floes should be an acknowledged factor in redistribution theory for pack ice thickness, which is so far not the case.

Forcing. As underlined by the 1D studies, the incoming radiative LW and turbulent oceanic forcing should be better studied and constrained. First, in current simulations with **LIM3**, clouds come from a climatology, though synoptic and interannual variations are significant. In addition, the interactions of the changing ice and clouds are not well known, but studies indicated that clouds have a tremendous impact on the radiative forcing in both hemispheres (see, e.g., Guest (1998); Key and Minnett (2006); Gorodetskaya and Tremblay (2008)). Furthermore, clouds modify the light properties which results in apparent changes in sea ice radiative properties (Grenfell and Maykut, 1977). Second, the ice-ocean heat and radiation exchanges should be better studied and validated. Finally, in some

²⁸In their parameterization, the lateral melt rate depends on a prescribed mean floe diameter and on the difference between the mixed layer temperature and the seawater freezing point.

²⁹The marginal ice zones are the sea ice regions close to the ice edge.

regions, even the reanalysed winds and temperatures are not good, for example winds along the Antarctic Peninsula (Timmermann *et al.*, 2005) and temperatures in some regions of the Arctic (Hunke and Holland, 2007).

Data. The most significant gap in the sea ice large-scale, continuous datasets is the ice thickness, though as detailed in the introduction, this should change in the near future.

1.7 Salinity variations in sea ice

Thanks to a representation of the mechanisms of brine entrapment and drainage, LIM1D is the first model able to simulate the seasonal cycle of the Arctic sea ice salinity profile, which provides a general understanding of the different salinity profiles observed in the field. In addition, the different categories of ice thickness are characterized by different salinities. By combining the ice thickness distribution theory to brine drainage, LIM3 generated the first assessment of the large-scale distributions of the sea ice salinity.

1.7.1 Ice desalination and brine drainage mechanisms.

Field and laboratory studies indicate that, due to brine entrapment and drainage, the sea ice salinity varies in time. The two main stages of sea ice desalination were already identified in the pioneering study of Malmgren (1927), see Fig. 1.8. The first one is an intense, initial desalination progressively decreasing in intensity throughout the winter. The second one is a rapid desalination most pronounced at the ice surface, occurring in summer.

Brine entrapment occurs during initial sea ice growth, through congelation at the ice base and snow ice formation at the ice surface. Brine drains due to brine expulsion, gravity drainage and flushing. Brine expulsion occurs as follows. When a brine pocket cools, the thermal contraction of the ice around the pocket is greater than that of the liquid inside, producing a pressure gradient that may lead the pocket wall to rupture (Cox and Weeks, 1986). Gravity drainage gathers all processes where brine is expelled from the ice under the influence of gravity through an interconnected brine network (Cox and Weeks, 1975, 1988). The mechanism behind gravity drainage is not known (downward advection of brine due to increase in freeboard during ice growth, diffusion or convection). Flushing

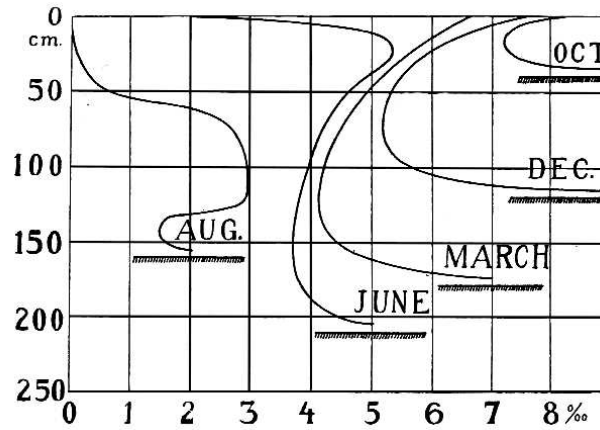


Figure 1.8: Schematic desalination pattern in Arctic sea ice, reproduced from Malmgren (1927).

is a special kind of gravity drainage, where the pressure head and water are provided by the fresh surface meltwater going through the permeable summer brine network (Untersteiner, 1968). Brine drainage mechanisms are reviewed extensively in **VBF07, sect. 4.3.2**.

1.7.2 Vertical variations in ice salinity simulated by LIM1D

In **VFB06**, the model simulates a continuous desalination pattern for Arctic sea ice, which clearly reproduces the transition from FY to MY sea ice outlined by Malmgren (1927). The FY ice desalination was divided into initial, winter and summer stages of desalination. The *initial desalination* is due to gravity drainage throughout all the layers. During *winter drainage*, the ice slab is split into two parts. The lower third is warm and saline enough to be permeable to gravity drainage. The upper layers are impervious. During *summer regime*, if the whole ice column is permeable, flushing occurs, which leads to the typical MY ice profile, which is characterized by low salinity near the surface. Then, the intensity of brine drainage decreases from year to year. Following this desalination pattern, the MY ice salinity profile is found to be sensitive to flushing near the ice surface and to gravity drainage near its base. The role of brine expulsion is negligible. The snow cover, plays an insulating role which and may prevent flushing if deep enough.

The validation of LIM1D at Point Barrow (**VBF07**) consisted in, first, a documentation of the summer desalination sequences. Based on this, the halodynamic module was developed and validated against data from Point Barrow, with a particular focus on summer processes. In winter, adjustment to the original Cox and Weeks (1988) gravity drainage parameterization was necessary to give reasonable model-observation agreement. Our results indicate that the Cox and Weeks (1988) parameterization – which is based on laboratory-controlled ice growth over about two weeks – is valid only for the earliest stages of desalination, a period for which there is some evidence that convection is active (see, e.g., Notz (2005)). After two weeks, the gravity drainage coefficient has to be reduced to maintain model-observation agreement. It is arguable that changes in convection (or even an interruption of it) weaken the desalination and explain the necessary reduction of the original Cox and Weeks (1988) gravity drainage coefficient. In summer, the salinity profile simulated by **LIM1D** closely resembles observations. Flushing induces a quick transition in the salinity of the upper 50 cm of ice: in 12 days, the salinity of the upper ice layers decreases from 8–10‰ to below 3‰.

Sensitivity experiments suggest that the summer desalination is triggered by solar radiation penetrating inside the ice, which enhances internal melting and increases the brine volume until the permeability threshold is reached, allowing some of the meltwater produced at the surface to percolate through the ice. In addition, the snow depth modulates the insulating role of snow and controls the timing of the desalination onset.

The simulated summer desalination confirms the observed summer pathways of meltwater through the ice. In early summer, about 25% of the meltwater production is stored into the ice. The remainder of the meltwater production refreezes on top of the ice (superimposed ice), collects into melt ponds or is expelled in the ocean through cracks and leads (Eicken *et al.*, 2002).

1.7.3 Large-scale variations in ice salinity simulated by LIM3

LIM3 offers the first opportunity to investigate the large-scale, seasonal and interannual variations of sea ice salinity (see **V08-2, sect. 7.5**). It was found that, due to brine drainage, salinity variations reflect the sea ice age variations: the older, the less saline the ice.

According to our simulations and observations (see Gow and Tucker (1991), for a review), the sea ice mean salinity is higher in the SH (6.3 ‰) than in the NH (3.6 ‰). This is due to, first, the larger contribution of FY ice in the SH than in the NH. Second, in the SH, surface melt and associated flushing are not as large as in the NH. Finally, in the SH, contrary to the NH, snow ice formation significantly contributes to the global ice mass balance. Geographical variations in the NH reflect the regional variations in FY-MY ice coverage, with lower values in the central Arctic and larger values towards the ice edge.

The simulated ice salinity also varies seasonally, with a maximum in fall-winter, when the abundance of more saline, FY ice is maximum, and a minimum in summer. In addition, the model simulates interannual variations, particularly in the NH, which reflect the decreasing trend in ice area. From 1979 on, the NH MY ice coverage decreased, the relative importance of FY ice increased accordingly, and, in turn, the average ice salinity increased. The seasonal and geographical ice salinity variations compare well to observations³⁰.

The salinity variations in thickness space were found to compare well to the Cox and Weeks (1974) relations. Thinner ice is younger and therefore more saline than thicker ice. In summer, in the NH, the salinity contrast between thin and thick ice disappears because of flushing which percolates through ice of any thickness indifferently.

1.7.4 Discussion

Simulation of ice salinity is only beginning and numerous issues remain. First, the fluid transport mechanisms behind gravity drainage are not well understood, which is a problem to assess the transport of nutrients through the brine network. Second, flooding, gap layer and snow ice formation are neither well understood.

Recent development of nondestructive methods to measure the ice salinity in situ, based on impedance and capacitance measurements (Notz *et al.*, 2005; Backstrom and Eicken, 2006) should bring significant new knowledge in the near future. In addition, based on new visualization techniques of the sea ice microstructure, Golden *et al.* (2007) developed a unified theory of the fluid permeability of sea ice and its dependence on microstructural evolution with temperature. These recent

³⁰In order to provide an observational reference to the simulation of salinity in **LIM3**, a collection of ice bulk salinity data from more than one thousand ice cores was collected from various sources (see **V08-2**, sect. 7.5.1.1).

progresses on the sea ice permeability and on new observation techniques should, in the long run, lead to significant improvements in sea ice models.

1.8 Impact of salinity variations on sea ice

Salinity variations may affect the large-scale sea ice mass balance through : (1) a direct impact on heat transfer and storage through the ice; (2) a direct impact on the exchanges of salt between ice and ocean; (3) an indirect impact through oceanic feedbacks. LIM1D and LIM3 enabled to carefully examine the individual role of these three important mechanisms.

1. Salinity affects the sea ice thermodynamics as follows. In the formulation of sea ice thermodynamics of Bitz and Lipscomb (1999), the vertical variations of the temperature in the ice (T) are computed by numerically solving the heat diffusion equation in 1 layer of snow and 20 layers of sea ice, characterized by their salinity S . In this equation, the sea ice specific heat c and the thermal conductivity k control heat transfer through and storage in sea ice. The boundary conditions state that growth or melt results from the imbalance between the internal (conductive) and external heat fluxes. Growth and melt rates are computed by dividing the heat loss (gain) at the interfaces F by q , the energy of melting, defined as the energy required to melt a unit volume of sea ice. Sea ice salinity affects phase equilibrium in sea ice and, in turn, all its physical properties. In particular, the sea ice thermal properties (c , k , and q) depend on S and T .

2. Sea ice and ocean exchange salt and freshwater. These exchanges are extremely important by shaping the density structure of the upper ocean, which in turn regulates the global THC (Aagaard and Carmack, 1989). Following Tartiville *et al.* (2001), if a surface process affects the ocean salinity without a net gain or loss of water for the ice-ocean system (e.g., through ice melting or freezing), this internal exchange is transformed into an equivalent salt flux that is applied to the ocean. This equivalent ice-ocean salt flux F^s , positive for a salt flux and negative for a freshwater flux, is given by:

$$F^s = \rho_i [v^i \frac{\partial S}{\partial t} + (S^w - S^{g/m}) \frac{\partial v^i}{\partial t}], \quad (1.3)$$

where the first term on the right-hand side refers to salinity variations and the second one refers to formation / melt of saline ice. ρ_i is the sea ice density, v^i is the sea ice volume per unit oceanic area, S^w the ocean salinity and $S^{g/m}$ is the salinity of growing or melting ice. In case of ice growth, it refers to the salinity

of the newly grown ice and varies significantly depending on the growth process. In case of ice melt, it is simply the salinity of the melting ice layer. The salt flux due to brine drainage is always positive. The salt flux due to ice growth / melt is positive if ice grows and negative if ice melt. Through their effect on the ocean, changes in the ice-ocean salt fluxes may induce oceanic feedbacks.

3. There are two main feedbacks between ice and ocean that are of importance here. The first one is the ice growth and vertical mixing feedback. It couples the ice growth, the ice-ocean salt flux, vertical mixing in the upper ocean and the ocean-ice heat flux (Martinson, 1990). In the SH notably, a thick, well mixed surface layer is maintained by surface mechanical stirring. At its base rests a sharp pycnocline (10-40 m thick) characterized by a thermal contrast of 2-3°C. The strong diffusive heat flux (around 20-25 W/m²) across this interface satisfies a large fraction of the approximately 35 W/m² of heat loss to the atmosphere. The remaining heat loss must be provided by latent heat of fusion released through ice growth. Let us assume that the system is in steady state. Any further cooling will drive new ice growth, which in turn implies an ice-ocean salt flux and a salinization of the mixed layer. This, therefore, becomes prone to static instability which will, in turn, cause the erosion of the pycnocline, a warming of the mixed layer, an increase in the sensible heat flux and a decrease in ice growth. The second feedback mechanism is the ice-albedo feedback, which is positive and tends to amplify summer melt. In summer, the complete removal of thin ice (through melt) replaces a reflective ice surface by a much darker oceanic surface. Consequently, more solar radiation is trapped inside the ocean, which warms the upper ocean and, in turn, further increases the sea ice basal melt.

1.8.1 Local impact of salinity studied with LIM1D

A series of experiments aimed at isolating the individual effect of each of the thermal properties – summarized in **VFB05, sect. 3.3.1** – has shown the following. (1) The role of salinity variations on the ice thickness through its impact on c and q is much more influential than their effect on k . (2) At high temperatures (summer), the higher c of saline ice requires more energy to be apportioned to internal warming and melting rather than to direct melt at the interface. In turn, this reduces the melt rate. (3) The reduction of q for saline ice enhances melt and growth rates. c and q have comparable and competing effects that dominate or not depending on several factors.

In **VBF05**, the bulk salinity and the shape of the vertical sea ice salinity profile were varied while the temporal changes were neglected, and the impact on sea

ice thickness and salt/freshwater exchanges was studied. First, a higher bulk salinity increases c and reduces surface melt, which leads to thicker ice. Using a vertically-varying salinity profile, typical of MY ice, with reduced ice salinity near the surface, increases surface melt and leads to thinner ice, compared to a vertically constant profile with comparable bulk salinity. Therefore, capturing the FY-MY ice salinity transition is important in sea ice mass balance simulations. Regarding the salt/freshwater exchanges between ice and ocean, our results indicate that changing the ice salinity only slightly affect the salt/fresh water ice-ocean fluxes. This study stressed the effect of salinity variations on **surface melt**.

In **VFB06**, the effect of the temporal evolution of sea ice salinity was studied. The halodynamic module of **LIM1D** was compared to several prescribed, steady-state salinity profiles in terms of ice thickness and ice-ocean salt/freshwater fluxes. Including an interactive salinity allows to grow new ice at relatively high salinity (6-12‰) compared to when salinity is prescribed, which reduces q and increases ice growth. The absolute magnitude of bottom growth and surface melt rates are larger when salinity variations are taken into account compared to a vertically-constant salinity profile. During the first 5 years of the run, the difference in growth dominates. Then, the difference in melt becomes larger. The ice is thicker using the interactive salinity module until year 25. Finally, our results indicate that the differences in ice-ocean salt/freshwater fluxes are governed by differences in growth/melt rates rather than differences in ice salinity. This study underlines the effect of sea ice salinity variations on **basal growth** of ice.

1.8.2 Large-scale impact of salinity variations studied with LIM3

In **V08-2, sect. 7.6**, the same type of sensitivity experiments are repeated with **LIM3**. A simulation with the interactive salinity distribution module is compared to simulations using prescribed, steady-state salinity profiles. In addition, an experiment designed to study the role of oceanic feedbacks is run. In the latter, only the ice thermodynamic component "sees" the salinity variations, while the ice-ocean salt/freshwater flux is computed as if the salinity was constant.

In the NH, the 3D results confirm the 1D findings. Including an interactive salinity distribution enhances both ice growth and melt rates through a direct impact of brine inclusions on ice thermodynamics. The difference in ice growth dominates. Thus, ice is thicker if an interactive salinity distribution is used. A difference from the 1D findings is that, due to the contribution of the larger oceanic heat flux in the seasonal ice zones, melt at the ice base has the largest contribution

in summer. The varying salinity facilitates thin ice melt and increases oceanic absorption of solar radiation and the ocean-ice heat flux, which in turn results in larger basal melt. The reduction in ice-ocean salt flux associated to winter ice growth freshens the upper ocean by up to 1 ‰ in zones of high sea ice production. This reduces vertical mixing and deep convection, but not enough to significantly feedback on sea ice.

Around Antarctica, the effect of an interactive salinity distribution is even larger. But, in contrast to the NH, because of the weaker oceanic stratification, ice-ocean interactions are important. Using a variable salinity allows to maintain significant ice growth rates with relatively small ice-ocean salt fluxes. This significantly reduces vertical mixing and the mixed layer depth. Through the ice growth - vertical mixing feedback, this in turn further enhances ice growth and results in thicker ice than if any of the prescribed salinity profiles is used.

In summary, salinity variations significantly affect both sea ice, ocean and their interactions and should be considered in future sea ice components in climate models. In the next chapters, the interested reader will find the technical elements and the complete scientific arguments that provide the scientific basis for this assertion.

Obigies, le 31 décembre 2007.

Chapter 2

Sea ice halo-thermodynamics in LIM1D

Sea ice traps some salt as it forms. A fraction of oceanic salt is trapped in the ice during formation, while the remainder is rejected into the ocean. As the crystalline network of H_2O cannot incorporate salt, sea ice can be seen as a porous matrix of pure ice filled with liquid inclusions of brine. The thermal properties of sea ice reflect the phase composition of sea ice, which depend on the ice temperature and salinity. In turn, brine entrapment and drainage induce variations in ice salinity, which affect heat transfer and storage as well as sea ice growth and melt. In this short chapter, we describe the theoretical basis of brine physics in LIM1D.

2.1 Heat diffusion in the snow-sea ice system

Here, we assume that sea ice as an infinitely wide plate covered by a snow layer and that heat transfer is purely vertical. The purpose of sea ice thermodynamic modelling is to compute changes in snow depth (h_s), ice thickness (h_i), surface temperature (T_{su}) and inner vertical temperature profiles inside snow (T_s) and ice (T_i), given the changes in the external atmospheric and oceanic forcings. In numerous aspects, the thermodynamics described in this section follows from Bitz and Lipscomb (1999).

Vertical heat conduction and storage in the sea ice-snow system are governed by the heat diffusion equation:

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(k \frac{\partial T}{\partial z} \right) + \kappa I_0 e^{-\kappa z}, \quad (2.1)$$

where ρ , c and k are the sea ice density, specific heat and thermal conductivity, respectively. I_0 is the penetrating solar radiation and $\kappa = 1.0 \text{ m}^{-1}$ is the extinction coefficient from Beer's law.

The boundary conditions applied to (2.1) at the interfaces with atmosphere and ocean express that there is a balance between the internal conduction and external heat fluxes. Any imbalance in the heat fluxes is converted in ice growth or melt. At the air-ice interface, the energy balance is given by:

$$F_{net}(T_{su}) = F_{sw}(1 - \alpha)(1 - i_0) + F_{lw}^\downarrow - \epsilon \sigma T_{su}^4 + F_{sh} + F_{lh} + k \frac{\partial T}{\partial z}, \quad (2.2)$$

where F_{sw} is the shortwave downwelling radiation arriving at the ice surface, α the surface albedo, i_0 the fraction of the penetrating solar radiation which penetrates inside the ice layers (i.e., which does not contribute to the surface energy balance), $F_{lw}^\downarrow$ is the downwelling longwave radiation flux, ϵ is the emissivity of the surface, σ is the Stefan-Boltzmann constant, F_{sh} and F_{lh} are the turbulent sensible and latent heat fluxes, respectively, and $k \partial T / \partial z$ is the conductive flux from the sea ice/snow interior towards the top surface. If $F_{net} \geq 0$, then the temperature of the ice surface is fixed at the melting point (equal to 0°C) and ablation occurs according to:

$$F_{net}(T) = -q(S, T) \frac{dh_x}{dt}, \quad (2.3)$$

where q is the sea ice energy of melting (defined as the energy required to melt a unit volume of sea ice) and the subscript x refers either to snow or ice. The temperature at the ice base equals the freezing point of seawater. At the ice base, the energy balance is given by:

$$F_w - k \frac{\partial T}{\partial z} = -q(S, T) \frac{dh_i}{dt}, \quad (2.4)$$

where F_w is the heat flux from the ocean and $k \partial T / \partial z$ is the conductive flux from the ice base towards the interior.

2.2 Phase equilibrium in sea ice

The sea ice phase composition depends on state variables, the temperature T and the salinity S . This follows from two assumptions:

- The freezing temperature of a saline solution is a linear function of its salinity S : $T_f = -\mu S$, $\mu = 0.054^\circ\text{C} \cdot \%^{-1}$ (Assur, 1958).
- Brine pockets are at their freezing point and in local thermal equilibrium with the surrounding ice and permanently at their freezing temperature.

Therefore, brine pockets adapt their salinity σ by changing size to the local ice temperature T (in $^\circ\text{C}$):

$$T = -\mu\sigma. \quad (2.5)$$

In this context, σ is the salinity of the brine pockets, while S is defined as the salinity of the ice if brines were distributed uniformly throughout the ice. Thus, σ and S are related to each other by :

$$\sigma e = S. \quad (2.6)$$

The relative volume of brine inclusions e (dimensionless, hereafter referred to as *brine volume*) is consequently only a function of S and T (Schwerdtfeger, 1963):

$$e = -\mu \frac{S}{T}, \quad (2.7)$$

which is plotted in panel (a) of Figure 2.1. Close to the sea ice melting point, the brine volume increases drastically. Above brine volumes of 5%, the sea ice brine network connects and sea ice becomes permeable to fluid transport (Golden *et al.*, 1998).

2.3 Sea ice thermal properties

The sea ice thermal properties (specific heat, thermal conductivity and energy of melting) depend on the sea ice phase composition, hence salinity and temperature. The sea ice specific heat c is defined as the amount of heat ∂Q required to warm a unit mass of sea ice by a small temperature increment dT , divided by dT , at constant salinity. ∂Q can be expressed as:

$$\partial Q \equiv cdT = c_0dT(1 - e) + c_wdT_e + L_id e, \quad (2.8)$$

where $c_0 = 2067 \text{ J} \cdot \text{kg}^{-1} \cdot ^\circ\text{C}^{-1}$ and c_w are the specific heat of pure ice and seawater, respectively. The terms on the right-hand side refer to the heat required to warm pure ice, the liquid fraction and to the latent heat necessary to melt brine pockets internally. L_i is the latent heat of pure ice, which depends on temperature:

$$L_i = L_0 + (c_w - c_0)T, \quad (2.9)$$

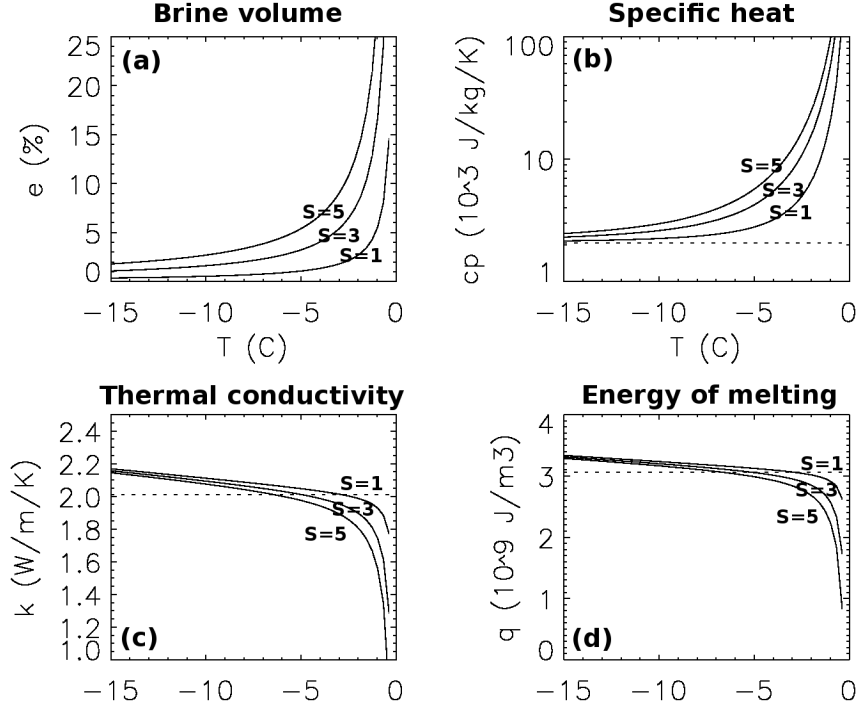


Figure 2.1: (a) Brine volume, (b) specific heat, (c) thermal conductivity and (d) energy of melting of sea ice as functions of ice salinity and temperature. Dotted lines represent the values corresponding to pure ice.

where $L_0 = 334 \times 10^3 \text{ J.kg}^{-1}$ is the latent heat of pure ice at 0°C . Differentiating (2.7) with respect to temperature, we get:

$$de = \frac{\mu S}{T^2} dT. \quad (2.10)$$

Using (2.10) and (2.9) in (2.8) and simplifying the terms, one finally obtains an expression for the sea ice specific heat:

$$c(S, T) = c_0 + \frac{L_0 \mu S}{T^2}, \quad (2.11)$$

which corroborates earlier field (Malmgren, 1927) and theoretical studies (Ono, 1967). As can be seen in panel (b) of Figure 2.1, at relatively low temperatures, the sea ice specific heat is close to the specific heat of pure ice. At relatively high temperatures, due to internal melt, the sea ice specific heat increases by more than one order of magnitude.

Integrating (2.11) between T and the seawater freezing point, we obtain the energy required to melt a unit volume of sea ice, also called the sea ice specific energy of melting q (always positive):

$$q(S, T) = \rho_i c_0 (-\mu S - T) + \rho_i L_0 (1 + \frac{\mu S}{T}). \quad (2.12)$$

ρ_i is the sea ice density. q is an equivalent of the latent heat of fusion for multiphase materials and also serves as a measure of internal energy for sea ice. In panel (d) of Figure 2.1, one can see that, at relatively low temperatures, q is very close to the pure ice latent heat. At higher temperatures, due to the liquid phase, q is on average 10 % lower than the standard latent heat of fusion of pure ice.

The sea ice thermal conductivity also depends on the sea ice phase composition. Following Pringle *et al.* (2007) :

$$k_i(S, T) = k_0 + \beta_1 \frac{S}{T} - \beta_2 T. \quad (2.13)$$

$k_0 = 2.01 \text{ W.m}^{-1}\text{.K}^{-1}$ is the thermal conductivity of pure ice and $\beta_1 = 0.09 \text{ W.m}^{-1}\text{.‰}^{-1}$ and $\beta_2 = 0.011 \text{ W.m}^{-1}\text{.°C}^{-2}$ are empirical constants. Since the S - T -dependence of the thermal conductivity of brine is unknown, a derivation of this equation cannot be readily obtained. Despite this, the second term on the right-hand side can be understood as the contribution of brine volume. At relatively high temperatures, the thermal conductivity of sea ice decreases by up to 20% compared to pure ice (see Figure 2.1, panel (c)). In principle, the sea ice density also depends on salinity and temperature. Nevertheless we neglected density variations in **LIM1D**.

2.4 Thermodynamic transformations of saline ice

As suggested by equations (2.7) and (2.12), any change in ice salinity will be followed by a change in brine volume :

$$de = -\frac{\mu}{T} dS + \frac{\mu S}{T^2} dT \quad (2.14)$$

and in internal energy :

$$dq = -\rho_i (c_s dS + c dT), \quad (2.15)$$

where $c \equiv \partial q / \partial T$ is the thermal specific heat given by (2.11) and $c_s \equiv \partial q / \partial S = \mu(c_0 - L_0/T)$ is the saline specific heat.

Four types of idealized thermodynamic transformations can occur and help to understand the thermal impact of brine drainage: isosaline ($dS = 0$), isothermal ($dT = 0$), energy-conserving ($dq = 0$) and isophasic ($de = 0$).

In isosaline transformations, a positive increase in ice temperature will be followed by an increase in brine volume and a decrease in internal energy. In isothermal transformations, a loss of salt will result in a decrease in brine volume and in an increase of internal energy. In energy-conserving transformations, a loss of salt results in an increase in ice temperature and in a reduction in brine volume. During isophasic transformations, a loss of salt results in an increase in ice temperature and in a loss of thermal energy. In the remainder of the text, we assume that sea ice desalination is an energy-conserving transformation.

2.5 Sea ice desalination

The theory describing the evolution of multiphase materials is frequently referred to as "mushy-layer" theory. Mushy-layer is a metallurgic term standing for two-phase, two-component reactive porous medium. The mushy-layer equations include the conservation of heat, brine salinity and momentum (Notz, 2005). A coupling term between fluid and heat transport is included. The bulk sea ice thermal properties are weighted means of the respective contributions from the solid and liquid phases. Written in terms of brine salinity σ , with $S = e\sigma$, the vertical salt conservation equation reads:

$$\frac{\partial}{\partial t}(e\sigma) = -\frac{\partial}{\partial z}(\sigma v_z) + \frac{\partial}{\partial z}\left(D\frac{\partial \sigma}{\partial z}\right), \quad (2.16)$$

where v_z is the vertical brine velocity in the brine network and D the molecular or turbulent diffusivity of brine. In other words, the evolution of the sea ice salinity profile is governed by transport (i.e., flushing) and diffusive or convective mixing processes (usually referred to as gravity drainage). Presently, the mushy-layer equations suffer from the lack of a formulation for gravity drainage.

The **LIM1D** halodynamic module (see **VBF07, sect. 3.2**) combines empirical formulae and mushy-layer physics. Based on an analysis of both literature and Point Barrow sea ice salinity data, the temporal and vertical changes in the salinity of Arctic sea ice were formulated. In **LIM1D**, the salinity profile $S(z, t)$ evolves following:

$$\frac{\partial S}{\partial t} = T_e K_e \frac{\partial^2}{\partial z^2} \left(E \left[\frac{dT}{dt} \right] \right) + G \left(e, \frac{\partial T}{\partial z} \right) - \frac{Q}{\rho_b} \frac{\partial \sigma}{\partial z}, \quad (2.17)$$

where ρ_b is the brine density. E gives the expulsion rate as a function of the temporal derivative of ice temperature and G refers to gravity drainage and is formulated as a function of the vertical temperature gradient, using the empirical formulae of Cox and Weeks (1988). K_b and T_b are associated time scales and diffusivity for brine expulsion, which product is taken to unity in our model. Q refers to the freshwater flux which percolates through the brine network in summer.

The brine network is assumed interconnected if $e > 5\%$. The bottom boundary condition applied to (2.17) specifies the new ice salinity as a fraction of seawater salinity (Cox and Weeks, 1988) and gives more brine entrapment for larger freezing rates. The surface boundary condition accounts for snow ice formation¹. For more details on the halodynamic formulation, see **VPF07, sect. 4.3.2**.

Resolved numerically and forced by atmospheric and oceanic forcings, the latter equations give the temporal evolution of snow depth, ice thickness and the vertical profiles of ice temperature and salinity that can be compared to observed values.

¹Since snow ice formation rarely occurs in the Arctic, this was not activated in the Point Barrow study.

Chapter 3

On the sensitivity of undeformed Arctic sea ice to its vertical salinity profile (VFB05)

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Abstract.

The temporal evolution of sea ice salinity affects the temperature profile and vertical growth and decay of the ice cover, as well as many other important properties. Here, we use a one-dimensional thermodynamic sea ice model to explore the sensitivity to the vertical profile of ice salinity of (1) Arctic first-year and equilibrium multiyear sea ice thickness, and (2) the salt/freshwater flux at the ice/ocean interface. Results indicate that increasing the mean salinity induces a higher thermal inertia reducing summer melt and finally increasing ice thickness. The shape of the profile is also important, since low salinity at the surface must be captured to produce enough surface melt. This study gives accurate hints on what the minimum complexity of a parameterization of the temporal evolution of sea ice salinity should be.

3.1 Introduction

Improving the treatment of thermodynamics in sea ice models is motivated by the importance of sea ice to high latitude climates, which themselves have a prominent role in the global climate. Currently, most climate models include a sea ice thermodynamic component, such as the one of Semtner (1976) or the more recent one of Bitz and Lipscomb (1999) (hereafter referred to as BL99), simulating the sea ice temperature profile and thickness as main prognostic variables. Some models have considered the ice salinity as a constant, only used while computing the salt/freshwater flux at the ice bottom interface (Fichefet and Maqueda, 1997). The impact of salinity on the ice properties has been accounted for in models either with a vertically varying salinity profile, fixed in time (BL99), or with an isosaline profile varying in time (Ebert and Curry, 1993). To improve this has frequently been reported as a next important step in future model development (Eicken, 2003).

Sea ice haline and thermodynamic processes are interrelated, and thus are characterized by unevaluated potential feedbacks. The first aspect of their twofold relationship is that sea ice thermodynamics control the relative rate of several brine drainage mechanisms responsible for the temporal evolution of the vertical salinity profile (Untersteiner, 1968; Cox and Weeks, 1988). The brine drainage leads, in the Arctic, to different salinity profiles for first-year (FY) and multiyear (MY) ice. The FY salinity profile is C-shaped with mean salinities often around 5‰ (Nakawo and Sinha, 1981). MY ice salinity is practically zero close to the surface and increases with depth, with a mean salinity around 2‰ (Schwarzacher, 1959).

Determining the level of complexity of a sea ice model with prognostic salinity requires a preliminary investigation of the second aspect of sea ice halothermodynamics (i.e., the interactions between haline and thermal processes in sea ice): the influence of sea ice salinity on the ice thermodynamics. Together with temperature, salinity controls the volume of internally trapped brine (Frankenstein and Garner, 1967). Consequently, sea ice thermal properties (which regulate ice growth and melt as well as the internal vertical temperature profile) depend on salinity and temperature, particularly close to the melting point where the greatest changes in brine volume occur. Thus, an accurate knowledge of the vertical salinity profile is important for capturing the main features of Arctic MY ice, as mentioned by BL99. But, to our knowledge, no detailed study of the impact of sea ice salinity on predicting sea ice growth and melt exists.

In this paper, we try to quantify the influence of the sea ice vertical salinity profile on the ice thermodynamics and on salt/freshwater exchanges in the Arctic. To do this, we use the BL99 energy-conserving one-dimensional (1D) thermodynamic sea ice model. In a forcing configuration typical of Arctic conditions, for undeformed congelation FY and MY ice, we run it with different salinity profiles, and analyse how this affects ice thickness, growth/melt rates, and, in turn, freshwater/salt flux to the ocean. We conclude by recommending directions towards the development of a halo-thermodynamic sea ice model.

3.2 Model description

The sea ice model used here is the BL99 1D thermodynamic energy-conserving model. The internal temperature profile is computed by numerically solving the heat diffusion equation in one layer of snow and 10 layers of sea ice, characterized by their own prescribed salinity. The thermodynamic effect of salinity, through its control of brine pockets, is represented by thermal properties depending on salinity (S) and temperature (T). The specific heat (c_p), which depends upon the internal storage of heat, increases by more than one order of magnitude close to the melting point. The thermal conductivity (k), which regulates the heat transfer, is lowered by up to 20%. The energy of melting, which governs the ice growth/melt rate (q , defined as the energy required to melt a unit volume of sea ice) is on average 10% lower than the standard latent heat of fusion.

The model is forced by classical idealized climatological atmospheric and oceanic conditions widely used in the literature. The upper surface forcing data (radiative and turbulent heat fluxes) are typical of the central Arctic (Fletcher, 1965). The oceanic heat flux is set to 2Wm^{-2} and the seasonal snow accumulation is as in Maykut and Untersteiner (1971). The model was initiated with zero-thickness sea ice and run for 50 years with a one-day time step. We consider that the first and last years of the simulation represent FY and MY ice, respectively.

In order to explore the sensitivity of the model to the ice salinity, different idealized salinity profiles are prescribed (see Figure 3.1). We use a non-linear profile, "SCHW", with salinity varying from 0‰ at the top of the ice to 3.2‰ at the bottom, and a mean salinity of 2.3‰. It is derived from the observations of Schwarzscher (1959). Being the most realistic profile, it is considered as the reference for the MY ice simulations. Furthermore, we use 3 vertically constant profiles, "I-0", "I-2.3" and "I-4.6". They correspond to salinity values of 0.1‰, 2.3‰ and 4.6‰, respectively. I-4.6 is chosen as a reference for FY ice since it

corresponds to the FY ice winter equilibrium value mentioned by Kovacs (1996). Finally, we also use 3 profiles situated between the vertically constant I-2.3 and the vertically varying SCHW profiles. Inside the ice, "L-PR" and "7-PR" are similar to I-2.3, and at the lower and upper interfaces respectively, they have the same salinity as SCHW. "SMS" profile is a linear profile with 0‰ at the surface and the same mean salinity as SCHW.

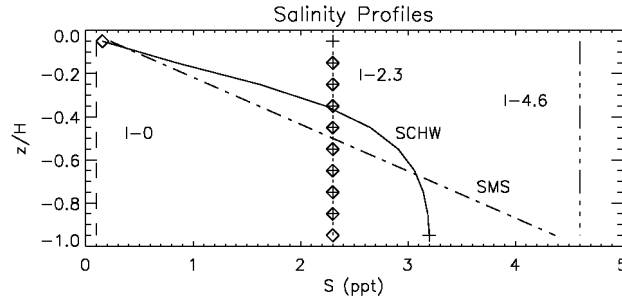


Figure 3.1: The different salinity profiles used in the simulations. 7-PR and L-PR are represented by squares and crosses, respectively. The remaining profiles are referenced on the plot : vertically constant profiles I-0, I-2.3 and I-4.6 (vertical dashed lines); vertically varying profile of Schwarzscher (1959) (SCHW, solid line); and a linear profile with the same mean salinity as SCHW (SMS, dashed-dotted line).

3.3 Model Results

3.3.1 Impact of bulk salinity

First, we only deal with vertically constant salinity profiles (i.e., isosaline ice). Figure 3.2a indicates that the model simulates a thicker ice for greater salinities. The MY thickness range among the different simulations is maximum in summer, reaching 144 cm. The values shown in Table 3.1 also indicate that, with increasing salinity, the surface melt period is shorter, as is the amplitude of the thickness seasonal cycle, and the bottom melt onset is delayed by up to two months.

Summer melt proves to be responsible for these differences. All along the 50 years of the simulation, total annual melt difference between I-0 and I-2.3 exceeds annual growth differences (see Figure 3.2b), with the major contribution from surface melt. Fresher ice melts more than higher salinity ice. Ice thickness, more than salinity, controls bottom growth, and since fresher ice is permanently

thinner, it also grows more; but the difference in melt dominates the behavior. As a consequence, the equilibrium thickness is smaller for fresher ice, while the amplitude of the seasonal cycle is larger.

Why does fresher ice melt more than more saline ice? Higher melt rates and longer melt periods share responsibility for this. They reflect the fundamental impact of salt on heat conduction and storage. Additional experiments (not shown here) aimed at isolating the individual effect of each of the thermal properties indicate the following. 1) Increasing salinity in the thermal conductivity $k(S,T)$ in isolation (i.e., while holding the salinity fixed in other salinity-dependent parameters) only modestly slows bottom growth through its impact on the heat conduction flux. 2) The impact of increasing salinity in the specific heat $c(S,T)$ in isolation is to delay melt onset at the surface by increasingly resisting warming (i.e. overestimating thermal inertia) as the ice approaches melting, thus diminishing the length of the melt period. With higher salinity, the ice preferentially apportions more energy to warming and melting internally rather than melting at the interfaces. Related to this, we noted greater downward heat conduction flux in summer at the ice surface, driven by stronger temperature gradients. Thus the increase in $c(S,T)$ with S has the effect of diminishing the rates of both winter growth and summer surface melt. Overall, the experiments indicate that accounting for salinity in c is much more influential than in k , in particular during summer. 3) As shown by BL99, the effect of accounting for salinity in the energy of melting $q(S,T)$, compared to assuming a freshwater latent heat of fusion, enhances melt and growth rates - the opposite of the effect of accounting for salinity in $c(S,T)$. Because Figure 3.2b indicates the melt rate decreases with increasing salinity, we know that the salinity effect on $c(S,T)$ must dominate in summer. In summary, higher salinity increases the thermal inertia of sea ice, which leads to lower melt rates, especially at the surface, and feedbacks with slower growth in winter. This finally leads to thicker ice, with a smaller amplitude seasonal cycle.

3.3.2 Impact of the shape of the salinity profile

On the one hand, FY ice thickness (see Figure 3.2a) and, more generally, all simulated physical FY ice characteristics, appear to be insensitive to the ice salinity, except for a slight tendency of faster growth for more saline ice, due to the dominant $q(S,T)$ effect. We expect the same behavior if salinity varies in time, since ice thickness rather than salinity controls ice growth. On the other hand, MY ice characteristics significantly depend not only on the bulk salinity (see Section 3.3.1) but also on the shape of the salinity profile. Figure 3.2a and Table 3.1 show the impact of progressively moving from a constant profile l-2.3 to the

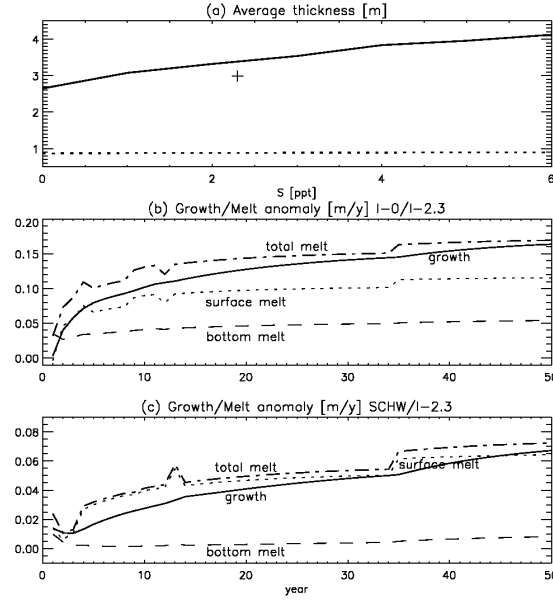


Figure 3.2: (a) Average MY (solid line) and FY ice thickness (dashed line) versus ice salinity, for simulations with a vertically constant salinity profile. The cross indicates the simulation with SCHW profile simulation. (b-c) 50-year evolution of differences in annual amounts of growth and total melt between simulations with (b) I-0 and I-2.3 profiles and (c) SCHW and I-2.3 profiles. Total melt is split into surface and bottom contributions.

non-linear SCHW profile, giving differences in annual mean ice thickness (-47cm), in the amplitude of the seasonal cycle (+7cm) and in the summer surface melt period duration (-4 days).

The importance of the low near-surface salinity is further stressed by the analysis of the transition from I-2.3 to SCHW. First, using the L-PR profile hardly changes the results, while on the contrary, decreasing the top layer salinity value to the almost-zero observed surface value (7-PR) divides nearly by a factor of five the thickness difference between the I-2.3 and SCHW cases. If we keep on getting closer to the SCHW profile, using the SMS profile, the annual mean thickness approaches the SCHW profile thickness value. The importance of surface salinity in increasing the summer surface melt is further clearly shown in Figure 3.2c, showing that throughout the 50 years differences between SCHW and I-2.3 cases are dominated by surface melt. In conclusion, using a constant salinity profile for Arctic multiyear ice yields an overestimation of surface thermal inertia, which results in an overestimation in thickness.

Salinity profile	Mean ice thickness (m)	Bottom ice growth (m/yr)	Bottom ice melt (m/yr)	Surface ice melt (m/yr)	Length of surface melt period (days)	Day of onset of bottom melt	Length of bottom melt period (days)
<i>SCHW</i>	2.99	0.43	-0.06	-0.38	49	199	112
<i>I-0</i>	2.65	0.53	-0.11	-0.43	52	169	101
<i>I-2.3</i>	3.46	0.36	-0.04	-0.31	45	206	116
<i>I-4.6</i>	3.93	0.26	-0.01	-0.26	42	246	106
<i>SMS</i>	2.96	0.45	-0.07	-0.38	49	192	111
<i>L-PR</i>	3.45	0.37	-0.05	-0.31	45	208	117
<i>7-PR</i>	3.11	0.42	-0.06	-0.36	48	200	111

Table 3.1: Summary of results for year 50 of the sensitivity experiments.

3.3.3 Salt/freshwater flux

The contribution from sea ice growth and decay to the salt/freshwater flux between ice and ocean can be represented by the bulk salt flux (BSF) from the ice to the ocean, proportional to the growth rate and salinity difference between ice and ocean (see, e.g., Tartinville *et al.* (2001)). In this representation, we implicitly assume that the ice rejects some salt into the ocean (BSF positive) while it grows, and "accumulates" some salt when it melts (BSF negative). The latter assumption is somewhat unphysical, but reflects that there actually are, during the melt period, two simultaneous fluxes. The first one is a salty brine flux expelled by the ice into the ocean, while the other is a freshwater flux coming from the ice melting. The negative BSF therefore represents the combination of a salt flux and a freshwater flux, dominated by the freshwater exchange and giving a "negative" BSF. By comparing Figures 3.3a and 3.3b, we infer that the growth rate is the main factor responsible for the differences in the integrated bulk salt flux (IBSF, defined at time t as the time-integral of the BSF calculated between the first day of the year and t).

However, because the ice salinity is much smaller than the ocean salinity, the dependence of IBSF on the ice salinity for both FY and MY ice is weak. A more saline ice implies a smaller IBSF. The salinity-induced relative difference in IBSF for FY and MY ice is about 10-20% and is principally caused by different growth rates. From Figure 3.3b, one can also see how the IBSF looks if the SCHW profile is used. The IBSF seasonal cycle appears very similar to the I-2.3 one (see Figure 3.3b). Nevertheless, the SCHW profile leads to an artificial sink of salt, since the ice can melt (at the surface) at a different salinity from the one it has been formed (at the bottom), stressing the need for a salt drainage model.

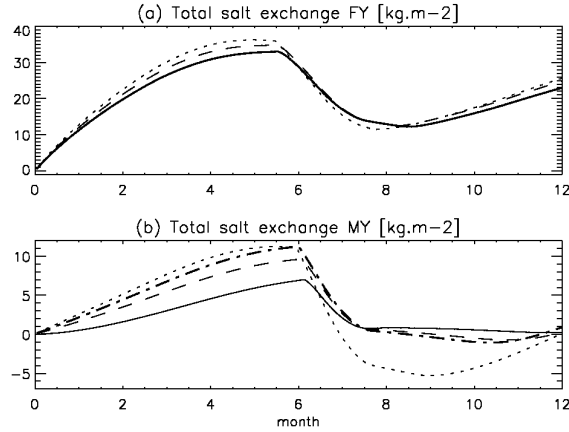


Figure 3.3: (a) Seasonal cycles of the integrated bulk salt flux (IBSF) between ice and ocean (positive to the ocean), for FY ice and for different salinity profiles, i.e. I-4.6 (solid line), I-2.3 (dashed line) and I-0.0 (dotted line). (b) Seasonal cycles of IBSF for MY ice and for different salinity profiles, i.e. SCHW (dashed-dotted line), I-4.6 (solid line), I-2.3 (dashed line) and I-0.0 (dotted line) profiles.

3.3.4 Discussion on application

It is important to remember that the model we used proved to be the most sensitive in summer, when it is known to be less accurate (Vancoppenolle and Fichefet, 2005). Our study further stresses the prominent importance of summer processes on the equilibrium thickness. The model uncertainties during summertime are not surprising, since melt ponds, porosity, brine drainage, meltwater flowing in the ice matrix and other processes specific to the summer are not represented. Finally, the conclusions of our study should be restricted to the Arctic, and not extended to the Southern Hemisphere, where the ice regime is different, with summer melt almost absent and a physically different sea ice (Eicken, 1998). Using more layers in the snow or reducing the number of ice layers to 3 does not significantly affect the simulated ice thickness. Nevertheless, thin ice growth on synoptic timescales was not evaluated.

3.4 Conclusions

We have shown how salinity affects the thermodynamics of Arctic sea ice and the freshwater/salt exchange between ice and ocean with an energy-conserving

thermodynamic sea ice model. About the thermodynamic influence of ice salinity, our results stress that the presence of salt in the ice increases its thermal inertia, and that this thermal inertia mostly affects summer surface melt rate. Capturing the low surface thermal inertia appears to be the key aspect, since it controls - through a specific heat effect damping temperature variations - the magnitude of summer melt, which controls the equilibrium thickness of the ice. A result similar in all points (Vancoppenolle and Fichefet, 2005) was obtained with high-frequency SHEBA forcing including a more realistic ocean heat flux provided by Huwald *et al.* (2005a). Regarding the salt/freshwater flux between ice and ocean, our results indicate that changing the ice salinity from 0 to 4.6‰ lowers the integrated bulk salt exchange by up to 50% between ice and ocean in the perennial ice zone.

Consequently, since the intense summer desalination by percolation of meltwater (flushing) drives the low surface salinity and the shape of the MY salinity profile in the Arctic (Untersteiner, 1968), but rarely occurs in the Antarctic (Eicken, 1998), a prerequisite to the development of a full halo-thermodynamic sea ice model would be to parameterize the flushing. Using a vertically constant salinity profile, even varying in time (instead of a MY ice salinity profile) would result in too thick ice through the overestimation of surface thermal inertia. The results shown here thus indicate that the simplest treatment of the temporal evolution of sea ice vertical salinity profile in thermodynamic models would be to assume a simple shape for the salinity profile, vertically constant for FY ice and linear for MY ice. An improved freshwater budget could be inferred with the time variations of the bulk salinity.

Acknowledgments.

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Chapter 4

Summer landfast sea ice desalination at Point Barrow, Alaska: modeling and observations (VBF07)

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Abstract.

Landfast sea ice cores from two sites in Point Barrow, Alaska, between 1999 and 2001 show a progressive desalination and a corresponding shape transition in the salinity profile starting at snow melt onset, around June 1. The vertical percolation of fresh surface melt water through the permeable ice matrix (flushing) has long been supposed to control this transition.

A parameterization of flushing in bare ice is formulated. It is incorporated into the semi-empirical winter sea ice desalination model of Cox and Weeks (1988), coupled to the one-dimensional thermodynamic sea ice model of Bitz and Lipscomb (1999) and forced under Point Barrow conditions. Adjustment to the original Cox and Weeks parameterization of gravity drainage was necessary to give reasonable agreement with observations in winter. With this change the model has the potential to simulate the full seasonal cycle of the Arctic salinity and mass balance of typical Arctic ice.

In summer, the model salinity profile closely follows the observations. The model upper ice temperatures are slightly too cold. At the thin-snow (Chukchi Sea) site, the model simulates the observed snow depth and ice thickness well. At the thick-snow (Elson Lagoon) site, the snow disappears 7 days earlier than observed, which results in underestimating ice thickness. Sensitivity experiments suggest that a more realistic representation of snow and melt water physics would significantly improve the simulation. Because of the effects of brine drainage on the sea ice mass balance and oceanic circulation, including the time dependence of the ice salinity profile would significantly improve the next generation of sea ice models.

4.1 Introduction

4.1.1 Importance of sea ice and its salinity

Sea ice acts as a damping membrane between atmosphere and ocean, modulating exchanges between them in polar regions. Transfer of radiation (through high albedo), momentum, heat (through low thermal conductivity) and mass (through freshwater and tracers) exchanges are affected.

Sea ice is a saline medium. A fraction of oceanic salt is trapped in the ice during formation, while the remainder is rejected into the ocean. The salt in sea ice is not locked within the ice crystalline lattice, but dissolved within liquid inclusions of brine. As this brine is progressively drained, the ice desalinates (Cox and Weeks, 1974; Weeks and Ackley, 1986).

The ice salinity affects sea ice thermodynamic properties. First, each part per thousand (‰) of sea ice salinity lowers the seawater freezing point by 0.054°C (Assur, 1958). Second, the phase composition of sea ice depends on salinity and temperature. The saltier and warmer the ice, the larger the relative brine volume (Frankenstein and Garner, 1967). Thus, the sea ice thermal properties depend on salinity and temperature (Malmgren, 1927; Untersteiner, 1964; Ono, 1967). The thermal properties regulate growth and melt rates at the ice interfaces, as well as the vertical temperature profile in the ice. In particular, in the Arctic, the sea ice salinity close to the surface seems to be a key factor, since it influences the summer melt rate and equilibrium multi-year ice (MY) thickness through a prominent control of heat transfer and storage in the ice (Vancoppenolle *et al.*, 2005). Near the surface, MY ice is fresher, has thus a lower specific heat, which yields weaker vertical temperature gradients and, in turn, reduces heat conduction through the ice. Finally, internal brine-pocket melting is reduced and surface ablation is increased accordingly.

The scope of the role of sea ice salinity actually goes beyond its influence on sea ice thermodynamics, as it is now largely accepted that the brine volume connectivity influences the nutrient supply for the microorganisms and algae living in sea ice (Ackley and Sullivan, 1994). In addition, the size and shape distribution of brine inclusions affect shortwave radiation scattering (Perovich, 1998; Light *et al.*, 2003) and strength and mechanical behavior (Timco and Frederking, 1990; Kovacs, 1997) in the ice. Sea ice salinity indirectly controls growth/melt rates and thus freshwater and brine exchanges between ice and ocean (Vancoppenolle *et al.*, 2005). The salt/freshwater fluxes associated with ice growth/melt have a strong impact on the oceanic salinity stratification, critical to the circulation of the high-latitude ocean and the World Ocean thermohaline circulation (Aagaard and Carmack, 1989). In summer, the melt water pathways and storage at the ice surface, driven by the ice salinity-dependent permeability structure, regulate the size of melt ponds, and, in turn, surface albedo (Eicken *et al.*, 2002, 2004). Finally, the sea ice salinity affects the carbon cycle in polar regions. Semiletov *et al.* (2004) reported that open brine channels form an important spring/summer CO₂ pathway to the Arctic Ocean. This is corroborated by recent observations in the ice pack off the Adélie-Wilkes Coast (spring) and in the Weddell Sea (summer), which show significant carbon fluxes between atmosphere and ocean through the sea ice, primarily modulated by a very dynamic biogeochemistry. The latter appears to be extremely sensitive to the brine network connectivity, which shows drastic changes around the 5% brine volume threshold (J.L. Tison and B. Delille, personal communication).

4.1.2 Observations of ice salinity

The sea ice salinity profile evolves in space and time (Malmgren, 1927; Eicken, 1998). The bulk salinity decreases with time as the ice thickens, as indicated by the significant salinity versus thickness relationships (Cox and Weeks, 1974; Kovacs, 1996).

In the Arctic, first-year ice has a C-shaped salinity profile (see, e.g., Nakawo and Sinha (1981). A first, rapid desalination stage is observed to occur in the first two weeks of ice formation, bringing the ice salinity from around 20 ‰ to roughly 10 ‰, by the time the ice is around 30 cm-thick. Then, a second, slower, desalination stage, occurs during the remainder of winter, bringing the ice salinity to around 5-6 ‰, by the beginning of its first melt season (Kovacs, 1996).

MY ice, which has experienced summer melt at least once, has a different salinity profile. The upper 50 cm of the ice are almost fresh. Below the fresh layer, the

salinity increases quickly with depth and reaches a maximum typically of 4 ‰. Very similar MY ice salinity profiles have been found at different locations (e.g., Schwarzacher (1959), Tucker *et al.* (1987), Eicken *et al.* (1995). Therefore, the MY ice salinity profile has been hypothesized to be very stable in time and often has been referred to as the *multi-year equilibrium salinity profile* (Schwarzacher, 1959; Untersteiner, 1968). Although the distinction between FY and MY salinity profiles is well documented, the desalination mechanism that causes the transition has received little attention, as outlined by Weeks (1998). Only a few Arctic early summer salinity profiles are documented in the literature. Based on 51 FY ice cores taken from the Fram Strait region, Tucker *et al.* (1987) show a gradual and intense desalination takes place from June to July.

In the Southern Ocean, the situation is quite different. Overall, Antarctic sea ice is slightly more saline (between 0.5 and 1.0 ‰) than its Arctic counterpart, as observed in the Weddell Sea (Gow *et al.*, 1982, 1987). In addition, salinity profiles around Antarctica have more variable shape (Eicken, 1992), probably due to intense rafting, surface flooding and snow ice formation (Maksym and Jeffries, 2000). Furthermore, very little difference between FY and MY salinity profile shapes have been seen, as indicated by a study of 89 sample cores extracted from winter Weddell Sea ice by Eicken (1998). The lack of surface melt in summer (Andreas and Ackley, 1982) alone might keep the surface layer in Antarctic relatively saline compared to Arctic MY ice surface layer in Antarctic MY ice. However, Haas *et al.* (2001) have recently suggested that the picture is more complex in the Weddell Sea, where the heavy snow layer encourages superimposed ice formation as well as the occurrence of liquid gap layers at the snow-ice interface. Superimposed ice is formed by snow surface melting, percolation, and refreezing on the top of the ice (Gerland *et al.*, 1999).

4.1.3 Modeling heat transfer and salinity variations in sea ice

First attempts in modeling the vertical heat transfer and storage and mass balance in sea ice thermodynamics were made in the 1970s (Maykut and Untersteiner, 1971; Semtner, 1976). Recently, improvements in the representation of thermodynamic processes in sea ice models (Bitz and Lipscomb, 1999; Taylor and Feltham, 2004; Huwald *et al.*, 2005b; Notz, 2005) have been motivated by the importance of sea ice on high latitude climates, which themselves have a prominent role in the global climate system. Currently, most climate models include at least a sea ice thermodynamic component, such as the one of Semtner (1976) or the more recent one of Bitz and Lipscomb (1999) simulating sea ice temperature

profile and thickness as main prognostic variables. Only the latter has salinity-dependent thermal properties but hold the vertical sea ice salinity profile fixed in time. To correct this has frequently been reported as a next important step in their development (Leppäranta, 1993; Bitz and Lipscomb, 1999; Eicken, 2003).

Modeling the temporal evolution of the sea ice salinity profile has been approached several times. Untersteiner (1968) theoretically tried to find which brine drainage mechanism was most responsible for the characteristic shape of the equilibrium MY salinity profile in the Arctic (see, e.g., Schwarzacher (1959)). He suggested that intense summer desalination induced by downward percolation of surface melt water through the permeable ice cover, what he called *flushing*, was the likely best candidate.

A numerical model of winter desalination of FY sea ice was developed by Cox and Weeks (1988) (hereafter referred to as CW88), based on laboratory observations of Cox and Weeks (1975). It relies on parameterizations of *salt entrapment* during ice growth, and desalination by *gravity drainage* and *brine expulsion*. The simulated salinity profiles looked roughly in agreement with current knowledge. Nevertheless, an attempt by Cox (1990) to validate the model against Arctic observations was unsuccessful. The modeled desalination rates were too strong and an improbable suction of salt had to be added to gain a better agreement with observations. In contrast to Cox's study, Eicken (1992), tested the CW88 model for sea ice in the Weddell Sea and successfully compared the simulated profiles to the observed ones. The agreement was independent of the growth mechanism (i.e., congelation or frazil), but the date of ice growth onset was found to be of primary importance. In a study of Ross Sea ice using the CW88 model, Maksym and Jeffries (2000) needed to add the impact of seawater flooding during snow-ice formation to have reasonable agreement with observations.

The CW88 model only applies to non flushing situations. Until now, no parameterization of the flushing mechanism has been proposed. Thus, previous sea ice salinity models could not simulate a complete seasonal cycle of the ice salinity in the Arctic. In addition, there is no coupling between salinity and temperature in the CW88 model (i.e., the sea ice thermal properties do not depend on salinity and temperature). Furthermore, in the CW88 model, the ice desalination rates have empirical, prescribed functional dependence on model diagnostics (e.g., vertical temperature gradient or temperature temporal derivative). These functions derive from the fitting data taken in cold room experiments by Cox and Weeks (1975).

Recently, it has been proposed to apply mushy-layer theory to sea ice. It provides an integrated approach of modeling sea ice halo-thermodynamics Worster and Wettlaufer (1997). "Mushy-layer" is a metallurgic term standing for two-phase, two-component reactive porous medium. The mushy-layer equations include the conservations of heat, brine salinity and momentum. A coupling term between fluid and heat transport is included. The bulk sea ice thermal properties are weighted means of the respective contributions from the solid and liquid phases. Within the mushy-layer framework, Taylor and Feltham (2004) solved the heat conservation equation. In their study, the salinity profile was constant in space and time and the vertical fluid transport was prescribed. Therefore, in practise, inside the ice, their thermodynamical approach was equivalent to Bitz and Lipscomb (1999). But, since their purpose was to model the evolution of melt pond depth on sea ice at SHEBA (Surface HEat Budget of the Arctic ocean), their model also included sophisticated surface melt water and radiative transfer components. Notz (2005), with an enthalpy-based approach, has modeled both heat and salt conservation equations, but so far limited his application to very few test cases.

4.1.4 Goals of this study

This paper addresses the following questions. (1) How does the vertical desalination proceed in undeformed Arctic sea ice, especially in summer ? (2) Is it possible to reproduce the temporal evolution of salinity by using a sea ice model of reasonable complexity ? What degree of accuracy can we expect from this type of model ? (3) What are the physical processes and interactions between heat transfer, brine volume and desalination described by the model. (4) Can modeling help us infer the actual contribution of surface melt water to desalination by downward percolation in undeformed bare ice?

In order to answer those questions, we first analyse a series of salinity profiles observed in 87 fast ice cores from the Point Barrow area, near the north coast of Alaska (Section 4.2). We compare uncertainty due to errors to extraction and horizontal variability. The fast ice has the advantage of being driven mostly by thermodynamic processes.

Then, we propose a new, semi-empirical one-dimensional halo-thermodynamic sea ice model (Section 4.3). The thermodynamic component Bitz and Lipscomb (1999) accounts for the effect of the salinity profile on heat transfer and storage. The halodynamic component (i.e. computing the ice salinity profile) is an extended version of the CW88 model. We add to CW88 a formulation of flushing so that the model is able to work over a complete seasonal cycle. The model

components are coupled through a brine volume function of salinity and temperature. Over a 5 % brine volume threshold, the brine network connects and sea ice is permeable to fluid transport. Our model includes several aspects of mushy-layer physics. We chose the thermodynamic model to have a simplified representation of snow physics (single layer, no salinity, no penetration of shortwave radiation through) in order to reduce the number of complex processes in the model.

Next, the model is run under Point Barrow conditions (Section 4.4). We first try to match observed and modeled ice and snow mass balance (Section 4.4.1). Huwald *et al.* (2005a) have shown an accurate mass balance comparison between model results and observations is thwarted by high variability at small horizontal scales in snow depth and ice thickness. In addition, desalination by flushing depends sensitively on the ice melt onset date. Keeping this in mind and the fact that the observations we use for atmospheric forcing and the initial conditions of the ice salinity profile, ice thickness and snow depth were not made at the exact same site, we emphasize the strength of this study is in exploring the salinity model (Section 4.4.2, 4.4.3) and its sensitivity to snow physics (Section 4.4.4) and internal parameters (Section 4.4.5) rather than attempting to simulate the mass balance. Also, we trace the summer freshwater pathways in the model and compare it to existing observations (Section 4.5). Then we discuss our results and sum up the conclusions of our study (Section 4.6). In a companion manuscript (Vancoppenolle *et al.*, 2006), we study the implications of the time-dependent ice salinity in large-scale ice mass balance simulations.

4.2 Observation of fast ice desalination at Point Barrow, AK

The salinity and mass balance data we use in this study (Perovich *et al.*, 2001) come from 2 seasons of yearlong observations (1999-2000 and 2000-2001) of FY landfast sea ice at 2 different sites in northern Alaska: Chukchi Sea (CS) and Elson Lagoon (EL) (Figure 4.1). The combination of seasons and data sets correspond to four subsets, which we refer to as CS2000, CS2001, EL2000 and EL2001.

The ice from the two sites exhibit some differences. CS is an open site where the wind tends to blow away most of the snow, while EL is more inland, with a local wind pattern tending to trap snow. Thus, snow is deeper at EL than at CS. The ice tends to form significantly earlier at the EL site, but the maximum thickness



Figure 4.1: Map of Point Barrow neighborhood. Both sampling sites are indicated : Chukchi Sea (CS) and Elson Lagoon (EL).

is smaller than in the CS site since the deep snow reduces ice growth rates at EL. The ice at EL is overall slightly more saline than in CS since the water from which ice forms is more saline at EL (*Hajo Eicken*, personal communication]. The ice at EL is often highly sediment-laden, which is rarely the case at the CS site (*Frey et al.*, 2001; *Stierle and Eicken*, 2002).

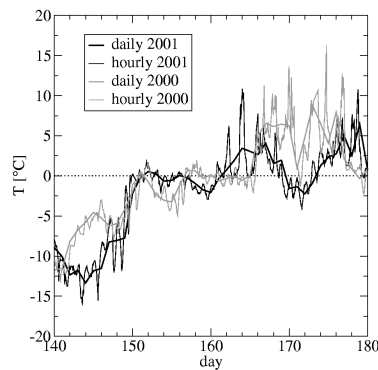


Figure 4.2: Summer evolution of 2m air temperature in 2000 and 2001 according to meteorological station data. June 1st is day number 153 (152) in 2000 (2001).

The two studied seasons are also different. Their winter temperatures (from the Point Barrow meteorological station) are similar, but spring and summer were warmer in 2000 than in 2001 (Figure 4.2). The average 2 m air temperature in May 2000 (2001) was -10.7°C (-14.3°C), and in June it was 2.6°C (1.7°C). The surface melt appears discontinuous during early summer, with temperature being alternatively above (from 0 to 15°C) and below (between -3 and -1°C) the

freezing point (period around 5 days). Superimposed ice was observed to form at the CS site in 2000, but not in 2001, which is coherent with much thicker snow in 2000 than in 2001 (Perovich *et al.*, 2001; Eicken *et al.*, 2002, 2004). We also suspect some superimposed ice formation in EL in late winter / early summer 2000 and 2001, since the snow was thicker than 30 cm both years, but no clear mention of superimposed ice was made in the data record.

4.2.1 Mass balance

Ice/snow mass balance was monitored at both sites and is presented in Table 4.1. The instrumentation at the sites typically consisted of a datalogger, snow stakes, thickness gauges and an above-ice acoustic sounder measuring the height above sea level every hour to within 5 mm. A thermistor string measured vertical profiles of temperature. The thickness gauges consisted of an ablation stake measuring mass loss at the surface and a hot-wire gauge measuring bottom accretion or ablation. Data was manually collected every 1–2 weeks during winter and every 2–4 days during the melt season. Four stake and gauge combinations were installed at each site. Accuracies of stake and gauge readings were about 0.5 cm. EL ice typically forms earlier and has more snow. Since deeper snow takes longer to melt, the onset of sea ice surface melt occurs later at EL. For more details on the data, see Perovich *et al.* (2003).

4.2.2 Salinity profiles

Ice salinity profiles were measured from ice cores. The number of cores and their origin is detailed in Table 4.2. Pond ice samples were only taken at the CS site in 2000 (7 cores). Salinity was measured every 5 cm within a core, and cores were taken as often as every two days in summer. In summer, the cores were mostly confined to the upper 50 cm of ice.

4.2.2.1 Extraction errors

The ice coring approach is known to underestimate salinity since brine drains during core extraction, especially in the lower (warm) parts of the cores (Notz *et al.*, 2005), and in high salinity core sections (Eicken *et al.*, 1991). In summer, when the ice is warmer, the brine loss is potentially higher. In winter, when the ice

Table 4.1: Observed mass balance.

	Chukchi Sea		Elsa Lagoon	
Initial ice thickness, cm	1999-2000	2000-2001	1999-2000	2000-2001
Initial snow thickness, cm	37	22	54	30
First observation date	3	3	3	3
Maximum ice thickness, cm	Nov 12, 99	Nov 2, 99	Nov 10, 00	Nov 3, 00
Maximum snow thickness, cm	154.3	162.8	148.6	157.8
Maximum snow thickness date	36	11	47.8	30.1
Ice surface melt initial date	May 18	May 21	May 26	May 18
Bottom melt initial date	Jun 14	Jun 4	?	Jun 14
	Jun 23	Jun 4	Jun 13	Jun 21

en 50 cm.

TABLE 1.2. ICE CORE DATA. ALL SUMMERY CORE LENGTHS ARE GREATER THAN 50 CM.							
Period	Site	Number of Cores			Bulk Salinity, ‰		Comments
		Total	Summer	Sum. Bare Ice	Winter	Summer	
1999-2000	CS	29	18		11	5.34	3.53
	EL	9	3		3	5.69	3.02
2000-2001	CS	17	9		9	5.69	3.89
	EL	13	6		6	5.73	3.52
		68	30		29	5.61	3.49

is cold, the error in salinity due to extraction should be of a few tenths of permils at most ($<0.2\text{ ‰}$ for ice salinities around 5 ‰). However, near the bottom, the ice is always relatively warm and can easily have a loss in brine salinity of more than 1 ‰ . On the other extreme, in summer, when the ice is warm, the loss of brine can be substantial and depends on how quickly the core is processed after extraction. For high salinities ($>5\text{ ‰}$), the brine loss may be as high as 0.5 to 1 ‰ . However, for warm, low salinity ice, the impact of brine loss on salinity will not be as high (e.g., around 0.2 ‰ for a 1 ‰ salinity). In both cases, the salinity deficit due to brine loss represents from 10 to 20% of the initial salinity.

4.2.2.2 Horizontal variability

We briefly describe the local horizontal variability of the salinity profiles as an estimate of their uncertainty. The discrete nature of brine drainage channels is a source of horizontal variability below about 10 cm (Cottier *et al.*, 1999). This variability is overcome by the core diameter (10 cm). Larger-scale variability, at 2-20 m scale, is controlled by variability in brine drainage (Tucker *et al.*, 1984; Eicken *et al.*, 1991).

In the year 2001, a few synchronous salinity profiles were measured at 10-m intervals at both sites in late winter and early summer. The mean and standard deviation of the intra-site bulk salinities for each time period is given in Table 4.3. The bulk salinities reported in early summer are only for the top 50 cm due to brine drainage difficulties in the lower parts of the cores. The standard deviation of the intra-site bulk salinity ranges from $0.22\text{--}0.28\text{ ‰}$ in late winter and $0.05\text{--}0.44\text{ ‰}$ in summer.

Naturally the salinity at a particular depth varies more than the bulk salinity across the ice cores. Figure 4.3 illustrates the depth-dependent variations in salinity. The late winter cores have a characteristic C-shape, as described in Section 4.1.2, with variations in the overall amplitude of bulk ice salinity that depend mostly on the thickness. When profiles are normalized by dividing the vertical coordinate by the core length, as shown in Figure 4.3, the standard deviation for a 5-cm layer at a given normalized depth is on average 0.76 ‰ at CS and 0.88 ‰ at EL in late winter.

In early summer, rapid desalination at melt onset adds to the difficulty of estimating the horizontal variability in the salinity profiles. We compare cores taken at most 2 days apart, but it is still possible that our horizontal estimates are made somewhat larger by temporal variability as well and hence are more conservative.

Table 4.3: Horizontal variability of ice bulk salinity, 2001. σ refers to the inter-core standard deviation. The single-core mean salinities are weighted according to the length of each core section. The inter-core means are arithmetic means, computed independently of core length.

	Winter - Spring					
	Mar 12	Mar 12	May 3	May 4	May 4	σ
<i>Chukchi Sea</i>						
Core length, m	1.13	1.29	1.40	1.54	1.45	0.14
Mean salinity, ‰	5.17	5.47	5.04	5.64	5.50	0.22
<i>Elson Lagoon</i>						
	Mar 11	Mar 14	May 1	May 5		σ
Core length, m	1.11	1.13	1.26	1.42		0.12
Mean salinity, ‰	5.99	5.49	5.22	5.48		0.28
	Early summer					
<i>Chukchi Sea</i>						
	Jun 7	Jun 9	Jun 9			σ
Core length, m	0.50	0.50	1.46			0.45
Mean salinity (upper 50 cm), ‰	3.36	2.29	2.97			0.44
<i>Elson Lagoon</i>						
	Jun 8	Jun 9				σ
Core length, m	0.45	1.38				0.47
Mean salinity (upper 50 cm), ‰	3.20	3.29				0.05

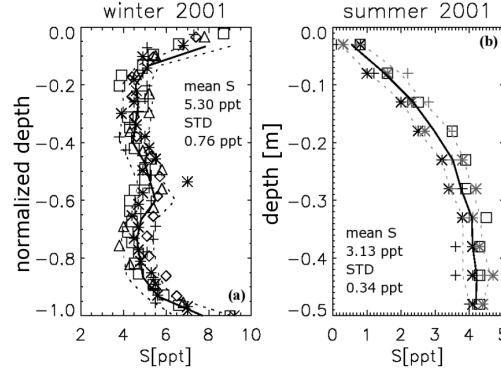


Figure 4.3: Horizontal variability of vertical salinity profile, 2001. (a) Chukchi Sea site, winter. Five normalized profiles are shown, with vertical axis corresponding to normalized depth : two close profiles, March 12th (crosses, stars), at a distance of about 10 m; one profile, May 3rd (squares); and two close profiles, May 4th (triangles and diamonds). The mean profile (solid line) and the one standard deviation range (dotted lines) are shown. (b) Chukchi Sea and Elson Lagoon sites, early summer. Five surface salinity profiles (i.e., restricted to upper 50 cm) are shown. Three profiles from the Chukchi Sea site (black symbols) : one profile, June 7th (squares), and two profiles from sites close to each other, June 9th (crosses and stars). Two profiles from Elson Lagoon Site (grey symbols) : two profiles, June 8th (crosses) and June 9th (stars). The mean upper 50 cm salinity profile (solid line) and the one standard deviation range (dotted lines) are shown.

The standard deviation for a 5-cm layer at a given normalized depth is on average 0.34 ‰ in summer, which is comparable to the standard deviation of the bulk salinity for the whole 50-cm layer.

In summary, we consider that the average error due to horizontal variability in bulk salinity is 0.25 (0.30) ‰ in winter (summer) and that the error at a particular depth is 0.82 (0.34) ‰ in winter (summer). In late winter, the uncertainty due to horizontal variability exceeds the error due to brine loss upon extraction, except near the bottom where the extraction error is considerable. In early summer, these two sources of error are comparable in the upper 50 cm, and we expect the extraction error dominates below this layer.

4.2.2.3 Summer desalination sequences

We have a sufficient number of salinity profiles to visualize bare ice desalination sequences for CS2000, CS2001 and EL2001. Furthermore, for CS2000, we also

have 7 salinity profiles of ponded ice, so we can contrast salinity in bare ice and ponded ice.

In summer 2000 at the CS site, the average salinity of bare ice (3.38 ± 0.65 ‰) was significantly higher than the salinity of ponded ice (2.50 ± 0.55 ‰). The shapes of the profiles were also different (Figure 4.4). In bare ice, the typical more or less linear profile with low surface salinity is well established (i.e., the bulk salinity of the uppermost 50 cm of ice is under 3 ‰) after 11 days of surface melt, if we assume that surface melt started on June 1st (cfr. Figure 4.2). On June 15th though, a special event increased the surface salinity by 2 ‰. This event could be due to lateral or vertical intrusions of salty water from various origins (seawater flooding, lateral drainage of brine). After this event, progressive desalination was typical again. In ponded ice, the desalination sequence is mostly continuous and the ice reaches a roughly vertically constant profile, with a salinity slightly above 2 ‰ on June 23rd.

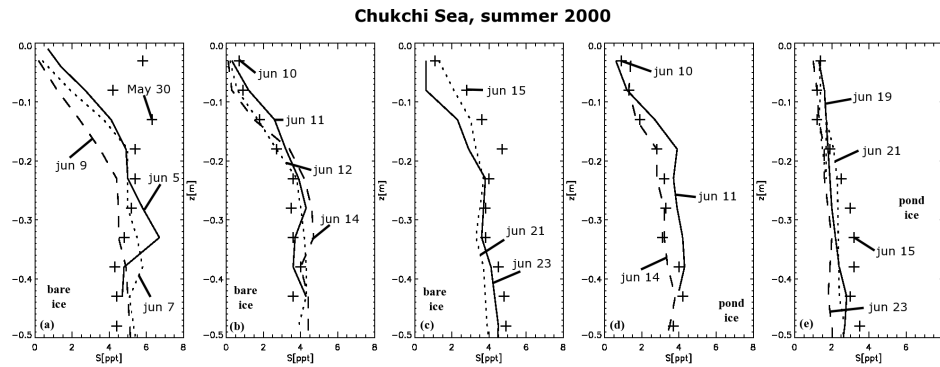


Figure 4.4: Chukchi Sea site, summer 2000. Observed surface desalination sequence during the early melt period for bare ice (a-c) and ponded ice (d-e).

In 2001, the bare ice desalination sequences observed at CS (7 profiles, Figure 4.5a) and EL (6 profiles, Figure 4.5b) sites show progressive, practically monotonic, desalination patterns, with no apparent seawater flooding as in 2000. Unfortunately the first core at CS was taken on June 2nd, after the onset of flushing. At EL, the rapid desalination started between June 1st and June 5th. At both sites (particularly visibly at EL), we can see the ice salinity profile rapidly switch from a C-shaped to a more or less linear one with low surface salinity. After 13 (12) days of surface melt at CS (EL), considering that surface melt started on June 1st (see Figure 4.2), the uppermost 50 cm of ice had a salinity under 3 ‰.

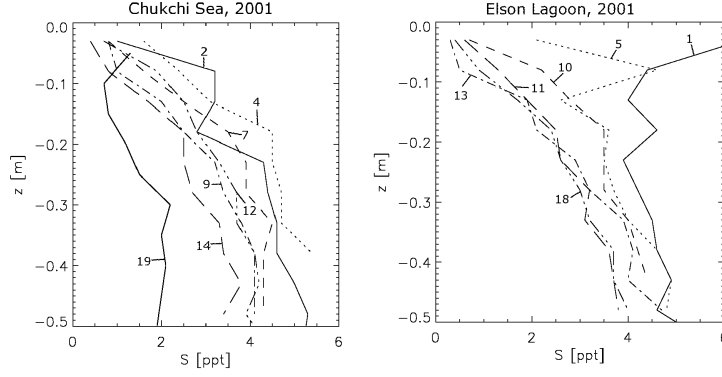


Figure 4.5: Desalination sequences at Chukchi Sea and Elson Lagoon sites during the early melt period, 2001. The numbers correspond to the days of June.

4.3 The model

The model we use in this study is a semi-empirical one-dimensional model of undeformed, bare (i.e., unpounded), Arctic sea ice, resulting from the coupling of thermodynamic and halodynamic components. We are currently working on the model to extend it to Antarctic situations. The thermodynamic component computes heat transfer and storage (i.e. temperature profile) as well as ice thickness and snow depth. The halodynamic component resolves brine entrapment and drainage mechanisms, which lead to the computation of the vertical salinity profile. In addition, the haline component treats melt water flushing, but the thermodynamic component ignores melt water thermal effects like ponding and refreezing.

4.3.1 Thermodynamic component

The thermodynamic sea ice model used here is the one-dimensional energy-conserving model of Bitz and Lipscomb (1999). Model prognostic variables are the ice thickness h_i , the snow depth h_s , the vertical temperature profile $T(z, t)$ in the ice-snow system and the surface temperature $T_{su}(t)$, where t is the time and z is the vertical coordinate. $T(z, t)$ is computed by numerically solving the heat diffusion equation. $T_{su}(t)$ and the growth/melt rates dh_x/dt (where suffix x refers either to snow or ice) are deduced from the heat balance between inner conduction fluxes and atmospheric (respectively oceanic) external fluxes at

the upper (respectively lower) interfaces. The thermodynamic effect of salinity, and its control on the size of brine pockets, is represented by the functional dependence of the thermal properties on salinity and temperature (which affect heat diffusion and growth/melt rates). The atmospheric fluxes include radiative shortwave (SW), radiative longwave (LW), and turbulent sensible and latent heat components.

The ice cover is represented with one layer of snow on top of N layers of sea ice. The snow layer has a temperature T_s and each layer ($l = 1, \dots, N$) of sea ice has a specific temperature T_i^l and salinity S_i^l (see Figure 4.6).

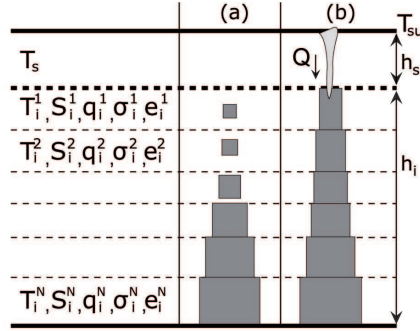


Figure 4.6: Vertical grid of the sea ice when snow is present. Typical brine network configuration in winter (a) and summer (b). h_s and h_i are the snow and ice thicknesses. T_{su} , T_s , and T_i^j ($j=1, \dots, N$), are the surface, snow and ice layers temperatures, respectively. S_i^j , q_i^j , σ_i^j and e_i^j ($j=1, \dots, N$) are the ice layers bulk salinity, energy of melting, brine salinity and brine volume, respectively. When flushing occurs, a desalinating mass flux Q , prescribed as a fraction ϕ of the surface melt, penetrates the interconnected brine network.

The theoretical basis of this model relies on two assumptions involving model state variables, the temperature T and the salinity S . (1) The approximate freezing temperature of a saline solution, T_f (in $^{\circ}\text{C}$), is a linear function of its salinity σ given by

$$T_f = -\mu\sigma, \quad (4.1)$$

where $\mu = 0.054^{\circ}\text{C} \text{ } \text{‰}^{-1}$, as determined by Assur (1958). (2) The brine pockets are in local thermal equilibrium with the surrounding ice and permanently at their freezing temperature, so that they adapt their salinity σ by changing size to the local ice temperature:

$$T = -\mu\sigma. \quad (4.2)$$

In this context, σ is the salinity of the brine pockets, while S is the salinity of the ice if brine were distributed uniformly throughout the ice. Thus, σ and S are

related to each other by :

$$\sigma e = S. \quad (4.3)$$

The relative volume of brine inclusions e (dimensionless, hereafter referred to as *brine volume*) is consequently only a function of S and T (Schwerdtfeger, 1963) :

$$e = -\mu \frac{S}{T}. \quad (4.4)$$

The brine density ρ_b is given from Zubov (1945) by assuming a linear dependence on brine salinity,

$$\rho_b = \rho_w(1 + c\sigma), \quad (4.5)$$

where $\rho_w = 1000 \text{ kg m}^{-3}$ is the fresh water density and $c = 0.8 \cdot 10^{-3} \text{ } \text{‰}^{-1}$ is an empirical coefficient.

The mathematical formulation of the thermal properties of the ice relies on the brine volume. First, the specific heat, $c_p(S, T)$, describes the magnitude of heat storage in the ice. It is given according to Malmgren (1927) and Ono (1967), and can at warm temperatures be two orders of magnitude larger than its asymptotic limit for cold ice. Second, the thermal conductivity, $k(S, T)$, regulating the internal heat transfer, is specified according to Untersteiner (1964). $k(S, T)$ is decreased at most by 20% at warm temperature. The third S-T dependent thermal property is the sea ice energy of melting $q(S, T)$, defined as the energy required to melt a unit volume of sea ice of salinity S and temperature T . It modulates the growth/melt rate of sea ice at its interfaces with atmosphere and ocean. It is significantly lower than the standard latent heat of fusion of pure freshwater ice, because it takes into account the internal melt. The model conserves energy, salt and ice mass, but not water mass.

The radiation scheme assumes that a fraction i_o of the incoming net solar radiation does not contribute to the surface energy balance and penetrates inside the snow/ice system and is absorbed and transmitted according to Beer's law. The details of the thermodynamic model can be found in Bitz and Lipscomb (1999).

The simplification of snow physics is certainly one of the limits of the model. Snow cover is highly complex. Its depth and properties are highly variable on small horizontal scale (Sturm *et al.*, 2002), even on the same floe. Since snow is not the main focus of the paper, we only use one layer of snow, with no penetration of solar radiation, as in, for example Maykut and Untersteiner (1971), or more recently Taylor and Feltham (2004). The snow is assume to be of zero salinity, though it was observed to be slightly saline (especially in the lower layers) in landfast FY ice of the Canadian Archipelago (Mundy *et al.*, 2005). Snow compaction and melt water refreezing are also neglected.

4.3.2 Halodynamic component

The halodynamic component of the model computes the temporal evolution of the vertical salinity profile $S(z, t)$. Experiments performed in the sixties and the seventies (see Weeks and Ackley (1986) for a review) led Untersteiner (1968) and CW88 to formulate the changes in the salinity profile $S(z, T)$ by

$$\frac{dS}{dt} = \left. \frac{dS}{dt} \right|_{exp} + \left. \frac{dS}{dt} \right|_{gd} + \left. \frac{dS}{dt} \right|_{flu}. \quad (4.6)$$

The different terms on the right hand side correspond to brine expulsion, gravity drainage and flushing. The bottom boundary condition depends on how salt is trapped in the ice during growth and is done as in CW88: each new layer formed at the bottom of sea ice has a salinity given by the multiplication of the fractionation coefficient f of CW88 by the salinity of the underlying ocean S_w . The more rapid the ice growth, the more salt trapped inside the ice. New sea ice is assumed to form at the sea-water freezing point ($T_b = -\mu S_w$). Thus, the initial brine volume of new ice is $e_b = f$. For the first two tendency terms of the salinity equation, we use slightly modified versions of CW88 equations. For the flushing term, we propose a new formulation.

4.3.2.1 Brine expulsion

When a brine pocket cools, the thermal contraction of the ice around the pocket is greater than that of the liquid inside, producing a pressure gradient that may lead the pocket wall to rupture. Actual observations of brine expulsion are still lacking, so that the direction of the expulsion flow is not well known, although indications of both downward and upward flow are suggested (Cox and Weeks, 1975; Cottier *et al.*, 1999). Saline snow (Mundy *et al.*, 2005) and the presence of highly saline frost flowers on top of new ice (Martin *et al.*, 1995) also support the upward migration of brine in sea ice. Nevertheless, the mechanism through which this happens is not yet clear.

Through volume conservation arguments, Cox and Weeks (1986) provide an expression for the subsequent expulsion desalination rate of a piece of ice of uniform temperature. The expulsion rate is specified as a function of the cooling rate through temperature-dependent formulas of brine density $\rho_b(T)$ and salinity $\sigma_b(T)$. For each layer, we assume local equilibrium and use the functional expression E of Cox and Weeks :

$$\left. \frac{dS}{dt} \right|_{exp} = E \left[\frac{dT}{dt} \right], \quad (4.7)$$

though with different formulas (equations 4.4 and 4.5) for $\rho_b(T)$ and $\sigma(T)$, to be consistent with our thermodynamic model component. The impact of this difference on the model results is not significant, since the magnitude of this process proves to be very small. The temperature tendency (dT/dt) is computed by the thermodynamic component. Compared to CW88, we assume that the brine is not expelled outside the ice. Instead, we assume that the brine is transported equally upwards and downwards, with a barrier to flow at the upper surface and free passage into the ocean beneath. This is motivated by the fact that brine expulsion leads more to a redistribution than to a net loss of salt in the ice (Notz, 2005).

4.3.2.2 Gravity drainage

Gravity drainage, the main reason for early winter desalination rates, gathers all processes where brine is expelled from the ice under the influence of gravity through an interconnected brine network (Weeks and Ackley, 1986).

The actual mechanism beyond gravity drainage is not clear. First, ice growth-driven gravity drainage was proposed by Untersteiner (1968) and Cox and Weeks (1975). The upward advection of ice layers driven by ice growth transports brine above sea level and creates a pressure head having the potential of expelling the brine out of the ice. The second suggested mechanism is a temperature-driven gravity drainage. In winter, the ice is the warmest at its base, which creates an unstable brine density profile. If the ice is porous enough, convection ensues, causing net downward salt transport and a local decrease in salinity. Similarly, the downward temperature-induced brine salinity gradient induces molecular diffusion of salt inside the permeable region. If the brine tubes are thin enough, the capillarity forces are strong enough to compensate the destabilizing buoyancy forces, as suggested by the data from *Mercier et al.* [2005].

Cox and Weeks (1975) indirectly observed gravity drainage as a residual between observed desalination rates and computed brine expulsion. Then, CW88 devised an empirical regression formula linking the desalination rate, brine volume e and temperature T gradient. As in CW88, we assume that gravity drainage depends on the temperature gradient and brine volume. If $\partial T/\partial z < 0$ and $e > 5\%$, the desalination rate is

$$\left. \frac{\partial S}{\partial t} \right|_{gd} = \delta(1 - \eta e) \frac{\partial T}{\partial z}, \quad (4.8)$$

and 0 otherwise. δ and η are empirical coefficients. δ controls the magnitude of the desalination flow. Contrary to CW88 who proposed to use $\delta_{lab} = 1.68 \cdot 10^{-7}$

$\text{‰}^\circ\text{C}^{-1}\text{m s}^{-1}$, based on lab observations, we recommend to use $\delta_{mod} = 5.88 \cdot 10^{-8} \text{‰}^\circ\text{C}^{-1}\text{m s}^{-1}$ (suitable for our modeling study), which is 65% smaller than their value, for reasons we explain in section 4.4.2. The value $\eta = 20$ just reflects the fact that gravity drainage is 0 for $e = 5\%$ and then its intensity increases (i.e., $\partial S/\partial t|_{gd}$ decreases towards more negative values) for greater brine volumes. We allow no gravity drainage for any layer above an impermeable layer. Since (4.8) implicitly assumes vertical brine flow, it is mass conserving. In other words, the downwelling salt flux from each layer is implicitly transported to the underlying layer.

4.3.2.3 Flushing

Flushing is a special kind of gravity drainage, where the pressure head and water are provided by the fresh surface melt water going through the permeable summer brine network. First mentioned by Untersteiner (1968), it is believed to strongly contribute to the summer surface desalination of pack ice and to be responsible for the shape of the MY salinity profile, known to be important in controlling the surface melt rate and sea ice equilibrium thickness (Vancoppenolle *et al.*, 2005). The picture is actually more complex than a simple one-dimensional vertical downward melt water flow, as shown by observations of summer ice hydrology at SHEBA and later at Point Barrow by Eicken *et al.* (2002). Lateral melt water flow directed towards the lowest topographic features was found to be highly significant, and even dominant during the early stages of surface melt. Our model only considers the vertical component of the melt water flow through bare ice.

Field experiments by Eicken *et al.* (2002) suggest that once a permeability threshold is reached (i.e. when brine channels can be assumed to be connected), any available melt water is almost instantly transferred through the ice matrix. The summer production of melt water at the ice surface is around $10^{-4} \text{ kg m}^{-2} \text{ s}^{-1}$ while the flow the ice can sustain once permeable much higher and therefore flushing can be seen as instantaneous. We thus make the following hypothesis: brine drainage processes are much more rapid than the thermal reequilibrium processes. This prevents the brine channels from enlarging through advection of heat by melt water.

Once the permeability threshold is reached ($e_T = 5\%$, mentioned both by theoretical (Golden *et al.*, 1998) and experimental studies (Eicken *et al.*, 2004)) and the surface is melting ($T_{su} \geq 0^\circ\text{C}$), we assume that flushing occurs. A certain fraction of the melt water flows through the brine channels, replacing salty brine by almost fresh melt water coming from the top of the ice. If these requirements

are not fulfilled, the melt water is assumed to contribute to lateral drainage to ocean through cracks and leads, or to collect in the lowest gravity areas and form melt ponds, as recently observed by Eicken *et al.* (2002).

The ice salinity profile is converted into the vertical profile of brine salinity (using equation 4.3) and brine volume (using equation 4.4). The melt water mass flow Q per unit area of sea ice [$\text{kg m}^{-2}\text{s}^{-1}$] through the ice matrix is then specified as a fraction of the surface mass flux

$$Q = \phi \rho_x \left. \frac{dh_x}{dt} \right|_{su}, \quad (4.9)$$

if $e \geq e_T = 5\%$ and is set to zero otherwise. The subscript x refers to either snow or ice. ρ_x is the density of ice/snow and $dh_x/dt|_{su}$ is the surface melt rate computed by the thermodynamic component. ϕ is the prescribed fraction of the melt water allowed to percolate vertically through the ice matrix. The salinity of this melt water mass is $S_Q = 0\text{‰}$ if melt water comes from snowmelt and $S_Q = S_i^1$ (the salinity of the first ice layer) if the ice surface is melting.

During flushing, Q penetrates the brine network and pushes brine downwards. To compute the new brine salinities after flushing, we solve the brine mass conservation equation:

$$\frac{\partial \sigma}{\partial t} = - \frac{Q}{e \rho_b} \frac{\partial \sigma}{\partial z}. \quad (4.10)$$

We assume that the mass flow is the same through all horizontal brine network sections and that the brine volume is fixed during the process. Once σ is computed, we recompute S (using equation 5.3).

ϕ , the melt water partitioning coefficient and e_T , the brine volume permeability threshold, dominate this parameterization. ϕ controls the impact of the desalination flow on salinity, while e_T controls the flushing onset date. i_0 is an important indirect parameter, since it controls the penetration of net SW radiation to the ice interior, which can connect the brine pockets.

We numerically solve equation (4.10) with an implicit scheme, which is unconditionally stable, but not conservative. Consequently, the solution has to be corrected at each time step to ensure mass conservation. We divide the numerical loss of salt by the number of layers and add the equivalent salinity loss to each layer salinity. The solution has been shown to be practically insensitive to this numerical correction.

At each time step, the vertical grid is recomputed to adapt to the new ice thickness. We linearly redistribute the mass of salt (before and after desalination) and

heat content of each layer onto the new grid. Then we infer the new ice layers temperatures and salinities.

At this time our model only attempts to represent the summer melt water pathways and storage in sea ice for an unponded slab of sea ice. Taylor and Feltham (2004) have developed a model of melt pond evolution in sea ice, but prescribed the vertical flow of melt water through the ice in summer. Our respective approaches could be complementary in a 2-cell model with cells of bare and ponded ice connected by a freshwater flux proportional to $1 - \phi$. Furthermore, an account for snow-ice formation would also make our model more general. Specifying the exact source of flooding seawater during such events might be difficult though, since seawater pathways in sea ice are not well understood. The study of Maksym and Jeffries (2000) indicates that the flooding of the bottommost snow layers during snow-ice formation events is probably not performed by upward percolation of seawater through a permeable brine network. Nevertheless, other potential pathways for seawater to flood the ice, as lateral floe borders, or wider vertical channels (as cracks or large moats) are not, to our knowledge, well documented.

4.3.2.4 Forcing and experimental design

We made three control experiments, corresponding to conditions at CS in 1999-2000 (CS2000-1) and 2000-2001 (CS2001-1) as well as at EL in 2000-2001 (EL2001-1). The model is initialized with observed values of ice thickness and snow depth, temperature and salinity profiles. The objective while setting the experimental design was to calibrate the snow / ice mass balance in order to provide a suitable background to compare the model and observed ice desalination.

It is considered that the mass balance and salinity data have a spatial horizontal scale on the order of the floe size, since 1) four gauges and stakes were used at each site and 2) spatial variability in ice salinity was assessed. The most important sources of forcing variations at scales comparable to the floe size come from variability in clouds, in snow depth and in seawater state (due to local geographic and bathymetric features) for which we do not have enough or any data. Therefore, some tuning was required to adjust the snowfall rate, the downwelling LW radiation and the oceanic heat flux.

We force the model with a hybrid transient and climatological forcing. For temperature, pressure and wind speed, we take daily averages of hourly observations made at the meteorological station at Point Barrow, provided by the CMDL (Climate Monitoring and Diagnostics Laboratory). Cloudiness and relative humidity

come from daily interpolations of widely used monthly climatologies (Berliand and Strokina, 1980; Trenberth *et al.*, 1989).

The incoming SW radiation flux is prescribed as a function of latitude, cloudiness and humidity (Zillmann, 1972). The incoming LW radiative flux is parameterized according to Berliand and Berliand (1952) as a function of air temperature, cloudiness and specific humidity. Mass balance simulations show that the incoming LW is the most likely forcing component to be underestimated. The cloudiness climatology we use is a source of error to which the LW flux is especially sensitive at high temperatures. Since Lindsay (1998) mention that errors in the incoming LW flux with such parameterizations can be as large as 20 Wm^{-2} , taking into account uncertainties linked to clouds, we increased the LW radiation by 10 % to bring the melt onset date to the observed value. The turbulent sensible and latent heat fluxes are computed with bulk aerodynamic formulas described by Goosse (1997).

Since snow depth is highly variable even at sub-kilometer scales (Sturm *et al.*, 2002), we derived 3 specific snow precipitation temporal data sets, one for each case study, from the snow depth temporal evolution data. We interpolated the change in snow depth on a daily time step to infer daily snowfall values. If snow depth happened to decrease, then the precipitation rate was set to 0. In other words, we specify depth when snow accumulates but let snow melt in the model.

The oceanic heat flux F_w is empirically calibrated to improve the agreement between observed and model snow / ice mass balance. At CS (EL), a value of $F_w = 4.5$ (3.0) W m^{-2} is used. These values lie in the range proposed by Krishfield and Perovich (2005). The difference in the F_w optimal values between the two sites suggests that their geographical configuration (CS, open sea; EL, lagoon) affects heat transfer between seawater and ice.

Observations by Eicken *et al.* (2004), made at the same site during similar field campaigns, suggest albedo values of 0.80, 0.65 and 0.50 for dry snow, melting snow, and bare ice, respectively. The fraction of net solar radiation penetrating below the surface i_0 is taken to be 0 in the presence of snow and 0.30 without snow, an average of Grenfell and Maykut [1977] values. i_0 is a key parameter since it controls the amount of internal warming, especially in the upper layers of the ice (where flushing occurs) and also the increase in brine volume. Thermal conductivity of the snow is set to $k_s = 0.31 \text{ W m}^{-2}$, snow density to $\rho_s = 300 \text{ kg m}^{-3}$, as in Bitz and Lipscomb (1999). $N = 20$ layers in the ice and a time step of $\Delta t = 1$ day are used.

4.4 Model results

In this section, we first examine the modeled snow / ice mass balance (Section 4.4.1). Second, we analyse the model desalination in detail for winter (Section 4.4.2) and summer (Section 4.4.3). Finally, the sensitivity of the salt and mass balance to the model representation of the snow (Section 4.4.4) and to the main parameters of the flushing parameterization (Section 4.4.5) is studied.

4.4.1 Mass balance

Figure 4.4.7 shows the modeled and observed mass balance for the control runs CS2001-1 and EL2001-1.

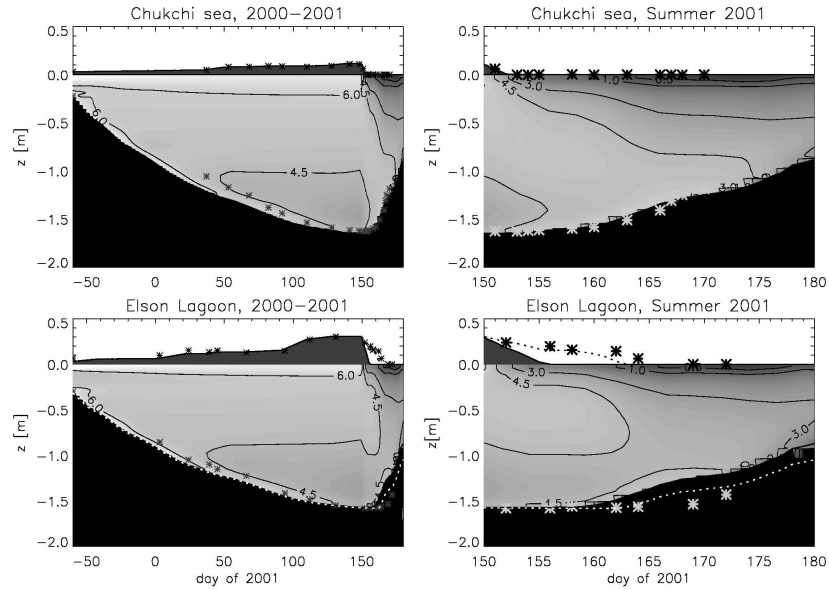


Figure 4.7: Mass balance and contours of salinity [%] as simulated by the model at both sites in 2001 (simulations CS2001-1 and EL2001-1). The stars refer to observed values of ice and snow thicknesses. On the left, the whole ice growth/melt season is shown. On the right, the plots are restricted to the summer season. The dotted curves on the lower plots represent the results of the EL2001-6 simulation (including improved representation of snow physics).

Overall, the differences between observed and simulated ice thickness are within 16 cm (5.2 cm) in winter (i.e., before the onset of surface melt) at CS (EL).

The average error is 0.2 cm (-9.4 cm) after the melt onset (see Table 4.4 for details). In winter, at both sites, the prescribed snow depth matches the available observations. In summer, at the CS site, the average difference in model and observed snow depth equals 0.2 cm. The simulated and observed snow melt period are consistently short (3 days). On the contrary, at the EL site, the snow cover disappears 7 days too early in the model, leading to an average error in model snow depth equal to -7.6 cm. This problem is further investigated in Section 4.4.4.

The relatively large error in the modeled winter ice thickness at CS2001 may be due either to errors in snow depth or physical properties, for which data are lacking, or in the oceanic heat flux F_w , which was not measured at this site. Snow and oceanic heat supply control the ice growth at the bottom interface. We use a constant value of F_w , while its standard deviation has been shown to be up to 15 W m^{-2} by Krishfield and Perovich (2005). Much larger average error in summer ice thickness is found at EL since the snow disappears too early in the model.

For CS2000-1, since summer 2000 salinity data were contaminated by a flooding event, we only focus on the winter period. The ice thickness model-data agreement is within 2 cm.

4.4.2 Winter desalination

Since in winter the salinity profile has a predictable C-shape, we compare the model and observed winter salinity profile normalized by ice thickness. We perform the comparison for winter 2000 at the CS site (CS2000) for which we have the greatest amount of data (10 profiles), for different model configurations (see Figure 4.8).

The original parameterization of CW88 leads to substantial problems while simulating the winter ice salinity. In agreement with the study of Cox (1990), the simulated desalination is much too high, inducing an underestimation of ice salinity of about 2-3 ‰ in the ice interior. Gravity drainage is thus too strong in CW88, and we chose to adapt the gravity drainage intensity coefficient δ . Our value, δ_{mod} , is 65% lower than the CW88 value δ_{lab} . This value gave the best balance between a low enough surface salinity and a high enough bottom salinity. If $\delta < \delta_{mod}$, the salinity close to the ice bottom is correct, but it is too high in the upper layers. If δ is higher, the salinity minimum near the ice bottom is substan-

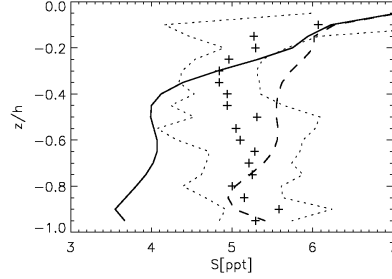


Figure 4.8: Comparison between model and observed normalized average winter salinity profiles, CS2000. Average over 10 observed profiles, taken between February 5th and May 31st, (crosses) and one standard deviation range (dotted lines) are shown. Average simulated profiles (mean taken over the same days) are also presented for different values of δ , which controls the intensity of gravity drainage: *Cox and Weeks* [1988] value δ_{lab} (solid line) and δ_{mod} (dashed line).

tially too low, as shown by the simulation with the original δ_{lab} value. Reducing δ is of limited utility though. Something is clearly missing from the CW88 model.

No small-scale variability in the vertical salinity profile occurs in the model, while it is often reported in observations. In addition, brine expulsion plays almost no role, a role even smaller than in Eicken (1992), since our expelled brine is redistributed in the ice. Its main impact is to slightly increase surface salinity by less than 1‰.

4.4.3 Summer halo-thermodynamics

We present the results of the comparison for summer 2001 at both sites. According to the picture given by Eicken *et al.* (2002), our case study falls in their first and second stages of the melt period, characterized by low ice permeability, snow melt lateral drainage, and some vertical melt water transport.

Figure 4.9 shows the evolution of modeled / observed surface bulk salinity (S_{up}), defined as the salinity in the uppermost 50 cm of the ice. At both sites, the simulated melt water flushing starts on day 151, simultaneous with surface melt. Then, at the CS site, the flushing continues, while it stops between days 154-158 at the EL site where the initial loss of salinity lowers the permeability ($e < e_T$) in the upper layers. Thicker snow at EL contributes to lowering e by delaying warming. The model surface desalination rate at CS (EL) (i.e., computed in the uppermost 50 cm of ice), is equal to -0.27 (-0.23) ‰/day, which is higher than

the bulk (i.e. computed over the total thickness of the ice cover) desalination rate, equal to 0.12 (0.08) ‰/day. As can be seen in Table 4.4 the model-data agreement is very good at both sites (error smaller than 1 ‰ and correlations higher than 0.9). At EL, S_{up} is on average overestimated by 0.57 ‰, which is of the same order of magnitude as the observation error.

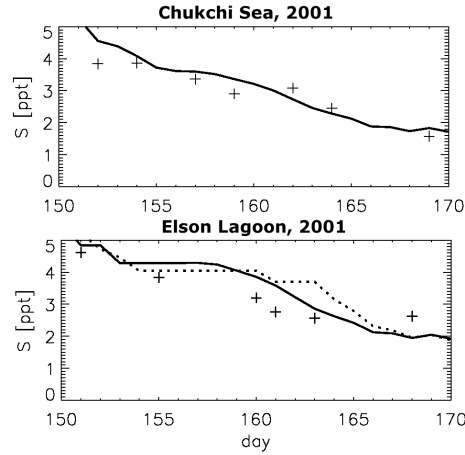


Figure 4.9: Observed (crosses) and simulated (lines) surface bulk salinities S_{up} (i.e., restricted to the upper 50 cm of ice) in summer 2001 at CS (upper panel, simulation CS2001-1) and EL (lower panel). On the lower plot, the results of simulations EL2001-1 (solid line) and EL2001-6 (dashed line) are shown.

Figure 4.10 and 4.11 display the simulated and observed temporal evolutions of the salinity, temperature and brine volume vertical profiles in summer at CS and EL. The overall agreement of the modeled salinity profile with observations is very good. At the CS site, the individual modeled and observed salinities agree within the range of errors, except on day 153. The initial surface salinity decrease is not strong enough in the model. At the EL site, the first modeled profile, typical of the end of winter situation, significantly differs from observations (error of interior salinity is up to 2 ‰). During the melt period, the shape of the salinity profile is well captured, but, for days 156, 161, 162 and 164, the individual ice salinities are overestimated by up to 1.2 ‰. On day 169, the observed and simulated salinity profiles agree within the range of observations. While the flushing parameterization seems to give good results, the maximal error occurs just after the onset of flushing, which points towards the importance of the early desalination.

The model-data agreement for temperatures corroborates the salinity pattern. At CS, the ice temperature profile is slightly too cold but agrees well with observa-

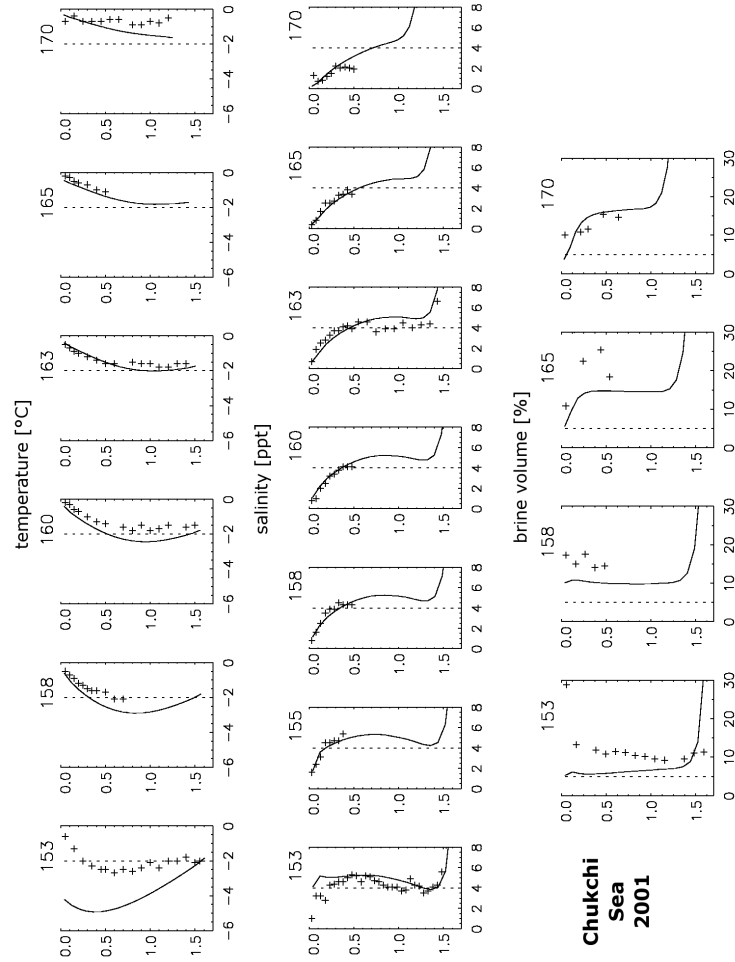


Figure 4.10: Observed (stars) and simulated (lines) vertical temperature [°C], salinity [‰] and brine volume [%] profiles at CS during the early melt period of 2001 (simulation CS2001-1). Vertical axis refer to depth in the ice [m]. The numbers at the top of the individual plots indicate the number of the day in the year. Brine volume is computed from equation (4.4).

tions (errors less than 0.7°C). Again, the maximal error occurs on day 153, with significant, widespread underestimation of the temperature by the model (by up to 3°C). Similarly, at the EL site, the ice temperatures are underestimated by the model at the beginning of the melt period, and then agree with observations within 0.5°C .

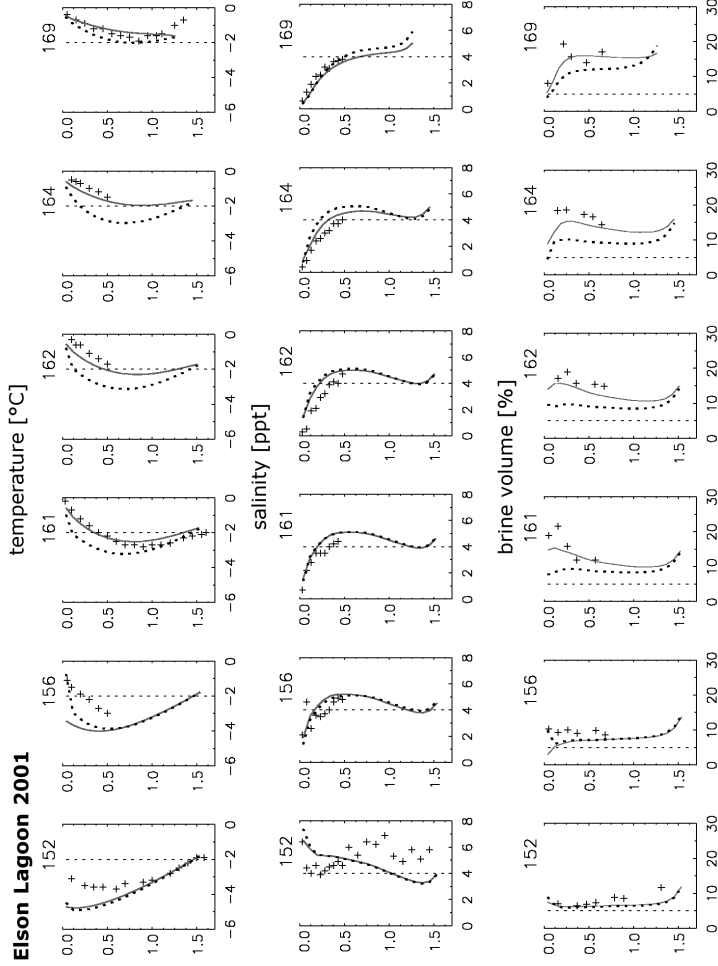


Figure 4.11: Same as figure 4.10, but at EL. EL2001-1 (gray solid curves) and EL2001-6 simulations (black dashed curves) are shown

The agreement between observed and model brine volume vertical profiles is not as good as what we had for salinity and temperature. Differentiating (4), we get :

$$\Delta e = -\frac{\mu}{T} \Delta S + \mu \frac{S}{T^2} \Delta T, \quad (4.11)$$

where T is a temperature in $^{\circ}\text{C}$. Equation (4.11) indicates that the error in brine volume is dominated by temperatures close to the seawater freezing point. The overall underestimation in brine volume at both sites indicate that ΔT dominates the error in brine volume at both sites. Nevertheless, the model captures well the time when the ice first reaches the permeability threshold ($e > e_T$).

In summary, when the snow cover is thin and disappears quickly (CS2001), the model manages to simulate the ice mass balance, salinity and temperature profiles. If the snow cover is thick and melts slowly, then the model fails to reproduce the snow and ice mass balance but still does a reasonable job in simulating the salinity and temperature profiles. The upper ice layers temperatures are consistently underestimated at the onset of the melt period.

4.4.4 Sensitivity of summer halo-thermodynamics to the representation of the snow in the model.

In this section, we study the sensitivity of the simulation to snow physics at EL in summer 2001. We investigate whether it is possible to improve the model summer snow mass balance while maintaining or improving the desalination process. We perform five sensitivity experiments (EL2001-2,3,4,5,6) presented in Table 4.4.

EL2001-2: We estimated the snowfall rate P from the derivative of the snow depth above, which omits P events when snow is melting. However, when an extra 2 cm/day of snowfall is added to days 155-165 for the EL site, the results change very little. The absolute value of the average error in snow depth is only reduced by 1.3 cm (from 7.6 cm). Therefore the omission of P during the melt period is unlikely to be responsible for the model-data discrepancy.

EL2001-3: Several studies [e.g., Sturm *et al.* (2002); Huwald *et al.* (2005a)] indicate that the deepest snow layers tend to be denser and therefore more conductive, due to snow compaction and refreezing. These layers persist in summer after a few days of melt. In EL2001-3, we use $\rho_s = 400 \text{ kg/m}^3$ and $k_s = 0.40 \text{ Wm}^{-1}\text{K}^{-1}$ to take this effect into account. In order to maintain the same bottom ice growth as in the control run, we only modify the snow thermal conductivity and density when snow melts.

EL2001-4: Snow transmits radiation, as suggested by Grenfell and Maykut (1977). In EL2001-4, we solve Beer's law in the snow. As they suggest, we use $i_o = 0.15$ and use an extinction coefficient $\kappa_s = 15 \text{ m}^{-1}$ in the snow.

EL2001-5: Meltwater percolates and refreezes, leading to superimposed ice formation. Superimposed ice formation is much more likely when the snow layer is thick [Hajo Eicken, personal communication]. Superimposed ice was present at the EL site, at the snow-ice interface from June 5th to 11th (day 156-161), but it was not observed at CS at the same period. In our model, when the snow

Table 4.4: Differences between simulated and observed ice thickness (h_i), snow depth (h_s), bulk salinity and average temperature of the uppermost 50 cm of ice (S_{up} , T_{up}) at CS and EL, summer 2001. The comparison is done for the period between May 31 and June 19 (days 151-170). The date of onset of ice surface melt is also indicated. The experiments EL2001-2 to EL2001-6 are sensitivity experiments to the representation of snow physics in the model (further described in section 4.4.4.). The average error between model and observations (E), the standard deviation of error during the melt period and the correlation coefficient (c.c) are shown.

Run	Description	h_i (cm)		h_s (cm)		S_{up} (‰)		T_{up} (°C)		ice surf. melt onset
		E	c.c.	E	c.c.	E	c.c.	E	c.c.	
CS2001-1	Control run	0.2 ± 0.4	0.98	0.2 ± 0.5	0.99	0.01 ± 0.40	0.91	-0.53 ± 0.67	0.94	4-Jun
EL2001-1	Control run	-9.4 ± 11	0.83	-7.6 ± 8.0	0.62	0.57 ± 0.59	0.87	-0.77 ± 0.70	0.93	7-Jun
EL2001-2	Incr. snowfall (1)	-5.8 ± 7.4	0.86	-6.3 ± 6.5	0.75	0.69 ± 0.73	0.78	-1.33 ± 0.66	0.86	12-Jun
EL2001-3	Incr. k_s and ρ_s (2)	-7.1 ± 9.3	0.85	-6.2 ± 6.8	0.74	0.34 ± 0.80	0.85	-1.11 ± 0.64	0.89	11-Jun
EL2001-4	SW pen. in snow (3)	-4.1 ± 7.2	0.86	-4.5 ± 5.7	0.83	0.85 ± 0.86	0.74	-1.63 ± 0.75	0.81	13-Jun
EL2001-5	Meltwat. refreez. (4)	-8.0 ± 10	0.84	-6.8 ± 7.0	0.71	0.04 ± 0.73	0.91	-0.72 ± 0.43	0.96	10-Jun
EL2001-6	(2) + (3) + (4)	-2.7 ± 5.8	0.87	-1.7 ± 3.6	0.94	0.68 ± 0.68	0.80	-1.15 ± 0.43	0.93	14-Jun

melts, the resulting melt water is immediately sent to the ocean, instead of being trapped in the snow or at the ice surface. We thus lose melt water, which might otherwise refreeze in the snow or at the ice surface and form superimposed ice. Consequently, it may increase the snow density and the required energy to melt the snow. In addition refreezing melt water increases heat transport to the ice. In the EL2001-5 simulation, when snow starts to melt, we assume half of the melt water refreezes at the ice surface, as suggested by superimposed ice fractions derived from $\delta^{18}\text{O}$ values in the uppermost ice layers by Eicken *et al.* (2004) and direct observations of Nicolaus *et al.* (2003). Effectively the rate of snow loss is reduced and instead the uppermost ice layers are warmed by an equivalent amount of heat. This amount of energy per unit of area Q_{rf} , representing the thermal role of refreezing on the snow-ice interface heat balance, is given by:

$$Q_{rf} = 0.5\rho_s L_0 \Delta h_s^{<0}. \quad (4.12)$$

ρ_s ($=330 \text{ kg m}^{-3}$) is the snow density and L_0 ($=334 \text{ kJ kg}^{-1}$) is the massive latent heat of pure ice. $\Delta h_s^{<0}$ is the daily decrease in snow melt.

In the EL2001-6 simulation, we combine the three additional snow physical features used in the simulations EL2001-3, 4, 5. Taken separately, the extra snow features reduce snow melt, maintain the snow cover during a longer period, reducing errors. Improving the snow depth simulation consistently results in better ice thickness, since ice does not melt instead of snow. Including the penetration of radiation through the snow (EL2001-3) induces the most remarkable reduction in the error in snow depth (3.1 cm). The effects of combined extra snow features add up (EL2001-6) and lead to an overall reduction in snow depth error by 5.9 cm, leading to a much smaller average error (1.7 cm). In Figure 4.7, the temporal evolution of snow depth in EL2001-6 can be compared to EL2001-1.

The impact of the extra snow features on temperature and salinity profiles is not always positive. Compared to the control run, in EL2001-3 and EL2001-4, the model temperature profile is deteriorated. The heat supply to the upper ice layers increases, due to larger heat conduction (transmitted radiation) in EL2001-3 (EL2001-4). In these two simulations though, the increased heat supply is smaller than the reduced supply of sensible heat due to the deeper insulating snow cover. This implies colder temperature in the uppermost ice layers. Increasing the snow thermal conductivity and density (EL2001-3) does not lead to any remarkable change in the simulation of the ice desalination. Adding the penetration of radiation through the snow (EL2001-4) significantly increases the error in ice salinity. The significantly deeper snow cover in EL2001-4 induces much colder temperatures, lower brine volume and fewer desalination.

Table 4.5: Description of the sensitivity experiments to internal parameters i_0 , ϕ , and e_T .

Run	i_0	ϕ	e_T
CS2001-1	0.30	0.30	5%
CS2001-2	0.00	0.30	5%
CS2001-3	0.50	0.30	5%
CS2001-4	0.30	0.20	5%
CS2001-5	0.30	0.50	5%
CS2001-6	0.30	0.30	8%
CS2001-7	0.30	0.30	10%

In EL2001-5, superimposed ice formation slightly warms the temperature of the upper ice layers. Though the ice is still too cold in EL2001-5, the simulation of the salinity profile is much better than in the control run. The warmer uppermost ice layer is more permeable and therefore more prone to desalination.

The combined snow features (EL2001-6) only slightly deteriorate the temperature and salinity profile, as can be seen in Figure 4.11. The overall quality of the simulation of the salinity and temperature profile is conserved. While penetration of radiation seems to play a key role in maintaining deeper snow, the thermal effect of melt water refreezing is of primary importance for temperature and salinity. It provides an efficient means to transfer heat to the uppermost ice regions and to maintain warm and permeable ice at the surface. Other source of errors, like the uncertainty in SW radiation due to errors in cloud cover, may also affect the upper ice layers temperatures, desalination and melt.

4.4.5 Sensitivity of summer desalination to melt water flushing parameters (i_0 , ϕ , e_T).

In order to get a better physical understanding of the summer sea ice desalination, we assess in this section the model sensitivity to the parameters i_0 , ϕ and e_T . Details about the experiments (run in the CS2001 conditions) are shown in Table 4.5. Results are shown in Figure 4.12.

According to equation (4.10), the desalination rate by flushing depends on Q (the desalinating flow), on $e(z)$ (the vertical profile of brine volume) and on $d\sigma/dz$ (the vertical profile of brine salinity gradient). Depending on the minimum brine

volume e_{min} , flushing occurs or not. Q (in m^3/s) is the same throughout all ice layers and is proportional to ϕ and the surface melt rate. For a given Q , ice layers with larger brine volume desalinate more slowly. $d\sigma/dz$ is conditioned by the temperature profile, according to equation (4.2). Depending on the shape of the temperature profile, $d\sigma/dz$ can be positive or negative, which in turn modulates the sign of dS/dt .

i_0 is the fraction of solar radiation penetrating inside the ice. Solar radiation penetrating inside the ice warms the inner ice layers, instead of contributing to the surface energy budget and surface melt. A higher i_0 therefore induces less surface melt, smaller Q but warmer upper ice layers.

In the CS2001-2 simulation ($i_0=0$), the surface melt and desalination are initially very intense (see Figure 4.12, a1). After 5 days of melt, the uppermost ice layers have a relatively low salinity and are relatively cold, such that $e < e_T$ (see Figure 4.12, a2). In other words, an impervious cold surface layer has formed. The desalination is then discontinuous, alternating between the melt and the formation of a new surface impervious layer. After 20 days into the melt period, the salinity is very low at the surface, but remains high in the bottom layers. First, the shape of the brine volume profile (see Figure 4.12, a5) with low brine volumes at the surface and high brine volumes at the ice bottom implies that the desalination effect is the strongest close to the ice surface. Second, the absence of penetrating radiation allows the C-shaped temperature profile to persist longer (see Figure 4.12, a3). The brine salinity gradient therefore remains negative (positive) in the upper (lower) ice layers (see Figure 4.12, a4). Therefore, the brine salinity always decreases (increases) in the uppermost (bottommost) ice layers. Note that this kind of impervious layer also forms in the EL2001-1 simulation but only during days 154-158 and might be an incentive to the accumulation of melt water at the surface and the formation of superimposed ice.

In the CS2001-3 simulation ($i_0=0.5$) simulation, surface melt and Q are smaller. Therefore, the desalination is less intense and lasts longer. Higher salinity and brine volume can be found close to the surface. There is a period of 10 days during which the temperature profile is C-shaped. Therefore, the brine salinity gradient is negative (positive) in the uppermost (bottommost) layers. This period has a similar length in CS2001-1 and CS2001-3. Since Q is smaller in CS2001-3, in the uppermost (bottommost) layers the brine salinity will be higher (lower) than in CS2001-1.

ϕ , the melt water partitioning coefficient has a direct impact on the desalinating flow Q . A smaller ϕ reduces Q . Therefore, the simulation CS2001-4 ($\phi=0.20$)

Table 4.6: Simulated total mass of surface meltwater percolating through the ice (total flow and specific contributions from snow and ice) in early summer 2001 (June 1-20) and daily average percolating flux. The percentages represent the ratio of total (respectively snow, ice) percolating meltwater flow to the total (respectively snow/ice) meltwater mass formed at the surface. Simulations CS2001-1 and EL2001-4 are used.

	Snow/Ice (kg)	Snow (kg)	Ice (kg)	Av. Perc. Flux
<i>Chukchi Sea</i>	116.9	10.8	106.1	5.56 kg m ⁻² d ⁻¹
	28.8%	27%	29.1 %	0.56 cm d ⁻¹
<i>Elson Lagoon</i>	76.4	17.8	58.0	3.63 kg m ⁻² d ⁻¹
	24.0%	17.8%	27.6%	0.36 cm d ⁻¹

has qualitatively the features of the CS2001-3 simulation described in the previous paragraph. A larger ϕ induces a larger Q and stronger and quicker desalination. This results in smaller (larger) ice salinity in the uppermost (bottommost) ice layers. Compared to the CS2001-2 ($i_0 = 0$) simulation described earlier, owing to the penetration of solar radiation, the inner ice layers remain warm enough and no surface impervious layer is formed.

e_T , the permeability threshold affects mainly the time of the onset of flushing (see Figure 4.12 c). With $e_T \leq 5\%$, flushing starts when surface melt starts. If $e_T = 8\%(10\%)$, the flushing starts 5 (10) days after surface melt.

4.5 Summer melt water budget

According to our model, during the first 21 days of the melt period (days 151-171), surface melt water percolates through bare ice at the rate of 0.36–0.56 cm/day or 76–117 l/m² (see Table 4.6). In other words, on bare ice, on average 26.4% of the surface melt water (produced on bare ice) must flush through the ice to simulate the observed salinity profiles during the first 21 days of the melt period. The remainder of the melt water is either discharged into the mixed layer through cracks and leads, percolates through ponded ice, or accumulates in surface and underwater melt ponds by lateral drainage. By comparison, from observations, Eicken *et al.* (2002) deduce that over the summer 12 % of the total meteoric water is retained by bare ice. This tends to corroborate that much of the vertical percolation occurs during the early stage of melt.

4.6 Discussion and conclusions

4.6.1 Observations of early summer desalination

We examined two years (2000 and 2001) of early summer sequences of salinity profile measured from landfast ice cores at two sites in Point Barrow, AK. We see a quick desalination, starting at the onset of melt. This desalination for bare ice causes a transition between the C-shaped FY ice salinity profile and the typical MY salinity profile, characterized by the upper 50 cm of almost fresh ice. After 12 days (averaged over the CS2000, CS2001 and EL2001 desalination sequences), the bulk salinity of the upper 50 cm of ice drops under 3 ‰ and the typical MY ice profile forms, and then remains nearly unchanged through the end of the investigated period (first 21 days of melt). A longer desalination sequence of melting FY ice described by Eicken *et al.* (2002) shows a further desalination, leading to practically fresh ice after 20 more days of melt. This further desalination is prompted by convective overturning of freshwater coming from the ice bottom into the brine network.

In ponded ice, the salinity profile tends to a vertically constant profile with $S=2$ ‰. This points to a different halo-thermodynamic regime, which is beyond the scope of the present study.

The observed desalination signal is significantly larger than observation errors. In addition to systematic extraction errors (about 10-20% of the initial salinity in summer), the fact that subsequent ice cores were taken from different places was certainly not ideal. During early summer, horizontal variability in individual ice salinity (0.34 ‰) was found to be 2-3 times smaller than in winter. We attribute it to the fact that the ice desalination occurs everywhere and therefore homogenizes the profiles. We suspect summer horizontal variability in ice salinity is larger than what was directly observed. In particular, snow cover variability can induce variability in superimposed ice formation which affects the amount of percolating water. In addition, other kinds of flooding events may scramble the pattern. New instruments able to measure the sea ice salinity, as for example, based on impedance measurements (see, e.g., Notz *et al.* (2005)) should be very useful at monitoring the horizontal variability of sea ice salinity at longer time scales.

4.6.2 Modeling the winter sea ice salinity profile

We formulated a prognostic salinity model for Arctic undeformed bare ice, based on the model of CW88 regarding the winter desalination. The simulation of winter salinity profiles reveals that the original CW88 model has some deficiencies, as already mentioned by Cox (1990). As it is proposed, CW88 leads to correct desalination in the first two weeks of ice growth. After that, the gravity drainage is too intense and leads to excessively small salinity close to the ice-ocean interface. The δ factor, describing the impact of the unstable temperature gradient on the ice desalination, had to be reduced by 65 % to give better bottom salinity. The fact that the CW88 parameterization of gravity drainage is valid only for early stages of ice growth is not surprising. The data on which this parameterization was based (Cox and Weeks, 1975) were taken from laboratory ice, only up to 20 days old, and probably are not representative of the overall winter period. The corrected δ value suggested by our study is more representative of the average winter behavior, but not of the initial desalination, which is too weak.

In CW88, gravity drainage is empirical. A feature absent in our simulations is the double-C-shape that we see in the observations (see Figure 4.3a). Notz (2005) gives an explanation of this feature, based on the concept of a permeability-associated mush-Rayleigh number that cycles through sub- and supercritical regimes.

The present data show that, in the ice interior, the salinity is practically constant with a value of about 5 ‰, which suggests that the winter desalination mechanisms stop when this value is reached. Cox (1990) suggested to add seawater suction (the reverse process of brine expulsion) to the CW88 model. When the ice is warmed, brine would be sucked up from the underlying seawater, which has never been observed. Furthermore, this mechanism is not well motivated and not supported by observations. Still, the question of the mechanism beyond gravity drainage (brine convection, diffusion, capillarity effects in brine tubes) leading to a stable C-shaped profile is not elucidated. Further field and model studies are required to answer this question. Mercier *et al.* (2005) show that diffusion might dominate in certain years in the Southern Ocean. A more physical modeling approach, based on mushy-layer theory, should also help to answer this question (see, e.g., Oertling and Watts (2004) and Notz (2005)).

Our results suggests that the CW88 parameterization – which is based on laboratory-controlled ice growth over around two weeks – is valid only for the earliest stages of desalination, when convection is active (see, e.g., Notz (2005)). After two weeks, the gravity drainage coefficient has to be reduced to maintain model-observation agreement. It is arguable that changes in convection (or even an

interruption of it) weaken the desalination and explain the necessary reduction of the original Cox and Weeks (1988) gravity drainage coefficient.

4.6.3 Modeling the summer Arctic bare ice mass balance and salinity profile

The external forcing (LW radiation, and oceanic heat flux) was tuned to provide a reliable simulation of the sea ice salinity profile. In summer, when the snow cover is thin (at the CS site), the model and observed ice thickness and snow depth are in close agreement (average error smaller than 0.5 cm). When snow is deep (at the EL site), the snow melts 10 days too early and the ice is on average 9.4 cm too thin. In both cases, the salinity profile is well captured by the model, but the upper ice interior is slightly too cold during the first 10 days of the melt period. Even though the model captures the brine volume vertical profile relatively well, the relative errors in brine volume are higher than in salinity and temperature because the error in brine volume, depending on $1/T^2$, are maximal close to the freezing point. Direct observations of brine volume would help to further investigate this problem.

Regarding summer desalination, our model includes an original parameterization of summer flushing. Flushing is defined as the vertical component of the translation movement of the brine salinity structure in the brine network, triggered by an interconnected brine network and the availability of freshwater at the surface. Because the hydraulic flow depends mostly on ice permeability rather than melt water production, we suggest that flushing can be considered instantaneous compared to one-day time step.

The simulated summer bare ice desalination is in agreement with observations, when the fraction ϕ of available melt water percolating into the brine network was set to 30%. This tuning parameter partitions the available melt water between lateral and vertical percolation. Its value is probably dependent on the time step. We imagine, for example, that if the diurnal cycle is resolved, the ice could be permeable at noon and impervious at night. The value used here represents the daily average amount of melt water used in vertical desalination.

A significant part of errors in the model mass balance, temperature and salinity at EL is also due to the errors in the forcing (in particular the ocean heat flux). On the other hand, the rough representation of the snow cover in the model is certainly of primary importance. The sensitivity experiments in Section 4.4.4

suggest that a better representation of snow physics and of the fate of surface melt water in the model would improve not only the simulation of the snow cover but also the simulation of ice mass, heat and salt balance. Penetration of radiation through the snow, snow compaction and subsequent increase in snow thermal conductivity, as well as superimposed ice formation, slow the melt of snow and increase the heat transfer to the ice interior. Penetrating radiation has the largest impact on snow depth. Refreezing of surface melt water is important to maintain the uppermost ice warm under a thick, insulating snow cover. The simulations show at EL that a sufficient amount of heat transferred through the snow is compulsory to open the brine network in the upper ice layers.

4.6.4 Flushing and superimposed ice formation

In our simulations the first flushing event occurs at the same time as the onset of surface melt. At the CS site, the brine network is always open and flushing is continuous. At the EL site, helped by the initial decrease in S and insulated from the surface by a thick snow cover, the upper brine network closes for a few days, leaving no way to melt water and favoring superimposed ice formation. These differences suggest a series of desalination regimes that depend on snow depth as follows:

(1) A very thick snow cover is present and there is no surface melt. No percolation of melt water occurs, the profile remains C-shaped. Such profiles were among second-year ice profiles observed in the Weddell Sea by Eicken (1998).

(2) A thick snow cover is present, melting at the surface. The upper ice layers are efficiently insulated and thus cold enough at the snow melt onset so that they are impervious to the incoming melt water percolating through the snow. In contact to the impervious, relatively cold surface, the percolating melt water has the potential to form a fresh layer of superimposed ice. This leads to a profile with a "7"- shape (i.e. fresh at the surface and with more or less constant salinity inside the ice). Profiles from the Weddell Sea taken in summer by Haas *et al.* (2001) correspond to this description. This is a reason why more superimposed ice is found under thick snow layers (see, e.g., Eicken *et al.* (2004)). The thick superimposed ice layers observed in the Baltic Sea (see, e.g., Granskog *et al.* (2006)) may also be explained because the ice is almost fresh there and therefore impervious to melt water percolation.

(3) A thick snow cover is present, melting at the surface, but the upper sea ice layers are permeable at the melt onset. Initially, melt water percolates and

creates a profile with a more or less linear shape at the surface. Since the upper ice layers are well insulated from the surface by the thick snow cover, they are relatively cold and the desalination may render the upper ice layers impervious. Superimposed ice formation may then start for a few days, until the ice layers are warm enough again to be permeable. This is the situation found at CS2000 and EL2001.

(4) A thin snow cover is present, melting at the surface. The upper sea ice layers are permeable at the melt onset. Flushing is continuous in these situations. Meltwater percolates but does not form superimposed ice. This is the situation found at CS2001.

4.6.5 Meltwater balance

The modeled desalination is compatible with melt water storage in the ice data of Eicken *et al.* (2002). Our model suggests that 24.5% of the basin-scale melt water produced at the surface is stored in bare ice in the early stages of summer (20 days), while Eicken *et al.* found a value for the whole summer of 12%. The discrepancy between the two values might signify that the vertical seepage of melt water in the ice decreases, favouring lateral transport, after the early stage of melt because bare ice becomes much more permeable as it desalinates. Another explanation is that some of the bare ice melts and releases its melt water content to the mixed layer. Thus, the freshwater flow coming from the bare ice surface melt is excessive to desalinate the ice.

4.7 Perspectives

The present study underlines that further field campaigns with a large focus, joining several aspects of the ice cover physics (e.g., snow and ice mass balance, freshwater pathways, ice temperature, salinity and brine volume, ...) are a powerful tool to constrain thermodynamic sea ice models.

In a parallel study (Vancoppenolle *et al.*, 2006), we show that the halo-thermodynamic model presented here is able to simulate MY ice salinity profile and reaches a quasi-equilibrium after 10 years of simulation (i.e., the present model does not drift). Therefore, it should be able to simulate the features of the Arctic bare ice

salinity profile in climate models, where particular care should be taken to ensure heat and salt conservation.

Nevertheless, the present model is only a step towards a better representation of the thermodynamics of the polar snow-ice cover. Future model development should focus on the physics of the snow cover, the melt water pathways and storage and the gravity drainage of brine. Another extension of the model could be to include a meltpond scheme, such as from Taylor and Feltham (2004).

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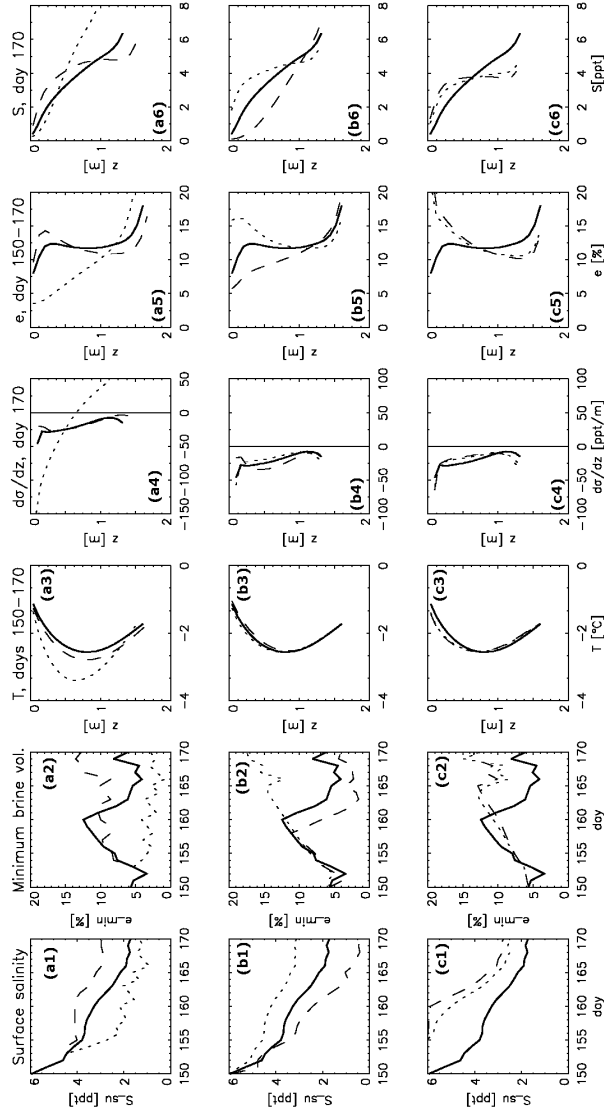


Figure 4.12: Sensitivity experiments (a) to i_0 : CS2001-2 (dotted line); CS2001-1 (solid line); CS2001-3 (dashed line). (b) Sensitivity to ϕ CS2001-4 (dotted line); CS2001-1 (solid line); CS2001-5 (dashed line). (c) Sensitivity to e_T : CS2001-1 (solid line); CS2001-6 (dotted line); CS2001-7 (dashed line). For each sensitivity experiment, the temporal evolution of surface salinity (i.e., bulk salinity of the uppermost 50 cm of ice) and of minimum brine volume is shown. The vertical profiles of temperature (averaged over days 150-170), of $\partial\sigma/\partial z$ (day 170), of brine volume (averaged over days 150-170) and of salinity (day 170) are also plotted.

Chapter 5

Modeling the salinity profile of undeformed Arctic sea ice (VFB06)

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Abstract.

The salinity of sea ice affects its physical and ecological properties. Here, a multilayer one-dimensional halo-thermodynamic sea ice model is used to simulate the vertical salinity profile of undeformed Arctic sea ice. The model successfully reproduces the desalination pattern observed in first-year (FY) and multi-year (MY) ice. The model can also be integrated with a prescribed, time-independent salinity profile. Substantial differences in the simulated mass balance and ice-ocean salt flux arise depending on the salinity. After 10 years into the simulation, the annual mean ice thickness is 2.85 m with the interactive halodynamic component, compared to 2.53 m (2.29 m) with a prescribed, time-independent, vertically varying (constant) salinity profile. Modelling sea ice salinity is especially important when sea ice is transitioning from a MY to FY ice regime. Thus including a halodynamic component in sea ice models would significantly improve simulations of future climate.

5.1 Introduction

Sea ice is a major component of the climate system and a preferential site for the development of microorganisms and algae (Ackley and Sullivan, 1994). Consequently, the changes in the physical state of the sea ice cover projected for the next century (e.g., Arzel *et al.* (2006)) are likely to have serious repercussions on polar environments and ecosystems.

Sea ice traps some salt as it forms. This salt is contained in brine pockets, while the ice itself is nearly salt-free. The salinity of sea ice S for the combined ice and brine pockets is usually much less than seawater, although the brine pocket salinity alone can be much higher.

The salinity of sea ice plays a significant role in its mass balance and the salt / freshwater exchange between ice and ocean (Vancoppenolle *et al.*, 2005). The relative volume of salty liquid (brine) inclusions (e) in sea ice strongly depends on ice salinity and temperature. In turn, the sea ice thermal properties depend on e . Above an $e = 5\%$ threshold, the brine network becomes interconnected, rendering the sea ice permeable to fluid transport (Golden *et al.*, 1998). This structural change directly influences the nutrient supply for microorganisms and algae living in sea ice (Ackley and Sullivan, 1994) and the atmosphere-ocean exchange of dissolved gases (i.e., CO_2) through sea ice (Delille, 2006).

Brine drainage causes the sea ice salinity to vary in time. First-year (FY) ice is initially quite saline (from 6 to 15‰), but it desalinates quickly and acquires a C-shape vertical salinity profile (Nakawo and Sinha, 1981). Multi-year (MY) ice is much less saline and is characterized by a fresh surface layer (Schwarzacher, 1959). Despite these striking differences in salinity over time, only prescribed, steady and simplified sea ice salinity profiles have been included into the standard sea ice models used in climate studies (e.g., Fichefet and Maqueda (1997); Bitz and Lipscomb (1999)).

In the present work, we use the newly developed Vancoppenolle *et al.* (2007) (hereafter referred to as VBF07) halo-thermodynamic sea ice model, forced by climatological forcing data, to study the temporal evolution of the salinity and brine volume profiles of undeformed Arctic FY and MY ice. In addition, we discuss results from sensitivity experiments to investigate the main features of the modeled desalination. We also examine how the ice thickness is affected by the representation of the salinity profile in the model. Finally, we assess the contributions of ice growth and brine drainage to the ice-ocean salt flux.

5.2 The model

The model is described in details in VBF07. It is a multilayer one-dimensional halo-thermodynamic sea ice model. Its prognostic variables are: ice thickness, snow thickness and both ice temperature and salinity vertical profiles. Brine volume is diagnosed from temperature and salinity. Warm temperatures and/or high salinities result in high brine volume. The temporal evolution of salinity and temperature are coupled in the following way. The evolution of the salinity profile depends upon the thermal properties of sea ice (i.e., specific heat, thermal conductivity and energy of melting, defined as the energy required to melt a unit volume of sea ice). These thermal properties control the heat transport and storage (thus the temperature profile) as well as the growth/melt rates at the ice interfaces (thus the ice thickness).

The model ice salinity profile $S(z, t)$ varies due to 1) brine entrapment during ice growth (larger growth rates yield more salt entrapment), 2) brine expulsion (depending on the temporal derivative of ice temperature), 3) gravity drainage (assumed to depend on the vertical temperature gradient, and active if the brine network is open), and 4) flushing. Brine entrapment, expulsion and gravity drainage are modeled following *Cox and Weeks* [1988] with small differences described in VBF07. The flushing parameterization is newly developed by VBF07, so we describe it briefly here. If the brine pocket network is interconnected, corresponding to a brine volume $e \geq e_T = 5\%$, where e_T is the permeability threshold mentioned by Golden *et al.* (1998), and if (2) the surface temperature $T_{su} = 0^\circ\text{C}$, we suppose that a fraction ϕ of the available meltwater (inferred from the surface melt rate) instantaneously flows through the brine channels, replacing salty brine by surface meltwater. The meltwater salinity is given by the salinity of the snow/ice from which it formed. The snow is assumed to be of zero salinity. ϕ ($=0.30$ if flushing occurs and zero otherwise) was calibrated in VBF07. The remainder of the meltwater is assumed to contribute to lateral drainage to ocean through cracks and leads, or to collect in the lowest gravity areas and form melt ponds. The model was shown to be sensitive to ϕ and e_T values.

The model is forced by idealized climatological atmospheric and oceanic conditions, which are widely used in the literature (e.g., Maykut and Untersteiner (1971); Bitz and Lipscomb (1999)). The monthly values of upper surface forcing data (downwelling radiative and turbulent heat fluxes) are typical of the central Arctic. The ocean-ice heat flux is set to 2 Wm^{-2} and the seasonal snow accumulation allows the snow to reach a depth of 37 cm at the end of the ice growth season. The model is initialized with 10 cm-thick sea ice and no snow. The ice is initially isosaline with $S=12 \text{ ‰}$ and isothermal with $T=270 \text{ K}$. The initial

Table 5.1: Description of the simulations.

Experiment	Description
SZT	prognostic halodynamic component used
PR	steady-state, vertically-varying salinity profile
BK	steady-state, isosaline salinity profile

salinity is obtained from empirical thickness–bulk salinity relationships (Kovacs, 1996). The key parameters of the thermodynamic component are as in Bitz and Lipscomb (1999). The only exception is i_0 , the coefficient partitioning the solar radiation into surface versus internal warming equals zero if snow is present and 0.17 otherwise. 10 layers in the ice and 1 layer in the snow are used.

In addition to a simulation with the prognostic halodynamic component (referred to as SZT), we also run simulations with prescribed, time-independent salinity profile. We either let the ice be isosaline with $S=4.6$ ‰ (case BK), a value representative of typical mid-minter FY ice, or we prescribe a vertically varying salinity profile (case PR, see Figure 5.1e). The latter is taken from observations of MY ice in central Arctic (Schwarzacher, 1959). The model simulations are summarized in Table 5.1.

The model is run for 50 years. The annual net ice growth is less than 1 cm after 30 years for SZT, after 48 years for BK and after 27 years for PR. Annual bulk salinity variations are less than 0.01 ‰ after 10 years. At this time, the ice salinity reaches a quasi-equilibrium, but only stabilizes when the ice thickness is in equilibrium. Since the age of Arctic sea ice rarely exceeds 10 years (e.g., Belchansky *et al.* (2005b)), we analyze the results during the first 10 years, when the model is not in equilibrium.

VBF07 verified the model with a 2-year integration by forcing with daily-varying meteorological data and comparing with sea ice measurements taken from nearby landfast cores. In this study we investigate the longer-term response instead, so we force the model by repeating a climatological mean annual cycle and compare with basin-average estimates of sea ice salinity. Though neglecting ice motion and variable forcing, this idealized study offers insight into the long-term thermal and haline coupling in isolation. Future studies will investigate the influence of sea ice dynamics and synoptic variability on the coupling.

5.3 Results

5.3.1 FY ice salinity variations

The monthly desalination for the entire first year in the model is shown in Figure 5.1(a-d). Figure 5.2a illustrates the tendency terms of the salinity equation. The desalination of FY ice occurs in 3 phases. First, a rapid (15 days) and intense (5 ‰) desalination occurs during the initial rapid ice growth. This *initial desalination* (see Figure 5.1a, January profile) is due to gravity drainage (99 % of the total desalination) throughout all the ice layers. The brine network is permeable and filled with brine denser than the seawater underneath, an inherently unstable situation. This phase lasts until an inner layer becomes cold and fresh enough so that its brine volume falls under the 5 % permeability threshold. The bulk salinity decreases by 4.62 ‰ in this phase. This initial drainage was observed in the field by Nakawo and Sinha (1981) and in lab experiments by Worster and Wettlaufer (1997).

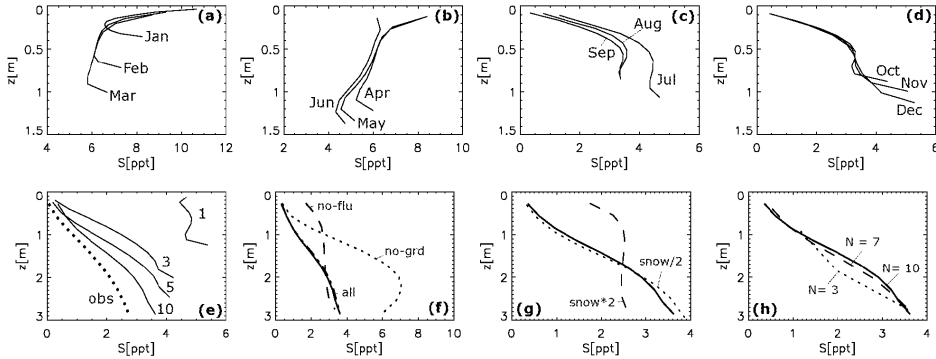


Figure 5.1: (a-d) Salinity profiles for the first year of the simulation, on the 15th of each month. (e) Simulated annual mean salinity profiles for years 1, 3, 5 and 10. The dotted line represents the observed MY ice salinity profile of Schwarzscher (1959). (f-h) Sensitivity of the annual mean salinity profile for the 10th year of the simulation to different model features. (f) Sensitivity to the different brine drainage mechanisms. Gravity drainage and flushing are turned off independently. The effect of turning off brine expulsion is too small to be visible. (g) Sensitivity to snowfall rate. Salinity profile with normal snowfall rate (solid line), half snowfall rate (dotted line) and double snowfall rate (dashed line) are shown. (h) Sensitivity to the number of vertical layers in the ice N . $N=3$ (dotted line), $N=7$ (dashed line) and $N=10$ are plotted.

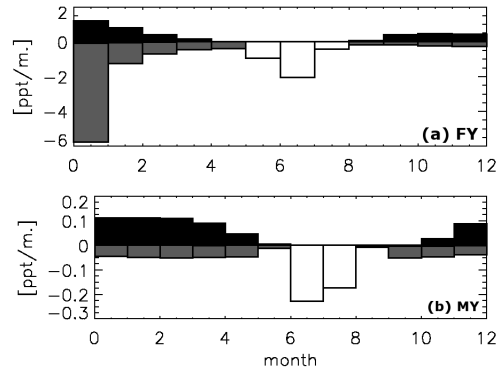


Figure 5.2: Monthly tendency terms of the salinity equation averaged over all vertical layers [$\text{ppt}/\text{m.}$] for FY (a) and 10-year-old MY ice (b). Salt entrapment (black), gravity drainage (grey) and flushing (white) are plotted. Brine expulsion is too small to be visible (max. $0.03 \text{ ‰}/\text{month}$).

In the second phase, the desalination pattern moves to a steady *winter regime* (see Fig 1a-b, February to May profiles). At this stage, the ice slab is split into two parts. The lower third is warm and saline enough so that the brine volume is above the permeability threshold and brine continues to be expelled from the ice by gravity drainage. During this phase, the bulk salinity decreases by 3.91 ‰ . The bottom permeable region can be seen as a macroscopic representation of the brine desalination network, with smaller brine tubes joining each other in larger brine channels, as observed by Bennington (1967).

In contrast to the lower permeable part, the upper part is impervious and therefore not connected anymore to the ocean. Only brine expulsion is active there. Though in this part brine expulsion has a high relative importance, it lowers the bulk salinity by only 0.25 ‰ over all winter months. However, in reality this upper part might be prone to stronger expulsion episodes associated with cold events unresolved by the monthly forcing. Such episodes could lead to the formation of frost flowers (see, e.g., Martin *et al.* (1995)). Entrapment of salt at the ice bottom progressively slows as the ice growth rate decreases. This winter regime lasts until the end of May, resulting in a C-shaped profile as observed by Nakawo and Sinha (1981).

The onset of surface melt and flushing marks the beginning of the third phase, or *summer regime*, in mid-June (day 162) (see Fig 1b-c, June to September profiles), when the whole ice layer becomes permeable. The transition between the FY C-shaped profile to the typical MY ice profile with a fresh surface layer takes around 15 days and continues to evolve slowly until mid-August (day 220),

when the surface melt stops. The surface desalination pattern looks very similar to the one observed in the field by Tucker *et al.* (1987) and was observed in lab experiments by Cottier *et al.* (1999). At freeze onset (near the end of September), the desalination process returns to the *winter regime*, but the salinity remains low at the surface (see Fig 1d).

5.3.2 The MY ice salinity profile and its sensitivity to several model features

After vigorous desalination in the first years, the salinity profile simulated by the model reaches a quasi-equilibrium by 10 years into the simulation, and the amplitude of the annual cycle is less than 0.5 ‰ in any layer. The simulated annual mean MY ice salinity profile is slightly more saline than the observed profile (see Fig 1e). Since errors in salinity from drainage during ice coring may be up to 1 ‰ (Notz *et al.*, 2005), there is little hope of distinguishing error in the model physics from undersampling and observational issues.

Desalination slows over time as there is less salt to lose. In the second year, the total bulk salinity tendency is only 12 % of its value for the first year. This amount decreases to 4 % in year 5 and to 2 % in year 10. Gravity drainage loses its effectiveness more quickly than flushing. In the second year, the total bulk salinity tendency due to flushing (gravity drainage) is equal to 56 % (6 %) of its value in year 1. Nevertheless, in the bottommost ice layers, gravity drainage remains of significant importance even for MY ice.

The relative importance of brine drainage mechanisms in shaping the MY ice profile is investigated by successively eliminating each from the model (see Fig 1f). Expulsion of brine proves again to be of very minor importance, with a contribution smaller than 1 %. Gravity drainage is important in winter in the lower portion of the ice, where layers are permeable. Without it, the ice is more saline in the lower two-thirds of the slab. Summertime flushing freshens the surface and transports salinity downwards.

In order to assess the influence of the snow cover, we successively divide and multiply the snowfall rate by two (see Fig 1g). If snowfall is smaller, then the ice grows faster in winter and the entrapment of salt at the ice base is larger than in the control run. The desalination mechanisms are affected by a negligible amount so the bottom salinity is slightly higher in the half-snowfall simulation. Conversely, if snowfall is increased, the surface layers remain cold in summer

and the permeability threshold is reached rather late in summer. The surface desalination associated with flushing operates only over a limited time (two weeks instead of two months). Thus, the MY surface salinity is greater in the double-snowfall simulation. The number of vertical ice layers can be as low as three and the model still captures the MY ice salinity profile reasonably well (see Fig 1h).

5.3.3 Impact of a prognostic salinity on ice thickness

In this section, we study how the account for salinity variations in the model affects other model variables. Figure 5.3a shows the temporal evolution of ice thickness in the SZT, PR and BK simulations. The salinity-induced thickness variations are dominated by the ice thickness-growth feedback (Vancoppenolle *et al.*, 2005). Because the sea ice thermal properties are salinity dependent, the differences in the ice salinity profile result in small differences in ice thickness. These differences are later amplified through a growth rate strongly dependent on ice thickness.

After 10 years into the simulation, the annual mean ice thickness in PR, BK and SZT is equal to 2.29, 2.53 and 2.85 m, respectively. PR has the thinnest ice throughout the simulation. The amplitude of the ice thickness seasonal cycle is larger in PR than in SZT and BK. The absolute difference in total melt (largely dominated by surface melt) between PR and the two other runs dominate the absolute difference in growth throughout the simulation.

During the first 25 years, the ice thickness in SZT is larger than in BK. After this, the ice thickness in BK becomes larger, through the end of the simulation. The SZT integration traps more brine during growth, so it grows thicker at a faster rate than the BK case. The total (and surface) melt in SZT is less than in the BK simulation, because the fresher ice in SZT has a higher energy of melting close to the surface. But, after 5 years, the reduction in the ice surface specific heat (induced by surface freshening) in SZT dominates, and therefore, the ice in SZT melts faster.

5.3.4 Salt flux to the ocean

Here we evaluate the seasonal cycle of ice-ocean brine exchange. In coupled ice-ocean models, an ice-ocean salt flux is used to ensure salt conservation in the ice-ocean system. No actual water mass is exchanged between ice and ocean

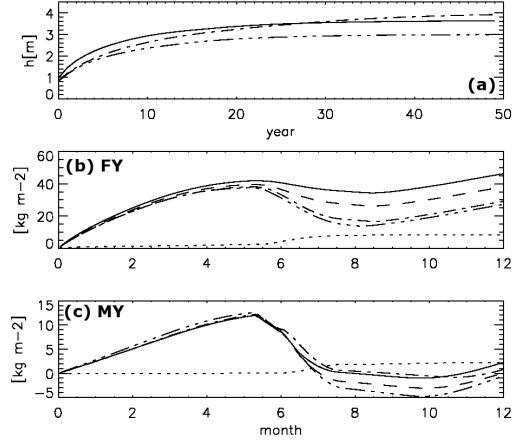


Figure 5.3: (a) 50-year long time series of annual mean ice thickness in the SZT (solid line), BK (dash-dotted line) and PR (dash-3x dotted line) simulations. (b-c) Time-integrated massive salt flux I^s from the ice to the ocean for FY (b) and 10-year-old MY ice (c). Total salt flux (solid line), salt flux from brine drainage (dotted line), equivalent salt flux from freshwater storage in the ice (dashed line) for SZT simulation and total salt flux for BK (dash-dotted line) and PR (dash-3x dotted line) simulations are shown.

when a phase change occurs. Instead fresh water or brine exchange is converted into an equivalent salt flux. In practise, dynamical effects also must be taken into account as sea ice may grow in one place and melt in another. Our 1-d model estimates only takes into account the separation of growth and melt in time not in space. However, the following equations apply more generally.

The ice-ocean salt flux F_{tot}^s (> 0 from the ice to the ocean, in $\text{kg m}^2 \text{s}^{-1}$) can be divided into two parts : $F_{tot}^s = F_{bd}^s + F_{up}^s$. The first component is a direct salt flux from all processes which we can refer to as *brine drainage* :

$$F_{bd}^s = -\rho[h_i \frac{dS}{dt}|_{bd} + (S_w - S_b) \frac{dh_i}{dt}|^{>0} - S \frac{dh_i}{dt}|^{<0}]. \quad (5.1)$$

ρ is the sea ice density, h_i is the ice thickness, S_w is the seawater salinity and S_b is the salinity of new ice. The first term in (5.1) corresponds to all brine drainage mechanisms. The next two account for salt lost by the ocean via brine entrapment during ice growth and for salinity returned to the ocean during ice melt. The second component of the ice-ocean salt flux is from freshwater exchange between the sea ice slab (including snow) and ocean due to melting or growing, which we refer to as *net freshwater uptake*:

$$F_{up}^s = \rho_s S_w (1 - \phi) \frac{\partial h_s}{\partial t}|^{<0} + \rho_i S_w (1 - \phi) \frac{\partial h_i}{\partial t}, \quad (5.2)$$

where ρ_s is the snow density, h_s is the snow depth, and ϕ is the partitioning coefficient for flushing. The first term in (5.2) represents the direct freshwater input from snow melt. The second term corresponds to the ice growth-induced freshwater removal from the ocean (or ice melt-induced freshwater return to the ocean).

We associate these salt fluxes with the *integrated massive salt flux* at day d , $I_{tot}^s(d) = \int_{1Jan}^d F_{tot}^s(t)dt$. Similarly, the two components $I_{bd}^s(d)$ and $I_{up}^s(d)$ are defined as the temporal integrals of F_{bd}^s and F_{up}^s . Their contribution to the net ice-ocean I_{tot}^s are very different for FY ice and MY ice (see Fig 3). For FY ice surviving one summer, the total amount of salt rejected from the ice is around 45 kg m^{-2} over a year, and almost four fifths of this amount is supplied by the net freshwater uptake in the ice. For MY ice, over a year, there is almost no net salt rejection from the ice to the ocean. But integrated over winter only, the freshwater uptake in MY ice rejects almost 10 kg m^{-2} of salt. In contrast, the brine drainage in MY ice has an annual contribution of 2 kg m^{-2} .

In Figures 5.3b-c, the total salt flux to the ocean is shown for the PR and BK simulations. For FY ice surviving one summer, the total salt rejection over a year in SZT is 15 kg m^{-2} higher than in PR and BK. The contribution of I_{up}^s accounts for 68% of the average difference in I_{tot}^s , while the remainder is from brine drainage. The differences essentially build over summer. For MY ice, the differences between the three simulations are only significant in summer, and their magnitude is always smaller than 5 kg m^{-2} .

5.4 Discussion and conclusions

We presented here a simulation of the salinity profile of undeformed Arctic sea ice with a multi-layer halo-thermodynamic sea ice model. The model successfully reproduces the FY ice desalination, which we divided in three steps, referred to as initial, winter and summer stages of desalination. The intensity of brine drainage decreases from year to year and the desalination continues until a quasi equilibrium is reached on year 10. The modeled MY ice salinity profile compares well with observations.

The modeled desalination is sensitive to gravity drainage in the bottom layers and to flushing in the uppermost layers. If snowfall increases by a sufficient amount, the upper ice layers are thermally insulated and cannot reach the permeability threshold of 5 %, preventing flushing to occur, and keeping the upper part of the

ice slab saline. Using 5 vertical ice layers is largely sufficient to simulate the key features, which is the maximum current number used in large-scale sea ice models (e.g., Hunke and Lipscomb (2004)). To incorporate our halo-thermodynamic model in a large-scale sea ice model would require to include the mass of salt for each vertical ice layer in the advection scheme and to add the brine drainage contribution to the ice-ocean salt flux.

Substantial differences in ice thickness occur when different parameterizations of salinity are used. They are driven by differences in salinity at both interfaces, which affect the growth and melt rates. The transient and equilibrium sensitivities of the modeled ice thickness are different. The effect of modeling sea ice desalination on ice-ocean salt/freshwater accounts for 15 kg m^{-2} (a third) of the total annual salt rejected by FY ice. This effect is an order of magnitude smaller for MY ice.

Our results also indicate a limited picture of how the ice-ocean salt exchanges could be affected by a transition to a seasonal ice cover, often cited as a possible scenario for the next century (e.g., Stroeve *et al.* (2005)). The increased winter ice growth would enhance the total ice-ocean salt flux compared to now, and the summer meltwater inflow would also be larger during summer melt, amplifying the sea ice-induced seasonal cycle of mixed-layer salinity. In conclusion, our study demonstrated that including a halodynamic component would significantly improve the next generation of large-scale sea ice models.

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Chapter 6

Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 1. Model description and validation (V08-1)

The text of the present chapter was submitted as as : Vancoppenolle, M., T. Fichefet, H. Goosse, S. Bouillon, G. Madec and M.A. Morales Maqueda (2008a), Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 1. Model description and validation (V08-1), Ocean modelling.

Abstract

This paper is the first part of a twofold contribution dedicated to the new version of the Louvain-la-Neuve sea ice model LIM3. In this part, LIM3 is described and validated. LIM3 is a C-grid dynamic-thermodynamic model, including the representation of the subgrid-scale distributions of ice thickness, enthalpy, salinity and age. Brine entrapment and drainage as well as brine impact on ice thermodynamics are explicitly included. LIM3 is embedded into the ocean modelling system NEMO, using OPA9, a hydrostatic, primitive equation, finite difference ocean model in the $2^\circ \times 2^\circ \cos\phi$ configuration ORCA2. Model performance is evaluated by performing a hindcast of the Arctic and Antarctic sea ice packs, forced by a combination of daily NCEP/NCAR reanalysis data and various climatologies. The annual cycle of sea ice growth and decay is realistically captured with ice area, thickness, drift, snow depth in reasonable agreement

with observations. In the Arctic, the simulated geographical distributions of ice thickness and concentration significantly improved compared with earlier versions of LIM. Model deficiencies feature an overestimation (underestimation) of ice thickness in the Beaufort Gyre (around the North Pole) as well as an underestimation of ice thickness in the Southern Ocean. The simulated first-year / multiyear sea ice contrast agrees with observations. Nevertheless, the values and distribution of sea ice age in the perennial ice zone are different from satellite-derived values, which is attributed to the different definitions of ice age. LIM3 is thus found appropriate for large-scale sea ice and climate simulations.

6.1 Introduction

Sea ice is a major component of the climate system (e.g., Serreze *et al.* (2007)). It acts as an insulating layer, damping exchanges between atmosphere and ocean. Transfer of radiation, through high albedo (e.g., Perovich *et al.* (2002); Brandt *et al.* (2005)), sensible heat, through low thermal conductivity (e.g., Pringle *et al.* (2007)) and momentum are affected. In addition, salt and freshwater releases respectively associated to the growth and melt of sea ice (e.g., Widell *et al.* (2006); Vancoppenolle *et al.* (2006)) have a significant impact on the oceanic stratification and hence on the World Ocean thermohaline circulation (e.g., Aagaard and Carmack (1989); Goosse and Fichefet (1999)). Sea ice is also an habitat for a group of bacteria and microscopic plants encased in the ice matrix (e.g., Ackley and Sullivan (1994)). It carries considerable amounts of iron (Lannuzel *et al.*, 2007) and modulates the polar CO_2 exchanges (e.g., Delille (2006)) through the intermittently open brine network and, more importantly, through the opening of leads and polynyas. This places sea ice as a major actor of the polar and global carbon cycle. Finally, sea ice is, on the one hand, sensitive to small external perturbations but also, on the other hand, associated to a series of important feedbacks (e.g., Ebert and Curry (1993)). Therefore, sea ice acts not only as an early indicator of climate change but also as an amplifier of climatic perturbations.

In this context, the changes recently observed in the Northern Hemisphere sea ice cover have to be considered with the greatest attention. Passive microwave (PMW) records consistently show that the annual mean Arctic sea ice extent has been decreasing since 1979 by around 3 % per decade (Cavalieri *et al.*, 2003). The trend is higher in summer and accelerated in the 2000s (-9.1% per decade for 1979-2006, see Stroeve *et al.* (2007)), and was characterized by a recent series of minima of extent (Stroeve *et al.*, 2005; Comiso *et al.*, 2008). A downward winter trend is also observed for a few years, reaching -2.9% per decade (Comiso, 2007). Much less is known about ice thickness (and volume) changes since observations

are still sparse. Nevertheless, data from upward-looking sonars (ULS) onboard submarines (e.g., Rothrock *et al.* (1999), Wadhams and Davis (2000)) and electromagnetic soundings (Haas, 2004) consistently point to a significant decrease in ice draft between 22.5 and 32% in several regions of the Arctic basin in the 1990s. Satellite observations of ice thickness data are at their beginning in the Arctic. They give results in agreement with ULS (Laxon *et al.*, 2003; Kwok *et al.*, 2006), though coverage, validation and availability are limited. However, this recent ice thickness trend and its significance are not firmly established because of high interannual and interdecadal variability indicated by altimeter data (Laxon *et al.*, 2003) as well as historical landfast ice measurements (Polyakov *et al.*, 2003) that could not be precisely taken into account in the available observations.

Given the magnitude of the recent changes and their potential impact through feedbacks, it is clear that understanding the mechanisms responsible for those changes and performing reliable projections are crucial issues of current climate studies. Large-scale sea ice models developed in the last decades (e.g., Fichefet and Maqueda (1997); Bitz *et al.* (2001); Hunke and Lipscomb (2004); Zhang and Rothrock (2003)) are included in the coupled models used to perform climate change projections (see, e.g., Arzel *et al.* (2006); Zhang and Walsh (2006)). Those models are based on the same dynamic-thermodynamic structure, discretized using finite differences on structured grids. Typically, the model thermodynamic component is based on the Semtner (1976) 0- or 3-layer models, or on the more sophisticated model of Bitz and Lipscomb (1999). The dynamic component solves the ice momentum equation assuming that sea ice behaves as a two-dimensional continuum with a viscous-plastic (VP) rheology, using either the original VP numerical method (Hibler, 1979), the elastic-viscous plastic (EVP) method (Hunke and Dukowicz, 1997) or a semi-implicit alternate method (Zhang and Hibler, 1997). In addition, some models include the ice thickness distribution (ITD) physics (Thorndike *et al.*, 1975) with various physical (Flato and Hibler, 1995; Zhang and Rothrock, 2003) and numerical (Lipscomb, 2001) improvements. An interactive ITD enables to simulate the faster growth and decay of thin ice compared to thick ice (Maykut, 1986), which in turn leads to a more realistic ice-albedo feedback (Holland *et al.*, 2006b). Nevertheless, the large uncertainties in, on the one hand, coupled ice-ocean hindcast simulations of the 20th century (e.g., Gerdes and Köberle (2007), and in IPCC simulations for the 20th and 21st centuries (e.g., Arzel *et al.* (2006); Lefebvre and Goosse (2008)), on the other hand, stress the need for further model development and validation.

Sea ice is a saline medium. The salt cannot lock into the ice crystalline lattice, but is rather dissolved into liquid inclusions of brine. Due to brine entrapment and drainage, the ice salinity varies in space and time (Malmgren, 1927). Neverthe-

less, presently, the sea ice salinity is either prescribed or ignored in sea ice models. Salinity, in association with temperature, controls the relative brine volume encased in the ice, which in turn affects all ice properties (thermal, mechanical and radiative). Recently, a 1-D energy-conserving halo-thermodynamic model (i.e., including a prognostic salinity profile) was developed and validated (Vancoppenolle *et al.* (2007), hereafter referred to as VBF07). The brine impact on the ice thermal properties influences the seasonal cycle of ice thickness, which in turn affects the exchanges of salt and freshwater between ice and ocean Vancoppenolle *et al.* (2006). Therefore, including salinity in global sea ice models is potentially important.

In this contribution, we describe and validate LIM3¹, a new, large-scale, dynamic-thermodynamic sea ice model, combining the physics of the ITD and brine pockets. It includes a representation of the subgrid-scale distributions of ice thickness, enthalpy, salt and age distributions. Brine pockets and drainage are treated explicitly, following a simplification of the VBF07 1-D model. LIM3 is included in the ocean modelling system NEMO². Our purpose is (1) to document model behavior, in particular the improvement compared with prior versions as well as changes associated with the main originalities of LIM3 and (2) to identify the main shortcomings and orient future research. To accomplish this goal, we analyse the results of LIM3 in a 1970-2006 hindcast performed with NEMO-LIM3 and compare them with several sets of observational data. In the companion paper (Vancoppenolle *et al.*, 2008a) we focus on the importance of the sea ice salinity processes.

6.2 Model description

LIM is originally a dynamic-thermodynamic sea ice model developed by Fichefet and Maqueda (1997). The basis of the coupling with ocean models was introduced by Goosse and Fichefet (1999). LIM was then coupled to OPA v8.2, leading to the ice-ocean model configuration ORCA2-LIM³ (Timmermann *et al.*, 2005), which we will refer to as LIM1 in the remainder of the text. The code was then rewritten formally without any significant physical change, which resulted in the so-called LIM2. The new sea ice model described below has subsequently taken the name of LIM3.

¹LIM stands for Louvain-la-Neuve Ice Model, developed at the Institut d'Astronomie et de Géophysique Geroges Lemaître, Louvain-la-Neuve, Belgium

²NEMO stands for Nucleus for European Modelling of the Ocean

³ORCA2-LIM is a configuration of the OPA (Océan Parallélisé, developed at LOCEAN, Paris, France) model.

The model used in this study is a global coupled ice-ocean model that consists of the ice model LIM3, a C-grid dynamic-thermodynamic sea ice model with thickness, enthalpy, salinity and age distributions, and the ocean model OPA v9.0 (Madec, 2008), in the framework of the NEMO ocean modelling system. The representation of sea ice is done through a velocity vector and a series of ice state variables described further below. A list of model parameters and constants is given in Table 1.

6.2.1 Ocean general circulation model

OPA is a finite difference, hydrostatic, primitive equation ocean general circulation model, with a free sea surface and a non-linear equation of state in the Jackett and McDougall (1995) formulation. Lateral tracer mixing is done along isopycnals. Eddy-induced tracer advection is parameterized following Gent and McWilliams (1990) with the thickness diffusivity decreased in the tropics. Partial steps, i.e., specifying the depth of the bottom cell variable and adjustable to the real depth of the ocean (Pacanowski and Gnanadesikan, 1998), induce a better representation of steep topographic gradients and oceanic general circulation (Barnier *et al.*, 2006). Surface boundary layer mixing and interior vertical mixing are parameterized according to a turbulence closure model (order 1.5) adapted to OPA by Blanke and Delecluse (1993). In case of static instability, a viscosity/diffusivity enhancement up to $10 \text{ m}^2 \text{ s}^{-1}$ is used.

6.2.2 Ice dynamics

The ice velocity $\mathbf{u}$ is determined from the momentum balance :

$$m \frac{\partial \mathbf{u}}{\partial t} = \tau_a + \tau_w - m f \mathbf{k} \times \mathbf{u} - m g \nabla \eta + \nabla \cdot \sigma, \quad (6.1)$$

where m is the mass of snow and ice per unit area, τ_a and τ_w are the contributions of air and water stresses and $\nabla \cdot \sigma$ is the internal stress term. f , g , η and $\mathbf{k}$ are respectively the Coriolis parameter, the acceleration of gravity, the sea-surface elevation and a unit vector pointing upwards.

The latter assumes that sea ice is a two-dimensional VP continuum, following Hibler (1979), with an elliptical yield curve aspect ratio $e = 2$. The momentum equation is solved using the new C-grid formulation (S. Bouillon *et al.*, The Elastic-Viscous-Plastic rheology on different grid types, manuscript in preparation,

Table 6.1: List of model parameters and constants.

M	Number of ice categories	5	-
N	Number of vertical ice layers	5	-
C_a	Atmospheric drag coefficient	1.4×10^{-3}	-
C_w	Ice-ocean drag coefficient	5.0×10^{-3}	-
P^*	Ice strength	40000	N.m^{-1}
C_s	Fraction of energy dissipated by shear	0.5	-
T_e	Elastic waves time scale	9600	s
e	Elliptical yield curve eccentricity	2	-
a^*	e-folding participating area coefficient	0.05	-
H^*	Maximum ridged ice thickness coefficient	100	m
h_P	Maximum rafting ice thickness	0.75	m
C_{ra}	Ridging-rafting transition coefficient	5	m^{-1}
p	Initial ridge porosity	0.3	-
ρ_s	Snow density	330	kg.m^{-3}
ρ_i	Sea ice density	917	kg.m^{-3}
ρ_w	Reference seawater density	1020	kg.m^{-3}
k_s	Snow thermal conductivity	0.31	$\text{W.m}^{-1}.\text{K}^{-1}$
k_0	Pure ice thermal conductivity	2.01	$\text{W.m}^{-1}.\text{K}^{-1}$
c_0	Pure ice specific heat	2067	$\text{J.kg}^{-1} \text{K}^{-1}$
L_0	Pure ice latent heat of fusion	334×10^3	J.kg^{-1}
L_s	Pure ice latent heat of sublimation	2834×10^3	J.kg^{-1}
μ	Sea ice melting point coefficient	0.054	$\text{‰}^\circ\text{C}^{-1}$
β_1	First thermal conductivity coefficient	0.09	$\text{W.m}^{-1}.\text{‰}^{-1}$
β_2	Second thermal conductivity coefficient	0.011	$\text{W.m}^{-1}.\text{K}^{-2}$
ϵ	Sea ice surface emissivity	0.97	-
S^G	Winter equilibrium sea ice salinity	5	‰
S^F	Summer equilibrium sea ice salinity	2	‰
S^1	First salinity transition coefficient	3.5	‰
S^2	Second salinity transition coefficient	4.5	‰
T^G	Winter sea ice salinity time scale	20	days
T^F	Summer sea ice salinity time scale	10	days

hereafter referred to as Bouillon et al., 2008, in preparation) of EVP (Hunke and Dukowicz, 1997). We use 300 sub-iterations for ice dynamics, so that the time step used in the dynamic component is $\Delta t_{dyn} = 960$ s. An artificial regularization of small deformation rates is made by introducing damped elastic waves, with the damping time scale set to 9760 s, which is high enough to ensure model stability Hunke (2001) and small enough so that elastic waves are damped when ice dynamics have converged. The stress tensor is kept in memory at the end of each time step, which improves convergence and global model stability. Stress tensor invariants used to force the mechanical redistribution module are computed at the end of the dynamic time step. For the ice strength, we follow Hibler (1979) formulation and use $P^* = 40,000$ N m⁻¹, close to the satellite-derived value of Tremblay and Hakakian (2007).

Compared to the former B-grid VP rheology, experiments with LIM2 show that EVP has higher stability and responds more rapidly to changes in the wind forcing. Because it allows to drastically reduce the numerical viscous flow limit, using EVP gives a better solution of the ice momentum equation. In addition, EVP is explicit, which allows easier parallelization. Furthermore, the C-grid has better dispersion properties than a B-grid and straits are better represented. Finally, the coupling to the OPA model, written on a C-grid, is more natural. Nevertheless, the C-grid requires the interpolation of velocities in order to calculate the Coriolis term.

6.2.3 Ice state variables

The thickness distribution is numerically discretized into several ice thickness categories. The numerical formulation of the thickness categories follows Bitz *et al.* (2001) and Lipscomb (2001). A fixed number M of thickness categories ($M = 5$) is imposed. Each thickness category has a mean thickness h_m^i ($m = 1, 2, \dots, M$) which is constrained to lie between H_{m-1} and H_m . H_0 is set to zero. The other boundaries are chosen with greater resolution for small h (see Fig. 6.1). Sea ice in each category is divided in N vertical layers of ice and one layer of snow.

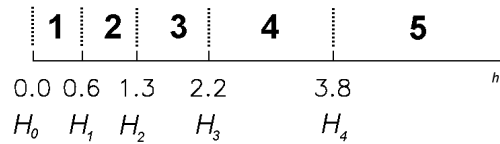


Figure 6.1: Boundaries of the model ice thickness categories (m).

Table 6.2: Sea ice state variables in LIM3.

Symbol	Description	Units
$\mathbf{u}$	Sea ice velocity	$[m.s^{-1}]$
g_m^i	Concentration of sea ice in category m	$[-]$
v_m^i	Volume of sea ice per unit area in category m	$[m]$
v_m^s	Volume of snow per unit area in category m	$[m]$
$e_{m,k}^i$	Sea ice enthalpy per unit area in category m and layer k	$[J.m^{-2}]$
e_m^s	Snow enthalpy per unit area in category m	$[J.m^{-2}]$
M_m^s	Sea ice salt content in category m	$[\% . m]$
O_m	Sea ice age content in category m	$[days.m]$

Each ice category has its own set of global state variables (see Table 6.2 for a list). The global ice state variables are extensive variables which include the ice concentration (g_m^i), volume per unit area (v_m^i) and enthalpy per unit area ($e_{m,k}^i$, $k = 1, 2, \dots, N$), salt content (M_m^s) and age content (O_m), as well as the snow volume per unit area (v_m^s) and enthalpy per unit area (e_m^s). The global state variables are used by all but thermodynamic model routines, in which they are, for convenience, converted into equivalent state (intensive) variables (i.e., thickness, temperatures, see Table 6.3 for a list). In the remainder of the text, the m index runs for ice thickness categories and k for the vertical ice layers.

The global ice variables undergo changes that are classified into changes due to transport, thermodynamic and mechanical redistribution processes. Extending equations (4) to (6) of Bitz *et al.* (2001), we impose that all the global ice state variables (X) follow a conservation equation of the form :

$$\frac{\partial X}{\partial t} = -\nabla \cdot (X \mathbf{u}) + \Psi^X + \Theta^X, \quad (6.2)$$

where Θ^X and Ψ^X represent the effects of thermodynamical and mechanical redistribution processes on X , respectively.

6.2.4 Transport

As in Fichefet and Maqueda (1997), the global ice state variables are transported horizontally using the advection scheme of Prather (1986). This scheme is based on the conservation of second-order moments of the advected quantity. It is weakly diffusive and avoids unphysical negative values.

Table 6.3: Equivalent variables in LIM3. To each state variable corresponds an equivalent variable. Equivalent variables are used only in 1-D modules.

Symbol	Description	Units
$h_m^i = v_m^i/g_m^i$	Ice thickness	[m]
$h_m^s = v_m^s/g_m^i$	Snow depth	[m]
$q_{m,k}^i = e_{m,k}^i/(h_m^i/N)$	Ice specific energy of melting	[J.m ⁻³]
$q_m^s = e_m^s/h_m^s$	Snow specific energy of melting	[J.m ⁻³]
$T_{m,k}^i = f(q_{m,k}^i)$	Ice temperature	[K]
$T_m^s = f(q_m^s)$	Snow temperature	[K]
$S_m^i = M_m^s/v_m^i$	Ice salinity	[‰]
$o_m^i = O_m/v_m^i$	Ice age	[days]

6.2.5 Mechanical redistribution

Divergence and shear open the ice pack and create ice of zero thickness. Convergence and shear consumes thin ice and create thicker ice by mechanical deformation through rafting (piling of two ice sheets on top of each other) and ridging (piling of a series of broken ice blocks into ridges). The redistribution functions Ψ^X describe how opening and deformation affect the global ice state variables. In the model, they are computed as functionals of the stress rate tensor and the thickness distribution itself.

6.2.5.1 Redistribution function

A general expression of Ψ^g , the mechanical redistribution function associated to the ice concentration, was proposed by Thorndike *et al.* (1975):

$$\Psi^g = |\dot{\epsilon}|[\alpha_o(\theta)\delta(h) + \alpha_d(\theta)w_d(h, g)], \quad (6.3)$$

The first and second terms on the right-hand side correspond to opening and deformation, respectively. $|\dot{\epsilon}| = (\dot{\epsilon}_I^2 + \dot{\epsilon}_{II}^2)^{1/2}$, $\dot{\epsilon}_I$ and $\dot{\epsilon}_{II}$ are the strain rate tensor invariants, $\theta = \text{atan}(\dot{\epsilon}_{II}/\dot{\epsilon}_I)$ and $w_d(h, g)$ is the deforming mode. $|\dot{\epsilon}|\alpha_o$ and $|\dot{\epsilon}|\alpha_d$ are the lead opening and closing rates, respectively, and they are related by $|\dot{\epsilon}|\alpha_o - |\dot{\epsilon}|\alpha_d = |\dot{\epsilon}_I|$. The closing rate can be written as a sum of two terms representing the energy dissipation by shear and convergence Flato and Hibler (1995):

$$|\dot{\epsilon}|\alpha_d(\theta) = C_s \frac{1}{2}(\Delta - |\dot{\epsilon}_I|) - \min(\dot{\epsilon}_I, 0), \quad (6.4)$$

where $\Delta = (\dot{\epsilon}_I^2 + \dot{\epsilon}_{II}^2/e^2)^{1/2}$ for the elliptical yield curve. The factor C_s is added to allow for energy sinks other than ridge building (e.g., sliding friction) during shear.

Babko *et al.* (2002) concluded that rafting plays a significant role during initial ice growth in fall. Therefore, we assume that the deforming mode consists not only of ridging (which redistributes participating ice on a various range of thicknesses) but also rafting (which doubles the participating ice thickness):

$$w_d(h, g) = -[b^{ra}(h) + b^{ri}(h)]g(h) + n^{ra}(h) + n^{ri}(h), \quad (6.5)$$

where $b^{ra}(h)$ and $b^{ri}(h)$ are the rafting and ridging participation functions. They determine which regions of the ice thickness space participate in the redistribution. $n^{ra}(h)$ and $n^{ri}(h)$, called transfer functions, specify how thin, deforming ice is redistributed onto thick, deformed ice. Participation and transfer functions are normalized in order to conserve area.

6.2.5.2 Participation functions

We assume that the participation of ice in redistribution does not depend upon whether the deformation process is rafting or ridging. Therefore, the participation functions can be written as follows:

$$b^{ra}(h) = \beta(h)b(h), \quad (6.6)$$

$$b^{ri}(h) = [1 - \beta(h)]b(h), \quad (6.7)$$

where $b(h)$ is an exponential weighting function with an e-folding scale a^* (Lipscomb *et al.*, 2007), which preferentially apportions the thinnest available ice to ice deformation. It is numerically more stable than the original version of Thorndike *et al.* (1975). $\beta(h)$ partitions deforming ice between rafted and ridged ice. $\beta(h)$ is formulated following Haapala (2000), using the Parmeter (1975) law, which states that, under a critical participating ice thickness h_P , ice rafts, otherwise it ridges:

$$\beta(h) = \frac{\tanh[-C_{ra}(h - h_P)] + 1}{2}, \quad (6.8)$$

where $C_{ra} = 5 \text{ m}^{-1}$ and $h_P = 0.75 \text{ m}$ (Haapala, 2000; Babko *et al.*, 2002). The $\tanh$ function is used to smooth the transition between ridging and rafting. In spite of its relative simplicity (see, e.g., Tuhkuri and Lensu (2002)), our formulation is a step towards a better representation of mechanical redistribution in models.

6.2.5.3 Transfer functions

The rafting transfer function assumes a doubling of ice thickness :

$$n^{ra}(h) = \frac{1}{2} \int_0^\infty \delta(h - 2h')b(h')g(h')dh, \quad (6.9)$$

where δ is the Dirac delta function, while the ridging transfer function is :

$$n^{ri}(h) = \int_0^\infty \gamma(h', h)(1 + p)b(h')g(h')dh. \quad (6.10)$$

The redistributor $\gamma(h', h)$ specifies how area of thickness h' is redistributed on area of thickness h . We follow Hibler (1980) who constructed a rule, based on observations, that forces all ice participating in ridging with thickness h' to be linearly distributed between ice that is between $2h'$ and $2\sqrt{H^*h'}$ thick. This in turn determines how to construct the ice volume redistribution function Ψ^v . Volumes equal to participating area times thickness are removed from thin ice. They are redistributed following Hibler's rule. The factor $(1+p)$ accounts for initial ridge porosity p , defined as the fractional volume of seawater initially included into ridges. In many previous models, the initial ridge porosity has been assumed to be 0, which is not the case in reality since newly formed ridges are porous, as indicated by in-situ observations Leppäranta *et al.* (1995); Hoyland (2002). In other words, LIM3 creates a higher volume of ridged ice with the same participating ice. $p = 0.3$ is chosen following the observations of Leppäranta *et al.* (1995).

6.2.5.4 Mechanical redistribution for other global ice variables

The other global ice state variables redistribution functions Ψ^X are computed based on Ψ^g for the ice age content and on Ψ^{v^i} for the remainder (ice enthalpy and salt content, snow volume and enthalpy). The general principles under this derivation are described in Appendix A of Bitz *et al.* (2001). A fraction $f_s = 0.5$ of the snow volume and enthalpy is assumed to be lost during ridging and transferred to the ocean. The contribution of the seawater trapped into the porous ridges is included in the computation of the redistribution of ice enthalpy and salt content (i.e., Ψ^{e^i} and Ψ^{M^s}). During this computation, seawater is supposed to be in thermal equilibrium with the surrounding ice blocks. Ridged ice desalination induces an implicit decrease in internal brine volume, and heat supply to the ocean, which accounts for ridge consolidation as described by Hoyland (2002). The inclusion of seawater in ridges does not imply any net change in ocean salinity. The energy used to cool down the seawater trapped in porous ridges until the seawater freezing point is rejected into the ocean.

6.2.5.5 Role of ridge porosity and rafting

Sensitivity experiments indicate that assuming a non-zero ridge porosity increases the simulated Northern Hemisphere ice volume by a few percent, which suggests that an additional melt of the warm porous ridges compensates the additional ice volume created initially. No significant difference is found in the Southern Hemisphere.

Including rafting in the model increases the total average simulated ice volume in the Southern Hemisphere by 25%. In this region, thinner rafted ice occupies more area than thicker ridged ice, in particular in the winter-spring marginal ice zone. In spring, less solar radiation is absorbed and less bottom melt occurs than if only ridging is included, which leads to a summer ice extent larger by $1 \times 10^6 \text{ km}^2$. In the Northern Hemisphere, where the deformed ice volume is higher, thinner rafted ice results only in a 2% decrease in the simulated ice volume.

6.2.6 Ice thermodynamics

The thermodynamic processes (Θ^X) include vertical diffusion of heat in the snow/ice system, snow/ice growth and decay, creation of new ice in open water, brine drainage and ice natural ageing. After thermodynamics are computed, the linear remapping scheme of Lipscomb (2001), which computes the transport in thickness space among categories due to ice growth/melt, is applied to all state variables. This scheme is weakly diffusive, converges with a few thickness categories and its computational cost is minimal. In LIM3, there is no explicit account for lateral melting. Lateral melting is controlled by floe size (Steele, 1992), which is not a model variable. Nevertheless, as in other models including ITD, there is some implicit lateral melting accounted for by the melt of thin ice (Bitz *et al.*, 2001).

6.2.6.1 Vertical growth and decay

The formulation of ice vertical thermodynamics follows from Bitz and Lipscomb (1999) and VBF07. Vertical heat conduction and storage and vertical temperature profile in the sea ice are governed by the heat diffusion equation. Surface energy balance and seawater freezing point provide the boundary conditions at the top and bottom interfaces, respectively. The albedo is a function of ice thick-

ness and snow depth (Shine and Henderson-Sellers, 1985). A fraction i_0 of the shortwave radiation absorbed at the surface, function of cloudiness, is transmitted inside the ice, as in Fichet and Maqueda (1997). Radiation is absorbed in the ice cover following Beer's law, with an attenuation coefficient $\kappa_i = 1 \text{ m}^{-1}$. Radiation does not penetrate in snow. For each thickness category, the heat diffusion equation is solved in the snow-ice system in one layer of snow and $N = 5$ layers of ice. This number can decrease to 2 without significantly deteriorating the results. The imbalance between external and inner heat conduction fluxes at the interfaces are used to compute growth / melt rates. The salinity evolution is then computed. Finally, to account for changes in ice/snow thickness and salinity in an energy-conserving way, the new temperature profile is remapped on the updated vertically uniform grid.

6.2.6.2 Brine pockets

The impact of brine pockets on heat transfer and storage is done following the lines drawn by Maykut and Untersteiner (1971); Bitz and Lipscomb (1999), by formulating sea ice thermal properties as functions of salinity S and temperature T (in °C). The sea ice specific heat is given by (Malmgren, 1927) :

$$c_i(S, T) = c_0 + L_0 \mu \frac{S}{T}, \quad (6.11)$$

where c_0 and L_0 are the specific heat and latent heat of fusion of fresh ice, respectively, μ is an empirical constant relating the seawater freezing point to its salinity $T_m = -\mu S$. The sea ice thermal conductivity follows (Pringle *et al.*, 2007) :

$$k_i(S, T) = k_0 + \beta_1 \frac{S}{T} - \beta_2 T, \quad (6.12)$$

where k_0 is the thermal conductivity of pure ice and β_1 and β_2 are empirical constants. Following Bitz and Lipscomb (1999), we assume that the sea ice specific energy of melting q (i.e., the energy required to melt a unit volume of sea ice) is also a function of salinity and temperature :

$$q(S, T) = \rho_i c_0 (-\mu S - T) + \rho_i L_0 (1 + \frac{\mu S}{T}), \quad (6.13)$$

where ρ_i is the pure ice density. The enthalpy per unit area in any ice layer of thickness $\Delta h = h_i/N$ is $q(S, T)\Delta h$. With such a formulation, saline ice grows more than fresh ice for a given heat loss.

6.2.6.3 Sources and sinks of ice mass

Ice mass can be created by congelation growth at the ice bottom, by new ice growth in open water areas and by snow ice formation. Ice mass can be lost by melting at the top and bottom interfaces.

Bottom growth. If heat is lost at the ice base, ice grows. The ice growth rate in the m^{th} category is given by :

$$\frac{\partial h_m^{i,b+}}{\partial t} = -\frac{F_m^b}{q(S_m^{i,b}, T^w)}, \quad (6.14)$$

where F_m^b is the heat loss at the bottom surface in the m^{th} category. T^w is the ice-ocean interface temperature, which is equal to the seawater freezing point. $S_m^{i,b}$, the salinity of new ice in the m^{th} category, is given by :

$$S_m^{i,b} = \nu \left(\frac{\partial h_m^{i,b+}}{\partial t} \right) S^w, \quad (6.15)$$

where ν is a fractionation coefficient depending on ice growth (Cox and Weeks, 1988) and S^w is the ocean salinity. The system of 3 nonlinear equations (6.13), (6.14) and (6.15) is solved iteratively. The associated enthalpy per unit area is added at the ice bottom.

Snow ice. The formation of snow ice results from the transformation of snow into ice at the surface. As in Fichefet and Maqueda (1997), we assume that when snow is thick enough to depress the snow-ice interface below the sea level, seawater infiltrates and freezes into the snow, so that snow ice formation occurs :

$$\Delta h_m^{si} = \frac{\rho_s h_m^s - (\rho_w - \rho_i) h_m^i}{\rho_s + \rho_w - \rho_i}, \quad (6.16)$$

where ρ_w and ρ_s are the reference density of seawater and of snow, respectively. The newly created ice is assumed to have $T = T^w$ and $S = S_m^{si} = (\rho_i - \rho_s) / \rho_i S^w$. A corresponding enthalpy per unit area $q(S_m^{si}, T^w)$ is added on top of the sea ice surface and merged with the remainder of the sea ice column. The residual energy (used to warm the snow to T^w) is removed from the top ocean layer.

Ice melt. Ice melts following a convergence of heat. The ice melt rate at the top and bottom interfaces in the m^{th} category is computed by solving $\partial h_m^{i-} / \partial t = -F_m / q(S_N, T_N)$, where F_m is the net heat flux, computed from the difference between external fluxes and internal conductive heat flux. $q(S_N, T_N)$ is the energy of melting of the interfacial ice layer.

New ice formation. If the top ocean layer is at its freezing point and that the surface is losing heat, then new ice forms. The volume of ice forming in open water areas, which has a thickness h^0 , is given by :

$$\frac{\partial v^{i,0}}{\partial t} = -\frac{F^{ow}}{q(S^0, T^w)}, \quad (6.17)$$

where F^{ow} is the heat loss in open water areas. The salinity of new ice S^0 is a function of ice thickness, following the observation-based regression of Kovacs (1996) :

$$S^0 = 4.606 + \frac{0.91603}{h^0}. \quad (6.18)$$

The thickness of new ice (or frazil ice collection thickness) h^0 is computed using a parameterization based on Biggs *et al.* (2000). In this parameterization, the conservation of momentum and energy of the frazil-pack ice system gives the thickness of new ice as a nonlinear function of ice velocity, wind speed and pack-ice thickness. It gives new ice thicknesses ranging from 5 to 15 cm. Finally, the fractional area of new ice is diagnosed, and new ice (i.e., all its state variables) is merged in the category where h^0 stands. This category has generally $m = 1$. This value is well suited for the representation of new ice growth in the calm waters of the Arctic Ocean, but is not ideal for dynamical (pancake) ice growth processes in turbulent conditions, prevalent in the Southern Ocean. In LIM1, to overcome this, $h^0 = 1$ m was used.

6.2.6.4 Snow

Snow has constant density, specific heat, thermal conductivity and latent heat, one temperature point and no salinity. Snow depth increases when snow falls and decreases by snow ice formation, sublimation and melt. Clearly, the representation of snow is simple and should be improved in future model versions.

6.2.6.5 Salinity variations

The global sea ice state variable associated to the ice salinity in the m^{th} category is the sea ice salt content $M_m^s = S_m^i v_m^i$, where S_m^i is the vertically averaged (hereafter referred to as "bulk") sea ice salinity in the m^{th} category. The evolution of M_m^s follows an evolution equation of the type (6.2). Θ^{M^s} includes (1) brine entrapment and drainage and (2) transport in thickness space due to ice growth/melt.

The representation of brine entrapment and drainage is based on a simplification of the more complex VBF07 model. Field data (Malmgren, 1927), laboratory experiments (Cox and Weeks, 1975) and 1-D simulations with the VBF07 model (Vancoppenolle *et al.*, 2006) indicate that the sea ice desalination can be divided into initial, winter and summer stages. Gravity drainage and flushing dominate winter and summer desalinations, respectively. In addition, salinity data from ice cores drilled in some of the Southern Ocean deep snow-covered regions suggest that salt concentration occurs at the ice surface during snow ice formation (Jeffries *et al.*, 1997). Therefore, in the thermodynamic module, the following equation for S_m^i is solved:

$$\frac{\partial S_m^i}{\partial t} = \frac{\nu S^w}{h_m^i} \frac{\partial h_m^{i,b+}}{\partial t} + S_m^{si} \frac{\partial h_m^{i,si}}{\partial t} - \left(\frac{S_m^i - S^G}{T^G} \right) I^G - \left(\frac{S_m^i - S^F}{T^F} \right) I^F. \quad (6.19)$$

The terms on the right-hand side refer to brine entrapment during ice growth, snow ice formation, gravity drainage and flushing, respectively. T^G and S^G (T^F and S^F) are the winter (summer) desalination time scale and equilibrium salinity. $I^G = 1$ if the difference between surface and bottom ice temperatures is negative and 0 otherwise. This corresponds to a temperature gradient directed downwards, which, since brine salinity increases toward low temperatures, leads to an unstable brine density gradient in the brine network and a subsequent desalination (Cox and Weeks, 1988). $I^F = 1$ if surface melts and 0 otherwise.

We assume that salinity has a vertical distribution $S_{m,k}^i$ only in thermodynamic computations. Observation of the vertical sea ice salinity profile from ice cores suggest that the shape of the profile is a function of the vertically averaged salinity because flushing affects both the mean salinity and profile shape (VBF07). Numerical experiments have shown that a linear profile well approximates the sea ice salinity profile (Vancoppenolle *et al.*, 2005). Therefore, we assume the following. At high mean salinity, i.e., if $S_m^i > S_2 = 4.5\%$, the profile is isosaline: $S_{m,k}^i = S_\infty(z) = S_m^i$. At low mean salinity, i.e., if $S_m^i < S_1 = 3.5\%$, the profile is linear: $S_{m,k}^i = S_0(z)$. The profile is constrained by its mean salinity equal to S_m^i , given by (6.19), and by $S = 0\%$ at the surface :

$$S_0(z) = 2 \frac{S_m^i}{h_m^i} z. \quad (6.20)$$

At intermediate salinities, i.e., between S_1 and S_2 , $S_{m,k}^i$ is a combination of S_0 and S_∞ . In summary,

$$S_{m,k}^i(z) = \alpha(S_m^i) S_0(z_k) + [1 - \alpha(S_m^i)] S_\infty(z_k), \quad (6.21)$$

where

$$\alpha(S_m^i) = \begin{cases} 1 & \text{if } S_m^i < S_1 \\ (S_m^i - S_2)/(S_1 - S_2) & \text{if } S_1 \leq S_m^i \leq S_2 \\ 0 & \text{if } S_m^i > S_2 \end{cases}.$$

Local thermodynamic equilibrium is assumed during brine drainage. Therefore, an increase in temperature compensates any desalination. Finally, assuming thermal equilibrium in the brine inclusions, the relative brine volume $e_{m,k}$ can be diagnosed from salinity and temperature by the relation $e_{m,k} = -\mu S_{m,k}/T_{m,k}$.

6.2.7 Ice age.

For diagnostic purposes, the ice age o_m^i of each thickness category is computed. Utilisation of age in marine modelling has shown that the values and interpretation of model age are strongly dependent on the age definition (see, e.g., Deleersnijder *et al.* (2001)). We assume that the age is associated to the areal sea ice age content $O_m = o_m^i g_m^i$, which verifies:

$$\frac{\partial O_m}{\partial t} = -\nabla \cdot (O_m \mathbf{u}) + \Theta_m^O + \Psi_m^O. \quad (6.22)$$

For coherence, first, the mechanical redistribution function associated to ice age Ψ^O is constructed based on Ψ^g (see earlier). Second, in contrast to Harder and Lemke (1994), vertical growth and melt do not affect the ice age (i.e., vertically accreted new ice is assumed to have the age of existing ice). Nevertheless, new ice formed in open water has an age equal to zero. Therefore, our value reflects an areal residence time, larger than the actual ice age, and the thermodynamic ageing term reads:

$$\Theta_m^O = g_m^i - \frac{\partial(f_m O_m)}{\partial h}, \quad (6.23)$$

where f_m is the sea ice growth/melt rate in the m^{th} category. The first term on the right-hand side represents ice ageing. The second term accounts for the formation of new ice in open water and for the transport in thickness space of the age content due to ice growth and melt.

6.2.8 Ice-ocean coupling

The ice-ocean coupling is conceptually similar to Goosse and Fichefet (1999). Ice and ocean exchange momentum through an ice-ocean stress deriving from a quadratic bulk formula assuming an oceanic drag coefficient of $C_w = 5.0 \times 10^{-3}$.

The oceanic heat flux, function of the temperature of the ocean and turbulent mixing, derives from McPhee (1992) and Goosse and Fichefet (1999):

$$F_w = \rho_w c_w C_h u_{io}^* (T^w - T^{ml}), \quad (6.24)$$

where c_w is the seawater specific heat, $C_h = 0.006$ is a heat transfer coefficient (McPhee, 1992), T^{ml} is the mixed layer temperature and u_{io}^* is the friction velocity, which depends on the ice-ocean RMS velocity difference.

Freshwater fluxes for ocean and sea ice use the approach of Tartinville *et al.* (2001). Adaptations were made to account for time-varying sea ice salinity. From Tartinville *et al.* (2001), we adopt the following convention. First, if liquid or solid water is added to or extracted from the ice-ocean system (e.g., through precipitation or evaporation), then it is introduced as a freshwater flux. Second, if a surface process affects ocean salinity without a net gain or loss of water for the ice-ocean system (e.g., through ice melting or freezing) this internal exchange is transformed into an equivalent salt flux that is applied to the ocean. In this approach, sea ice acts as a negative reservoir of salt inside the ice-ocean system. As ice is formed, salt is released into the ocean. As ice melts, salt is taken out of the ocean.

The freshwater flux at the surface of the ocean F^f (in kg/m²/s) is given by :

$$F^f = P - E - \frac{\rho_s}{\rho_w} \sum_{m=1}^M \frac{\partial v_m^{s,-}}{\partial t}, \quad (6.25)$$

where P is the precipitation, E the evaporation, $\partial v_m^{s,-}/\partial t$ is the snow volume loss per unit area in the m_{th} category due to snow melt and snow ice formation. The salt flux F^s reads :

$$F^s = F_b^s + F_{eq}^s. \quad (6.26)$$

F_b^s is the salt flux due to brine drainage :

$$F_b^s = \rho_i \sum_{m=1}^M v_m^i \left(\left. \frac{\partial S_m^i}{\partial t} \right|_G + \left. \frac{\partial S_m^i}{\partial t} \right|_F \right). \quad (6.27)$$

The partial derivatives on the right-hand side refer to the loss of salt by gravity drainage and flushing (next-to-last and last terms of (6.19)), respectively. F_{eq}^s is the equivalent salt flux due to ice growth and melt:

$$F_{eq}^s = \rho_i (S^w - S^n) \frac{\partial v^{i,n}}{\partial t} + \rho_i \sum_{m=1}^M g_m^i [(S^w - S_m^{i,b}) \frac{\partial h_m^{i,b+}}{\partial t} + (S^w - S_m^{si}) \frac{\partial h_m^{i,si}}{\partial t} + (S^w - S_m^i) \frac{\partial h_m^{i,-}}{\partial t}] \quad (6.28)$$

The terms on the right-hand side respectively correspond to the salt rejected during new ice formation in open water, to bottom congelation growth, to snow ice formation and to the melt of saline ice.

6.3 Forcing and experimental design

6.3.1 Model grid

The ocean and sea ice models both run over the same domain, extending from 78°S to 90°N. The model uses the tripolar ORCA2 grid with 2° zonal resolution and a meridional resolution varying from 0.5° at the equator to 2° $\cos\phi$ at a latitude ϕ (see Timmermann *et al.* (2005) and Madec (2008); for details). The grid features two points of convergence in the Northern Hemisphere, both situated on continents. The ocean model has a time step ($\Delta t_o = 5760$ s = 1/15 d) and the ice model is called every 5 ocean time steps ($\Delta t_i = 28800$ s = 1/3 d).

6.3.2 Simulation setup

The simulation is conducted over the period 1948-2006. Unless otherwise stated, years 1979-2006 are analysed. The initial ocean temperature and salinity fields are specified using the January data from the Polar Science Center Hydrographic Climatology Steele *et al.* (2001)). Initial mean sea ice thicknesses (snow depths) of 3.5 m (0.3 m) in the Arctic and 1 m (0.1 m) in the Southern Ocean are imposed in regions with a sea surface temperature below 0° C. Initial ice concentration is assumed to be 0.95 (0.90) in the Arctic (Antarctic). Initially, the ice thickness distribution is assumed to be Gaussian, age and velocity are prescribed to 0, salinity to 6 ‰ and ice and snow temperatures to 270 K.

6.3.3 Model forcing

Atmospheric forcing fields are a combination of NCEP/NCAR daily reanalysis data of 10-m wind speed and 2-m temperature (Kalnay *et al.*, 1996), and monthly climatologies of relative humidity (Trenberth *et al.*, 1989), total cloudiness (Berland and Strokina, 1980) and precipitation (Large and Yeager, 2004).

Reanalysis and climatological data are interpolated from a 2° × 2° latitude-longitude grid to the ORCA2 grid with a 3rd order moments conserving algorithm and interpolated linearly in time to get a value at each oceanic time step. Surface radiative and turbulent heat fluxes are calculated as in Goosse (1997). Surface wind stress derives from quadratic bulk formulae assuming an atmospheric drag

coefficient of $C_a=1.40\times 10^{-3}$. Evaporation/sublimation is derived from the latent heat flux. River runoff rates are prescribed using the climatological Baumgartner and Reichel (1975) dataset combined with a mean seasonal cycle derived from the Global Runoff Data Centre (GRDC) data. Contrary to earlier versions of the model, no salinity restoring is used.

6.4 Results

In this section, we discuss the simulated general ice physical state: the ice areal coverage, volume and drift, the snow volume, and the ice age and salinity fields. We compare the model results with available observations and to LIM1 and AOMIP (Arctic Ocean Model Intercomparison Project) hindcast simulations. The LIM1 control simulation setup and forcing configuration is similar to the LIM3 simulation discussed hereafter, except that the run spans two times the 1977-1999 period, and that an older precipitation dataset is used (Timmermann *et al.*, 2005). LIM1 is a single-category model with VP dynamics and Semtner (1976) 3-layer thermodynamics. The AOMIP (see Johnson *et al.* (2007); Gerdes and Köberle (2007)) simulations are a series of 1948-2001 hindcast runs with 9 different ice-ocean models, including several physical parameterizations at various spatial resolutions. There is not yet an equivalent intercomparison project in the Southern Hemisphere.

6.4.1 General view of the ice physical state

The physical state of the winter ice cover is summarized in Fig. 6.2 and several statistics of the simulation are shown in Table 6.4. The model simulates significantly different ice packs in the Northern and Southern Hemispheres (hereafter referred to as NH and SH). Arctic ice is on average (1979-2006) older (2.6 y) than its Antarctic counterpart (100 days). Therefore, it is also thicker (1.9 m in the NH vs. 41 cm in the SH), has a deeper snow cover (24 cm vs. 18 cm), is less saline (3.6‰ vs. 6.3‰) and has a lower brine volume (6.3% vs. 7.5%). The sources and sinks of ice mass also differ markedly (see later on). This corresponds generally to in-situ observations of ice thickness and snow depth, and textural analysis of ice cores from the Arctic and Antarctic regions (see, e.g., Tucker *et al.* (1987); Eicken (1998)).

Table 6.4: Global statistics for the control simulation.

Variable (units)	Hem.	Mean	Av. max	Av. min	T_{mean} (units/decade)	T_{max}	T_{min}
Area (10^6 km^2)	N	10.2 ± 0.38	14.2 ± 0.26	4.4 ± 0.77	-0.32 ± 0.007	-0.18 ± 0.005	-0.62 ± 0.01
Thickness (cm)		190 ± 24	269 ± 38	141 ± 22	-24.5 ± 3.4	-34.0 ± 6.0	-22.4 ± 3.2
Velocity (cm/s)		4.48 ± 0.33	5.20 ± 0.62	3.33 ± 0.74	0.27 ± 0.06	0.003 ± 0.13	0.39 ± 0.02
Salinity (‰)		3.59 ± 0.20	4.59 ± 0.31	2.04 ± 0.01	0.20 ± 0.03	0.30 ± 0.04	0.005 ± 0.004
Age (years)		2.62 ± 0.34	3.97 ± 0.38	2.06 ± 0.35	-0.24 ± 0.07	-0.13 ± 0.09	-0.27 ± 0.07
Area (10^6 km^2)	S	11.8 ± 0.36	19.5 ± 0.45	0.93 ± 0.55	0.25 ± 0.007	0.24 ± 0.01	0.41 ± 0.01
Thickness (cm)		41 ± 1.3	58 ± 1.6	22 ± 3.8	0.6 ± 0.3	-0.5 ± 0.4	2.8 ± 0.7
Velocity (cm/s)		6.34 ± 0.20	8.64 ± 0.80	2.31 ± 0.37	0.09 ± 0.04	-0.01 ± 0.19	0.22 ± 0.007
Salinity (‰)		6.29 ± 0.11	9.13 ± 0.61	4.30 ± 0.46	-0.08 ± 0.01	-0.45 ± 0.12	0.27 ± 0.10
Age (years)		0.31 ± 0.03	0.94 ± 0.13	0.10 ± 0.04	0.03 ± 0.004	0.04 ± 0.03	0.04 ± 0.007

The means are successively computed over categories (weighted by the area of each of the categories for the computation of area, velocity and age; by volume for thickness and salinity), then over the ice covered domain and finally over the 1979-2006 time series. Maxima and minima refer to the extremum values over the mean seasonal cycle. The trends (T_{mean} , T_{max} and T_{min}) and the uncertainty are computed using the time series of the annual mean, minimum and maximum values.

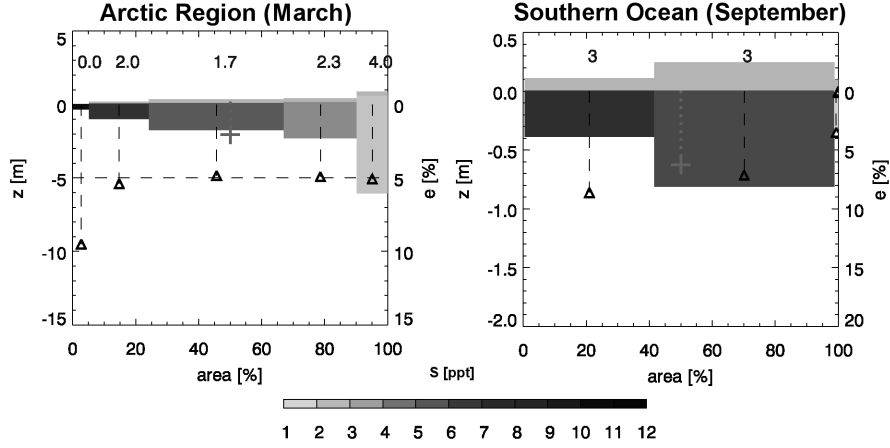


Figure 6.2: Average winter sea ice physical state (1979-2006), for the Arctic regions (including the Arctic Ocean and the surrounding continental seas) and the whole SH, detailed for each of the ice thickness categories. The ice concentrations are represented horizontally. The ice thicknesses (negative values) and snow depths (positive values) are shown vertically. The grey colors refer to the ice salinity. The black triangles indicate the relative brine volume (computed from temperature and salinity), averaged over the vertical ice layers. The numbers on top are ice ages (years for the Arctic and months for the Antarctic). The cross indicates the mean ice thickness.

6.4.2 Ice areal coverage

In this section, the mean seasonal cycle and interannual variations of the global hemispheric ice coverage as well as the geographical distribution of ice concentration are discussed.

6.4.2.1 Mean seasonal cycle

Averaged over 1979-2006, in the NH, the simulated maximum (minimum) global hemispheric area is equal to 14.2 (4.4) $\times 10^6$ km^2 , and is reached on Feb 23 (Aug 26). The mean seasonal cycle of ice area (Fig. 6.3) closely follows the Bootstrap PMW recordings Comiso (2007). The minimum area is however underestimated by 1×10^6 km^2 and is reached 15 days earlier than observed. The ice extent (i.e., the area enclosed within the 15% concentration contour) is underestimated all year long by 1 - 1.5×10^6 km^2 . In the SH, the area maximum (minimum) is 19.5 (0.9) $\times 10^6$ km^2 and is reached on Aug 15 (Feb 15). The seasonal cycle is

reasonably well represented. Areal growth is faster than observed. The maximal area is reached 20 days in advance and is significantly overestimated, by 2.5×10^6 km². The simulated seasonal cycle of ice extent is much more realistic, which suggests that the model SH ice concentration is higher than in the observations.

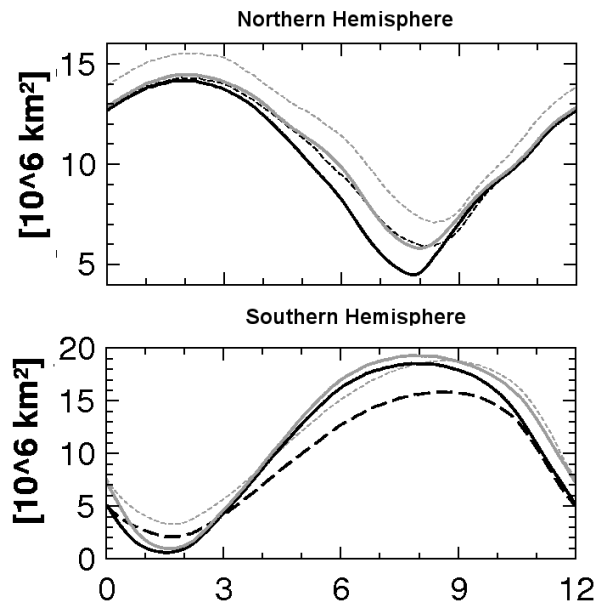


Figure 6.3: Mean seasonal cycles of NH and SH sea ice areas (black) and extents (grey) in the NEMO-LIM3 reference simulation (solid) and in observations derived from satellite Bootstrap PMW recordings (dashed; Comiso, 2007). Ice extent is the area enclosed in the 15% concentration contour, while ice area is the total area of ice.

6.4.2.2 Spatial distribution

The ice edge is generally well simulated by the model and the geographical distributions of ice concentration are generally in agreement with observations (Fig. 6.4 and 6.5). There also are notable regional differences. However, the comparison with observations has to be considered with caution since they are subject to significant errors. Kwok (2002), comparing PMW retrievals of ice concentration to synthetic aperture radar (SAR) data (with higher resolution, therefore more accurate), found that the PMW overestimates the winter open water coverage by 3-5 times. In summer, the observed PMW summer ice concentration is subject to large uncertainties due to melt pond effects, reaching up to 15-20 % in the Bootstrap data that are used here Comiso (2007).

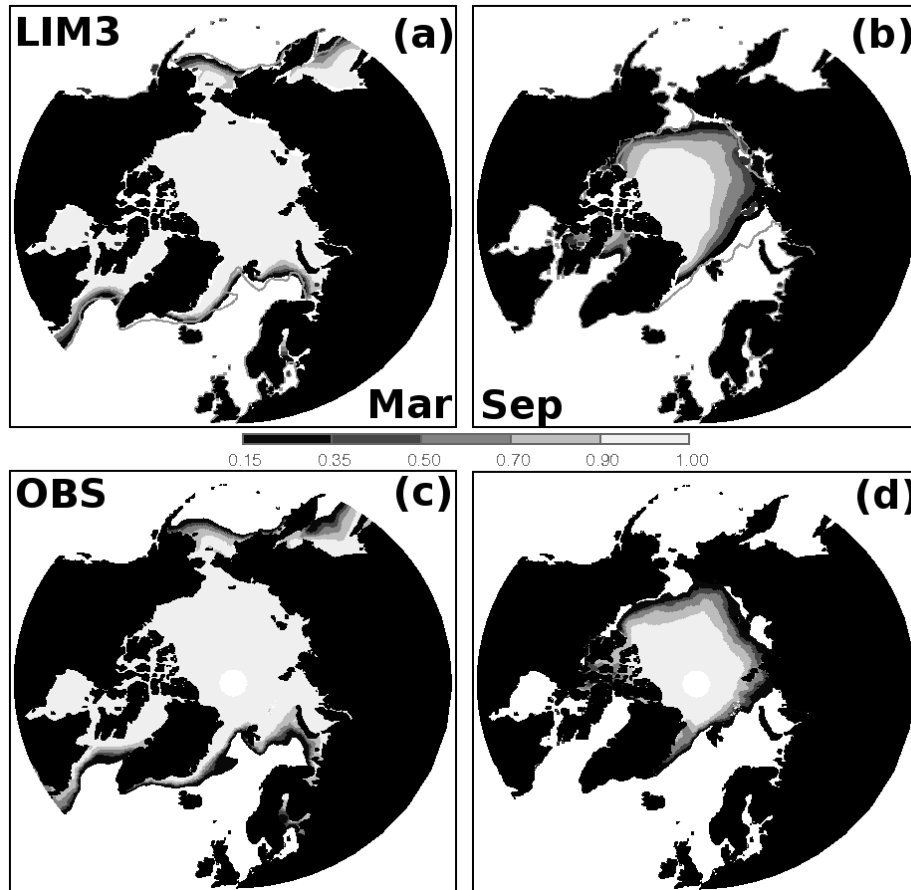


Figure 6.4: Simulated March and September spatial distributions of mean NH (a,b) ice concentrations (1979-2004), indicating the extrema of ice coverage. Observed ice edges (solid grey lines) derived from the PMW Bootstrap dataset (Comiso, 2007) are depicted. The corresponding observed PMW ice concentration data, for the same period and regions, are shown in panels (c,d).

Northern Hemisphere. First, in winter, the simulated NH ice concentrations (see Fig. 6.4) inside the pack are overestimated by a few percent. Nevertheless, the observation error has the same order of magnitude.

Second, the simulated winter width and concentration of the ice stream along the east coast of Greenland is overestimated, as in LIM1 and virtually all AOMIP models. Therefore, this must be due to the absence of specific atmospheric processes in the NCEP/NCAR reanalyses (e.g., katabatic winds) or to a common model feature (ice deformation field, absence of ocean waves). Nevertheless,

compared with LIM1, the ice-covered region in LIM3 is narrower and the ice pile-up in Denmark Strait disappears. This seems to be due to the introduction of the ITD. The transition from VP to EVP minorly affects the results there (S. Bouillon, personal communication). In ITD models, in the marginal ice zones, during early summer, thin ice disappears fast, which leads to lower ice concentrations than in single-category models, promotes higher absorption of shortwave radiation in the ocean and intensifies the ice bottom melt. This more realistic ice-albedo feedback was also found in the global climate model simulations of Holland *et al.* (2006b). Here, the ice-ocean part of this mechanism is included.

In addition, as in LIM1, the Odden/Nordbukta structure is absent. Therefore, following Timmermann *et al.* (2005), we argue that the Odden feature would only cover 3-4 grid cells in zonal extent, so that none of the hypothesized mechanisms for Odden formation (e.g., cold air outbursts, instabilities of the East Greenland Current) is covered in enough detail. This is supported by AOMIP results, which indicate that models with sufficient spatial resolution and at least viscous-plastic dynamics⁴ manage to capture the Odden ice tongue.

In summer, the simulated geographical distribution of ice concentration features high concentrations in the central Arctic and a sharp decrease toward lower latitudes. This is more realistic than LIM1 and most AOMIP models, which underestimate (overestimate) the ice concentration in the central Arctic (in the Siberian seas). Over the three AOMIP models including an ITD, two of them show a similar, realistic latitudinal decrease in ice concentration⁵. Yet this latitudinal decrease is not sharp enough in LIM3. LIM3, as LIM1 and more than half of the AOMIP models, including the most realistic ones, does not reproduce the observed summer ice retreat in the Laptev and East Siberian Seas. This suggests that the ice models miss some important processes in the shelf regions (tides, river runoff, ...). In addition, the ice retreats too far in the northeastern Kara Sea, and the ice concentration is underestimated in the Atlantic sector of the Arctic Ocean (i.e., between Spitsbergen, the North Pole and Severnaya Zemlya). This is associated with an underestimation of ice thickness in those areas (see later on). Finally, ice unrealistically subsists in summer in the northern Baffin and Hudson Bays, as in LIM1 and in some AOMIP models. In these regions, there are numerous mesoscale features which cannot be resolved by our model.

Southern Hemisphere. In the SH, as in LIM1, the location of the simulated winter ice edge is slightly too far northward off East Antarctica and in the Amundsen and Bellingshausen seas (Fig. 6.5). In contrast to this, the winter ice extent is slightly underestimated in the Weddell and Ross seas. Northward, the ice extent

⁴e.g., the UW and NPS models.

⁵The LANL and RAS models.

is limited by the westerly winds and the position of the Antarctic Circumpolar Current (ACC) which, even in winter, has surface temperatures above the freezing point and acts as a permanent heat source. Its course is largely determined by bottom topography which, especially in a z-coordinate model at 2° mean resolution, does not necessarily capture all details of the real bathymetry.

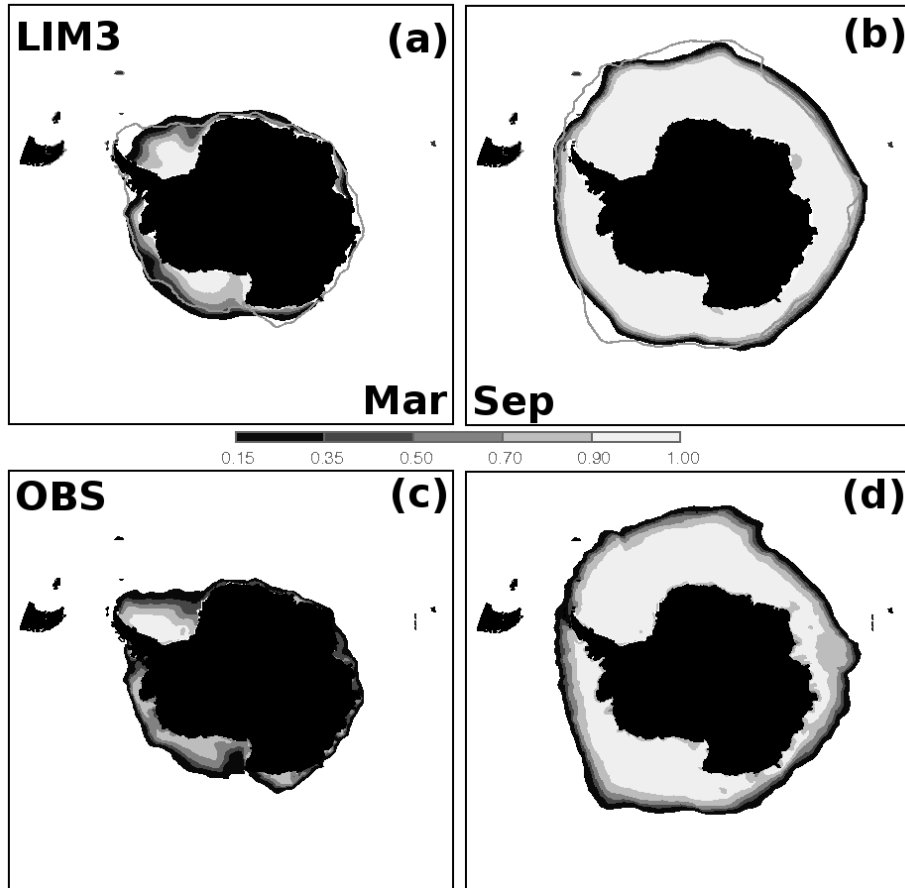


Figure 6.5: Same as Figure 6.4, but for the SH.

The winter ice concentrations are overall overestimated in the SH. In addition, the width of the marginal ice zone is underestimated. This differs from LIM1. In LIM1, the high h^0 prevents the fast closure of open water and regulates the fall advance of the ice edge. We will come back to this later on.

In summer, the simulated sea ice distribution shows a lack of sea ice near the tip of the Antarctic Peninsula and along the coast of East Antarctica. Following

Timmermann *et al.* (2005), we attribute the first mainly to a poor representation of the Antarctic Peninsula in the atmospheric forcing fields and the latter to the model's relatively coarse resolution and inability to simulate the landfast ice along Antarctica, especially between 90 and 180°E. The width of the landfast ice band is on average 45-65 km but can reach 270 km in some regions, as pointed out by Russian long-term monitoring in the second half of the 20th century (Fedotov *et al.*, 1998).

6.4.2.3 Interannual variations

Considering the model annual mean ice extent, for the 1979-2006 period, we find, in the NH, a linear trend of $-0.30 \pm 0.006 \times 10^6 \text{ km}^2/\text{decade}$ (obs. $-0.33 \pm 0.07 \times 10^6 \text{ km}^2/\text{decade}$, 1979-2005, Lemke *et al.*, 2007). In the SH, there is an increasing trend in the model of $0.25 \pm 0.007 \times 10^6 \text{ km}^2/\text{decade}$ (obs. $0.06 \pm 0.09 \times 10^6 \text{ km}^2/\text{decade}$). Trends for minimal extent are the largest in both hemispheres. See Table 6.4 for more information.

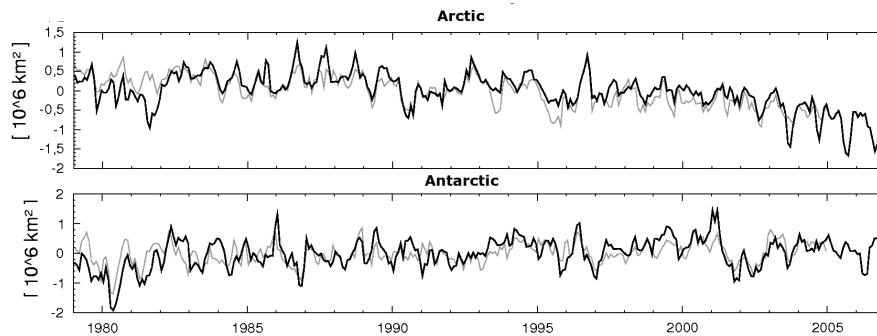


Figure 6.6: Time series of monthly mean anomalies of sea ice area for the NH (a) and SH (b) as simulated by the model (black) and in observations derived from satellite PMW recordings (grey; Comiso, 2007).

In the NH, the time series of monthly mean ice area anomalies (i.e., differences of monthly means from the mean seasonal cycle, Fig. 6.6) are well represented by the model (correlation with obs.= 0.74). The simulated area anomalies feature, first, an increase until 1988, then a series of alternating maxima and minima and, finally, a gradual decrease until 2006. In the 2000s, the model simulates the observed series of lower than average minima of area Stroeve *et al.* (2005). The typical spatial distribution of 2000s concentration anomalies (see Fig.6.7, b and d) indicates that the negative concentration anomalies, observed in the East Siberian and Beaufort Seas, are slightly shifted toward the Siberian shelf.

Some years are not well simulated. First, in 1980 and 1981, the NCEP/NCAR winter air temperatures are warmer than observations from the Russian meteorological stations located on the Siberian Coast (B. Tartinville et al., unpublished manuscript). Spurious minima of NH ice extent are simulated accordingly during these two years. Second, in the 1990s, the model overestimates the 1993 and 1995 minima. Spatial distributions of concentration anomalies (see Fig. 6.7, a and c) show that the observed retreat is not strong enough on the Siberian shelf.

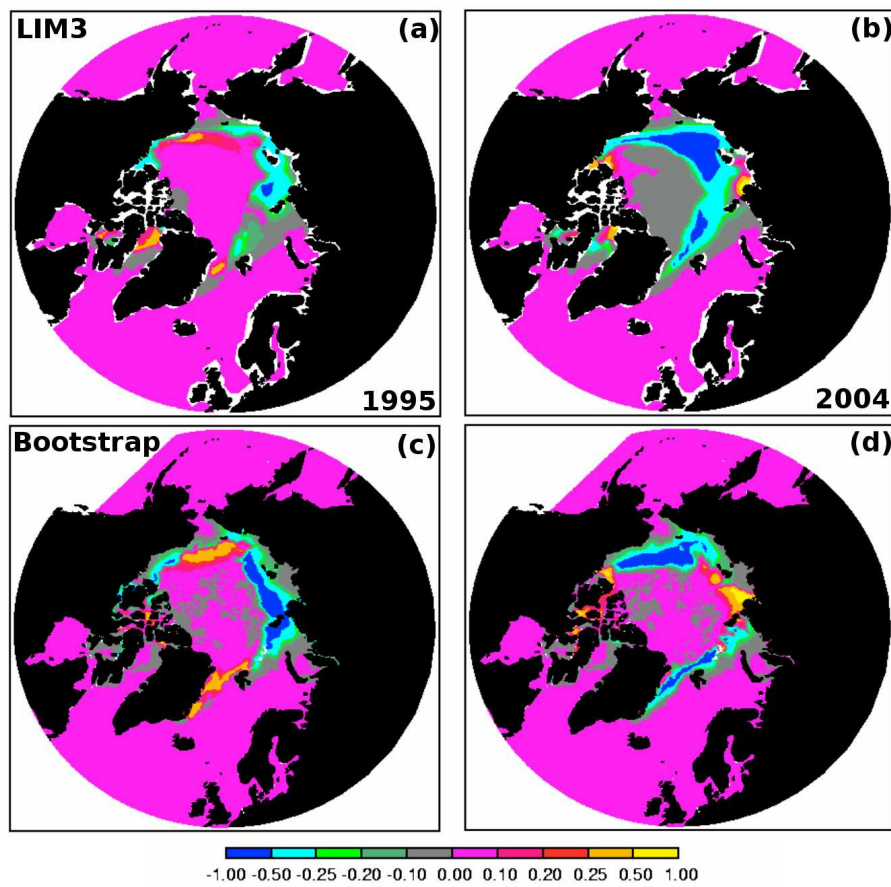


Figure 6.7: Simulated September ice concentration anomalies for two years in the NH (1995, a; 2004, b). (c-d) Same, but from PMW Bootstrap observations (Comiso, 2007). Anomalies are computed using 1979-2004 as the reference period.

In the SH, the simulated interannual variations are larger than in the Arctic and generally well captured by the model (correlation with obs. 0.65). In winter, their spatial distribution is usually correct, though their amplitude is generally underes-

timated, presumably because atmosphere-ice/ocean feedbacks are not included in the model. In summer, their location is correct, except where the ice retreats too far in the model (e.g., in the central Weddell Sea in 1986, see Fig. 6.8). In particular, in 2001, the simulated ice retreat is too slow. This triggers a high anomaly discrepancy between model results and observations.

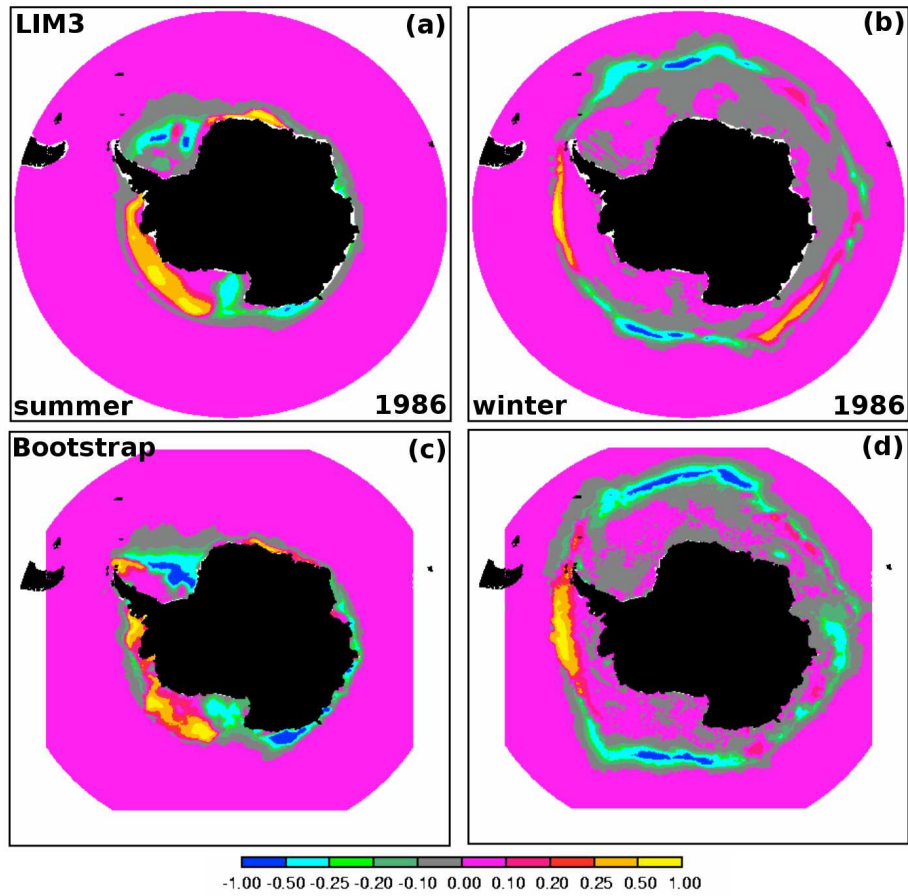


Figure 6.8: Simulated September ice concentration anomalies for September (a) and March (b) 1986 in the SH. (c,d) Same, but from PMW Bootstrap observations (Comiso, 2007). Anomalies are computed using 1979-2004 as the reference period.

6.4.3 Ice thickness and volume

6.4.3.1 Ice thickness

The model 1979-2006 NH (SH) mean ice thickness equals 1.90 ± 0.25 (0.41 ± 0.01) m. The interannual variability in ice thickness (STD equals 12% of the mean) is higher than for ice area (3%). In the SH, the ice thickness variability is not as high (4%).

Northern Hemisphere. Simulated winter sea ice thickness features a gradient of ice thickness from 0-2 m on the west Siberian shelf to 3.5 m at the North Pole and 5-6 m maximum off Canada (see Fig. 6.9), and is consistently thin (0-1.5 m) in the seasonal ice zone. Qualitative comparison to the Bourke and Garrett (1987) 1960-1982 ULS-derived climatology indicates a good model representation of the large-scale geographical ice distribution. Qualitative comparison to the October-March 1993-2001 ice thickness data derived from satellite altimetry (Laxon *et al.*, 2003) shows also good agreement. The largest model-data differences (estimated visually) are negative and located in the northern Barents Sea ($\Delta = -1.5$ m) and along the east coast of Greenland ($\Delta = -2$ to -3 m). They indicate an underestimation of the volume of ice exported out of the Arctic Ocean through Fram Strait and Barents Sea.

Quantitative comparison to the submarine ULS ice drafts (Wensnahan *et al.* (2007); see Fig. 6.9b) confirms and refines the latter underestimation. The observed ice draft, averaged over 3478 50 km-long transects from 37 cruises spanning the years 1976-2000, were used. Observed ice drafts were converted into thicknesses using the snow depth climatology of Warren *et al.* (1999). The results indicate an average model underestimation of ice thickness ($\Delta = -0.55 \pm 1.04$ m). First, the temporal evolution of the annual mean model-ULS difference shows no trend over the years. Most of the years, the absolute error is small and comprised between -0.8 and 0.2 m (average -0.41 m). From 1981 to 1983, the model error is larger (from -0.88 to -2.22 m) because of the error in the NCEP/NCAR temperatures (see earlier). In 1989 and 1998, the model underestimation is also larger ($\Delta = -1.5$ m), in particular close to the North Pole. Second, averaged on a monthly basis, independently of the year of observation, and looking at the seasonal perspective, the model-ULS difference is clearly larger in summer ($\Delta = -1.54$ m in September) than in winter ($\Delta = -0.06$ m in April-May). Finally, the spatial distribution of the model-ULS differences shows a maximum in the Beaufort Gyre and a minimum in the Atlantic sector of the Arctic Ocean between Spitzbergen and the North Pole. This pattern is stable over the years and is

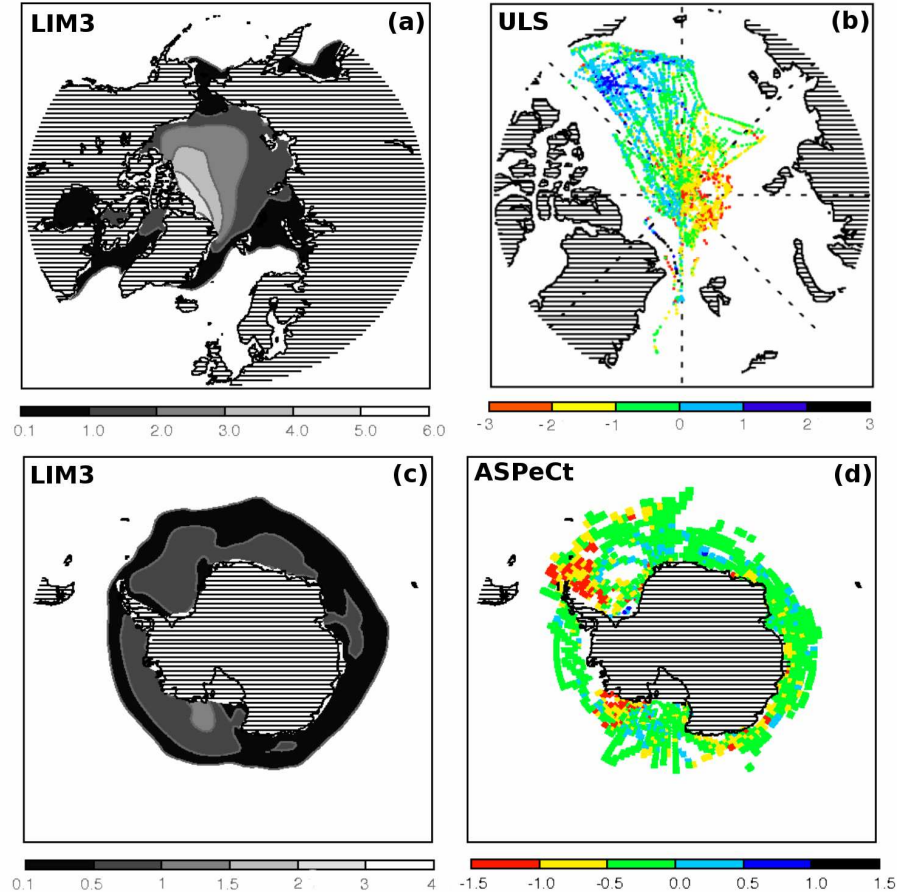


Figure 6.9: Simulated geographical distribution of ice thickness (m), in the NH (January, a) and SH (September, c). On plate (b), the differences between model and ULS data in the NH are shown. The differences are computed using ULS draft values, averaged over 3478 transects spanning 37 submarine cruises between years 1976-2000 (NSIDC, 2006). Drafts are converted into thicknesses using the snow depth climatology of Warren et al. (1999). Each ULS thickness is compared to the monthly mean 1979-2006 model value which month and grid coordinates are chosen as the closest to the actual ULS observation. Since ULS averages include open water, the model ice thickness used accounts also for open water fraction. On plate (d), the differences between model and ship-based ASPeCt observation dataset (Worby et al., 2007) are shown. The differences are computed by first gridding the ASPeCt observations to the ORCA2 grid then by computing the average months of observation and, finally, by comparing the resulting value to a monthly mean 1979-2006 model value which month and grid coordinates are chosen as the closest to the corresponding ASPeCt observation.

remarkably similar to results of simulations performed with other multi-thickness sea ice models (see Rothrock *et al.* (2003), Fig. 9; Wilchinsky *et al.* (2006), Fig. 7). Therefore, it must be due to a common model feature. The error magnitude, typically comprised between -2 and 2 m and also consistent over the different models, suggests that the origin of the error lies in the model dynamic component. Wilchinsky *et al.* (2006) proposed a modification of the yield curve which takes sliding friction into account. This model change has a better physical basis and also reduces model error.

Compared to LIM1, LIM3 has a more realistic spatial distribution of ice thickness. The Siberia-Canada thickness gradient is larger. Ice is thinner in the Barents, Kara and Laptev Seas ($\Delta = -1$ m), in the East Siberian Sea ($\Delta = -0.5$ m) and in the Beaufort and Chukchi Seas ($\Delta = -1$ m). This is due to, first, lower h^0 , which promotes younger new ice and second, to more intense summer melt due to the stronger ITD-associated ice-albedo feedback. In addition, ice is thicker along Canada ($\Delta = 0-1.5$ m).

Southern Hemisphere. The simulated geographical sea ice thickness distribution in the SH has a typical winter sea ice thickness between 0.5 and 1 m, increasing southwards, with a local maximum of 1.5 m in the Ross Sea. Along the east coast of East Antarctica and in the Ross Sea, the model features a series of polynyas locally reducing ice thickness

In the SH, the data are even more sparse than in the Arctic. The only ice thickness dataset in the Southern Ocean is ASPeCt (Antarctic Sea ice Processes and Climate, see Worby *et al.* (2008)), which is a compilation of visual ship-based observations of ice thickness, covering almost all the Southern Ocean but with some potential biases due to preferential ship track crossing thin ice and to varying date of observation. Comparison of model results to the ASPeCt dataset shows reasonable agreement, though model ice thickness is underestimated. The typical first-year (FY) ice thickness is underestimated by 0.-0.5 m. In older-ice areas, the ice thickness is further underestimated: along the East Coast of the Peninsula (1.5 m), along the West Pacific Coast of Antarctica (0.5 m) and in the Amundsen and Ross Seas (0.5-1.5 m).

We attribute the underestimation of typical FY thickness to the faster than observed advance of the ice edge and the related overestimation in closing rates due to the small h^0 . Comparison to LIM1 supports this. LIM1 has a higher h_0 than LIM3. Therefore, in open water, for a given heat loss, thicker ice, with smaller coverage is created, which maintains high oceanic heat loss during longer periods and ice production. As a result, LIM1 has thicker and less concentrated

ice than LIM3. In LIM3, the thinner than real ice melts too fast in summer, which prevents the maintenance of a thick MY ice cover in coastal areas. Along the Antarctic Peninsula, following Timmermann *et al.* (2005), we argue that the further overestimation of the ice thickness is due to the excessive melting due to a warm bias in the forcing data.

6.4.3.2 Ice volume

The ice volume is the domain integral of the product of ice thickness by ice area. In the NH, on average, the ice volume has its maximum at $25 \times 10^3 \text{ km}^3$ at the beginning of April and reaches its minimum in mid-August at $11.8 \times 10^3 \text{ km}^3$, which lies in the the range of the different available simulations over the same period (see, e.g., Rothrock *et al.* (2003); Gerdes and Köberle (2007)). Interannual variations are significant. First, from 1979 to 1982, an initial decrease in annual mean NH ice volume ($-2.03 \times 10^3 \text{ km}^3/\text{y}$) is simulated. Nevertheless, the latter may well be not realistic since the NCEP/NCAR air temperatures are warmer than observed during that period (see earlier). Then, from 1983 to 1987, the simulated NH annual mean volume increases at a rate of $1.04 \times 10^3 \text{ km}^3/\text{y}$, reaching $29.4 \times 10^3 \text{ km}^3$ in 1989. Finally, from 1988 to 2006, the NH annual mean volume decreases at a rate of $-0.46 \times 10^3 \text{ km}^3/\text{y}$. In the SH, the ice volume on average reaches its maximum in early September at $8.8 \times 10^3 \text{ km}^3$ and then decreases to 300 km^3 at the end of January. The annual SH ice volume shows no trend.

6.4.3.3 Sources and sinks of ice mass

Northern Hemisphere. The different components of the simulated sea ice mass balance differ from one hemisphere to the other (see Fig. 6.10). In the NH, averaged over the 1979-2006 period, annual freezing at the ice bottom contributes by 89% to the total annual growth. New ice formation in open water only contributes slightly (10%), while snow ice formation is negligible (0.1%). Analyzing the structure of ice cores extracted from ice exiting the Arctic basin through Fram Strait, Tucker *et al.* (1987) found 90% of columnar ice (indicating bottom congelation process), in agreement with our simulations. Simulated annual bottom melt contributes by 59% to the total, which is slightly higher than surface melt (41%), especially due to the thinner ice categories. This is comparable to

the SHEBA⁶ station observations, where roughly equivalent contributions from bottom and surface melt were observed (Perovich *et al.*, 2003).

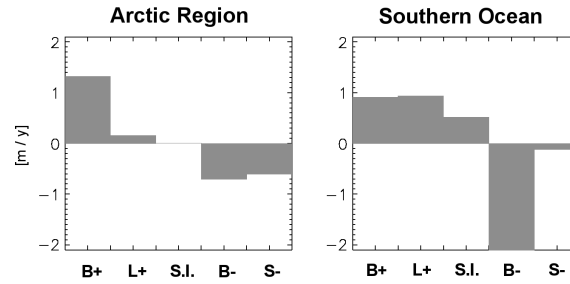


Figure 6.10: Annual 1979-2006 averages of sources terms of ice volume per unit area due to different mechanisms : bottom growth (B+), new ice growth in open water (L+), snow ice formation (S.I.), bottom melt (B-) and surface melt (S-) for the Arctic regions (including Arctic Ocean and surrounding continental seas, left) and the whole Southern Ocean (right).

Southern Hemisphere. In the SH, the simulated bottom congelation is important (37% of the annual volume created). Nevertheless, in contrast to the Arctic, new ice formation in open water (38%, mostly in fall) and snow ice growth (25%, mostly in spring) have a significant impact on the ice mass budget. Ice core structural analyses by Jeffries *et al.* (1997); Worby *et al.* (1998), performed in-situ in the Ross and Amundsen Seas and in the east Antarctic region, respectively, found 26% (resp. 39%) of columnar ice, 44% (resp. 47%) of granular ice, which indicates frazil ice formation in turbulent waters, and 24% (resp. 13%) of granular ice with snow ice origin. This suggests that our model probably overestimates the bottom growth compared to frazil ice formation in the Southern Ocean. The simulated large oceanic heat flux (10-40 W/m²) induces that bottom melt tremendously dominates the decay of the ice cover (95 % of the total), while surface melt is practically absent, which corresponds to available observations (Andreas and Ackley, 1982; McPhee *et al.*, 1998).

The underestimation of creation of new ice in open water confirms that the SH dynamic processes of ice growth are not accurately represented in the model. In the SH, new ice is known to grow in wavy, turbulent waters, in what is called the pancake cycle (Lange *et al.*, 1989). During this process, significant open water areas are created and destroyed at very short time scales by wave-ice interactions.

⁶The SHEBA (Surface Heat Budget of the Arctic Ocean) was a yearlong drift station in the Beaufort Sea between October 1997 to September 1998.

6.4.3.4 Ice thickness distribution

We compared the simulated ITD with 4 sets of observed ITD from both hemispheres (see Fig. 6.11, a-d). Observations come from different sources and techniques : (1) Witte and Fahrbach (2005), moored ULS in the Fram Strait region; (2) Haas (2004), electromagnetic sounding and drilling in the Transpolar Drift region; (3) Worby *et al.* (1996), drilling and ship observations in the Amundsen and Bellingshausen seas; (c) Lange and Eicken (1991), drilling in the northwestern Weddell Sea.

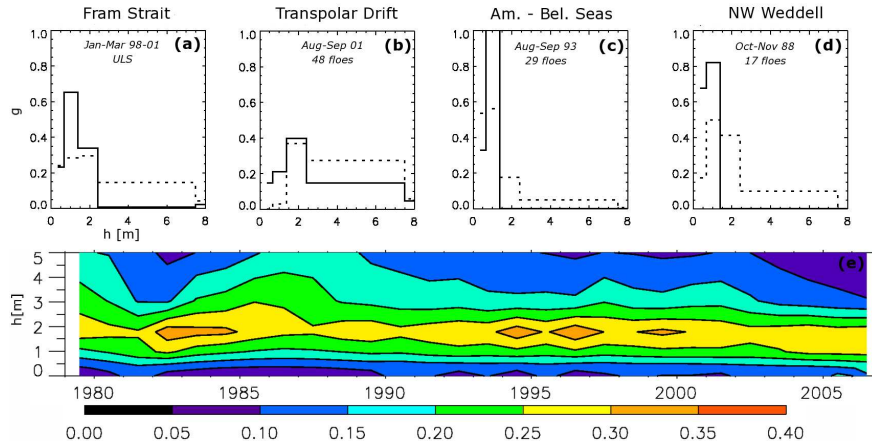


Figure 6.11: (a-d) Comparison of simulated (solid) and observed in-situ (dashed) ITD, excluding open water. Observations come from (a) Witte and Fahrbach (2005), moored ULS (b) Haas (2004), electromagnetic sounding and drilling (c) Worby *et al.* (1996), drilling and ship observations and (d) Lange and Eicken (1991), drilling. The simulated distribution is a 2-month mean taken from the closest grid point, at the corresponding time of the year and year. (e) Simulated Arctic (including Arctic Ocean, Beaufort, Chukchi and all Siberian shelf seas) annual mean ice thickness distribution, including open water, 1979-2006 (contours).

The simulated thickness distribution follows the commonly observed features. NH ITD is a mixture of level FY and MY ice as well as thick ridged ice. In relation to the mean ice thickness underestimation discussed earlier, the model ITD features an excess of thin ice and a lack of thick ice. In the SH, the most representative ice type is level FY ice. Nevertheless, the simulated SH ITD has a lack of thick MY and ridged ice.

The interannual variations in NH annual mean ITD (panel e) are quite significant. In the 1980s, until 1987, the modal thickness as well as the relative contribution of thick categories increase. In 1988, the coverage of the thickest ice decreases and,

in 1989, the modal ice thickness also starts to decrease, which, as pointed out by several authors (see, e.g., Lindsay and Zhang (2005)), was triggered by a switch towards positive Arctic Oscillation (AO) index. Defined as the first the leading EOF of sea-level pressure (SLP) above 60°N, the AO is the dominant pattern of non-seasonal SLP variations in the Arctic Thompson and Wallace (1998), which affects the large-scale ice motion in the Arctic basin. Weaker (stronger) Beaufort Gyre and higher (lower) than normal values of ice export through Fram Strait are associated to high (low) values of AO index, and results in low coverage over the Siberian shelf seas. Higher AO index in the 1990s resulted in an increase in the flushing of old, thick ice out of the Arctic Basin through Fram Strait. From 1996 on, the AO index recovered back to more normal values but the thinning continued. As hypothesized by Lindsay and Zhang (2005), the Arctic sea ice cover seems to have reached a tipping point.

6.4.4 Ice drift

6.4.4.1 Drift patterns

The large-scale simulated sea ice drift (see Fig. 6.12) reflects the general patterns of atmospheric circulation. Ice motion is more intense in the SH, due to higher winds and lower continental constraints. The 1979-2006 average simulated ice velocity for the NH (SH) is 5.2 (8.7) cm/s, which corresponds well to satellite observations of ice motion (Martin and Gerdes, 2007; Heil *et al.*, 2006). The standard deviation is higher in the NH (34% of the mean) than in the SH (19%). From 1979 to 1986, the NH simulated ice velocity decreases by 0.8 cm/s per decade, and then increases until 2006 at a rate of 0.5 cm/s per decade. No remarkable evolution is found in the SH.

The drift patterns are very similar to those shown by Timmermann *et al.* (2005) and Bouillon *et al.* (2008) (using LIM1 with VP and EVP respectively) and compare to the observations of Fowler (2003). In the NH, the mean ice drift field shows the divergent motion over the Siberian shelf (1-2 cm/s), the Beaufort Gyre (1-3 cm/s), the Transpolar Drift Stream (1-4 cm/s) and the increased ice drift across Fram Strait (4-8 cm/s) and along the east coast of Greenland (5-10 cm/s); as well as large nearly motionless ice zone northward of the Canadian Archipelago.

In the SH, the model drift features the coastal (3-13 cm/s) and offshore circumpolar (1-6 cm/s) currents, as well as the Weddell and Ross Gyres. Spurious

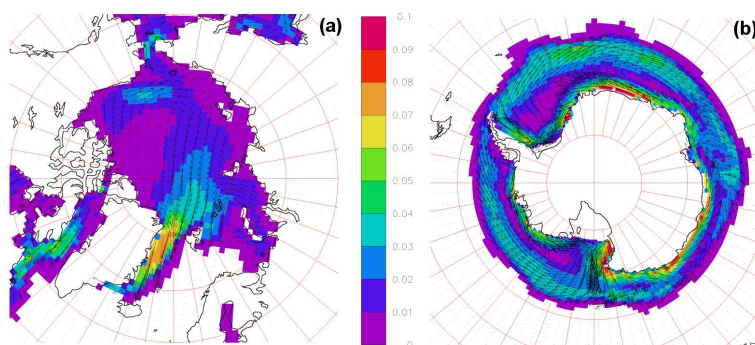


Figure 6.12: Simulated mean (1979-2006) ice velocity in the NH (a) and in the SH (b).

maxima of ice velocity are found along the coastal areas of the Indian (4-10 cm/s) and West Pacific (5-10 cm/s) sectors of the Southern Ocean, as well as in the Western Ross Sea (5-10 cm/s), where satellite observations show that ice is almost at rest, which is supported by in-situ observation of large bands of landfast ice in the West Pacific sector Fedotov *et al.* (1998). The position of the transition zone between coastal and offshore currents is well captured between 67 and 72° S. A more quantitative analysis of sea ice drift is done in Bouillon *et al.* (2008).

6.4.4.2 Fram Strait export

Fram Strait, situated between Greenland and Svalbard, is the main output gate for sea ice out of the Arctic Ocean and has therefore an important impact on the NH ice volume. The simulated annual areal export through Fram Strait is 0.9×10^6 km², i.e., 12 % of the Arctic Ocean area, which compares well to the data of Kwok *et al.* (2004) and the 13% in the Lindsay and Zhang (2005) simulation (Fig. 6.13). The seasonal variations (with high winter and low summer exports) also follow the observations. The area flux has no significant trend over the simulation. The simulated volume flux has an average value of 1.2×10^3 km³. This value seems underestimated by 30% compared to Kwok *et al.* (2004) data (combining radar and ULS information) and Lindsay and Zhang (2005) simulation. This indicates that the exported ice (i.e., ice in the Atlantic sector of the Arctic Ocean) is too thin, as previously mentioned in the analysis of the ice thickness fields north of Greenland. LIM1 overestimates ice export by about 50%, because it also overestimates ice thickness in the NH. The volume export

has larger interannual variations than the areal export. As in Lindsay and Zhang (2005), the winter volume flux has maxima in 1989 and 1990, which is caused more by the large mean thickness than by a large area flux. The winter volume flux declines in the late 1990s because the mean exported thickness declines.

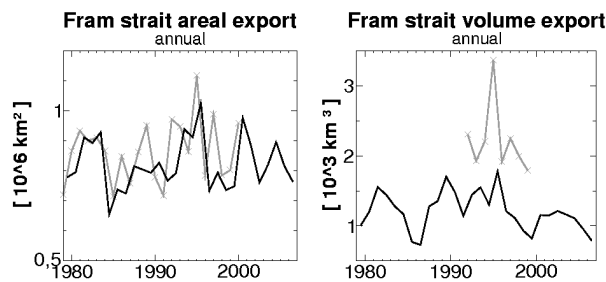


Figure 6.13: Simulated (black) and observed (gray) time series of annual mean sea ice areal (left) and volume (right) exports through Fram Strait. Observations combine recordings from synthetic aperture radar and moored ULS data (Kwok *et al.*, 2004).

6.4.5 Snow volume and depth

In spite of its important impact on surface heat fluxes (Fichefet *et al.*, 2000), the mass of snow on sea ice and its interannual variations are not well known, and this for several reasons. First, large-scale observations of snow depth are recent and the knowledge of the sources of error is not yet optimal. Second, the observation of snowfall is extremely difficult to carry out accurately. In particular, precipitation gauges underestimate snowfall significantly, by wind undercatch, wetting and evaporation losses (Goodison *et al.*, 1998). In addition, the link between precipitation and snow depth is not yet understood. In particular, the role of wind on snow resuspension and deposition on a variable thickness ice cover is poorly documented and therefore neglected in models (Massom *et al.*, 2001).

Northern Hemisphere. Averaged over 1979-2006, the simulated NH snow volume peaks at $2.8 \times 10^3 \text{ km}^3$ in early April and falls to zero in mid-July. The simulated snow depth is around 30 cm in the Central Arctic, which corresponds to the climatology of Warren *et al.* (1999) from Russian drift stations. In the Arctic, unfortunately, no satellite snow depth record exists, because the snow microwave signal is not discriminable from the one of MY ice (Markus and Cavalieri, 1998).

Southern Hemisphere. In the SH, algorithms allowing snow depth retrieval from PMW recordings have been developed (Markus and Cavalieri, 1998, 2006), though their level of validation is not as high as for ice concentration and the magnitude of errors is unknown. Averaged over 1979-2006, the model (observed) snow volume oscillates between $0.12 (0.62) \times 10^3 \text{ km}^3$ in early February and $2.65 (2.34) \times 10^3 \text{ km}^3$ in early September. In winter, the maximum snow volume occurs 1 month later than observed.

Attributing errors would require an extensive study. The small winter overestimation may be linked to the absence of the model representation of snow metamorphism, in particular compaction (Sturm *et al.*, 2002), or blowing snow and blowing snow sublimation effects (Déry and Tremblay, 2004), or both. It can also be linked to the precipitation forcing which is poorly constrained by observations. The summer underestimation has to be interpreted with even more caution since data are subject to more errors in summer. First, the underestimation of ice cover induces a smaller surface for snow. Second, this underestimation is consistent with the fact that 1-D models including a representation of snow cover similar to LIM3 prove to melt snow too quickly. As pointed by 1-D studies (Nicolaus, 2006; Vancoppenolle *et al.*, 2007), improving the model snow physics (e.g., snow metamorphism, radiative effects and thermodynamics) would likely give smaller snowmelt.

The spatial distribution (Fig. 6.14) of snow depth compares quite well with observations. The negative northward gradient and the regional accumulations (i.e., the maxima in the Weddell Sea, Amundsen and Bellingshausen seas, and the minimum in the Ross Sea) are well simulated. The lack of accuracy in the exact location of the maxima and in their amplitude probably reflects (1) the lack of a stable MY ice cover along the Antarctic Peninsula in the model (see above), (2) the absence of redistributing wind effects and (3) errors in the forcing.

6.4.6 Sea ice age

6.4.6.1 Hemispheric mean values and interannual variations

Fig. 6.15 (panel a) shows time series and spatial distributions (panels b-c) of the simulated mean ice age (i.e., averaged over categories) in both hemispheres. Ice is older in the NH (average 2.6 y) than in the SH (0.3 y), which well represents the higher prevalence of MY in the NH (Cavalieri *et al.*, 1997; Kwok, 2004). The mean ice age seasonal cycle is characterized by a difference between maximum

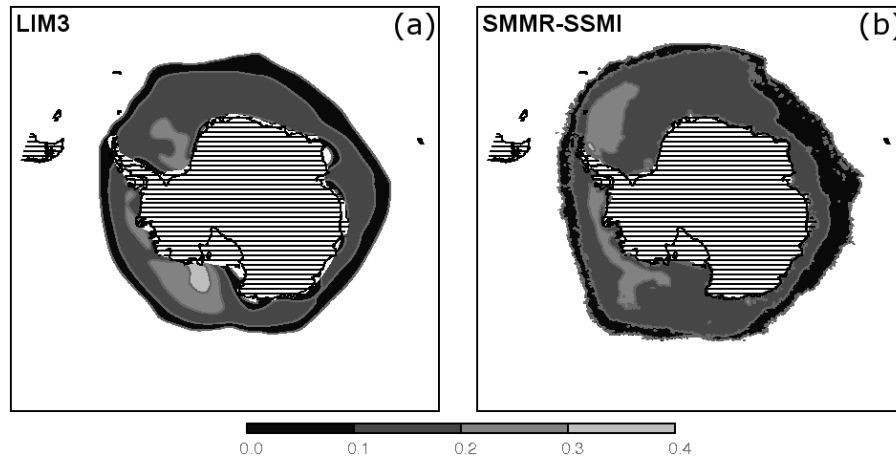


Figure 6.14: Simulated (a) and observed (b) spatial distributions (September) of mean snow depth (m) in the SH. Observations are derived from satellite passive microwave recordings (Markus and Cavalieri, 2006). The means, computed over 1979-2004, take into account the weighted contributions of the ice thickness categories, but exclude open water.

equal to 1.91 (0.84) y in the NH (SH). The summer maximum, reached on Aug 23 (Feb 18) in the NH (SH), reflects the higher concentration of old ice. The time of the minimum, reached on Feb 27 (Apr 14) in the NH (SH), reflects the time of the highest coverage of new ice.

Due to the sea ice age definition, the interannual variations in sea ice age mostly reflect the variations in ice area. The correlation between mean ice age and total ice area is 0.72 (0.58) in the NH (SH), but it is only -0.26 (-0.39) for ice volume. The NH sea ice age increases during the 1980s until 1989, remains stable until 1995, and then decreases monotonically, as also shown by observation-based computations of Rigor and Wallace (2004). The trend in ice age over the 1979-2006 period is significant in the NH with -0.18 years / decade. The decrease is more pronounced for the annual minimum, which indicates an increasing winter contribution of young ice to the overall ice pack. No such trend is found in the SH.

6.4.6.2 Geographical distribution.

Northern Hemisphere. In the NH, the simulated geographical distribution of ice age indicates FY ice (0-1 y) in the Hudson and Baffin Bays and in the East

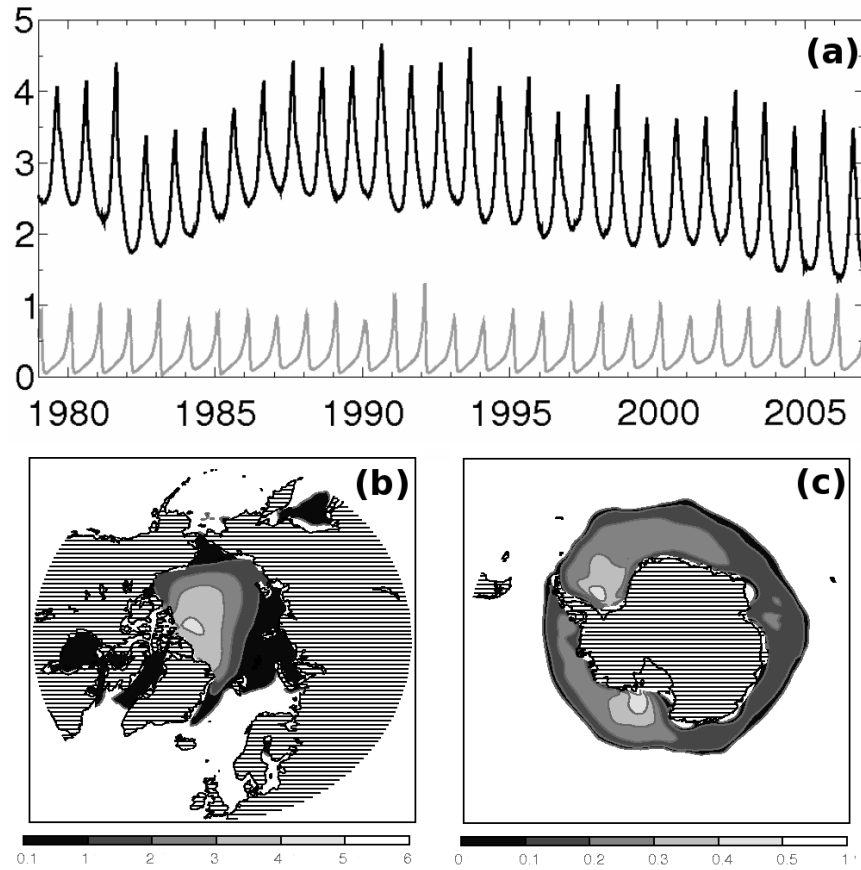


Figure 6.15: Sea ice age (years), area-weighted average over all categories, excluding open water. (a) Simulated time series of global hemispheric means for the NH (black) and SH (gray). (b,c) Spatial distribution for the NH (March, b, years) and the SH (September, c, months). The observed MY ice edge derived from neural network analysis of passive microwave data (Belchansky, 2005a) is drawn in white on (b) panel.

Greenland, Barents, Kara, Chukchi and Okhotsk seas. The intermediate ages (1-3 y) that are found in Laptev and Beaufort Seas, as well as in the Eurasian basin of the Arctic Ocean reflect a mixture of young, thin, FY ice and old, thick, MY ice. In the Canadian basin, the age of thick, deformed sea ice reaches 4-5 years. The average over ice categories hides the variations of age among ice of different thickness. The age of the thickest ice reaches up to 10 years in deformed regions.

The simulated NH geographical distribution of sea ice age qualitatively agrees with available observation-based inferences from PMW ice motion and concentration (Rigor and Wallace, 2004; Belchansky *et al.*, 2005b). First, the computed MY ice edge compares well with the 15% MY ice concentration contour of PMW neural network-derived data (Belchansky *et al.*, 2005a). Nevertheless, the model seems to underestimate MY ice in the Atlantic sector of the Arctic Ocean, i.e., where the model underestimates ice thickness. Second, while the location of the simulated maximum is the same as in observations, its magnitude (4-6 y) is much smaller. The observed maximum is larger than 10 years, and reaches up to 20 years in some cases. Observation-based inferences only take ice transport into account. In the model, three mechanisms decrease the mean and maximal sea ice age compared to satellite-derived values. First, new ice growth inside the pack adds areas of age zero. Second, ice deformation merges the ages of thin deforming ice (young) with thick deformed ice (old). Finally, merging of ice age between neighbour cells due to ice transport diffuses the maxima of ice age. This explains why the observation-based inferences of sea ice age are higher. Finally, the location of the sea ice age maximum occurs near the North Pole in the model, while it is situated in both datasets in the Beaufort Gyre. More work should be done to attribute the origins of these differences.

Southern Hemisphere. In the SH, the simulated ice age rarely exceeds two years, and averaged over 1979-2006, it does not exceed one year. Typically, the sea ice age equals zero at the ice edge and increases southwards. The ice age shows maxima in the Weddell and Ross Seas, where the ice is on average 6 months old. In the Amundsen and Bellingshausen seas, the ice is younger (2-3 months). Associated to the aforementioned underestimation of ice thickness in the Weddell, Amundsen, Bellingshausen and Ross Seas, we believe the sea ice age is also resultingly underestimated. This is corroborated by the observation of second-year or even MY ice in those regions (e.g., Eicken (1998)). The ice is the youngest (0-2 months on average) off East Antarctica. The near zero sea ice age in western Ross Sea and in several places along East Antarctica reveals the presence of simulated polynyas. There is no large-scale sea ice age dataset in the SH.

Despite the hemispheric ice age is not well correlated with the total hemispheric volume, the similarity between age and thickness geographical distributions is striking. Point to point correlations between single-category age and thickness averaged over 1979-2006 in the NH (SH) give correlations of 0.69 (0.52). Computed over smaller time periods, the correlation reaches up to 0.8, but the value of the age-thickness regression coefficient changes with season and from year to year. Therefore, it is possible that satellite estimates of ice age could be a proxy for ice thickness, but this requires a deeper analysis.

6.4.7 Ice salinity

Since the simulated sea ice salinity field is validated and analyzed in detail in the companion paper, we only describe it briefly here. Fig. 6.16 shows the simulated time series of global hemispheric means of ice salinity. Mean ice salinity is 3.59 (resp. 6.29) ‰ in the NH (SH). The sea ice salinity is higher in the SH because of (1) the higher relative contribution of young ice, (2) the much lower frequency of surface melt, and therefore of flushing and (3) the higher contribution of snow ice to the global ice mass balance.

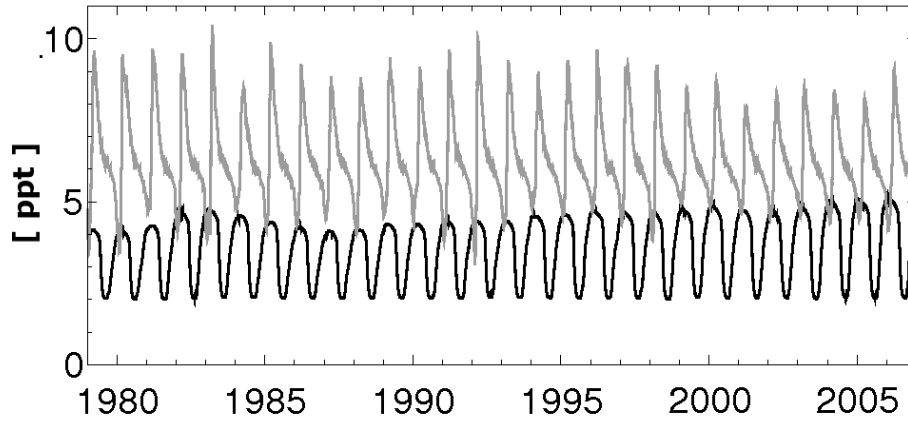


Figure 6.16: Simulated time series of hemispheric mean sea ice salinity (volume-weighted average over all categories, excluding open water) for the NH (black) and SH (gray).

The seasonal cycle of sea ice salinity has an average amplitude of 2.55 ‰ in the NH (4.83 ‰ in the SH), with a maximum in winter (fall) and a minimum in summer (summer). Interannual variability in ice salinity is mostly driven by the maximum value, which reflects the relative contribution of young ice to the total ice volume, itself depending on the amount of summer melt. Salinity and age seem to be highly related to each other, as indicated by the high correlation between the daily mean salinity and age time series of -0.83 (-0.71). The trend in global hemispheric mean ice salinity, governed by the trend in the maximum, is significant only in the NH.

6.5 Discussion and conclusions

We have presented results from LIM3, a new dynamic-thermodynamic sea ice model, including the representation of subgrid-scale distributions of ice thickness, enthalpy, salinity and age. LIM3 is embedded in the ocean modelling system NEMO in a global configuration.

Despite the model is global and the parameter values the same for both hemispheres, the simulated NH and SH sea ice covers are characterized by different physics. In the NH, the simulated sea ice is thicker, older and less saline than in the SH. The ice thickness increases mostly by bottom congelation growth and melts by comparable amounts at the top and bottom interfaces. In the SH, the three different growth mechanisms included in the model (bottom congelation growth, new ice formation in open water areas and snow ice formation) contribute significantly. This north-south contrast in the sources and sinks of ice mass is quite realistic compared to the information from the textural analysis of ice cores.

In the NH, the model provides a very good simulation of the large-scale sea ice features, in line with the most advanced AOMIP models (Gerdes and Köberle, 2007; Johnson *et al.*, 2007). The mean seasonal cycle of ice extent as well as the mean geographical distribution of ice concentration are rather realistic compared to observations derived from satellite PMW recordings. The geographical distribution of ice thickness and concentration is better simulated in LIM3 than in earlier versions of LIM due to a more sophisticated ice-albedo feedback associated to the introduction of the ITD in the model. Local model errors are found in the Odden region in winter and in the Atlantic sector of the Arctic Ocean in summer. The 1979-2006 interannual variations of ice coverage are well captured by the model. The simulated ice thickness generally agrees with ULS observations, though the model overestimates ice thickness in the Beaufort Gyre and underestimates it southwards of the North Pole, which is presumably due to an underestimation of shear strength in the model. The simulated drift is in agreement with PMW inferences of ice motion. The areal export through Fram Strait is well simulated. Nevertheless, as a result of the underestimation of ice thickness around the North Pole, the volume export through Fram Strait is underestimated. Information is too limited to draw any conclusion regarding the simulated snow depth in the Arctic.

In the SH, the seasonal variations of ice extent are rather well represented compared to PMW observations. The winter ice extent is well simulated, though the winter ice area and concentrations are overestimated. In summer, the ice re-

treats too far southward in the model. The simulated geographical distribution of ice thickness resembles the observations, with a tendency towards underestimation, in particular in coastal seas. Compared to PMW observations, snow depth is quite realistic in the model. In addition, the general features of ice drift are well simulated, though the coastal currents are significantly overestimated by the model. The formation of new ice into open water areas is underestimated when compared to the granular ice frequency (i.e., ice originating from open water) in ice cores from several Antarctic subregions. The overestimation of winter ice concentration, the yearlong underestimation of ice thickness and the underestimation of new ice growth in open water all suggest that the representation of dynamical (pancake) processes of ice growth should be improved in the model.

The age of sea ice was a new feature included in the model. The NH FY-MY ice contrast is well simulated. In the SH, the simulated ice age is much younger than in the Arctic and reflects the lack of a model MY ice cover. In the NH, the values and the geographical distribution of sea ice age within the perennial ice zone differ from satellite observation-based inferences. Based on the use of age in marine modelling Deleersnijder *et al.* (2001), we argue that this reflects mostly the fact that the definition of observed and model ice age are different. Nevertheless, this could also indicate model misbehaviour, though this is difficult to disentangle at this stage. Consequently, the simulation of sea ice age deserves an entire study which is beyond the scope of the present paper.

Another new feature included into the model is the sea ice salinity. The simulated ice salinities show strong hemispheric and seasonal differences, which reflect the spatio-temporal variations in the thermodynamic regime and will be investigated in detail in the second part of the present paper.

To conclude, we have shown that LIM3 is suitable for large-scale sea ice and climate simulations. LIM3 results compare well to observations, though some deficiencies and several open questions remain and should be addressed in the future.

6.6 Acknowledgements

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Chapter 7

Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 2. Sensitivity to salinity processes (V08-2)

The text of the present chapter was submitted as as : Vancoppenolle, M., T. Fichefet, H. Goosse (2008c), Simulating the mass balance and salinity of Arctic and Antarctic sea ice. 2. Sensitivity to salinity processes (V08-2), Ocean modelling.

Abstract.

The theory of sea ice thickness distribution is generalized to include the sea ice salinity distribution. The formulation of the thermodynamic changes in ice salinity is done through a simple parameterization of brine entrapment and drainage, which is validated against a more complex sea ice halodynamic scheme in a one-dimensional test case. The salinity distribution is included in LIM3, a three-dimensional dynamic-thermodynamic sea ice model including ice thickness, enthalpy and age distributions, as well as the impact of ice salinity on ice growth and decay. LIM3 is coupled to the ocean model OPA9 and run over 1979–2006, forced by a combination of daily

NCEP/NCAR reanalysis data and various climatologies. The seasonal cycle of the simulated sea ice mean salinity agrees well with available ice core salinity data and its distribution follows the salinity-thickness relations of Cox and Weeks (1974). Due to differences in the physical forcings, the model simulates Arctic and Antarctic sea ice salinity fields that profoundly differ, in agreement with observations. The simulated large-scale sea ice mass balance is found quite sensitive to the representation of ice salinity. In the Arctic, including an interactive salinity distribution enhances ice growth / melt through a direct impact on ice thermodynamics, leading to changes in ice thickness up to one meter in some regions. Around Antarctica, the effect of an interactive salinity distribution affects ice volume up to 28%, mostly through changes in ice-ocean interactions and feedbacks. Using a variable salinity maintains large ice growth rates with relatively small salt fluxes to the Southern Ocean which reduces vertical mixing and, in turn, further enhances ice growth. Given the importance of salinity on the simulated sea ice characteristics, the salinity distribution should be included in assessments of the response of the high-latitude oceans to ongoing and future climate change.

7.1 Introduction

Sea ice is both an actor in and an indicator of climate change (Serreze *et al.*, 2007). It affects the radiative, heat and momentum exchanges between polar atmosphere and ocean (Holland *et al.*, 2006b). Sea ice is also an important habitat for marine organisms which lie at the basis of the polar foodweb (Thomas and Dieckmann, 2002). Its permeability structure and the presence of leads and polynyas modulate the atmosphere-ocean exchanges of gases such as CO₂ (Delille, 2006). Therefore, the role of sea ice in the carbon cycle is considered with increasing attention (e.g., Bates (2006)). The Arctic ice coverage (Cavalieri *et al.*, 2003; Comiso, 2006) and draft (Rothrock *et al.*, 2003) have strongly decreased over the last decades. Climate projections for the next century (e.g., Arzel *et al.* (2006); Holland *et al.* (2006a)) suggest that the decrease will continue in the future leading to a less extensive, thinner and younger Arctic sea ice cover. These changes should in turn feedback on global climate and major biogeochemical cycles. Sea ice models, coupled to atmosphere and ocean general circulation models, are one of the key tools to refine present understanding of recent changes and projections of future state of the sea ice cover.

Sea ice thickness is highly variable (0–20 m), even at subfloe scales, which is important, notably because the ice growth rate and the atmosphere-ocean heat exchanges nonlinearly depend on ice thickness (Maykut, 1986). In large-scale sea ice models, this has been considered through the use of a subgrid-scale ice thickness distribution (ITD) (see, e.g., Bitz *et al.* (2001)). ITD physics combine

numerous complex processes. Ice thermodynamics govern ice growth and melt (Maykut and Untersteiner, 1971) and induce smooth changes in the ITD. Mechanical processes induce sharper transitions in ITD, creating open water under diverging motion, and deforming the weakest thin ice by rafting or piling it into thick pressure ridges under converging motion. The summer ice-albedo feedback critically depends on the melt of thin ice, which can only be represented through the use of an ITD. Hence, resolving the ITD greatly improves the simulated mean sea ice state and sensitivity and, in turn, is required in climate simulations (Holland *et al.*, 2006b).

In addition, sea ice is not totally fresh. Salt cannot lock into the ice crystalline lattice but rather dissolves into liquid inclusions of brine. Brine remains trapped in the ice during its formation and then drains back to the ocean as ice thickens and gets older (Cox and Weeks, 1988; Vancoppenolle *et al.*, 2006). Because of brine drainage, the vertically averaged (hereafter referred to as "bulk") sea ice salinity varies in time (Cox and Weeks, 1974) and also vertically and horizontally. First, driven by differences in ice thickness, snow cover and distribution of brine channels (Cottier *et al.*, 1999), the sea ice bulk salinity exhibits significant horizontal, subfloe-scale variations (Eicken *et al.* (1991); Vancoppenolle *et al.* (2007), hereafter referred to as VBF07). In addition, due to the variable efficacy of brine entrapment and drainage, there are vertical variations in ice salinity (Malmgren, 1927; Schwarzacher, 1959; Nakawo and Sinha, 1981; Eicken, 1992; Jeffries and Porter, 1995). Thirdly, the ice salinity should logically present large-scale geographical variations, but to our knowledge, these have not yet been studied. The global ice cover salt content is determined by the combination of the salinity-thickness relation and the large-scale ITD.

Resolving sea ice salinity variations in large-scale sea ice models could be important, as, in tight association with temperature, it determines the brine-solid ice phase equilibrium (Frankenstein and Garner, 1967), the ice thermal properties (Malmgren, 1927; Pringle *et al.*, 2007) and the permeability transitions (Golden *et al.*, 1998), and affects shortwave radiation scattering (Light *et al.*, 2003) and mechanical behaviour (Timco and Frederking, 1990) in the ice. Finally, salt and fresh water fluxes, respectively associated to ice growth and melt, strongly influence the oceanic density stratification and the World Ocean circulation (Aagaard and Carmack, 1989; Goosse and Fichefet, 1999). Despite this, the role of salinity variations on the evolution of the sea ice cover and on ice-ocean interactions are presently unknown.

The main reason of these gaps in the knowledge related to large-scale sea ice salinity lies in its crude representation in present large-scale three-dimensional (3D) sea ice models. In most models, the ice salinity is considered as a constant,

only used to compute the salt / fresh water flux at the ice-ocean interface (e.g., Fichefet and Maqueda (1997)). In some of them, the impact of sea ice salinity on ice thermodynamics has been considered either through a vertically varying salinity profile, fixed in time (e.g., Bitz *et al.* (2001); Hunke and Lipscomb (2004)), or through an isosaline profile varying in time (e.g., Schramm *et al.* (1997)).

The purpose of this paper is to introduce, validate and analyze the impact of a sea ice salinity distribution formulation in the 3D sea ice model LIM3¹. In the companion paper (Vancoppenolle *et al.* (2008c), this issue, hereafter referred to as V08), the general features of LIM3 are described and the simulated sea ice characteristics are critically compared to observations. Here, we focus on the sea ice salinity aspects. First, we generalize the ITD theory to include the ice salinity distribution. Second, we validate the formulation of the thermodynamically driven changes in ice salinity produced by a simple parameterization and compare its results to the ones of a more complex, one-dimensional (1D), halodynamic (i.e. computing salinity variations) sea ice model. Then, we validate the ice salinity simulated by LIM3 using this parameterization and analyze the resulting geographical salinity distribution. Finally, we study the impact of an interactive salinity distribution on the simulated sea ice characteristics and, briefly, on the upper ocean.

7.2 Ice thickness and salinity distribution theory

As a preliminary step, it is necessary to extend the ITD theory to include the ice salinity. In order to account for unresolvable variations in ice thickness (h) in large-scale sea ice models, Thorndike *et al.* (1975) proposed to introduce an ITD function defined by :

$$\int_{h_1}^{h_2} g(h)dh = \frac{A(h_1, h_2)}{R}, \quad (7.1)$$

where R is the total area of the region of interest, $A(h_1, h_2)$ is the area in R covered by ice with thickness ranging between h_1 and h_2 . Invoking continuity (i.e., assuming that $g(h)$ is valid for any subregion of R), Thorndike *et al.* argued that the conservation of ice mass can be written in the following way :

$$\frac{\partial g}{\partial t} = -\nabla \cdot (g\mathbf{u}) - \frac{\partial}{\partial h}(fg) + \psi^g, \quad (7.2)$$

where $\mathbf{u}$ is the two-dimensional horizontal ice velocity vector, f is the thermodynamic vertical growth rate (m/s) and ψ^g is the so-called mechanical redistribution

¹Louvain-la-Neuve sea Ice Model, developed at the Institut d'Astronomie et de Géophysique G. Lemaître, Louvain-la-Neuve, Belgium.

function, which is a function of h and $g(h)$. ψ^g determines the changes in $g(h)$ during deformation events, redistributing ice area between the different thickness categories. The terms on the right-hand side of (7.2) respectively describe the effects on the ITD of (1) advection by the velocity field, (2) thermodynamic processes (growth and melt) and (3) mechanical processes (opening, rafting and ridging).

Any sea ice-related state variable has an associated distribution function which follows a conservation equation of the type (7.2), as suggested by Bitz *et al.* (2001) and Zhang and Rothrock (2001). Following similar arguments, we define the salinity distribution function s by :

$$s(\lambda, \phi, h, t) = hg(\lambda, \phi, h, t)S(\lambda, \phi, h, t), \quad (7.3)$$

where S is the sea ice bulk salinity, λ is the longitude and ϕ the latitude. During transport and deformation, vertical variations in sea ice salinity are neglected. As g , s follows a conservation equation :

$$\frac{\partial s}{\partial t} = -\nabla \cdot (s\mathbf{u}) + \Theta^s + \psi^s. \quad (7.4)$$

The terms on the right-hand side of (7.4) respectively represent the effects of (1) horizontal advection, (2) halo-thermodynamics and (3) mechanical redistribution. The halo-thermodynamic term reads :

$$\Theta^s = -\frac{\partial(fs)}{\partial h} + \theta^s. \quad (7.5)$$

The first term on the right-hand side is the transport in thickness space due to ice growth and melt, while the second one refers to brine entrapment and drainage. The latter is the main source of salinity variations and will be discussed in the next section.

7.3 One-dimensional fomulation of the thermodynamic changes in ice salinity

In a 1D test case, we compare two formulations of θ^s (SZT-1 and SZT-2) using the same sea ice thermodynamic component. SZT-1 is the sophisticated halo-dynamic module of the VBF07 model, which would be computationally expensive in a large-scale model with ITD. SZT-2 is a simpler parameterization of brine

entrapment and drainage. First, we briefly describe the thermodynamic component, then the two parameterizations and, finally, we discuss the results of a 1D simulation using them alternatively.

7.3.1 Thermodynamic component

The thermodynamic component used here is the Bitz and Lipscomb (1999) 1D energy-conserving model. Prognostic variables are the ice thickness h^i , the snow depth h^s and the vertical temperature profile $T(z)$ in the ice-snow system. Temperature variations are computed by numerically solving the heat diffusion equation in 1 layer of snow and 20 layers of sea ice, characterized by their salinity S . In order to represent the salinity effect on heat transfer and storage, through its impact on the sea ice relative brine content, the sea ice thermal properties depend on S and T (i.e., the specific heat c , thermal conductivity k and energy of melting q , defined as the energy required to melt a unit volume of sea ice). The growth (melt) rates at the interfaces are computed by dividing the heat loss (gain) F by q .

7.3.2 Reference salinity model

The reference formulation used to compute the spatio-temporal variations in ice salinity is the halodynamic module of VBF07 (SZT-1). It is based on laboratory and field experiments and has been validated against sea ice core data. In this model, the evolution of the sea ice salinity profile is governed by brine expulsion (function of the temporal derivative of temperature), gravity drainage (function of the vertical temperature gradient and active when the brine network is open) and flushing (i.e., the percolation of surface meltwater through the open brine network). The relative brine volume $e(z, t)$ is diagnosed as a function of S and T . The brine network is assumed to be interconnected if $e > 5\%$. The boundary condition at the ice-ocean interface specifies the salinity of new ice as a fraction of seawater salinity as indicated by laboratory measurements (Cox and Weeks, 1988) and gives more brine entrapment for larger basal congelation rate.

7.3.3 Simple parameterization of brine entrapment and drainage

The second parameterization of salinity variations used here is a simplification of the VBF07 halodynamic module, hereafter referred to as SZT-2. It is designed for use in large-scale 3D models. An equation for the bulk salinity $\bar{S}$ is solved and the shape of the profile is computed as a function of the bulk salinity. This has the advantage that only the bulk salinity has to be transported and redistributed in the 3D model.

Field data (Malmgren, 1927), laboratory experiments (Cox and Weeks, 1975) and 1D simulations with the VBF07 model (Vancoppenolle *et al.*, 2006) indicate that the sea ice desalination can be divided into initial, winter and summer stages. Gravity drainage and flushing dominate winter and summer desalinations, respectively. In addition, salinity data from ice cores drilled in some of the Southern Ocean deep snow-covered regions suggest that salt concentration occurs at the ice surface during snow ice formation (Jeffries *et al.*, 1997; Perovich *et al.*, 2004). Based on this, the equation for bulk salinity includes brine entrapment during basal growth and snow ice formation, gravity drainage and flushing.

Observation of the vertical sea ice salinity profile from ice cores suggests that the shape of the profile is a function of the bulk salinity. At high salinity, the profile has a C-shape, while at low salinity, the profile has an almost linear shape with low salinity near the surface. This is mainly because of flushing, which affects salinity near the surface (VBF07). In addition, numerical experiments have shown that for mass balance simulation purposes, a linear profile well approximates the realistic, nonlinear sea ice salinity profile (Vancoppenolle *et al.*, 2005). Based on this, a profile shape which is function of the bulk salinity is used. At high salinity, the salinity is assumed to be vertically constant. At low salinity, the salinity profile is assumed to be linear, with $S = 0\%$ at the surface. At intermediate salinity, the salinity profile is a linear combination of the linear and vertically-constant profiles. For more details on this parameterization, see V08, section 6.2.6.5.

7.3.4 Experimental setup

The 1D model is used to simulate the growth and decay of landfast sea ice at Point Barrow (see Figure 7.1), on the northern Alaskan coast (Chukchi Sea site, 2000-2001), using different halodynamic modules (see Table 7.1). The VBF07

model was already validated at that site against observations of ice thickness, snow depth, salinity and temperature profiles.

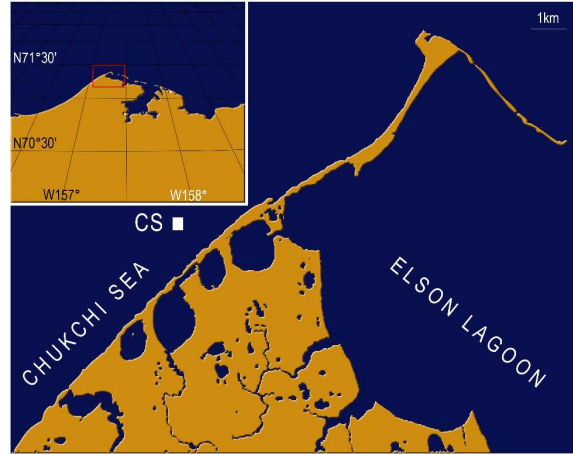


Figure 7.1: Map of Point Barrow neighborhood. Chukchi Sea (CS) sampling site is indicated.

The runs were initialized using in-situ observations of sea ice physical characteristics (Grenfell *et al.*, 2006) and forced with surface heat fluxes derived from Point Barrow meteorological station data and with a prescribed ocean heat flux (see VBF07, section 4.3.2.4, for details).

Five simulations were run using different salinity parameterizations. SZT-1 (control run) and SZT-2 are compared with simpler, prescribed salinity profiles. In CW, the salinity is vertically constant and prescribed as a function of ice thickness, using the relation of (Cox and Weeks, 1974), as in , as in Ebert and Curry (1993). In PR, the vertically varying salinity profile of Schwarzacher (1959), typical of Arctic multiyear (MY) ice, is used, as in Bitz and Lipscomb (1999). Finally, in BK5, an isosaline profile with $S = 5\text{‰}$ is used.

Table 7.1: *Description of the simulations performed with the 1D model.*

Experiment	Description
SZT-1	VBF07 sophisticated halodynamic module
SZT-2	Simple parameterization of brine entrapment and drainage
PR	Steady-state, vertically varying salinity profile of Schwarzacher (1959)
CW	Salinity function of ice thickness (Cox and Weeks, 1974)
BK5	Steady-state, isosaline salinity ($S = 5\text{‰}$)

7.3.5 Results

Ice salinity. The results of the simulations are shown in Figure 7.2. Compared to SZT-1, $\bar{S}$ is very well reproduced in SZT-2. CW has a correct winter $\bar{S}$, but is not able to simulate flushing. The salinity profile in SZT-2 well approximates the non-linear SZT-1 profile. In summer, the simulated near-surface salinity (i.e., taken over the uppermost 50 cm) in SZT-1 and SZT-2 are very close. Desalination in SZT-1 is slightly too slow during the first days of summer.

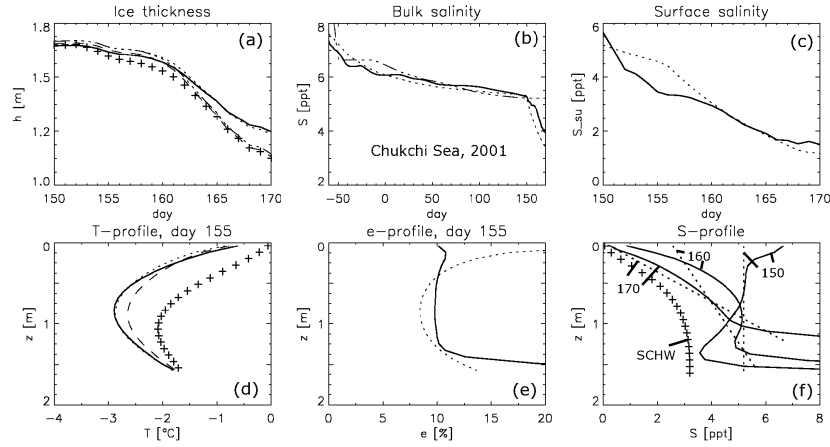


Figure 7.2: Impact of the different parameterizations of salinity variations on several simulation features at Chukchi Sea site, 2001. When present, solid lines refer to SZT-1 simulation, dotted lines to SZT-2 simulation, dashed lines to the BK simulation, dashed-dotted lines to the CW simulation and crosses to the PR simulation. (a) Temporal evolution of ice thickness, (b) temporal evolution of bulk salinity, (c) temporal evolution of near-surface salinity (i.e., taken over the uppermost 7 layers), (d) temperature profile on day 155, (e) brine volume on day 155 and (f) temporal evolution (days 150, 160 and 170) of the vertical salinity profile in the SZT-1, SZT-2 and PR.

Temperature. The temperature profile is similar in all simulations but PR. PR is the only case in which the near-surface salinity is low when the melt period starts. The near-surface specific heat is therefore much lower. This yields faster upper ice warming and melt.

Ice thickness. The ice thickness in SZT-2 compares very well with SZT-1, which itself agrees best with observations. The other simulations underestimate the ice thickness by 10 cm. There are two competing effects of salinity on ice thickness Vancoppenolle *et al.* (2005). First, the increase in specific heat (c) with S favours

lower melt rates for more saline ice. Second, the decrease in energy of melting (q) with S yields higher melt rates for more saline ice Bitz and Lipscomb (1999).

The difference in ice thickness between PR and SZT-1 is due to a larger surface melt rate in PR between days 150 and 160. The warmer upper ice layers in PR during that period suggest that the c effect dominates. The decrease in c associated to low S and high T occurs 10 days later in SZT-1 than in PR. At this time, the surface melt rates in SZT-1 and PR become similar.

The differences in ice thickness from SZT-1 in CW and BK5 are due to larger surface melt after day 160. Between days 150 and 160, the near-surface temperature evolves similarly in all three simulations. After day 160, CW and BK have a higher near-surface salinity than SZT-1. Since the temperature profile is steady and similar in all simulations, only q can play a role here. Due to lower bulk salinity in SZT-1 than in CW and BK5, q is larger near the surface, which yields smaller melt rates. This result contradicts the 1D simulations of Vancoppenolle *et al.* (2006). We will come back to this later on in Section 6.

Brine volume. An exception to the agreement between SZT-1 and SZT-2 is the brine volume vertical profile. The near surface salinity in SZT-1 is smaller than in SZT-2, therefore brine volume is also smaller. Nevertheless, the simulated brine volume in SZT-1 has large errors compared to brine volume derived from observations. "Observed" brine volume is derived from observed salinities and temperatures, but the actual value is not really well known. Clearly, direct observations of brine volume (as a tomography-based method, see Golden *et al.* (2007) would help to better constrain the models. Brine volume is important for fluid transport and biogeochemistry.

7.3.6 Summary

In conclusion, SZT-1 is well approximated by the simple parameterization SZT-2. The other parameterizations lead to errors in the simulated salinity, temperature and ice thickness. For the Point Barrow conditions, a time-varying salinity favours lower surface melt rates. In the first 10 days of summer, high near-surface S and c keep the temperature relatively cold near the surface and inhibit the melt. Then, after intense surface drainage, higher q maintains lower melt rates. Similar experiments – not shown here – designed to simulate Arctic MY ice were performed using SZT-1 and SZT-2 in a Central Arctic configuration. In these experiments the seasonal cycle of surface heat fluxes from a central Arctic climatology is repeated over 50 years as model forcing (see Vancoppenolle *et al.* (2006) for a

detail of the setup). As for Point Barrow conditions, thicknesses and bulk salinities (transient and equilibrium) are similar in the two runs. In conclusion, we argue that SZT-1 can be used instead of SZT-2 to analyse the impact of sea ice salinity in a large-scale sea ice model.

In addition, using SZT-1 was compared to prescribed, constant profiles in the same central Arctic MY ice configuration (Vancoppenolle *et al.*, 2006). These experiments indicated the following: taking temporal and vertical variations of ice salinity into account (1) increases basal ice growth, because the energy required to form a unit volume of sea ice is lower and (2) increases surface melt. The second item apparently contradicts the simulations run here, but reflects the fact that for MY, thicker ice, the salinity near the surface is already quite low in early summer, which reduces c near the surface and facilitates surface melt. There are also presumably differences due to snow and radiative forcings.

7.4 Representation of the salinity distribution in OPA9-LIM3

In this section, we present how the conservation equation for the salinity distribution (7.4) is solved, using SZT-2 to calculate θ^s , in a large-scale sea ice model.

7.4.1 LIM3 overview

The ice-ocean model used here is developed in the framework of the ocean modelling system NEMO² that combines LIM3, a C-grid, dynamic thermodynamic sea ice model, with ice thickness, enthalpy, salinity and age distributions and OPA³ v9.0, a hydrostatic, primitive equation, ocean general circulation model included in NEMO. Hereafter, LIM3 is only briefly described. For more details, see V08.

The ice in LIM3 is represented as a series of $M = 5$ thickness categories. In each of them, the ice cover is vertically divided into $N = 5$ layers of sea ice covered by one layer of snow. Each ice category has its own set of global state variables: the ice concentration, volume per unit area, enthalpy per unit area,

²Nucleus for European Modelling of the Ocean, developed at LOCEAN, Paris, France.

³Ocean Parallélisé; Madec (2008).

salt content and age content, as well as the snow volume per unit area and enthalpy per unit area. Every state variable X has a definite value in all the thickness categories and follows a conservation equation including horizontal transport, transport in thickness space due to ice growth / melt and effects of the thermodynamical and mechanical redistribution processes. The ice momentum equation is solved on a C-grid with an elastic-viscous-plastic rheology (S. Bouillon *et al.*, The Elastic-Viscous-Plastic Sea Ice Model on Different Grid Types, manuscript in preparation). The ice state variables are advected horizontally with the second-order moment conserving scheme of Prather (1986). The thickness distribution formulation is Lagrangian in thickness space and follows Bitz *et al.* (2001). The transport in thickness space is computed using linear remapping (Lipscomb, 2001). The formulation of the mechanical redistribution is based on the ideas of Thorndike *et al.* (1975) and Bitz *et al.* (2001) and includes rafting and porous ridging.

The thermodynamic module of LIM3 includes an exhaustive representation of brine-related processes (see Section 2.6 of V08 for details). The sea ice salt content M_m^s in category m ($m = 1, 2, \dots, M$) is the extensive variable numerically representing the salinity distribution s in LIM3. It is defined, for each category, as $M_m^s = S_m^i v_m^i$, where S_m^i and v_m^i are the sea ice bulk salinity and volume per unit area in category m , respectively. As for other variables, a conservation equation (i.e., equation 7.4) is solved. The term θ^s is computed using the simple parameterization SZT-2 described in Section 3.2. The variations in vertical salinity in the N vertical layers $S_{m,k}^i$ ($k = 1, 2, \dots, N$) are only "seen" by the thermodynamic module. The formation of snow ice introduces salt into the ice cover. Newly forming ice in open water has a salinity given by the salinity-thickness relation of Kovacs (1996). Finally, following earlier studies, the mechanical redistribution function ψ^s is built upon the fundamental ψ^g function (see V08, Section 2.5.4; equation (9) in Zhang and Rothrock (2001); Appendix A of Bitz *et al.* (2001)).

In LIM3, salinity variations affect sea ice thermodynamics as well as ice-ocean salt and fresh water fluxes. Sea ice thermodynamics in LIM3 include the thermal effect of brine inclusions, through thermal properties that depend on S and T . Sea ice mass source and sink terms include basal and snow ice growth, basal and surface melt, and the growth of new ice in open water. Vertical growth and decay is determined using a model very close to Bitz and Lipscomb (1999) (see Section 3.1). For more details, see V08, Section 2.6.2. The ice-ocean salt and fresh water flux (F^s) has two components. The first one is the direct effect of ice growth and melt (F_{eq}^s) and accounts for the salinity of growing (or melting) ice. The second one is brine drainage (F_b^s) and is directly related to salinity variations. See V08, Section 8 for more details.

7.4.2 Forcing and experimental conditions

The ocean and sea ice models both run on the same global domain and use the tripolar ORCA2 C-grid with 2° zonal resolution and a meridional resolution varying from 0.5° at the equator to $2^\circ \cos\phi$ (see, for details, Timmermann *et al.* (2005)). A spinup simulation was run over the period 1948-1969. The model runs presented here were restarted from the end of the spinup and run over 1970-2006. They are analysed over the period 1979-2006 for which satellite data are available. Atmospheric forcing fields are a combined data set consisting of NCEP/NCAR daily reanalysis data of 10-m wind speed and 2-m temperature (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). More details can be found in V08 (Section 3.3). River runoff rates are prescribed using the climatological Baumgartner and Reichel (1975) dataset combined with a mean seasonal cycle derived from the Global Runoff Data Centre (GRDC) data. Contrary to earlier versions of the model, no salinity restoring is used.

7.5 Results of the control simulation

In this section, the simulated salinity field is validated and analysed.

7.5.1 Mean seasonal cycle of bulk sea ice salinity averaged over the ice thickness categories

7.5.1.1 Observation data

Many in-situ measurements of sea ice salinity have been performed. They were typically obtained from the melt of sea ice cores sections (see, e.g., Malmgren (1927); Eicken *et al.* (1991)). In order to validate the salinity field simulated by LIM3, a part as large as possible of available salinity data, unfortunately excluding the Soviet sources was reviewed (see Tables 7.2 and 7.3). For each single data set, we either took the data from tables or digitized them from plots in the original sources. In the majority of the cases, bulk salinity, inter-core standard deviations and the number of drilled cores were available.

Table 7.2: *Summary of the Arctic salinity data. S is the sea ice mean salinity, σ_S the inter-core standard deviation of S , N is the number of cores. BS stands for Beaufort Sea, AO for Arctic ocean, FS for Fram Strait, BB for Baffin Bay, SV for Svalbard Archipelago (landfast ice), LapS for Laptev Sea and LabS for Labrador Sea.*

Month	Year	Season	Sector	S	σ_S	N	Reference
Oct.-Dec.	22-25	fall	AO	8.6	3.3	17	Malmgren (1927)
Jan.-Mar.	22-25	winter	AO	7.8	3.6	9	Malmgren (1927)
May-Sep.	58	summer	AO	2.2	-	40	Schwarzacher (1959)
Aug.-Sep.	91	summer	AO	2.1	0.7	66	Eicken <i>et al.</i> (1995)
Jul.-Aug.	94	sum. FY	AO	1.3	0.1	4	Tucker <i>et al.</i> (1999)
Jul.-Aug.	94	sum. MY	AO	2.0	0.4	17	Tucker <i>et al.</i> (1999)
Sep.	22-25	summer	AO	2.1	0.3	3	Malmgren (1927)
Jun.-Jul.	84	sum. FY	FS	4	-	20	Tucker <i>et al.</i> (1987)
Jun.-Jul.	84	sum. MY	FS	2.1	-	31	Tucker <i>et al.</i> (1987)
March	71	winter	BS	4.0	1.6	3	Cox and Weeks (1974)
Mar.-April	72	winter	BS	3.7	0.5	30	Cox and Weeks (1974)
Jan-May	00-01	win.-spr.	BS	5.6	0.2	32	Vancoppenolle <i>et al.</i> (2007)
April	69	spring	BS	6.8	0.4	4	Cox and Weeks (1974)
April	83	spring	BS	4.7	0.4	5	Tucker <i>et al.</i> (1984)
June	00-01	summer	BS	3.4	0.3	36	Vancoppenolle <i>et al.</i> (2007)
Jun.-Aug.	58	summer	BS	1.8	0.6	19	Cox and Weeks (1974)
June-Aug.	98	summer	BS	2.0	1.7	14	<i>SHEBA</i>
Sep.-Oct.	69	summer	CA	2.2	0.3	9	Cox and Weeks (1974)
October	69	fall	CA	9.1	2.0	15	Cox and Weeks (1974)
Apr.-May	99	spring	LapS	3.8	2.0	91	Eicken and A. Proshutinsky (2005)
May	98	spring	SV	4.5	-	6	Gerland <i>et al.</i> (1999)
Oct.-Nov	56	fall	BB	7.75	-	-	Weeks and Lee (1962)
Nov-May	77-78	winter	BB	6	-	13	Nakawo and Sinha (1981)
Dec.-Jan.	55	fall	LabS	8.5	2.1	21	Weeks and Lee (1958)
Jan.-Mar.	56	winter	LabS	6.7	0.8	15	Weeks and Lee (1958)
Mar.-May	56	spring	LabS	4.3	0.8	14	Weeks and Lee (1958)
April	72	spring	BS	9.6	2.6	24	Cox and Weeks (1974)
Feb.-Mar.	70	winter	BerS	5.5	0.4	8	Cox and Weeks (1974)

Table 7.3: *Summary of Antarctic salinity data. S is the sea ice mean salinity, σ_S the standard deviation of S , N is the number of cores. WP stands for West pacific sector, I for Indian Sector, RS for Ross Sea, BeS for Bellingshausen Sea and WS for Weddell Sea.*

Month	Year	Season	Sector	S	σ_S	N	Reference
Apr	93	fall	WP	7.6	1.9	18	Worby <i>et al.</i> (1998)
Aug	99	winter	WP	13.0	4.1	43	V. Lytle, unpub.
Oct	91	spring	WP	5.5	1.8	4	Worby <i>et al.</i> (1998)
Sep	94	winter	I	6.5	2.0	33	Worby <i>et al.</i> (1998)
Nov	91	spring	I	5.7	1.3	15	Worby <i>et al.</i> (1998)
Nov	92	spring	I	6.7	3.3	24	Worby <i>et al.</i> (1998)
Jul-Aug	01	winter	BeS	7.1	1.1	20	Perovich <i>et al.</i> (2004)
Aug-Nov	93-94	winter	BeS	5-6	-	13	Jeffries and Porter (1995)
Oct-Nov	80	spring FY	RS	5.9	0.7	26	Gow <i>et al.</i> (1998)
Nov	04	spring MY	RS	3.86	2.17	13	Eicken, unpub.
Sep-Oct	89	win.-spr. FY	WS	6.0	1.2	22	Eicken (1998)
Sep-Oct	89	win.-spr. MY	WS	4.8	1.4	9	Eicken (1998)
Oct-Nov	88	spring	WS	4.2	0.3	8	Eicken <i>et al.</i> (1991)
May-Dec	86-89	win.-spr.	WS	6	-	129	Eicken (1992)
Dec	04	sum.	WS	5.5	1.0	14	Tison <i>et al.</i> (2008)
							Hellmer <i>et al.</i> (2006)
Feb-Mar	80	sum.-fall FY	WS	4.5	0.9	62	Gow <i>et al.</i> (1987)
Feb-Mar	80	sum.-fall MY	WS	3.4	0.5	14	Gow <i>et al.</i> (1987)

These data, collected by different teams, at different places and times, are neither coherent in space nor in time. Furthermore, they are subject to significant spatial variability. Therefore, the following data processing was applied. In order to remove subfloe- and larger-scale spatial variability, average salinities were computed over cores which are coherent in space and time. The inter-core average salinities were then sorted by hemisphere and region. This resulted in 28 (Arctic) and 17 (Antarctic) regional salinity values, corresponding to a total of 1063 individual ice cores. Despite the present record is subject to errors of several origins and could certainly be improved by adding more data, the large-scale pattern shown by the data should be considered as robust thanks to the already large amount of data compiled.

7.5.1.2 Comparison model-observations

For several regions in each hemisphere, each observed salinity was plotted versus the corresponding mean sampling month, independently of the sampling year, and compared with the corresponding mean salinity (weighted over the thickness categories) taken from the seasonal 1979-2006 model climatology (see Figure 7.3). The regions considered in the Northern Hemisphere (hereafter referred to as NH) are the Arctic Ocean, its peripheral seas (Kara, Laptev, East Siberian, Chukchi and Beaufort Seas) and the outer Arctic, which gathers the remainder of the ice-covered regions of the NH. The Arctic region corresponds to the Arctic Ocean and its peripheral seas.

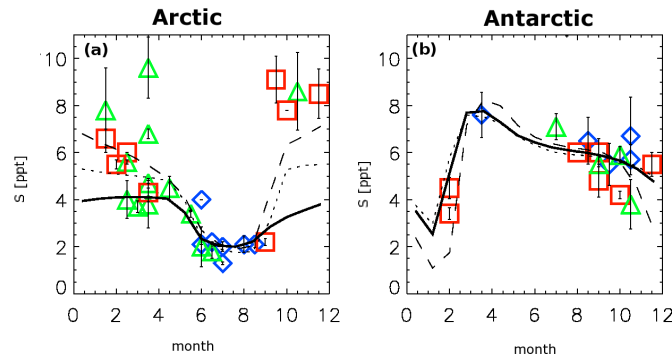


Figure 7.3: Simulated and observed 1979-2006 mean seasonal cycle of ice salinity in several regions of the Arctic (a) and Antarctic (b). The model means take into account the volume-weighted contributions of the ice thickness categories. The observations are inter-core means computed from different field campaigns (see text for details). Regions for model values in the NH are : Arctic Ocean (solid); peripheral seas (dot); and outer Arctic (dash). For observations, regions are : Arctic Ocean and Fram Strait (blue diamonds); peripheral seas (green triangles) and outer Arctic (red squares). Regions for model values in the SH are : global Southern Ocean (solid); Weddell Sea (dots) and West Pacific sector (dash). For observations, regions are : West Pacific and Indian sectors (blue diamonds), Ross, Amundsen and Bellingshausen Seas (green triangles), and Weddell Sea (red squares).

The model-data agreement is generally good, though there are a few data points in the NH characterized by a very high salinity for which the model-data agreement is poor. For those points, error is mostly due to the low statistical representativity of the observed value. For example, the high salinity March-April point in the NH (see top of Figure 7.3a, data point from Cox and Weeks (1974)) is typical of newly formed thin ice, while the simulated value – averaged over thickness categories – is dominated by thick ice.

Field studies report significant differences between Arctic and Antarctic sea ice salinities (see, e.g., Gow and Tucker (1991) that are simulated by the model. First, both data and model indicate that sea ice is more saline in the Southern Hemisphere (SH) than in the NH. This is because, in the SH, (1) the relative contribution of young ice, more saline, is higher, (2) the amount of surface melt and flushing is much lower and (3) the contribution of snow ice formation, which traps large amounts of salt, is higher (see V08) than in the NH.

Figure 7.3 reveals that there is a large amplitude seasonal cycle of ice salinity in both hemispheres, characterized by a maximum in fall and a minimum in summer. The annual minimum salinity occurs thus when the ice is the oldest (summer). This minimum reaches around 2‰ in the NH and 4‰ in the SH. The seasonal maximum salinity takes place when the abundance of newly formed, saline ice is the largest (fall).

7.5.2 Ice salinity distribution in thickness space

The second model validation test is to compare the salinity distribution in thickness space with the observations of Cox and Weeks (1974). Cox and Weeks' relations were compiled from NH data and show a winter decrease in ice salinity with thickness. In summer, ice has salinity around 2‰ for all ice thicknesses.

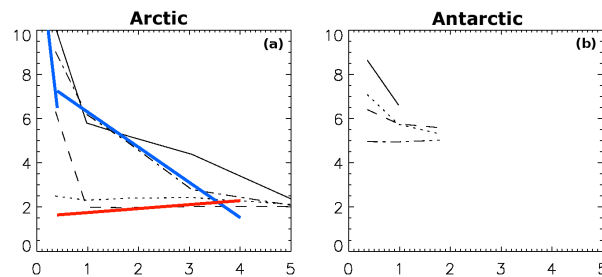


Figure 7.4: Simulated mean bulk salinity S (‰) versus ice thickness h (m) in the Arctic (a) and in the Antarctic (b) in March (solid), June (dot), September (dash) and December (dash-dot). On the left panel, the Cox and Weeks (1974) $S - h$ relations for winter (blue) and summer (red) – which were derived from cores in the Arctic Region – are depicted. Means are ice volume weighted and computed category per category over the whole domain.

The seasonal cycle of the salinity distribution in thickness space is shown in Figure 7.4. In the NH, the model agrees well with the Cox and Weeks' relations. The

salinity distribution is characterized, on the one hand, by large differences in winter, with thin (young) ice being more saline than thick ice, and on the other hand, by an isosaline ice cover in summer. In addition, the winter NH simulated salinity and age are strongly correlated (-0.78). In the SH, differences in ice salinity are lower. The SH simulated age-salinity correlation is low, because of the low surface melt and flushing, which prevents the formation of a low-salinity MY ice cover.

7.5.3 Geographical variations of ice salinity

Figure 7.5 depicts the geographical variations in mean ice salinity simulated by OPA9-LIM3. In the NH, the regional variations reflect the FY-MY contrast, with larger values in the seasonal ice zone (SIZ), associated to young ice, and lower values in the perennial ice zone (PIZ), associated to old ice. In the peripheral seas, intermediate salinities are found and reflect the mixture of young and old ice. In contrast to the other Arctic regions, the simulated winter sea ice salinity in the Baltic Sea lies within 0-2‰. Granskog *et al.* (2004) reported similar values (0–1.5‰). In the Baltic Sea, sea surface salinity ranges from 6 to 8‰. Therefore, fractional entrapment of brine during growth leads to much lower salinities, which is well reproduced by the model. In the SH, the regional variations in ice salinity are relatively small compared to the Arctic, though salinity is higher than average where ice forms intensively, i.e., along the ice edge and in coastal polynyas in the Weddell and Ross Seas as well as along East Antarctica.

7.6 Sensitivity experiments

In order to investigate the impact of salinity variations on the simulated sea ice and upper ocean physical characteristics, we run several sensitivity experiments which are summarized in Table 7.4.

Table 7.4: *Description of the simulations performed with OPA9-LIM3.*

Experiment	Description
SZT-IO	Prognostic ice salinity used
SZT-I	Prognostic ice salinity used in thermodynamic computations only
BK5	Prescribed, vertically constant salinity ($S = 5‰$)
PR	Steady-state, vertically varying salinity profile (Schwarzacher, 1959)

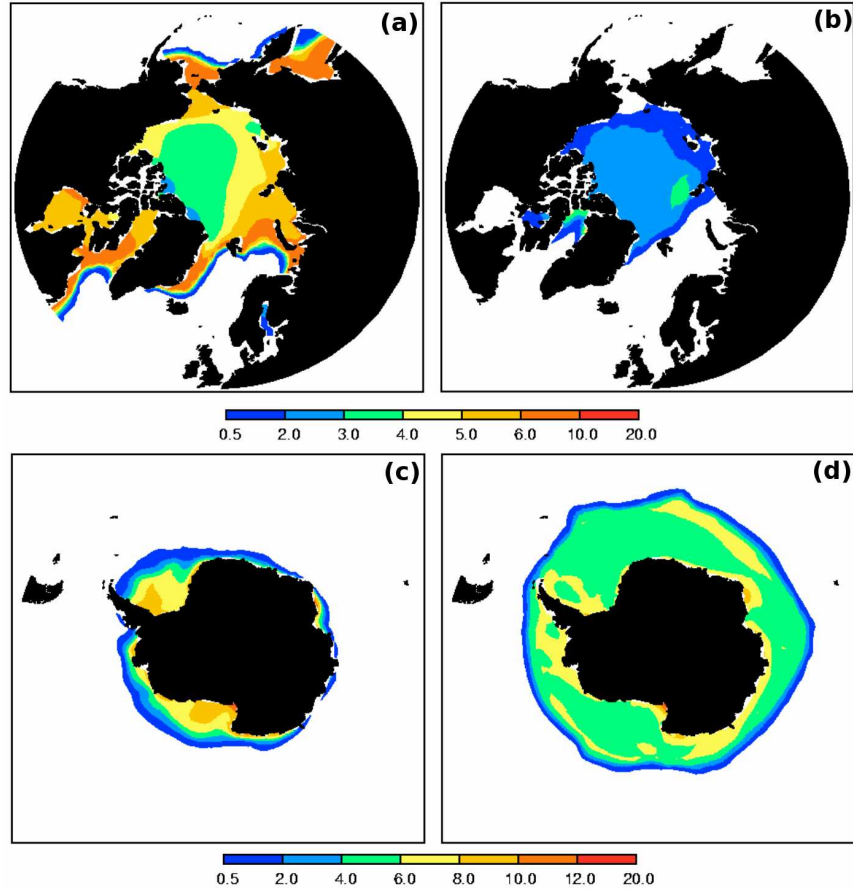


Figure 7.5: Simulated spatial distributions of mean sea ice salinity (‰) in the NH, March (a) and September (b), and in the SH, March (c) and September (d), 1979-2006. The means take into account the weighted contributions of the ice thickness categories, but exclude open water.

In SZT-IO (control simulation), (1) the temporal and vertical variations in ice salinity (equations (6.19-6.21) of V08) are included and affect the sea ice thermodynamic processes. (2) The salt flux due to brine drainage (equation (26) of V08) is applied. (3) The ice-ocean salt / fresh water flux due to ice growth / melt is computed using a varying salinity (equation (6.26) of V08). If the ice is growing, this salinity depends on the formation mechanism (i.e., new ice formed in open water, at the ice base and through snow-ice formation). If the ice is melting, this salinity corresponds to the salinity of the melting ice layer.

In SZT-I, (1) the variations of ice salinity are included in the sea ice thermodynamic computations. (2) The salt flux due to brine drainage is set to 0. (3) The ice-ocean salt / fresh water flux due to ice growth / melt is computed using a prescribed salinity ($S = 5\text{‰}$) if ice grows or melt. Though this run is not salt-conserving, its interest is to isolate the effects of salinity variations on sea ice thermodynamics. In contrast, the impact of salinity variations on ice-ocean salt and fresh water fluxes and their influence on ocean and on sea-ice, through oceanic feedbacks, are excluded.

In BK5, the ice is supposed isosaline with $S = 5\text{‰}$ in the sea ice thermodynamic computations as well as in the computation of the ice-ocean salt / fresh water flux. Representing salinity in that manner is done frequently (e.g., Fichefet and Maqueda (1997); Zhang and Rothrock (2001)).

In PR, the steady state, vertically varying profile of Schwarzacher (1959), typical of central Arctic MY ice, is used in the computation of the ice-ocean salt / fresh water flux and the corresponding vertically averaged salinity ($S = 2.3\text{‰}$) is used to compute the ice-ocean salt / fresh water flux. This representation of ice salinity was used in several instances (e.g., Bitz *et al.* (2001); Hunke and Lipscomb (2004)).

7.6.1 Northern Hemisphere

In the Arctic, over 1979-2006, regarding sea ice, the simulations SZT-IO and SZT-I are quite similar, while they differ from BK5 and PR. This suggests that in the NH, most of the changes in sea ice are due to changes in the thermal properties of sea ice. On the contrary, regarding the ocean, differences from SZT-IO are at least qualitatively similar among the three sensitivity runs. Therefore, the direct effect of varying salinity through changes in the ice-ocean salt flux has an effect on the ocean that is presumably too small to induce feedbacks on sea ice that are significant at a large-scale.

7.6.1.1 Sea ice characteristics

In light of these remarks, in the NH, the discussion will be focused on SZT-IO and BK5. Significant differences in PR and SZT-I will only be reported when necessary.

Ice area and concentration. A summary of various run statistics is given in Table 7.5. The differences in total ice extent between the four simulations are within $0.2 \times 10^6 \text{ km}^2$. Comparison to passive microwave data of ice coverage (Comiso, 2007) indicates that the correlation of anomalies of ice extent is higher in SZT-IO.

Table 7.5: *Statistics for control and sensitivity runs. $V(10^3 \text{ km}^3)$ and $E(10^6 \text{ km}^2)$ are the total ice volume and extent over the hemisphere of interest, respectively. $\Delta h^i(\text{m})$ is the bias in ice thickness between simulated and observed values. The observed ice thickness are derived from upward-looking sonar (ULS) observations of ice draft (Wensnahan et al., 2007). See V08, Section 4.3.1 for details on the procedure. C_c is the correlation coefficient between the simulated and observed monthly mean anomalies of ice extent. Observations of ice coverage derive from passive microwave observations (Comiso, 2007).*

Source	Hem.	V_{min}	V_{mean}	V_{max}	Δh^i	E_{min}	E_{mean}	E_{max}	C_c
DATA	N	-	-	-	-	5.9	12.0	16.3	-
SZT-IO		11.8	18.8	25.1	-0.55 ± 1.04	5.8	10.9	14.4	0.74
SZT-I		12.0	18.9	25.2	-0.53 ± 1.01	5.9	10.9	14.4	0.73
BK5		12.1	18.5	24.4	-0.58 ± 1.00	6.0	10.9	14.5	0.72
PR		10.4	17.2	23.2	-0.83 ± 0.98	5.8	10.9	14.5	0.73
Source	Hem.	V_{min}	V_{mean}	V_{max}	Δh^i	E_{min}	E_{mean}	E_{max}	C_c
DATA	S	-	-	-	-	3.3	11.9	18.8	-
SZT-IO		0.2	4.8	8.9	-	1.0	11.8	19.3	0.65
SZT-I		0.1	3.4	6.4	-	0.4	10.1	17.4	0.63
BK5		0.1	3.8	7.3	-	0.5	10.9	18.6	0.62
PR		0.1	3.6	6.8	-	0.5	10.7	18.5	0.64

Differences in ice concentration between runs are rather small. In winter (see Figure 7.6, panels a-b, for BK5), the differences are located in the marginal ice zones (MIZ) and show qualitatively similar patterns for the three sensitivity runs, including SZT-I. This is an exception to the general pattern and underlines the role of ice-ocean interactions in the MIZs. As noticed for the ice area, the ice edge in SZT-IO compares slightly better to passive microwave observations (Comiso, 2007). In the Labrador Sea, ice concentration is up to 10% higher in SZT-IO than in BK5, which is associated to less intense convective activity in SZT-IO (see Section 6.1.3 for a discussion of changes in the oceanic circulation). The ice tongue along Greenland is narrower, with ice concentration up to 0.05–0.1 lower in SZT-IO, which is linked to a shift to the west of convection site located in the Greenland Sea. In the Baltic Sea, S is locally lower in SZT-IO than in BK5, which

induces a larger energy required for ice formation (q), thus less new ice formation and, therefore, lower ice concentrations (by up to 0.1). In the Barents Sea, the differences are not clear and change of sign from month to month. In the Pacific region, changes in sea ice are associated to small changes in the surface oceanic circulation that bring more (less) heat, induce SST warmer by 0-0.1°C (colder by 0-0.2°C) and in turn, lead to more (less) ice in the Okhotsk (Bering) Sea in SZT-IO compared to BK5.

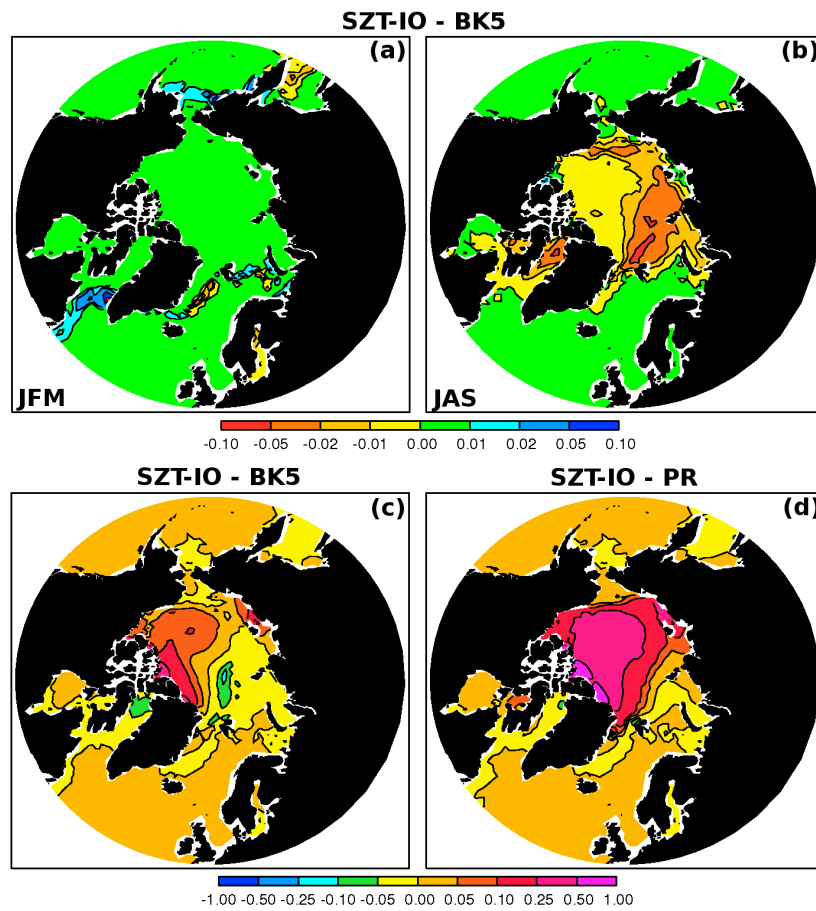


Figure 7.6: SZT-IO - BK5 differences in ice concentration in winter (a) and summer (b), in the NH, averaged over 1979-2006. Annual mean 1979-2006 SZT-IO - BK5 (c) and SZT-IO - PR (d) differences in ice thickness for the NH.

Ice volume and thickness. The NH ice volume confirms the 1D findings summarized in the last paragraph of Section 3.6. In SZT-IO, the seasonal cycle of the NH sea ice volume is similar to SZT-I, very close to BK5 and significantly different

from PR (see Table 7.5). In the NH, in SZT-IO, the ice grows more rapidly than in BK5 and, therefore, the seasonal maximum of ice volume is 1.5% larger. Ice in SZT-IO also melts more rapidly. Consequently, the minima of ice volume in the SZT-IO and BK5 differ only by 0.8%. Similarly, ice grows more rapidly in SZT-IO than in PR. Despite ice melts more rapidly in SZT-IO than in PR, the difference in growth dominates and the volume in PR is, on annual average, 11% smaller than in SZT-IO. Interannual variations of the difference in ice volume between SZT-IO, PR and BK show larger contrasts than mean differences. During years of net ice growth (in the 70's particularly), the differences between SZT-IO and BK5 increase with time, reaching 5% in April 1980. During years of net ice melt, i.e., from 1987 on, they decrease. At the end of the simulation, the ice volumes in the different runs are back to values that are similar to each other.

The spatial distribution of the ice thickness differences between SZT-IO and BK5 (see Figure 7.6, panels c-d), averaged over 1979-2006, shows that the largest differences are located north of the Canadian Archipelago and Greenland (10–50 cm) and in the Beaufort Gyre (5–15 cm). Differences between SZT-IO and PR are even larger, lying in the 10–50 cm range in most of the Arctic Ocean and from 50 to 100 cm north of the Canadian Archipelago. Differences in ice thickness between SZT-IO and SZT-I are within 10 cm. Comparison between modelled ice thickness and ULS data of ice draft (Wensnahan *et al.* (2007); see V08 for the comparison procedure and Table 7.5 for details) indicates a reduction of the bias in ice thickness by 5 cm (30 cm) from SZT-IO to BK5 (PR).

Sources and sinks of ice mass. The differences in NH ice thickness between the runs are due to changes in growth and melt that follow from changes in sea ice thermal properties. In turn, this affects sea ice growth and melt. The spatial distribution of the SZT-IO–BK5 differences in the main growth and melt terms is shown in Figure 7.7. The average seasonal cycles of those terms, averaged over the Arctic Ocean (where the differences are the largest), are shown, for all the simulations, in Figure 7.8. Unless otherwise stated, all the numbers presented here are annual averages computed over the ice-covered regions of the NH.

Ice growth is 5% larger in SZT-IO than in BK5, in the PIZ, due to basal growth and, in the SIZ, to the growth of new ice in open water, 75% of the difference in growth between SZT-IO and BK5 are due to ice growth in open water, while the remainder is due to basal ice growth. The interactive salinity of new ice forming at the ice base and in open water in SZT-IO is significantly larger than the 5‰ used in BK5, which yields a smaller energy required to form ice (q). In turn, for a given heat loss, this results in a higher growth rate in SZT-IO. In fall, the amount of basal growth is 2% (1%) larger in PR compared to BK5 (SZT-IO) because of thinner ice in PR. Heat conduction through ice is larger and compensates the

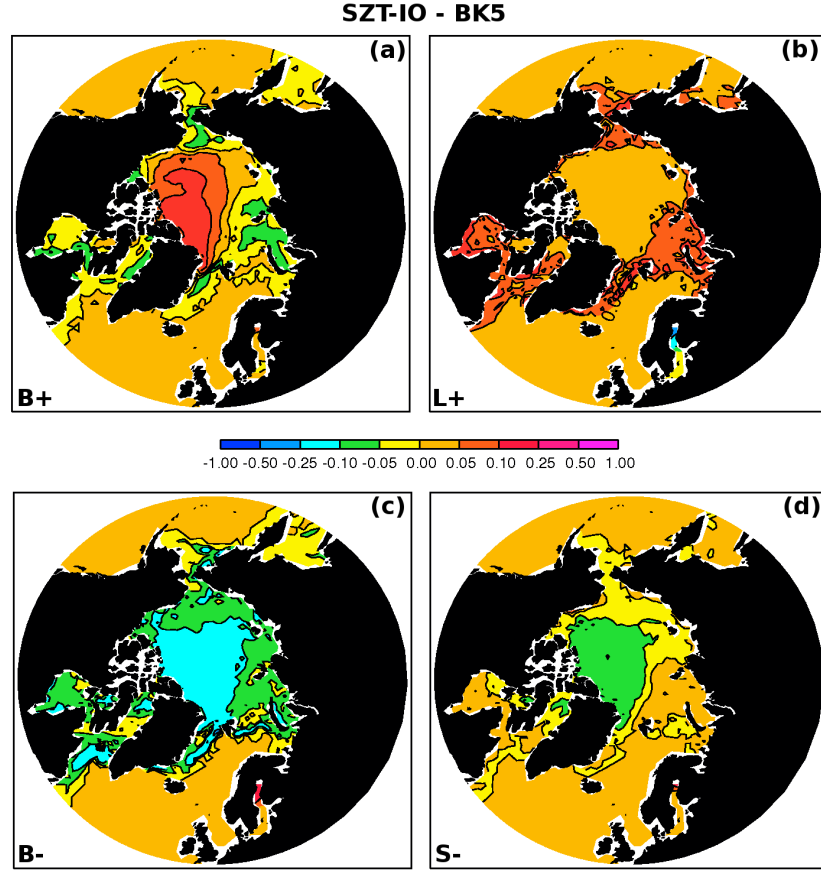


Figure 7.7: SZT-IO - BK5 differences in the different thermodynamic source / sink terms of ice mass (m/y), NH, 1979-2006 : (a) basal growth; (b) growth of new ice in open water; (c) basal melt; (d) surface melt.

effect of the larger q (due to lower S) effect, which tends to reduce basal growth in PR.

In SZT-IO, the annual amount of ice melt is 11% larger in SZT-IO compared to BK5. 80% of the differences come from changes in basal melt (all over the pack). The remainder is due surface melt in the PIZ. Ice melts by 13% more at the ice base in SZT-IO compared to BK5 for two reasons. First, in SZT-IO, associated to the lower S , c is lower and, therefore, the heat coming from the ocean is preferentially apportionned to melting at the ice base rather than to internal warming and melting. Second, associated to a stronger ice-albedo

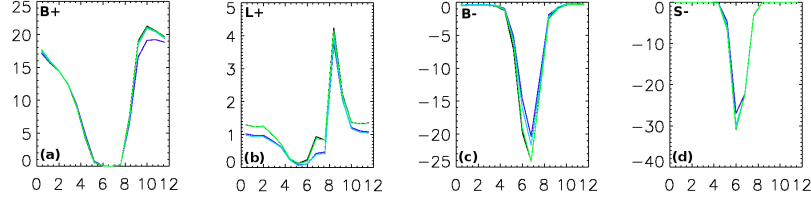


Figure 7.8: 1979-2006 mean seasonal cycle of the different thermodynamic source / sink terms of ice mass (cm/month), Arctic Ocean, for SZT-IO (black), SZT-I (green), BK5 (dark blue), PR (light blue) : (a) basal growth; (b) growth of new ice in open water; (c) basal melt; (d) surface melt. Horizontal axis represents months over the year.

feedback in SZT-IO, the summer oceanic heat flux is 4 W/m^2 higher. Due to the larger coverage of thin ice and to its easier melt in early summer in SZT-IO, the ice concentration decreases more rapidly, which increases the absorption of solar radiation in the ocean and, in turn, feedbacks on the ice through larger heat flux and basal melt. This argument, based on coverage and thermal properties of thin ice, is corroborated by the similarity between SZT-I and SZT-IO. At the surface, in the Arctic Ocean, the amount of melt is 10% larger in SZT-IO compared to BK5. As explained in Vancoppenolle *et al.* (2005), in summer, the shape of the salinity profile, characterized by low salinity near the surface, largely reduces the specific heat c near the surface and facilitates melt. This effects dominates here the effect of the larger q in SZT-IO, which tends to reduce surface melt. For the same reasons, in the Arctic Ocean, the amount of surface melt is 6% higher in PR than in BK5.

7.6.1.2 Storage of fresh water and salt in sea ice

In this section, we analyse how the representation of salinity affects sea ice fresh water and salt contents as well as the sea ice dilution potential on the oceanic upper layer. In a given region R , the sea ice fresh water content C_{fw}^i (kg H_2O) can be defined as :

$$C_{fw}^i = \int_R \int_0^\infty \rho_i h g(h) dh dR, \quad (7.6)$$

ρ_i being the sea ice density; the sea ice salt content C_s^i (kg $NaCl$) can be defined as :

$$C_s^i = \int_R \int_0^\infty \rho_i h g(h) \frac{S}{1000} dh dR \quad (7.7)$$

and the sea ice dilution potential as $C_{dil}^i = C_{fw}^i S^w / 1000 - C_s^i$, where S^w is the seawater salinity. C_{dil}^i is a measure of the dilution effect on the first oceanic layer if all ice melted, accounting for both fresh water and salt contents in sea ice.

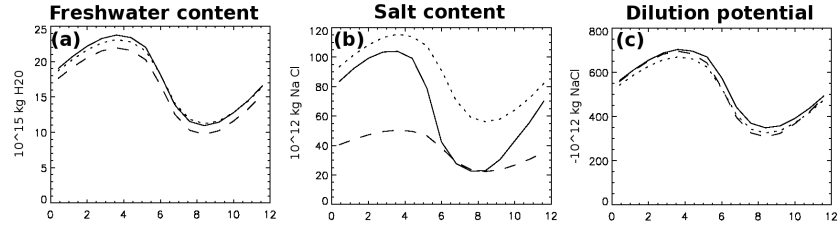


Figure 7.9: Mean 1979-2006 seasonal cycle of the total sea ice fresh water content (a), salt content (b) and dilution potential (c) as defined in Section 6.1.2 for the entire NH. Solid lines correspond to SZT-IO, dotted lines to BK5 and dashed lines to PR. Since SZT-I does not conserve salt, it makes no sense to show the associated diagnostics.

In Figure 7.9, the mean seasonal cycle of these three diagnostics for the NH is depicted. For fresh water content, differences are due to ice volume, while for salt content, they are due to both ice volume and salinity. Though, on annual average, differences in fresh water contents between SZT-IO and BK5 are within 1%, the salt content is overestimated by 25% in BK5, with maximal differences in summer. Consequently, BK5 underestimates the sea-ice dilution potential by 5%. The relatively small impact on the dilution potential is due to rather small sea-ice salinity values compared to oceanic values. Finally, compared to SZT-IO, PR underestimates the fresh water content by 8% and salt content by 42%, on annual average. Differences tend to counterbalance each other and result in an underestimation of the dilution potential by 4% in PR compared to SZT-IO.

7.6.1.3 Ice-ocean interactions and characteristics of the ocean

In this section, we will describe how the ice-ocean salt and fresh water fluxes differ in the present simulations and briefly outline how this affects the upper ocean.

Ice-ocean salt and freshwater fluxes. Using a one-dimensional model, Vancoppenolle *et al.* (2005, 2006) showed that the differences in ice-ocean salt flux are dominated by differences in growth / melt rate, but also that the relatively high salinity of growing ice at the base, typically 6 – 10‰, reduces the ice-ocean salt flux. In the simulations presented here, this effect is enhanced due to the contribution of highly saline ice growing in open water (up to 13‰).

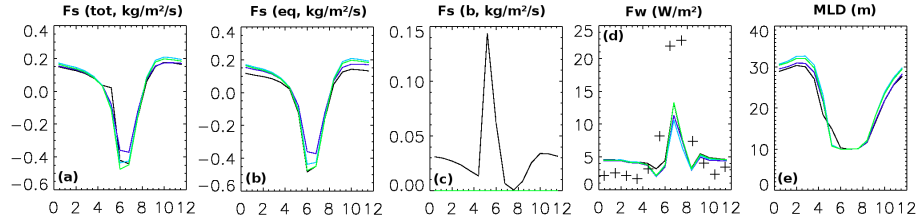


Figure 7.10: 1979-2006 mean seasonal cycles of (a) the total ice-ocean salt flux (F^s), (b) the ice-ocean salt flux due to ice growth / melt (F_{eq}^s), (c) the ice-ocean salt flux due to brine drainage (F_b^s), (d) the oceanic heat flux (F_w) and the mixed layer depth averaged over the Arctic Ocean, for SZT-IO (black), SZT-I (green), BK5 (dark blue) and PR (light blue). Horizontal axis represents months over the year. The crosses on plate (d) refer to the data of Krishfield and Perovich (2005).

During the ice growth season, averaged over the NH, the total ice-ocean salt flux (F^s) is 7% smaller in SZT-IO than in BK5 (see Figure 7.10a). Its components differ more markedly. Even with higher ice growth rate in SZT-IO compared to BK5, the salt flux due to ice growth (F_{eq}^s) is 29% smaller in SZT-IO, which underlines the role of the relatively high salinity of growing ice. Nevertheless, this salt is rapidly rejected back to the ocean by gravity drainage (F_b^s), with a characteristic time scale of twenty days. In regions of concomittant ice production and export (e.g., on the Siberian shelf), during the growth season, this reduces the total salt flux at the production site and increases it away from the production site.

In summer, at the beginning of the melt season, sea ice quickly loses a significant amount of salt through flushing with a characteristic time scale of ten days. This yields a relatively high F^s and maintains a 3m deeper mixed layer in early summer in SZT-IO compared to BK5. This effect can be related to the convection plumes which have been observed in early summer melting sea ice close to Spitsbergen by Widell *et al.* (2006). As bottom melt rates are practically similar in SZT-I and SZT-IO, this effect, localized in time, does not seem to have a large-scale impact. This is because, in summer, averaged over the the ice-covered areas of the NH, the negative ice-ocean salt (i.e., fresh water) flux due to ice melt is more than 8 times higher than the salt input due to flushing.

In turn, due to larger amounts of melt in SZT-IO compared to BK5, the absolute value of the total negative salt flux is 6% larger. Differences between SZT-IO and PR are due to summer drainage of brine, which explains why the absolute

value of the total negative salt flux is 7% smaller in SZT-IO, while ice melts more in SZT-IO.

Characteristics of the ocean. Figure 7.11 depicts the March salinity differences at the surface and at 200m depth in the Arctic Ocean, between SZT-IO and BK5, averaged over 1990-1999. The averaging period was changed from the remainder of the analysis in order, first, to be late enough after initial conditions and, second, to exclude the 2000s, which were characterized by a series of exceptional minima of sea ice extent. In the Arctic Ocean, at the surface, waters are 0 – 0.8‰ fresher in SZT-IO than in BK5, while at 200 m depth, they are slightly more saline, characterized by differences within 0.1‰. Differences in sea surface salinity (SSS) are maximum in the intense ice formation zones: on the East Siberian shelf (0.5 – 0.8‰), in the Beaufort Gyre (0.3 – 0.5‰) and in the Hudson Bay (0.15 – 0.25‰). In the remainder of the Arctic Ocean, differences are within 0.2‰. On the East Siberian shelf, comparison with SZT-I indicates that more than half of the differences can be attributed to the direct effect of varying salinity on salt flux. Compared to the PHC climatology (Steele *et al.*, 2001), SZT-IO reduces the bias in SSS on the Siberian shelf, in the Beaufort Sea and along the northern coast of Greenland, though not enough.

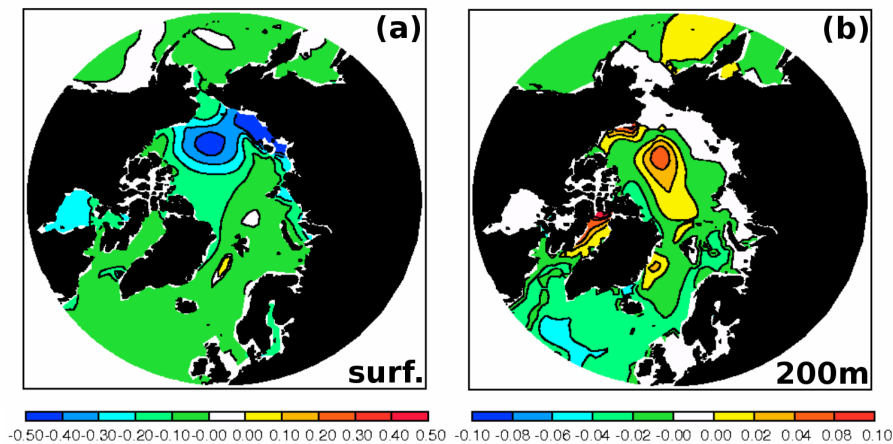


Figure 7.11: Differences in oceanic salinity (March) in the NH at the surface (a) and at 200 m depth (b), averaged over 1979-2006.

In addition, there are also differences in SSS near the model deep convection sites. First, the Irminger convection site, between Iceland and Svalbard, is displaced to the west. In March, the month of maximum convective activity, 25 – 50 m deeper mixed layers are found more to the west, closer to Greenland, in SZT-IO compared to BK5, associated to higher SSSs (0.1‰). The subsequent increase in

oceanic heat flux ($1\text{--}5\text{ W.m}^{-2}$) induces changes in sea ice that in turn stabilize the pattern: reduced ice concentrations decrease the basal melt and the associated fresh water flux. Conversely, shallower mixed layers and fresher surface waters are found to the east.

The second NH convection site, located in the Labrador Sea, is comparatively more affected. In March, mixed layers are $100 - 600\text{ m}$ shallower in SZT-IO compared to BK5 and SSSs are lower by $0.1 - 0.4\text{‰}$. In this location, the ocean has close characteristics in SZT-IO and SZT-I. Therefore, changes must be due to the direct impact of salinity variations on the ice-ocean salt flux. In fall, during the preconditioning of upper seawater for deep convection, there is the same amount of new ice formed in the Labrador Sea and Baffin Bay in SZT-IO and in BK5. Nevertheless, in SZT-IO, the associated salt rejection to the ocean is smaller (by $5\text{--}10\%$). Thus, the water column is fresher and more stable in SZT-IO than in other runs, which in turn weakens deep convection. In summer, the NH differences between SZT-IO and BK5 in terms of seawater salinity closely resemble the winter pattern except in the deep convection zones.

Using a variable salinity in SZT-IO inhibits the vertical exchanges of salt in the ocean. In addition, changes in deep convection in the Labrador Sea are associated to slight changes in global oceanic circulation. Compared to BK5, SZT-IO features reduced convective activity, which corresponds to a slightly weaker Labrador Gyre, a stronger North Atlantic Current and colder temperatures in the North Atlantic, with differences typically up to -0.4°C . The magnitude of the zonally integrated meridional streamfunction in the North Atlantic is smaller in SZT-IO than in BK5 (maximal differences 0.7 Sv). Nevertheless, our runs are too short to draw firm conclusions on the impact of salinity variations on the global oceanic circulation.

7.6.2 Southern Hemisphere

In the SH, the physics of ice, ocean and their interactions differ remarkably from the NH situation. The Southern Ocean is characterized by a water column that is less stable compared to the NH and therefore more sensitive to changes in the buoyancy forcing from sea ice growth / melt. Consequently, in the Southern Ocean, ice-ocean interactions and the associated feedbacks are important. This is confirmed by the magnitude of the changes between SZT-I and SZT-IO, which is much larger in the SH than in the NH.

7.6.2.1 Sea ice characteristics

Ice area and concentration. The seasonal cycle of ice extent is represented optimally in SZT-IO, though it is not perfectly simulated. In particular, in SZT-IO, the mean winter ice extent is 4–10% higher than in the sensitivity runs, and, on annual average, the closest to observations. Correlation between model and observed monthly mean anomalies of SH total ice extent are slightly higher in SZT-IO than in the other simulations.

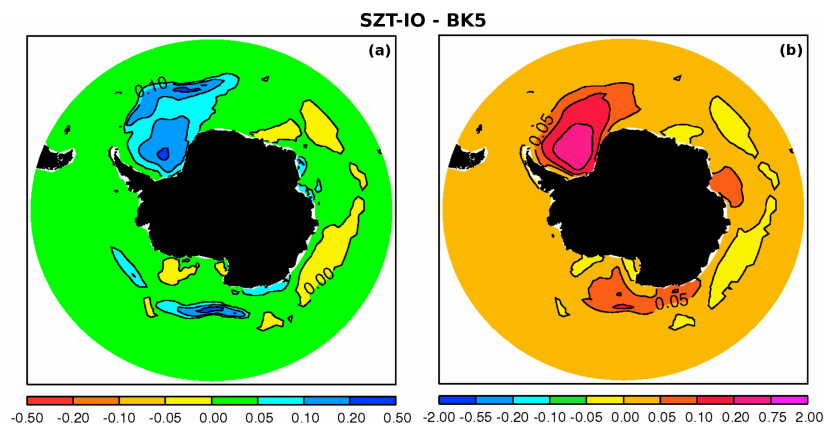


Figure 7.12: Annual mean 1979-2006 SZT-IO - BK5 differences in (a) ice concentration and (b) thickness (m) in the SH.

Ice concentration is typically higher by 0.1 – 0.2 in SZT-IO compared to the sensitivity runs. The amplitude of the differences among runs is markedly higher in the SH than in the NH. The geographical distribution of the differences is qualitatively similar in the three sensitivity runs (see Figure 7.12, panel a). In SZT-IO, associated to higher new ice salinity, the energy required to form new ice in open water and at the ice base is lower. This leads to higher open water closure rates which result in an increased sea ice concentration. In addition, the amount of winter bottom melt in SZT-IO is smaller than in other runs. The largest differences are found in the central Weddell Sea (0.1-0.5, spring) and close to the ice edge in the Weddell, Bellingshausen, Amundsen and Ross Seas (0.1-0.2, winter). In SZT-IO, the differences contribute to improve the simulated ice edge compared to passive microwave observations.

Ice volume and thickness. SH sea ice volume in the sensitivity experiments partly confirms the 1D findings described in the last paragraph of Section 3.6

(see Table 7.5) and underlines the importance of ice-ocean interactions. The ice growth rate is the largest in SZT-IO. This yields a seasonal maximum volume which is 18 – 28% larger in SZT-IO compared to the other simulations. The minima of ice volume are quite similar in all simulations, since, in SZT-IO, the ice melts more rapidly than in the other simulations. In spite of this, SZT-IO has the largest summer minimum volume. The interannual variations of the differences in ice volume between runs are small.

The patterns of the control-sensitivity differences in mean ice thickness are also similar among the sensitivity runs (see Figure 7.12, panel b). In SZT-IO, the ice thickness is generally larger, typically by 10-50 cm, than in the sensitivity runs, with a maximum difference just after melt onset in spring. Differences are the largest in the central Weddell Sea (10-50 cm), are slightly smaller in the Ross Sea (5-25 cm), extend along the coast of Adélie Land and are rather weak along the coast of East Antarctica (5-15 cm).

Sources and sinks of ice mass. In the SH, sources and sinks of ice mass are more diversified than in the NH. All along winter, ice not only grows in open water, at the ice base and through snow-ice formation but also surprisingly melts at the ice base. This occurs because of the combination of the large magnitude of the oceanic heat flux (see V08) and of the synoptic weather variability, which is characterized by the alternance between the transport of cold (warm) air masses from the continent (ocean). Though, in reality, basal ice growth / melt can only work alternatively, they artificially seem to occur simultaneously in the climatological monthly means.

In terms of sources and sinks of ice mass, differences among the runs are larger in the SH than in the NH and are dominated by basal melt. Though the mean seasonal cycles of the terms of tendency of ice thickness are shown for the Weddell Sea (see Figure 7.14) – where the differences among the runs are the largest and contribute the most to differences in SH ice volume – the following discussion will cover all regions of the Southern Ocean. The control-sensitivity differences are characterized by geographical distributions that are qualitatively similar among the sensitivity runs. Therefore, in Figure 7.13, only the SZT-IO - BK5 differences are shown. Unless otherwise stated, all the numbers presented hereafter refer to annual averages computed over the ice-covered regions of the SH.

The total amount of ice growth, i.e., due to basal growth, new ice formed in open water and snow-ice formation, is 1% larger in SZT-IO than in BK5 and PR and 11% smaller than in SZT-I. At the ice base, ice grows 5% (9%) more in SZT-IO than in BK5 and PR (SZT-I). The differences are due not only to smaller q in

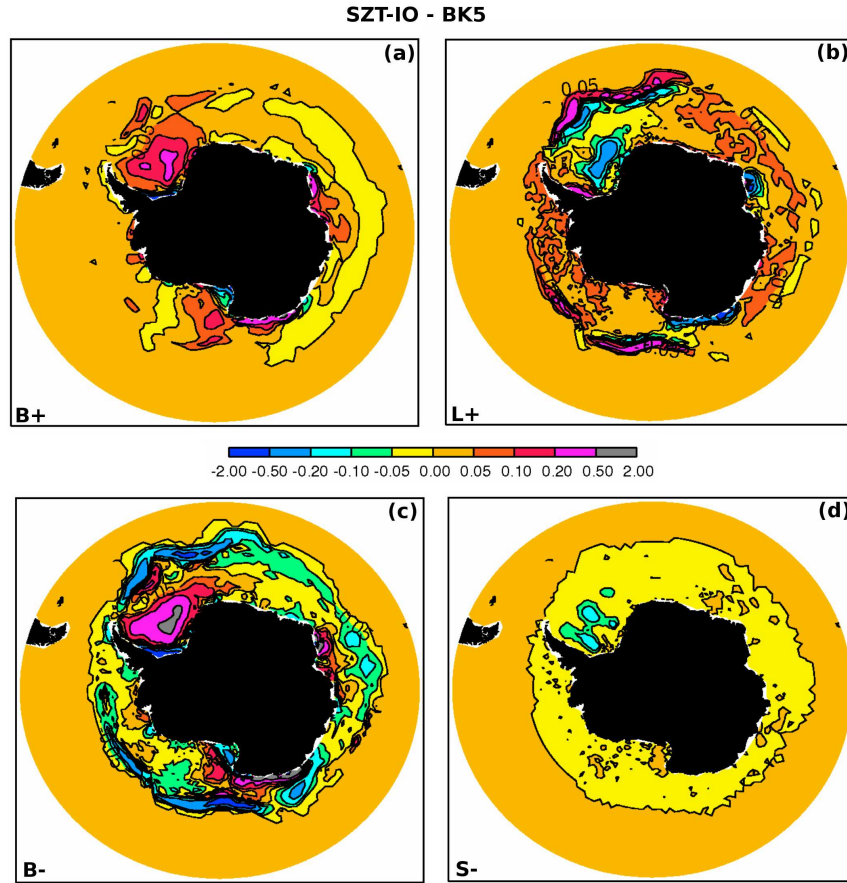


Figure 7.13: SZT-IO - BK5 differences in the different thermodynamic source / sink terms of ice mass (m/y), SH, 1979-2006 : (a) basal growth; (b) growth of new ice in open water; (c) basal melt; (d) surface melt.

SZT-IO, caused by higher new ice salinity, but also because of the oceanic heat flux, which is 11 – 15% smaller in SZT-IO compared to the sensitivity runs.

Differences in the amount of new ice growth in open water are within 2% among SZT-IO, BK5 and PR. These differences in SZT-IO from BK5 and PR are spatially averaged and result from the balance between, on the one hand, positive differences close to the ice edge and negative differences inside the pack and along the coasts of Antarctica, on the other hand. The former reflect the effect of the lower q and oceanic heat flux in SZT-IO and result in larger winter ice extent. The latter reflect the larger summer extent and the higher winter ice concentration in

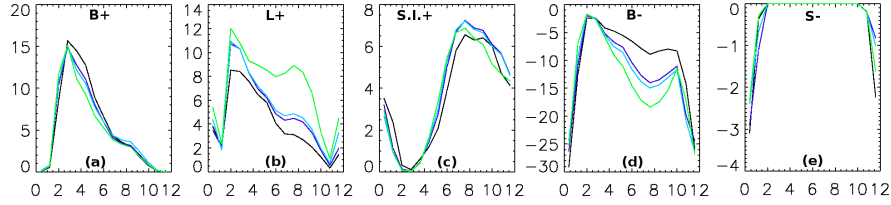


Figure 7.14: 1979-2006 annual mean seasonal cycles of the different thermodynamic source /sink terms of ice thickness (cm/month), averaged over the Weddell Sea, for SZT-IO (black), SZT-I (green), BK5 (dark blue), PR (light blue) : (a) basal growth; (b) growth of new ice in open water; (c) snow ice formation; (d) basal melt; (e) surface melt.

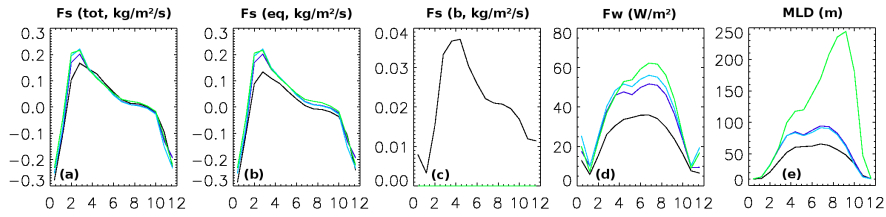


Figure 7.15: Same as Figure 7.10, but in the Weddell Sea.

SZT-IO, leaving less free open water available for new ice growth. The amount of new ice growth in SZT-IO is 37% smaller than in SZT-I. This combines the effects of lower q due to higher new ice salinity and smaller open-water areas in SZT-IO. Since the largest differences in ice growth from SZT-IO are in SZT-I, the role of the interactive salinity through changes in ice-ocean interactions must be important.

Snow ice growth is 3 – 5% smaller in SZT-IO than in the three sensitivity runs. This is particularly true in the Weddell Sea. In SZT-IO, ice is thicker than in the sensitivity runs and, hence, negative freeboards are less frequent.

The total amount of ice melt – due by 95% to basal melt and for the remainder to surface melt – is 6% smaller in SZT-IO than in BK5 and PR and 21% smaller than in SZT-I. Bottom melt is 6% (22%) smaller in SZT-IO than in BK5 (PR). Differences are the largest in fall, winter and spring and are associated to the smaller winter oceanic heat flux in SZT-IO. The amount of surface melt does not show differences among the runs that are significant at a large-scale.

7.6.2.2 Storage of fresh water and salt in sea ice

In Figure 7.16, the mean seasonal cycle of the total sea ice fresh water content, salt content and dilution potential are depicted for the SH. Due to the higher sea ice volume in SZT-IO, on annual average, the SH sea ice fresh water content is 20% (25%) larger than in BK5 (PR). Due to higher sea ice volume and higher mean salinity in SZT-IO, on annual average, the SH sea ice salt content is 34% (72%) larger than in BK5 (PR). The differences in fresh water contents dominate the differences in salt contents, and therefore BK5 (PR) underestimates the sea ice dilution potential by 16% (15%). As in the NH, differences in salt content dominate between BK5 and PR.

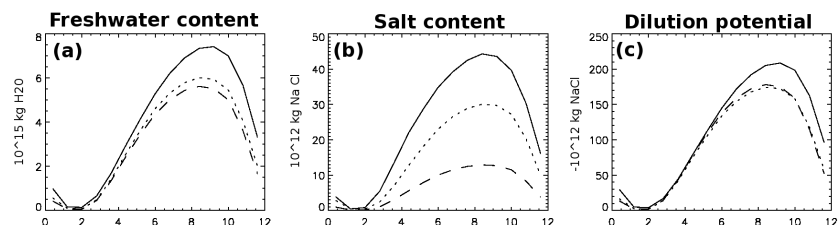


Figure 7.16: Mean 1979-2006 seasonal cycle of the total sea ice fresh water content (a), salt content (b) and dilution potential (c) as defined in Section 6.1.2 for the entire SH. Solid lines correspond to SZT-IO, dotted lines to BK5 and dashed lines to PR. Since SZT-I does not conserve salt, it makes no sense to show the associated diagnostics.

7.6.2.3 Ice-ocean interactions and characteristics of the ocean

As witnessed from the results described hereabove, the role of ice-ocean interactions and feedbacks is rather important in the SH.

Ice-ocean salt flux. First, during the growth season, the total ice-ocean salt flux (F^s) in SZT-IO is 12%, 18% and 36% lower than in BK5, PR and SZT-I (see Figure 7.15). In SZT-IO, the ice-ocean salt flux due to ice growth (F_{eq}^s) accounts for 64% of the total flux and brine drainage (F_b^s) accounts for the remainder. Differences in F_{eq}^s are larger than in F^s : compared to BK5, PR and SZT-I, F_{eq}^s is larger in SZT-IO, respectively by 66%, 73% and 80%. Though, as discussed for the NH, this salt is partly rejected back to the ocean by gravity drainage, this may be important in regions of concomitant ice production and export (i.e., in the Weddell Sea, along the Antarctic Peninsula, in the coastal polynyas along East Antarctica and along the Ross Ice Shelf). Finally, in summer, due to the

higher dilution potential in SZT-IO, the fresh water flux due to ice melt is larger by 9 – 13% than in the sensitivity runs.

Characteristics of the ocean. In the SH, the sea surface salinity (SSS) is large and geographical differences are small compared to the Arctic, situated between 33 and 35‰.

Creation and transport of ice with smaller dilution potential in SZT-IO stabilize the water column, inhibit winter deep convection and reduce the vertical exchanges of salt. This is the case in the Central Weddell Sea where the September mixed layer is reduced locally by 400 m in SZT-IO compared to BK5. Away from the ice edge in the South Atlantic and in the South East Pacific, the September mixed layer is shallower by 200 – 300 m in SZT-IO compared to the sensitivity runs, which is witnessed by the higher salinities at 200 m depth found in SZT-IO compared to BK5. In turn, the weakening of deep convection slightly reduces the strength of the Antarctic circumpolar current and may reduce the formation of bottom water masses.

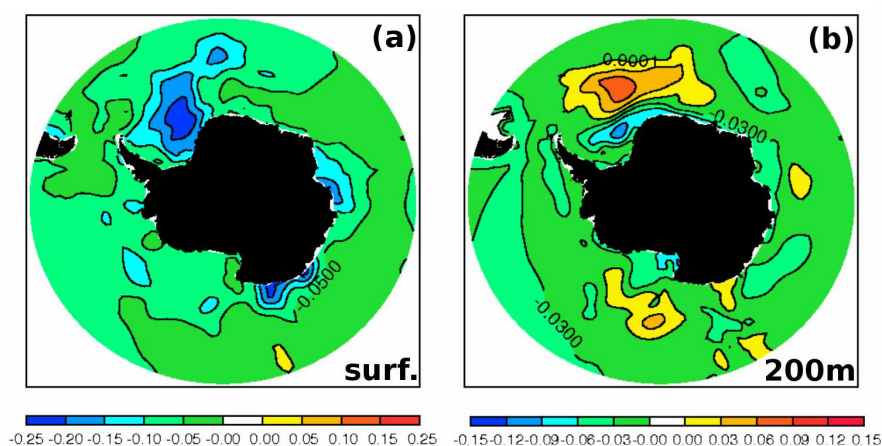


Figure 7.17: Differences in oceanic salinity (March) in the SH at the surface (a) and at 200 m depth (b), averaged over 1979-2006.

Figure 7.17 depicts the March oceanic salinity differences at the surface and at 200 m depth in the Southern Ocean, between SZT-IO and BK5, averaged over 1990-1999. In the SH, at the surface, waters are 0 – 0.25‰ fresher in SZT-IO than in the sensitivity runs. Differences among runs are maximum in spring and summer, in the zones where ice melts the most (i.e., Weddell Sea and off East Antarctica). This is associated to the later spring/summer retreat of sea ice pack with a higher dilution potential in SZT-IO, which increases the summer meltwater

input . In winter, fresher surface salinities are maintained by the reduced salt flux associated to ice growth in SZT-IO.

Along the ice edge, most differences between SZT-IO and BK5 are found in spring. This corresponds to the later retreat of thicker sea ice in SZT-IO. Typically, in spring, the negative (positive) SSS anomalies are found to the north (south) in SZT-IO where ice melts (is not melting yet).

7.6.2.4 Summary

In summary, in the SH , the relatively high salinity of new ice has two effects. First, it reduces the required energy to form new ice, which enhances the ice growth rate at the ice base and in open water. Second, it reduces the ice-ocean salt flux, maintains a shallow mixed layer and a relatively low oceanic heat flux, which in turn further increases (reduces) bottom growth (melt). Locally, in SZT-IO, the intense salt flux due to brine drainage deepens the mixed layer and induces bottom melt close to the coasts. This is the case along the Antarctic Peninsula and along Ross Ice Shelf.

7.7 Discussion, conclusions and perspectives

In this paper, we generalized the sea ice thickness distribution theory to include the ice salinity distribution. Large-scale variations in ice salinity result from horizontal transport, transport in thickness space, thermodynamic processes and mechanical redistribution. The most important term is associated to the halo-thermodynamic changes in ice salinity (i.e., brine entrapment and drainage). It is here formulated using a simple parameterization. In this parameterization, the sea ice vertically averaged ("bulk") salinity variations are due to salt entrapment during growth and snow ice formation, and brine drainage due to gravity drainage and flushing. New ice has a prescribed salinity following a field-derived salinity-thickness relation (Kovacs, 1996). The vertical salinity profile is computed as a function of the bulk salinity. In a 1D landfast sea ice test case, the results of the simple parameterization were found in close agreement with the sophisticated, multi-layer halo-dynamic VBF07 module, which was itself carefully validated against observations (Vancoppenolle *et al.*, 2006, 2007).

The salinity distribution was included in LIM3, a global, large-scale, dynamic-thermodynamic sea ice model with ice thickness, enthalpy and age distributions, coupled to the ocean model OPA9. OPA9-LIM3 was run in a 1970-2006 hind-cast, forced by a combination of daily NCEP/NCAR reanalysis data and various climatologies. The simulated salinity field was validated in two ways. First, the simulated mean seasonal cycle is in good agreement with available observations from ice cores. The main source of model-data discrepancy is due to the sometimes low statistical significance of the observed value. Second, the simulated salinities were compared and found in good agreement with the salinity-thickness relation of Cox and Weeks (1974), both in winter and summer. Scarcity of data renders an accurate model validation difficult. Nevertheless, the general features of the simulated spatio-temporal variations in sea ice salinity seem to be well captured and in agreement with present knowledge.

The simulated NH and SH salinity fields are quite different. In the NH, in winter, there is an important contrast in ice salinity driven by the relative importance of FY (thin) and MY (thick) ice, which are characterized by high and low salinities, respectively. In summer, this contrast is wiped out by the percolation of meltwater through the open brine network (flushing), which drastically reduces the sea ice salinity to around 2‰. In the SH, the salinity is higher and the amplitude of the seasonal and spatial variations is lower. This is due to the fact that, in the SH, (1) the contribution of young ice is more important, (2) the frequency of surface melt, and hence of flushing, is much lower, and (3) the contribution of snow ice formation (which traps large amounts of salt) to the global ice mass balance is higher.

The sensitivity of the simulated sea ice mass balance and exchanges with the ocean to the representation of the ice salinity in the model was studied. In the NH, the role of ice salinity variations mainly consists of a direct impact on the ice thermodynamic processes. Using an interactive salinity distribution reduces the energy required to melt a unit volume of sea ice. This resultingly increases winter bottom and lateral growth rates and produces thicker winter sea ice (from 10 cm to 1 m) in the Arctic Ocean. In summer, using an interactive salinity distribution facilitates the melt of new ice, which has a lower thermal inertia. This leads to higher open water areas, traps more solar radiation, increase the oceanic heat flux and enhances basal melt significantly. Finally, this yields differences in ice volume up to 11%. In addition, considering salinity variations reduces the ice-ocean salt flux during sea ice growth, which freshens the upper ocean, reduces vertical mixing and slightly weakens deep convection in the North Atlantic. Nevertheless, these changes are not sufficient to induce significant feedbacks on sea ice.

In the SH, because of the weaker vertical oceanic stratification, the influence of a variable salinity on ice-ocean interactions and feedbacks dominates. Using an interactive salinity distribution allows to grow more ice while rejecting less salt into the ocean. This maintains a relatively shallow mixed layer and a low oceanic heat flux, which, in turn, further enhances ice growth. This results in differences in ice volume up to 28%.

We have shown that the sea ice volume and coverage as well as the geographical distribution of ice thickness are sensitive to the model representation of ice salinity. Arctic sea ice is on a track toward a seasonal cover (see, e.g., Serreze *et al.* (2007)). According to climate models used for the last IPCC projections, this trend should keep on in the 21st century (e.g., Arzel *et al.* (2006); Zhang and Walsh (2006)). Therefore, as the contribution of FY becomes larger and larger, the Arctic sea ice salinity is increasing and will continue to increase in the future. In addition, the subsequent increase in sea ice growth will give more winter salt rejection to the ocean, deepen the mixed layer and, hence, reduce the amount of ice growth. Nevertheless, in present sea ice models, salinity is tuned for conditions that prevailed over the 20th century, but which are not adequate for the future sea ice cover. The models used for simulations of future climate should therefore take into account salinity variations. The present study encourages to maintain the effort to improve the representation of physical processes in large-scale sea ice models.

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Chapter 8

Conclusion

Parlez-moi de la pluie et non pas du beau temps,
Le beau temps me dégoûte et me fait grincer les dents,
Le bel azur me met en rage, ...

Georges Brassens

What are the recent sea ice mass and salinity variations and their interrelations? In order to answer this question, we progressed from an analysis of processes at local scales to their synthesis at large scales.

In this framework, two unique modelling tools were developed. The first one, **LIM1D**, is a one-dimensional halo-thermodynamic sea ice model designed to simulate the variations of snow depth and sea ice characteristics (ice thickness, vertical profiles of salinity and temperature) on a point, given the external forcing. The second one, **LIM3**, is a three-dimensional, dynamic-thermodynamic model including ice thickness, enthalpy, salinity and age distributions, as well as a complete description of brine physics, which is designed to simulate the global, large-scale sea ice characteristics given the external forcing. **LIM3** is coupled to the ocean general circulation model OPA9.

The study of the sea ice mass balance at local scales with **LIM1D** demonstrated the possibility to simulate, with a high level of accuracy, the seasonal cycle of the sea ice characteristics as well as the equilibrium MY ice thickness. Nevertheless, spatial variability in the validation datasets, simplicity of the snow model

and uncertainties in the longwave and oceanic forcings still constitute important shortcomings of the simulation. Large-scale simulation of sea ice characteristics with **LIM3** is significantly improved compared to earlier versions of the model, in particular in the NH, with snow depth, ice coverage, thickness, salinity and drift in good agreement with available observations over the last 30 years. There are several significant remaining uncertainties. Research should focus on, first, improving model physics (rheology, snow and meltwater, young ice), second, on collecting more data for model validation (in particular ice thickness) and, three, on reducing uncertainties in the atmospheric forcing.

Before the present thesis, not much had been done on the simulation of ice salinity. There are two exceptions: the work of Cox and Weeks (1988), which led to their empirical winter FY ice desalination model, and the first applications of mushy-layer framework to sea ice (Notz, 2005). Based on data analysis and on a physical-empirical modelling approach, the present study confirms quantitatively the scheme generally accepted to explain the ice salinity profiles found in-situ by Weeks and Ackley (1986). The FY, C-shape ice salinity profile results from the balance between brine entrapment at the ice base and gravity drainage. The shape of the MY equilibrium salinity profile reflects the impact of the summer percolation of meltwater through the open brine network. Basic understanding on desalination physics, in particular on gravity drainage and on flooding, snow ice and gap layer formation is needed for further model development.

The understanding of local salinity processes was simplified and summarized in a parameterization of brine entrapment and drainage that was included in **LIM3**. This generated the first assessment of the large-scale distribution of ice salinity, which shows hemispheric, seasonal, regional and subfloe-scale variations. In addition, the simulated salinity field is in agreement with information from the available ice core records and with the well-known thickness-salinity relationships of Cox and Weeks (1974). Nevertheless, the current parameterization is quite simple. Given the importance of the ice salinity on the simulated sea ice mass balance, the model would clearly gain in accuracy from a better representation of salinity.

Despite the role of ice salinity on its large-scale characteristics has been suspected from the very beginning of sea ice investigations, the present work provides its first systematic assessment. Using **LIM1D** in a central Arctic configuration, it was shown, first, that the salinity of new ice, higher than with a prescribed salinity profile, results in larger winter bottom growth. Second, fresher ice near the surface, due to flushing, increases the summer surface melt. For MY ice, the difference in growth dominates during the 5 first years, which leads to thicker ice. 3D studies with **LIM3** underline the role of feedbacks. In the Arctic, the

general 1D findings remain true. In addition, using a salinity distribution enhances the intensity of the ice-albedo feedback. Thin ice melts quicker, which increases the oceanic absorption of solar radiation and, in turn, enhances bottom melt. In addition, a varying salinity reduces vertical mixing, freshen the upper ocean, but not enough to significantly feedback on sea ice. In the Southern Ocean, the ice-ocean interactions are important. Using an interactive salinity distribution allows to grow ice faster, while keeping the ice-ocean salt flux relatively smaller than if salinity is prescribed. This reduces winter vertical mixing, the ocean-to-ice heat flux and, in turn, further increases ice growth.

In summary, including the variations in ice salinity in 1D and 3D sea ice models affects the simulated mean ice state. Furthermore, it alters, in a nonlinear way, the model sensitivity to changes in the external forcing as well as ice-ocean interactions. In light of this, an interactive salinity distribution should be included in future climate projections. Furthermore, with the advent of prognostic salinity schemes for sea ice models and recent knowledge on sea ice biogeochemistry gained in-situ, the perspective of including interactive biogeochemical components in global sea ice models, as well as their interactions with oceanic biogeochemistry and associated carbon cycle, becomes realistic.

Finally, another side of the present and future of the Arctic ice cover is suggested in this study. As the contribution of FY becomes larger and larger, the Arctic sea ice salinity is increasing and will continue to increase in the future. In relation to the latter, the freshwater / salt exchanges are and will keep on intensifying, with larger salt flux in winter (as more ice grows) and larger freshwater flux in summer (as more ice melts). This intensification of ice-ocean interactions provides previously unreported incentives to the positive feedbacks that are acting to accelerate the Arctic sea ice retreat. First, a more saline Arctic ice cover presumably enhances the ice-albedo feedback through the easier summer melt of thin ice. Second, the more intense salt rejections associated with the larger growth of new ice associated to high new ice salinity may excite the ice growth and vertical mixing feedback. In other words, the Arctic ice cover is not only becoming less extensive and thinner, it is also becoming characterized by different physics.

Appendix A

Acronyms

EVP	Elastic-viscous-plastic
FY	First-year
ITD	Ice thickness distribution
IPCC	Intergovernmental Panel on Climate Change
LIM	Louvain-la-Neuve ice model
LW	Longwave
MY	Multi-year
NEMO	Nucleus for european modelling of the ocean
NH	Northern hemisphere
OPA	Ocean parallélisé
PIZ	Perennial ice zone
PMW	Passive microwave
SAR	Synthetic aperture radar
SH	Southern hemisphere
SIZ	Seasonal ice aone
SW	Shortwave
THC	Thermohaline circulation
ULS	Upward-looking sonars
VP	Viscous-plastic

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