

## **Advances in understanding and parameterization of small-scale physical processes in the marine Arctic climate system: a review**

Timo Vihma<sup>1,2</sup>, Roberta Pirazzini<sup>1</sup>, Ilker Fer<sup>3</sup>, Ian A. Renfrew<sup>4</sup>, Joseph Sedlar<sup>5,6</sup>, Michael Tjernström<sup>5,6</sup>, Christof Lüpkes<sup>7</sup>, Tiina Nygård<sup>1</sup>, Dirk Notz<sup>8</sup>, Jerome Weiss<sup>9</sup>, David Marsan<sup>10</sup>, Bin Cheng<sup>1</sup>, Gerit Birnbaum<sup>7</sup>, Sebastian Gerland<sup>11</sup>, Dmitry Chechin<sup>12</sup>, and Jean Claude Gascard<sup>13</sup>

<sup>1</sup>Finnish Meteorological Institute, Helsinki, Finland

<sup>2</sup>The University Centre in Svalbard, Longyearbyen, Norway

<sup>3</sup>University of Bergen, Bergen, Norway

<sup>4</sup>University of East Anglia, Norwich, United Kingdom

<sup>5</sup>Bert Bolin Center for Climate Research, Stockholm, Sweden

<sup>6</sup>Department of Meteorology, Stockholm University, Sweden

<sup>7</sup> Alfred-Wegener-Institut Helmholtz-Zentrum für Polar- und Meeresforschung, Bremerhaven, Germany

<sup>8</sup>Max Planck Institute for Meteorology, Hamburg, Germany

<sup>9</sup>LGGE, Université de Grenoble, CNRS, Grenoble, France

<sup>10</sup>ISTerre, Université de Savoie, CNRS, Le Bourget-du-Lac, France

<sup>11</sup>Norwegian Polar Institute, Tromsø, Norway

<sup>12</sup>A. M. Obukhov Institute of Atmospheric Physics, Russian Academy of Sciences, Moscow, Russia

<sup>13</sup>Université Pierre et Marie Curie, Paris, France

## 1 **Abstract**

2 The Arctic climate system includes numerous highly interactive small-scale physical processes in the  
3 atmosphere, sea ice, and ocean. During and since the International Polar Year 2007–2009, significant  
4 advances have been made in understanding these processes. Here, these recent advances are reviewed,  
5 synthesized and discussed. In atmospheric physics, the primary advances have been in cloud physics,  
6 radiative transfer, mesoscale cyclones, coastal and fjordic processes, as well as in boundary-layer  
7 processes and surface fluxes. In sea ice and its snow cover, advances have been made in understanding  
8 of the surface albedo and its relationships with snow properties, the internal structure of sea ice, the  
9 heat and salt transfer in ice, the formation of super-imposed ice and snow ice, and the small-scale  
10 dynamics of sea ice. In the ocean, significant advances have been related to exchange processes at the  
11 ice-ocean interface, diapycnal mixing, double-diffusive convection, tidal currents and diurnal  
12 resonance. Despite this recent progress, some of these small-scale physical processes are still not  
13 sufficiently understood: these include wave-turbulence interactions in the atmosphere and ocean, the  
14 exchange of heat and salt at the ice-ocean interface, and the mechanical weakening of sea ice. Many  
15 other processes are reasonably well understood as stand-alone processes but the challenge is to  
16 understand their interactions with, and impacts and feedbacks on other processes. Uncertainty in the  
17 parameterization of small-scale processes continues to be among the greatest challenges facing climate  
18 modeling, particularly in high-latitudes. Further improvements in parameterization require new year-  
19 round field campaigns on the Arctic sea ice, closely combined with satellite remote sensing studies and  
20 numerical model experiments.

21

## 1 **1. Introduction**

2

3 Small-scale physical processes play an important role in the Arctic atmosphere – sea ice – ocean  
4 system, in particular at the interfaces and within boundary layers. Here, we define small-scale  
5 processes as such processes that need to be parameterized in climate or meteorological/oceanographic  
6 forecast models, with their current horizontal resolutions typically of the order of 1 to 100 km. These  
7 processes include (a) turbulent mixing in the atmosphere and ocean, (b) cloud and aerosol physics, (c)  
8 radiative transfer in the atmosphere, snow, ice, and ocean, (d) exchange of momentum, heat, and matter  
9 at air-sea, air-snow, air-ice, snow-ice, and ice-water interfaces, (e) small-scale mechanics in sea ice, (f)  
10 sea ice growth and melt, (g) formation of snow ice, super-imposed ice, and frazil ice, as well as (h)  
11 topographic effects on the atmosphere and ocean in coastal and continental shelf regions.

12 Better understanding and modelling of the Arctic sea-ice decline requires comprehensive, synthetic  
13 knowledge of small-scale processes in the atmosphere, snow, ice, and ocean. Such knowledge and  
14 related modelling capabilities are also prerequisites for a better understanding of the Arctic  
15 amplification of climate warming (Serreze and Barry, 2011), for which several processes have been  
16 proposed. Among them, the snow/ice albedo feedback has received most attention (e.g. Flanner et al.,  
17 2011; Hudson, 2011); in addition to its direct effect, it enhances the Arctic amplification by  
18 strengthening the water-vapour and cloud radiative feedbacks (Graversen and Wang, 2009). Further,  
19 feedbacks related to the shape of the temperature profile (Pithan and Mauritsen, 2014), the small heat  
20 capacity of the shallow stably-stratified boundary layer (Esau and Zilitinkevich, 2010) and increased  
21 fall-winter energy loss from the ocean (Overland et al., 2008; Screen and Simmonds, 2010a) tend to  
22 amplify the Arctic warming, as do the effects of aerosols. Black carbon aerosols have been suggested to  
23 reduce the surface albedo (e.g. Hadley and Kirchstetter 2012) and to warm the atmosphere (e.g. Quinn  
24 et al., 2008), while other aerosols affect the optical properties of the clouds and precipitation processes  
25 (e.g. Fridlind et al., 2012; Solomon et al., 2011). In addition to the above-mentioned small-scale  
26 processes, an increase in the advection of heat and moisture from lower latitudes also contributes to the  
27 Arctic amplification (Graversen et al. 2008; Kapsch et al., 2013). The relative importance of the above-  
28 mentioned processes in the Arctic is not well known.

29 Small-scale processes are most active and important in a layer that starts from the base of the ocean  
30 pycnocline and extends up to the top of the boundary-layer capping inversion in the atmosphere, as  
31 schematically illustrated in Figure 1. This layer extends down to 300 m into the ocean (Dmitrenko et  
32 al., 2008) and typically up to 100-1000 m in the atmosphere (Tjernström and Graversen 2009), but  
33 seasonal and regional variations are large. This layer includes large vertical gradients in temperature,  
34 salinity, air humidity, and wind/current speed; these gradients are generated by a complex interaction of  
35 large-scale circulation and small-scale processes. The large gradients are the driving force for turbulent  
36 and conductive exchange processes in a vertical direction. Further, the layer bounded by the ocean  
37 pycnocline and air temperature inversion includes major variations in radiative transfer. Compared to a  
38 dry atmosphere, the ocean, sea ice, snow, and clouds have a much higher longwave emissivity and a  
39 much lower shortwave transmissivity (Perovich et al., 2007a,b).

40 Over the central Arctic Ocean, small-scale processes are somewhat more tractable than near the coasts  
41 and continental shelves. In the latter regions, processes have a more profound three-dimensional  
42 structure, including orographic influences on the air flow (Renfrew et al., 2008) and, likewise,  
43 influences of the bottom topography and river discharge on local stratification and circulation in fjords  
44 and coastal waters (Cottier et al., 2007). In all regions, small-scale processes (e.g., radiative transfer,

1 cloud physics, and turbulent mixing) naturally include three-dimensional structures, but their net effect  
2 is mostly related to fluxes in the vertical; except for sea ice dynamics where many important small-  
3 scale processes act horizontally.

4 Processes on different scales are strongly interactive. On one hand, large-scale circulation and related  
5 lateral advection of heat and water vapour/freshwater in the atmosphere (Graversen et al., 2011; Sedlar  
6 and Devasthale, 2012; Kapsch et al., 2013) and ocean (Mauldin et al., 2010; Lique and Steele, 2012)  
7 strongly affect the boundary conditions for small-scale processes in the Arctic. On the other hand,  
8 small-scale processes modify the large-scale circulation via a number of interactive processes. For  
9 example, frictional convergence in the atmospheric boundary layer (ABL; see Table 1 for acronyms)  
10 affects the evolution of cyclones; while brine release from sea ice affects deep convection and vertical  
11 stratification in the ocean and, hence, the global thermohaline circulation. From the point of view of  
12 climate and operational modelling, the wide spatial and temporal range of important processes is a  
13 major restriction. The important scales range from micrometres (e.g. cloud physics) to thousands of  
14 kilometres (planetary waves). As models cannot resolve all scales of motion, many fundamentally  
15 important processes need to be parameterized using simplified physics and empirical relationships to  
16 resolved grid-scale variables. Variability on the ‘mesoscale’ (approximately 5 - 500 km in scale) is at  
17 the boundary of what is resolved and what must be parameterized in global numerical weather  
18 prediction and climate models. In the Arctic, this includes polar mesoscale cyclones, fronts, and  
19 orographic flows while there are also a wide range of oceanographic processes at these scales.

20 In subgrid-scale parameterizations the small-scale processes are presented as functions of those  
21 variables that can be resolved by the model grid. Subgrid-scale parameterization is one of the issues in  
22 climate models that are most prone to uncertainties and errors. This is for several reasons: (a) processes  
23 are often so complicated that it is not possible to accurately describe them solely on the basis of  
24 resolved variables, (b) models have errors in the resolved variables, (c) the resolved variables represent  
25 a large volume (grid cell) but there are large variations in the sub-grid scale processes inside the grid  
26 cell, (d) the physics of small-scale processes is often not sufficiently well known, (e) parameterizations  
27 require experimental data to constrain closure assumptions and the amount of such data may not be  
28 sufficient (in volume or in range), and (f) parameterizations are often tuned to make the overall  
29 performance of models better, according to Steeneveld et al. (2010), even when this makes the  
30 description of the particular small-scale process worse. The latter is a source of compensating errors  
31 and further inhibits model development, since improvements in one particular process via tuning often  
32 results in degradation in the overall model performance.

33 Present-day climate and numerical weather prediction (NWP) models as well as atmospheric reanalyses  
34 include large errors in small-scale processes. For example, in a validation of six regional climate  
35 models against year-round observations at the drifting ice station of the Surface Heat Budget of the  
36 Arctic Ocean (SHEBA), Tjernström et al. (2005) observed that the turbulent heat fluxes were mostly  
37 unreliable with insignificant correlations with observed fluxes and annual accumulated values an order  
38 of magnitude larger than observed. The downward shortwave and longwave radiation in the six models  
39 were systematically biased negative. Tjernström et al. (2008) showed that the radiation errors were  
40 strongly related to errors in cloud occurrence, heights, and properties (such as water and ice content and  
41 their vertical distribution). In an evaluation of the latest atmospheric reanalyses against independent  
42 tethered sounding data from the Central Arctic sea ice, Jakobson et al. (2012) showed that all five  
43 reanalyses included in the evaluation had large systematic errors. Even the best one (ERA-Interim of  
44 the ECMWF; Dee et al., 2011) suffered from a warm bias of up to 2°C in the lowermost 400 m layer  
45 and significant moist bias throughout the lowermost 900 m. The observed biases in temperature,

1 humidity, and wind speed were in many cases comparable or even larger than the climatological trends  
2 during the latest decades. This represents a major challenge for investigations of the recent Arctic  
3 warming, which are often based on atmospheric reanalyses. If the errors are solely systematic, then  
4 reanalyses may still yield useful information on trends, but for many variables and regions we lack the  
5 observations to determine if the errors are systematic or not.

6 The above-mentioned evaluation studies have addressed reanalyses and climate models, but little is  
7 known about the quality of operational weather forecasts in the central Arctic. Nordeng et al. (2007)  
8 reviewed the challenges in the field, and Jung and Leutbecher (2007) evaluated the ECMWF  
9 forecasting system, but quantitative comparisons between operational forecasts and observations taken  
10 at ice stations, research vessels and aircraft in the central Arctic have been very limited (Birch et al.,  
11 2009). More studies have been carried out for the Arctic marginal seas and coastal areas (Hines and  
12 Bromwich, 2008, Lammert et al., 2010, Renfrew et al., 2009b). About forecasting of polar lows and  
13 other mesoscale cyclones, see Section 2.3.2.

14 The most essential sources of information available from the Arctic Ocean are in-situ field  
15 observations, ice/ship-based or satellite remote sensing observations, operational analyses from NWP  
16 models, reanalyses, and results from model experiments dedicated to studies of small-scale processes.  
17 However, all these sources of information include uncertainties. Observations and data analyses  
18 focusing particularly on the Arctic are essential for an improved representation of processes within the  
19 Arctic, since the understanding obtained from lower latitudes may often not be valid for the Arctic. It  
20 should be noted that both atmospheric and ocean models apply parameterizations that are developed  
21 mostly on the basis of observations from low- and mid-latitudes. For example, stable stratification in  
22 the Arctic winter ABL is often long-lived, in contrast to the nocturnal stable ABL at lower latitudes; the  
23 latter is separated from the free atmosphere by the residual layer (Zilitinkevich and Esau, 2005). This  
24 makes the Arctic ABL more liable to the effects of propagating internal gravity waves. Also, the  
25 common presence of mixed-phase clouds in the Arctic marks a drastic difference from lower latitudes;  
26 observations of liquid water present in clouds at temperatures down to  $-34^{\circ}\text{C}$  during SHEBA (Beesley  
27 et al., 2000; Intrieri et al., 2002) demonstrated the need to develop better parameterization schemes for  
28 the ice and liquid water fractions (Gorodetskaya et al., 2008). In the past, forecast centres running  
29 global climate or NWP models have not necessarily paid enough attention to problems in physical  
30 parameterizations in the Arctic, but the situation is improving with the Arctic coming more into focus,  
31 driven by the worldwide attention to Arctic climate change and the increasing need for operational  
32 services in the Arctic.

33 During the International Polar Year 2007-2009 (IPY), a large effort was made in new field  
34 observations, data analyses and model experiments addressing small-scale processes in the Arctic  
35 atmosphere – sea ice – ocean system. One of the major efforts was the European project “Developing  
36 Arctic Modeling and Observing Capabilities for Long-term Environmental Studies” (DAMOCLES, in  
37 2005-2009), for which this Special Issue is dedicated. The project included an extensive amount of in-  
38 situ observations in the Arctic, supported by remote sensing, data analyses, and model experiments.  
39 During DAMOCLES, the drifting ice station Tara was a platform for oceanographic, sea ice, and  
40 meteorological research (Gascard et al., 2008). In addition, oceanographic and sea ice observations  
41 were carried out by several ships, meteorological research was made at ships, including short drift  
42 stations, by a research aircraft, and at coastal sites. Furthermore, drifting buoys, underwater gliders,  
43 and moorings collected extensive sets of oceanographic, sea ice, and meteorological observations. A  
44 DAMOCLES synthesis paper on the large-scale state and change of the Arctic climate system is  
45 presented in Döscher et al. (2014), while our focus is on small-scale processes. Small-scale physical

1 processes in the Arctic Ocean were reviewed already by Padman (1995), and other reviews on certain  
2 aspects on small-scale processes in high latitudes have been published more recently. Bourassa et al.  
3 (2013) focused on radiative and turbulent surface fluxes and remote sensing observations, and Heygster  
4 et al. (2012) addressed the DAMOCLES advances in sea ice remote sensing, which is related to micro-  
5 scale processes in snow and ice. Hunke et al. (2011) and Notz (2012) focused on sea ice physics and  
6 modelling, and Meier et al. (2014) reviewed the recent changes in Arctic sea ice and their impacts on  
7 biology and human activity. Rudels et al. (2013) reviewed the ocean circulation and water mass  
8 properties in the Eurasian Basin of the Arctic Ocean.

9 In this review we focus on the advances in research on small-scale processes in the Arctic since the  
10 start of the IPY, addressing physical processes only, and defining small-scale processes as those that  
11 need to be parameterized in climate models. Due to the above-mentioned recent papers, we will not  
12 address issues related to remote sensing of the ocean surface and sea ice. This review is organized in  
13 separate sections for small-scale processes in the atmosphere (Section 2), sea ice and snow (Section 3),  
14 and ocean (Section 4), with a cross-disciplinary synthesis, discussion, conclusions, and outlook in  
15 Sections 5 and 6. A reader not interested in specifics of all fields can skip some of Sections 2, 3, or 4.

## 16 17 **2. Atmosphere**

### 18 19 **2.1 Vertical structure and boundary-layer processes**

20 Many of the small-scale processes in the Arctic atmosphere closely interact with the vertical structure  
21 of the atmosphere, modifying it and being constrained by it. The vertical structure of the Arctic  
22 atmosphere is characterized by an ABL capped by temperature and specific humidity inversions  
23 (hereafter ‘humidity inversions’). The inversions are generated by the combined effects of the negative  
24 radiation balance of the sea ice surface, the direct radiative cooling of the air, and the horizontal  
25 advection from lower latitudes (Figure 1). The temperature inversion layer has a strong stable  
26 stratification, whereas the ABL stratification is typically stable or near-neutral, the latter stage is most  
27 often due to wind shear but, in conditions of large downward radiation, also due to surface heating.  
28 Above the ABL, mixed layers can also occur inside and below clouds (Section 2.2).

#### 29 30 2.1.1 Temperature and humidity inversions

31 In the vertical, temperature and humidity inversion layers are considered to be small-scale features  
32 although their spatial and temporal coverage can be extensive. Before the IPY, the knowledge on  
33 temperature inversion statistics over the Arctic Ocean was mostly based on radiosonde sounding data  
34 from coastal stations and the Russian drifting stations whose tracks were mostly in the sector of 120-  
35 240°E. The main findings were that surface-based inversions prevail during winter, extending to a  
36 height of typically 1200 m, with a typical temperature increase of 10-12 K (Kahl, 1990; Serreze et al.,  
37 1992). More recent ship and aircraft data show that in winter and early spring, especially during low  
38 temperatures, strong surface-based inversions exist also in the Atlantic sector of the Arctic Ocean  
39 (Lüpkes et al., 2012b). In summer, a slightly stable or near-neutral ABL prevails over sea ice with a  
40 capping inversion of variable depth. Tjernström et al. (2012) analysed soundings from four summer  
41 expeditions in the central Arctic, including SHEBA, and found a very persistent picture of near-neutral  
42 boundary-layer conditions with the layer depths ranging from ~ 200 to ~ 400 m. Tjernström and

1 Graversen (2009) analysed all the soundings from SHEBA and concluded that virtually all temperature  
2 inversions fall into either surface-based inversions or elevated inversions capping a near-neutral ABL,  
3 with no intermediary state. In winter, shifts between the two states are rapid, presumably depending on  
4 the presence of stratocumulus clouds, in which radiative processes and in-cloud turbulent dynamics  
5 together cause the shift of the inversion base from the surface to aloft (Tjernström and Graversen  
6 2009). There is also a pronounced annual cycle; in SHEBA data surface-based inversions were most  
7 common in winter and autumn, accounting for roughly 50% of the cases, whereas in summer  
8 practically all inversions were elevated ones on top of a near-neutral ABL. Since SHEBA, however, the  
9 occurrence of surface-based inversions in autumn has most probably decreased due to the sea ice  
10 decline.

11 Using the Atmospheric Infrared Sounder data, Devasthale et al. (2010) estimated that the area-averaged  
12 (70 to 90°N) clear-sky temperature inversion frequency is 70–90% for summer and approximately 90%  
13 for winter. Raddatz et al. (2011) found similar temperature inversion frequencies for a Canadian  
14 polynya region, whereas Tjernström and Graversen (2009) reported, based on SHEBA, that inversions,  
15 either surface based or elevated, are practically always present in the central Arctic. The spatial  
16 distribution of temperature inversions is inhomogeneous and strongly controlled by the surface type,  
17 the prevailing large-scale circulation conditions and by coastal topography (Pavelsky et al. 2011;  
18 Wetzel and Brummer 2011; Kilpeläinen et al. 2012).

19 The strongest temperature inversions are most often found in the lowermost kilometre, whereas the  
20 subsequent weaker inversions are nearly randomly distributed in the lowest 3 kilometres (Tjernström  
21 and Graversen 2009). The frequency, depth, and strength of temperature inversions have been found to  
22 correlate positively among each other, both spatially and temporally, and correlate negatively with the  
23 surface temperature (Devasthale et al. 2010; Zhang et al. 2011). However, the negative correlation  
24 between the inversion strength and surface temperature is noticeably weaker in summer (Figure 2)  
25 presumably due to a different formation mechanism: the summer inversion formation is probably  
26 dominated by warm air advection from lower latitudes, while in winter the inversions are often  
27 generated due to radiation loss at the surface (Devasthale et al. 2010). Vihma et al. (2011) reported that  
28 temperature inversions on the coast of Svalbard are strongly affected by the synoptic-scale weather  
29 conditions such as 850-hPa geopotential, temperature and humidity. In addition, during winter the  
30 temperature inversion strength over the ocean has a negative correlation with sea-ice concentration  
31 (Pavelsky et al., 2010).

32 A particular feature in the Arctic atmosphere that rarely, if ever, occurs at lower latitudes is that  
33 specific humidity very often increases across the ABL capping inversion, even for cases where the  
34 relative humidity in fact drops in the vertical (Tjernström et al., 2004). Importantly, this causes the  
35 entrainment of free troposphere air into the ABL to be a source of moisture, rather than a sink which is  
36 the case practically everywhere else on Earth. This contributes to the very moist conditions prevailing  
37 in the Arctic ABL. The frequency of specific humidity inversions has been found to be more than 80%  
38 throughout the year in the coastal Arctic, excluding the slightly lower summer frequencies on the  
39 Russian coast (Nygård et al. 2013). Vihma et al. (2011), for example, found humidity inversions to be  
40 present in all their tethered profiles taken in spring on the coast of Svalbard. Although summertime  
41 humidity inversions are slightly less frequent, they are stronger than in winter due to higher summer  
42 temperatures (Devasthale et al. 2011, Nygård et al. 2013). Humidity inversion climatologies based on  
43 radio-sounding data (Nygård et al. 2013) and satellite observations (Devasthale et al. 2011) differ  
44 notably, especially in the seasonal cycle of inversion properties, due to differences in the vertical  
45 resolution and methodology. Humidity inversions are nearly always found at multiple levels

1 (Devasthale et al. 2011; Nygård et al. 2013). Vihma et al. (2011) reported that, compared to  
2 temperature inversions, humidity inversions were on average thicker and had their base at a higher  
3 level. They concluded that this was mostly due to the role of the snow and sea ice surface as a sink for  
4 heat but commonly not for humidity (see also Persson et al., 2002). In other studies, however, humidity  
5 inversions have been found to usually coincide with temperature inversions (Sedlar et al. 2012;  
6 Tjernström et al. 2012). Differences in the observations may at least partly originate from different  
7 seasons (early spring in Vihma et al., (2011) and late summer in Tjernström et al. (2012)), while Sedlar  
8 et al. (2012) include SHEBA and several years of data from Barrow, hence possibly indicating that  
9 there may also be regional differences. A nonlinear relationship between humidity and temperature  
10 inversion strength is clearly found in all seasons except during summer (Devasthale et al. 2011).

11 Temperature and humidity inversions also have notable implications for the longwave radiation.  
12 Bintanja et al. (2011) and Pithan and Mauritsen (2014) demonstrated that atmospheric near-surface  
13 cooling efficiency decreases markedly with the temperature inversion strength, as the inversion layer  
14 damps the infrared cooling to space, and Boé et al. (2009) obtained analogous results for the role of air  
15 temperature inversion in reducing the radiative cooling of the ocean surface. Humidity inversions, in  
16 turn, can contribute up to 50% of the total amount of condensed water vapour in a relatively dry  
17 atmosphere in winter and spring, which can significantly influence the longwave radiative  
18 characteristics of the atmosphere (Devasthale et al. 2011), and they are presumably vital for the  
19 formation and maintenance of Arctic clouds (Section 2.2.1).

20 Inversions are a robust metric to evaluate the reproducibility of ABL processes in numerical models  
21 (Devasthale et al. 2011). Currently, Arctic temperature and humidity inversions are not realistically  
22 captured with respect to strength, depth and base height by operational weather forecasting models  
23 (Lammert et al. 2010), climate models (Medeiros et al. 2011), high-resolution mesoscale models  
24 (Kilpeläinen et al. 2012), or even reanalyses (Lüpkes et al. 2010; Jakobson et al. 2012; Serreze et al.  
25 2012). In particular, it is the nature of the Arctic atmosphere to contain multiple inversion layers and  
26 this is not reproduced in the models (Kilpeläinen et al. 2012). The errors in temperature inversion  
27 characteristics are related to deficits in the simulation of stable boundary layer (SBL) turbulence,  
28 clouds, radiative transfer, and surface energy budget (Lammert et al. 2010; Kilpeläinen et al. 2012) but  
29 are also sensitive to vertical resolution in models.

30

### 31 2.1.2 Stable boundary layer

32 Over sea ice in the central Arctic, the ABL is typically stably stratified during six winter months and is  
33 near-neutral or weakly stable during the other months (Persson et al., 2002; Section 2.1.1). Although  
34 cases of near-neutral stratification occur throughout the year, from the point of view of understanding  
35 and parameterization of the ABL over sea ice, the main challenges are related to stable stratification,  
36 and will be our focus here. The inner part of the Arctic Ocean, where the ice concentration is high and  
37 the surface is relatively flat and homogeneous, is ideal for SBL studies (e.g. Heinemann, 2008).  
38 Research on the Arctic SBL is strongly motivated by the major problems that climate models and  
39 reanalyses have in stably stratified conditions. Further, there are important feedback mechanisms  
40 related to the temperature inversion (Section 5.3).

41 A large part of the recent advance is still based on analyses of data from the SHEBA experiment.  
42 Important issues addressed in recent research include (a) scaling of SBL turbulence and (b) presence of  
43 turbulence under very stable stratification. Related to both (a) and (b), one of the main sources of  
44 uncertainty in SBL data analyses and modelling is the large scatter between experimental functions that

1 describe the stability-dependent relationships between vertical gradients and fluxes. Until recently,  
2 these formulae have not been based on Arctic data, but Grachev et al. (2007a, b) derived new formulae  
3 for stable stratification on the basis of SHEBA data. Considering (a), the traditional scaling, based on  
4 the Monin-Obukhov similarity theory, is such that the flux-gradient relationships depend on the  
5 stability parameter  $z/L$ , where the Obukhov length  $L$  depends on the turbulent fluxes. Mauritsen and  
6 Svensson (2007) and Grachev et al. (2012) demonstrated that, for moderately and very stable  
7 conditions, a scaling simply based on the vertical gradients (expressed in terms of the gradient  
8 Richardson number,  $Ri$ ) is better, because in such conditions the vertical gradients are large and their  
9 errors are relatively small. Further, there is no self-correlation between fluxes and  $z/L$ .

10 Considering (b), on the basis of SHEBA and mid-latitude data, Sorbjan and Grachev (2010) concluded  
11 that the necessary condition for the presence of continuous turbulence is that  $Ri < 0.7$ , which is a much  
12 larger value than expected on the basis of older studies. Intermittent turbulence is, however, present in  
13 the atmosphere even under very stable stratification with  $Ri \gg 1$ . This is related to the anisotropy of  
14 turbulence, which allows enhanced horizontal mixing, and to internal waves, which preserve vertical  
15 momentum mixing (Galperin et al., 2007; Mauritsen and Svensson, 2007). The energy of internal  
16 waves is associated with the turbulent potential energy (TPE), the importance of which has recently  
17 been better understood (Mauritsen 2007; Zilitinkevich et al., 2013), in addition to well-known  
18 importance of the turbulent kinetic energy, TKE. If TPE is taken into account, it follows that there is no  
19 critical  $Ri$ , and turbulence can survive in the very stable boundary layer. Another approach to treat the  
20 very stable stratification is based on the Quasi-Normal-Scale Elimination (QNSE) theory, which also  
21 takes into account waves and the turbulence anisotropy (Sukoriansky et al., 2005). This is enabled by  
22 the spectral nature of QNSE, based on ensemble averaging over infinitesimally thin spectral shells.  
23 Implemented in the NWP model HIRLAM, the QNSE approach yielded promising results for the  
24 Arctic, compared against SHEBA data (Sukoriansky et al., 2005).

25 Related to the division between weakly stable and strongly stable ABL, Lüpkes et al. (2008a) found  
26 that during SHEBA the lowest near-surface temperatures did not occur under calm conditions, but at a  
27 wind speed of about  $4 \text{ m s}^{-1}$ . Based on the results of a column (atmosphere and sea ice) model they  
28 found that this value can be considered as a lower threshold to generate sufficient mixing maintaining a  
29 close thermal coupling between the snow surface and near-surface air. Also Sterk et al. (2013)  
30 simulated the lowest near-surface temperatures in conditions of non-zero wind speed.

31 A low-level jet (LLJ) is a distinctive feature of the SBL; it is often generated by inertial oscillations  
32 related to the establishment of stable stratification, and it affects the SBL turbulence via top-down  
33 mixing due to the large wind shear below the jet core. An analytical model for a LLJ was presented  
34 already by Thorpe and Guymer (1977). Recently, ReVelle and Nilsson (2008) improved the description  
35 of frictional effects in such a model, and obtained promising results for the Arctic Ocean. New  
36 observations of LLJs over the Arctic Ocean include the work of Jakobson et al. (2013) based on  
37 tethered soundings at Tara. In their data, baroclinicity related to transient cyclones was the most  
38 important forcing mechanism for LLJs. On average, the baroclinic jets were strong and warm,  
39 occurring at lower altitudes than other jets, related among others to inertial oscillations and gusts.

40 Considering ABL modelling, it is well known that the ABL schemes commonly applied in climate  
41 models and NWP yield excessive heat and momentum fluxes in the SBL (Cuxart et al., 2006;  
42 Tjernström et al. 2005) typically resulting in a warm bias near the surface (Atlaskin and Vihma, 2012).  
43 In the Arctic, Byrkjedal et al. (2007) demonstrated the importance of a high vertical resolution: not  
44 surprisingly, model experiments with 90 levels in the vertical yielded much better results than those  
45 with 31 levels, the latter being typical for climate models contributing to the IPCC AR4. The high-

1 resolution simulations significantly reduced the warm bias and the excessive turbulent fluxes of heat  
2 and momentum that were present in the coarse resolution results over the Arctic Ocean.

3 A major challenge in ABL modelling is to better understand the interaction of turbulence, radiation,  
4 cloud physics, and thermodynamics of sea ice and snow. The work of Sterk et al. (2013), applying a  
5 single column version of the Polar Weather Research and Forecasting (Polar WRF), has yielded  
6 methodological advance in this respect. They used so-called process diagrams to indicate how the  
7 variations in parameter values in the schemes for various physical processes were related to differences  
8 in the model output.

### 10 2.1.3 Convection over leads, polynyas, and the open ocean

11 Although the Arctic ABL has a predominantly stable or near-neutral stratification, convection occurs as  
12 well. This is mostly due to the coexistence of ice and open water surfaces causing strong gradients in  
13 the surface temperatures. The influence of open water on the atmosphere strongly depends on the  
14 season, being largest in winter and smallest in summer (Bromwich et al., 2009; Kay et al., 2011).  
15 Convection may appear over leads, polynyas, and over the open ocean during cold air outbreaks. Thus  
16 there is a large variability in the involved spatial scales, and different parameterizations of turbulence  
17 are required. Convection over leads and polynyas (Figure 3) has been studied since 1970s (e.g. Andreas  
18 et al., 1979). As summarized by Lüpkes et al. (2012b) progress has been made during recent decades  
19 mainly with respect to the parameterization of energy fluxes at the lead surface. For example, the  
20 Andreas and Cash (1999) parameterization states that the transport of sensible heat is more efficient  
21 over small leads than over large leads due to the combined effect of forced and free convection.  
22 Recently, based on the lead distribution as analysed from a SPOT satellite image, Marcq and Weiss  
23 (2012) found that this dependence can increase heat fluxes over a large region of the Arctic by up to 55  
24 % since the small leads are dominating. Also Overland et al. (2000) (observations) and Lüpkes et al.  
25 (2008a) (1D air-ice modelling) point to the strong potential impact of atmospheric convection over  
26 leads on the surface energy budget. Both found that the net heat flux over an ice-covered region in the  
27 inner Arctic was close to zero due to a balance of downward fluxes during slightly stable near-surface  
28 stratification and upward fluxes from leads.

29 Although the effect of a single lead on the temperature is small, the integral effect of convection over  
30 leads can be very large: according to the model simulations by Lüpkes et al. (2008a), during polar night  
31 under clear skies, a 1% decrease in sea ice concentration results in up to a 3.5 K increase of the near-  
32 surface air temperature, if the air-mass flows over the sea ice long enough (48 hours). Polar WRF  
33 experiments by Bromwich et al. (2009) revealed that in winter over a region with an ice concentration  
34 of about 60%, the grid-averaged surface temperature increased by 14 K compared to an experiment  
35 with 100% ice concentration. For Antarctic winter, Valkonen et al. (2008) obtained a maximum of 13  
36 K sensitivity of the 2-m air temperature to the sea ice concentration data set applied (all based on  
37 passive microwave observations). A related modeling challenge is the formation of new ice in leads  
38 and polynyas (Figure 3; Section 4.1), which strongly affects the surface temperature, the release of  
39 latent and sensible heat, and further the evolution of the ABL (Tisler et al., 2008). Especially, the  
40 modelling of thin ice growth is difficult due to the required resolution, but also the relation between the  
41 transfer coefficients of momentum and heat/humidity still requires future work (Fiedler et al., 2010).

42 The height reached by convective plumes strongly depends on the width of the lead/polynya, wind  
43 speed, surface-air temperature difference, and the background stratification against which the  
44 convection has to work (e.g. Liu et al., 2006). On the basis of airborne observations and high-resolution

1 modelling, Lüpkes et al. (2008b, 2012b) concluded that convection over 1-2 km wide leads reached  
2 altitudes of 50 – 300 m depending on the boundary layer structure on the upstream side of leads. On the  
3 basis of aircraft in-situ, drop sonde, and lidar observations, Lampert et al. (2012) observed that over  
4 areas with many leads the potential temperature decreased with height in the lowermost 50 m, and then  
5 was nearly constant due to convective mixing up to the height of 100-200 m. When the leads were  
6 frozen and their fraction was small, however, an SBL extended up to a height of 200-300 m.

7 Ebner et al. (2011) showed by a modelling study that convective plumes generated over the Laptev Sea  
8 polynya influence atmospheric turbulence even 500 km downstream of the polynya, and Hebbinghaus  
9 et al. (2006) found that cyclonic vortices can be generated or intensified over polynyas due to  
10 convective processes. Such processes over large polynyas may be important with respect to the drastic  
11 changes in sea ice cover observed in recent years.

12 In models, difficulties arise in the treatment of plumes generated over leads, which interact with the  
13 stable or near-neutral environment when the convective internal boundary layer is growing (Figure 3).  
14 Only first attempts have been made to account for the nonlocal character of turbulent fluxes in the  
15 plume regions at higher ABL levels (Lüpkes et al., 2008b). Processes in the upper ABL need to be  
16 investigated in future also with the help of Large Eddy Simulation (LES). For example, Esau (2007)  
17 found that the structure of turbulent regimes over leads can be extremely complicated under light winds  
18 as often found in Arctic regions. This finding forms a challenge for future improved parameterizations  
19 of energy transports.

20 Compared to the conditions over leads and polynyas, deeper convection in the Arctic atmosphere takes  
21 place in cold-air outbreaks (CAOs) over the open ocean. Due to the Arctic warming, the atmospheric  
22 boundary layer temperatures during CAOs have increased (Serreze et al., 2011), but Vavrus et al.  
23 (2006) found by a modeling study that the number of CAOs will increase during the 21st century in  
24 several regions as, for example, over the Atlantic Ocean. On the basis of reanalysis data, Kolstad et al.  
25 (2009) concluded that seasonal and inter-annual variability of CAOs is mostly governed by the  
26 variability of the 700 hPa air temperature, T700, rather than by the sea surface temperature. Using a  
27 rough measure of CAO occurrence based e.g. on T700, Kolstad and Bracegirdle (2008) concluded that  
28 climate models broadly capture the observed climatology of CAOs, but differences from observations  
29 occur in areas where models have excessive sea ice cover. As energy fluxes are very large in CAOs and  
30 extensive ocean regions are affected, small differences in the CAO occurrence and properties may  
31 cause a large effect on the regional ocean-atmosphere heat flux. Furthermore, strong off-ice winds as  
32 being typical for CAOs have a large impact on the drift of sea ice in the marginal ice zone (MIZ),  
33 which in turn affects the CAO development. Thus it is important to investigate small-scale physical  
34 processes in CAOs such as ABL turbulence in strong convective regimes as well as cloud physics.

35 Lüpkes et al. (2012b) pronounce that the simplest possibility to successfully parameterize turbulent  
36 transport in a strong convective regime is to use closures allowing counter-gradient transport of heat.  
37 Applying a mesoscale model with different grid sizes, Chechin et al. (2013) found for idealized cases  
38 that the strength of the ice breeze developing in CAOs over open water downstream of the MIZ was  
39 strongly affected by the grid sizes: models with grid sizes larger than 20 km tend to underestimate the  
40 wind speed close to the ice edge. This finding confirms earlier results by Renfrew et al. (2009a,b) and  
41 Haine et al. (2009). Since the ice breeze occurring in a region of roughly 100 km width along the polar  
42 ice edges influences the energy fluxes, there might be a systematic underestimation of surface energy  
43 fluxes in large scale models.

1 One of the most striking small scale features during CAOs is the occurrence of roll convection, which  
 2 has been extensively studied in the last decades (Liu et al., 2006). There are, however, still fundamental  
 3 questions under discussion. Gryschka et al. (2008) found by an LES study that in case of strong surface  
 4 heating and weak wind shear, surface inhomogeneity in the MIZ is an important factor for the  
 5 generation of convection rolls. This finding stresses also the importance of a close-to-reality treatment  
 6 of the MIZ processes including the near surface-fluxes (see Section 2.1.4).

#### 7 8 2.1.4 Surface roughness and momentum flux

9 The drift speed of Arctic sea ice has increased during recent decades (Rampal et al., 2009; Spreen et al.,  
 10 2011). Increased wind speeds have contributed to the drift acceleration between 1950 and 2006  
 11 (Häkkinen et al., 2008), but not between 1989 and 2009 (Vihma et al., 2012). Instead, the recent  
 12 increasing trend in drift speeds is mostly due to ice becoming thinner and mechanically weaker  
 13 (Section 3.3.1). To reliably model the ice drift velocity field and ice export out of the Arctic, it is  
 14 essential to accurately parameterize the transport of momentum from the atmosphere to the sea ice.  
 15 Moreover, the friction at the surface determines the atmospheric cross-isobaric mass flux, sometimes  
 16 called Ekman transport, that is very important for the proper simulation of the lifetime of synoptic-scale  
 17 weather systems.

18 The momentum flux depends on the wind velocity, thermal stratification in the ABL, and aerodynamic  
 19 roughness of ice/snow surface, which can be expressed as a roughness length ( $z_0$ ) or drag coefficient  
 20 ( $C_{D10N}$  referring to that at 10 m height under neutral stratification). In addition to the skin friction over  
 21 smooth ice/snow surface, the aerodynamic roughness of sea ice is affected by factors generating form  
 22 drag: ridges, floe edges, and sastrugi (Andreas et al., 2010a,b; Andreas, 2011; Lüpkes et al., 2012a;  
 23 2013). This generates a challenge for operational modelling: the above-mentioned characteristics of sea  
 24 ice surface vary rapidly in time and often over small spatial scales, but they are difficult to observe by  
 25 remote sensing. Over broken sea ice cover, however, the form drag is mostly caused by floe edges,  
 26 whose occurrence is related to the sea ice concentration, which can be observed by remote sensing.

27  $z_0$  of sea ice can be calculated on the basis of tower or aircraft observations. However, the results are  
 28 not directly comparable, as tower observations are not necessarily representative for the wider  
 29 surroundings, where the occurrence of ice ridges, floe edges, and sastrugi may differ from that in the  
 30 footprint area of the tower. New results for the Arctic sea ice, based on the tower observations from  
 31 SHEBA, include those by Andreas et al. (2010a,b). A significant advance has been the better  
 32 understanding of the differences between  $z_0$  in winter and summer. For winter conditions, Andreas et  
 33 al. (2010a) propose a constant  $z_0$  for a large range of friction velocities, and argue that a former  
 34 stronger dependence on friction velocity found by Brunke et al. (2006) might have occurred due to a  
 35 fictitious self-correlation. Andreas et al. (2010b) addressed the Arctic summer, when open water is  
 36 present due to melt ponds and leads, and proposed  $C_{D10N}$  with a dependence on the sea ice  
 37 concentration. Lüpkes et al. (2012a) revised this dependence by including a drag partitioning concept  
 38 distinguishing between skin drag over sea ice and open water in melt ponds and leads and form drag  
 39 caused by the edges of ponds and leads. They proposed a hierarchy of drag parameterizations whose  
 40 complexity depends on the used background model (e.g., stand-alone atmosphere or coupled ocean-sea  
 41 ice-atmosphere model). Compared to pre-IPY results, the role of melt ponds in the parametrizations by  
 42 Andreas et al (2010) and Lüpkes et al., 2012a) is a new aspect. Lüpkes et al. (2013) showed on the  
 43 basis of sea ice concentration and melt pond fraction data obtained by MODIS (Rösel et al., 2012) that

1 the inclusion of the melt pond effect on roughness has a significant impact on the drag coefficients to  
2 be used in climate models.

3 It should be noted that NWP and climate models often apply  $z_0$  values over sea ice that are much larger  
4 than those suggested as mean values by field observations. Further, to avoid decoupling, models often  
5 apply some threshold values, e.g. a lower limit for the friction velocity. In general, a high  $z_0$  and other  
6 means to enhance turbulent mixing yield more Ekman pumping and a better evolution of synoptic-scale  
7 systems (Beare, 2007; Svensson and Holtslag, 2009). Few studies exist where the momentum flux in  
8 climate models is systematically evaluated. Tjernström et al. (2005) concluded that the momentum flux  
9 is systematically overestimated for five evaluated regional models. This overestimation leads to an  
10 enhanced mixing and is a root cause for many other systematic problems in NWP and climate models.

11 Compared to the large number of studies related to aerodynamic roughness, only few studies have  
12 addressed the effect of stratification on the wind stress over Arctic sea ice. Considering differences  
13 between sea ice and open water, the effects of stratification and roughness usually tend to compensate  
14 each other. At least for low wind speeds open water (leads, polynyas, and the open ocean) usually has a  
15 lower  $z_0$  than sea ice but for most of the year the stratification over open water is unstable, which  
16 enhances the vertical transport of momentum. Demonstrating the dominating effect of stratification, a  
17 larger momentum flux over open water than sea ice has been observed (Brümmer and Thiemann, 2002)  
18 and obtained in modelling studies (Tisler et al., 2008; Kilpeläinen et al., 2011). At a global scale,  
19 advances have also been made in studies of momentum flux over the open ocean (see Bourassa et al.  
20 (2013) for a review).

21 The surface momentum flux also affects drifting/blowing snow. Most of the recent research advances  
22 originate from Antarctica and Greenland, but the issue is relevant also for the Arctic sea ice: via  
23 redistributing the snow thickness, drifting/blowing snow further affects the locations of melt pond  
24 formation (Section 3.1). Andreas (2010a) showed that, under wind speeds strong enough for the  
25 occurrence of drifting snow, the  $z_0$  of snow-covered sea ice is independent of the friction velocity (see  
26 above), which is in contrast to many commonly applied parameterizations.

## 28 **2.2 Clouds and radiation**

### 30 2.2.1 Cloud physics

31 Clouds are ubiquitous in the Arctic. As mentioned in Section 2.1, clouds interact with the temperature  
32 and humidity inversions and affect the ABL stratification (Figures 1 and 4), and fog (sea smoke) is  
33 often formed over leads and polynyas (Figure 3). The cloud fraction has an annual cycle with a  
34 maximum in early autumn and minimum during late winter (e.g. Curry et al., 1996; Shupe et al. 2011).  
35 This has been observed since the beginning of the satellite era (Liu et al., 2012), yet atmospheric  
36 models continue to struggle with even this most first-order cloud property. An ensemble average of  
37 state-of-the-art CMIP3 climate models generally agree with satellite observations of the Arctic cloud  
38 fraction annual cycle. Individually, however, models display a substantial inter-model spread, largest  
39 during winter and smallest in summer, which dramatically biases their ability to capture the correct  
40 annual cycle amplitude and some models even have an inverse annual cycle with less clouds in summer  
41 and more in winter (Karlsson and Svensson, 2011). Summer clouds posed problems also for the  
42 Community Atmospheric Model version 4 (CAM4) (Kay et al., 2011), and simulation of clouds was  
43 one of the main problems in testing of the Polar WRF model against SHEBA data (Bromwich et al.,

1 2009) and recently against the Arctic Summer Cloud Ocean Study (ASCOS) data (Wesslen et al.,  
 2 2013) as part of the Arctic System Reanalysis effort. Models also have difficulties in representing the  
 3 correct amount and vertical distribution of cloud hydrometeor phase partitioning over polar regions,  
 4 under a wide range of annual temperatures. These biases lead to direct consequences for the surface  
 5 radiation budget, near-surface temperature and the lower ABL thermal stability and turbulent structure  
 6 (Tjernström et al., 2008; Birch et al., 2009; Karlsson and Svensson, 2011; Kay et al., 2011; Cesana et  
 7 al., 2012; Liu et al., 2012).

8 The difficulties in modelling clouds over the Arctic are related to the numerous interactive processes,  
 9 schematically illustrated in Figure 4 for mixed-phase stratocumulus (MPS) clouds. Even though cloud  
 10 fraction is relatively high year-round, Shupe (2011) has clearly shown that seasonally-dependent,  
 11 vertical cloud phase preferences exist. Liquid-only clouds rarely exist above 2 km above ground level,  
 12 occurring predominantly during the sunlit portions of the year. Unlike the rest of the globe, MPS clouds  
 13 tend to be the most common in the lower Arctic troposphere, except during winter and early spring  
 14 when ice-only clouds are somewhat more frequent. The MPS clouds have a profound impact on the  
 15 surface energy balance, since liquid water generates significantly more longwave radiation to the  
 16 surface than do ice clouds (Tjernström et al., 2008; Sedlar et al. 2011; Wesslen et al., 2013), and hence  
 17 on the surface melt and freeze (Figure 4). Hence, MPS clouds will be a focus here.

18 An obvious connection between cloud phase and atmospheric temperature is present. MPS clouds are  
 19 often the preferential cloud class when temperatures range between  $-15$  to near  $0^{\circ}\text{C}$  (Shupe, 2011; de  
 20 Boer et al., 2009), but liquid water has been observed in clouds at temperatures as low as below  $-34^{\circ}\text{C}$   
 21 (Intrieri et al., 2002). Complicating the matter, the presence of liquid droplets and ice crystals together  
 22 forms an unstable equilibrium due to the saturation vapour pressure differences of ice and liquid, the  
 23 Wegener-Bergeron-Findeisen (WBF) process (c.f. Morrison et al., 2012). Despite this instability,  
 24 liquid-topped clouds with ice and/or drizzle precipitating from this layer are the norm within the lower  
 25 Arctic troposphere from spring through autumn (Tjernström et al., 2004; de Boer et al., 2009; Shupe,  
 26 2011; Sedlar et al., 2011). Shupe et al. (2011) observed mean duration times of the order of 10 hours  
 27 for these cloud systems, but they may also occur as quasi-stationary systems persisting for days (Shupe  
 28 et al., 2008; Sedlar et al., 2011; Shupe, 2011).

29 The generally long lifetime of MPS clouds suggests that relative humidity with respect to liquid ( $\text{RH}_{\text{liq}}$ )  
 30 is kept high within and near the cloud layer. If  $\text{RH}_{\text{liq}}$  becomes sub-saturated in the presence of ice  
 31 crystals, liquid droplets must evaporate following the WBF process, and hence would cause a rapid  
 32 depositional ice growth and cloud layer glaciation. Instead Shupe (2011) has shown that in-cloud  $\text{RH}_{\text{liq}}$   
 33 and temperature distributions at a number of Arctic stations are in fact surprisingly similar, lending  
 34 support for a system that is both conditioned for, and dependent upon, mixed-phase clouds. In general,  
 35 stratiform clouds do not need large-scale updrafts, e.g. convection, to sustain them. Instead, these  
 36 clouds rely on cloud-driven (in-cloud production of) vertical motion where the small-scale dynamics  
 37 (turbulence) both depends on the presence of liquid, through the cloud-top cooling, but also supplies  
 38 the moisture that sustain that liquid layer.

39 Cloud top radiative cooling is typically very efficient as near-adiabatic liquid water content (LWC)  
 40 profiles are common in the Arctic (Curry, 1986; Shupe et al., 2008). Arctic MPS droplet radii generally  
 41 also increase with height (e.g. Curry, 1986) and droplet effective radii often range between 4 to 15  $\mu\text{m}$ .  
 42 Typical LWC in MPS peaks between  $0.1 - 0.2 \text{ g m}^{-3}$  (McFarquhar et al., 2007) and together with  
 43 relatively thin liquid layers (Shupe et al., 2008; Shupe 2011), cloud liquid water path (LWP) is often  
 44 below  $100 \text{ g m}^{-2}$  (de Boer et al., 2009; Sedlar et al., 2011; Shupe et al. 2011). In-cloud ice water

1 contents (IWC) are generally largest between cloud mid-level and base, decreasing upwards towards  
2 cloud top where they are initially formed (Shupe et al., 2008). Recent campaigns report a wide  
3 spectrum of ice crystal effective diameters, ranging from 20 – 60  $\mu\text{m}$  (McFarquhar et al., 2007; Shupe  
4 et al., 2008) and upwards of 100  $\mu\text{m}$  when falling through the subcloud layer (de Boer et al., 2009).

5 The ratio of LWC to total water content is often larger than 0.8 (McFarquhar et al., 2007; Shupe et al.,  
6 2008) indicating the resilience of cloud liquid despite near-constant drizzle and ice precipitation. In  
7 fact, de Boer et al. (2011) find evidence that liquid saturation occurs prior to ice crystal development  
8 even in a supersaturated environment with respect to ice. The authors suggest that ice nucleation  
9 mechanisms in Arctic MPS thus tend to be controlled by processes that rely on the presence of liquid  
10 condensate, further emphasising the importance of cloud motions in controlling the resilience of MPS.

11 In contrast to subtropical stratocumulus where decoupling between the surface and the cloud layer  
12 occurs during daytime as a part of a diurnal cycle, the Arctic ABL and sub-cloud thermodynamic  
13 structure often feature a persistent decoupling between the surface and the cloud layers (Shupe et al.  
14 2013) and the mechanisms are different. This decoupling appears to be most common during the cold,  
15 dark months, but occurs also during the transition and summer seasons (Kahl, 1990; Tjernström et al.,  
16 2004; Sedlar et al., 2011, 2012; Solomon et al., 2011; Tjernström et al., 2012; Shupe et al. 2013). Thus  
17 the surface-based moisture source for Arctic MPS is often missing (Figure 4). Sedlar and Tjernström  
18 (2009) and Sedlar et al. (2012) identified a common, persistent Arctic MPS cloud regime over the  
19 Arctic where the cloud layer is decoupled from the surface, a liquid cloud top extending above the  
20 stably stratified temperature inversion base, and ice crystals precipitating from the cloud. They  
21 hypothesize that the presence of specific humidity inversions, a common Arctic phenomenon (see  
22 Section 2.1.1), are vital to Arctic MPS survival. Surface turbulent heat and moisture fluxes are  
23 generally small over sea ice (Persson et al., 2002; Tjernström et al. 2005, 2012), and ice crystals falling  
24 from the cloud into the sub-saturated sub-cloud layer will further enhance decoupling due to cooling  
25 from ice crystal sublimation (Figure 4; Harrington et al., 1999). Thus instead of moisture originating  
26 from the surface, the increased humidity within the inversion structure may be the moisture source  
27 which sustains the cloud system (Solomon et al. 2011; Sedlar et al. 2012).

28 Turbulent kinetic energy is generated near cloud top (Shupe et al., 2012, 2013) due to parcel buoyancy  
29 differences initiated by radiative cloud-top cooling, causing top-down overturning circulations and  
30 vertically turbulent motions. Within these turbulent eddies, condensation and evaporation compete  
31 (Figure 4), often with condensation (evaporation) occurring in turbulent updrafts (downdrafts) near  
32 cloud top (Shupe et al., 2008). These mechanisms also occur within, and sustain, warm subtropical  
33 stratocumulus. The key difference in the Arctic is the presence of liquid and ice simultaneously. Shupe  
34 et al. (2008) show that ice production is generally limited to cloud-generated updrafts that increase the  
35 supersaturation with respect to ice. When downdrafts were observed, ice production generally ceased  
36 and fewer ice crystals grew to large sizes and fell from the still-present, yet slightly more tenuous,  
37 liquid layer. Hence the coexistence of liquid and ice is intimately linked to cloud scale motions, which  
38 in turn depends on the presence of liquid water.

39 Tjernström (2007) suggested that most of the boundary-layer turbulence in the Arctic is in fact  
40 generated by boundary-layer clouds, at least in summer. If the in-cloud turbulence production is strong  
41 and stratification below the cloud layer is weak, the cloud-induced turbulent eddies may penetrate to  
42 the surface, hence affecting the surface fluxes of momentum, heat, and moisture (Figure 4). Cloud-  
43 generated mixing is found beneath cloud base, but the extent to which these turbulent motions reach the  
44 surface is often limited by a sub-cloud stable layer (Shupe et al., 2013; Sedlar and Shupe, 2014) and is

1 also dependent on the distance from the cloud base to the surface and the sublimation of precipitation  
2 in the layer below the cloud base (Figure 4). Hence the strongest but also most variable turbulence  
3 generation is due to buoyant cloud overturning due to cloud top cooling, which generates eddies that  
4 often persist below the cloud base. Mechanical generation of turbulence at the surface, on the other  
5 hand, is seldom very strong and intense buoyant mixing is essentially absent over sea ice (other than  
6 over winter leads/polynyas), and the ABL is therefore most often shallow. Coupling, or the lack  
7 thereof, of MPS clouds to the surface and surface fluxes is therefore more often dependent on if the  
8 cloud generated turbulence can reach down to the ABL or not, rather than the other way around. This in  
9 turn is sensitive to the cloud generated turbulence but also to the cloud base height (Figure 4;  
10 Tjernström et al. 2012; Shupe et al. 2013; Sotiropoulou et al. 2014).

11 Spectral analysis of in-cloud vertical velocities reveals only modest changes to the cloud-generated  
12 temporal frequencies and horizontal wavelengths of vertical velocity when the cloud layer transitions  
13 between a surface-cloud coupled and decoupled state (Sedlar and Shupe, 2014); the authors conclude  
14 that the surface-cloud coupling state is therefore a result of the cloud processes and not dependent on  
15 the turbulence generated near the surface. Analysis of winter soundings from SHEBA in Tjernström  
16 and Graversen (2009) additionally shows how the boundary layer structure changes are almost binary  
17 between a well-mixed state, similar to summer conditions when clouds containing liquid water are  
18 present, and a distinct surface inversion structure when clouds are either absent or optically thin.

19 In terms of temperature, the radiative cooling from the liquid cloud top (Harrington et al., 1999)  
20 dominates over other local processes and hence, in the absence of frontal passages or other large-scale  
21 controls, cloud droplets will continuously form to replace the water that precipitates out. Cloud droplets  
22 can persist as long as a moisture source is present. The presence of humidity inversions near cloud top  
23 provide such a source (Figure 4), and Solomon et al. (2011) describe how cloud-generated vertical  
24 motions, and small but appreciable droplet condensation above the temperature inversion base, create  
25 the link between the cloud layer and the stable upper entrainment zone. This is a feature unique to the  
26 low-level Arctic thermodynamic structure, not observed in lower latitudes where large-scale subsidence  
27 generally prohibits humidity increases near cloud top. Furthermore, this situation is maintained by ice  
28 crystal formation and fallout (Shupe et al., 2008), effectively limiting the LWC near cloud top.

29 In addition to moisture, clouds need aerosol particles on which to condense and freeze (Figure 4).  
30 These cloud condensation nuclei (CCN) and ice nuclei largely determine the cloud's microphysical  
31 structure and hence its radiative properties. Over the Arctic, where local sources of pollution generally  
32 do not exist, transport in the region is considered a large contributor to the concentration and  
33 composition of CCN and ice nuclei (e.g. Shaw, 1975). In winter, when the ocean is ice covered, there is  
34 a substantial transport of aerosols and aerosol precursor gases into the Arctic (Barrie, 1986; Garrett and  
35 Zhao, 2006; Lubin and Vogelmann, 2006). In summer, the meridional transport is smaller and the  
36 formation of low clouds and fog at the MIZ, as sub-Arctic marine air adjusts to the frozen or melting  
37 surface, forms an effective filter for the transport of aerosols in the lower troposphere. Thus in the  
38 summer boundary layer the aerosol concentrations are generally very low compared to further south  
39 (Tjernström et al. 2014), while transport of aerosols from lower latitudes may occur at higher  
40 elevations (Lance et al., 2011). While the ocean surface is more exposed in summer, local production  
41 of aerosols may be important (Tjernström et al., 2014). Low aerosol concentrations and low  
42 temperatures both contribute to a preference for optically thin clouds and also promote precipitation  
43 formation.

44

1 Historically many models, especially weather forecast models, such as the ECMWF model, distinguish  
2 between cloud liquid and ice based only on temperature, having often failed to maintain liquid in very  
3 cold winter clouds (e.g. Beesley et al. 2000; Tjernström et al., 2008). Recently more advanced moist  
4 physics has made its way into state-of-the-art climate and weather forecast models (Meehl et al., 2013).  
5 However, while being more physically based, it has been difficult to properly tune such schemes to  
6 work well in all seasons and under all conditions. Tjernström et al. (2008) showed that models with  
7 more advanced cloud physics schemes generally did not perform better than those with simple  
8 temperature-schemes. In an evaluation of ERA-Interim and two versions of the Arctic System  
9 Reanalysis (ASR) against the ASCOS data, it was found that ERA-Interim more faithfully retained the  
10 observed Arctic MPS in spite of its much simpler temperature dependent formulation, albeit not  
11 necessarily for the right reasons (Wesslén et al., 2013).

### 13 2.2.2 Cloud-radiation interaction

14 The central Arctic imposes unique boundary conditions on both shortwave (solar) and longwave  
15 (infrared) radiative transfer, controlled by the large seasonal variations in the incoming fluxes and a  
16 wide range of surface albedo conditions (Section 3.1.2). The presence of cloud cover impacts radiation  
17 reaching the surface in two competing ways. First, cloud hydrometeors absorb longwave radiation,  
18 increasing the emissivity relative to a clear-sky atmosphere. This results in a net warming effect at the  
19 surface, especially over the Arctic where clear-sky effective emissivity is generally low, but  
20 simultaneously leads to cooling of the upper portion of the clouds. Conversely, clouds reflect incoming  
21 shortwave radiation to space resulting in a net surface cooling effect. Over the Arctic, the efficiency of  
22 shortwave cloud cooling is further limited by relatively large solar zenith angles (SZAs) and surface  
23 albedos; the latter is often as high as that of the overlying cloud. In fact, it still remains uncertain  
24 whether the net radiative effect of clouds in summer is to cool the surface over the large-scale Arctic  
25 basin, even though observations from SHEBA suggest a net cloud cooling effect during June and July  
26 (Intrieri et al. 2002; Shupe and Intrieri 2004). In an Arctic-wide sense, this net cloud effect is  
27 significantly connected to time of year, geographic location and surface albedo, notwithstanding the  
28 cloud physical properties.

29 The surface energy residuals, available for melting or freezing of the ice, are therefore strongly  
30 modified by the cloud radiative forcing. Surface energy budget analysis during the end of the 2008 melt  
31 season, towards the initiation of freeze up, during ASCOS demonstrated the delicate interplay of  
32 clouds, radiation, turbulence, and heat conduction in snow and ice (Sedlar et al. 2011). A week-long  
33 delay of the autumn freeze-up was realized through the manifestation of a positive longwave cloud  
34 radiative forcing of about  $70 \text{ W m}^{-2}$ , while the shortwave radiative cooling was limited to about  $-40 \text{ W}$   
35  $\text{m}^{-2}$  by surface albedo and SZA constraints. Net surface energy residuals, however, were significantly  
36 reduced by redistribution of heat and moisture via near-surface turbulence and heat conduction in  
37 snow/ice. The increase of the surface albedo, that eventually put the energy balance beyond recovery,  
38 was not gradual but a result of heavy frost formation and melt pond freezing during a short colder  
39 period with new snowfall (Sedlar et al. 2011; Sirevaag et al. 2011; Tjernström et al. 2012). The onset of  
40 freeze up was not realized until the low-level Arctic MPS became tenuous and cloud LWP decreased  
41 below  $20 \text{ g m}^{-2}$  – essentially diminishing the cloud greenhouse effect.

42 Comparing various climate models, the monthly averaged spread in LWP and ice water path (IWP) in  
43 the Arctic can be as large as a factor of three (Karlsson and Svensson 2011). Such variability inherently  
44 results in differences in cloud fraction as well as in the cloud-radiation interaction (Karlsson and

1 Svensson 2011). Tjernström et al. (2008) identified significant biases in several regional climate model  
 2 simulations of surface radiative fluxes during SHEBA. Both downwelling shortwave and longwave  
 3 radiation were negatively biased, while the bias magnitudes varied depending on the model. Tjernström  
 4 et al. (2008) found a significant underestimation (overestimation) in cloud LWP above (below) 20 g  
 5 m<sup>-2</sup>. Conversely, nearly all models underestimated the IWP and there were clear biases in the model  
 6 simulations of liquid to total cloud water path. The authors speculated that the biases in downwelling  
 7 longwave radiation might be due to an absence of sufficient liquid water in winter and that the  
 8 downwelling shortwave radiation bias was due to too opaque clouds, i.e. too high cloud albedo.  
 9 However, even when the actual errors in LWP and IWP were cancelled in the analysis a bias remained.  
 10 Thus, even if the distribution of ice and liquid were properly resolved, the modeled cloud-radiative  
 11 interaction tends to be misrepresented, and this error will propagate to surface radiation balance errors  
 12 for the ice and the ocean in coupled Earth System Models. These results point at the importance of a  
 13 proper handling of the aerosol/cloud/radiation feedback in resolving the proper radiation balance at the  
 14 surface (Section 5.3).

## 17 **2.3 Partly resolved processes**

### 18 2.3.1 Coastal and fjordic features

19 Coastal regions and in particular coastal mountain ranges can have a pronounced impact on the  
 20 mesoscale and boundary-layer meteorology of the adjacent coastal waters. This impact arises from the  
 21 combined effects of orography and spatial differences between the surface temperatures of snow/ice-  
 22 covered land, sea ice, and the open ocean. Considering orographic effects, when the wind is flowing  
 23 towards a barrier it must either rise over it or is distorted by it, i.e. it turns to flow along the coast as a  
 24 barrier wind or related feature such as a tip jet (common near the southern tip of Greenland). On the  
 25 downstream side of a barrier there is often some sort of orographic-forcing mechanism leading to  
 26 mesoscale features such as gap winds, katabatic winds, foehn winds or wake effects. The surface  
 27 temperature differences affect the thermodynamics of the ABL and further the wind field, sometimes  
 28 also generating mesoscale circulations. All of these mesoscale phenomena are only partially resolved in  
 29 current climate models and global NWP models; although NWP models can adequately simulate these  
 30 features if appropriate parameterizations are used and the grid size is sufficiently small.

31 Complex small-scale processes over Arctic coastal regions, including fjords, have received increasing  
 32 attention, especially around Greenland and Svalbard. During the IPY, the Greenland Flow Distortion  
 33 Experiment (GFDex) (Renfrew et al., 2008) and the Norwegian IPY-Thorpex Experiment (Kristjansson  
 34 et al., 2011) both examined such coastal phenomena through aircraft observations and numerical  
 35 simulations. The first comprehensive observations of barrier winds off southeastern Greenland are  
 36 documented in Petersen et al. (2009). They find barrier-effect enhancements of up to 20 m s<sup>-1</sup> and peak  
 37 wind speeds of up to 40 m s<sup>-1</sup>. The structure of the barrier winds was strongly dependent on the  
 38 synoptic-scale situation, often consisting of a cold barrier jet undercutting a warmer maritime air mass  
 39 and generally with a significant ageostrophic component of the flow. A climatology of these barrier  
 40 winds shows that they occur typically once a week, but with a large interannual variability determined  
 41 primarily by the broader-scale situation (Harden et al. 2011). Off SE Greenland there are two distinct  
 42 areas of occurrence (Harden et al. 2011). Idealized numerical simulations (Harden and Renfrew 2012)  
 43 and reanalyses work (Moore 2012) have shown that these two areas are related to two areas of steep  
 44 topography, separated by a major fjord. In SE Greenland barrier winds are known to play a key role in

1 generating a fjordic ocean circulation leading to submarine melting and thus the rapid retreat of ice  
2 shelves that is now being seen there (Straneo et al. 2010).

3 The first in situ observations of a tip jet off Cape Farewell, Greenland, documented near-surface winds  
4 of over  $35 \text{ m s}^{-1}$  and peak jet winds of almost  $50 \text{ m s}^{-1}$  (Renfrew et al. 2009a); while a dynamical  
5 analysis of these events showed their characteristic curve around the ‘tip’ was associated with a  
6 collapse in the cross-jet pressure gradient as the barrier decreases in height (Outten et al. 2009, 2010).  
7 Tip jets are also found off Svalbard (e.g. Reeve and Kolstad 2011), and over the Bering Sea (Moore  
8 and Pickart 2012); while gap flows were observed by an instrumented aircraft in the Svalbard region  
9 during the Norwegian IPY-Thorpex experiment (Barstad and Adakudlu 2011).

10 There are generally very high winds associated with all of these coastal jet features, so consequently  
11 there are elevated momentum fluxes and often elevated heat and moisture fluxes, depending on the  
12 source of the air, i.e. the air-sea temperature difference. Petersen and Renfrew (2009) provide  
13 observations from six GFDex flights into tip jets and barrier winds using the eddy covariance method  
14 and find fluxes up to  $1.9 \text{ N m}^{-2}$  (momentum)  $300 \text{ W m}^{-2}$  (sensible heat) and  $300 \text{ W m}^{-2}$  (latent heat).  
15 These are amongst the highest fluxes ever directly measured and certainly significant enough to lead to  
16 enhanced ocean mixing, water mass changes and potentially circulation changes in the ocean (e.g.  
17 Våge et al. 2008; Haine et al. 2009, Sproson et al. 2010). Although large air-sea heat fluxes are not  
18 always the case; associated with Greenland’s easterly tip jets the heat fluxes tend to be more moderate  
19 and are not associated with the deep open ocean convection events that tend to occur in the SE  
20 Labrador Sea (Sproson et al. 2008).

21 The spatial variability of atmospheric variables within a fjord may be very large (Figure 5). For  
22 Svalbard fjords, Kilpeläinen et al. (2011) reported that variability can reach levels comparable to the  
23 synoptic-scale temporal variability. The contribution of surface type to the spatial variability of  
24 turbulent heat fluxes increases with increasing air-sea temperature difference and typically dominates  
25 over topographic effects. On the other hand, the effect of topography dominates over surface type for  
26 the spatial variability of wind speed and momentum flux (Kilpeläinen et al. 2011). Realistic  
27 parameterization of turbulent fluxes is a challenge in a fjord as the Monin-Obukhov similarity theory  
28 has limitations in this environment. The combination of topographic effects and wave influence often  
29 causes significant crosswind momentum transfer, and sometimes also upward momentum transfer,  
30 which invalidate conventional stability and scaling parameters (Kilpeläinen and Sjöblom 2010; Kral et  
31 al. 2013). Monin-Obukhov similarity theory has, however, been found to be applicable during  
32 moderate or high wind speeds when the wind direction is along the fjord axis (Kilpeläinen and Sjöblom  
33 2010; Mäkiranta et al. 2011; Kral et al. 2013), which resembles results from valleys. The non-  
34 dimensional wind gradients in Arctic fjords have been found to be smaller than predicted by traditional  
35 empirical similarity functions, indicating a higher momentum flux than expected from the vertical wind  
36 shear in the surface layer (Kilpeläinen and Sjöblom 2010; Mäkiranta et al. 2011; Kral et al. 2013). The  
37 non-dimensional temperature gradients, in turn, have generally higher values than suggested by the  
38 traditional empirical similarity functions in unstable conditions, indicating less efficient sensible heat  
39 transport over fjords (Kilpeläinen and Sjöblom 2010; Kral et al. 2013). In stable conditions, however,  
40 more efficient mixing of sensible heat than predicted has been reported in a fjord environment by  
41 Mäkiranta et al. (2011). They suggest that in stable conditions the wind shear above the boundary layer  
42 provides a non-local source for the turbulence which enhances the mixing over the fjord. Their  
43 interpretation was supported by tethered observations of Vihma et al. (2011): LLJs were often lifted  
44 above the cold-air pool on an ice-covered fjord (Kongsfjorden). The presence of sea ice cover was  
45 found as a very important factor determining whether a katabatic flow can reach the fjord surface or be

1 elevated above the stable boundary layer (Vihma et al. 2011). Effects of sea ice cover on spatial  
2 variations in the ABL over a Svalbard fjord were also detected by Láska et al. (2012).

3 Orographic effects are sometimes responsible for the genesis of polar mesoscale cyclones, e.g. in the  
4 case of lee cyclones southeast off Greenland. In most cases, however, polar mesoscale cyclones are not  
5 directly related to orographic forcing and are discussed in a separate section below.

### 6 7 2.3.2 Meso-scale cyclones

8 Polar mesoscale cyclones are vortices north of the main polar frontal zone, with the most intense ones  
9 (near-surface wind speeds more than  $15 \text{ m s}^{-1}$ ) being classed as polar lows. They are typically short-  
10 lived (12-48 hours in duration) and generally occur over the subpolar seas. They fall broadly into two  
11 classes: those that are fundamentally convective, i.e. forced by large air-sea heat fluxes, and those that  
12 are fundamentally baroclinic, i.e. instabilities of a horizontal temperature gradient, often associated  
13 with Arctic fronts. In reality most polar mesoscale cyclones have a mixture of these forcing  
14 mechanisms at different stages of their life cycle. Polar mesoscale cyclones tend to occur over the sub-  
15 polar seas, e.g. the Greenland, Norwegian, Iceland, Barents, Irminger, Labrador, and Bering Seas, the  
16 Sea of Japan and the Gulf of Alaska in the Northern Hemisphere. Further background can be found in  
17 e.g. Renfrew (2003) and Rasmussen and Turner (2003).

18 In recent years there has been an upsurge of interest in polar lows. The IPY was a focal point for a  
19 number of field campaigns which observed polar lows, including GFDex (e.g. Renfrew et al. 2008) and  
20 the Norwegian IPY-Thorpex campaign (Kristjánsson et al. 2011). In the latter arguably the most  
21 comprehensive set of observations of a polar low to date were obtained for a case over the northern  
22 Norwegian Sea, enabling studies of the structure, dynamics, lifecycle, simulation accuracy and  
23 predictability of this event (e.g. Linders and Saetra 2010; Føre et al. 2011; Føre and Nordeng, 2012;  
24 McInnes et al. 2011; Wagner et al. 2011; Irvine et al. 2011; Aspelien et al. 2011; Kristiansen et al.  
25 2011). Finding, for example, that this case had critical upper-level forcing (Føre et al. 2011), and was  
26 more accurately simulated with convection-permitting grid resolution of 4 or 1 km (McInnes et al.  
27 2011). Operational weather forecasting systems have now reached the state where polar lows should be  
28 able to be predicted routinely. Numerical weather prediction grid sizes have been adequate for some  
29 time, but observing and data assimilation systems have not always been able to consistently provide  
30 suitable initial conditions, for example in Irvine et al. (2011) there was strong sensitivity to the initial  
31 conditions. Regional high-resolution ensemble prediction systems (EPS) provide a realistic prospect of  
32 robust predictions at the mesoscale, tackling initial condition sensitiveness for example. These regional  
33 EPS systems are still being developed and optimizing their setup for polar lows is a current challenge  
34 (Aspelien et al. 2011; Kristiansen et al. 2011). For example, Kristiansen et al. (2011) find a crucial  
35 dependence on EPS domain size and location, as well as on certain parameterization settings.

36 Polar mesoscale cyclones are not explicitly resolved by the current generation of global climate  
37 models. Due to their high impact, predictions of any changes in frequency or location of occurrence are  
38 important. A couple of recent studies address this: Kolstad and Bracegirdle (2008) use marine cold-air  
39 outbreaks as a proxy for polar low activity; while Zahn and von Storch (2010) use dynamical  
40 downscaling to simulate polar mesoscale cyclones. In both studies a migration northwards is found,  
41 following the retreating sea-ice pack, and consequently there is a decrease in the frequency of polar  
42 lows through the 21<sup>st</sup> Century.

43 Polar lows are highly coupled phenomena. Large fluxes of heat, moisture and momentum from the  
44 relatively warm ocean are usually crucial for their development. Hence they also provide a strong

1 forcing for the ocean, e.g. deepening the mixed-layer so bringing warmer waters to the surface (Saetra  
 2 et al. 2008) and changing water mass properties and consequently the ocean circulation (Condrón et al.  
 3 2008; Condrón and Renfrew 2013). In a set of high resolution ocean modelling experiments with and  
 4 without polar lows, Condrón and Renfrew (2013) find adding polar lows significantly increases the  
 5 depth of deep convection (Figure 6), spins up the Greenland Sea gyre, and increases the momentum  
 6 and heat transported north in the North Atlantic's subpolar gyre as well as the frequency of dense water  
 7 flowing south out of the Nordic Seas. The impact of polar lows on the coupled climate system is still  
 8 uncertain: their occurrence is subject to changes in both the atmosphere and ocean, and any changes  
 9 will potentially feedback on both the atmosphere and ocean.

10

### 11 **3. Sea Ice and Snow**

12

#### 13 **3.1 Radiative processes and properties**

14

##### 15 3.1.1 Melt onset

16

17 Based on the SHEBA data from the Beaufort Sea, Persson (2012) analysed the links between the spring  
 18 onset of snow melt and free-tropospheric synoptic variables, clouds, precipitation, and in-ice  
 19 temperatures. He found that the melt onset is primarily determined by large increases in downwelling  
 20 longwave radiation and modest decreases in the snow surface albedo. These changes in the radiative  
 21 fluxes are related to synoptic events and seasonal warming of the free troposphere. The work of  
 22 Persson (2012) benefited from detailed observations, but only addresses a single spring in a limited  
 23 region. Maksimovich and Vihma (2012) utilized the ERA-Interim reanalysis, which are far less reliable  
 24 than observations but allowed the study of the inter-annual differences in the circumpolar Arctic. They  
 25 found that the anomaly in net surface heat flux 1–7 days prior to the snow melt onset explains up to  
 26 65% of the inter-annual variance in the melt onset in the central Arctic. Among the terms of the net  
 27 heat flux, the downward longwave radiation most strongly controlled the variability of snow melt  
 28 onset. Statistically, solar radiation by itself is not an important factor, but together with other fluxes it  
 29 improves the explained variance of melt onset. In accordance with the above-mentioned results, the  
 30 early melt onset in 2007 was preceded by an exceptionally warm spring (Vihma et al., 2008) with a  
 31 large advection of warm, cloudy marine air masses from the Pacific sector (Graversen et al., 2011).  
 32 After the melt onset, the evolution of the snow surface albedo and the transmissivity of the snow-ice  
 33 system is crucial for the surface energy budget.

34

##### 35 3.1.2 Snow and ice albedo; observations and parameterizations

36

37 A schematic illustration of snow and ice thermodynamic processes and interactions, with focus on the  
 38 role of surface albedo, is provided in Figure 7.

39

40 The detailed and complete datasets of snow/ice and atmospheric quantities that were collected during  
 41 SHEBA have still been used during and after IPY to thoroughly evaluate and compare many snow and  
 42 ice albedo schemes (Liu et al., 2007; Wyser et al., 2008; Pedersen et al. 2009). Several Arctic field  
 43 campaigns carried out after SHEBA (including the Tara campaign of DAMOCLES) were crucial to  
 44 monitor and deepen the understanding of the processes controlling the snow and ice albedo in a rapidly

1 changing environment. Altogether, these observations have shown that the seasonal evolution of the  
2 Arctic sea ice albedo follows the surface metamorphism and change of phases, from dry snow to  
3 melting snow, pond formation, pond drainage, pond evolution, and fall freeze-up (Perovich et al., 2009;  
4 Nicolaus et al., 2010a; Perovich and Polashenski, 2012). Seasonal ice has a lower albedo than  
5 multiyear ice, because (a) it has a thinner and therefore faster melting snow layer, (b) the ice itself is  
6 thinner, containing a much lower fraction of scattering bubbles, and (c) melt ponds are more extensive  
7 due to less ice deformation and a smaller freeboard (Perovich and Polashenski, 2012). The area-  
8 averaged surface albedo results from a complex combination of the albedos of open water, melt ponds,  
9 snow-free sea ice, and snow-covered sea ice (Perovich et al., 2009).

10 As snow/ice albedo is the key factor affecting the surface energy budget over the Polar areas, a large  
11 number of recent modeling studies have addressed the improvement of the snow and ice albedo  
12 representation, also with the goal of simulating the various climate feedback mechanisms affected by  
13 changes in snow/ice albedo. The climate models used in the IPCC AR4 systematically overestimated  
14 the sea ice albedo in summer, by as much as 0.05 (Wang et al., 2006), and failed to incorporate the  
15 recently observed rapid reduction of Arctic sea ice into their predicted ranges of variability. Small  
16 changes in the ice albedo scheme may lead to significant changes in the simulation of summer sea ice  
17 extent (e.g. Dorn et al., 2007; 2009). This result called for a reconsideration of the physical basis of the  
18 sea ice albedo models, which might explain in part why the rapid reduction of Arctic sea ice is better  
19 captured by the models used for the latest assessment report AR5 (Stroeve et al., 2012; Massonet et al.,  
20 2012).

21 An accurate albedo calculation requires a radiative transfer model in the atmosphere and in the  
22 snow/ice layer, coupled with a snow/ice model that represent the snow/ice crystals with their optical  
23 properties and the snow/ice layering (Peltoniemi, 2007; Kaempfer et al., 2007). The size and shape of  
24 the crystals determine their optical properties, thus the crystal metamorphism is the principal driver of  
25 the albedo evolution. However, in climate and NWP models albedo is usually expressed as a function  
26 of the bulk snow/ice/atmospheric properties that more or less directly affect the snow metamorphism  
27 (surface temperature, snow age) or are affected by it (snow and ice thickness, snow density), the form  
28 of the equation and the values of the included coefficients resulting from the best fit with observations  
29 or with detailed radiative transfer calculations (Gardner and Sharp, 2010). The degree of complexity  
30 varies a lot among these models, NWP models traditionally have much less detailed surface schemes  
31 than climate models. Prognostic snow and ice albedo parameterizations, which include a time-  
32 dependent albedo decay, gave the best results when their performance was compared with simpler  
33 temperature-dependent parameterizations (Essery et al., 2012; Wyser et al., 2008). Among the  
34 prognostic schemes, one of the most sophisticated is the model introduced already by Dickinson et al.  
35 (1993), which accounts for the albedo dependence on spectral bands and direction of the illumination.  
36 It has been implemented in many climate models (Bitz et al., 2012; Goosse et al., 2009), and it has also  
37 been coupled to an explicit treatment of melt pond albedo (Pedersen et al., 2009).

38 Variations in the areal melt pond coverage are a major driver of albedo changes on melting Arctic sea  
39 ice. Considering observations of melt ponds, the drift of Tara in DAMOCLES offered a valuable  
40 opportunity to observe the temporal change of multiyear sea ice at very high latitudes. Sankelo et al.  
41 (2010) quantified the areal melt pond coverage at about 88°N, which was higher than expected on the  
42 basis of previous observations, with maximum pond coverage of 32-42% in mid-August. Rösel et al.  
43 (2012) presented the first satellite derived Arctic-wide, multi-annual melt pond data set. The study for  
44 the time period from 2000 to 2011 was based on Moderate Resolution Image Spectroradiometer  
45 (MODIS) data. Since there is an ongoing shift in the Arctic sea ice cover from multiyear ice to seasonal

1 ice (Perovich and Polashenski, 2012), melt pond studies for first-year ice become more and more  
2 important. Recent sophisticated field studies of melt ponds on seasonal sea ice were conducted on land-  
3 fast ice in the Chukchi Sea during the summer melt seasons of 2008, 2009, and 2010 (Polashenski et  
4 al., 2012). Ice surface topography and melt water balance are found to both play key roles in melt pond  
5 evolution.

6 Substantial efforts have already been made to formulate physically based models of melt pond  
7 formation and evolution to predict melt pond coverage (Scott and Feltham, 2010; Skyllingstad et al.,  
8 2009; Flocco and Feltham, 2007) and to incorporate explicit melt pond parameterizations/models into  
9 albedo calculations of global and regional sea ice and climate models (Holland et al., 2012; Flocco et  
10 al., 2010; Hunke and Lipscomb, 2010; Pedersen et al., 2009; K $\ddot{o}$ ltzow, 2007). The explicit  
11 consideration of melt pond effects has a huge impact on the simulated Arctic sea ice cover as shown,  
12 e.g., by Flocco et al. (2012) who incorporated their pond model into the Los Alamos CICE sea ice  
13 model. Simulations for the period 1990 to 2007 are in good agreement with satellite-based ice  
14 concentration. In comparison to simulations without ponds, the September ice volume is nearly 40%  
15 lower.

16 In the melt water accounting conceptualization, a melt pond can be represented as a volume of water  
17 determined by the balance of inflows and outflows, distributed in the lowest points of local topography  
18 (Polashenski et al., 2012). The general approach of the GCM melt pond parameterizations by Holland  
19 et al. (2012), Hunke and Lipscomb (2010), and Pedersen et al. (2009) is based on this concept.  
20 ECHAM5 (Pedersen et al., 2009) and the CCSM CICE 4.0 (Holland et al., 2012; Hunke and Lipscomb,  
21 2010) use functional relationships to relate pond depth to pond area fraction. CICE 4.0 uses a linear  
22 function, and the ECHAM5 version applied by Pedersen et al. (2009) used a more complex function.  
23 The linear function is based on SHEBA data. However, Polashenski et al. (2012) show that the  
24 relationship between melt pond depth and area fraction is not unique. Polashenski et al. (2012) suggest  
25 that a better solution to compute both quantities would be to relate components of the melt water  
26 balance to ice properties already being calculated in the GCM's, and to collect data representing the  
27 topography of various ice types to better parameterize the areal distribution of melt water. The results  
28 of their field studies identify links between the temporal evolution of pond coverage and ice  
29 temperature, salinity, and thickness. Hence, measurement results provide new opportunities to  
30 realistically parameterize ponds within sea ice models.

31 The simulation of surface albedo is also related to the representation of the thermal insulation of the  
32 snowpack, which is coupled to the modeling of snow mass and density. Compared to observations,  
33 more consistent results are obtained from those snow schemes that include a prognostic representation  
34 of snow density, and take some account of the storage and refreezing of liquid water within the snow  
35 (Essery et al., 2012; Dutra et al., 2012). Presently, snow albedo schemes are more advanced over land  
36 than over sea ice. The reason is related to the complexity of the sea ice surface types, especially during  
37 melting conditions (Figure 7). The sea ice model LIM2, recently implemented into the ECMWF  
38 forecasting system (Molteni et al., 2011), has a sea ice albedo parameterization which includes several  
39 snow and ice categories, depends on snow and ice thickness and cloudiness, does not retain any melt  
40 water and implicitly accounts for a constant melt pond fraction when the surface is melting. However,  
41 in the ongoing development there is the implementation of a more comprehensive snow model that  
42 includes a variable vertical resolution based on the density stratification, the representation of melt  
43 ponds and superimposed ice formation. Also in the case of CCSM, the land snow scheme (CLM4,  
44 Lawrence et al. (2011)) has a more advanced snow thermodynamic treatment than the latest version of  
45 the sea ice scheme (CICE4.0, Hunke and Lipscomb (2010)), which has fixed snow and ice density and

1 thermal conductivity. This oversimplification was partly responsible for positive biases in snow  
2 thickness over the Arctic, and excessive late autumn and early winter snow density, with feedbacks on  
3 the albedo (Blazey et al., 2013).

4  
5 The widely applied NWP and research model WRF is often used with an oversimplified snow albedo  
6 parameterization (a constant value of 0.8), which leads to large errors in the summer shortwave  
7 radiative fluxes (Porter et al., 2011). To simulate the Arctic atmospheric conditions during the SHEBA  
8 experiment, a simple idealized albedo model based on the SHEBA observations (Perovich et al.,  
9 2007a) and a satellite dataset was used in Polar WRF (Bromwich et al., 2009)). This albedo model was  
10 then applied to the entire Arctic Ocean to simulate the one year period from December 2006 to  
11 November 2007 (Wilson et al., 2011). Simulated annual mean temperatures had, however, a cold bias  
12 of -1 to -2°C (Wilson et al., 2011).

13  
14 In some occasions some of the most sophisticated prognostic albedo parameterizations in GCM and  
15 NWP models have been defined as “physically based” to distinguish them from even simpler albedo  
16 schemes (Essery et al., 2012), but in fact, they do not allow the coupling between penetration of solar  
17 radiation into the snow and ice layer, the micro-scale characteristics of the ice crystals and the surface  
18 albedo. The gap between the snow albedo formulated in detailed radiative transfer and snow models  
19 and the albedo parameterizations applied in GCM and NWP models has recently been narrowed by the  
20 development of a prognostic parameterization of snow grain metamorphism, which links the snowpack  
21 microphysics to the albedo evolution (Flanner and Zender, 2006). In this SNow and ICe Aerosol  
22 Radiation (SNICAR) model albedo is calculated from the inherent scattering-absorption properties of  
23 snow crystals and included absorbers. SNICAR has recently been implemented in sea ice models with  
24 detailed radiative transfer schemes and high vertical resolution (for instance the CCSM CICE4.0,  
25 Holland et al. (2012)), contributing to a significant improvement in the simulation of the Arctic albedo  
26 and sea ice concentration (Gent et al., 2011).

27  
28 Many of the recently developed snow and ice albedo parameterizations have not yet been thoroughly  
29 evaluated against field observations. High quality, complete datasets of radiation and snow and ice  
30 properties are extremely rare and still their acquisition requires a large effort. Because of uncertainties  
31 in the forcing data and oversimplifications in representing many physical processes, increasing the  
32 complexity of the schemes may lead to severe simulation errors, and existing biases in the driving  
33 parameters will propagate to the processes that depend on them. Thus, even the simplest  
34 parameterizations can give equally good or bad results as the most complex ones (Essery et al., 2012;  
35 Brun et al., 2008). Recent advances in the remote sensing retrieval techniques of surface albedo over  
36 the Arctic allowed the collection of a 28-year time series of albedo estimations in all sky conditions  
37 (Riihelä et al., 2012), offering a valuable reference dataset to analyze spatial and temporal albedo  
38 variability.

39  
40 The transfer of solar shortwave radiation under cloudy skies in the boundary zone of the open sea and  
41 snow/ice cover is a complex process that has not yet received much detailed attention. Pirazzini and  
42 Räisänen (2008) found that under overcast skies with multiple reflections between the cloud base and  
43 the snow/ice surface, the local value of downwelling solar radiation also depends on the albedo of the  
44 neighbouring surface type. They further derived a simple parameterization for the broadband effective  
45 albedo, defined as the albedo of a homogeneous surface that would result in the same downwelling  
46 irradiance as locally observed in the presence of a heterogeneous surface.

### 3.1.3 Aerosol deposition on snow and ice

Aerosol deposition on snow and ice is an issue that has attracted very much recent research. As black carbon (BC) effectively absorbs visible radiation, it causes acceleration in the growth of snow grains, and therefore an overall decrease in albedo. In particular, Hansen et al. (2005) suggested that the effect of BC on snow albedo contributes substantially to rapid warming and sea ice loss in the Arctic, although recent measurements (Forsström et al. 2009; 2013; Doherty et al., 2010) have shown substantially lower levels of BC than was observed in the 1980s (Clarke and Noon 1985). In view of these findings, parameterizations of BC and soot concentration in snow have been recently developed (Flanner and Zender, 2006; Yasunari et al., 2011; Aoki et al., 2011). Evaluations of these parameterizations have revealed their capability to better reproduce the observed snow albedo and snow depth (Yasunari et al., 2011; Hadley and Kirchstetter, 2012). Moreover, it has been found that the BC/snow radiative forcing in the Arctic is at maximum coincidentally at the time of snowmelt onset (Flanner et al., 2007), triggering strong snow-albedo feedback in local springtime. For this reason, although the magnitude of the climate response from light-absorbing particles on snow is much smaller than the impact of doubling CO<sub>2</sub>, the sensitivity of the atmosphere to the BC/snow forcing (i.e, the temperature change per unit of forcing) is three times larger than the sensitivity to the CO<sub>2</sub> forcing (Goldenson et al., 2012; Flanner et al., 2007).

Flanner et al.'s estimation of global annual mean BC/snow surface radiative forcings (0.054 and 0.049 Wm<sup>-2</sup> during strong (1998) and weak (2001) boreal fire years) was in line with the IPCC AR4 estimation (IPCC, 2007) and was later confirmed by other studies (Wang et al., 2011; Goldenson et al., 2012). Over large areas of the Arctic Ocean and sub-Arctic seas, the autumn and winter near-surface warming resulting from this radiative forcing is 1-2°C (Goldenson et al., 2012). Through 20th century equilibrium climate experiments, Koch et al. (2009) obtained a 0.5°C mean Arctic surface warming due to the BC-snow albedo effect. In equilibrium climate experiments, the effect of present-day aerosol deposition on sea ice thickness was estimated to be a thinning of about 30 cm (averaged over the year) compared to a scenario without aerosol deposition (Goldenson et al., 2012; Holland et al., 2012). Nevertheless, since the BC content in Arctic snow has decreased since the 1980s, it is improbable that the present aerosol load has contributed to the recently observed rapid decline of Arctic sea ice. Koch et al. (2011) attributed about 30-50% of the Arctic warming and ice melt that occurred in early 20<sup>th</sup> century to the BC-albedo effect, but determined that later in the century the reduction in Arctic BC contributed to Arctic cooling and increased snow/ice cover, so that on average, over the 20<sup>th</sup> century, only about 20% of Arctic warming and ice melting was attributable to the BC-albedo effect.

Through idealized experiments, Flanner (2013) concluded that the current simulated distribution of Arctic atmospheric BC slightly cools the surface with a sensitivity of  $-0.21 \pm 0.32 \text{ K (Wm}^{-2}\text{)}^{-1}$  supporting an earlier study (Shindell and Faluvegi, 2009), while the atmospheric and cryosphere-deposited BC originating from the Arctic (mostly Siberian forest fires) warms the Arctic with a sensitivity of  $+0.5 \pm 0.4 \text{ K (Wm}^{-2}\text{)}^{-1}$ . Flanner et al. (2009) argued that, in springtime, the radiative effect of the reduction of surface-incident solar energy (dimming) caused by atmospheric aerosols containing BC and organic matter has been smaller than the effect of the reduction of snow albedo caused by deposition of such aerosols (darkening), resulting in a warming. However, this is probably

1 true only for the first half of the last century, as in the more recent decades the dimming effect (causing  
2 atmospheric cooling) has likely dominated over the darkening (Koch et al., 2011).

#### 4 3.1.4 Transmittance of sea ice and snow

5 Knowledge about the transmittance of sea ice for solar radiation is crucial when assessing the surface  
6 energy balance, and within that the contribution of atmospheric versus oceanic forcing to ice melt, and  
7 the radiation available for the ecosystem in and below the sea ice. Transmittance of the sea ice system  
8 depends on snow and ice properties, and on possible content of algae in the ice (e.g. Mundy et al.,  
9 2007). Spectral radiometer surveys during recent years have yielded substantial advances in resolving  
10 characteristics of transmittance of sea ice in time and space. Light et al. (2008) summarized SHEBA  
11 transmittance measurements under different ice types at different stages of the seasonal evolution of sea  
12 ice. Autonomous setups (Nicolaus et al., 2010b; Wang et al., 2014) have been installed on drifting ice  
13 floes, measuring transmittance continuously over periods covering the entire transition from freezing to  
14 melt and back to freezing conditions (Nicolaus et al., 2010a; Wang et al., 2014). With this, the nature,  
15 timing and length of the period of increased transmittance during summer, related to snow  
16 metamorphism, snow melt and ice properties, could be quantified. Such measurements are limited  
17 regarding information in space. Despite the fact that the ice floe with the autonomous setup is drifting,  
18 and thus covers a larger geographical area, the ice floe remains the same. New studies worked on  
19 investigating the spatial variability of sea ice, and herein especially of first-year ice, the ice type that  
20 increases in relative portion over the Arctic as a whole at the cost of multiyear sea ice. Frey et al.  
21 (2011) studied an ice floe with a number of individual measurements under locations with different  
22 surface characteristics, and quantified the role of melt ponds for the radiation balance below the ice. By  
23 combining surface measurements from a sledge based system (Hudson et al. 2012) with measurements  
24 carried out by divers beneath the ice, the complete radiation balance of the first-year sea ice system  
25 could be quantified, for a given case and stage (Hudson et al. 2013). Similar observations were also  
26 done over land-fast ice near Barrow, Alaska, but with the under-ice radiation measured from a sledge  
27 that slides along the underside of the fast ice, pulled with a rope (Nicolaus et al. 2013).

28 On larger scales, models can help to estimate the amount of light penetrating the ice and its heating  
29 effect (e.g. Itoh et al. 2011). This requires, however, a good vertical resolution. Climate and NWP  
30 models have traditionally used a single snowpack layer, but a high vertical resolution in snow and ice  
31 models has been revealed to be important to correctly simulate light scattering coefficients (Light et al.,  
32 2008), surface albedo (Aoki et al., 2011), the onset of ice melt (Cheng et al., 2008b), sub-surface grain  
33 metamorphism and melt (Dadic et al., 2008; Cheng et al., 2008a,b), the vertical profile of thermal  
34 conductivity (Dadic et al., 2008), and deep snowpack conditions (Dutra et al., 2012). The increase in  
35 vertical resolution has yielded a fundamental improvement in the treatment of the penetration of  
36 shortwave radiation in snow and sea ice (Briegleb and Light, 2007; Light et al., 2008). By accounting  
37 for the ice layering, Light et al. (2008) concluded that much less radiation is absorbed in the uppermost  
38 highly-scattering layer, and more light is predicted to penetrate deep into the ice and into the ocean than  
39 was previously accounted for. This modeling progress is parallel to the increased effort in simultaneous  
40 measurements of snow/ice spectral albedo and transmittance (Nicolaus et al., 2010a,b; Perovich, 2007;  
41 Ehn et al., 2008a; 2011), which have also revealed the impact of some biological processes on sea ice  
42 transmittance in the Arctic central (Nicolaus et al., 2010b) and on land-fast ice (Ehn et al., 2008a,b;  
43 2011).

44 Radiative processes in sea ice and snow closely interact with sea ice structure and other processes, such  
45 as snow and ice melt, heat conduction, refreezing of melt water, and gravity drainage of salt (Figure 7).

## 1 **3.2 Sea ice structure and non-radiative processes**

2

### 3 3.2.1 Internal structure of sea ice: salinity and gravity drainage

4 The internal structure of sea ice consists of a mixture of solid fresh-water ice, liquid salty brine and gas  
5 inclusions, whose interaction on the millimetre scale crucially affects the large-scale behaviour of sea  
6 ice. This interaction defines the evolution of the solid fraction within sea ice, which in turn defines  
7 virtually all large-scale properties of sea ice; these include the heat capacity, heat conductivity,  
8 mechanical strength, and susceptibility to percolation of surface melt water to name but a few. In  
9 addition, the small-scale processes governing the interior structure of the ice define how efficiently  
10 brine can drain from the ice, which in turn contributes to shaping the large-scale circulation of the  
11 world ocean.

12 Most of our recent progress in modelling the small-scale structure of sea ice has come from application  
13 of the so-called mushy-layer theory (e.g., Feltham et al., 2006). This theory describes any multi-  
14 component, multi-phase reactive porous medium of which sea ice is but one example. This theory has  
15 in particular allowed us to better understand the temporal evolution of sea-ice salinity (Notz and  
16 Worster, 2009). This understanding is crucial because the salt content and temperature of sea ice  
17 define, together with the amount of entrapped gas, the solid fraction of the ice as the most fundamental  
18 parameter to describe the state of a specific sea-ice sample. We now know that, initially, all salt that is  
19 contained within sea water is also contained in newly formed sea ice. Much of this salt then rapidly  
20 drains out by convective overturning, which in the interior of the ice replaces dense, salty brine with  
21 less salty sea water (so-called gravity drainage). This leads to a rapid reduction of the salinity of sea ice  
22 and in turn increases the solid fraction of the ice. Additional loss of salt then occurs in summer through  
23 the slushing of fresh surface melt water that percolates through the ice. Measurements from warm first-  
24 year sea ice exposed to an increased oceanic flux show substantial desalination (Widell et al. 2006).  
25 Based on this understanding, models are starting to simulate in a physically consistent way the  
26 evolution of the bulk salinity of sea ice from its initial formation to its complete melt. Such models  
27 range from specialised models of gravity drainage (Wells et al., 2011; Rees Jones and Worster,  
28 2013a,b) to more applied models that present simplified parameterisations of this major desalination  
29 process for the use in large scale models (Turner et al., 2013; Griewank and Notz, 2013). Based on  
30 these models, a more realistic representation of the interaction between the small-scale structure of sea  
31 ice and the ocean and the atmosphere has now become possible.

32 For more details on this topic, we refer to the recent dedicated review article by Hunke et al. (2011).

33

### 34 3.2.2 Formation of superimposed ice and snow ice

35 Snow-ice and superimposed ice are generated by refreezing of snow-slush (Figure 7). The slush layer is  
36 created by either ocean flooding or snow melting. In the case of ocean flooding, the product of  
37 refreezing is called snow ice, whereas in the case of snow melt and percolation of the melt water down  
38 to the snow-ice interface, the refreezing generates superimposed ice. Already long before the IPY, the  
39 generation of snow-ice has been taken into account in sea ice models (e.g. CICE, LIM) with a  
40 simplification of the Archimedes' principle, with more detailed modelling for seasonal sea ice  
41 presented e.g. by Cheng et al. (2006).

1 The contribution of snow ice and superimposed ice to the total ice mass in the Arctic has, however, not  
2 received much attention so far. This is partly due to the fact that snow ice has been rarely formed in the  
3 Arctic, since the ratio of snow thickness to ice thickness has usually been low. Superimposed ice has  
4 been observed to occur in Arctic sea ice (e.g., Nicolaus et al., 2003; Wang et al., 2014), but it is usually  
5 rapidly deteriorated in the following melting season. Pre-IPY work in modelling of snow ice and  
6 superimposed ice has mainly focused on sub-Arctic seas (Baltic Sea, Sea of Okhotsk) and to some  
7 extent on the Chukchi Sea (Cheng et al., 2008b). In Semmler et al. (2012) the modeled ice thickness on  
8 an Arctic lake showed a large improvement when snow-ice and superimposed ice were taken into  
9 account.

10 The source term for snow-ice and superimposed ice is the total precipitation available on ice. Accurate  
11 information on precipitation is critical for modelling, particularly in early winter. Detection of snow  
12 thickness in the Arctic is challenging because it is subject to large spatial and temporal variations, due  
13 to wind drift, etc. The effects of wind also make the *in situ* precipitation measurements liable to errors,  
14 which can be as large as 200% (Aleksandrov et al., 2005). Further, *in situ* measurements are rare,  
15 making NWP models the primary source of atmospheric forcing for snow and ice modeling. Cheng et  
16 al. (2013) introduced a simple snow parameterization scheme connected to the precipitation from an  
17 NWP model to account for the snow accumulation in the early winter season.

18 The snowfall declines in the Arctic summer, which has mainly been due to the change of precipitation  
19 from snowfall to rain with very little change in total precipitation (Screen and Simmonds, 2012).  
20 However, considering the total annual precipitation, climate models project an increase (e.g., Overland  
21 et al., 2011). This together with the thinning of sea ice will likely result in a more extensive occurrence  
22 of snow ice and superimposed ice in the Arctic, with their larger contributions to the total ice mass  
23 (their contributions are already large e.g. in the Baltic Sea and, for snow ice, in the Antarctic).

24

### 25 3.2.3 Heat conduction

26 The mass balance of sea ice and its snow cover largely depend on the heat conduction through snow  
27 and ice (Figure 7). The conductive heat flux contributes to the surface energy budget, and the  
28 melt/growth at the ice bottom is controlled by the difference between the conductive heat flux and the  
29 ice-water heat flux. Heat conduction is vitally important also for consolidation of raft ice (Bailey et al,  
30 2010). The thermal conductivity of snow is usually parameterized as a function of snow density, and  
31 that of sea ice as a function of ice temperature and salinity (Maykut and Untersteiner, 1971). Pringle et  
32 al. (2007) presented a new parameterization for sea ice on the basis of amended data analysis; the heat  
33 conductivity was higher than that based on Maykut and Untersteiner (1971): by 5-10% for multi-year  
34 ice and by 5-15% for first-year ice. For snow, a micro-tomographic study by Calonne et al. (2011)  
35 indicated that the effective thermal conductivity increases with decreasing temperature, mostly  
36 following the temperature dependency of the thermal conductivity of ice. Accordingly, a temperature  
37 and density dependent heat conductivity of snow should be used in models (Lecomte et al, 2011).

38 The temperature dependence of snow and ice heat conductivity is a bulk effect, as indeed conductivity  
39 depends on the micro-structural and mechanical properties of the snow and ice texture, which change  
40 when subjected to temperature gradients. This became evident in temperature gradient - snow  
41 metamorphism experiments at a constant density: the heat conductivity increased as much as twice its  
42 initial value in response to changes in structure and texture (Scheebeli and Sokratov, 2004), showing a  
43 strong anisotropic behaviour (Shertzer and Adams, 2011). Moreover, Dominé et al. (2011) observed

1 that thermal conductivity of snow can be expressed as a function of snow density and shear strength  
2 alone.

3 In the Arctic, the spatial inhomogeneity of snow distribution has a major impact on the regional heat  
4 conductivity, especially when the snow depth is less than 0.4 m. When the snowpack is thin on  
5 average, bare ice is likely present because of the effect of wind in redistributing the snow thickness.  
6 Hence, the effective snow heat conductivity would be a mixture of heat conductivity of snow and ice  
7 (Semmler et al., 2012).

### 9 **3.3 Small-scale dynamics of sea ice**

10 Sea ice dynamics is closely tied to the processes discussed above; it is forced by the air-ice momentum  
11 flux (Section 2.1.4), and affects the regional albedo (Section 3.1.2), heat fluxes from the ocean to the  
12 atmosphere via leads and polynyas (Section 2.1.3), as well as sea ice growth via rafting and ridging  
13 (Figure 8), which further affects sea ice thermodynamics (Sections 3.1 and 3.2).

#### 15 3.3.1 Sea ice deformation

16 The much-faster-than-expected drift of the Tara in 2006-2007 along the Transpolar Current was among  
17 the first signs of ongoing profound changes in Arctic sea ice mechanics and kinematics (Gascard et al.,  
18 2008). A systematic analysis covering 30 years of buoys' drift data revealed a significant increase of  
19 both sea ice drift speeds and deformation rates over this period within the Arctic basin (Rampal et al.,  
20 2009), with obvious consequences in terms of sea ice export, negative mass balance, and decline  
21 (Rampal et al., 2011). This accelerated kinematics does not simply result from sea ice shrinking and  
22 thinning, but is also the consequence of a recent mechanical weakening of the Arctic sea ice cover in  
23 both winter and summer (Gimbert et al., 2012b). This mechanical weakening is likely related to an  
24 intensification of sea ice fracturing and fragmentation. This calls for a better understanding of these  
25 processes from local to regional scales. Indeed, through lead opening, sea ice fracturing partly control  
26 energy fluxes between the ocean and the atmosphere (see Section 2.1.3), and to an extent momentum  
27 fluxes through a modification of surface roughness and drag coefficients (Section 2.1.4).

28 Mechanical waves travel within the Arctic sea-ice cover, generated by ocean surface waves as well as  
29 sea ice fracturing, ridge build-up, and floe collisions (Figure 8). While in-situ stress measurements  
30 (Weiss et al., 2007) and aerial/satellite observations are essential to explore sea ice mechanics, a high  
31 frequency monitoring of sea ice fracturing and faulting, i.e. at the timescale of crack propagation, was  
32 not available until recently, except for short-duration (week-long) experiments that only investigated  
33 high-frequency noise (e.g. Dudko et al., 1998). During the DAMOCLES field campaign in spring 2007,  
34 a network of broad-band (100 Hz-60s) three-component seismometers was installed around Tara,  
35 recording signals dominated by ice swell (Marsan et al., 2011). Marsan et al. (2012) exploited the  
36 dispersion of this ubiquitous signal, i.e. the fact that the higher the frequency the faster the wave  
37 propagation, and its dependence on the ice thickness, to invert the average thickness of the Tara's floe.  
38 The results agreed well with electromagnetic measurements and drill-hole profiles conducted on the  
39 same floe (Haas et al., 2011), thus validating the use of a classical concept (the dependence of wave  
40 propagation on ice thickness) to passively monitor sea ice thickness on a regular basis over horizontal  
41 scales from  $10^0$  to  $10^2$  km.

1 These original seismic observations open the way towards a more systematic recording and analysis of  
2 waves in ice over larger space and time scales, in order to (i) monitor average ice thickness and its  
3 evolution at the regional scale, and (ii) to complement satellite measurements of sea ice deformation by  
4 providing a much more detailed temporal sampling and therefore a better characterization of sea ice  
5 fracturing processes. This should help to constrain the parameterization of sea ice strength in sea ice  
6 models. Indeed, sea ice strength is still poorly constrained, either at the local or pan-Arctic scale, and  
7 an analysis of the response of sea ice to the Coriolis forcing is a way to estimate it.

8

### 9 3.3.2 Relationships of inertial oscillations and sea ice rheology

10 As mentioned in the previous section, the weaker the sea-ice cover, the easier its fracturing and  
11 fragmentation. Consequently, when sea ice becomes more mobile, it is characterized by larger speeds  
12 and deformation rates. To measure such possible mechanical weakening at the global scale is difficult.  
13 This has been performed recently from the analysis of the response of sea ice to the well-defined  
14 Coriolis force, i.e. of inertial oscillations (Gimbert et al., 2012a; Gimbert et al., 2012b).

15 In ice-covered waters, the amplitude of inertial oscillations depend on the ice state (thickness,  
16 concentration) as well as on ice rheology. For an ice cover consisting of a loose assembly of floes, such  
17 as south of Fram Strait (Lammert et al., 2009), we expect ice internal stresses to vanish, ice floes to  
18 move nearly in free drift, and therefore inertial oscillations to be strong. In contrast, in a compact ice  
19 cover, strong internal stresses immediately damp the oscillations, which become undetectable (Gimbert  
20 et al., 2012a). Therefore, the measurement of the average amplitude of these oscillations from ice  
21 drifter data can be used to estimate the amount of mechanical dissipation within the ice cover as well as  
22 its degree of cohesiveness and mechanical strength. Averaging must be done both in space, to mitigate  
23 sparse sampling, and in time, especially as inertial oscillations are particularly large after (episodic)  
24 strong winds.

25 Such quantitative analysis was performed by Gimbert et al. (2012a) on the basis of the buoy trajectory  
26 dataset of the International Arctic Buoy Programme covering 30 years (1979-2008). It was found that  
27 (i) the amplitude of the inertial oscillations follows an annual cycle in agreement with the  
28 corresponding annual cycles of sea ice concentration, thickness, and kinematics, i.e. stronger  
29 oscillations in summer, (ii) oscillations are stronger in peripheral zones of the Arctic (the Beaufort Sea,  
30 eastern Arctic, and south of the Fram Strait) corresponding nowadays to first-year sea ice or to a loose  
31 ice pack, and (iii) their average amplitude has significantly increased, especially in summer (Figure 9).  
32 While the first two observations suggest that the use of inertial oscillations is relevant as a proxy for  
33 cohesion, and therefore for mechanical strength, the last points to a mechanical weakening over the  
34 latest 30 years at the global scale.

35 To discriminate the effects of the ice state (thickness, concentration) from those related to the sea ice  
36 mechanical behaviour per se, Gimbert et al. (2012b) built a coupled analytical ocean boundary layer –  
37 sea ice dynamical model and applied it to Arctic sea ice motion in the frequency domain around the  
38 inertial period. This model was able to explain the above-mentioned observations and trends obtained  
39 by Gimbert et al. (2012a). In particular, it was demonstrated that the strengthening of inertial  
40 oscillations in recent years was partly the result of a genuine mechanical weakening of the ice cover,  
41 with a winter ice cover that nowadays mimics the mechanical behaviour of summer sea ice 20 to 30  
42 years ago. From the same model, a significant thinning of the Arctic ocean boundary layer was also  
43 obtained, consistent with an enhanced stratification of the upper halocline triggered by sea ice melt or  
44 increasing river runoffs.

1

## 2 **4. Ocean**

3

### 4 **4.1 Ice-ocean interface; exchange of momentum, heat, and salt**

5 The exchanges of momentum, heat and salt between sea ice and the underlying ocean are small-scale  
6 processes that must be parameterised in large-scale models. That these exchange processes depend on  
7 truly small-scale properties of the interface becomes particularly apparent for the exchange of heat and  
8 salt during sea-ice melting. Here, early measurements showed that the melt rate of sea ice that drifts in  
9 comparably warm water is far less than would be expected from the turbulent exchange of heat and salt  
10 (McPhee et al., 1987). These small melt rates can be explained by the fact that during sea-ice melting, a  
11 thin layer of meltwater with a very low salinity forms underneath the ice, which leads to a locally very  
12 stable stratification. Therefore, the far-field ocean cannot interact turbulently with the interface, but all  
13 transport is governed by diffusion across the thin sublayer underneath the retreating ice (Figure 1; Notz  
14 et al., 2003).

15 Because usually the water temperature is still below 0°C, the phase transition of the ice at the ice-ocean  
16 interface is not governed by a physical melt process, but rather by a dissolution process. Therefore, the  
17 double-diffusive transport of heat and salt (due to the lower molecular diffusivity of salt than heat)  
18 across the thin sublayer ultimately determines the ablation rate at the bottom of the ice. These processes  
19 can be parameterised for large-scale models based on a three-equation approach (Notz et al., 2003;  
20 MCPhee, 2008), where three equations are solved that return the interfacial temperature, salinity and  
21 ablation rate. A crucial parameter for these equations is the ratio of the exchange coefficients for heat  
22 and salt transfer across this interface. Here, recent measurements point towards a value of about 35  
23 (Sirevaag, 2009; MCPhee, 2008). The physical mechanisms that determine this value are, however,  
24 currently not well understood.

25 During freezing, the salty brine that is released from the ice prevents the formation of a stable  
26 stratification. Hence, as long as the ice is growing, the exchange of heat and salt is exclusively  
27 governed by turbulent exchange, and double-diffusive effects can be neglected (Figure 1; MCPhee,  
28 2008). If the effect of the buoyancy flux is negligible, the main unknown then becomes the  
29 determination of the friction velocity, which in turn reduces primarily to a determination of the  
30 hydrodynamic roughness length  $z_{0B}$  at the ice-water interface. Only relatively few measurements of  $z_{0B}$   
31 at the bottom of sea ice exist, and it remains a major challenge to parameterise  $z_{0B}$  as a function of ice  
32 type in large-scale models. To our knowledge, most models prescribe a constant value and do not vary  
33  $z_{0B}$  depending on the ice-thickness distribution within a particular grid cell. This is despite the fact that  
34  $z_{0B}$  ranges from 1 mm for undeformed sea ice (MCPhee et al., 1999) to several centimetres for heavily  
35 deformed ice (Shaw et al., 2009) and ice in the MIZ.

36 The roughness length, stratification, and velocity of ice relative to the ocean together determine the  
37 exchange of momentum between the ocean and the sea ice. Lu et al. (2011) found that for example in  
38 MIZ, most of the momentum transfer may occur through the form drag along the floe edge. In MIZ in  
39 the Barents Sea, Fer and Sundfjord (2007) observed dissipation rates in the upper ocean elevated  
40 above the levels expected from the wind-stress scaling, down to 2.5 times the keel depth, associated  
41 with the pressure-ridge keels. Hence, it is essential that the effects of form drag are accounted for,  
42 either via a larger value of  $z_{0B}$  or separately. This requires information or assumptions on the geometry  
43 of individual ice floes. The increasing availability of remotely sensed distribution of ice floes can, in

1 the years to come, aid the inclusion of such distribution into large scale models and allow for the  
2 parameterization of the related small-scale processes.

#### 4 **4.2 Brine formation in the Arctic Ocean**

5 The salinity (S) of sea ice depends on the ice age and thickness (Notz and Worster, 2009) and rarely  
6 exceeds  $S = 15$ , measured in the practical salinity scale, whereas the average salinity of polar surface  
7 water is about  $S = 30$ . Accordingly, half of the salt contained in sea water is retained in sea ice and the  
8 other half is drained out (Section 3.2.1). The dense brines precipitate and convect through the surface  
9 mixed layer down to a certain depth depending on the vertical stratification and the water depth. In  
10 Storfjorden, Svalbard, a major brine factory (Harpaintner et al 2001), brines have two major effects  
11 depending on where they are formed. One effect is to increase salinity of the upper 100 m in  
12 Storfjorden in the deepest part of the fjord and the second effect is to form a benthic layer originating  
13 from the shallowest parts of the fjord and overflowing at sill depth into the Barents Sea  
14 (Storfjordrenna). The first effect results from dilution into the underlying water masses provided that  
15 the water is deep enough to dilute the brines entirely before they reach the bottom of the fjord. The  
16 second effect results from the fact that brines precipitate to the bottom of the fjord because of shallow  
17 bottom depth. These two effects associated with brine formation can be related to the Arctic Ocean  
18 stratification.

19 Different processes contributing to the formation and evolution of the cold halocline layer (CHL) are  
20 described and discussed in Rudels et al. (1996). Salinization of cold water by brine rejection over  
21 shelves produces waters of varying salinities which can sink along the slope and interleave at their  
22 corresponding density levels (Aagaard et al, 1981). Depending on the density deficit, this process  
23 contributes partly to the formation and maintenance of the cold halocline, or to the ventilation of the  
24 deeper waters. Middag et al. (2009) used dissolved aluminium concentrations in the Eurasian Basin that  
25 indicate deep reaching convection of shelf waters. Paleoclimatologists (e.g. Dokken and Jansen 1999)  
26 argued that this type of ventilation was predominant in the Arctic Ocean during ice age in contrast with  
27 warm period where ocean deep convection is the dominant ventilation factor for deep waters. Because  
28 of the strong upper layer stratification of the Arctic, brine rejection in the central Arctic (in e.g., leads)  
29 cannot lead to deep reaching convection. This process, however, can contribute to the stratification in  
30 the upper CHL; an example from the Laptev Sea is given in Figure 10. During the IPY, Bauch et al  
31 (2011) collected an extensive data set on the oxygen isotope ratio  $\delta^{18}\text{O}$  in the Eurasian and Makarov  
32 Basins that led them to identify layers of the CHL influenced by brine release in coastal polynyas and  
33 layers of the CHL influenced by sea ice formation over the open ocean where vertical convection is  
34 more dominant. Both processes are active in the present climate but it is not clear if one process  
35 dominates over the other.

36 Brine rejection occurs all over the Arctic Ocean but it is much more active in open water areas  
37 (polynyas) than in pack ice. In autumn and winter over polynyas, the sensible heat flux is usually the  
38 dominant part of the surface energy budget, with smaller contributions from the latent heat flux and net  
39 radiation (Lüpkes et al., 2012b). Consequently, the upward sensible heat flux is the main forcing term  
40 for frazil ice formation and brine release in polynyas. Due to the large Arctic sea-ice retreat in summer,  
41 young sea ice expands very fast in the Arctic Ocean and multiyear sea-ice floes vanish. This tendency  
42 significantly enhances frazil ice formation and brine release in the Arctic Ocean, which can partly  
43 contribute to the CHL as described by Bourgain and Gascard (2011).

### 1 **4.3 Diapycnal mixing in the Arctic Ocean**

2 Subsurface layers with above zero temperatures in the Arctic Ocean, originating from the Atlantic and  
3 the Pacific Ocean, form a considerable heat reservoir. The inflow of warm Atlantic Water (AW)  
4 through the Fram Strait alone would be enough to melt 1 m ice per year, if brought to the surface  
5 (Turner, 2010). Diapycnal mixing in the ocean is the main mechanism by which this interior oceanic  
6 heat can be fluxed to the surface, contributing to melting from the ice bottom. Mixing in the stratified  
7 interior ocean is related to internal wave energy, which tends to be low under the Arctic Ocean ice  
8 cover (Levine et al., 1985). Microstructure measurements conducted during the IPY show that the  
9 Arctic Ocean is a quiescent environment with background mixing rates close to molecular levels  
10 (Rainville and Winsor, 2008; Fer, 2009). Efficient vertical mixing and upward oceanic heat fluxes  
11 occur, however, along the continental rise and over topographic features where the warm boundary  
12 current is guided (Sirevaag and Fer, 2009; Fer et al., 2010).

13 An illustration of the main forcing mechanisms and physical processes leading to diapycnal mixing are  
14 summarized in Figure 11. The reader is also referred to Fig. 2 of Padman (1995) and to Fig. 2 of  
15 Rainville et al (2011) for a sketch of the processes. The latter also contrasts the dominant mixing  
16 processes for an Arctic Ocean with relatively small and large seasonal ice-free areas. In the central  
17 basins of the Arctic Ocean, the typical hydrography of the upper ocean is characterized by a 10-30 m  
18 thick mixed layer below the ice-ocean interface with temperature near the freezing point, overlaying a  
19 cold isothermal layer where salinity increases with depth (CHL), followed by the deeper pycnocline  
20 where both temperature and salinity increases to the relatively warm and saline core of AW. The core  
21 of AW gradually deepens as the water circulates along the margins and into the deep basins of the  
22 Arctic (Dmitrenko et al., 2008); in the Amundsen Basin, close to the North Pole, the core of the AW-  
23 derived water resides at around 300 m depth. Direct microstructure measurements in the Amundsen  
24 Basin, conducted during IPY show that the vertical mixing of heat is suppressed by the strong density  
25 stratification in CHL (Fer, 2009). In the central Canada Basin, subsurface temperature maxima due to  
26 intrusions of Pacific Summer Water are located at about 50 m, i.e., closer to the ice. Utilizing the  
27 microstructure measurements made during the drift of the SHEBA ice camp, Shaw et al. (2009)  
28 reported that the strong stratification limited the thickness of mixing zone at the mixed layer base.  
29 Observations made from ice-tethered profilers deployed during the IPY echo these findings (Toole et  
30 al., 2010). In addition, efficient lateral mixed layer re-stratification also impedes mixed layer deepening  
31 (Toole et al., 2010). Re-stratification as a result of submesoscale (order of 1 km) instabilities within the  
32 surface layer is reported using ice-tethered profiler measurements from the Canada Basin  
33 (Timmermans et al., 2012). Previous and subsequent estimates of vertical diffusivity and heat transport  
34 therefore suggest that the warm subsurface layers in the central basins cannot contribute to significant  
35 ice melt. Above the subsurface temperature and salinity maxima of AW, the stratification is favorable  
36 for double-diffusive convection (Section 4.4), which leads to diffusive fluxes up to an order of  
37 magnitude more efficient than the molecular diffusion (Sirevaag and Fer, 2012). Given the quiescent  
38 interior and the large-scale lateral extent of diffusive staircases, the heat flux from double-diffusive  
39 convection can be significant for the average heat loss of the AW layer in the deep basins.

40 The competition between the role of diffusive mixing and the advection of the AW in the boundary  
41 current is decisive on the seasonality of the AW signal. The advective time scale for circum-Arctic  
42 transport of AW from the Santa Anna Trough to the southern Canada Basin, inferred from transient  
43 tracer data, is 7.5 years (Mauldin et al., 2010). The mixing rate between Barents Sea Branch Water in  
44 the boundary current and the interior of the Arctic is slow (5-10 years) allowing the advected  
45 interannually varying tracer signals to dominate over diffusion. At the Lomonosov Ridge where the

1 boundary current bifurcates, however, the mixing rates are elevated, leading to gradual disappearance  
2 of the seasonal AW signal. Modelling results (Lique and Steele, 2012) support this; the seasonal AW  
3 signal survives over order 1000 km distance in the Nansen Basin along the continental slope whereas it  
4 is absent in the Canada and Makarov Basins.

5 The oceanic heat is found to affect the sea ice growth and melt primarily in the MIZ (Polyakov et al.,  
6 2010; Steele et al., 2010). Heat accumulated in the upper ocean will largely be lost to the atmosphere,  
7 delaying the onset of the freezing season and sea ice growth, as well as affecting the heat and moisture  
8 fluxes. Numerical model results of Steele et al. (2010) show that approximately 80% of upper ocean  
9 warming in the Pacific Sector arises from surface heat flux whereas the remaining originates from  
10 ocean lateral heat flux convergence. Melting as a result of upper warming induced by atmospheric  
11 fluxes, comprising of melting on the ice surface and also lateral and basal melting from local warming  
12 of the ocean surface, is responsible for about 60% of summertime melting; dynamical ocean processes,  
13 such as heat flux convergence and vertical mixing, account for the rest of the melting, with an  
14 increasing role of the vertical diffusion (hence bottom melt) in late summer. In the Atlantic sector,  
15 positive temperature anomalies in the AW layer during 2007 coincided with a significant shoaling of  
16 this layer in the Central Arctic (Polyakov et al., 2010) and an estimated increase in the oceanic heat  
17 flux to the ocean surface, despite a coincident increase in stratification in the Makarov and Eurasian  
18 basins (Bourgain and Gascard, 2012). Observations from the drifting ice station ASCOS show a  
19 transition toward a more seasonal ice cover with a more pronounced freezing and melting cycle  
20 (Sirevaag et al., 2011). The heat and fresh water content in the mixed layer and upper cold halocline  
21 were significantly more and the winter mixed layer salinity was significantly larger than those observed  
22 in the early 1990's. The ocean mixed layer was found to be heated from the top and heat was  
23 redistributed downwards by turbulent mixing.

24 Microstructure measurements made during IPY in the central Arctic Ocean show enhanced turbulence  
25 dissipation rates following a storm, correlated with near-inertial frequency band motions that appear in  
26 shear and strain in the upper ocean (Fer, 2014). The study emphasizes the importance of near-inertial  
27 internal wave energy and its role in mixing in the CHL and deeper Arctic stratification, primarily by  
28 modulating the Richardson number to favor shear production of turbulence kinetic energy. While the  
29 diapycnal mixing in the interior Arctic Ocean is quiescent, primarily due to weak internal wave field,  
30 recent studies have shown a correlation between the absence of sea ice and increased near-inertial shear  
31 and internal wave content (Rainville and Woodgate, 2009; Rainville et al., 2011). Retreating ice cover  
32 is thus suggested to lead to an increase in background mixing levels; the MIZ, in particular, can be a  
33 hot-spot of mixing with consequences for the ice extent. Recent studies show enhanced heat fluxes and  
34 turbulent mixing in the MIZ north of Svalbard (Fer and Sundfjord, 2007; Fer et al., 2010). In the wind-  
35 forced stratified Laptev Sea continental shelf, episodic intermittent diapycnal mixing was observed  
36 when baroclinic tides and inertial currents gave rise to a rotating shear vector in the pycnocline that is  
37 amplified on semidiurnal time scales (Lenn et al., 2011). The effect of decreasing ice cover on the  
38 internal wave energetics, however, is not well established. Comparisons of internal wave energy  
39 between modern and historical data, reanalyzed in identical fashion, reveal no trend evident over the  
40 30-year period in spite of drastic diminution of the sea ice (Guthrie et al., 2013). The possible increase  
41 in internal wave forcing due to reduced sea ice cover may be offset by increased stratification by  
42 meltwater, which amplifies the dissipation of internal wave energy in the under-ice boundary layer.

43 The tidal mixing over topography controls the northward extension of temperate AW and thus sea ice  
44 cover variability (Holloway and Proshutinsky, 2007), and enhances dense water formation  
45 (Postlethwaite et al., 2011). Recent numerical model results show that there is significant internal tidal

1 wave generation in the Arctic Ocean, with baroclinic tidal energy dissipation structures similar to but  
 2 two-three orders of magnitude less than that observed on mid-Atlantic and Hawaiian ridges (Kagan et  
 3 al., 2011). The average coefficient of diapycnal diffusion is found to be less than the canonical value of  
 4 the vertical eddy diffusivity in the deep ocean prescribed in models of global ocean circulation, but  
 5 significant enough to influence the Arctic Ocean climate.

#### 7 **4.4 Double Diffusive convection in the Arctic Ocean.**

8 The role of double diffusion at the ice-ocean interface is discussed in Section 4.1. Here we address  
 9 double diffusion deeper in the ocean, far from the effects of the ice-ocean boundary layer (Figure 1).

10 The Arctic Ocean is very quiescent (Section 4.3). The level of turbulent kinetic energy is very low, and  
 11 this is a very favourable environment for double diffusion processes to occur. Double diffusion in the  
 12 ocean is due to different molecular diffusivities of temperature and salinity (Kelley et al., 2003). There  
 13 are two types of double diffusion in the ocean: Salt fingers occur when warm and salty water lies over a  
 14 cold and fresh water. In contrast, a cold and fresh water laying above a warm and salty water as it  
 15 occurs in the Arctic Ocean, is the preconditioning for the diffusive convection process. Steps like micro  
 16 structures in the vertical distribution of temperature, salinity and density are a manifestation of double  
 17 diffusion. Mixed layers alternate with sharp interfaces both in temperature, salinity and density.

18 Measurements during IPY revealed the ubiquitous nature of double diffusive steps in the Canada Basin  
 19 characterized by a surprisingly large spatial coherency of the steps over several hundreds of kilometers  
 20 (Timmermans et al., 2008). The mixed layers interleaving with the sharp interfaces were described as  
 21 small features of limited vertical extension (few meters) and related limited vertical heat fluxes ( $0.05 \text{ W}$   
 22  $\text{m}^{-2}$  to  $0.3 \text{ W m}^{-2}$ ). Detailed microstructure measurements in the central Arctic show a persistent  
 23 thermohaline staircase above the AW temperature maximum with an inferred average vertical heat flux  
 24 of  $0.6 \text{ W m}^{-2}$  (Sirevaag and Fer, 2012). The lateral coherency seen in the Canada Basin was, however,  
 25 absent in the Amundsen Basin (Sirevaag and Fer, 2012).

26 The main parameter characterizing double diffusion is the density ratio. This is the ratio between  $\beta$   
 27  $\delta S/\delta z$  and  $\alpha \delta\theta/\delta z$ , where  $\beta$  is the haline contraction coefficient,  $\alpha$  is the thermal expansion coefficient  
 28 of sea water, and  $\delta S/\delta z$  and  $\delta\theta/\delta z$  are the vertical salinity and temperature gradients, respectively. The  
 29 deepest part of the Arctic halocline was defined by Bourgain and Gascard (2011) as the depth where  
 30 the density ratio is equal to 20. At greater depth within the main thermocline density ratios are typically  
 31 between 1 and 10. The most favourable conditions for double diffusion to occur correspond to density  
 32 ratios approaching 1. In such conditions unstable temperature gradient develop through interfaces,  
 33 leading to more active convection in the mixed layers (e.g. Kelley et al 2003).

34 A structure of small steps in temperature and salinity profiles is also characteristic of double diffusion,  
 35 as observed during IPY (Timmermans et al., 2008; Sirevaag and Fer, 2012); an example from the  
 36 Amundsen Basin is shown in Figure 12a. In the continental slope of the Laptev Sea, Polyakov et al.  
 37 (2012) observed profiles with larger steps, which were remarkably persistent in time despite internal  
 38 waves, eddies and strong AW pulses increasing significantly the level of kinetic turbulent energy. This  
 39 large-step structure might be a result of a degenerative form of a double diffusion process, and it might  
 40 not be correct to calculate the vertical heat fluxes associated with those large steps applying the double  
 41 diffusion theory of Kelley et al. (2003). These large steps have not been observed in the past. The  
 42 vertical scales of the steps that are much larger than the typical diffusive layer thicknesses, however,  
 43 are comparable to the double diffusive, thermohaline intrusions frequently observed in the Arctic

1 (Carmack et al, 1997, Rudels et al, 1999, Kuzmina et al 2011). The intrusions are laterally coherent  
2 over thousands of km, with nested temperature-salinity structure, and are proposed to be driven and  
3 organized by double-diffusive processes (Walsh and Carmack, 2003). The intrusions emanate from the  
4 core of the AW in the slope current, and spread into the interior basin propagating heat and salt over  
5 long distances. An example of the intrusive features at three stations taken across the Lomonosov  
6 Ridge is shown in Figure 12c.

#### 9 **4.5 Submesoscale eddies, fronts, and other processes**

10 Submesoscale processes, here defined as on the order of Rossby deformation radius, which is typically  
11 several km in the upper Arctic water column, provide the link between mesoscale features (such as large  
12 frontal and current systems and large eddies, of the order of 100 km) and fine- and small-scale  
13 processes that contribute to diapycnal mixing in the ocean (Sect. 4.3). Submesoscale eddies, also  
14 referred to as submesoscale coherent vortices (SCV), are frequently observed in the Arctic, particularly  
15 in the Canadian Basin and along the ice edges and along the West Spitsbergen Current in Fram Strait  
16 (see Padman (1995) for a review). Using ice-tethered profilers covering as far north as 79°N,  
17 Timmermans et al (2008) analysed encounters of SCVs in the Canada Basin, and found their formation  
18 mechanism consistent with the instability of a surface front. Arctic SCVs isolate and transport  
19 anomalous water properties, and have implications for transport and lateral dispersion in the Arctic.  
20 Furthermore, Timmermans et al. (2012) observe re-stratification in the upper layers that can be  
21 attributed to lateral processes associated with submesoscale features. This has consequences for  
22 maintaining the insulating stratification of the CHL. The SCVs in ice covered waters of the Arctic  
23 Ocean are relatively shallow (300-500 m) and differ from those involved in open ocean deep  
24 convection, e.g. in the Greenland Sea. Observations from drifting floats in the Greenland Sea revealed  
25 the existence of SCVs composed of very homogeneous newly formed Greenland Arctic Intermediate  
26 Waters extending from near the surface down to 3000 m depth (Gascard et al., 2002). These SCVs had  
27 a 5 km diameter anticyclonic core with a time period of 2 to 3 days. They are transferring  
28 homogeneous oxygen rich waters from the shallow mixed layer deeper down through the main  
29 pycnocline to renew deep ocean layers and contribute to the large scale thermohaline circulation. These  
30 submesoscale deep convective SCVs are among all the eddies, those having the longest lifetime (several  
31 years) and this is the reason why they are so called SCV. They can only live where the ocean is deep  
32 enough (> 3000 m depth).

33 In pan-Arctic and global models, the SCVs are yet not resolved and must be parameterized. Their  
34 dynamics and resulting impact on vertical mixing are not properly understood or accounted for in the  
35 numerical models. Recent progress include the promising implementation by Fox-Kemper et al. (2011),  
36 however, the application in Arctic, under sea ice merit further research.

37 More attention is also needed for continental shelf waves trapped above the continental shelf break  
38 region all around the Arctic Ocean where resonance occurs during spring tides. This mechanism has a  
39 great potential to trigger sea-ice break up during springtime in MIZ and consequently to enhance sea-  
40 ice melting and retreat.

41

42

## 1 **5. Discussion**

### 3 **5.1 Main advances and remaining challenges in individual research fields**

4 Considering research on ABL processes and the vertical structure of the lower troposphere, much of the  
5 advance has been based on field experiments. For the SBL, the SHEBA observations have still been the  
6 starting point for a major part of recent advances. This demonstrates the high quality and uniqueness of  
7 the data set but, due to the major changes in the lower boundary conditions for the ABL since SHEBA  
8 in 1997-1998 (decrease in sea ice concentration and thickness), it simultaneously urgently calls for new  
9 year-round drifting stations with sophisticated ABL observations. In SBL research, major challenges  
10 remain in understanding and modelling of conditions of very stable stratification, in particular the  
11 interaction of waves and turbulence. Considering convective ABL over leads and polynyas, part of the  
12 recent advance has been based on utilization of improved remote sensing products on the ice  
13 concentration (e.g. Marcq and Weiss, 2012) and on coupled atmosphere – sea ice – ocean modeling  
14 (e.g. Ebner et al., 2011). Challenges remain in the high sensitivity of winter air temperatures to sea ice  
15 concentration (Lüpkes et al., 2008a, Tetzlaff et al. 2013), in the representation of new, thin ice in  
16 atmospheric models (Tisler et al., 2008) and in the interaction of convective plumes with the capping  
17 stable or near-neutral environment (Lüpkes et al., 2008b). In the dynamics of cold-air outbreaks over  
18 the open ocean, the new results linking the occurrence of roll convection with surface inhomogeneities  
19 in upwind sea ice (Liu et al., 2006; Gryschka et al., 2008) are an interesting advance, although the links  
20 are still under discussion. This work also demonstrates the need for close collaboration of atmospheric  
21 and sea ice scientists. Considering the occurrence and properties of temperature and humidity  
22 inversions, recent advance has been partly due to availability of new remote sensing data (Devasthale et  
23 al., 2010; 2011) but also simply due to increased interest on the issue (Nygård et al., 2013). Improved  
24 estimates on large-scale moisture advection and surface evaporation (Boisvert et al., 2012; 2013) are a  
25 prerequisite to better understand the processes controlling the vertical profile of air humidity.

26 Much of the advance in understanding and modelling Arctic clouds has been based on recent field data,  
27 above all in the circum-Arctic coastal observatories (Shupe et al., 2011) and during the I/B Oden  
28 expeditions in summers 2001 and 2008. The main advances have been related to the amounts of and  
29 partitioning between cloud liquid water and ice, radii of cloud droplets and ice crystals, decoupling  
30 between the surface and cloud layers, moisture sources from below and above the clouds, and  
31 production of turbulence in clouds. Challenges remain in improving our understanding of Arctic cloud  
32 physics (including the coupling of clouds, aerosols, radiative transfer, ABL turbulence, and cloud-  
33 generated turbulence) and even more in representing it in climate and NWP models. Limited horizontal  
34 and vertical resolution as well as a general lack of binned microphysical parameterizations mean that  
35 models will continue to rely on moist physics parameterizations based on more well-understood, lower  
36 latitude systems – which are likely not representative of Arctic conditions (e.g. Prenni et al., 2007). In  
37 the field of radiative transfer in the atmosphere, advance has taken place with respect to a better  
38 understanding of the interaction of radiation with cloud properties, condensation nuclei, surface albedo,  
39 near-surface turbulence, and heat conduction in snow and ice (Sedlar et al., 2011; Mauritsen et al.,  
40 2011). Comparisons against SHEBA data showed, however, that negative biases prevail in both  
41 shortwave and longwave downward radiation in several regional climate models (Tjernström et al.,  
42 2008). A better handling of the aerosol/cloud/radiation feedback is a prerequisite to improve model  
43 results for radiation balance at the sea ice and open ocean surface.

1 Considering fjordic and coastal processes, the advance has been supported by new aircraft  
2 observations, tethered sounding campaigns, and model experiments. Recent studies include the first  
3 comprehensive observations on barrier winds off southeastern Greenland (Petersen et al., 2009) and the  
4 first in situ observations of a tip jet off Cape Farewell (Renfrew et al., 2009a), and investigations of the  
5 governing dynamics of these flows. The presence of sea ice in Svalbard fjords has been found  
6 important for the dynamics of katabatic winds. It is now well demonstrated that various coastal and  
7 fjordic features can be accurately simulated with a sufficient model resolution of the order of kilometre,  
8 but it will take long before climate models can reach such a resolution.

9 The IPY was a focal point for extensive campaigns during which polar lows were observed.  
10 Operational weather forecasting systems have now reached the state where polar lows should be able to  
11 be predicted routinely. The recent development is above all related to better observing and data  
12 assimilation systems. Challenges remain, however, in the optimization of regional high-resolution  
13 ensemble prediction systems for Polar lows (Kristiansen et al. 2011).

14 Recent studies have demonstrated the importance of downward longwave radiation for the spring onset  
15 of snow melt on the Arctic sea ice (Persson, 2012; Maksimovich and Vihma, 2012). After the onset, the  
16 amount of melt is primarily controlled by the absorbed shortwave radiation. The albedo of snow  
17 evolves following the surface metamorphism and change of phases, from dry snow to melting snow,  
18 pond formation, pond drainage, pond evolution, and fall freeze-up (Perovich et al., 2009; Nicolaus et  
19 al., 2010a; Perovich and Polashenski, 2012). Numerous studies during and after IPY have addressed the  
20 snow and sea ice albedo (more than 70 papers cited here), the actual research topics including the  
21 spectral differences, spatial variations between various surface types, and effects of impurities such as  
22 black carbon. Further development of albedo parameterizations in climate and NWP models has been  
23 guided by the development of microscale models of the snow metamorphism (Flanner and Zender,  
24 2006), which allow the coupling between penetration of solar radiation into the snow and ice layer, the  
25 micro-scale characteristics of the ice crystals and the surface albedo. A proper validation of these  
26 parameterizations is, however, still missing. The development of new observation techniques for  
27 radiation (Nicolaus et al., 2010b; Hudson et al., 2012) and snow and ice properties (Arnaud et al., 2011;  
28 Gallet et al., 2009) has the potential to facilitate the future collection of high quality and complete  
29 datasets. There is also need for more realistic melt pond parameterizations, which, in addition to  
30 albedo, account for the latent heat, which has impact on the timing of fall freeze up. Further, more  
31 sophisticated snow aging parameterizations are needed, based on the inherent snow microphysical  
32 properties and accounting for the effects of liquid melt water on optical and thermal snow properties.

33 New results on sea ice structure have been largely based on application of the mushy-layer theory  
34 (Notz and Worster, 2009). This theory has proven particularly useful for better understanding the  
35 temporal evolution of sea ice salinity, in which the gravity drainage of salty brine and its replacement  
36 by less saline ocean water is essential. Process models work well for this desalination, and simplified  
37 parameterizations have been developed to describe it in large-scale models. Challenges remain in  
38 particular in realistically representing the fate of the draining brine in the oceanic boundary layer, and  
39 in realistically modeling the evolution of sea-ice salinity during periods of melt water flushing in  
40 summer. Regarding the basic issue of heat conduction in snow and ice, the need to take into account the  
41 effects of temperature and density on the snow heat conductivity is now better understood (Lecomte et  
42 al., 2011). Further, due to the spatial inhomogeneity of the snow cover, the need to use an effective heat  
43 conductivity of snow is well demonstrated (Semmler et al., 2012). Future perspectives with thinner sea  
44 ice and increasing precipitation suggest an increasing contribution of snow ice and superimposed ice in  
45 the Arctic sea ice mass balance. Modelling of these granular ice types has received attention, but

1 snow/ice models suffer from considerable inaccuracy in precipitation forcing (Cheng et al., 2008b;  
2 2013).

3 Considering small-scale dynamics of sea ice, the first estimates on mechanical weakening of sea ice in  
4 pan-Arctic scale were made via analysis of the response of sea ice to the Coriolis force. On the basis of  
5 buoy data and model experiments, Gimbert et al. (2012a,b) demonstrated that the strengthening of  
6 inertial oscillations in recent years (Figure 9) was partly a result of a genuine mechanical weakening of  
7 the ice cover, with a winter ice cover that nowadays mimics the mechanical behavior of summer sea ice  
8 20 to 30 years ago. The mechanical weakening of the ice has contributed to the accelerated drift.  
9 Seismometers installed on sea ice have allowed high-frequency monitoring of sea ice fracturing and  
10 faulting. The propagation speed of seismic waves has been found to depend on the ice thickness,  
11 allowing a novel method to estimate the latter in a regional scale (Marsan et al., 2012). Seismic  
12 observations also allow complementing satellite measurements by providing a much more detailed  
13 temporal sampling and therefore a better characterization of sea ice fracturing processes. Consequently,  
14 the next challenge is to extend the explorative DAMOCLES sea ice seismic survey to longer durations  
15 (at least a winter season) and to a broader-scale range, from the km scale to the regional (100 km)  
16 scale. In addition, an analysis of seismic noise induced by ocean-wave energy and recorded by land-  
17 based seismic stations installed at the periphery of the Arctic basin might be a way to monitor a proxy  
18 of the ice strength on a perennial basis (Tsai and McNamara, 2011).

19 The sea ice cover of the Arctic Ocean strongly reduces the energy input from the atmosphere, and  
20 thereby the mixing of the underlying water masses. Hence, mixing processes that do not play a large  
21 role elsewhere are often important in the Arctic. New results have demonstrated that above the  
22 subsurface temperature and salinity maxima of the Atlantic Water, the stratification is favourable for  
23 double-diffusive convection, which leads to vertical fluxes up to an order of magnitude larger than the  
24 molecular diffusion (Sirevaag and Fer, 2012). Apart from scarce direct microstructure measurements,  
25 our present quantification of double-diffusive fluxes depends on laboratory-based flux laws that may  
26 not be sufficiently accurate for geophysical environments. Recent observations following a storm event  
27 suggest that near-inertial response beneath the mixed layer can contribute significantly to vertical  
28 mixing within and below the CHL (Fer, 2014). The fraction of the near-inertial energy flux penetrating  
29 deep into the ocean and contributing to mixing, and particularly how it would change with ice cover, is  
30 uncertain. Challenges in understanding and modeling diapycnal mixing include the presence of large  
31 spatial variations: mixing is much more efficient along the continental rise and over topographic  
32 features, and the interplay between horizontal advection and diffusive mixing depends a lot on the  
33 location. Challenges also remain in quantitative understanding on the role of the ocean heat that reaches  
34 the surface: how large a portion escapes to the atmosphere and how much is used to melt the sea ice?  
35 Important topics that have not received enough attention in the recent years include deep ocean  
36 convection, continental shelf waves and the role of near-inertial forcing. These processes should be  
37 considered in large-scale modelling of the Arctic Ocean by developing appropriate parameterizations.

38

## 39 **5.2 Cross-disciplinary analogies**

40

41 Small-scale processes in the Arctic atmosphere, snow, sea ice, and the ocean cover a broad range of  
42 research areas. In some fields addressed here, such as turbulence in the atmosphere and ocean, the  
43 recent advances build on work that was started several decades ago, whereas some other issues, such as  
44 propagation of seismic signals in sea ice, represent very recently opened research fields. The older

1 research fields of atmospheric and ocean turbulence have a lot of analogy in recent advances and  
2 challenges. The interaction of waves and turbulence is an acute research topic both for the atmosphere  
3 and ocean. New evidence has been obtained that turbulence prevails in the atmosphere even under very  
4 stable stratification, which is related to the anisotropy of turbulence and to internal waves, which  
5 preserve vertical momentum mixing (Galperin et al., 2007). In the Arctic Ocean, the weakness of the  
6 internal wave field is a primary reason for the quiescent diapycnal mixing. Reduction of the sea ice  
7 cover is, however, expected to increase the background mixing levels (Rainville et al., 2011). Although  
8 the measurements in the MIZ are in support of this hypothesis (Fer et al., 2010), the effect of  
9 decreasing ice cover on the internal wave energetics is not yet well established (Guthrie et al., 2013).

10 During ice growth, the main uncertainty in modeling the turbulent exchange of heat and salt at ice-  
11 water interface originates from the roughness length  $z_{0B}$ . The observational values include a large  
12 scatter, and a major question is how to parameterize the role of form drag due to flow edges and keels.  
13 In the atmosphere, the new parameterizations for  $z_0$  have dealt with the same issue: the role of ridges,  
14 flow edges, melt pond edges, and sastrugi in generation of form drag (Andreas et al., 2010a,b; Andreas,  
15 2011; Lüpkes et al., 2012a; 2013). The  $z_0$  values applied in large-scale atmospheric models, however,  
16 sometimes strongly differ from the results of field experiments, because  $z_0$  is used as a tuning  
17 parameter. In ocean models, the angle between the ice-ocean stress and ice drift vectors is often used  
18 similarly (Uotila et al., 2014).

19 Furthermore the dominant vertical structures controlling stratification in the Arctic atmosphere and  
20 ocean, the temperature inversion and ocean halocline, have an analogy in the sense that both are  
21 strongly affected by the horizontal advection (of heat and salt, respectively). Challenges remain in  
22 better quantifying these advective fluxes, their vertical profiles, and their interaction with small-scale  
23 processes. Differences between the atmosphere and ocean include double diffusion that only occurs in  
24 the ocean and the strong stabilizing role of melt water at the ice bottom. The latter makes double  
25 diffusion an important limiting factor in the OBL during the melt season (in addition to its importance  
26 in the quiescent interior of the ocean).

### 29 **5.3 Feedback mechanisms**

30 Understanding the role of small-scale processes in the Arctic climate system is complicated by  
31 numerous feedback effects. Positive feedbacks are essential in explaining the observed Arctic  
32 amplification of the climate warming (Serreze and Barry, 2011; Pithan and Mauritsen, 2014; Döscher  
33 et al., 2014), and feedbacks related to small-scale processes are often interacting with changes in large-  
34 scale transports in the atmosphere (Langen et al., 2012) and ocean (Bitz et al., 2006). Here we focus on  
35 the feedbacks related to small-scale processes, which include the albedo, water vapour, aerosol-cloud-  
36 radiation, Planck, and lapse-rate feedbacks. Several recent studies have stressed the close connections  
37 between these processes.

38 The surface albedo feedback (SAF) mechanism is reputed to be an important contributor to the loss of  
39 Arctic sea ice over the last few decades (Screen and Simmonds, 2010b; Crook et al., 2011; Taylor et  
40 al., 2013). By synthesizing a variety of remote sensing and field measurements, both Flanner et al.  
41 (2011) and Hudson (2011) concluded that the change in the radiative impact of the Arctic sea-ice at the  
42 top of the atmosphere in the period 1979-2008 has been a reduced cooling of about  $0.1 \text{ Wm}^{-2}$ .  
43 Combining this finding with the observed Northern Hemisphere warming, the Northern Hemisphere

1 sea ice albedo feedback is between  $0.17$  and  $0.54 \text{ W m}^{-2} \text{ K}^{-1}$  (or between  $0.33$  and  $1.07 \text{ W m}^{-2} \text{ K}^{-1}$  if the  
2 effect of land-based snow is included) (Flanner et al., 2011). These values are substantially larger than  
3 comparable estimates obtained from 18 climate models of the CMIP3 dataset (Flanner et al., 2011).  
4 Considering future climate projections of Arctic sea-ice, Hudson (2011) estimated that in an ice-free  
5 summer scenario the radiative forcing caused by the albedo reduction would be about  $0.3 \text{ W m}^{-2}$ , similar  
6 to the present-day anthropogenic forcing caused by tropospheric ozone pollution or by halocarbon  
7 emissions (Forster et al., 2007). Several studies have concluded that the Arctic climate system does not  
8 have an irreversible tipping point behaviour associated with the SAF (Stranne and Björk, 2011; Armour  
9 et al., 2011; Tietsche et al., 2011). However, Müller-Stoffels and Wackerbauer (2012) showed that the  
10 shape of the albedo parameterization near the melting temperature differentiates between reversible  
11 continuous sea ice decrease under atmospheric forcing and a hysteresis behaviour.

12 The SAF is strongly linked to the change in the phase of precipitation. The observed decline in summer  
13 snowfall and increase in rain over the Arctic Ocean and Canadian Archipelago has resulted in a  
14 substantial decrease in the surface albedo (Screen and Simmonds, 2012). Further, melt ponds enhance  
15 the SAF because of enhanced melt pond coverage in a warmer climate, while aerosol deposition on ice  
16 (when kept constant) reduces the SAF, because of enhanced melt-out of aerosols in a warmer climate  
17 (Holland et al., 2012). Thus, the impact of particulate impurities on snow and sea ice is expected to  
18 decrease in a doubling- $\text{CO}_2$  scenario (Holland et al., 2012; Goldenson et al., 2012). Finally, the SAF  
19 can be enhanced by mechanical processes: a thinner, less concentrated sea ice cover is weaker (Gimbert  
20 et al., 2012b), which results in increasing fracturing and lead opening. These have an indirect effect on  
21 albedo, as splitting up of the ice field increases lateral melt and, hence, decreases the area-averaged  
22 albedo.

23  
24 Although SAF has received most attention, it is not certain if it is the strongest feedback in the Arctic  
25 climate system. One of the major problems in understanding SAF is its close interaction with cloud  
26 changes (Figure 7). Sedlar et al. (2011) observed that sea ice albedo is a strong modulator of the cloud  
27 shortwave radiative forcing (which decreases with increasing surface albedo) and of the near-surface  
28 temperature. Graverson and Wang (2009) estimated that most of the polar amplification of the surface-  
29 air temperature is not directly attributable to the SAF itself, but rather to the SAF strengthening of the  
30 water-vapour and cloud feedbacks, which have a greenhouse effect that is larger in the Arctic than at  
31 lower latitudes. On the other hand, the presence of clouds over sea-ice reduces the radiative forcing due  
32 to changes in sea ice concentration and albedo. Indeed, Hudson (2011) showed that the present-day  
33 cloud cover manages to mask approximately half of the clear-sky sea-ice albedo feedback, while  
34 Mauritsen et al. (2013) found a dominating role of the water-vapour feedback. Generally, a reduction in  
35 sea-ice extent is expected to cause an increase in cloud cover, but this relationship seems quite weak in  
36 summer (Eastman and Warren, 2010; Kay and Gettelman, 2009), when the sea-ice albedo feedback is  
37 most important.

38  
39 In addition to albedo, the cloud radiative forcing and related feedback are sensitive to the number of  
40 CNN available. During ASCOS, even at 100% relative humidity, Mauritsen et al. (2011) observed  
41 clouds optically thin enough to be undetectable by the eye: “tenuous clouds”. Two regimes were found  
42 with an approximate division at CCN concentrations near  $10 \text{ cm}^{-3}$ . When CCN was lower than this  
43 threshold, clouds would be “gray” in the infrared, and an increase in CCN would lead to an increase in  
44 downwelling radiation that far outweighed the simultaneous decrease in downwelling shortwave  
45 radiation; this gives rise to a warming effect at the surface. Conversely, when CCN concentrations were  
46 higher, further increases in CCN concentrations instead lead to reduced downwelling shortwave

1 radiation causing a cooling effect at the surface, while clouds are already black in the infrared resulting  
2 in little or no change in the longwave. Perusing CCN observations from four expeditions to the summer  
3 Arctic, Mauritsen et al. (2011) speculate that the tenuous clouds regime may occur up to 30% of the  
4 time in summer; also see Tjernström et al. (2014).

5  
6 The lapse-rate feedback is related to the vertical structure of the warming. In the tropics, due to the  
7 deep convection and strong release of latent heat during cloud condensation throughout the  
8 troposphere, a small temperature increase enough to compensate for a certain radiative imbalance at the  
9 top of the atmosphere. In the Arctic, however, due to the prevailing stable stratification, vertical mixing  
10 is limited and surface warming does not reach high altitudes. Hence, a larger near-surface temperature  
11 increase is needed to compensate for the same radiative imbalance as in the tropics (Bintanja et al.,  
12 2012; Pithan and Mauritsen, 2014). The often overlooked Planck feedback results from the fact that the  
13 longwave radiation emitted by the Earth's surface and atmosphere is proportional to the fourth power  
14 of the absolute temperature. Hence, a certain increase in emitted longwave radiation corresponds to a  
15 larger temperature increase in the Arctic than at lower latitudes (Pithan and Mauritsen, 2014). Even  
16 without any other feedback mechanisms, an increase in the greenhouse gas concentrations would cause  
17 a small Arctic amplification. On the basis of CMIP5 climate model results, Pithan and Mauritsen  
18 (2014) argue that the largest contribution to Arctic amplification originates from the combined effects  
19 of the lapse-rate and Planck feedbacks, the former being more important. Their net effect is that when  
20 the Earth surface warms, less energy is radiated back to space in the Arctic than at lower latitudes.

21 An issue not to be confused with the lapse-rate feedback is the small heat capacity of a shallow ABL  
22 (typically SBL). A certain heat input results in a larger temperature increase in a shallow than in a deep  
23 ABL. As the ABL is typically shallow in the Arctic, this may have contributed to the Arctic  
24 amplification of climate warming (Esau and Zilitinkevich, 2010; Esau et al., 2012). It is, however, not a  
25 positive feedback, as heating of the ABL tends to increase its thickness.

26 The diapycnal mixing in the Arctic Ocean, in addition to double diffusion where favourable, is  
27 primarily driven by breaking internal waves that are forced by tides or wind. In an Arctic Ocean with a  
28 larger fraction of open water areas, the internal wave field is expected to be energized through more  
29 input of wind and near-inertial energy, which in turn leads to enhanced mixing. Increased amounts of  
30 oceanic heat from the AW layer can thus reach the under-ice boundary. Resulting increase in melting  
31 rates may lead to a positive feedback that needs to be studied. The implications may be more  
32 significant near the shelf break where the increased wind-driven energy can influence the AW  
33 boundary current dynamics and cross-slope exchange processes.

34 Feedbacks also occur in partly-resolved scales, e.g. related to the occurrence of Polar lows or ocean  
35 eddies. An accurate representation of feedbacks continues to be one of the major challenges in  
36 modeling of the Arctic and global climate change. For example, the nature of sea ice loss – whether it  
37 will be reversible or not – is sensitive to the parameterization of feedbacks (Eisenman and Wettlaufer,  
38 2009; Müller-Stoffels and Wackerbauer, 2012). The present level of uncertainty is characterized by the  
39 fact that recent improvements in the ECMWF land snow scheme have resulted in doubling of the snow-  
40 albedo feedback (Dutra et al., 2012). Further, the net effect of all the feedbacks taking place in the  
41 Arctic is difficult to assess because they operate in different spatial and temporal scales (Callaghan et  
42 al., 2012).

43

44

## 1 **5.4 Representativeness of results**

2 The results reviewed here are based on observations and model experiments, but the former are not  
3 uniformly distributed in space and time. The climatological representativeness of observations has been  
4 studied a lot (e.g. Bourassa et al., 2013). The representativeness of observations from the point of view  
5 of process understanding is, however, a different issue. In some respect, spatial and temporal variations  
6 are less crucial for process understanding than for climatology; as soon as a process is physically  
7 understood, gaps in data are no more a problem. However, it is often difficult to know if the state of  
8 sufficient physical understanding has been reached, or if the process is sensitive to changes in some  
9 boundary conditions that require further observations. This makes it difficult to quantify the  
10 representativeness of observations from the point of view of process understanding.

11 Various spatial and temporal scales are relevant here, but the most serious issue is the very limited  
12 amount of data available from winter and late autumn, when many small-scale processes are certainly  
13 different due to the lack of solar radiation. The only significant winter and late autumn in-situ data  
14 sources originate from SHEBA, the Russian drifting stations, and coastal observatories, with the  
15 majority of literature relevant for this review based on SHEBA. When most results for winter processes  
16 are based on a single campaign, it raises the question of how sensitive the small-scale processes were to  
17 the conditions that happened to occur during that particular winter. Considering other seasons, it is not  
18 clear if the temporal unevenness in the amount of data has significantly affected the understanding and  
19 parameterization of processes, but the availability of data varies also between other seasons, often due  
20 to logistical reasons. Examples of these include an easier access to sea ice by aircraft and helicopter  
21 during spring than other seasons, and an easier access to the northern parts of the Arctic Ocean by  
22 research vessels in late summer and early autumn than other seasons. It is clear that the observation  
23 method affects the representativeness and interpretation of the result (see Section 2.1.4 for sea ice  
24 roughness). In some respects, buoy observations build a bridge between research vessel and airborne  
25 surveys (Richter-Menge et al., 2006), but not for all variables that are needed in studies of small-scale  
26 processes.

27 In the coastal and archipelago areas, the representativeness of observations is naturally a major issue  
28 but even in the central Arctic, far from direct influence of land and sea-floor orography, the boundary  
29 conditions for small-scale processes are affected by the large-scale flow in the ocean and atmosphere  
30 and related advection of heat, moisture, and salt. Hence, it is difficult to estimate how representative  
31 our observationally based knowledge on small-scale processes truly is, bearing in mind that a large  
32 portion of the best data sets have been gathered from rather limited regions, such as the Beaufort /  
33 Chukchi Sea and the Atlantic sector of the Arctic. A new challenge is that observations may get less  
34 representative when the amount of thick ice is decreasing. In addition to sea ice and snow research, this  
35 is a problem also for meteorology and oceanography. Due to obvious safety reasons, manned ice  
36 stations and expensive automatic measurement devices are typically deployed on fairly thick sea ice.  
37 Not much information are available on the quantitative effects of these observational biases, but Inoue  
38 et al. (2009) have suggested that accuracy of reanalyses may decrease due to smaller sea ice areas  
39 available for buoy deployments.

40 Sea ice and snow thermodynamics is one of the processes most liable to small-scale spatial variations.  
41 Due to sastrugi, melt ponds, ice ridges and keels, rafted floes, cracks, and small leads, significant  
42 variations are present already in scales of less than a metre. In the case of measurements at manned ice  
43 stations, such variations can be mapped (e.g. Hudson et al., 2012), but in the case of buoys (e.g. ice  
44 mass-balance buoys) uncertainty often remains on the small-scale surroundings of the measurement  
45 site. Although buoys are typically deployed on sites as representative as possible (Richter-Menge et al.,

1 2006), these sites may gradually change to become less representative, especially during the melting  
2 season. It is therefore essential that studies on sea ice and snow thermodynamic processes are based on  
3 a large amount of in-situ data, preferably supported by remote sensing data and model experiments.

## 5 **6. Conclusions and Outlook**

6 We have reported advances in the development of parameterizations for the surface albedo, melt ponds,  
7 turbulent surface fluxes, desalination of sea ice, snow thermal conductivity, ablation rate at the ice  
8 bottom, double-diffusive transport, and submesoscale coherent vortices. In cloud physics, radiative  
9 transfer in the atmosphere, sea ice small-scale dynamics, and diapycnal mixing in the ocean, the recent  
10 advance in physical understanding has not yet yielded remarkable improvements in parameterizations.  
11 Ideally, the advance in physical understanding and parameterization should progress hand in hand:  
12 large model errors may suggest that something is wrong or insufficient in the physical understanding,  
13 which generates a need for more process studies, which improve the physical understanding and further  
14 result in improved parameterizations. In practice, however, the improvement of large-scale models  
15 often takes place after some delay. The reasons for this are manifold, including (a) the limited  
16 computational power, (b) the need to prioritize among the large number of issues that need  
17 improvements in models, (c) too little communication between observationalists and large-scale  
18 modellers, (d) too little communication between disciplines, and (e) compensating errors in models,  
19 which stop balancing each other out. The development of parameterizations is further complicated by  
20 the lack of understanding on how much complexity is cost-effective.

21 A key difference between partially resolved processes (such as polar lows, orographic flows, and ocean  
22 mesoscale eddies) and processes that are only parameterized is that further increases in grid resolution  
23 will eventually enable good representation of the former in NWP and climate models. In the mean-time  
24 (next decade or two), however, parameterizations of processes on both scales remain necessary. Hence,  
25 on both scales, we have to accept the fact that uncertainty and errors will remain in parameterizations.  
26 Future challenges include to quantitatively understand how much these errors are related to (a) the fact  
27 that many recent findings on small-scale physics have not yet been (fully) implemented in model  
28 parameterizations, (b) our lack of understanding of the processes, and (c) our inability to parameterize  
29 them using grid-resolved variables. Further, accepting the fact that parameterizations will always have  
30 errors, more work is needed to develop and apply methods such as stochastic physics in ensemble  
31 prediction systems, as already done in some climate (Palmer and Williams, 2010) and NWP models  
32 (Krasnopolsky et al., 2013).

33 Considering climate modelling for this century, the sources of uncertainty can be roughly divided into  
34 three groups: (1) internal variability of the system, (2) model uncertainty, and (3) scenario uncertainty.  
35 According to Hawkins and Sutton (2009), the uncertainty related to internal variability dominates over  
36 the first decade of a model run, the model uncertainty dominates over the fourth decade, and the  
37 scenario uncertainty dominates over the ninth decade, except in high-latitudes. There the model  
38 uncertainty is so large that it still dominates over the ninth decade. A major challenge for the Arctic  
39 research community is to reduce the dominating model uncertainty.

40 The concrete path towards better understanding and parameterization of small-scale physical processes  
41 in the Arctic is multifaceted. First, further advance can be made via more systematic and cross-  
42 disciplinary analyses of existing observations supported by model experiments devoted to improvement  
43 of parameterizations, applying both large-scale and process models (including LES). Large-scale

1 operational and climate models are essential to evaluate how well the interaction of individual  
2 processes at different temporal and spatial scales is reproduced, preserving process relationships as  
3 diagnosed from observations. Attention should also be paid on the optimal utilization of new recent  
4 remote satellite sensing products, such as the SMOS (Soil Moisture and Ocean Salinity) data on thin  
5 ice thickness, new generation Radio Occultation instruments and sounders for atmospheric remote  
6 sensing, as well as fully exploiting the potential of MODIS, Calipso, Cloudsat, and EarthCARE data on  
7 (mixed-phase) clouds. The WMO Polar Prediction Project (PPP) is expected to have a major role in  
8 coordination of the data analyses and modeling activities. PPP will include an intensive phase: the Year  
9 of Polar Prediction (YOPP) in 2017-2019.

10 Second, after sixteen years since the end of SHEBA, we desperately need more year-round field  
11 observations, including both in situ and ship/ice/aircraft-based remote sensing observations. It is  
12 essential that the observations are made in extensive, multi-disciplinary campaigns, so that the  
13 interaction of different variables and processes can be observed. Expectations for new process-level  
14 observations on the Arctic atmosphere-sea-ice-ocean system are laid at the doorstep of MOSAiC  
15 (Multidisciplinary drifting Observatory for Studies of Arctic Climate, described at  
16 [www.mosaicobservatory.org/](http://www.mosaicobservatory.org/)), a year-round field campaign planned for the time frame of 2017-2019.  
17 MOSAiC will overlap with YOPP, which will provide excellent possibilities for coordination of  
18 observations and model experiments. To improve the representativeness of observations (Section 5.4) a  
19 large spatial coverage of observations will be essential, so that observations at the main ice station will  
20 need to be supported by a network of autonomous ice-based stations, airborne observations (research  
21 aircraft, helicopters, unmanned aerial vehicles), underwater gliders, other research vessels, and  
22 intensive campaigns at coastal stations.

23 It is essential to develop novel observational methods focusing on the “New Arctic”, characterized by,  
24 among others, larger areas of open water and thin ice, longer periods of snow and ice melt, and more  
25 rain instead of snow fall. Increasingly important processes to be studied include the autumn freeze-up,  
26 snow on sea ice, wave-ice interaction, and storm effects. Observations over thin ice will generate  
27 challenges for instrument deployment. Hence, further development of remote sensing methods is  
28 essential to obtain a good spatial and temporal coverage, and the role of unmanned aerial vehicles (e.g.  
29 Inoue et al., 2008; Reuder et al., 2012), dropsondes, controlled meteorological balloons (Voss et al.,  
30 2013), and autonomous underwater vehicles (e.g. Doble et al., 2009) is expected to increase.  
31 Underwater gliders have recently proven to be a suitable platform for ocean microstructure  
32 measurements (Fer et al., 2014). Coordinated planning of new observations is needed to maximize the  
33 utilization and mutual support of in-situ and remote sensing data. Observational requirements need to  
34 be well defined and to be communicated to space agencies for future mission design. In some fields,  
35 such as snow and ice physics, field experiments could also be more systematically supported by  
36 laboratory experiments.

37 It is noteworthy that better understanding and modeling of small-scale processes in the Arctic is  
38 essential not only for the Arctic climate system but also for the mid-latitudes. Sea ice decline in the  
39 Arctic has had some, although mostly poorly understood, effects on the large-scale atmospheric  
40 circulation (see Vihma (2014) and Walsh (2014) for recent reviews). The effects reaching mid-latitudes  
41 originate from changes in small-scale processes in the Arctic, including interaction of convection and  
42 baroclinic processes (Petoukhov and Semenov, 2010), destruction of the low-level temperature  
43 inversion (Deser et al., 2010), a deepening of the ABL (Francis et al., 2009), and destabilization of the  
44 lower troposphere (Jaiser et al., 2012). Bearing in mind the large errors still present in reanalyses and

1 climate models (see the Introduction), these findings call for more research on small-scale processes in  
 2 the Arctic.

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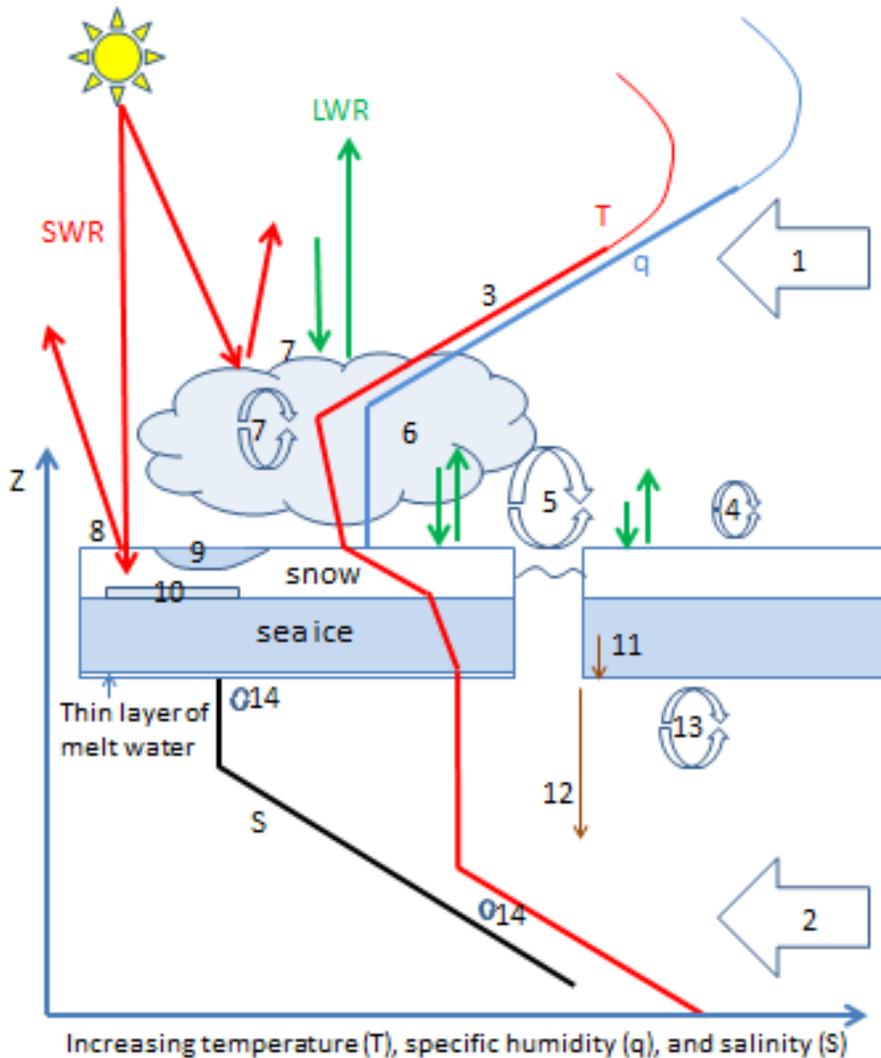
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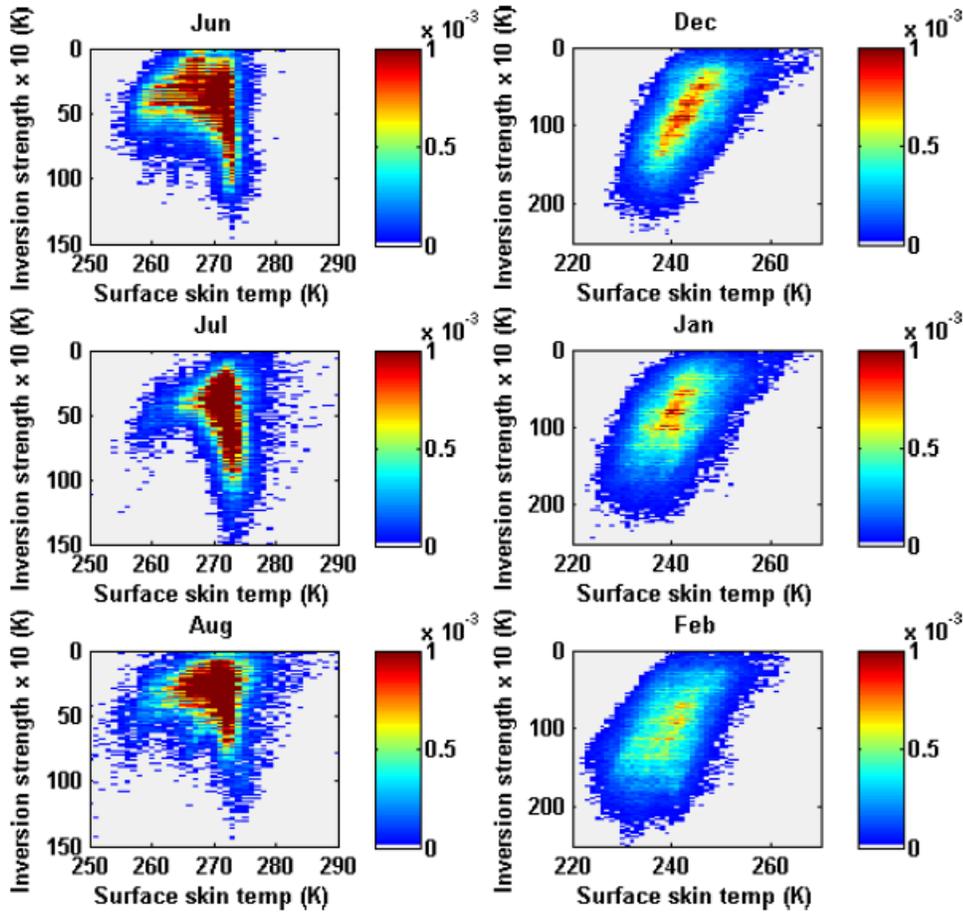
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1 Table 1. List of acronyms

Acronym	Definition
ABL	atmospheric boundary layer
ASCOS	Arctic Summer Cloud Ocean Study
ASR	Arctic System Reanalysis
AW	Atlantic Water
BC	black carbon
CAM4	Community Atmospheric Model version 4
CAO	cold-air outbreak
CHL	cold halocline layer
CICE	The Los Alamos sea ice model
CMIP3	Coupled Model Intercomparison Project
CNN	cloud condensation nuclei
ECHAM5	5th generation of the ECHAM general circulation model
ECMWF	European Centre for Medium-Range Weather Forecasts
EPS	ensemble prediction system
ERA-Interim	an atmospheric reanalysis by the ECMWF
GFDex	Greenland Flow Distortion Experiment
HIRLAM	High-Resolution Limited Area Model
IPCC AR4(5)	Intergovernmental Panel on Climate Change, Assessment Report 4(5)
IPY	International Polar Year 2007-2009
IWC	ice water content
IWP	ice water path
LES	large-eddy simulation
LIM (2)	Louvain-la-Neuve Sea Ice Model (two-level version)
LLJ	low-level jet
LWC	liquid water content
LWP	liquid water path
MIZ	marginal ice zone
MODIS	Moderate Resolution Imaging Spectroradiometer
MOSAiC	Multidisciplinary Drifting Observatory for the Study of Arctic Climate
MPS	mixed-phase stratocumulus
NWP	numerical weather prediction
PPP	Polar Prediction Project
QNSE	Quasi-Normal-Scale Elimination (method)
$RH_{liq}$	air relative humidity with respect to liquid water
SAF	surface albedo feedback
SBL	stable boundary layer
SCV	submesoscale coherent vortex
SHEBA	Surface Heat Budget of the Arctic Ocean
SMOS	Soil Moisture and Ocean Salinity (satellite)
SNICAR	Snow and Ice Aerosol Radiation (model)
SPOT	Satellite Pour l'Observation de la Terre (Satellite for Observation of the Earth)
SZA	solar zenith angle
TKE	turbulent kinetic energy
TPE	turbulent potential energy
WBF	Wegener-Bergeron-Findeisen (process in cloud physics)
WMO	World Meteorological Organization
WRF	Weather Research and Forecasting (model)
YOPP	Year of Polar Prediction



1  
 2 Figure 1. Simplified presentation of physical processes and vertical profiles of temperature (T), air  
 3 humidity (q), and ocean salinity (S) in the marine Arctic climate system. In reality the shape of the  
 4 profiles varies in time and space. The numbers indicate the following processes: 1. atmospheric  
 5 advection of heat and moisture to the Arctic, 2. oceanic advection of heat and salt to the Arctic, 3.  
 6 generation of temperature and humidity inversions, 4. turbulence in stable boundary layer, 5.  
 7 convection over leads and polynyas, 6. cloud microphysics, 7. cloud-radiation-turbulence interactions,  
 8 8. reflection and penetration of solar radiation in snow/ice, 9. surface melt and pond formation, 10.  
 9 formation of superimposed ice and snow ice, 11. gravity drainage of salt in sea ice, 12. brine formation,  
 10 13. turbulent exchange of momentum, heat and salt during ice growth, and 14. double-diffusive  
 11 convection. More detailed illustration of small-scale processes is given in Figures 2-12.



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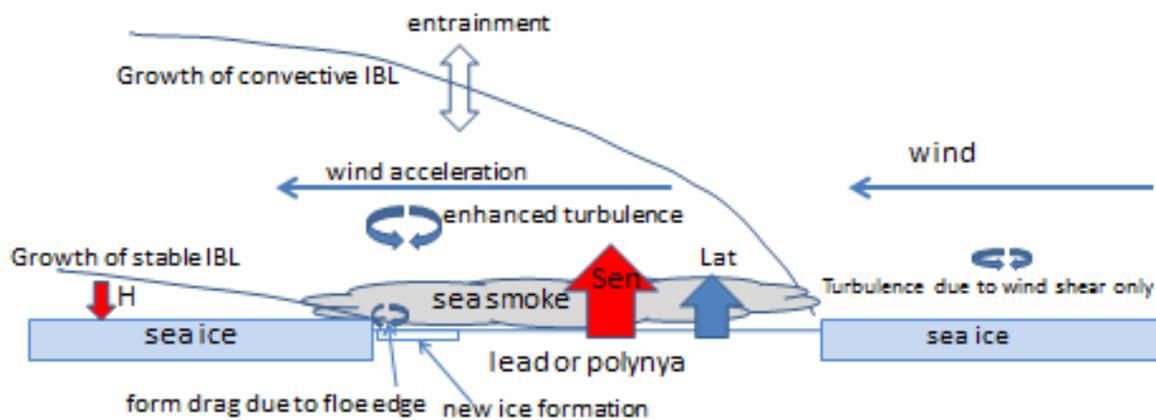
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3 Figure 2, Histograms of inversion strength and surface temperature for summer (left column) and  
 4 winter (right column) months in the Arctic, based on Atmospheric Infrared Sounder data. Note that the  
 5 x and y axes are different for summer and winter months and inversion strength is multiplied by 10.  
 6 Each temperature-temperature bin is normalized by the total number of observations in the entire  
 7 histogram. Reproduced with permission from Devasthale et al. (2010).

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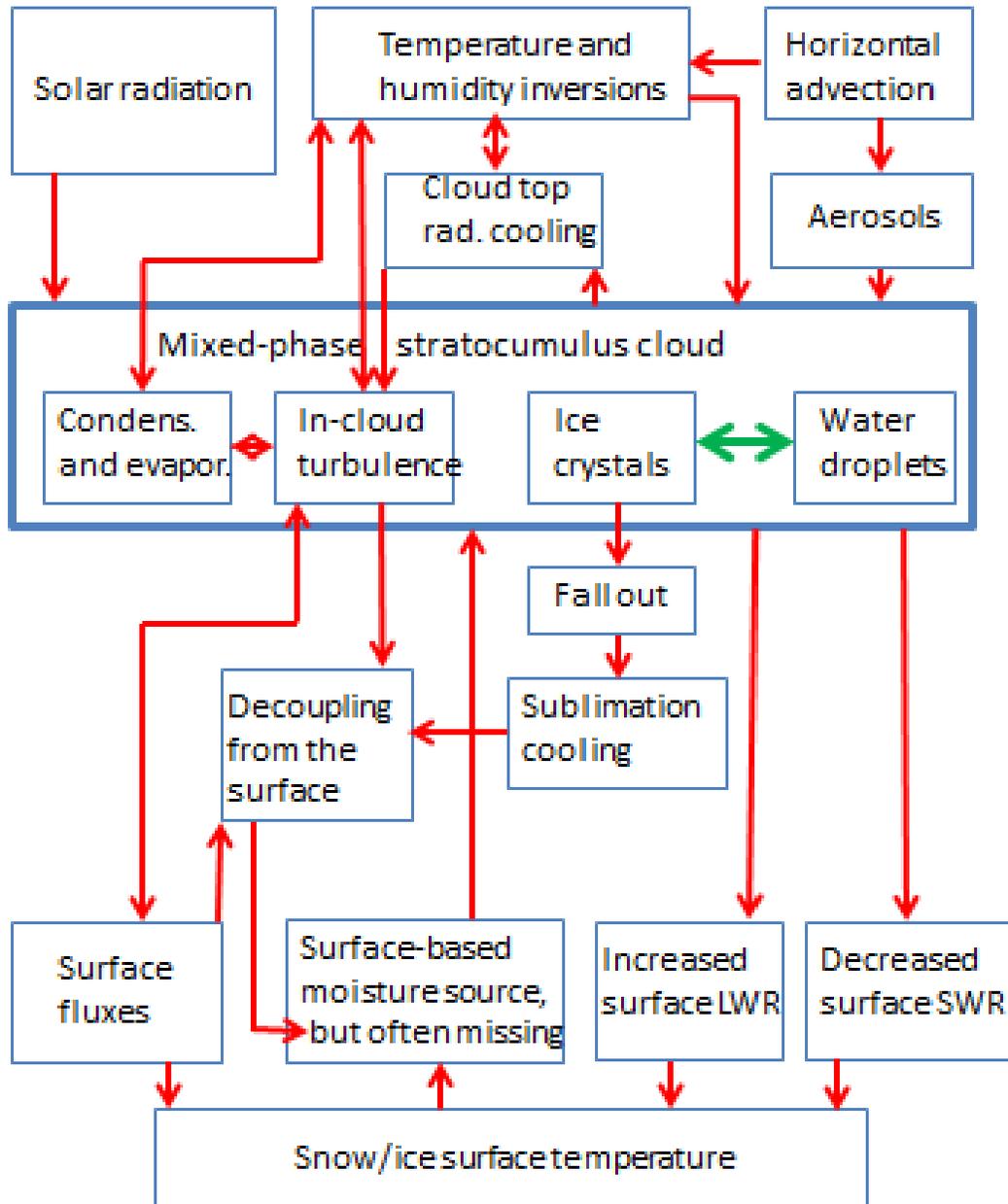


1 (a)  
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3 (b)



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5 Figure 3. Convection over leads and polynyas: (a) sea smoke originating from leads in the Fram Strait  
6 on 7 March 2013 (photo: C. Lüpkes), (b) schematic presentation of ABL processes over a lead /  
7 polynya. Sen and Lat are the turbulent fluxes of sensible and latent heat, respectively.

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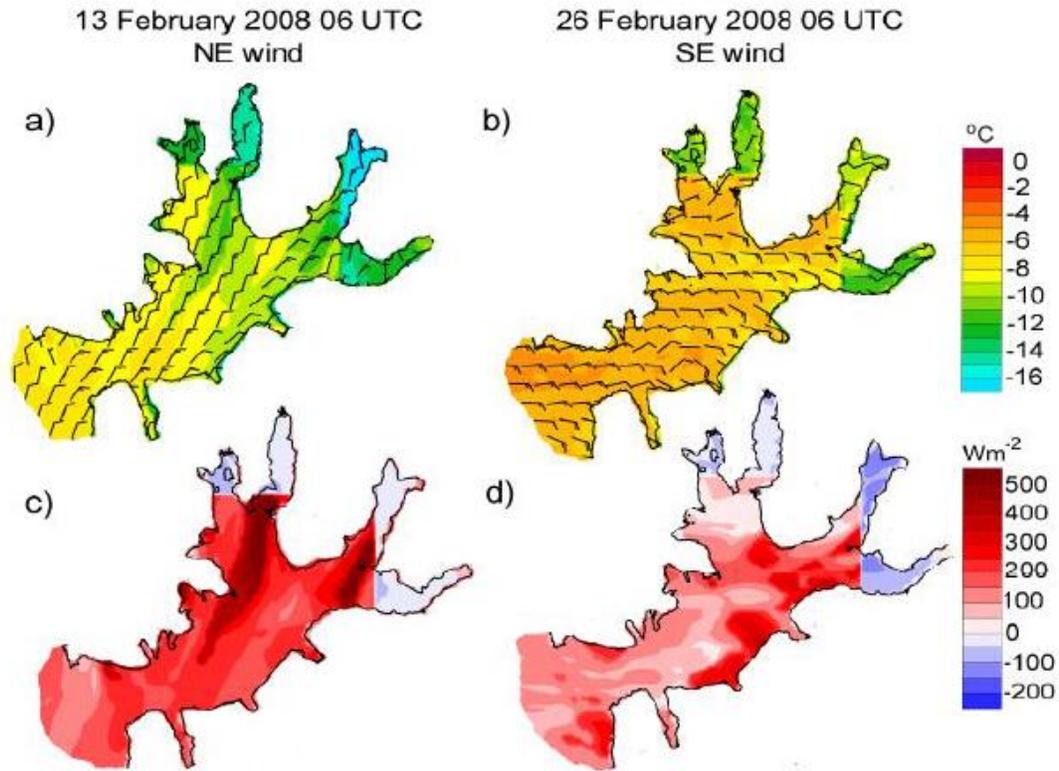


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2 Figure 4. Schematic diagram on the effects and interactions related to mixed-phase stratocumulus  
 3 clouds and radiative transfer. Macro- and microphysical processes and interactions are shown as  
 4 arrows, the green arrow representing numerous microphysical processes related to aerosols, nucleation,  
 5 evaporation, depositional ice growth, cloud layer glaciation, and effects of saturation vapour pressure  
 6 differences of liquid and ice (see e.g. Morrison et al. 2012).

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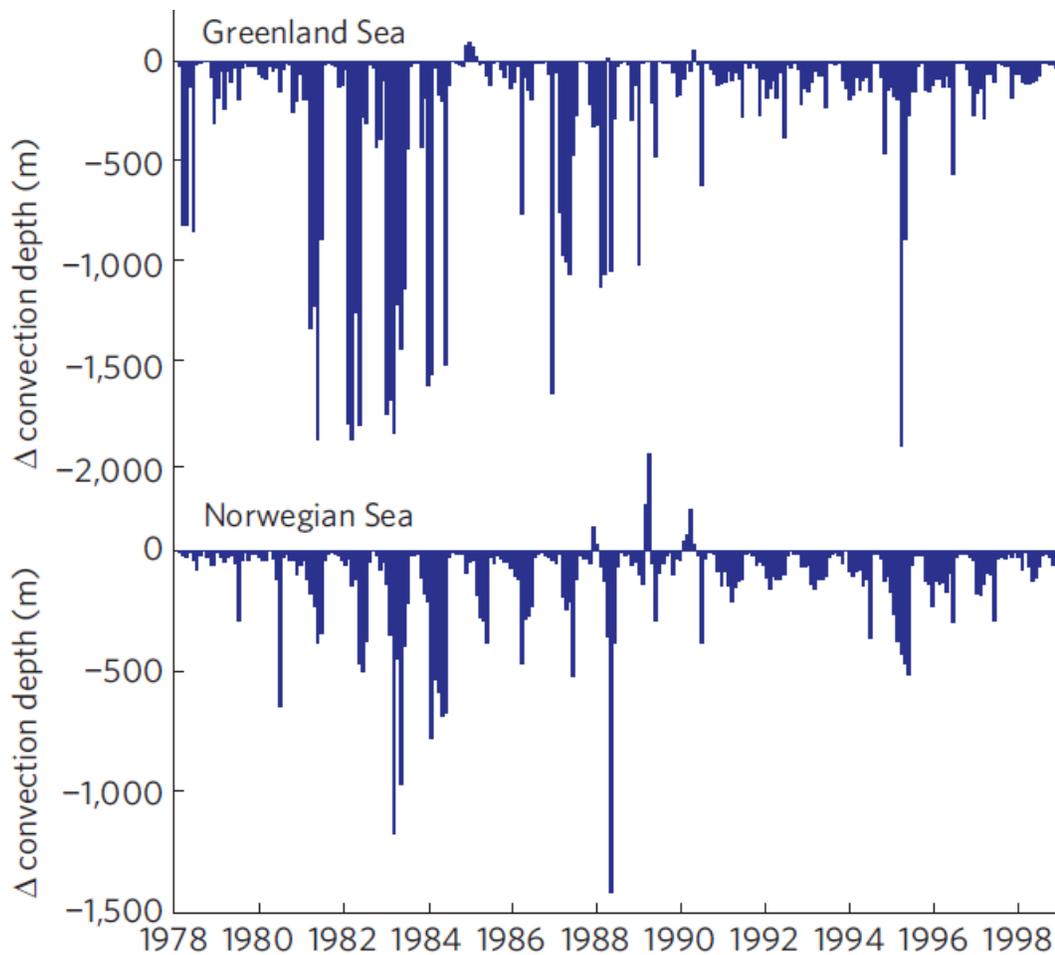


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4 Figure 5. Examples demonstrating large spatial variations in air temperature and wind (a and b) and  
 5 sensible heat flux (c and d) over a complex fjord (Isfjorden in Svalbard, length approximately 100 km),  
 6 as simulated applying a high-resolution atmospheric model. Redrawn with permission from Kilpeläinen  
 7 (2011).

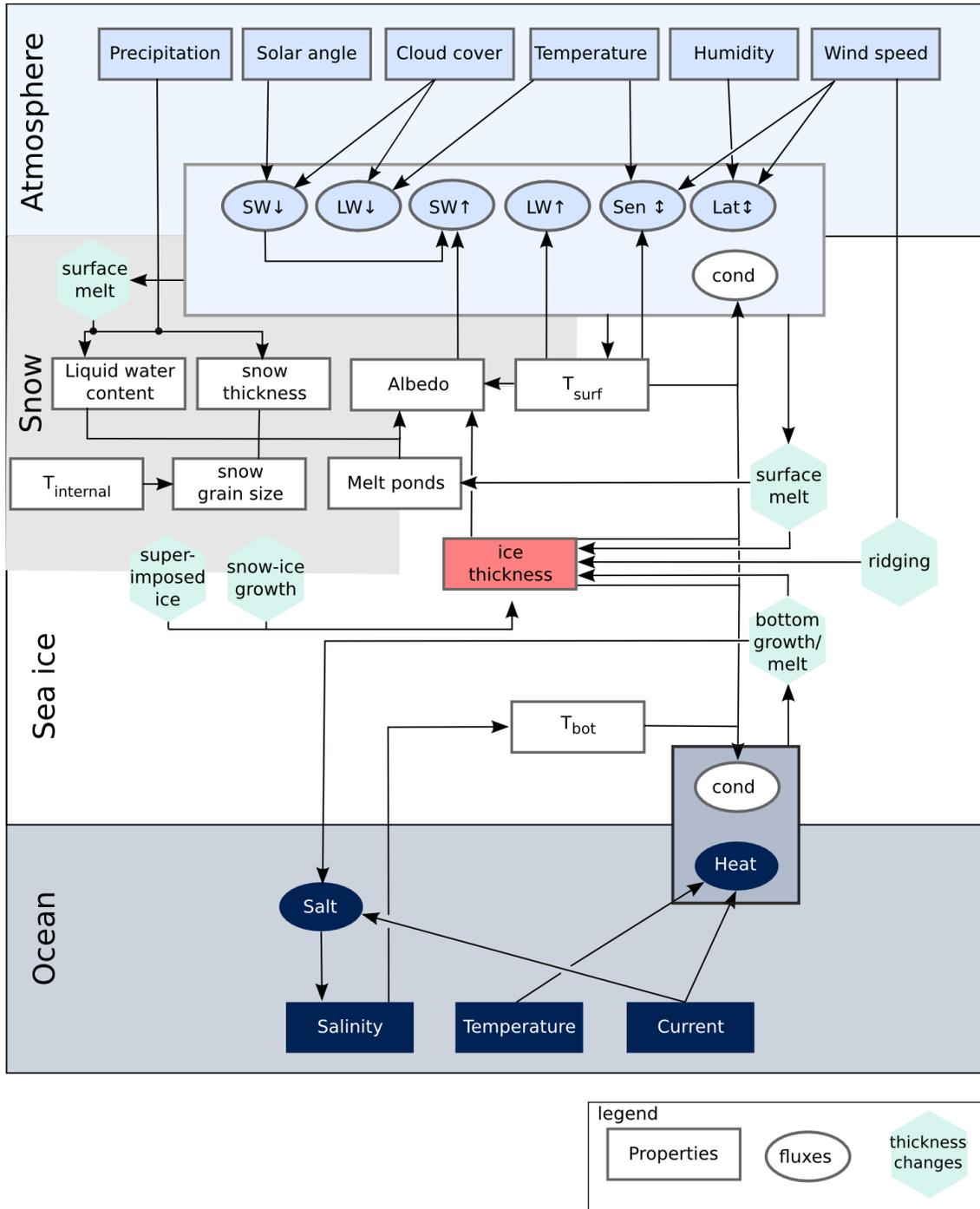
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3 Figure 6. Differences in the monthly maximum depth of open-ocean convection in ocean model  
4 experiments with and without polar lows included in the atmospheric forcing, (a) the Greenland Sea,  
5 (b) the Norwegian Sea. Reproduced with permission from Condron and Renfrew (2013).



1

2 Figure 7. Schematic overview of some of the processes that influence and that are influenced by the  
 3 growth and melt of sea ice. Only the most important pathways of interaction are shown. SW is  
 4 shortwave radiation, LW is longwave radiation, Lat is latent heat flux, Sen is sensible heat flux, Cond  
 5 is conductive heat flux in the ice, Heat is oceanic heat flux, Salt is oceanic salt flux, T<sub>bot</sub> is ice bottom  
 6 temperature and T<sub>surf</sub> is surface temperature.

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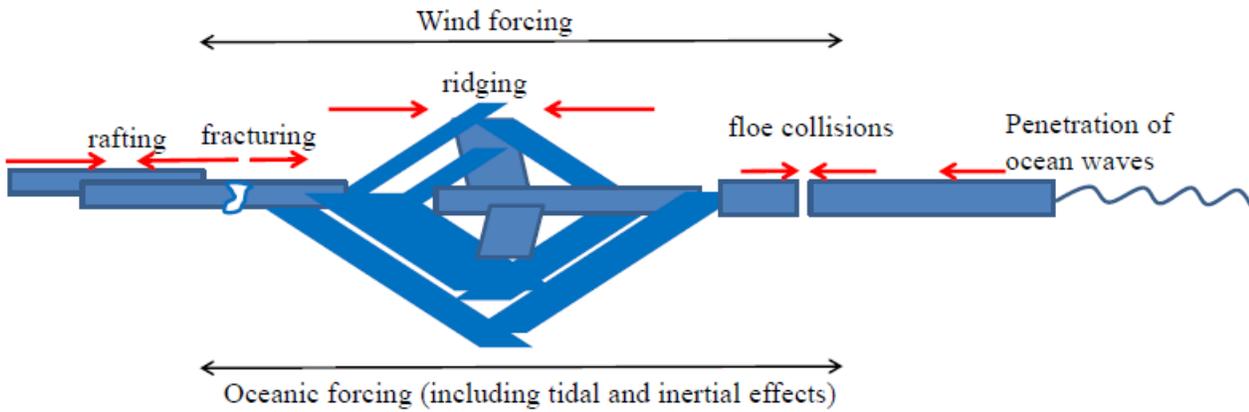


Figure 8. Processes generating mechanical waves travelling within sea ice.

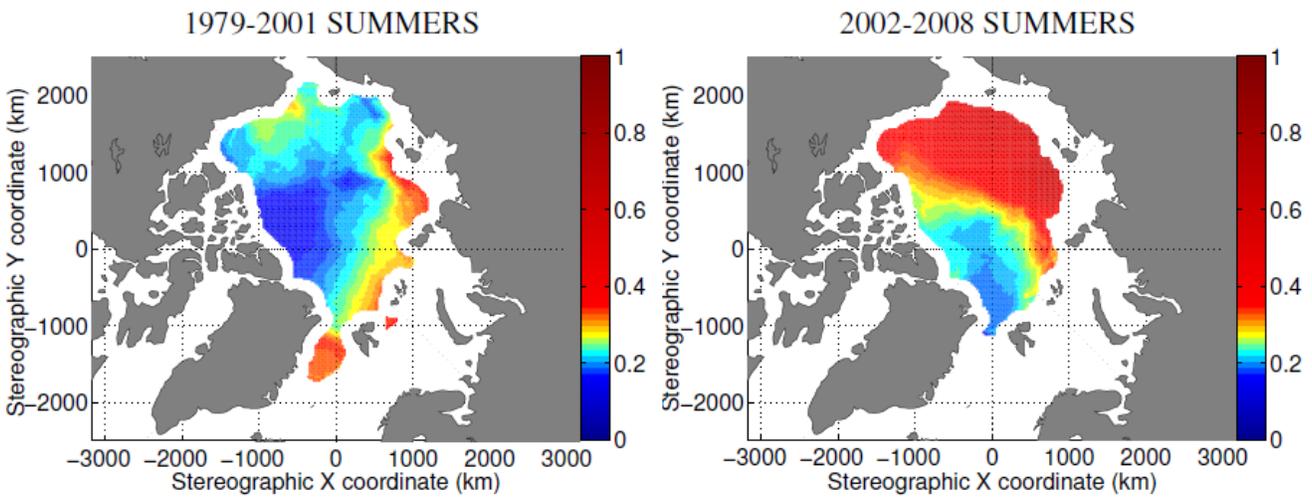
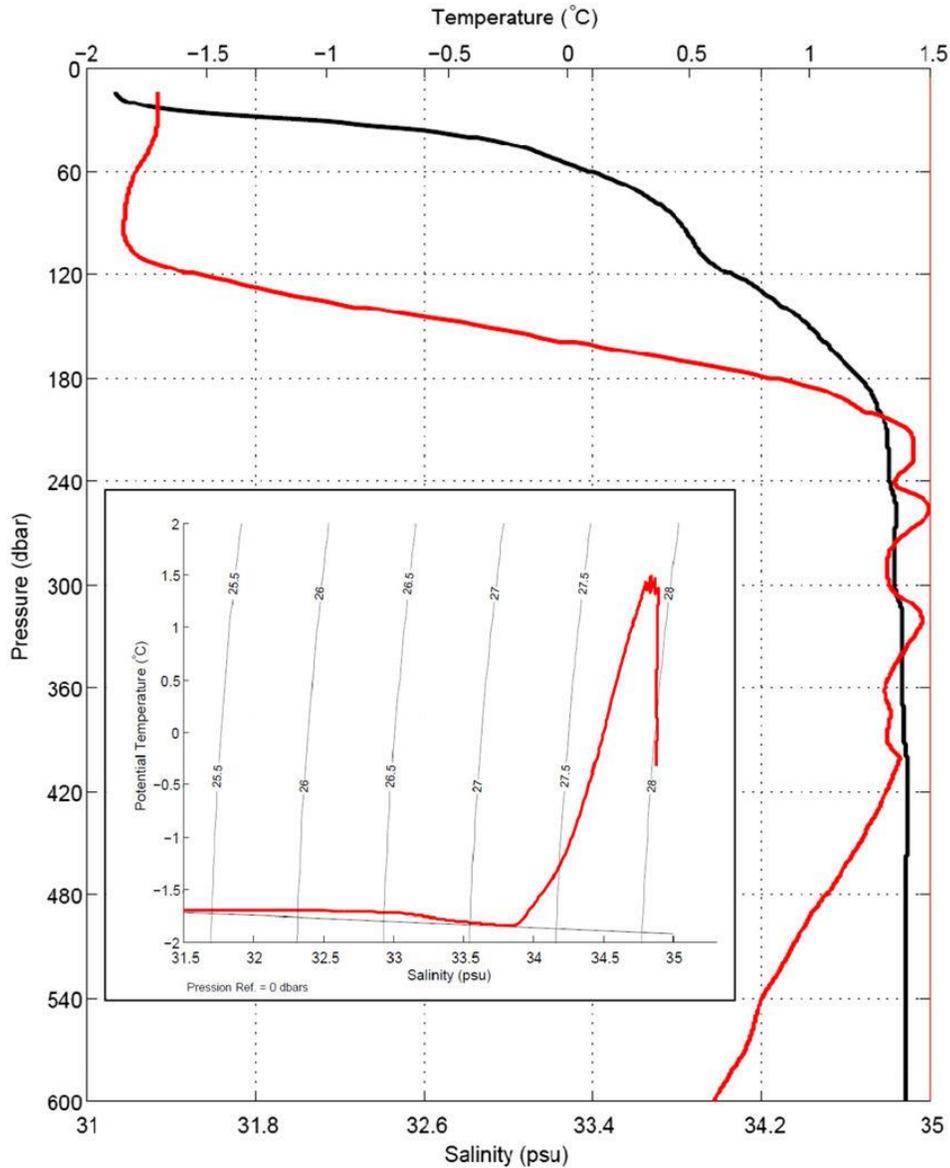


Figure 9. Spatial distribution of the magnitude of inertial oscillations in Arctic sea ice in summer in (a) 1979-2001 and (b) 2002-2008. The colour scale presents a non-dimensional parameter that Gimbert et al. (2012a) calculated to represent the magnitude of inertial oscillation. Reproduced with permission from Gimbert et al. (2012).



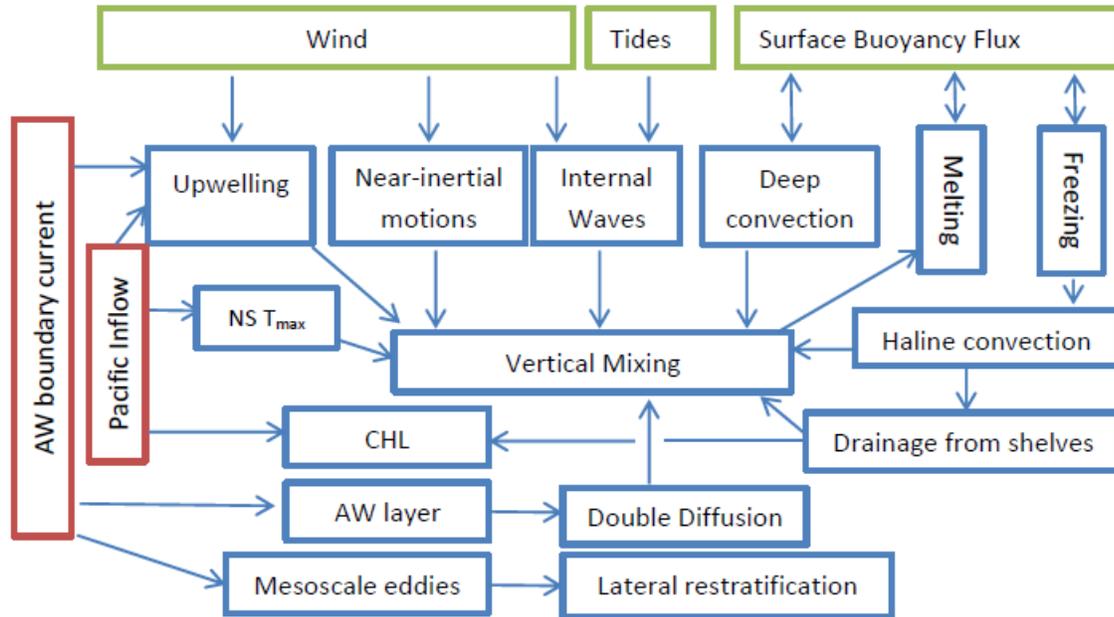
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2 Figure 10. Temperature (red) and salinity (black, in practical salinity units) profile and temperature-  
 3 salinity diagram (inner plot) of a sounding profile from 29 October, 2006, close to the Laptev Sea. The  
 4 isolines in the temperature-salinity diagram show the water density subtracted by  $1000 \text{ kg m}^{-3}$ .  
 5 Reproduced with permission from Bourgain and Gascard (2011).

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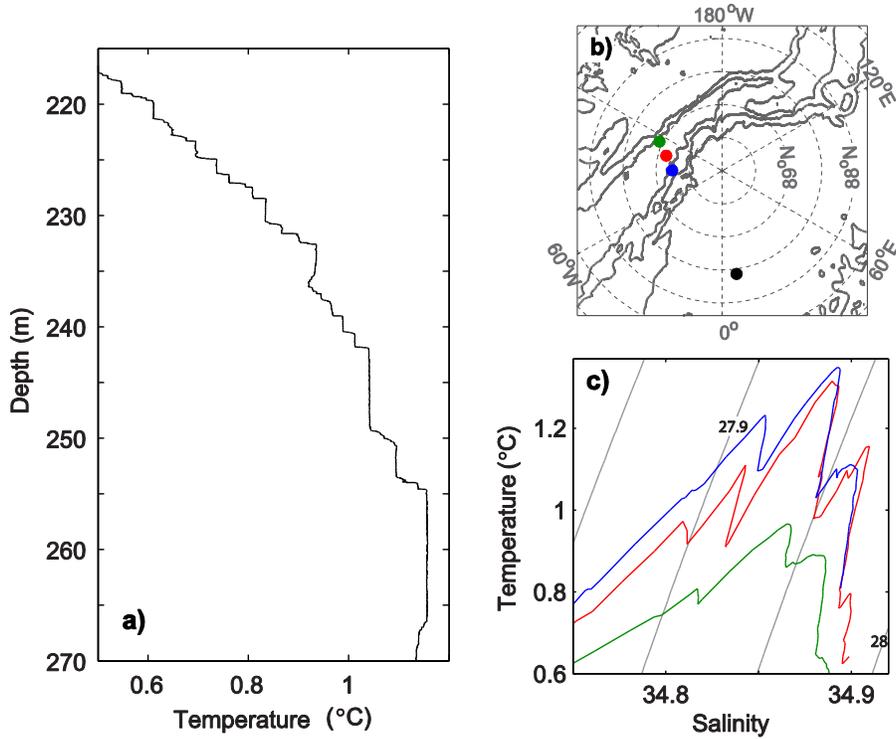
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Figure 11. Main forcing (green), oceanic heat input (red), physical processes (blue), and their relations (arrows) in the Arctic Ocean. CHL is the Cold Halocline Layer; AW is the Atlantic Water; and NS  $T_{max}$  is the near-surface temperature maximum.



1

2 Figure 12. a) An example temperature profile, collected using a microstructure profiler, showing the  
 3 staircase structure in the Amundsen Basin in the Arctic Ocean (station shown by black bullet in b). b)  
 4 Map showing the isobaths (1000 m contour interval) of the Lomonosov Ridge and station locations of a  
 5 and c in corresponding colors. c) Temperature and salinity diagram from three profiles across the  
 6 Lomonosov Ridge. The grey isolines show the water potential density subtracted by  $1000 \text{ kg m}^{-3}$ .