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ANTACTIC ICE SHELF MELTING AND ITS IMPACT ON THE GLOBAL SEA ICE-OCEAN SYSTEM

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ACADEMIC DISSERTATION

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Antarctic ice shelf melting and its impact on the global sea ice-ocean system

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abstract

Earth’s ice shelves are mainly located in Antarctica. They cover about 44% of the Antarctic coastline and are a salient feature of the continent. Antarctic ice shelf melting (AISM) removes heat from and inputs freshwater into the adjacent Southern Ocean. Although playing an important role in the global climate, AISM is one of the most important components currently absent in the IPCC climate model.

In this study, AISM is introduced into a global sea ice-ocean climate model ORCA2-LIM, following the approach of Beckmann and Goosse (2003; BG03) for the thermodynamic interaction between the ice shelf and ocean. This forms the model ORCA2-LIM-ISP (ISP: ice shelf parameterization), in which not only all the major Antarctic ice shelves but also a number of minor ice shelves are included. Using these two models, ORCA2-LIM and ORCA2-LIM-ISP, the impact of addition of AISM and increasing AISM have been investigated.

Using the ORCA2-LIM model, numerical experiments are performed to investigate the sensitivity of the polar sea ice cover and the Antarctic Circumpolar Current (ACC) transport through Drake Passage (DP) to the variations of three sea ice parameters, namely the thickness of newly formed ice in leads ($h_0$), the compressive strength of ice ($P^\star$), and the turning angle in the oceanic boundary layer beneath sea ice ($\theta$). It is found that the magnitudes of $h_0$ and $P^\star$ have little impact on the seasonal sea ice extent, but lead to large changes in the seasonal sea ice volume. The variation in turning angle has little impact on the sea ice extent and volume in the Arctic but tends to reduce them in the Antarctica when ignored. The magnitude of $P^\star$ has the least impact on the DP transport, while the other two parameters have much larger influences.

Numerical results from ORCA2-LIM and ORCA2-LIM-ISP are analyzed to investigate how the inclusion of AISM affects the representation of the Southern Ocean hydrography. Comparisons with data from the World Ocean Circulation Experiment (WOCE) show that the addition of AISM significantly improves the simulated hydrography. It not only warms and freshens the originally too cold and too saline bottom water (AABW), but also warms and enriches the salinity of the originally too cold and too fresh warm deep water (WDW). Addition of AISM also improves the simulated stratification. The close agreement between the simulation with AISM and the observations suggests that the applied parameterization is an adequate way to include the effect of AISM in a global sea ice-ocean climate model.

We also investigate the models’ capability to represent the sea ice-ocean system in the North Atlantic Ocean and the Arctic regions. Our study shows both models (with and without AISM) can successfully reproduce the main features of the sea ice-ocean system. However, both tend to overestimate the ice flux through the Nares Strait, produce a lower temperature
and salinity in the Hudson Bay, Baffin Bay and Davis Strait, and miss the deep convection in the Labrador Sea. These deficiencies are mainly attributed to the artificial enlargement of the Nares Strait in the model.

In this study, the impact of increasing AISM on the global sea ice-ocean system is thoroughly investigated. This provides a first idea regarding changes induced by increasing AISM. It is shown that the impact of increasing AISM is global and most significant in the Southern Ocean. There, increasing AISM tends to freshen the surface water, to warm the intermediate and deep waters, and to freshen and warm the bottom water. In addition, increasing AISM also leads to changes in the mixed layer depths (MLD) in the deep convection sites in the Southern Ocean, deepening in the Antarctic continental shelf while shoaling in the ACC region. Furthermore, increasing AISM influences the current system in the Southern Ocean. It tends to weaken the ACC, and strengthen the Antarctic coastal current (ACoC) as well as the Weddell Gyre and the Ross Gyre.

In addition to the ocean system, increasing AISM also has a notable impact on the Antarctic sea ice cover. Due to the cooling of seawater, sea ice concentration and thickness generally become higher. In austral winter, noticeable increases in sea ice concentration mainly take place near the ice edge. In regards with sea ice thickness, large increases are mainly found along the coast of the Weddell Sea, the Bellingshausen and Amundsen Seas, and the Ross Sea. The overall thickening of sea ice leads to a larger volume of sea ice in Antarctica.

In the North Atlantic, increasing AISM leads to remarkable changes in temperature, salinity and density. The water generally becomes warmer, more saline and denser. The most significant warming occurs in the subsurface layer. In contrast, the maximum salinity increase is found at the surface. In addition, the MLD becomes larger along the Greenland-Scotland-Iceland ridge.

Global teleconnections due to AISM are studied. The AISM signal is transported with the surface current: the additional freshwater from AISM tends to enhance the northward spreading of the surface water. As a result, more warm and saline water is transported from the tropical region to the North Atlantic Ocean, resulting in warming and salt enrichment there. It would take about 30–40 years to establish a systematic noticeable change in temperature, salinity and MLD in the North Atlantic Ocean according to this study.

The changes in hydrography due to increasing AISM are compared with observations. Consistency suggests that increasing AISM is highly likely a major contributor to the recent observed changes in the Southern Ocean. In addition, the AISM might contribute to the salinity contrast between the North Atlantic and North Pacific, which is important for the global thermohaline circulation.

**Key words:** Antarctic Ice shelf melting (AISM), Southern Ocean, Global sea ice-ocean system, Hydrography, Ocean circulation, North Atlantic Ocean, Global teleconnection.
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Chapter 1 Introduction

1.1. Cryosphere

The cryosphere is the frozen portion of the Earth’s surface. It includes snow, sea ice, river ice and lake ice, ice caps, ice sheets, glaciers, ice shelves, frozen ground and permafrost. In terms of water mass and heat capacity, it is the second largest component of the global climate system.

1.1.1. Importance of the cryosphere

Cryosphere plays an important role in the global climate. Its interaction with the other components of the global climate system is shown in Figure 1.1. In general, the cryosphere is strongly influenced by the atmospheric temperature, precipitation, solar radiation and clouds, and in turn, affects on the atmospheric properties. Owing to its high surface reflectance (albedo) and latent heat associated with phase changes, it has strong impact on the energy and moisture exchange between the Earth’s surface (land or ocean) and the atmosphere. The cryosphere on land stores about 75% of the world’s freshwater, and thus plays an important role in the freshwater budget. Changes in the ice mass on land affect the global sea level, as well as the oceanic and atmospheric circulations (e.g., Driesschaert et al., 2007).

![Fig. 1.1. Numerous interactions between the cryosphere and other major components of the global climate system. List in the upper boxes indicate important state variables, while list in the lower boxes indicate important processes involved in interactions. Arrows indicate direct interactions. Adapted from G. Flato (Online publ: EOS Sciences implementation plan (1999), Chapter 6, Cryospheric Systems (http://eospso.gsfc.nasa.gov/sci_plan/chapters.html).](image)

1.1.2. Area and volume of the cryosphere

Table 1.1 shows the area, volume and sea level equivalent (SLE) of the cryospheric components. It is seen that snow has the largest areal extent, with a maximum of about 47 million km². Snow is essentially seasonal with maximum coverage in winter and nearly absence in summer in both hemispheres. Because land surfaces at high latitudes are much larger in the northern hemisphere than in the southern hemisphere, the majority of snow is
located in the northern hemisphere, with a maximum area of $46.5 \times 10^6$ km$^2$ in January and a minimum area of $3.9 \times 10^6$ km$^2$ in August (Robinson et al., 1993). Although the areal extent of snow is the largest of the cryosphere components, its volume is the smallest, with a mean of 0.002 million km$^3$.

**Table 1.1. Area, volume and sea level equivalent (SLE) of cryospheric components.**

<table>
<thead>
<tr>
<th>Component</th>
<th>Area ($10^6$ km$^2$)</th>
<th>Ice volume ($10^6$ km$^3$)</th>
<th>SLE (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Max</td>
<td>Mini</td>
<td>Max</td>
</tr>
<tr>
<td>Snow on land</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NH (late Jan.)</td>
<td>46.5</td>
<td>3.9 (late Aug.)</td>
<td>0.002</td>
</tr>
<tr>
<td>SH (late July)</td>
<td>0.85</td>
<td>0.07 (early May)</td>
<td></td>
</tr>
<tr>
<td>Sea ice</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NH (late Mar.)</td>
<td>14.0</td>
<td>6.0 (early Sep.)</td>
<td>0.05</td>
</tr>
<tr>
<td>SH (late Sep.)</td>
<td>15.0</td>
<td>2.0 (late Feb.)</td>
<td>0.02</td>
</tr>
<tr>
<td>Mountain glaciers &amp; small ice caps</td>
<td>0.68</td>
<td>0.18</td>
<td>0.5</td>
</tr>
<tr>
<td>Permafrost*</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>continuous</td>
<td>7.6</td>
<td>0.03</td>
<td>0.08</td>
</tr>
<tr>
<td>discontinuous</td>
<td>1.73</td>
<td>0.07</td>
<td>0.18</td>
</tr>
<tr>
<td>Ice sheets</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Greenland</td>
<td>1.7</td>
<td>3.0</td>
<td>7.6</td>
</tr>
<tr>
<td>Antarctica</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>East</td>
<td>9.9</td>
<td>25.9</td>
<td>64.8</td>
</tr>
<tr>
<td>West</td>
<td>2.3</td>
<td>3.4</td>
<td>8.5</td>
</tr>
<tr>
<td>Ice shelves</td>
<td>1.5</td>
<td>0.7</td>
<td>~0</td>
</tr>
<tr>
<td>Total</td>
<td>37.78–101.76 (mean: 69.77)</td>
<td>33.302–33.352 (mean: 33.327)</td>
<td>81.66</td>
</tr>
</tbody>
</table>

* excluding Antarctica; NH: northern hemisphere; SH: southern hemisphere; East Antarctica and West Antarctica roughly correspond to the eastern and western hemisphere relative to the Greenwich meridian.

The areal extent of sea ice is the second largest among the components of the cryosphere, with a mean maximum of 29 million km$^2$. Although the sea ice coverage in both hemispheres is almost the same in winter, there is a large difference in summer, with more ice in the northern hemisphere due to the presence of perennial sea ice in the Arctic Ocean. Similar to snow, the sea ice volume is also quite small, with a maximum of 0.07 million km$^3$ and a minimum of 0.022 million km$^3$. Snow and sea ice totally only account for around 1% of the world’s ice mass.

Ice sheets cover about 14 million km$^2$, being the third largest in areal extent among the cryospheric components. The areal extent is much larger in Antarctica (11.7 million km$^2$) than in Greenland (1.7 million km$^2$). The total volume of the ice sheets is about 32.3 million km$^3$, accounting for 96.7% of the total global ice, and thus ranks the largest of the cryospheric components in volume. The Antarctic ice sheet holds the majority of ice in the world today. This freshwater storage corresponds to 80 m of the world sea level equivalent (SLE), with East Antarctica accounting for 79% and West Antarctica for 11%, totally 90% of the global ice mass. The large ice volume and area in Antarctica make this continent an important player in the global climate.

Ice shelves, the seaward extension of ice sheets, cover about 1.5 million km$^2$ in area and 0.7 million km$^3$ in volume. Although they only account for about 2% of the world ice, both in area and in volume, ice shelves play an important role in the global climate system as the
interface between the ice sheets and the oceans. The melting process at the base of ice shelf provides freshwater and takes heat from the adjacent ocean water. Loss of ice shelf would decrease the frictional force on the upstream ice sheet, thus accelerating the downward movement of the ice sheet. In addition, melting may reduce the ice shelf thickness, which could foster the breakup of the ice shelf and iceberg calving.

1.2. Ice shelves

An ice shelf is a huge sheet of ice, connected to the ice sheet on land but extending out into the ocean. It mainly develops from glaciers flowing slowly downhill toward the ocean, but also can be created by formation of marine ice and composite ice on the ice shelf (Jeffries, 2002). An ice shelf may be hundreds of kilometers across and hundreds of kilometers long. In thickness, it ranges from about 100 to more than 1000 meters.

An ice shelf has a grounding line, an ice front and a cavity (Fig. 1.4). The grounding line is the boundary between floating ice shelf and the grounded ice that feeds it, while the ice front is the seaward floating edge of an ice shelf. An ice shelf has its thickest part at the grounding line and thinnest part at the ice front. Between them is the cavity of the ice shelf, with the base of the ice shelf above and the seafloor below. So far, the geometry of sub-ice shelf cavities and the seafloor topographies are the most unknown parts of the Antarctic ice shelves due to the difficulty in accessing to make measurement.

An ice shelf is generally in a quasi-steady state. Its grounding line and ice front may advance and retreat. For example, iceberg calving from an ice shelf can result in the retreat of an ice front, while the thinning of ice shelf can lead to the retreat of the grounding line.

![Fig. 1.2. Map of the Canadian ice shelves (MODIS image, July 22, 2008)](image)

1.2.1. Distribution of Ice shelves

Ice shelves are only found in Antarctica, Greenland and Canada. In Greenland, there are no large ice shelves but large glacier outlets, since no broad front anywhere lets the unconfined ice sheet reach the sea. Thus, most of the ice shelves are located in Canada and Antarctica, with the majority in Antarctica.
1.2.1.1. Canadian ice shelves

All Canadian ice shelves lie north of 82°N and are attached to the northern Ellesmere Island. A century ago, the Canadian ice shelves were not separate as today, but one continuous ice shelf, up to 70 m thick and 8,900 km$^2$ in size covering the Ellesmere Island. In the 19th century and the beginning of the 20th century, only a band of thick floating ice was observed along the northern margins of Ellesmere Island in the Canadian high Arctic, with a total length of 500 km. During the course of the 20th century, about 90% of the Ellesmere Ice Shelf disappeared. By 1990s, the remnant of the Ellesmere Ice Shelf, which had been continuously fringed along the Ellesmere Island coastline, broke into six: Serson, Peterson, Milne, Ayles, Ward Hunt and Markham ice shelves (Fig. 1.2). The largest of these was Ward Hunt Ice Shelf, covering about 400 km$^2$, and the second largest was Milne Ice Shelf. They were located approximately 800 km south of the North Pole.

Fig. 1.3. The main ice shelves in Antarctica (from U.S. Geology Survey)

In the 21st century, the Canadian ice shelves have experienced remarkable loss due to climate change. For example, in 2005, the entire Ayles Ice Shelf broke free because of anomalously warmer temperature and persistent offshore and along shore winds. The break-up was believed to be the largest break-up of its kinds in the past 30 years, forming a 66 km$^2$ giant ice island. In 2008, more frequent break-up was witnessed. Ice shelves disintegrated an area of 214 km$^2$ in total. Markham Ice Shelf completely broke away from the coast of the Ellesmere Island with a disintegration area of 50 km$^2$. After that, Ward Hunt Ice shelf and Serson Ice Shelf lost 42 km$^2$ and 122 km$^2$ of area, respectively. The break-up from Serson Ice Shelf was the largest ice shelf disintegration event in that year, representing about 60% of its previous area. The remaining ice shelves in Canada are four: Serson, Peterson, Milne and Ward Hunt. The total area is about 930.4 km$^2$.

1.2.1.2. Antarctic ice shelves

Ice shelves are a salient feature of Antarctica (Fig. 1.3). They cover about 44% of the coastline (Drewry, 1983) with a total area approximately 1.54×10$^6$ km$^2$. According to Wikipedia, there are roughly 42 ice shelves surrounding the Antarctic continent and floating
in the Southern Ocean. Clockwise from the eastern side of the Antarctica Peninsula, they are
Larsen, Ronne, Filchner, Brunt, Riiser-Larsen, Quar, Ekström, Jelbart, Fimbul, Lazarev, Hannan, Zubchatyy, Wyers, Edward VIII, Amery, Publication, West, Shackleton, Moscow, Voyeykov, Cook, Slava, Gilett, Nasen, McMurdo, Ross, Swinburne, Sulzberger, Nickerson, Getz, Dotson, Crosson, Cosgrove, Abbot, Venable, Stange, Bach, George, Wilkins, Wordie, Jones, Müller and Prince Gustav ice shelves. Some of other small ice shelves, such as
Nivilisen in Dronning Maud Land (Horwath et al., 2006), are not listed by Wikipedia. Due to
regional atmospheric warming over the Antarctic Peninsula in the second half of the 20th
century, some of the ice shelves along its western and eastern coasts disappeared. The ice
shelves of Prince Gustav, Müller, Jones, and Woride, along the western coast of Antarctic
Peninsula, disappeared in 1995, 1999, 2003 and 2009, respectively. The ice shelves Larsen A
and B, along the eastern coast of the Antarctic Peninsula, collapsed in 1995 and 2002,
respectively.
(1) Major ice shelves
There are totally 14 major ice shelves in Antarctica at present. They are Filchner-Ronne
(Filchner, Ronne), Larsen C, Brunt, Riiser-Larsen, Ekström, Fimbul, Amery, West, Shackleton, Ross, Getz, Abbot, George VI and Wilkins. All the ice shelves have an area more
than 15×10^3 km^2 (Table 1.2).

### Table 1.2. Area (×10^3 km^2) of Antarctica’s major ice shelves

<table>
<thead>
<tr>
<th>Name</th>
<th>Area</th>
<th>Name</th>
<th>Area</th>
<th>Name</th>
<th>Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Larsen C</td>
<td>61</td>
<td>Ekström</td>
<td>21</td>
<td>Ross</td>
<td>494</td>
</tr>
<tr>
<td>Filchner-Ronne</td>
<td>438</td>
<td>Fimbul</td>
<td>30</td>
<td>Getz</td>
<td>34</td>
</tr>
<tr>
<td>Filchner</td>
<td>45</td>
<td>Amery</td>
<td>65</td>
<td>Abbot</td>
<td>26</td>
</tr>
<tr>
<td>Riiser-Larsen</td>
<td>E-Weddel</td>
<td>26</td>
<td>West</td>
<td>17</td>
<td>George VI</td>
</tr>
<tr>
<td>Brunt</td>
<td>45</td>
<td>Shackleton</td>
<td>30</td>
<td>Wilkins</td>
<td>18</td>
</tr>
</tbody>
</table>

Ross, Filchner-Ronne and Amery are the three largest ice shelves embayed in Antarctica.
Ross Ice Shelf is located at the head of the Ross Sea. It is areally the largest ice shelf in
the world, covering an area of about 494,000 km^2. Its mean thickness is 330 m near the ice edge
and about 700 m at the grounding line. Filchner-Ronne Ice Shelf lies in the southern Weddell
Sea, a combination of Filchner Ice Shelf in the east and Ronne Ice Shelf in the west, partly
separated by Berkner Island open to the sea. The area of Filchner-Ronne Ice Shelf is about
438,000 km^2, being the second largest ice shelf in the world. On the other hand, it holds the
largest ice volume in the world, possessing an average thickness of 700 m, around 1800 m at
the grounding line. Amery Ice Shelf is the third largest ice shelf, covering an area of 65,000
km^2, tiny compared with Ross and Filchner-Ronne Ice Shelves. Although small, it is the
largest ice shelf in East Antarctica, draining about 16% of the grounded East Antarctic ice
sheet through Lamber Glacier and other tributary glaciers (Allison, 1979). Amery Ice Shelf
has a thickness 300 m at the centre of the calving front, and 2500 m at the grounding line
(Roberts et al., 2007). These three major ice shelves are situated over broad continental shelf
and relatively far away from the continental break. They belong to “Type 1” ice shelves as
identified by Beckmann and Goosse (2003; hereafter referred to BG03).

The other 11 major ice shelves are relatively small, usually less than 65,000 km^2. They
belong to “Type 2” ice shelves according to BG03 and are exposed to warm deep water. They
are generally located over a narrow continental shelf and relatively close to the continental
break. Fimbul Ice Shelf is located in the eastern Weddell Sea, bordering the coast of Queen
Maud Land from 3°W to 3°E. Due to the quite narrow continental shelf, this ice shelf
occasionally overhangs the continental slope. It has a thickness between 160 m and 550 m,
with the ice front between 160 m and 250 m (Nøst, 2004). The base of the ice shelf is
relatively rough (Nicholls et al., 2006). The cavity beneath the ice shelf has a thickness up to 900 m in the central part and more than 2000 m near the ice shelf edge, with a series of sills connecting with the Weddell Sea (Nøst, 2004). The main sill, underlying the ice tongue, is at 1°W with a depth between 500 m and 600 m, while the eastern sill is at 3°E. Through these sills, the cavity is episodically flushed with relatively warm waters (Nicholls et al., 2006).

George VI Ice Shelf is located on the western side of the Antarctica Peninsula. It occupies the George VI Sound and is bordered by the Alexander Island and the Palmer Land with two openings at the end, one to the north and another to the south (Fig. 1.3). It is an extensive ice shelf covering an area of about 29,000 km². Brunt and Riiser-Larsen Ice Shelves are the major ice shelves fringing on the eastern coast of the Weddell Sea. They are usually together called Eastern Weddell Sea (E-Weddell) Ice Shelves (e.g., Beckmann et al., 1999; BG03), covering a total area of 71,000 km². They play an important role in the water mass preconditioning and formation in the Weddell Sea (Thoma et al., 2006).

| Table 1.3. Name, length and width of the minor ice shelves (unit: km) |
|----------------|----------------|----------------|----------------|----------------|----------------|
| Name           | Length*        | width*         | Name           | Length*        | width*         |
| Quar           | 40**           |                | Gillett        | narrow         |                |
| Jelbart        | 64             |                | Nansen         | 48             | 16             |
| Lazarev        | 90             |                | McMurdo        | Portion of RIS |                |
| Hannan         | 29             |                | Swinburne      | 32             | 8              |
| Zubchatyoy     | small          |                | Sulzberger     | 137            | 80             |
| Wyers          | small          |                | Nickerson      | 56             |                |
| Edward VIII    | small          |                | Dotson         | 48             |                |
| Publication    | 60             |                | Crosson        | 56             |                |
| Moscow         | narrow         | 193            | Cosgrove       | 56             | 40             |
| University     |                |                |                |                |                |
| Voyeykov       | 90**           |                | Venable        | 60             | 24             |
| Cook           | 89             |                | Stange         | 20**           | 70**           |
| Slava          |                |                |                |                |                |

*: adopted from Wikipedia;
**: estimated from the images of Antarctic ice shelves (http://nsidc.org/data/iceshelves_images).

(2) Minor ice shelves

As shown in Table 1.3, the minor ice shelves are those with length and width usually less than 90 km. The relatively large minor ice shelves include Sulzberger, Lazarev, Voyeykov and Cook (Fig. 1.3), with a width of about 90 km. Lazarev Ice Shelf is located at 69°37’S, 14°45’E, east of Fimbul Ice Shelf (Fig. 1.3). Between Lazarev and Amery Ice Shelves there are a series of very small ice shelves, including Hannan, Zubchatyoy, Wyers, and Edward VIII. Voyeykov and Cook Ice Shelves are located between 90–180°E. Between Cook and Ross Ice Shelves these is also a series of quite small ice shelves, including Slava, Gilett, Nasen, and McMurdo. The ice shelves of Sulzberger, Nickerson, Dotson, Crosson, Cosgrove and Stange are located in West Antarctica (Fig. 1.3). Compared with major ice shelves, much less information is available for these minor ice shelves.

1.2.2. Importance of ice shelves

Being the interface between the ice sheet and the ocean, ice shelves play an important role in the evolution of the ice sheet and the oceanic water masses. As the seaward extension of the ice sheet, they are important for the ice mass balance. Ice shelves, even the smaller ones, can buttress inland ice flow (e.g., Weertman, 1974; Dupont and Alley, 2005, 2006), and dam
up the ice moving off the continent and eventually into the ocean. Thus ice shelf thinning or ice loss from the ice shelf would accelerate the flow of the ice sheet and the discharge of the ice into the ocean. This has been observed in Greenland and Antarctica. A close relationship exists between the ice speed and the ice thickness based on the satellite observations of the Greenland’s Jakobshavn Isbræ Glacier (Joughin et al., 2004). The Drygalski Glacier in Antarctica accelerated threefold following the collapse of Larsen A ice shelf in 1995 (Rott et al., 2002). Antarctic Peninsula glaciers speeded up significantly after the break-up of Larsen B ice shelf in 2002 (Rignot et al., 2004; Scambos et al., 2004).

Shrinkage of ice shelves can potentially contribute to the sea level rise, even in the case of smaller ones, due to their supporting effect on the inland ice (Dupont and Alley, 2006). The West Antarctic Ice Sheet has long been considered to be potentially unstable as “marine ice sheet”, due to resting on a “bowl” like bed mostly below sea level (Bentley, 1964). If ice shelves fringing in the western Antarctica were removed, the West Antarctic Ice Sheet would collapse more quickly. The complete collapse would result in global sea level rise by 5 to 6 m (e.g., Mercer, 1978), or by 3.3 m as recently estimated by Bamber et al. (2009). It would be a major disaster for the people living near the coast.

Intimately contacting with the ocean, ice shelves interact with the ocean and play a significant role in the water mass formation and freshwater budget as described below.

![Image](image_url)

**Fig. 1.4.** Schematic map of the interaction processes between the ice shelf and ocean.

### 1.3. Interaction between the ice shelf and ocean

The interaction between the ice shelf and ocean is quite complicated. Figure 1.4 roughly illustrates our present general understanding of iceberg calving, ice shelf basal melting, and their consequences.

#### 1.3.1. Iceberg calving and its importance

Iceberg calving is a form of ice ablation or ice disruption. A chunk of ice is suddenly released or broken away from the seaward front of an ice shelf, being one of the primary mechanisms of mass loss from the ice shelf. Although it is possibly caused by a tidal or
seismic event, iceberg calving is considered to be a normal geological process, due to the tendency of the ice to spread out at the terminus of ice sheets and glaciers. After being calved, an iceberg drifts with ocean currents, mainly under the force of water drag and water advection (Bigg et al., 1997). It is entrained in the coastal current around Antarctica with the help of the Coriolis acceleration (Gladstone et al., 2001).

Iceberg is a major freshwater source and plays a significant role in the freshwater balance in the Southern Ocean. Compared with the ice shelf basal melting, meltwater from an iceberg is a non-negligible term in the freshwater balance. It is estimated to be 75.21 mSv south of 55°S (Schodlok et al., 2006) and 50.7 mSv south of 63°S (Silva et al., 2006), much larger than the estimates of 25 mSv (Jacobs et al., 1996) and 28 mSv (Hellmer, 2004) for basal melting. In addition, iceberg meltwater flux has a large spatial variability. Thereby, it has special significance in some areas, such as the Scotia Sea, the western Weddell Sea, and the Prydz Bay (Silva et al, 2006), which faces Amery Ice shelf.

Iceberg calving and the subsequent motion of iceberg have impact on the adjacent ocean. They can modify the flow pattern and water mass distributions, and affect the sea ice coverage (Grosfeld et al., 2001; Dinniman et al., 2007), regardless whether floating or grounding on the seabed. In 1986, three giant icebergs separated from Filchner Ice Shelf and subsequently stranded on the shallow Berkner Bank. This calving event and grounding of icebergs caused long-term disturbance to the hydrographic conditions (Nøst and Østerhus, 1998). In 2000–2004, several large icebergs calved from Ross Ice Shelf. They moved through the Ross Sea and caused dramatic interannual variability in sea ice extent in the Ross Sea during that period (Arrigo and van Dijken, 2003).

1.3.2. Ice shelf and ocean interaction within the sub-ice shelf cavity

Due to the pressure dependence of the freezing point of seawater, both freezing and melting occur at the base of ice shelf and drive the thermohaline circulation within the sub-ice shelf cavity (Fig. 1.4). The physical process can be simply described as follows. The principal external oceanographic forcing is the production of high salinity shelf water (HSSW) during winter (Nicholls, 1996), when seawater freezes over the continental shelf. During the course of sea ice formation, salt is released into the ocean, thus increasing the seawater density and generating the HSSW. This dense, saline water sinks down to the continental shelf. Portion of it penetrates into the sub-ice shelf cavity under gravity. Because of the depression of the seawater freezing point with pressure (Millero, 1978), the HSSW flowing toward the grounding line becomes warm enough to melt the deep basal ice. Consequently, the dense HSSW evolves into very cold but relatively fresh ice shelf water (ISW; Jacobs et al., 1992). The ISW is relatively buoyant and ascends following the base of ice shelf. Melting continues as long as the ascending ISW entrains enough warm water from beneath to maintain it at a temperature higher than the freezing point. However, if its temperature is lower than the freezing point, the ascending ISW becomes supercooled and marine ice likely forms on the base of the ice shelf (Robin, 1979). Marine ice formation has been found at the base of Larsen Ice Shelf (Holland et al., 2009), Filchner Ice Shelf (Grosfeld et al, 1998), and Amery Ice Shelf (Morgan, 1972; Fricker et al., 2001). The combination of entrainment of HSSW and deposition of marine ice causes an increase in the ISW density. With the help of local topography, some of the ISW can recirculate toward the grounding line, creating an internal circulation cell driven by the difference in freezing point between the deep and shallow parts of the ice shelf base (Gerdes et al., 1999). This cycle of melting at deep and refreezing in shallower areas is called the “ice pump” (Lewis and Perkin, 1983). Whenever its density matches the ambient stratified water, the ISW detaches from the base of ice shelf and leaves the cavity. It flows downwards over the continental shelf edge, and finally contributes to the formation of deep and bottom water masses.
The basal melting described above is the first mode of ice shelf melting (ISM) as identified by Jacobs et al. (1992). In the second mode, melt takes place when warm water at intermediate depths offshore enters the cavity as part of the general circulation. The circumpolar deep water (CDW), which is about 3°C warmer than the in-situ melting point beneath George VI Ice Shelf, is one such heat source in the Southern Ocean. It is also the major source of shelf thinning in the west Antarctica (Shepherd et al., 2004). The third mode is near the ice front, associated with the seasonally warmer upper ocean waters just north of the ice front in summer, which can be advected into the cavity by tidal currents and other mechanisms. Although basal melting is mostly concentrated near the grounding line as suggested by observations (Rignot and Jacobs, 2002) and models (Payne et al., 2007; Walker and Holland, 2007), it may also dominant near the ice front due to tidal actions (Joughin and Padman, 2003).

1.3.3. Importance of basal melting

Although both freezing and melting occur at the base of ice shelf, basal melting is more important. It removes heat from and injects freshwater into the adjacent ocean, i.e. cools and freshens the seawater (Beckmann et al., 1999; Hellmer, 2004; Thoma et al., 2006; Wang and Beckmann, 2007). Due to its connection with the ice sheet and its modification of the ocean hydrography, basal melting plays an important role in the global climate system.

Basal melting has a potential impact on the stability of the ice sheet (Walker et al., 2008) and the sea level elevation. It causes thinning and a reduction of the ice shelf. Increased melting has been suggested to be the main reason of the thinning of Pine Island Glacier in West Antarctica and the collapse of parts of Larsen Ice Shelf in the Antarctica Peninsula (Shepherd et al., 2003, 2004). In addition, basal melting also results in retreat of the ice shelf grounding line (Walker et al., 2008). Since ice shelves have buttressing effect on the inland ice (Weertman, 1974; Dupont and Alley, 2005, 2006), their reduction or loss could lead to acceleration of tributary glaciers (Scambos et al., 2004). Thus, although the melting of ice shelves has little direct effect on the sea level rise since ice shelves are already afloat, it has the potential to significantly affect the sea level through the acceleration of ice sheet flow.

The freshwater from the basal melting plays an important role in the adjacent ocean. It is a significant contributor to the freshwater fluxes in the Weddell Sea (Timmermann et al., 2001) and the Southern Ocean (Jacobs et al., 1992). Unlike other freshwater generated at the surface, this freshwater is released below the surface, usually deeper than 200 m. It affects the stability of the near-surface stratification (Hellmer, 2004), prevents deep ocean convection (Beckmann et al., 1999), and thickens the sea ice cover (Hellmer, 2004; Wang and Beckmann, 2007).

Basal melting might also important to the global ocean. It contributes to the deep and bottom water formation (Beckmann et al., 1999; Jenkins and Holland, 2002; Hellmer, 2004; Rodehacke et al., 2007) and is important for the water mass precondition and formation (Thoma et al., 2006) in the Southern Ocean. Due to the connections of the Southern Ocean with three major oceanic basins, the signal of the Antarctic ice shelf melting (AISM) might not be confined to the Southern Ocean but likely reaches to the global ocean. Toggweiler and Samules (1995) suggested that up to 75% of the deep ocean water might retain the signature of the Antarctic ice shelf meltwater input. Wang and Beckmann (2007) have revealed significant changes in mixed layer depth induced by the AISM in the northern hemisphere.

The overall importance of the Antarctic ice shelves attracts more attention. In this thesis, our focus is also on the Antarctic ice shelves, especially on their basal melting effects on the global sea ice-ocean system. In the following, we will first describe the state-of-the-art of the research on the Antarctic basal melting and then outline the configuration of this thesis.
1.4. Basal melting of the Antarctic ice shelves

Direct measurement of basal melting is extremely difficult due to the darkness and the difficulties in accessing the sub-ice shelf cavities. Our knowledge of basal melting is therefore mainly from oceanographic observations, glaciological measurements, and numerical modeling.

1.4.1. Oceanographic observations

Oceanographic observations have been mainly made within the sub-ice shelf cavity and north of the ice shelf front. To measure the oceanographic conditions within a sub-ice shelf cavity, the classic methods are hot water drilling access holes (e.g., Makinson, 1994; Nicholls et al., 1997) and installation of under-ice moorings (e.g., Nicholls and Makinson, 1998). Both approaches have their own advantages and disadvantages in terms of the ease of deployment, recalibration, recovery of equipment, as well as the quantity of the oceanographic data obtained (Nicholls, 1996). Although the measurement can be made anywhere on the ice shelf provided the drill can penetrate the depth of the ice encountered, there are totally less than 20 access points made across all of the Antarctic ice shelves due to logistic costs and laborious work involved. More recently, complex environmental conditions of the cavity were measured by use of an autonomous underwater vehicle (Nicholls et al., 2006). Although only one return mission was conducted in the cavity, it made great progress to our understanding of the extraordinary environment of the sub-ice shelf cavity, and provides us another possibility to make direct measurement of the oceanographic conditions within the cavity.

North of the ice front, oceanographic conditions can reflect what happens within the sub-ice shelf cavity. They are easier to be measured compared with those within the sub-ice shelf cavity. Oceanographic moorings can be deployed during ship’s cruises; for example, during Polarstern’s 1995 cruises to the western Ronne Ice Front, two moorings were deployed and they provided the first long-term oceanographic records of the western Ronne Ice Front (Woodgate et al., 1998). On the other hand, deployment of oceanographic moorings could be hampered by the presence of sea ice during winter and threatened or destroyed by iceberg calving.

Assuming a steady state for the process within the sub-ice shelf cavity, basal melting can be estimated from oceanographic measurements within the cavity or near the ice front. Inflow can be calculated assuming geostrophic balance. Then melt can be estimated, provided knowledge the existing potential temperature difference between the inflow and outflow and the total melting capacity for one degree Celsius (e.g., Nicholls et al., 1997), or knowing the freshening of the inflow (e.g., Jacobs et al., 1992; 1996). The estimated melt rate from the two methods is the effective melt rate due to not considering the refreezing process at the base of ice shelf.

In addition to the estimate from oceanographic measurements, chemical tracers, such as helium and neon, are ideal for estimating the basal melt, due to their high concentrations in glacial meltwater compared to other environmental sources. Aboard the Polar Sea in 1994 and aboard the NBP00-01 cruise in 2001, CFC-11, CFC-12 and CFC-113 were measured along the front of Ross Ice Shelf and used to estimate the basal melt rate of Ross Ice Shelf (Loose et al., 2009).

Oceanographic measurements along the ice front or under the ice shelf are mostly taken in summer months, due to the harsh weather in Antarctic winter and due to variable sea ice conditions and the threat of iceberg calving along the ice front. Therefore the estimated melt rate mainly represents the seasonal mean over some years. In addition, when making the estimate, the general assumption was an inflow at the western side and an outflow at the eastern side of the ice shelf front. This flow pattern could be changed by the location of the maximum HSSW on the continental shelf due to the sea ice formation (Timmermann et al.,
Furthermore, the measurements of temperature and salinity of inflow and outflow at the ice shelf front might be affected by local mixing (Nicholls et al., 2003). Therefore, the simple assumption of the circulation dynamics and the estimate using oceanographic observations along the ice front may lead to overestimate or underestimate of the basal melt rate.

### 1.4.2. Glaciological observations

Basal melt also can be estimated from glaciological observations. The standard method requires the data of ice thickness, surface accumulation and ice flow velocity (Jenkins and Doake, 1991). Assuming an ice shelf in steady state, the horizontal divergence of the volume flux equals the combined surface and basal accumulations (Jenkins and Doake, 1991). Thus, basal melt can be estimated knowing the velocity, thickness and surface accumulation. Two methods are usually applied. The first one is a box model. The flux divergence is simply calculated on the ice shelf perimeter. Using this method, Jacobs et al. (1992) estimated the basal melt of Ronne Ice Shelf according to the measurements along the Rutford flowline. The second one is the flux divergence integrated over the ice shelf. For example, it has been used to estimate the spatial distribution of melting and freezing beneath Filchner-Ronne Ice Shelf (Joughin and Padman, 2003) and under the Pine Island Bay’s Ice Shelf (Payne et al., 2007).

When the ice shelf is experiencing thinning, the net melt rate should include both the steady state melting as described above and the ice shelf thinning rate (Shepherd et al., 2004).

There are uncertainties in the estimate of the melt rate from the glaciological measurements. The sampling interval is important (Payne et al., 2007). Sparse sampling may miss the highly localized melt peaks (e.g., Shepherd et al., 2004) and result in an underestimate (Payne et al., 2007). The unclear upstream boundary may lead to underestimation of the inflow to the ice shelf (e.g., Jenkins et al., 1997) and hence to underestimation of the overall mass loss by melting (Payne et al., 2007), or, in opposite, to overestimation. When warm water erodes ice shelves and leads to ice shelf thinning (Shepherd et al., 2004), the steady state assumption apparently is invalid and likely results in underestimation of the basal melting.

### 1.4.3. Numerical modeling

Due to the difficulties in accessing to make measurements, observations described above are very limited. They mainly cover Filchner-Ronne, Ross, Amery, Fimbul, Ekström, Pine Island, and George VI Ice Shelves. Most of the Antarctic ice shelves have not been measured. Therefore, numerical model is another essential tool to understand the ice shelf-ocean interaction process. Compared with the estimate made from measurements, numerical models can provide not only basal melting, but also freezing as well as thermohaline circulation in space and time. So far, 1- to 3-dimensional models have been employed as reviewed by Williams et al. (1998).

#### 1.4.3.1. Plume models

Oceanographic observations (Jacobs et al., 1979; Nicholls et al., 1991) and glaciological observations (Jenkins and Doake, 1991) show a two-layer profile of temperature and salinity in the water column under the ice shelves. These are the base for plume models. The plume model was first developed by MacAyeal (1985) by assuming the melt water behaving as a turbulent, buoyant plume ascending the ice shelf base. Then it was further developed by Jenkins (1991), Nicholls and Jenkins (1993) and Lane-Serff (1993). However, these models did not include the formation of frazil ice, which was observed to contribution to most of the basal accumulation beneath ice shelves (Robin, 1979; Engelhardt and Determann, 1987: and Nicholls et al., 1991). By including the growth and deposition of frazil ice crystals suspended within the plume, the plume model of Jenkins (1991) was upgraded by Jenkins and Bombosch.
(1995) using one crystal size, and further developed by Smedsrud and Jenkins (2004) with multiple size classes of crystals. All these models are one-dimensional, depth-integrated models, and the path taken by each plume must be chosen beforehand. Holland and Feltham (2006) developed a two-dimensional, depth-integrated model by incorporating the Coriolis force and the formation of frazil ice. With the influence of Coriolis force incorporated into the plume model for the first time, they found that the characteristic of real ice shelf water plumes can only be captured using models with both rotation and a realistic topography.

1.4.3.2. Two-dimensional models

Since one-dimensional plume models generally ignore the variations perpendicular to the ice shelf gradients and the effect of Coriolis force, two-dimensional models were developed using Boussinesq and hydrostatic approximations in the momentum balance. Two-dimensional thermohaline circulation can be described with a single equation for the stream function, which is produced from the two-dimensional flow field coupled with the continuity equation. The two-dimensional model was firstly developed by Hellmer and Olbers (1989), then upgraded by Hellmer and Olbers (1991) by permitting flow through a channel by altering the boundary conditions of the stream function. Applying both versions to Amery Ice Shelf, they found that changing the slope of the ice shelf base near the grounding line changed the regional patterns of melting and freezing, but had little impact on the overall circulation. On the other hand, changing the sea bed topography had a greater impact on the circulation pattern.

1.4.3.3. Three-dimensional models

(1) Individual ice shelf

Based on the work of Bryan (1969) and Cox (1984), Determann and Gerdes (1994) developed the first three-dimensional model for the sub-ice shelf circulation for an idealized ice shelf-ocean configuration. Then this model was applied to Filchner-Ronne Ice Shelf (Determann et al., 1994; Gerdes et al., 1999), Amery Ice Shelf (Williams et al., 2001), Ekström Ice Shelf (Nicolaus and Grosfeld, 2004), and to an idealized ice shelf cavity geometry coupled with open ocean at the topographic ice shelf barrier (Grosfeld et al., 1997). Under the idealized ice shelf cavity, Determann and Gerdes (1994) and Grosfeld et al. (1997) found pronounced sensitivity of the ice shelf-ocean interaction to the ice shelf and bottom topographies. For real ice shelves, Determann et al. (1994) and Gerdes et al. (1999) derived typical circulation patterns within the sub-ice shelf cavity of Filchner-Ronne Ice Shelf. Williams et al. (2001) demonstrated that the circulation within the cavity was generally steered by the cavity topography and driven by the density gradient in the cavity, which was strongly influenced by the heat and salt fluxes at the ice-ocean interface and across the open ocean boundary. Nicolaus and Grosfeld (2004) indicated the importance of precise and high-resolution geometries in numerical models, especially in key regions such as across the narrow continental shelf.

The model used above was constructed in σ coordinates. This has some advantages. E.g., the ice shelf topography is more easily resolved since the vertical levels follow the base of the ice. On the other hand, the approach has disadvantages. Little effort is taken to include the ice shelf processes, because all the ice shelf-ocean interactions are applied at the surface level, which now is the base of ice shelf. In addition, many grid points are needed to resolve baroclinic structures, and pressure gradient errors near steep topography may result (Mellor et al., 1994), for example, near the ice shelf edges where σ coordinates are “bent” from surface values to approximately 200 m depth (Losch, 2008).

As an alternative, isopycnic models are employed, which is appropriate for well-stratified, deep ocean environments.
Holland et al. (2003) used an isopycnic model, based on the Miami Isopycnic Coordinate Ocean Model (MICOM, Bleck 1998), to study the ocean circulation beneath Ross Ice Shelf, and reproduced many of the observed and expected features of the sub-ice shelf circulation. They suggested that the simulated lower net melting over the whole ice shelf base might be more realistic as additional forcings are added to the model.

All the models above indicate that sub-ice shelf circulation is strongly sensitive to the shape of cavity, and that the actual melting or freezing rates are determined by the slope. The circulation is controlled by the topographies of the ice shelf base and the sea-bed. The combination of the ice shelf base and the sea-bed determine the water column thickness, which appears paramount in determining the pattern of circulation (Williams et al., 1998).

(2) Multi-ice shelves

Ocean general circulation models, regional or global, are used to simulate the interaction between the ice shelf and ocean. In these models, usually more than one ice shelf is included.

Beckmann et al. (1999) was the first to include the shallow shelf areas as well as the sub-ice shelf cavities of the inner Weddell Sea and Ross Sea in a large-scale regional stand-alone ocean model BRIOS-1 (Bremerhaven Regional Ice-Ocean Simulations). Filchner-Ronne, Ross, Larsen, E-Weddell, and Fimbul Ice Shelves were included. They found that the near-surface layer became colder and fresher due to the sub-ice shelf forcing. The water modified in the sub-ice shelf cavities contributed significantly to the deep and bottom water formation along the continental slope and affected the water mass characteristics throughout the Weddell Sea, by increasing the stability of the near-surface stratification and preventing deep convection.

Coupling the stand-alone ocean model BRIOS-1 to a dynamic-thermodynamic sea ice model, Timmermann et al. (2002) developed ice-ocean model BRIOS-2 to simulate ice-ocean dynamics in the Weddell Sea. They included the same ice shelves as Beckmann et al. (1999). Their results demonstrated that the sub-ice shelf circulation under Filchner-Ronne Ice Shelf is governed by sea ice formation in the southwestern continental shelf. The circulation fluctuated between two modes, cyclonic and anti-cyclonic. Although hardly affecting the area-averaged basal melt rates, it influenced the spatial distribution of freezing and basal melting.

Using the similar model as Timmermann et al. (2002), Hellmer (2004) studied the impact of freshwater originating from the ice shelf base. In addition to the ice shelves included by Beckmann et al. (1999), Hellmer (2004) also included Shackleton, Getz, Abbot and George VI Ice Shelves. He showed that if the freshwater from the caverns was absent, sea ice would be thinner, shelf waters would be warmer and saltier, and the Southern Ocean deep basins would be flushed by denser waters.

The models above are formulated in $\sigma$ coordinate because of being well suitable for studies of shelf dynamics and bottom boundary layer flows (Beckmann et al., 1999). However, many global ocean models have been constructed in z-coordinate to date. This leads to the work of Losch (2008).

Losch (2008) developed a new ice shelf cavity model for z-coordinate models and applied it to a nearly global ocean coarse resolution model with sea ice. Only Ross Ice Shelf in the Ross Sea and Filchner-Ronne Ice Shelf in the Weddell Sea were included. He showed that glacial meltwater from the Ross Sea could be traced as far as north as 15°S, while glacial meltwater from the Weddell Sea was confined to the ACC on a 100-year time scale. He again showed that the effects of ice shelf-ocean interaction ought to be included in ocean general circulation models as suggested by BG03.
1.4.3.4. Thermodynamic exchange at the ice shelf-ocean interface

The thermodynamic exchange at the ice shelf-ocean boundary associated with phase change has to be formulated in the numerical models described above. Various approaches were reviewed by Holland and Jenkins (1999). Most of the models treat the ice shelf as a fixed boundary and do not include the dynamics of ice shelf. With some prior assumptions about the ice shelf-ocean boundary, the thermodynamic exchange at the interface can be simply described by the equation of the freezing point of seawater only (e.g., MacAyeal, 1985; Jenkins and Doake, 1991); or by the equation of the freezing point of seawater together with the heat conservation law (e.g., Determann and Gerdes, 1994; Grosfeld et al., 1997). These two approaches are the so-called one-equation formulation or two-equation formulation by Holland and Jenkins (1999). The most sophisticated formulations contain three equations (Holland and Jenkins, 1999): the equation of the freezing point of seawater together with the heat and salt/freshwater conservation equations. They can be solved knowing the temperature of the model cells adjacent to the ice-water interface and the ice properties, without making any prior assumption about the ice shelf-ocean interface conditions.

There are a variety of treatments for the heat and salt conservation equations. The main differences are in the turbulence exchange coefficients for heat and salt, whether assumed to be constant (e.g., Hellmer and Obers, 1989, 1991; Hellmer and Jacobs, 1992; Jenkins et al., 2010) or functions of the friction velocity (e.g., Jenkins and Bombosch, 1995).

1.4.4. Necessity for parameterization of the ice shelf basal melting

Due to its important role in climate system, the effect of basal melting of ice shelves must be in some way included in global climate models (BG03; Losch, 2008). However, the progress has been very slow. In recent years, researchers (e.g., Beckmann et al., 1999; Timmermann et al., 2002; Hellmer, 2004; Thoma et al., 2006) have studied the local and regional impact of the ice shelf-ocean interaction in the Southern Ocean, mainly focusing on the Weddell Sea, through explicit inclusion of the three largest Antarctic ice shelves and part of the major ice shelves in regional oceanic general circulation models (OGCM).

Nevertheless, currently, there are no climate models including the sub-ice shelf cavities (see Griffies et al., 2000 for review). This is because inclusion of ice shelves would require substantial modification of the model code (e.g., Beckmann et al., 1999; Holland and Jenkins, 2001) and extension of the model domain far beyond 75° S (BG03). In addition, representation of the physical processes under ice shelves needs fine resolution, usually around 20 km, which most climate models obviously cannot fulfill. Modeling results show that the shape of cavity, the seabed topography and the water column thickness under an ice shelf control the sub-ice shelf circulation (e.g., Determann et al., 1994; Gerdes et al., 1999; Williams et al., 2001), and consequently the ocean-ice interaction at the base of ice shelf (Gerdes et al., 1999). However, our knowledge of the sub-ice shelf cavities is still very limited. For example, our traditional projection for the smooth ice shelf base is modified by the recent measurement conducted with an autonomous underwater vehicle under the Fimbul Ice Shelf (Nicolls et al., 2006). Although seismic reflection measurements provide information of ice thickness and seabed topography (Nøst, 2004; McMahon and Lackie, 2006), most of the geometric information under the Antarctic ice shelves still remain largely unknown so far. Thus, a major problem exists to realistically represent ice shelf cavities in models. In addition, explicit inclusion of the ice shelves would highly increase the computational time, which is obviously not suitable for the long-term integration in climate studies. Therefore, implicit inclusion of ice shelves is still a better choice.

Part of the ice shelf-ocean interaction has been implicitly included into climate models by nudging to surface salinity (e.g., DeMiranda et al., 1999) or by prescribing an additional freshwater flux on the continental shelf (e.g., Goosse and Fichefet, 2001). But both
approaches are not suitable for the study of climate variability, climate change or paleoclimate simulations since they have no temporal variations (BG03). In addition, the freshwater flux due to basal melting is actually released at the subsurface (at least deeper than 200 m), not at the surface, which could lead to different impact on the ocean as shown in this study. Thus, explicit inclusion of the ice shelf-ocean interaction is necessary for climate studies, even without knowing the details of sub-ice shelf conditions. The parameterization of BG03 for basal melting provides us such an opportunity, now described and introduced into the ORCA2-LIM model in this study. The details are presented in Chapter 4.

1.5. Thesis configuration

The thesis is organized as follows. In Chapter 2, we describe the model ORCA2-LIM used, together with the model initialization and forcing treatment. In Chapter 3, we investigate the model sensitivity, with the focus on the sea ice cover (sea ice extent, concentration and thickness) and the water transport through the Drake Passage (DP). In Chapter 4, we introduce the parameterization of BG03 for the ice shelf-ocean interaction into the model ORCA2-LIM (originally without ISM), which forms our model ORCA2-LIM-ISP (including ISM). Through comparisons with a series of observations, we validate the model ORCA2-LIM-ISP as well as ISM. In Chapter 5, we explore the impact of AISM on the Southern Ocean hydrography. We detail the hydrographic changes in the Southern Ocean due to the increasing AISM through comparing with the recent observations. In Chapter 6, we assess the impact of AISM on the Antarctic sea ice, with the focus on the seasonal and interannual variability of the Antarctic sea ice extent, concentration and thickness. In Chapter 7, we further investigate the impact of AISM on the sea ice-ocean system in the northern hemisphere, with focus on the North Atlantic Ocean and the Arctic Ocean. Then we discuss the mechanism for the inter-hemispheric teleconnection. Finally, in Chapter 8, we give discussion and conclusions as the end of this monograph.

Based on the results, one paper has been published before (Wang and Beckmann, 2007).
Chapter 2 Model Description

ORCA2-LIM is the model used in this study. It is a global sea ice-ocean coupled model with a mean horizontal resolution of 2°, developed in the framework of NEMO (Nucleus for European Modelling of the Ocean) for studying the components of the earth climate system ([http://www.nemo-ocean.eu](http://www.nemo-ocean.eu)). It is composed of two components: LIM (Louvain-la-Neuve sea ice model) and ORCA (Ocean PArallelise), where LIM is a dynamic-thermodynamic sea ice model and ORCA is a primitive equation ocean model. This chapter gives a detailed description of these two components and their coupling techniques, together with the forcing treatment and model initialization. The model has been validated by Timmermann et al. (2005) through process studies in high latitudes.

2.1. ORCA model

ORCA is the global configuration of OPA, a primitive equation OGCM developed at Laboratoire d’Océanographie DYnamique et de Climatologie (LODYC), for studying the ocean and its interactions with the other components of the earth climate system (Madec et al., 1999). Its uncoupled or coupled mode has been widely validated, for process studies in tropics (Raynaud et al., 2000; Guilyardi et al., 2001; Vialard et al., 2001; Lengaigne et al., 2002) and in high latitudes (Timmermann et al., 2005), for paleoclimate simulations (Braconnot et al., 1999), and for climate change scenarios (Barthelet et al., 1998; Friedlingstein et al., 2001). Its coupled code is used in the present study for investigating the evolution of the sea ice-ocean system in polar regions.

2.1.1. Governing equations

To solve the primitive equations, OPA assumes the Boussineq and hydrostatic approximations and incompressibility, and adopts the turbulent closure scheme to close the equations. Written for the horizontal velocity \( \mathbf{u} \), and the vertical velocity \( w \), \( \mathbf{U} = (u, w) \), the potential temperature \( T \), and salinity \( S \), the governing equation can be expressed as,

\[
\frac{\partial \mathbf{u}}{\partial t} = -(\zeta + f) \mathbf{k} \times \mathbf{u} - w \frac{\partial \mathbf{u}}{\partial z} - \frac{1}{2} \nabla_h (u^2) - \frac{1}{\rho_0} \nabla_h \mathbf{p} + D^U
\]

\[
\frac{\partial p}{\partial z} = -\rho g
\]

\[
\frac{\partial w}{\partial z} + \nabla_h \cdot \mathbf{u} = 0
\]

\[
\frac{\partial T}{\partial t} = -\nabla \cdot (T \mathbf{u}) - w \frac{\partial T}{\partial z} + D^T
\]

\[
\frac{\partial S}{\partial t} = -\nabla \cdot (S \mathbf{u}) - w \frac{\partial S}{\partial z} + D^S
\]

\[
\rho = \rho(T, S, \rho)
\]

Equations (2.1–2.6) are the horizontal momentum equation, the vertical momentum equation, the continuity equation, the conservation equations for heat and salt, and the equation of state, respectively. For notation, \( \zeta \) is the relative vorticity, \( \nabla_h \) is the horizontal gradients operators, \( \rho \) and \( \rho_0 \) are the in-situ and reference density of seawater, \( p \) is the local pressure, \( g \) is the gravitational acceleration, and \( t \) is the time; \( f \) is the Coriolis parameter defined as \( 2\Omega \sin \varphi \), where \( \Omega \) is the Earth’s rotation frequency and \( \varphi \) is the latitude; and \( D^U \), \( D^T \) and \( D^S \) account for the effects of small-scale processes, not explicitly represented by the model. They are parameterized as follows.

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2.1.2. Parameterization for small-scale processes

Considering the large difference of length scale in horizontal and vertical directions, the parameterizations of $D^U$, $D^T$, and $D^S$ are done separately.

2.1.2.1. Parameterization for lateral mixing

Mesoscale eddies typically have a scale of 50–100 km in the ocean. They play an important role in the climate change due to their influence on the circulation through: (i) transporting and mixing heat and salt; (ii) extracting potential energy from the mean flow; and (iii) exchanging momentum with the mean flow. Therefore their effects cannot be neglected in climate models. However, mesoscale eddies cannot be explicitly represented so far and usually parameterized in coarse and medium resolution climate models. These models are the so-called non-eddy resolving models.

(1) Lateral mixing of heat and salt

The lateral mixing of $D^{IT}$ or $D^{IS}$ can be approximated by a two-dimensional Laplacian operator,

$$D^{IT} = \nabla \cdot (A_T \mathcal{R} \nabla T)$$

(2.7)

where $A_T$ is the horizontal eddy diffusivity,

$$\mathcal{R} = \begin{pmatrix}
1 & 0 & -r_1 \\
0 & 1 & -r_2 \\
-r_1 & -r_2 & r_1^2 + r_2^2
\end{pmatrix}$$

(2.8)

Here $r_1$ and $r_2$ are the slopes between the surface along which the diffusive operator acts and the computational surface. In $z$-coordinate, if (i) $r_1=r_2=0$, the slopes are between the geopotential and computational surface. The lateral mixing is performed along geopotential surface; if (ii) $r_1 = \frac{\partial \rho}{\partial x} (\frac{\partial \rho}{\partial z})^{-1}$ and $r_2 = \frac{\partial \rho}{\partial y} (\frac{\partial \rho}{\partial z})^{-1}$, the slopes are between the isopycnal and geopotential surfaces. The lateral mixing is done along isopycnal surface.

The operator above, however, does not represent the effect of mesoscale-eddy extracting potential energy from the mean flow through baroclinic instability. This effect is parameterized as an additional eddy advection of tracers (Gent and McWilliams, 1990; Gent et al., 1995). By introducing an “eddy-induced velocity ($U^*$”)”, the lateral mixing of Eq. (2.7) becomes,

$$D^{IT} = \nabla \cdot (A_T \mathcal{R} \nabla T) + \nabla \cdot (U^* T)$$

(2.9)

where $U^*=(u^*, v^*, w^*)$ and

$$u^* = \frac{\partial}{\partial z} (A^{ev} r_1)$$

$$v^* = \frac{\partial}{\partial z} (A^{ev} r_2)$$

$$w^* = \left( \frac{\partial}{\partial x} (A^{ev} r_1) + \frac{\partial}{\partial y} (A^{ev} r_2) \right)$$

(2.10)

in which $A^{ev}$ is an eddy-induced velocity coefficient, $r_1$ and $r_2$ have the same meaning as in Eq. (2.8).

(2) Lateral mixing of momentum

The parameterization of momentum exchange between mesoscale eddies and the mean barotropic flow is still a challenge. Some studies show improvements in the simulations of boundary currents based on a statistical eddy field with bottom topography (e.g., Alvarez et al., 1994; Holloway et al., 1995; England and Holloway, 1998), based on the so-called eddy-
topography interaction parameterization (Griffies et al., 2000). However, no global climate model has employed such parameterization.

In the present model, the lateral mixing of momentum \( (\text{D}^u) \) is simply expressed as a Laplacian operator which separates the divergent and rotational parts of the flow.

\[ D^u = \nabla_h (A_u \chi) - \nabla_h \times (A_u \zeta) \quad (2.11) \]

where \( A_u \) is the lateral eddy viscosity, taken as a constant except in the tropics (2.5º–20º), \( \chi = \nabla \cdot \mathbf{u} \) is the divergence, and \( \zeta = \nabla \times \mathbf{u} \).

### 2.1.2.2. Parameterization for vertical mixing \( (\text{D}^v_u, \text{D}^v_T) \)

The vertical mixing is expressed as,

\[ D^v_u = \frac{\partial}{\partial z} \left( K_u \frac{\partial u}{\partial z} \right) \]

\[ D^v_T = \frac{\partial}{\partial z} \left( K_T \frac{\partial T}{\partial z} \right) \]

\[ D^v_s = \frac{\partial}{\partial z} \left( K_T \frac{\partial S}{\partial z} \right) \quad (2.12) \]

where \( K_u \) is the vertical diffusivity coefficient for momentum, and \( K_T \) is the vertical diffusivity coefficients for \( T \) and \( S \). Based on a 1.5-level turbulent closure model (Blanke and Delecluse, 1993), the vertical viscosity and diffusivity coefficients are written as,

\[ K_u = C_k l_k \sqrt{\bar{e}} \]

\[ K_T = K_u / P_r \]

where \( C_k \) is a constant (=0.1), \( P_r \) is the Prandtl number, which is a function of the local Richardson number \( (R_i) \), and \( \bar{e} \) is the turbulent kinetic energy. The mixing turbulent length scale \( l_k \) is denoted as,

\[ l_k = \sqrt{2 \bar{e} / N} \quad (2.13a) \]

Where \( N \) is the Brunt-Vaisälä frequency. The length scale is limited by,

\[ \left| \frac{\partial l_k}{\partial z} \right| \leq 1 \quad (2.13b) \]

to prevent the vertical variations of the length scale to be larger than the variations of depth. Thus the length scale is not only limited by the distance to the surface or to the ocean bottom, but also limited by the distance to a strongly stratified portion of the water column, such as, thermocline. When a portion of water column is unstable \( (N^2<0) \), the vertical eddy coefficients \( K_u \) and \( K_T \) are set to be large (up to 1 m\(^2\)/s). The turbulent kinetic energy \( \bar{e} \) evolves through vertical shear, its destruction through stratification, its vertical diffusion and its dissipation. It reads,

\[ \frac{\partial \bar{e}}{\partial t} = K_u \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 \right] - K_T N^2 + \frac{\partial}{\partial z} \left[ K_u \frac{\partial \bar{e}}{\partial z} \right] - c_e \frac{\bar{e}^{3/2}}{l_e} \quad (2.14) \]

where \( c_e \) is a constant, and \( l_e \) is the dissipation length scale, assumed equal to \( l_k \). More details can be found in Madec et al. (1999).

### 2.1.3. Barotropic dynamics and a non-linear free sea surface assumption

Barotropic assumption is usually used for the gravest mode in the vertical direction. The External Gravity Waves (EGWs) are barotropic because of the hydrostatic assumption. The speed of these waves is high with an order of 200 m/s. Taking a typical horizontal grid spacing of 100 km, explicit resolution of these waves would require a time step less than 10
minutes (Beckers and Deleersnijder, 1993). Therefore, such high speed strongly restricts the time step allowed in numerical models. To solve this problem, there are usually two methods used: one is the classical rigid-lid assumption; another is the free sea surface assumption. In the former, the EGWs are eliminated through considering the surface flat; while in the latter, the EGWs are kept and the surface is taken as undulated. The free surface assumption is used in the present study.

Representation of EGWs in the model is realized either by using a short time step or by filtering them. In climate models, the time step is usually of the order of hours. Such a time step does not resolve most part of the EGWs, neither does it resolve the wavelength of the EGWs smaller than several thousand kilometers. For the temporally unresolved EGWs, time splitting and implicit schemes are usually employed to damp the fast unresolved waves, which correspond to the explicit and implicit free surface methods, respectively (Griffies et al., 2000). In addition, Roullet and Madec (2000) introduced an additional force in the horizontal momentum equation and assuming a variable top cell thickness to filter the fast EGWs.

\[ \frac{\partial \mathbf{u}}{\partial t} = \mathbf{G} - g \mathbf{\nabla} \eta - g T_c \mathbf{\nabla} \frac{\partial \eta}{\partial t} \quad (2.15) \]

where \( \mathbf{G} \) represents the overall contributions from the Coriolis acceleration, hydrostatic pressure gradient, the nonlinear and viscous terms in Eq. (2.1). The last term in Eq. (2.15) represents the additional force characterized by a parameter \( T_c \). \( T_c \) larger than \( \Delta t \) (time step of the model) ensures the stability of the model and can damp all the temporally unresolved EGWs. Thereby, \( T_c = 2 \Delta t \) is adopted when numerical and physical modes join together (Roullet and Madec, 2000). As a result, the free sea surface (\( \eta \)) is now expressed as,

\[ \frac{\partial \eta}{\partial t} = -(\mathbb{Y} + P - E) \quad (2.16a) \]

in which

\[ \mathbb{Y} = \mathbb{V}[H + \eta \mathbb{W}] \quad (2.16b) \]

where the overbar denotes the vertical average over the whole water column, \( \mathbb{Y} \) is the divergence of the vertically integrated velocity, \( P \) and \( E \) are the precipitation and evaporation, and \( H \) is the ocean depth.

### 2.1.4. Boundary conditions

Through boundaries, ocean exchanges mass, momentum and energy with the other components of the Earth system. To represent the boundary conditions, we denote the sea surface height as \( \eta \), taking the mean sea level (\( z = 0 \)) as the reference height.

#### 2.1.4.1. Lateral boundary conditions

Assuming no flux across the lateral boundaries, the boundary conditions for \( T \) or \( S \) are,

\[ T = T_0 \]
\[ S = S_0 \quad (2.17) \]

For momentum, they are,

\[ \frac{\partial \mathbf{u}}{\partial n} = 0 \quad \text{(free-slip) or} \quad \mathbf{u} = 0 \quad \text{(no-slip)} \quad (2.18) \]

#### 2.1.4.2. Surface boundary conditions

At the sea surface, the kinematic boundary condition reads,

\[ w = \frac{\partial \eta}{\partial t} + \mathbf{u} \cdot \mathbf{\nabla}_h H + E \quad \text{if} \quad z = \eta \quad (2.19) \]

The dynamic condition is,
\[ \tau_s = \tau_{\text{wind}} + \tau_{\text{fresh}} \]  
(2.20)

in which the subscript \( s \) denotes surface. \( \tau_{\text{wind}} \) is the wind stress, and \( \tau_{\text{fresh}} \) is the momentum transfer in connection with the freshwater flux. The heat flux at the surface \((F_H)\) is,

\[ F_H = F_{sw} + F_{lw} + F_l + F_s \]  
(2.21)

where the subscriptions \( sw, lw, l \) and \( s \) denote the solar radiation, the longwave radiation, the latent heat and the sensible heat, respectively.

2.1.4.3. Bottom boundary conditions

The ocean sea floor forms the bottom boundary. Because of its solid state, there is no flow across this boundary. The fluxes of heat and mass are set zero as they are negligibly small.

The kinematic boundary condition is,

\[ w = -u \cdot \nabla_n H \quad \text{if} \quad z = -H \]  
(2.22)

and the dynamic boundary condition is,

\[ \tau_b = ru_b \]  
(2.23)

where \( r \) is a friction coefficient, and the subscript \( b \) denotes the bottom layer.

2.1.5. Numerical methods

Variables in the primitive equations are spatially arranged following Arakawa C grid (Fig. 2.1a): the velocity components \((u, v, w)\) are located at the center of different sides of one grid cell, and scalars, such as temperature, are located at the center of one grid cell as shown in Fig. 2.1a. Taking \( i, j, \) and \( k \) as the indexes in the horizontal and vertical directions, respectively, the partial derivative for a variable \( q \) is discretized as (such as in \( x \) direction),

\[ \frac{\partial q}{\partial i} = q_{i+1} - q_i \]  
(2.24a)

and the average of \( q \) is given by,

\[ \bar{q}' = \frac{q_{i+1} + q_i}{2} \]  
(2.24b)

Similar operators are used for indexation \( j \) and \( k \) (model level).

Finite differences are used to approximate the primitive equations. The diffusive and non-diffusive parts are solved using different schemes. For the non-diffusive parts, a three-time level “leap-frog” scheme is used, that is, for \( T \) and \( S \),
\[ T^{i+\Delta t} = T^{i-\Delta t} + 2\Delta t \text{ RHS}^{i} \]  
where RHS represents the non-diffusive part of the right hand side of a given equation, and \( \Delta t \) is the time step. For momentum,

\[ \mathbf{u}^{i+\Delta t} = \mathbf{u}^{i-\Delta t} + 2\Delta t \text{ RHS}^{i} + 2\Delta t \ g T \nabla E^{i} + 2\Delta t \ g T \nabla Y^{i+\Delta t} \]  
(2.26a)

in which, the last two terms are the introduced additional forces (Roullet and Madec, 2000). The last term is estimated from an elliptic equation below,

\[ Y^{i+\Delta t} = \left[ I - g T \Delta T^{i+\Delta t} \right] \cdot \nabla \left[ H + \eta^{i+\Delta t} \right] \mathbf{F} \]  
(2.26b)

where \( \mathbf{F} \) represents the first three terms on the right hand of Eq. (2.26a).

For the free sea surface height,

\[ \eta^{i+\Delta t} = \eta^{i-\Delta t} - 2\Delta t (Y^{i} + E^{i}) \]  
(2.27a)

where,

\[ Y^{i} = \nabla \left[ H + \eta^{i} \left| \mathbf{u}^{i} \right| \right] \]  
(2.27b)

which is diagnosed from \( \mathbf{u}^{i} \).

To avoid the divergence of two consecutive time steps, a Robert-Asselin time filter is applied on the horizontal velocity (Robert, 1966; Asselin, 1972):

\[ u_{f}^{i} = u^{i} + \gamma\left[ u^{i-\Delta t} - 2u^{i} + u^{i+\Delta t} \right] \]  
(2.28)

where the subscript \( f \) denotes the filtered values and \( \gamma \) is the Asselin coefficient.

Since the leap-frog scheme is not suitable for the representation of diffusion, an implicit time difference scheme is used for the diffusive term \( R \),

\[ \mathbf{u}^{i+\Delta t} = \mathbf{u}^{i-\Delta t} + 2\Delta t R^{i+\Delta t} \]  
(2.29)

### 2.2. LIM model

LIM is a dynamic-thermodynamic sea ice model designed specially for climate studies. Its thermodynamic part is based on Semtner’s three-layer model (1976), taking into account snow-ice formation and assuming brine pockets in the ice, and its dynamic part basically follows the viscous-plastic model of Hibler (1979). LIM has been used for sea ice studies in both polar regions through coupling with a one-dimensional upper ocean model (Fichefet and Morales Maquede, 1997), for investigation of the importance of sea ice-ocean interaction for the global ocean circulation (Goosse and Fichefet, 1999), and for process studies in high latitudes (Timmermann et al., 2005) through coupling with a global, free surface OGCM.

#### 2.2.1. Sea ice dynamics

The motion of sea ice can be described by a two-dimensional momentum equation. The advection term is usually neglected in the global climate studies according to the scaling arguments stated in the last part of this section.

\[
\left( \rho_h + \rho_s h_i \right) \frac{\partial \mathbf{u}_i}{\partial t} = -\left( \rho_h + \rho_s h_i \right) \mathbf{f} k \times \mathbf{u}_i + \tau_{ai} + \tau_{wi} - \left( \rho_h + \rho_s h_i \right) g \nabla \eta + \mathbf{F}
\]  
(2.30)

where the subscription \( i \) and \( s \) denote ice and snow, respectively, \( \rho \) is the density, \( h \) is the mean thickness, \( \left( \rho_h h_i + \rho_s h_s \right) \) is the ice and snow mass per unit area, \( \tau_{ai} \) and \( \tau_{wi} \) are the wind and water stresses, given by,

\[
\tau_{ai} = \rho_a C_u |\mathbf{u}_a| \mathbf{u}_a
\]

\[
\tau_{wi} = \rho_w C_w |\mathbf{u} - \mathbf{u}_i| \left[ (\mathbf{u} - \mathbf{u}_i) \cos \theta + k \times (\mathbf{u} - \mathbf{u}_i) \sin \theta \right]
\]  
(2.31)

where \( \mathbf{u}_a \) is the wind velocity, and \( \mathbf{u}_i \) is the ice velocity, \( \rho_a \) and \( \rho_w \) are the density for air and water, \( C_u \) and \( C_w \) are the drag coefficients for air and water, \( \theta \) is the turning angle in the ocean boundary layer beneath sea ice.
The internal force $F$ is given by,

$$F = \nabla \cdot \sigma$$

in which $\sigma$ is a two-dimensional internal ice stress tensor, depending on the relationship between stress and strain-rates. Hibler’s viscous-plastic rheology (1979) treats sea ice as linear viscous for very small strain rates and as plastic for large strain rates. This rheology has been widely used so far and is described as,

$$\sigma = 2\mu \dot{\varepsilon} + \left( (\zeta - \mu) T(\dot{\varepsilon}) - \frac{P}{2} \right) I$$

where $\mu$ and $\zeta$ are the shear and bulk viscosities, respectively, $T(\dot{\varepsilon})$ is the trace of the two-dimensional strain-rate tensor $\dot{\varepsilon}$, $P$ is the compressive strength of ice, and $I$ is the two-dimensional unity tensor.

The viscosities are given by,

$$\zeta = e_c^2 \mu = \frac{P}{2A}$$

in which $e_c$ is the ratio of the principal axes of the elliptical yield curve, and,

$$A = \max \left\{ \frac{\dot{\varepsilon}_0}{\dot{\varepsilon}} \left[ T^2(\dot{\varepsilon}) (1 + \frac{1}{e_c^2}) - D(\dot{\varepsilon}) \frac{4}{e_c^2} \right] \right\}^{1/2}$$

where $\dot{\varepsilon}_0$ is a parameter termed “creep limit”, and $D(\dot{\varepsilon})$ is the determinant of the strain rate tensor. The compressive strength of ice $P$ is given by,

$$P = P^* Ah e^{-C(1-A)}$$

where $P^*$ and $C$ are empirical constants, and $A$ is the ice concentration representing the grid cell area covered by ice. The formulation above illustrates that the ice strength is strongly dependent on the amount of ice. When the ice becomes thicker, sea ice is strengthened.

### Table 2.1. Scale analysis

<table>
<thead>
<tr>
<th>Terms</th>
<th>Scale</th>
<th>Order</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice advection</td>
<td>$\rho h U^2/L$</td>
<td>$10^{-4}$</td>
</tr>
<tr>
<td>Coriolis acceleration</td>
<td>$\rho h f U$</td>
<td>$10^{-2}$</td>
</tr>
<tr>
<td>Air stress</td>
<td>$\rho_a C_a U_a^2$</td>
<td>$10^{-1}$</td>
</tr>
<tr>
<td>Water stress</td>
<td>$\rho_w C_w (U_i - U)^2$</td>
<td>$10^{-1}$</td>
</tr>
<tr>
<td>Sea surface tilt</td>
<td>$\rho h g H/L (H \leq 10^{-1})$</td>
<td>$\leq 10^{-2}$</td>
</tr>
<tr>
<td>Internal friction</td>
<td>$\phi h L (P$ taking 10 kpa)</td>
<td>$10^{-1}$</td>
</tr>
</tbody>
</table>

The air and water stresses are independent on the study scale and the ice thickness, and the other terms are all dependent on the ice thickness. In addition, the ice advection, sea surface tilt and internal friction are also related to the study scale (Table 2.1). To make scale analysis for the ice momentum equation, the following typical scales are selected: sea ice thickness $h_i=1$ m, density $\rho_i=900$ kg/m$^3$, grid spacing $L=100$ km, wind speed $U_a=10$ m/s, current velocity $U=0.1$ m/s, and $f=10^{-4}$, water density $\rho_w=1000$ kg/m$^3$, air density $\rho_a=1.3$ kg/m$^3$. The air and water stresses are dominant forces together with the internal friction, in the order of $10^{-1}$. Coriolis acceleration and sea surface tilt are an order smaller ($10^{-2}$), and ice advection is even smaller ($10^{-4}$). Therefore, ice advection can be neglected.

The magnitudes of ice advection, Coriolis acceleration, sea surface tilt and internal friction are proportional to the ice thickness (Table 2.1). In the case of thin-ice cover (0.1 m), these terms drop down by one order of magnitude, while the air and water stresses keep unchanged. This results in a governing balance between the air and water stresses. In the case of thick-ice
cover (10 m), the Coriolis acceleration, sea surface tilt, internal friction and ice advection increase by one order of magnitude. As a result, ice advection is still negligible ($10^{-3}$). The internal friction is dominant in compact ice, while the Coriolis acceleration and sea surface tilt rank to the same order of magnitude as the air and water stresses. More discussion of the significance of the terms can be found in Leppäranta (2005).

### 2.2.2 Sea ice thermodynamics

Sea ice is thermodynamically considered to be a homogenous slab over which snow can accumulate. Because the temperature gradient is much smaller in the horizontal direction than in the vertical direction, the just vertical flow of heat is needed in the heat diffusion equation,

$$
\rho_c c_{pc} \frac{dT_c}{dt} = G(h_c)k_c \frac{\partial^2 T_c}{\partial z^2} + q_{st}
$$

(2.36)

where $\rho_c$, $c_{pc}$ and $k_c$ are the density, specific heat and thermal conductivity for snow and ice (including snow-ice), $T_c$ is the ice temperature, and $G(h_c)$ is a correction factor representing the effect of subgrid-scale snow and ice thickness distributions on the average heat conduction. In case of a uniform thickness distribution over the ice-covered part of the grid cell,

$$
G(h_c) = 1 + \frac{1}{2} \ln \left( \frac{2h_c}{e\alpha} \right)
$$

(2.37)

where $e$ is the base of the natural logarithms, $\alpha$ is a threshold thickness, $h_c$ is an effective thickness for heat conduction, defined as,

$$
h_c = \frac{k_c k_i \left( h_i + h_c \right)}{k_i + k_c \left( k_i + k_c \right)}
$$

(2.38)

when $h_c < e\alpha/2$, $G(h_c)=1$.

$q_{st}$ in Eq. (2.36) is an internal source term accounting for solar radiation penetrating into the ice when ice is free of snow. The penetrated solar radiation has an important effect on ice thickness (Maykut and Untersteiner, 1971). It is usually not for immediate surface melting, but for warming the subsurface interior of ice and initiating the internal melting (Semtner, 1976). The trapped brine pockets behave as a thermal reservoir to retard the heating or cooling of the ice (Maykut and Untersteiner, 1971). The effects of penetrated solar radiation and brine pockets in the ice are parameterized as (Fichefet and Morales Maqueda, 1997),

$$
\frac{\partial q_{st}}{\partial t} = i_0(1 - a_{su})F_{sw} \left[ 1 - e^{-5(h_i-h_c)} \right] - F_{lbp}
$$

(2.39)

where $a_{su}$ denotes the surface albedo, $F_{sw}$ is the solar radiation, $h_i$ is a constant equal to 10 cm, and $i_0$ is the fraction of net solar radiation penetrated into the snow or bare ice, and is parameterized as,

$$
i_0 = 0
$$

(2.40)

in case of snow-covered ice (Maykut and Untersteiner, 1971), and

$$
i_0 = 0.18(1 - c) + 0.35c
$$

(2.41)

in case of free of snow-covered ice (Ebert and Curry, 1993), where $c$ is cloud cover. $F_{lbp}$ is the heat flux used to prevent the upper ice layer from cooling below the melting point of the ice.

The variation of ice thickness due to thermal effects is determined by the heat balance at the top surface and at the base of the ice. At the ice base, the variation of ice thickness is controlled by,

$$
\rho_i L_i \left( \frac{\partial h_i}{\partial t} \right)_{acc-abl} = Q_{cb} - Q_w
$$

(2.42)

where $L_i$ is the volumetric latent heat of ice freezing, $Q_{cb}$ and $Q_w$ are the conductive heat flux within the ice and the oceanic heat flux, respectively.
At the top surface, the heat balance \( Q_{sw} \) is determined by,

\[
Q_{sw} = (1 - i_0)(1 - \alpha_s)F_{sw} + \varepsilon_{sw}(F_{lw} - \sigma T_{sw}^4) + F_s + F_i + F_c
\]  

(2.43)

where the subscripts \( su \) is the surface, \( sw, lw, l \) and \( s \) have the same meaning as in Eq. (2.21). Thus, the second term on the right hand is the net long-wave radiation (\( F_{lw}^{\text{net}} \)). \( \varepsilon \) is the emissivity, \( \sigma \) is the Stefan-Boltzmann constant, and \( F_c \) is the conductive heat flux from below the surface. The heat flux is assumed positive toward the surface. The energy imbalance reflects phase changes (freezing or melting) or warming or cooling (storage or release of heat) of the ice. When \( Q_{sw} > 0 \), the excessive heat is used for melting the top surface while keeping its temperature at the freezing point,

\[
\left( \frac{\partial h_i}{\partial t} \right)_{abl} = \frac{Q_{sw}}{L_s} h_s > 0
\]

(2.44)

where \( L_s \) is the volumetric latent heat diffusion of snow. The heat balance at the top and the base of sea ice is summarized as in Fig. 2.2.

Snow accumulated over the ice has two effects, as an insulator between ice and air and as a contributor to the ice growth through the snow-ice formation. When the mass of snow is heavy enough, the lower part of the snow layer submerges into the water. Slush develops and eventually forms snow-ice. Snow ice has remarkable contribution to the ice cover in the Baltic Sea (Leppäranta, 1983) and in the Southern Ocean (e.g., Massom et al., 2001 for review; Worby et al., 2008). Neglecting the density difference between snow-ice and sea ice, the variation of snow and ice thickness after snow-ice formation is given by,

\[
-(\Delta h_i)_{si} = (\Delta h_i)_{si} = \frac{\rho_i h_i - (\rho_0 - \rho_i) h_i}{\rho_s + (\rho_0 - \rho_i)}
\]

(2.45)

where the subscript \( si \) represents snow-ice.

The variation of \( A \) depends on the heat budget of the open water \( B_l \). It is parameterized as,
\[
\left( \frac{\partial A}{\partial t} \right)_{\text{acc-abl}} = \frac{A}{2h_i} \Gamma \left( \frac{\partial h_i}{\partial t} \right), \quad \text{if } B_i > 0
\]
\[
\left( \frac{\partial A}{\partial t} \right)_{\text{acc}} = -\Phi(A) \frac{(1-A)B_i}{L_i h_0}, \quad \text{if } B_i < 0
\]

where $\Gamma$ is the Heaviside unit function, and $\Phi(A)$ is a correction factor between 1 and 0, denoting as (Fichefet and Morales Maqueda, 1997),

\[
\Phi(A) = \left(1 - A^2\right)^{1/2}
\]

$h_0$ is the thickness of newly formed ice in the leads.

2.2.3. Sea ice thickness categories and transportation

There has been a variety of approaches to describe ice thickness distribution in sea ice models. The classical one is the two-level approach (Doronin, 1971), which includes open water (ice concentration) and mean ice thickness.

The conservation law for ice/snow is,

\[
\frac{\partial \left(A h_{(i,s)}\right)}{\partial t} = -\nabla \cdot (u_i A h_{(i,s)}) + S_{h_{(i,s)}}^{\text{th}} + D_i \nabla^2 (A h_{(i,s)}) + S_{h_{(i,s)}}^{\text{dy}}
\]

while the conservation law for ice concentration is,

\[
\frac{\partial A}{\partial t} = -\nabla \cdot (u_i A) + S_A^{\text{th}} + D_i \nabla^2 A + S_A^{\text{dy}}
\]

where $D_i$ is the horizontal diffusivity, used for numerical stability (Hibler, 1979; Fichefet and Morales Maqueda, 1997), $S$ represents the source term due to thermodynamic ($\text{th}$) and shear deformation ($\text{dy}$) processes. The dynamic source $S_{h_{(i,s)}}^{\text{dy}}$ in Eq. (2.48-2.49) is generally taken as zero in the standard two-level ice models, although the shear deformation contributes to the formation of open water and ice ridges.

2.2.4. Numerical methods and algorithms

Differing from the ocean model ORCA, Arakawa B grid is used for arrangement of the variables in space (Fig. 2.1b) in the sea ice model LIM, where the vectors are located at the corner of the grid cell and the scalars are situated at the center of the grid cell. Because of the development of numerical instabilities when snow or ice thickness becomes small (Fichefet and Morales Maqueda 1997), a fully implicit time stepping scheme is applied to solve the heat diffusion equation (2.36). In order to strictly satisfy the heat balance at the top surface of the snow-ice system, a Newton-Raphson iterative procedure is employed to solve Eq. (2.45). Basically similar to that of Hibler (1979), a semi-implicit predictor-corrector procedure is adopted to solve the ice momentum balance Eq. (2.30) with a simultaneous under-relaxation technique. While for the ice transportation equations (2.48)–(2.49), a forward time-stepping scheme is used. To conserve the second-order moment of the spatial distribution of the advected quantities within each grid cell, the Prather technique (1986) is adopted to solve the advection term. More details can be found in Fichefet and Morales Maqueda (1997).

2.3. Coupling of the sea ice and ocean models

The coupling between sea ice and ocean components is accomplished through fluxes of momentum, heat and salt/freshwater. Due to the discrimination of open water and ice, the fluxes ($FL$) have to be weighted with areal coverage of open water and ice, i.e.,

\[
FL = A \cdot FL_{wi} + (1 - A) \cdot FL_{aw}
\]

where the subscripts represent the interfaces between ice and ocean ($wi$), and between air and water ($aw$).
2.3.1. Momentum flux

The momentum flux is computed as shown in Eq. (2.31) using the usual quadratic bulk formula. The drag coefficients for water \( C_w \) is taken as a constant. For air, it is computed from an iteration procedure as the transfer coefficients below.

2.3.2. Heat flux

The turbulent heat fluxes between air and ocean are given by the bulk formulae,

\[
F_{aw,i} = \rho_a c_{pa} C_{ha} (T_a - T_{sa}) U_a
\]

\[
F_{aw,i} = \rho_a L_w C_{wa} (q_a - q_{wa}) U_a
\]

where \( c_{pa} \) is the specific heat of air, \( U_a \) and \( T_a \) are the air velocity and temperature at certain height close to the surface, \( L_w \) is the latent heat of evaporation or sublimation, \( q_a \) and \( q_{wa} \) are the specific humidities of the atmosphere and the ocean surfaces, respectively. The transfer coefficient \( C_{ha} \) and \( C_{le} \) are estimated based on the Monin-Obukhov similarity theory (Monin and Obukhov, 1954; Garratt, 1992).

The sensible heat flux between ice and ocean, following McPhee (1992), is expressed as,

\[
F_{si} = \rho_i c_{pi} C_{hi} u_i (T_{si} - T_f)
\]

where \( C_{hi} \) is the heat transfer coefficient, and the friction velocity \( u_i \) is computed from,

\[
u_i = \left( \frac{\tau_{vis}}{\rho_i} \right)^{1/2}
\]

2.3.3. Mass flux

Mass exchange between ice and ocean includes freshwater flux (\( F_f \)). Although the freshwater flux is rather weak on climatic time scales, its representation could influence the global salinity (Huang, 1993) and general circulation (Tartinville et al., 2001). There are two approaches to represent the freshwater flux at the ocean surface in numerical models. The classical one is the equivalent salt flux: salt is removed/introduced at the sea surface in order to take into account the dilution effect.

\[
F_f = 0
\]

\[
F_i = S_{ref} \left( \frac{\partial m_s}{\partial t} \right)_{abl} + (S_{ref} - S_i) \left( \frac{\partial m_i}{\partial t} \right)_{acc,abl} + (S_{ref} - S_i) \left( \frac{\partial m_s}{\partial t} + \frac{\partial m_i}{\partial t} \right)_{si} + S_i \left( \frac{\partial m_s}{\partial t} \right)_{si} + S_{ref} (E - P)
\]

where \( F_i \) is salt flux, \( S_{ref} (=34.7 \text{ psu}) \) is a reference sea surface salinity, and \( S_i (=6 \text{ psu}) \) is the ice salinity, \( m_s \) and \( m_i \) are the masses of snow and ice. The first term on the right-hand side of (Eq. 2.54) simulates the freshwater flux to the ocean due to snow melting (labeled \( abl \)). The second one is associated with ice formation/melting (labeled \( acc \) and \( abl \)). The third and forth ones are salt fluxes resulting from snow-ice formation (labeled \( si \)), and the last one represents the influence of evaporation, precipitation and runoff.

This approach has numerical shortcomings and possibly leads to a continuous increase in the global salinity (Huang, 1993). Tartinville et al. (2001) introduced a more physical approach to represent the surface freshwater flux in their ice-OGCM. In their model, sea ice acts as a negative reservoir of salt, and the brine rejection is considered as a salt-flux rather than a freshwater flux.
\[ F_f = P - E - \left( \frac{\partial m_{\text{abl,si}}}{\partial t} \right) \]
\[ F_s = (S_{\text{ref}} - S_r) \left( \frac{\partial m_{\text{abl,acc,si}}}{\partial t} \right) \]

This approach is employed in the present study.

2.4. Model configuration, initialization and forcing

The model domain extends from 78.19°S to 90°N with a zonal resolution of 2° and a meridional resolution of 2°×cos\( \phi \) (\( \phi \): latitude). The resolution is finer at high latitudes, about 65 km in Arctic and 50 km in Antarctic. To overcome the singularity at the North Pole, a tri-polar grid is used with the northern points of convergence situated on the North America and Asia Continent (Fig. 2.3). Local fine refinement in the Mediterranean Sea can be seen in Fig. 2.3, because of its important role in the water mass distribution in the North Atlantic and its potential impact on the World Ocean Circulation (Jia et al., 2007). In addition, grid refinements are also applied to some straits, i.e., the Gibraltar Strait, and the seas, i.e., the Red Sea and the Black Sea (Fig. 2.3). 31 vertical levels are used in the ocean model, with vertical grid spacing from 10 m in the top 100 m up to 500 m at the bottom (Table 2.2). 3 layers are used for the thermodynamic part of the sea ice model with 1 topmost layer for snow and 2 uniform layers for sea ice, and only a uniform layer is employed for the dynamical part of the sea ice model.

Bottom topography and coastline are computed from the data of Smith and Sandwell (1997), complemented by the ETOPO5 dataset. The coastline and bathymetry in Antarctic and Arctic are shown in Fig. 2.4. The Antarctic ice shelves covering about 44% of the Antarctic coastline (Drewry, 1983) are not represented in the model configuration (Fig. 2.4a). They appear as solid Antarctic coastline, such as, George VI Ice Shelf, or ocean, such as, Ross Ice Shelf. Thus, the southern boundary of the Southern Ocean is the Antarctic coastline, not the frigid ice shelves in the model. The water depth is deeper than 3000 m in most part of the Southern Ocean, except along the Antarctic coast, where the water depth is less than 1000 m (Fig. 2.4a). The islands in the Weddell Sea, such as, the South Georgia in the north, are not well
represented in the model due to the coarse resolution. Contrary to the Southern Ocean, which is circumpolar around the Antarctic continent, the Arctic Ocean is mostly bordered by continents (Fig. 2.4b). The water depth is generally deeper in the west than in the east. Along the eastern Arctic, the water depth is less than 1000 m (the broad Arctic continental shelf). In contrast, in the western Arctic, the water depth is up to 4000 m in the Eurasian Basin and Canadian Basin. To better represent the bottom topography, partial cell is employed at the bottom. Partial cell is rectangular like the full cell, but its thickness is allowed to be a function of latitude and longitude according to the bottom topography. To improve the representation of overflows and dense water spreading, the bottom boundary scheme of Beckmann and Döscher (1997) is employed to compute the near-bottom tracer transport.

Fig. 2.4. Bathymetry of ORCA2-LIM in (a) Antarctic, and (b) Arctic.

Lateral mixing of tracers and momentum is performed by a Laplacian operator. Momentum is horizontally mixed along the geopotential surface with a constant eddy viscosity coefficient of 40000 m$^2$/s between 20º and 90º, a varying coefficient from 20º to 2.5º, and a constant coefficient of 2000 m$^2$/s near the equator (< 2.5º). Lateral mixing of tracers is done along isopycnals, taking into account the eddy-induced velocity coefficients due to the baroclinic instability following Gent and McWilliams (1990). Vertical eddy diffusivity and viscosity coefficients are computed from the 1.5 level turbulent closure scheme based on a prognostic equation for the turbulent kinetic energy. The vertical eddy diffusivity is modified in case of double diffusive mixing (salt fingering and diffusive layering) following Merryfield et al. (1999) and in locations of statically unstable stratification.

Models were started to run from rest with the interannual CORE (Coordinated Ocean-ice Reference Experiments) atmospheric forcing data from 1958 through 2000 (Large and Yeager, 2004). The atmospheric data includes daily NCEP/NCAR reanalysis data of 10 m air temperature, wind speed and specific humidity (Kalnay et al., 1996), daily ISCPP data of the solar radiation and downwelling long wave radiation (http://isccp.giss.nasa.gov/), and monthly GPCP (south of 60ºS and 25ºS–25ºN; Huffman et al., 1997), Xie-Aarkin (65ºS–30ºS and 30ºN–65ºN; Xie and Arkin, 1996) and Serreze and Hurst (north of 70ºN; Serreze and Hurst, 2000) precipitation data. Prior to 1979, the precipitation data is in principle climatology.

Model runs were initialized using monthly mean temperature and salinity fields derived from Levitus dataset (1999). An initial sea ice thickness of 1 m in the Antarctic and 3 m in the Arctic was set for the regions of sea surface temperature below 0ºC. Above the ice, an initial snow cover with a thickness of 0.1 m in Antarctic and 0.5 m in the Arctic was assumed. To
avoid spurious model drift, the Levitus data (1999) was used to restore the sea surface salinity in the regions free of ice.

Table 2.2. Depth and thickness of the vertical layers in the ocean model

<table>
<thead>
<tr>
<th>Level</th>
<th>( T )</th>
<th>( w )</th>
</tr>
</thead>
<tbody>
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<td>Thickness of the level (m)</td>
</tr>
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<td>10.00</td>
</tr>
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<td>2</td>
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</tr>
</tbody>
</table>
Chapter 3 Control Run Simulations

The model ORCA2-LIM has been validated by Timmermann et al. (2005; hereafter referred as T2005) for process studies in high latitudes using the NCEP/NCAR forcing data. Their studies showed that, despite some deficiencies, the model could basically reproduce main ocean circulation patterns and realistic large-scale sea ice features in both hemispheres. Their studies also showed that the simulated sea ice coverage was very sensitive to the forcing data. This induces the necessity to explore the model results with another forcing dataset, which is performed here in the first section with the CORE dataset (Large and Yeager, 2004).

Due to the great importance of sea ice in the Earth climate system and particularly its sensitive response to climate change, the major task of this chapter is to explore the model sensitivity to sea ice parameters. The physical processes of LIM, e.g. vertical growth and decay, lateral growth and decay, and sea ice dynamics have been verified through series of sensitivity studies (Fichefet and Morales Maqueda, 1997). However, the individual parameters have not been thoroughly investigated, in particular those with particular difficulties to observe. In the second section of this chapter, we choose three parameters to conduct the sensitivity studies. The three parameters are the thickness of newly formed ice in leads \( h_0 \), the compressive strength of ice \( P^* \), and the turning angle in the oceanic boundary layer beneath sea ice \( \theta \).

The model configurations follow much of T2005. For example, in both studies the ocean model ORCA is run with a time step of 5760 s (96 min or 15 time steps per day); ORCA is coupled to LIM every five time steps. However, this study differs from T2005 in the following aspects:

i) This study uses the interannual CORE forcing data from 1958-2000, while T2005 used the interannual NCEP/NCAR forcing data from 1977-1999;

ii) This study uses monthly mean hydrographic fields from Levitus (1998) dataset for initialization and restoring, while T2005 used the January data from the Polar Science Center Hydrographic Climatology;

iii) This study adopts a weaker restoring for sea surface salinity with a timescale of 292 days, instead of 60 days used by T2005. This is because less artificial restoration of sea surface salinity is required when AISW is applied. The restoring freshwater flux decreases from \(-1.45 \times 10^{-5} \text{ kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}\) to \(-1.17 \times 10^{-5} \text{ kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}\) when the AISM is taken into account (Wang and Beckmann, 2007).

### Table 3.1. List of Experiments

<table>
<thead>
<tr>
<th>Case</th>
<th>( P^* ) (kPa)</th>
<th>( h_0 ) (m)</th>
<th>( \theta ) (º)</th>
</tr>
</thead>
<tbody>
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</tr>
<tr>
<td>P1</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>P2</td>
<td>30</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>H1</td>
<td>-</td>
<td>0.3**</td>
<td>-</td>
</tr>
<tr>
<td>H2</td>
<td>-</td>
<td>0.9**</td>
<td>-</td>
</tr>
<tr>
<td>A1</td>
<td>-</td>
<td>-</td>
<td>0</td>
</tr>
</tbody>
</table>

- : keeps same value as in the Std experiment;
**: only changes in the southern hemisphere.

For the sensitivity studies, a standard configuration is established in this chapter. This reference case is also prepared for a later comparative investigation of the impact of the AISW (Chapters 4–7). The values of \( P^*=20 \text{ kPa}, \ h_0=0.6 \text{ m}, \) and \( \theta = 15^\circ \) are used in the standard experiment (Std), with variations in one of them in the other five experiments as shown in Table 3.1. All the experiments start to run at rest for one pass (43 model year) forced with the

3.1. Standard experiment

The Std experiment is only roughly investigated in this chapter via one pass results. Further exploration will be performed in the later chapters through comparisons with the case that includes AISM after the long-term integration. For the sensitivity studies, the focus here is on the sea ice extent (the integrated area of all grid cells having an ice concentration at least 15%), sea ice concentration, sea ice thickness and water transport through the Drake Passage (DP).

![Fig. 3.1. Standard simulation of the Antarctic (upper) and Arctic (lower) sea ice extent (a) time series, and (b) seasonal cycle.](image)

3.1.1. Sea ice extent

Figure 3.1 shows the standard simulation of the Arctic and Antarctic sea ice extent. Similar to T2005 study, the model reaches a quasi-steady seasonal cycle after only a few years under the CORE forcing data (Fig. 3.1a). Seasonal cycle and interannual variability are clearly seen in both hemispheres. The seasonal sea ice extent ranges from $2.5\times10^6$ km$^2$ to $20\times10^6$ km$^2$ in the Antarctic, while the variation is from $9\times10^6$ km$^2$ to $16\times10^6$ km$^2$ in the Arctic.

![Arctic](image)

The simulated seasonal sea ice extent (Fig. 3.1b) increases to its maximum in August (February) and retreats to its minimum in February (August) in the Antarctic (Arctic). The maximum sea ice extent is $17.46\times10^6$ km$^2$ in the Antarctic and $16.21\times10^6$ km$^2$ in the Arctic, whereas the minima are $0.25\times10^6$ km$^2$ and $8.31\times10^6$ km$^2$ in the Antarctic and Arctic, respectively. The simulated minimum sea ice extent is, however, less consistent with the observations than the maximum sea ice extent in both hemispheres. In addition, the Antarctic winter sea ice extent approaches its maximum one month earlier (in observations, the maximum occurs in September). These results are again similar to those of T2005 where the NCEP/NCAR forcing data were used. As a whole, our simulated minimum sea ice extents for both hemispheres are smaller than those of T2005. The minimum sea ice extent in the Antarctic is here $0.25\times10^6$ km$^2$ but in T2005 it was $3\times10^6$ km$^2$, which was already in underestimation. On the other hand, the minimum sea ice extent in the Arctic is here $8.31\times10^6$ km$^2$, which agrees much better with the observations ($8.33\times10^6$ km$^2$) than that of T2005 ($10.5\times10^6$ km$^2$). These differences illustrate the sensitivity of the summer sea ice extent to the forcing dataset. The different responses of the Antarctic and Arctic summer sea ice extents
suggest that the summer sea ice extent should also depend on other factors, such as oceanic heat flux in addition to the surface forcing (Gordon and O'Farrell, 1997). Exceedingly high upward oceanic heat flux may result in excessive summertime melting (Marsland and Wolff, 2001) and thus lead to the low Antarctic sea ice extent in this study.

3.1.2. Sea ice concentration

The simulated Arctic sea ice retreats to its minimum in August (Fig. 3.2a). The Canadian and Eurasian Basins are fully covered by sea ice. The Kara Sea, the Laptev Sea, part of the Greenland Sea, the Baffin Bay and the Hudson Bay are also covered by sea ice. The sea ice extent in the bays and in the Kara Sea and Laptev Sea are overestimated because the model resolution is relatively coarse (T2005). In contrast, the sea ice concentration in the Canadian Basin and Eurasian Basin is somehow underestimated; it is about 0.9–1.0 from observations, but about 0.7–0.9 in the simulations.

![Image of Arctic and Antarctic sea ice concentration](image)

**Fig. 3.2.** Monthly mean Arctic sea ice concentration in (a) August and (b) February, and Antarctic sea ice concentration in (c) August in the Std experiment.

In February, Arctic sea ice grows to its maximum extent (Fig. 3.2b). The simulated results are generally in good agreement with the observations. The sea ice extent in the Canadian Basin and Eurasian Basin, most of the Arctic Marginal Seas, the Baffin Bay and Hudson Bay, is consistent with the observations. These regions are also covered by compact ice ($A > 0.8$), similar to the observations. The present study shows the same deficiencies as those in T2005: (i) The sea ice extent is overestimated in the Barents Sea; (ii) The width of the ice stream along the east coast of the Greenland Sea and in Denmark Strait is overestimated; and (iii) The Odden Ice Tongue in Greenland Sea is not reproduced by the model, and (iv) A wider ice stream is obtained in the Davis Strait. One improvement here is the presence of sea ice in the Baltic Sea compared with the study of T2005, although the area is overestimated.

Unlike Arctic sea ice confined by the circumpolar continent, the Antarctic sea ice expansion is limited by the polar front in the warm Antarctic Circumpolar Current (ACC). The simulated sea ice advances to its maximum extent in August (Fig. 3.2c). Basically, the sea ice concentration is smaller in the eastern Antarctic than that in the western Antarctic. Most parts of the Weddell Sea, Ross Sea, Amundsen Sea and Bellingshausen Sea are covered by compact ice, except for the regions near the ice edge. This feature agrees with observations, although sea ice concentration is a little bit underestimated in the Amundsen Sea, Bellingshausen Sea and Ross Sea. Sea ice shows northward expansion in the Ross Sea and Amery Sea, corresponding to the oceanic gyres in them. However, the reproduction of the northward intrusion of the Weddell Sea Gyre is unsatisfactory. In February, Antarctic sea ice reaches its minimum extent. Sea ice is reasonably reproduced in the Antarctica, except in the east. There, patches of sea ice are still seen in the observations along the coast, while they totally melt away in the simulation.
3.1.3. Sea ice thickness

![Images of sea ice thickness maps](a) (b) (c)

Fig. 3.3. Monthly mean Arctic sea ice thickness in (a) August and (b) February, and Antarctic sea ice thickness in (c) August in the Std experiment.

The simulated Arctic sea ice thickness fields (Fig. 3.3a, b) are generally consistent with observations (Bourke and McLaren, 1992). Due to sea ice convergence and ice ridging, thickest ice (more than 3.5 m) is located off the Canadian Archipelago and north Greenland, where the sea ice can survive through the summer. Drifting with the Beaufort Gyre, sea ice, whether in summer or winter, shows a pronounced gradient in thickness from the Siberian Shelf to the Canadian Archipelago. Sea ice thickness close to the Scandinavia exhibits a shape like “A” (Fig. 3.3b), due to the intrusion of the warm North Atlantic water.

In contrast to Arctic sea ice, Antarctic sea ice is much thinner with much less perennial ice. Even in austral winter, the thickness is less than 2 m in the simulation (Fig. 3.3c). On the whole, the thickness is larger in the west Antarctic than in the east Antarctic. Sea ice is generally less than 1 m in the east, and mostly more than 1 m in the west, except near the ice edge. In the Weddell Sea, sea ice thickness decreases from south to north. In the simulations, multi-year ice appears mainly along the southern coast of the Weddell Sea, rather than its actual presence along the eastern coast of the Antarctic Peninsula (e.g. Strass and Fahrbach, 1998; Harms et al., 2001). This deficiency also appeared in the studies of T2005 and Stössel et al. (2007). In austral summer, sea ice exists only in the southern Weddell Sea and Ross Sea, and in the southern part along the west coast of the Antarctic Peninsula. The monthly mean ice thickness is usually small (up to 0.2–0.4 m) except in the Weddell Sea. In the southern Weddell Sea, the simulations produce multi-year ice and its thickness often reaches 1 m.

3.1.4. Water transport through Drake Passage (DP)

The ACC dominates the Southern Ocean and transports large volume of water (about 130 Sv) eastward. Due to the connection with the three great ocean basins, Atlantic, Pacific and Indian, it transfers substantial amounts of heat, salt and other properties between the basins, thus playing a key role in the global circulation. To better estimate the global ocean circulation, the World Ocean Circulation Experiment (WOCE) has identified a number of sections in the Southern Ocean for repeated observations. One such section is the ACC across DP, which is the water body between the southern tip of South America at Cape Horn, Chile and the South Shetland Islands of Antarctica.

Figure 3.4 shows the time series of the monthly and annual means of the vertically integrated ACC volume transport through DP. The annual mean transport varies from 168 Sv to 193 Sv during the first 43 years of integration. It exhibits an increasing trend in the Std experiment, inconsistent with the steady transport between 1975–2000 from Cunningham et al. (2003). The increasing trend can be suppressed to a more steady state through introduction of AISM as shown in the next chapter, since buoyancy forcing is an important factor driving the
ACC (Gnanadesikan and Hallberg, 2000; Gent et al., 2001). The annual mean transport during
the 43-year integration is about 179.2 Sv in this simulation. It is largely consistent with the
simulated annual mean transport of 170 Sv during 23 years from T2005, but it is larger than
the real-world value suggested by observations, about 100 Sv from Orsi et al. (1995), 134 ± 13
Sv from Whitworth and Peterson (1985) and Whitworth (1983), and 134 Sv ± 15 to 27 Sv
from Cunningham et al. (2003).

![Fig. 3.4. Time series of the ACC volume transport through DP. The line of black: monthly
mean, and red: annual mean.]

### 3.2. Sensitivity Experiments

In this section, we investigate the model sensitivity to three parameters: the thickness of
newly formed ice in leads ($h_0$), the compressive strength of ice ($P^*$), and the turning angle in
the oceanic boundary layer beneath sea ice ($\theta$).

The first parameter $h_0$ is concerned with sea ice thermodynamic process (Eq. 2.47), a fixed
demarcation thickness between thin and thick ice. When a lead loses its heat to the
atmosphere, newly formed sea ice accumulates to the side of the thick ice. The accumulated
sea ice is usually assumed to have a thickness of $h_0$ in sea ice models. A value of 0.5 m was
assigned in the studies of Hibler (1979) and Fichefet and Morales Maqueda (1997). This
slightly underestimates the actual growth in the Arctic (see Hibler, 1979). Therefore, in this
chapter we take $h_0 = 0.6$ m as the standard (Std) value. For the sensitivity studies, three
experiments are performed for this parameter (see Table 3.1), with the focus on the southern
hemisphere. In addition to the standard experiment, in experiments H1 and H3 $h_0$ is taken as
0.3 m and 0.9 m for the southern hemisphere, but remained 0.6 m in the Arctic.

The compressive strength of ice $P^*$ (Eq. 2.35) concerns with sea ice dynamics. It is a key
parameter in sea ice simulations (e.g. Hibler, 1979), usually in the range of 5–30 kPa. Due to
the quadratic relationship between wind stress and wind speed (Eq. 2.31), it is not surprising
to use a lower $P^*$ for average wind over several days mean or a month, and a higher value for
daily winds (Wang, 2007). For the former case, 5 kPa and 7 kPa were used by Hibler (1979)
and Tremblay and Mysak (1997), respectively. For the later one, most of the previous studies
used 20–30 kPa. For example, Harder and Lemke (1994) used 20kPa, Hibler and Walsh
(1982), and Flato and Hibler (1991, 1992) employed 27.5 kPa, and Zhang (2000) and Wang et
al. (2003, 2006) adopted 30 kPa. Here three experiments are conducted for this parameter (see Table 3.1). In the Std experiment, 20 kPa is used due to taking daily wind field in the atmospheric forcing dataset, while 10 kPa and 30 kPa are set for $P^*$ in the other two experiments of P1 and P2, respectively.

The momentum exchange between ocean and sea ice is depicted by Eq. 2.31. The parameter $\theta$ describes the turning angle or Ekman angle in the oceanic boundary layer beneath sea ice. A variety of values has been adopted in the previous studies. For example, Hibler (1979) used 25º for the Arctic simulations, Fichefet and Maroles Maqueda (1997) adopted 25º for the northern hemisphere and -25º for the southern hemisphere, while T2005 used 0º in their studies for high latitudes in both hemispheres. In this chapter, two experiments are performed (see Table 3.1). Experiment A1 takes $\theta$ equal to 0º, while in the Std experiment it is taken 15º for the northern hemisphere and -15º for the southern hemisphere.

### 3.2.1. Experiments H1 and H2

The sensitivity of the Arctic sea ice to the variation in $h_0$ has earlier been studied by Hibler (1979). Therefore, the focus here is mainly on the Antarctic sea ice. As shown in Fig. 3.5, the variation of $h_0$ has little impact on the monthly mean sea ice extent, but has substantial influence on the sea ice volume. For example, the winter maximum ice volume is only about $10 \times 10^3$ km$^3$ when $h_0=0.3$ m; whereas it reaches $18 \times 10^3$ km$^3$ when $h_0=0.9$ m, increasing by 80% more compared with $h_0=0.3$ m. It is therefore a critical parameter for the sea ice mass balance in the Antarctic.

![Fig. 3.5. Response of the Antarctic sea ice extent (upper) and ice volume (down) to the variation of $h_0$.](image)

The high sensitivity of the sea ice volume to the selection of $h_0$ can be explained as follows. $h_0$ has both dynamic and thermodynamic effects on the simulated ice volume. On the thermodynamic aspect, it determines the initial ice thickness during lateral growth. Since the overall ice volume during a time step is determined by the heat flux, a higher $h_0$ would have a
smaller ice area formed laterally. This would leave more open water, in which new ice forms more rapidly. On the dynamic aspect, the resulting lower ice concentration due to this thermodynamic treatment would reduce the ice strength, thus enhancing the compressive deformation and forming thicker ice. Therefore, on both dynamic and thermodynamic aspects, increasing $h_0$ would enhance the increase in ice thickness, and therefore the total ice volume. The nearly linear dependence of ice volume on $h_0$ is clearly seen in Fig. 3.5 (lower panel), which has earlier been demonstrated by Hibler (1979). However, the dependence of summer ice volume on $h_0$ is much weaker since the main process of sea ice is ablation.

In contrast, sea ice extent is insensitive to the selection of $h_0$ (Fig. 3.5, upper panel). From thermodynamic point of view, the heat flux between the atmosphere and the ocean approximates zero at the ice edge. This means that there is little lateral growth there since sea ice extent is basically controlled by the heat budget distribution when dynamic processes remain similar.

The nearly linear dependence of ice volume on $h_0$ is clearly seen in Fig. 3.5 (lower panel), which has earlier been demonstrated by Hibler (1979). However, the dependence of summer ice volume on $h_0$ is much weaker since the main process of sea ice is ablation. In contrast, sea ice extent is insensitive to the selection of $h_0$ (Fig. 3.5, upper panel). From thermodynamic point of view, the heat flux between the atmosphere and the ocean approximates zero at the ice edge. This means that there is little lateral growth there since sea ice extent is basically controlled by the heat budget distribution when dynamic processes remain similar.

The impact of variation in $h_0$ on the spatial distribution of the Antarctic sea ice concentration and thickness is shown in Fig. 3.6. As can be seen, the regional responses of sea ice concentration to the variation of $h_0$ are different. An increase in $h_0$ generally leads to lower ice concentration in the west side, and to higher ice concentration in most regions in the east side of the Southern Ocean (Fig. 3.6a). Large changes in ice concentration mainly occur in the vicinity of the marginal ice zone, whereas in the inner part it is hardly affected because the ice concentration is already high and little new ice can be formed in open water. In contrast, sea ice thickness increases in most part of the Southern Ocean when $h_0$ is larger. In the Ross Sea and the Southeastern Weddell Sea, the thickness differences even reach more than 0.4 m when $h_0$ increases from 0.3 m to 0.6 m.

The transport through the DP increases with $h_0$. The mean transports are 168.2 Sv, 179.2 Sv and 189.7 Sv for $h_0$ equal to 0.3 m, 0.6 m and 0.9 m, respectively. A change in $h_0$ by 0.3 m is therefore likely to induce a change in transport of about 10–11 Sv. However, the long-term increasing trend (see Fig. 3.4) remains similar regardless of the change in $h_0$. The increase of the DP transport with $h_0$ is explained as follows. The DP transport depends on the thermohaline circulation off the Antarctic shelf (Gent et al., 2001), which is related to the brine rejection during sea ice formation (Weatherly et al., 1998). Increase in $h_0$ results in more sea ice formation (Fig. 3.6b), thereby more brine rejected into the ocean. As a consequence, it
increases the strength of the thermohaline circulation off the Antarctic shelf, and therefore the DP transport.

Fig. 3.7. The impact of ice compressive strength ($P^*$) on the seasonal mean sea ice extent in (a) the Antarctic, and (b) the Arctic, and on the seasonal sea ice volume in (c) the Antarctic, and (d) the Arctic.

3.2.2. Experiments P1 and P2

The variation in the compressive strength of ice $P^*$ has little impact on the sea ice extent in both hemispheres (Fig. 3.7a and b). However, its impact on the sea ice volume is pronounced (Fig. 3.7c and d). For example, the Arctic sea ice volume becomes 3–4$x10^3$ km$^3$ larger when $P^*$ takes 10 kPa instead of 30 kPa, which is equivalent to 10% of the annual mean sea ice volume (Fig. 3.7d). In the Antarctic, the difference in ice volume reaches 2–3$x10^3$ km$^3$ in winter and spring, which is equivalent to 20–30% of the seasonal mean (Fig. 3.7c). In addition, the temporal response of the Antarctic sea ice to the change in $P^*$ is different from that of the Arctic sea ice. In the Arctic, the seasonal difference remains relatively stable, whereas in the Antarctica the seasonal variation in the volume difference is clearly seen (cf. Fig. 3.7c and d).

Fig. 3.8. Monthly mean difference (Std – P1) in sea ice thickness (unit: m) (a) in August, Antarctic, (b) in August, Arctic, and (c) in February, Arctic.

Changes in sea ice thickness due to the variation of $P^*$ are more pronounced and systematical than those in sea ice concentration. Figure 3.8 shows monthly mean difference (Std – P1) in sea ice thickness in the Antarctic and in the Arctic. In Antarctic winter (Fig. 3.8a), changes in sea ice thickness are very little with larger $P^*$ in most marginal ice zones,
while in the inner part a clear thinning can be seen, especially, in the Weddell Sea, Ross Sea, Amundsen Sea and Bellingshausen Sea, where perennial sea ice exists. In the Arctic, the thinning is present mainly along the coasts of Greenland and Canadian Archipelago and on a part of the Siberian coast (Fig. 3.8b and c), whereas minor thickening occurs mainly along the marginal ice zone in the Barents Sea and around the New Siberian Island (Fig. 3.8b and c). In comparison, thinning in the Arctic is much larger than that in the Antarctic. The decrease in sea ice thickness is up to more than 0.8 m in the Arctic, while it is at most up to 0.3 m in the Antarctic. It implies that \( P^* \) is a more critical parameter in the simulation of the Arctic sea ice.

**Fig. 3.9.** Monthly mean wind in August in the Antarctic; the vector scale is 10.0 m/s.

**Fig. 3.10.** Impact of the turning angle on the sea ice extent in (a) Antarctic and (b) Arctic, and on the sea ice volume in (c) Antarctic and (d) Arctic.

The response of ice thickness distribution to \( P^* \) can be explained as a result of the momentum balance and the advection of ice mass. Due to the Coriolis force, sea ice tends to drift to the right of the wind in the northern hemisphere and to the left in the southern
hemisphere. However, in the marginal ice zonal, a zonal westerly wind in the Southern Ocean (Fig. 3.9) will induce a free drift in the marginal sea ice, resulting in no impact of $P^*$ on the drift, thereby little impact on the sea ice thickness distribution. In the inner part of the Weddell Sea and the Ross Sea (Fig. 3.6a), where the ice is thicker, decrease in $P^*$ would cause more severe ice ridging, therefore forming thicker ice. The ice thickening due to the decrease of $P^*$ in the Arctic can be similarly explained. On the whole, less ice compressed to the Canadian Archipelago and the northern Greenland causes the decrease in the ice volume in the Arctic (Fig. 3.8d).

Variation in $P^*$ has little effect on the transport through the DP. The total transport through the DP is 178.8 Sv and 179.6 Sv for $P^*$ equal to 10 kPa and 30 kPa, respectively. Their differences from the Std experiment (179.2 Sv) are very small (less than 1 Sv), due to the little change in the hydrography in the ACC region.

3.2.3. Turning angle in the ocean boundary layer beneath sea ice ($\theta$)

The impact of the turning angle $\theta$ on the Arctic and Antarctic sea ice is different. As shown in Fig. 3.10, a change in $\theta$ has little effect on the Arctic sea ice, either on sea ice extent or on sea ice volume (Fig. 3.10b and d). On the contrary, both sea ice extent (Fig. 3.10a) and sea ice volume (Fig. 3.10c) in Antarctica decrease noticeably with the omission of $\theta$ during the ice melting season from September to December (Fig. 3.10b and d). On the whole, the annual mean sea ice extent decreases $0.49\times10^6 \text{km}^2$ ($9.64\times10^6 \text{km}^2$ for Std and $9.15\times10^6 \text{km}^2$ for A1) in the Antarctic, but only $0.1\times10^6 \text{km}^2$ ($12.56\times10^6 \text{km}^2$ for Std and $12.46\times10^6 \text{km}^2$ for A1) in the Arctic.

![Fig. 3.11. Monthly mean difference of (A1 – Std) in sea ice concentration in (a) August in the Antarctic, (b) in August in the Arctic, and (c) in February in the Arctic.](image1)

![Fig. 3.12. Monthly mean difference (A1 – Std) in sea ice thickness (unit: m) in (a) August in the Antarctic, (b) in August Arctic, and (c) in February Arctic.](image2)
The changes in sea ice concentration due to the variation of the turning angle are not uniform in the two hemispheres. In the Arctic, the turning angle of $\theta$ generally has small effect on the distribution of sea ice concentration, whether in winter (Fig. 3.11c) or in summer (Fig. 3.11b). This is because most of the sea ice covered area in the Arctic undergoes relatively free drift. Relatively larger effects are mainly found near the ice edge. Nevertheless, variations in turning angle have negligible effect on the total sea ice extent in major part of the Arctic Ocean, since if the Arctic regions are all taken into account, the increasing area and the decreasing area are much more balanced (Fig. 3.11b, c). This also results in the quite small changes in the seasonal cycle of the sea ice extent (Fig. 3.10b). In contrast, sea ice generally becomes more compact in most part of the Southern Ocean when $\theta$ changes from -15° to 0° (Fig. 3.11a).

Omission of the turning angle induces regional changes in the spatial distribution of ice thickness in the Antarctic (Fig. 3.12a) and in the Arctic (Fig. 3.12b and c). Ice thickening is clearly visible in most part of the Southern Ocean, more pronounced in the Ross Sea and Weddell Sea, where the thickening is up to 0.4 m. Ice thinning is mainly found along the Antarctic coast and in the sector 135°E–180°E. In the Arctic, the thickening is mainly present in the Eurasian Basin, while in the Canadian Basin sea ice is experiencing thinning. On the whole, the ice volume keeps relatively balanced over the whole domain (Fig. 3.10d).

The oceanic boundary layer turning angle affects the transport through the DP. The transport increases after ignoring the turning angle in the experiment of A1. A rapid increasing trend is seen especially after 1975. The total transport through the DP is 186.4 Sv in the experiment of A1, 7.2 Sv more compared with that in the Std experiment.

### 3.3. Concluding remarks

This chapter has explored ORCA2-LIM model and its sensitivity to three ice parameters: the newly formed ice thickness in leads ($h_0$), the compressive strength of ice ($P^*$), and the ocean boundary layer turning angle ($\theta$), using one pass results. The motivation of the exploration is the model sensitivity to the forcing data as identified by T2005. It was found that the summer ice extent was especially sensitive to the forcing data, and that the Arctic and Antarctic sea ice responded differently to the present forcing data. The evolution of the sea ice in the Baltic Sea was successfully simulated in the present study, while it failed in T2005. Using the current configuration, the water transport through the DP was 179.2 Sv in this study. Although overestimated by about 30–42 Sv compared with the observations (e.g., Cunningham et al., 2003), it is very close to the 170 Sv in T2005, who used the interannual NCEP/NCAR forcing data instead of the interannual CORE forcing data.

In total, six experiments were conducted through changing one of the three parameters, with $h_0$ only changed in the southern hemisphere and $P^*$ and $\theta$ changed in both hemispheres. Variations of these three parameters affect the sea ice extent and volume. On the whole, sea ice volume is more sensitive than the sea ice extent; Antarctic sea ice is more sensitive than Arctic sea ice. In particular, the values of $h_0$ and $P^*$ have little impact on the seasonal sea ice extent but lead to large changes in the seasonal sea ice volume. Therefore, choosing the values of $h_0$ and $P^*$ is more important for the simulation of the seasonal sea ice volume. In contrast, the variation in the turning angle has little impact on the sea ice extent and volume in the Arctic, but tends to reduce them in the Antarctica if $\theta$ is ignored. Therefore, selection of $\theta$ is more important for the Antarctic sea ice extent and volume.

Variations in the three parameters have different impacts for the other aspects of the sea ice cover, such as the sea ice concentration and thickness. Choosing $P^*$ is not critical for the sea ice concentration but is a key factor for multi-year ice thickness. Variation of $h_0$ affects sea ice concentration only near the ice edge but influences the ice thickness over the whole Southern Ocean. Variation of $\theta$ only plays an important role in some ice regions, such as the Weddell...
Sea, the Bellingshausen Sea to the Ross Sea in the Antarctica, and the Greenland Sea and Barents Sea in the Arctic.

Variations of $h_0$, $P^*$ and $\theta$ influence the transport through the DP. Of the three parameters, the magnitude of $P^*$ in the present study had the least impact. On the other hand, the other two parameters have larger influence on the transport through the DP. A 0.3-m change in $h_0$ can result in 10–11 Sv changes in the transport through the DP, and an omission of 15° turning angle can induce a 7.2 Sv change. As a summary, the impacts of the three parameters on the sea ice cover and the DP transport are shown in Table 3.2.

Table 3.2. The impacts of the variations in the three study parameters.

<table>
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<td>the water transport through the DP</td>
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Note: +: important; -: not important; ±: important in some region.
Chapter 4 Addition of Antarctic Ice Shelf Melting and Validation

From Chapter 2 we know that the model domain only extends to 78.19°S in the south. However, the two largest Antarctic ice shelves, Ross and Filchner-Ronne, extend to 86°S and 82°S, respectively, both far beyond the model domain. Thus, the two Antarctic ice shelves as well as other Antarctic ice shelves are not included in the model. Due to its importance in the climate system (see Chapter 1), ice shelf-ocean interaction ought to be included in ocean models (BG03). However, explicitly including sub-ice shelf cavities into climate models still remains extremely difficult (see section 1.4.3). Is there any possibility to include the ice shelf-ocean interaction in a model but without explicitly including the ice shelf cavities? The parameterization of BG03 for ice shelf-ocean interaction provides one such opportunity. With this approach, the net effect of the complicate processes within sub-ice shelf cavities can be reasonably accounted for. This chapter presents the implementation of BG03 parameterization into ORCA2-LIM model and performs the model validation.

4.1. Parameterization of BG03 and its implementation

4.1.1. Parameterization of BG03

The parameterization of BG03 uses three equations to represent the thermodynamic interaction between the ice shelf and ocean, which includes the net heat and freshwater flux from the ISM and the pressure-dependent freezing point temperature. The net heat flux between the ice shelf and ocean is computed as,

$$ Q_{net} = \rho_w c_{pw} \gamma T (T_{ocean} - T_f) A_{eff} $$

where $\rho_w$ is the sea water density (=1000 kg m$^{-3}$), $c_{pw}$ is the specific heat (=4000 J (kg °C)$^{-1}$), $\gamma T$ is the transfer coefficient (=10$^{-4}$ m s$^{-1}$), $T_{ocean}$ is the averaged oceanic temperature, taken the temperature at the southern boundary of the model at the depth of 200–600 m according to BG03, and $T_f$ is the pressure-dependent freezing point temperature at the edge of the ice shelf base; $A_{eff}$ is an “effective area” for basal melting, described as,

$$ A_{eff} = WL_{eff} $$

where $W$ is the along-shelf width, and $L_{eff}$ is an effective cross-shelf length (penetrative length), generally much smaller than the actual cross-shelf extent of the ice shelf. It is relatively uniform, on an order of a few kilometers (typically ranging from 5 to 15 km), regardless of the size and geometric form of the ice shelf (BG03). This implies that at least on the first order the net melting can be determined by the ice shelf edge processes, even though a significant portion of melting occurs at or near the grounding line (BG03). Here the value of $L_{eff}$ is taken equal to 10 km following Wang and Beckmann (2007). The net freshwater flux (m$^3$/s) is computed from,

$$ \frac{\partial m_{net}}{\partial t} = \frac{Q_{net}}{\rho_0 L_{is}} $$

where $\rho_0$ is the density of the ice shelf (=920 kg m$^{-3}$), and $L_{is}$ is the latent heat of fusion of the ice shelf (=334,000 J kg$^{-1}$). The calculated heat and freshwater fluxes are then used as the lateral boundary conditions for the ORCA2-LIM model. This treatment avoids large modification of the model configuration, while maintains fairly well the physical processes of ISM. For more details about this parameterization, reference is given to BG03.
4.1.2. Implementation of the parameterization of BG03 into ORCA2-LIM

The original ORCA2-LIM model does not include the Antarctic ice shelves and their interaction with the ocean. AISM, which is important in the climate system, is here added into the ORCA2-LIM model according to the approach of BG03. ORCA2-LIM model with the parameterization of BG03 is named as ORCA2-LIM-ISP model. Large ice shelves, such as Filchner-Ronne ice shelf (FRIS) and Ross Ice Shelf (RIS), represent an important component for the ventilation of the world ocean abyss. In addition, small ice shelves may contribute significantly to the freshwater budget and the preconditioning of the Weddell Sea continental shelf water mass, such as the Eastern Weddell (EW: Riiser-Larsen and Brunt) Ice Shelves (Thoma, 2006). Furthermore, minor ice shelves might experience high melt rates, such as ice shelves in the Amundsen Sea (Jacobs et al., 1996). Therefore, we have included all the major ice shelves and a number of minor ice shelves in Antarctica (Fig. 4.1) in ORCA2-LIM-ISP model as was done by Wang and Beckmann (2007).

It is noteworthy that sub-ice shelf cavities are only implicitly included in the present arrangement, and the southern boundary of the model is still the solid continent. Additional heat and freshwater fluxes due to ISM are added to the model through modification of the lateral boundary condition where the ice shelves are considered (Fig. 4.1). For most ice shelves, it is assumed that the ice shelf edge is next to the continent; therefore the fluxes are taken into the model at the grid points next to the continent. For FRIS and RIS, however, the fluxes are taken at the model southern boundary where the ice shelf edge coincides with the model boundary. In general, the relatively coarse resolution of the model needs some compromises and simplifications. For example, George VI Ice Shelf at the western side of the Antarctic Peninsula, which shaped like an inverse “C”, is represented by only two separate points (Fig. 4.1). Although the size of the Pine Island Glacier is much smaller than the model grid size, it is included as one point because of its high melting rate (Rignot et al., 2008). The ice edge thickness, which is relevant to the in-situ freezing temperature and consequently the ice shelf-ocean interaction, must be specified. It is assumed to be 200 m for most of the ice shelves, except for FRIS and RIS whose thicknesses are taken from the dataset of Johnson and Smith (1997).

Fig. 4.1. The name, extent and position of the Antarctic ice shelves included in the ORCA2-LIM-ISP model. EW: ice shelves of Riiser-Larsen and Brunt.
Using the above configuration, ORCA2-LIM-ISP is run for two passes (totally 86 model years) forced with the CORE forcing data from 1958–2000. For comparison, another experiment without AISM is conducted using ORCA2-LIM, which already ran one pass in the control run simulations for the STD experiment, and now continues to run another pass. The other parameters are the same in the two experiments. Here the first pass is basically for the model spin-up, and the second pass is for analysis. Note the results with AISM in the following are from ORCA2-LIM-ISP and without from ORCA2-LIM.

Table 4.1. The WOCE sections, observation sites, observation periods and areas used for validation

<table>
<thead>
<tr>
<th>Sections</th>
<th>Observation site</th>
<th>area used to validate</th>
<th>Observation period</th>
</tr>
</thead>
<tbody>
<tr>
<td>P18</td>
<td>105°W</td>
<td>74° S – 30° S</td>
<td>Feb. 22 – Apr. 5, 1994</td>
</tr>
<tr>
<td>A23</td>
<td>30°W</td>
<td>75° S – 46° S</td>
<td>Mar. 20 – May 06, 1995</td>
</tr>
<tr>
<td>S2</td>
<td>0°E</td>
<td>70° S – 46° S</td>
<td>May 21 – July 05, 1992&lt;br&gt;Mar. 28 – May 21, 1996</td>
</tr>
</tbody>
</table>
As mentioned by Wang and Beckmann (2007), the parameterization of BG03 is a crude way to include ISM effect. Therefore, both overestimation and underestimation of ISM are possible. In the following, we shall show that addition of AISM generally improves the simulated hydrography in the Southern Ocean. In some regions, the improvement is significant.

4.2. Validation of $T/S$ against observations

In this section, the simulated hydrography of the Southern Ocean is validated with the WOCE Southern Ocean Atlas (Orsi and Whitworth, 2004). WOCE, a part of the World Climate Research Program, used resources from nearly 30 countries to make unprecedented in-situ and satellite observations of the global ocean during 1990s. In the Southern Ocean, the observations were made mostly during austral summer and autumn in 1990–1999 through 15 sections, of which only S4 in the circumpolar direction, the others all in the south-north direction (Fig. 4.2). More details about the WOCE Southern Ocean Atlas can be found in http://woceatlas.tamu.edu. The sections used here for validation are marked with red rectangles (Fig. 4.2) and their details are shown in Table 4.1. For validation, the climatological means of the simulations are computed for the same periods as the observations.

4.2.1. Section S4

Section S4 is taken around 65°S (Fig. 4.2). It is the only circumpolar section in the WOCE observations. Here we only take the part off Amery Ice Shelf. It is seen that the topography is highly variable in this region. There is a seamount between 78–86°E (Fig. 4.3a, b), whose topography is fairly well represented in the model (Fig. 4.3c–f).
In the observations, there is a core of Warm Deep Water (WDW) at about 500 m depth with the temperature over 1.0°C, enclosed by cold surface ($\leq -0.8^\circ$C) and deep ($\leq 0^\circ$C) waters (Fig. 4.3a). East of the seamount, the bottom water temperature is about $-0.4^\circ$C.

It is seen that addition of AISM significantly improves the simulation of the water temperature (cf. Fig. 4.3a, c, and e). The simulated bottom water temperature was about $-0.8^\circ$C before inclusion of the AISM (Fig. 4c), much colder than the observation (Fig. 4.3a). After addition of AISM the bottom water temperature is warmed up to $-0.3^\circ$C (Fig. 4.3e), much closer to the observations. In addition, the simulated lower lobe of the 0°C isotherm is very close to the observations (around 2500 m) after inclusion of the ISM; Prior to that it was much shallower (about 1500 m). While both simulations underestimate the temperature of the WDW, addition of AISM increases the core of the WDW to a temperature above 1.0°C, which is approximately 0.2°C higher than that without ISM (cf. Fig. 4.3c and e). In the east of the seamount, this improvement is even more obvious.

The addition of AISM also improves representation of salinity. Observations show a maximum salinity more than 34.72 psu at the depth of 1000 m, enclosed by fresher surface water (salinity < 34.2 psu) and deep water (salinity < 34.68 psu) (Fig. 4.3b). The depth of this saline water (salinity $\geq 34.7$ psu) is about 1000-1500 m (Fig. 4.3b). However, this saline water is fully missing in the simulation without AISM; instead, it is replaced by fairly homogeneous seawater with salinity about 34.69 psu (Fig. 4.3d). In the simulation with AISM, we can see that this saline water layer is fairly well captured (cf. Fig. 4.3b and f). In addition, the relatively fresh bottom water is also successfully reproduced, with the simulated salinity very close to the observations.

The topographic effect on the hydrographic structure is very strong in the observations (Fig. 4.3a, b), in particular, the apparent doming of the isotherm and isohaline in the regions above the shoulder of the seamount. The doming is well reproduced in the simulated temperature fields (Fig. 4.3c, e), but it is invisible in the simulated salinity field before inclusion of AISM (Fig. 4.3d). The domings of the isohalines are captured in the simulation with AISM (Fig. 4.3f), although the signal is much weaker than in the observations, most probably due to the coarse resolution of the model.

### 4.2.2. Section I8

Section I8 is also close to Amery Ice Shelf, but in the meridinal direction (Fig. 4.2). The part along 90°E is examined here. In the observations, the water temperature generally increases from the bottom (about $-0.3^\circ$C) to the surface, with the warmest water (above 12°C) located north of 42°S (Fig. 4.4a). In addition, a layer of cold water (down to $-0.8^\circ$C) can be
seen on the continental shelf above 500 m. This cold water extends from the continent to about 58°S at the depth of 100 m. In contrast, salinity shows a different structure south of 42°S, with fresher water at the surface and bottom, and more saline water in between (Fig. 4.4b). The most saline water (>35 psu) is found at the surface north of 42°S.

![Figure 4.4](image-url)

**Fig. 4.4.** Potential temperature (left) and salinity (right) in the (a, b) observations, (c, d) simulation without ISM, and (e, f) simulation with AISM along 18.

The different water masses are defined by salinity and temperature together in this area. The warm and saline North Atlantic deep water (NADW), with a temperature of 0–2°C and salinity 34.68–34.75 psu, comes from the north and transforms into the Antarctic circumpolar deep water (CDW) in the Southern Ocean, with the core of the maximum salinity ascending...
from the depth of 2500 m to 1000 m. In the Southern Ocean, the relatively thin layer of cold and fresh water at the surface is the Antarctic surface water (ASW), which transforms into the Antarctic intermediate water (AIW) further north [e.g., Aoki et al., 2005] with a temperature of 2–6°C and salinity of 34.4–34.68 psu in the observations.

The basic structures of potential temperature and salinity are captured by both simulations (Fig. 4.4c-f). However, there are obvious deficiencies in the simulation before inclusion of the AISM (Fig. 4.4c, d). First, the bottom water is too cold and too saline, with a temperature down to –0.9°C and salinity over 34.69 psu. In addition, the ASW is a little bit too warm and too thin. It is seen that inclusion of the AISM improves the simulations. The bottom water, in particular north of the seamount (around 60°S), becomes warmer and fresher, and the core of the maximum salinity of the NADW becomes more saline, both of which agree better with the observations (cf. Fig 4.4b, f). In addition, the ASW becomes colder with its volume agreeing much better with the observations.

A closer inspection of the simulated bottom water indicates that there are still deficiencies in the simulation with AISM, in particular, the simulated temperature. In the observation, 0°C isotherm is at the depth of 2500–4200 m north of the seamount, while 1500–3000 m south of the seamount (Fig. 4.4a). Before inclusion of AISM (Fig. 4.4c), although the 0°C isotherm is at the depth of 2500–3800 m north of the seamount, closer to the observation, it is too shallow south of the seamount, extending to 200 m depth. In contrast, addition of AISM deepens the 0°C isotherm a little bit north of the seamount, but too much south of it (Fig. 4.4e). North of the seamount, 0°C isotherm is at the depth of 3000–4000m, slightly deeper than in the observations. South of the seamount, it only spans a small depth, 2900–3100m, too deep compared with the observations. The unrealistic simulations south of the seamount with and without AISM are likely due to the topography here. Too cold water without AISM or the too warm water with AISM is trapped in the deep basin due to the big seamount in the south.

4.2.3. Section P16

Section P16 is along 150°W in the Ross Sea out of RIS (Fig. 4.2). The topography along this section is relatively flat compared with sections S4 and I8, with only the Pacific-Antarctic Ridge at about 60°S (Fig. 4.5). Observations (Fig. 4.5a, b) show that the temperature generally increases from the bottom to the surface, accompanied by the general decrease in the salinity. At the surface, there is a thin, cold and fresh surface water layer in the Southern Ocean. Near the bottom, the water temperature is below 0°C south of the ridge. Blocked by the ridge, this cold water cannot mix with the warm water north of the ridge, where the water temperature is about 0.8°C.
Fig. 4.5. Potential temperature (left) and salinity (right) in observation (a, b), in the simulation without ISM (c, d) and with AISM (e, f) along P16.

Due to the coarse resolution, the Pacific-Antarctic ridge in the model is strongly smoothed (Fig. 4.5c-f). This tends to weaken the blocking effect of the ridge, resulting in excessive mixing in the intermediate and deep waters between the two sides of the ridge. Consequently, the temperature and salinity differences in the bottom water are much smaller between the two sides of the ridge in both simulations compared with those in the observations. Therefore, the intermediate and deep waters are too warm south of the ridge while they are too cold north of it.

Addition of AISM improves the simulated bottom temperature and salinity in the mid-latitudes. For example, the benthic water without AISM has a temperature of 0.1–0.4°C, much colder than the observations (0.6–0.8°C). After inclusion of AISM, the water temperature increases to above 0.4°C, much closer to the observations. In addition, the thickness of the more saline water layer becomes larger at the bottom after inclusion of AISM, in much better agreement with the observations.

In the Southern Ocean, particularly close to the continental shelf, 1°C isotherm goes too deep and 2°C isotherm goes a little bit too south in the simulation with AISM. This may mean that the overall contribution from AISM is overestimated for this region.

4.2.4. Section P18

This section is along 105°W, close to Abbot Ice Shelf and Pine Island in the south (Fig. 4.2). Its hydrographic structure is similar to that of P16, with temperature increasing and salinity decreasing from the bottom to the surface, and the presence of cold and fresh surface water in the south (Fig. 4.6a, b). Although these general features are reasonably captured by the simulation without AISM, large differences are clearly visible (Fig. 4.6c, d). In general, in the
simulations the water is too cold in the Southern Ocean from the surface to the bottom. For example, 1°C isotherm extends to the continental shelf in the observations, but is limited to north of 67°S in the simulation. 0°C isotherm locates below 4500 m in the observation, but is around 3500 m in the simulation. Near the bottom, the water temperature is below –0.2°C in the simulation, but around 0°C in the observation. Another problem is that the surface water is a little bit too fresh in the simulation. Furthermore, more saline water (salinity >34.68 psu) should be located below 500 m, but is simulated to be below 1000 m.

Fig. 4.6. Potential temperature (left) and salinity (right) in the observation (a, b) and in the simulations without (c, d) and with (e, f) AISM along P18.

Inclusion of AISM significantly improves the results (Fig. 4.6e, f), coming to good agreement with the observations. The water becomes warmer in the whole Southern Ocean along this section, and the temperature structure agrees well with the observations. The 1°C
isotherm extends to the continental shelf and locates at the depth of 1500 m on the continental shelf as in the observations. The more saline water (salinity >34.68 psu) is found below 700 m depth, which is much closer to the observations. Thus the volume of the more saline water is comparable to the observations, although the water is still a little bit fresher.

Fig. 4.7. Potential temperature (left) and salinity (right) in the observation (a, b) and in the simulations without (c, d) and with (e, f) AISM along section A23.

4.2.5. Section A23

Section A23 is taken in the Weddell Sea along around 30°W, and across the South Georgia islands at about 55°S (Fig. 4.2 and Fig. 4.7a, b). In the observations, the water is generally colder and fresher south of the island than in the northern side (Fig. 4.7a, b). In the south, the water temperature is observed to be below 1°C in the Weddell Sea. The WDW has a core with the highest temperature of 0.8°C and maximum salinity of 34.695 psu at the depth of 400 m. This warm and saline water layer is enclosed by the colder and fresher surface and bottom.
waters. The surface water is the coldest water with a temperature down to –1.8°C due to sea ice formation and ice shelf-ocean interaction. The bottom water is also the coldest among all sections, with the temperature down to –0.9°C. North of the island, the warm, saline NADW intrudes from the north with a maximum salinity over 34.78 psu at the depth of 2000 m.

Inclusion of AISM improves the simulated hydrographic structure (Fig. 4.7c–f). Before inclusion of AISM, the WDW in the south was far from the continent and the water next to the continent at that depth was too cold and too fresh (Fig. 4.7c, d). The bottom water was too warm (–0.6°C) and too saline (34.68 psu). After inclusion of AISM (Fig. 4.7e, f), the simulated temperature and salinity structures are more comparable with the observed ones. The isotherms and isohalines in the intermediate layer obviously become doming as seen in the observation, and the WDW contacts the continent as observed. The bottom water becomes fresher, although still slightly more saline than the observations.

One problem here is the overestimation of the bottom water temperature, regardless whether with or without AISM (cf. Fig. 4.7a, c and d). This warm bias is common for ocean models with similar resolution and time scale (e.g., Gnanadesikan et al., 2006; Losch, 2008). A possible cause is the over-smoothed topography near the South Georgia Islands, which allows more warm and saline NADW to intrude into the area and thus increase the water temperature. Another possible cause is the advection of warm and saline water from the east, where the temperature of the bottom water is also overestimated as shown below along section S2. Finally, it is also possible that underestimation of the open water (leads or polynyas) formation during winter season (see Fig. 6.4c and d) plays a role, which would cause less brine rejection, thereby decreasing the source for bottom water formation and thus warming the bottom water.
4.2.6. Section S2

This section is along the 0° meridian, outside Fimbul ice shelf (Fig. 4.2). The observed hydrography is basically similar to that of section A23. The bottom water temperature is down to –0.825°C (Fig. 4.8a) and salinity is 34.65 psu (Fig. 4.8b). This section is across the mid-Atlantic Ridge at 50–55°S, which is very steep at about 55°S. This high and sharp topography has significant effect on the hydrographic structure (Fig. 4.8a, b). The warm and saline NADW is found at the depth of about 500m after overflowing this steep topography and remains then around that depth as WDW.

Similar to the simulation of section A23, the addition of AISM improves the hydrographic structures. The isotherms and isohalines become horizontal, and the bottom water becomes much fresher, in good agreement with the observations (Fig. 4.8). Similar to section A23, the water temperature in the Southern Ocean (south of 55°S) is overestimated regardless whether with or without AISM. Furthermore, the salinity field in the Southern Ocean is also overestimated in both simulations. On the whole, the volume of WDW in this region is highly overestimated. The reason for such overestimations is most probably due to the coarse model resolution, particularly the highly smoothed topography. As can be seen in Fig. 4.8, the high and sharp features of the mid-Atlantic Ridge are almost fully missing in the model. This allows more warm and saline NADW to the region, forming a larger volume of WDW. Consequently, the deep and bottom waters become much warmer, as clearly seen in both simulations. Such overestimation of warm water would cause substantially larger heat and freshwater fluxes from the ice shelf-ocean interface, as evident from the large freshwater flux from Fimbul Ice Shelf in our simulation.

Fimbul Ice Shelf is the largest ice shelf in the southeast Weddell Sea. It is located on a narrow, at most 40 km wide, continental shelf, and its floating Trolltunga ice tongue at 1°W overshoots the shelf break into deep water (Nøst, 2004). The narrow continental shelf and the close proximity of relatively warm waters may produce high basal melt rate and therefore a significant freshwater flux. However, there are contradicting views whether WDW (the heat source of ISM) has direct access to the sub-ice shelf cavity. Nøst (2004) concluded that WDW does not have direct access to the sub-ice shelf cavity, based on his seismic survey topographic data. His statement was supported by Nicholls et al. (2006) from an autonomous underwater vehicle, and later by Price et al. (2008) from a ship-based hydrographic survey. However, Price et al. (2008) did not exclude the possibility of warmer water episodically entering the shelf cavity. In their model runs, Smedsrud et al. (2006) demonstrated that WDW enters the main cavity of the ice shelf, when the easterly wind weakens sufficiently. This idea
is supported by Nicholls et al. (2008) with model runs and CTD data at the ice front and at a single access hole through the ice shelf. They found warm water episodically flushing into the under-ice cavity. Recently, Nøst et al. (2011) showed with an idealized numerical model that mesoscale eddies advect WDW onto the shelf and into the sub-ice shelf cavity.

Intrusion of warm water into the sub-ice shelf cavity of Fimbul gives rise to a maximum melt rate of 10 m year\(^{-1}\), with an average of 1.9 m year\(^{-1}\) (Smedsrud et al., 2006), corresponding to freshwater fluxes of 9.5 mSv and 1.8 mSv, respectively. Other estimates from numerical models (e.g., Hellmer, 2004; Beckmann et al., 1999) are basically in agreement with the upper limit. The estimate of 10 \(\pm\) 3 mSv obtained by Walkden et al. (2009) at 2.8ºW is also close to the upper limit. The estimate from our simulation is 19.8 mSv, approximately twice the other estimates. This large freshwater flux is likely due to the over-flattened ridges (as shown by section S2), which permits substantial warm WDW intruding onto the continental shelf (Fig. 4.10). As can be seen, the observed water temperature on the continental shelf at the depth of 200–600 m is about 0°C (Fig. 4.10a), whereas the corresponding simulated water temperature is about 0.9°C (Fig. 4.10e). Such high temperature would cause an extremely high melt rate for Fimbul Ice Shelf.

4.2.7. Section I6

Section I6 is along 30ºE (Fig. 4.2). The hydrographic structure of this section (Fig. 4.9) is similar to that of I8 (Fig. 4.4). The coldest and freshest water is ASW with a temperature down to –1.8°C and salinity of 33.8 psu. Besides ASW, the water temperature generally increases from the bottom to the surface, with the bottom water down to –0.7°C and surface water up to 20°C north of 40ºS (Fig 4.9a). The most saline water is found in the intermediate depths, enclosed by fresher surface and bottom waters (Fig. 4.9b). The bottom water has a salinity of about 34.65 psu.

![Images of hydrographic sections](a) (b) (c) (d)
Fig. 4.9. Potential temperature (left) and salinity (right) in the observation (a, b) and in the simulations without (c, d) and with (e, f) AISM along section I6.

Both simulations capture the basic temperature structure as in the observations. However, there are deficiencies in both simulations, with and without AISM. Before inclusion of AISM (Fig. 4.9c), the 0°C isotherm is as in the observations (Fig. 4.9a), whereas WDW is generally much colder than the observations. Introduction of AISM increases WDW water temperature and makes its structure closer to the observations, whereas the 0°C isotherm (lower limb) is located too deep (at depth 3000m; about 1800 m in observation), indicating more WDW in the simulation (Fig. 4.9e). The bottom waters in both simulations are slightly warmer than the observations.

The salinity structure is much better represented in the simulation with AISM than without AISM. The simulation without AISM misses the fresh bottom water (Fig. 4.9d). This is modified after inclusion of the AISM (Fig. 4.9f), although the salinity is still slightly more saline than the observations. In addition, inclusion of AISM increases the salinity of the core of NADW, which agrees much better with the observation.

4.3. Discussions

The circumpolar freshwater flux from AISM is a result of the interaction between the ice shelves and the Southern Ocean. It will therefore be subject to spatial and temporal variations due to changes in the ice shelves and the Southern Ocean hydrography. Owing to its significant impact on the Antarctic Ice Sheet mass balance, the global sea level rise and the global thermohaline circulation, an accurate estimate of this freshwater flux is critical to the understanding of ice shelf-ocean interaction and projection of the influence of ISM to global climate.

However, the circumpolar freshwater flux from AISM is difficult to be accurately estimated. First, measurements or numerical simulations are not done for all Antarctic ice shelves (see Table 4.2). Second, direct measurements are still very limited due to difficulties in accessing the sub-ice shelf cavity. Third, thus, our knowledge of the melt rate of ice shelves is mostly from estimations based on glaciological and oceanographic observations or numerical simulations (see section 1.4). These approaches may yield large differences due to the sampling regime (Payne et al., 2007), the accuracy of velocity data (Joughin and Padman, 2003), the used ice thickness data (Yu et al., 2010), the assumption of steady-state ice sheet (Shepherd et al., 2004; Horwath et al., 2006), the selection of outflow depth (Payne et al., 2007), and the south limit of the sub-ice shelf cavity (Galton-Fenzi et al., 2008).

The first estimate of the circumpolar melt rate was made by Jacobs et al. (1992), being about 18.8 mSv (1 mSv = 10^3 m^3 s^-1), based on a combination of physical and geochemical observations of the meltwater and glaciological field measurements. In their estimate,
different melt rates were estimated for Filchner-Ronne, Ross, Amery and George VI Ice Shelves, but the melt rate for the undifferentiated ice shelves was only roughly determined to be 0.30 m year\(^{-1}\) according to the basal mass balance within 100 km of the calving front of RIS.

The lack of *in-situ* observations, especially for the undifferentiated ice shelves would induce large uncertainties in the estimation of the circumpolar AISM freshwater flux. This was improved by an oceanographic survey in early 1994 for the southeast Pacific ice shelves, which revealed that the melt rate of the undifferentiated ice shelves had been highly underestimated (Jacobs et al., 1996). The melt rates were estimated to be 10 m year\(^{-1}\) for the Pine Island Glacier and 2 m year\(^{-1}\) for the others. This results in an updated circumpolar AISM of 26.1 mSv, which was considered still conservative, as some estimates of the past attrition were restricted to outflows with *in-situ* temperature below the sea surface freezing point (Jacobs et al., 1996). Later estimates, based on glaciological measurements (Rignot, 1998; Shepherd et al., 2004), suggest that the melt rate of the Pine Island Glacier may even be double the level assumed (Table 2). Similar increase may also be possible for the other undifferentiated ice shelves due to the restrictions in the estimates of Jacobs et al. (1996), although the magnitudes are subject to change for particular ice shelves.

The estimated circumpolar AISM based on model simulations has been mainly performed with the regional BRIOS group models of the Southern Ocean. Beckmann et al. (1999) were the first to explicitly include the ice shelves in their model. Through considering Filchner-Ronne, Ross, Amery, Larsen, Eastern Weddell (EW), Ekström and Fimbul Ice Shelves, they simulated the circumpolar AISM to be 31.6 mSv. Later estimates (Beckmann and Timmermann, 2001; Timmermann et al., 2002) obtained a very close circumpolar AISM flux, about 29.6 mSv and 29.47 mSv, respectively, since they considered the same ice shelves as Beckmann et al. (1999). In addition to the major ice shelves mentioned above (except Ekström), Hellmer (2004) considered more major ice shelves, including Abbot, George VI, Getz, and Shackleton Ice Shelves. He obtained a circumpolar ISM flux of 28.42 mSv. Nevertheless, Hellmer (2004) indicated that his estimate might be biased up since the freshwater fluxes from EW and Fimbul Ice Shelves were possibly overestimated due to the coarse resolution of his model. On the other hand, his estimate might be further increased with even smaller ice shelves added (Hellmer, 2004).

The estimated circumpolar freshwater flux is roughly 30 mSv according to the studies mentioned above. This value could be much higher or lower with more knowledge of the ice shelves. For example, with the grounding line further southward, the melt rate could be much higher since melting dominates near the grounding line. Amery Ice Shelf is such a case: the south limit of the sub-ice shelf cavity is regularly defined at 71.6° S (e.g., William et al., 2001), but it is found to be at 73.6° S with the technique development by Galton-Fenzi et al. (2008).

In this study, the simulated total circumpolar ISM flux is about 106 mSv, which is much higher than the other estimates mentioned above. However, a closer inspection of the individual ISM reveals that a large part of the estimates are comparable, particularly for the major ice shelves (Table 2). For example, the freshwater flux from FRIS is 6.52 mSv in this study, in good agreement with the earlier glaciological shelf-wide estimate (Jacobs et al., 1996) and the numerical results from cavity-resolving models (e.g., Timmermann et al., 2002; BG03). The freshwater fluxes from the other large ice shelves, e.g. Ross (8.36 mSv) and Amery (5.44 mSv), as well as EW (4.89 mSv) in this study are also comparable to the other studies from regional cavity-resolving models (e.g., Beckmann et al., 1999; Timmermann and Beckmann, 2002; Hellmer, 2004). This indicates that the applied parameterization generally works well for majority of the major ice shelves.
### Table 4.2. Freshwater flux from ISM in different studies.

<table>
<thead>
<tr>
<th>Name of ice shelf</th>
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<th>Observations</th>
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</tr>
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<tbody>
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<td></td>
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<td>J96</td>
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<td>Filchner-Ronne</td>
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<td></td>
</tr>
<tr>
<td>George VI</td>
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<td>1.52</td>
<td></td>
</tr>
<tr>
<td>Pine Island</td>
<td>0.97</td>
<td>2.28±0.38</td>
<td></td>
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<tr>
<td>Total</td>
<td>18.8</td>
<td>26.1</td>
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Note: J92-Jacobs et al. (1992); J96-Jacobs et al. (1996); R98-Rignot (1998); S04-Shepherd et al. (2004); B99-Beckmann et al. (1999); BT01-Beckmann and Timmermann (2001); T02-Timmermann et al. (2002); BG03-Beckmann and Goosse (2003); H04-Hellmer (2004); TH06-Thoma et al. (2006); P07-Payne et al. (2007); JJ08-Jenkins and Jacobs (2008).
Our large circumpolar AISM flux is mainly because we include more ice shelves, not only those major ice shelves as in Hellmer (2004) but also the other major ice shelves, including West and Wilkins, and a number of minor ice shelves, such as Lazarev, Voyeykov, Cook and Dotson. There have been few studies on these ice shelves. Therefore the reliability of these estimates would need further verification. The estimated freshwater flux from West and Wilkins Ice Shelves are 4.54 mSv and 6.67 mSv, respectively, close to the other major ice shelves (Table 2). The high melt rate in the Amundsen and Bellingshausen Seas is generally consistent with the available estimates. For example, the freshwater flux from George VI Ice Shelf is 5.35 mSv in this study, corresponding to a melt rate of 3 m year$^{-1}$, which is close to the recent measurement of 3–5 m year$^{-1}$ (Jenkins and Jacobs, 2008), although higher than the previous estimates of 1.9 m year$^{-1}$ (Jacobs et al., 1996) and 0.43 m year$^{-1}$ (Hellmer, 2004). The Pine Island Glacier is experiencing high melt rate due to the warm CDW flowing onto the continental shelf. Its estimated melt rates are in a wide range, from 10 m year$^{-1}$ (Jacobs et al., 1996) to 24±4 m year$^{-1}$ (Rignot, 1998). The freshwater flux from our simulation is 1.62 mSv, corresponding to a melt rate of 21 m year$^{-1}$, close to the simulation of 20.6 m year$^{-1}$ by Payne et al. (2007) with a plume model.

The large freshwater fluxes from Fimbul (19.80 mSv), Getz (16.4 mSv), Shackleton (8.56 mSv) and Larsen (6.33 mSv) Ice Shelves are likely overestimates. The estimated freshwater flux from Fimbul is higher than the previous estimates (e.g. Beckmann et al., 1999; Beckmann and Timmermann, 2001; Timmermann et al., 2002; BG03; Hellmer, 2004; Walkden et al., 2009) by a factor of roughly two, due to the warm bias in our simulation. Similarly, the estimated freshwater flux from Getz is also more than twice of the estimates through glaciological measurement (Shepherd et al., 2004). The reason for this difference needs further investigation. The model estimate in Hellmer (2004) is much lower than the others, perhaps due to the lack of detailed topographical data under Getz Ice Shelf. The freshwater flux from Larsen Ice Shelf is generally 3–5 times the previous model estimates of 1.2–2.1 mSv (e.g. Beckmann et al., 1999; Beckmann and Timmermann, 2001; Timmermann et al., 2002; Hellmer, 2004). The freshwater flux from Shackleton Ice Shelf is also much larger than the model estimate in Hellmer (2004). The freshwater flux from these four ice shelves is roughly overestimated by 30 mSv in total, which would suggest a reasonable estimate for the entire circumpolar AISM flux of about 60–70 mSv.

The different performances of the parameterization in estimating the freshwater flux indicates that a uniform characteristic scale $L_{eff}$ may be not valid for all ice shelves. In order to successfully model the freshwater flux, $L_{eff}$ should be changed for some ice shelves. However, specific $L_{eff}$ remains to be determined for a large number of ice shelves due to the largely unknown sub-ice shelf cavities. In addition, due to the over-flattened topography in the coarse resolution climate model, changing of $L_{eff}$ might not significantly improve the overall simulation of the Southern Ocean hydrography. This has been shown in the simulations with and without AISM for sections A23 and S2, since the simulation without AISM is actually the case of $L_{eff} = 0$. Further increasing the model resolution is therefore highly desirable; otherwise a special treatment must be performed for the topography.

4.4. Conclusions

The AISM removes heat from and injects freshwater into the adjacent Southern Ocean, possessing a significant role in the global climate system. It is one of the most important components currently absent in the IPCC climate models. The main purpose of this chapter was to assess the capability of the ISM parameterization, which was initially proposed by BG03 and later implemented in a global sea ice-ocean climate model (ORCA2-LIM) by Wang and Beckmann (2007), in simulating the ice shelf-ocean interaction for climate studies. For this purpose, we have performed detailed analyses of the simulations with and without AISM, compared the simulated Southern Ocean hydrography with the WOCE observations,
and compared the simulated circumpolar AISM freshwater flux with the previous estimates based on in-situ measurements and finer resolution regional models.

Comparisons with WOCE observations along a series of WOCE sections demonstrated that the addition of AISM significantly improved the simulated hydrography. It not only warms and freshe the originally too cold and too saline bottom water, but also warms and enriches the salinity of the originally too cold and too fresh WDW. The addition of AISM also enhances the topographic effect on the hydrographic structure. The resulting hydrographic structure of the Southern Ocean agrees much better with the observations after inclusion of AISM. Furthermore, most of the simulated individual ISM freshwater fluxes are generally comparable with the previous estimates based on in-situ measurements and high-resolution regional models. This overall suggests that the applied AISM parameterization is an acceptable way to include the ice shelf-ocean interaction in global sea ice-ocean models.

Representation of the bathymetry is important for the simulation of water masses in the Southern Ocean. The highly smoothed ridges in coarse climate models may result in overestimation of temperature and salinity. This is particularly evident in the Weddell Sea, where inclusion of AISM only improves the simulated salinity structure. The overall overestimation of the water temperature remains, regardless whether with or without AISM, accompanied by overestimation of ISM freshwater flux from Fimbul Ice Shelf. The most probable cause is the over-smoothed topography, particularly near the South Georgia Island and the mid-Atlantic Ridge, which allows more warm and saline NADW to the area and thus increasing the water temperature there.

The circumpolar freshwater flux needs further verification with more knowledge of the ice shelves. In-situ measurements and high-resolution regional modeling studies are highly desirable to further investigate the ice shelf-ocean interaction for the minor ice shelves, in particular in the Amundsen and Bellingshausen Seas. The large contribution of these ice shelves to the freshwater flux suggests that they may play a significant role in the oceanography of the Southern Ocean and also in the global climate.
Chapter 5 Impact of Increasing Antarctic Ice Shelf Melting on the Southern Ocean Hydrography and Circulation

In previous chapter, we have shown through detailed comparisons with the observations that inclusion of AISM significantly improves the simulated hydrography in the Southern Ocean. In this chapter, we will further investigate the impact of increasing AISM on the Southern Ocean. There is evidence for increasing AISM, particularly in the Bellingshausen, Amundsen and Ross Seas (e.g. Rignot and Jacobs, 2002; Shepherd et al., 2004). However, determination of the exact AISM (spatial distribution and temporal evolution) still remains extremely difficult. Therefore, in this chapter our focus is to identify the qualitative effect of increasing AISM on the Southern Ocean hydrography. This is achieved through a detailed analysis of the twin experiments with and without AISM which have partly been shown in the previous chapter. In principle, the simulation without ISM provides an extreme case for the situation of less AISM. Thus, the differences between the simulations with and without AISM provide a first idea regarding the changes induced by increasing AISM. In order to verify the reliability of the impact of increasing AISM on the Southern Ocean, we compare these differences with a series of observations dealing with the trend of the Southern Ocean temperature, salinity and density in several depths. Also illustrated in this chapter are the impacts of increasing AISM on the mixed layer depth (MLD) and on the circulation of the Southern Ocean, followed by the temporal variability of freshwater flux from the AISM and its connection with the water temperature.

5.1. Impact of increasing AISM on the Southern Ocean temperature, salinity and density: climatological annual mean

5.1.1. Impact of the AISM on temperature

As a result of increased AISM, the Southern Ocean water below about 1000 m unanimously becomes warmer (Fig. 5.1). However, the warming is spatially inhomogeneous. At the depth 1033 m (Fig. 5.1a), the most pronounced warming is found in the Weddell Sea, up to 1.2°C, and near the continent, generally more than 0.6°C. In other areas, the increase in temperature is much smaller, generally less than 0.4°C. Down to the bottom, the warming becomes weaker. At the same time, the warming pattern also changes. For example, at the depth of 3725 m (Fig. 5.1b), the largest warming (over 0.6°C) takes place in the Australia-Antarctic Basin in the Indian Ocean, not next to the continent as in shallower layers.

These simulated changes are consistent with the long-term in-situ observations. The warming of the deep and bottom waters in the Southern Ocean over the past half century has been reported by a number of recent studies. Fahrbach et al. (1998) observed that the bottom water in the central Weddell Sea became warmer during the interval of 1989–1995. This warming trend was shown to affect most of the Weddell Sea when the data were extended to the early 2000s (Fahrbach et al., 2004). Observations dating back to 1912 indicated that the warming of the Weddell Sea WDW had already occurred since 1970s (Robertson et al., 2002). The warming of the bottom water was also observed in the Indian sector (Whitworth, 2002; Johnson et al., 2008) and the Pacific sector (Ozaki et al., 2009) of the Southern Ocean. In particular, Ozaki et al. (2009) revealed a long-term warming of the bottom layer of the Ross Sea from 1969 to 2004. Johnson et al. (2008) reported warming of the abyssal waters in the eastern Indian Ocean between 1994/95 and 2007. As shown in Fig. 5.1, these warming trends are consistent with the effects of increased AISM.
The impact of increasing AISM is more complex above 1000 m. The water in the vicinity of the Antarctic ice shelves generally becomes colder (Fig. 5.2), due to the cold freshwater from the ISM. The cooling is most significant at the depth of 216 m (Fig. 5.2c), the base of most ice shelves’ edges in the model (except FRIS and RIS), where the heat and freshwater fluxes due to AISM are applied. Substantial cooling at this level is found in the coastal areas of the Amundsen, Weddell and Ross Seas, down to about -2.2°C, -1.5°C and -1.25°C, respectively. The smaller ice shelves in East Antarctica tend to cause much smaller temperature changes in the vicinity of the continent, generally less than 0.2°C (except around the region of Amery Ice Shelf).

The coastal cooling diminishes further down to 512 m (Fig. 5.2d), where cooling only occurs in the vicinity of the much thicker ice shelves FRIS and RIS. Closer to the surface, the coastal cooling affects a much larger area (Fig. 5.2a and b). At the depth of 55 m (Fig. 5.2b), the cooling becomes weaker but broader next to the continent, as a result of the rising cold buoyancy freshwater from the melted ice shelves. At the surface (Fig. 5.2a), the temperature change is very limited, implying a rapid mixing between the colder ice shelf water and the warmer CDW, and a fast exchange with the atmosphere. The most pronounced cooling at the surface is not next to the continent, but in the ACC region north of the Weddell Sea (down to –1.0°C). The cooling next to the continent is mostly less than 0.4°C except in two regions. One is outside of Amery Ice Shelf, and the other is close to the Cape Adare in the western Ross Sea, both down to –0.8°C.

In addition to the limited cooling shown above, the water is undergoing significant warming over most of the Southern Ocean above the depth of 1000 m due to increasing ISM (Figs. 5.1 and 5.2). This is because the rising of cold melted ice shelf water attracts more CDW to compensate the water masses (Orsi et al., 1993; Schröder and Fahrbach, 1999). The warming is generally not over 1.0°C, except in the Amundsen and Bellingshausen Seas and the eastern Weddell Sea due to the large cooling in the vicinity of the ice shelves. The largest warming occurs at the depth of 216 m (Fig. 5.2c), up to 1.2°C. Large warming is also seen in the Atlantic and Pacific sectors of the ACC regions. Such warming is consistent with a southward shift of the CDW, as inferred from recent observations (e.g. Swift, 1995). Further downward the magnitude of warming tends to decrease but the warming area tends to be larger (Fig. 5.1 and Fig. 5.2d). Since the WDW is the major source of bottom water, increase in the WDW upwelling rate would cause warming of the bottom water (Fahrbach et al., 2004). This is consistent with the simulated bottom water warming shown in Fig. 5.1b.
Fig. 5.2. Climatology annual mean differences (with-without) of temperature (ºC) at the depth of (a) 5 m, (b) 55 m, (c) 216 m and (d) 512 m.

The simulated warming pattern is consistent with the recent trend studies of the world ocean, particularly in the Southern Ocean, suggesting that these observed warming may largely be caused by the enhanced AISM. Gille (2002) reported the warming of Southern Ocean waters at 700–1100 m depth from the 1950s to the 1990s. Most warming occurred in the ACC region between 45°S and 60°S, especially in the Atlantic and Indian sectors. This is almost identical to the warming pattern shown above as a result of increasing AISM (Fig. 5.1a and Fig. 5.2d). Robertson et al. (2002) reported recent warming of about 0.38 ºC at the depth of WDW in the southeast and northwest Weddell Gyre. Again, our simulation show similar pattern due to increasing AISM (Fig. 5.1 and 5.2d). Aoki et al. (2003) showed another example at depths of 200–900 m around 30°E–150°E in the ACC region, where warming over three decades suggested that the Southern ACC Front had shifted southward. A similar warming is produced by our model. Levitus et al. (2005) reported changes in ocean heat content vs. time for 0–300, 0–700 and 0–3000 m integrations, finding that more than half of the increased heat content was from the Atlantic sector. They also noted the cooling above ca 600 m and between 1100 m and 1450 m at high-southern latitudes, which are the formation regions of the shelf and bottom waters, and the warming between those depths, where CDW reaches the Antarctic continental margin. The cooling above ca 600 m is reproduced near the coastal regions (Fig. 5.2), whereas the cooling below ca 1100 m is not
reproduced by the simulation, possibly due to increasing AISM in the cavities of the FRIS and RIS not resolved in the present model.

### 5.1.2. Impact of increasing AISM on salinity

The impact of increasing AISM on the Southern Ocean salinity on different levels is shown in Fig. 5.3. AISM inputs freshwater into the adjacent ocean. As for temperature, the largest effects are seen close to 216 m (Fig. 5.3c). The largest freshening is in the vicinity of the continent (up to –1 psu), where most of the ice shelf-ocean interaction was applied in the model. In addition, freshening is dominant over the whole Southern Ocean at this depth, with patches of slightly higher salinity (up to 0.1 psu) in the central Kerguelen Gyre, the central Weddell Sea and the Ross Sea, since the freshwater from AISM is transported with the ocean current.

At the surface, the freshening is substantial (Fig. 5.3a, b). This is because the fresh subsurface water rises to the surface due to its small density (Fig. 7.15). The freshening at the surface is particularly remarkable in the Ross and Weddell Seas (down to –0.5 psu), the Amundsen and Bellingshausen Seas (down to –0.4 psu), and the Prydz Bay (down to –0.3 psu). In the Indian Ocean and western Weddell Sea, the freshening is generally much weaker. In the region east of 45°E, the water even becomes more saline, up to 0.2 psu. In 55 m depth, almost all of the Southern Ocean becomes fresher (Fig. 5.3b). Unlike at the surface, the freshening is more pronounced next to the continent at this depth. Aside from the general freshening, small areas of more saline water up to 0.1 psu are found around the 45°E meridian and in the southern Ross Sea.

Downward to the depths of 512 and 1033 m, increasing AISM generally causes higher salinity in the Southern Ocean (Fig. 5.3d–e). Freshening is only found in the Weddell Sea and Ross Sea close to the continent, where the decrease in salinity is up to 0.3 psu. At the bottom (Fig. 5.3f), the major part of the Southern Ocean shows a universal trend of freshening due to the increasing AISM, although the changes are all very small (< 0.02 psu).

The large freshening in the surface water (Fig. 5.3a, b) is supported by recent observations. Wong et al. (1999) found a large-scale freshening of intermediate waters in the Pacific and Indian Oceans, and partly attributed this to surface water freshening in the source region of the Southern Ocean. Their estimates showed that although the freshwater gain in the source region of AAIW corresponded to the freshening in the entire South Pacific, the increase in precipitation could only explain about 1/3 of this freshwater gain. It is highly likely that the extra freshwater is from the increase in AISM. The simulated large salinity decrease in the vicinity of ice shelves is not inconsistent with the summer observations near RIS by Jacobs et al. (2002), who found a rapid freshening in the surface water in the Ross Gyre and much deeper close to the ice shelf edge. Jacobs et al. (2002) attributed this freshening to the increasing AISM in the Bellingshausen and Amundsen Seas, which is found also in later in-situ observations (e.g. Shepherd et al., 2004).

The spatial extent of the surface freshening down to 500 m between the Pacific, Atlantic and Indian sectors is generally consistent with the observed linear trend of salinity change between the 1950s and 1990s (Boyer et al., 2005). It is seen from the observations that strong freshening occurs in the Pacific and Atlantic sectors of the high-latitude Southern Ocean, with the most significant salinity decrease in the top surface waters. In the Indian sector of the Southern Ocean, however, it is seen that the overall salinity is slightly increasing (Boyer et al., 2005). This pattern is qualitatively depicted in the present study (Fig. 5.3), where the freshening in the Indian sector is much weaker than in the Pacific and Atlantic sectors. Consistent with the recent observations by Rintoul (2007), Johnson et al. (2008) and Ozaki et al. (2009), increasing AISM would cause freshening of the bottom water in the Southern Ocean (Fig. 5.1b and 5.3f). Such overall consistency presented in this study has not yet been simulated in any previous study.
Fig. 5.3. Climatology annual mean differences (with-without) of salinity at the depth of (a) 5 m, (b) 55 m, (c) 216 m, (d) 512 m, (e) 1033 m and (f) 3725 m.

5.1.3. Impact of increasing AISM on density

In general, density is determined by both temperature and salinity. However, in the cold Southern Ocean, density is mainly controlled by salinity. Therefore, decrease in salinity
results in decrease in density. Figure 5.4 shows the density changes due to increasing AISM. The general patterns of density change are similar to those of salinity at most of the vertical levels, e.g. at the depth of 5 m, 216 m and 3752 m (Fig. 5.4a, b, & d), where the density is reduced. The largest decrease in density (> 0.2 kg/m$^3$) is found at the surface (Fig. 5.4a), particularly in the Weddell, Bellingshausen, Amundsen and Ross Seas. Below the surface layer, a large decrease in density is mainly confined to the continent margins down to the depth of 216 m. At the depth of 1033 m, the water has a slight decrease in density in a fairly large area (Fig. 5.4c), where both temperature (Fig. 5.1a) and salinity (Fig. 5.3e) increase due to the increasing AISM at this level. Here, the increase in temperature plays the dominant role in the density change. At the depth of 3752 m the decrease is generally less than 0.1 kg/m$^3$.

The simulated density decrease in the bottom water is consistent with the recent observations (Withworth, 2002; Aoki et al., 2005; Rintoul, 2007; Johnson et al., 2008; Ozaki et al., 2009). Due to the rapid freshening of the bottom water, the decrease in density is also significant, resulting in loss of the densest water (e.g. Johnson et al., 2008). Additionally, warming of the intermediate water would also cause decrease in density. Such changes may reduce the stability of the water column, enhancing eddy formation due to baroclinic instability (e.g. Meredith and Hogg, 2006; Screen et al., 2009).

**Fig. 5.4.** Climatology annual mean differences (with-without) of density at the depth of (a) 5 m, (b) 216 m, (c) 1033 m and (d) 3725 m.
5.2. Seasonal variability of temperature and salinity changes due to increasing AISM

The seasonal variability of the impact of increasing AISM on the Southern Ocean temperature and salinity can be clearly seen in the mixed layer (Wang and Beckmann, 2007). Figure 5.5 shows the summer and winter water temperature and salinity differences in the Southern Ocean. There is little change in temperature in the Weddell Sea, the Amundsen and Bellingshausen Sea, and the Ross Sea in winter (Fig. 5.5b), but in summer, there is significant decrease in temperature in these regions (Fig. 5.5a). In contrast, the temperature in the ACC region decreases dramatically in winter, but shows small changes in summer (cf. Fig. 5.5a and b). The different seasonal variability of the temperature change in the ACC region and south of it is mostly related to the presence of sea ice in the Southern Ocean. The Antarctic Marginal Seas are sea ice covered in winter, and the seawater temperature is kept at or close to the freezing point. Thus, the additional ice shelf water would have little impact on the temperature of the surface water. However, during summer, most of the Southern Ocean is sea ice free. Thus, the rising cold glacial water which is advectively or diffusively brought to the surface can decrease the sea surface water temperature. It spreads northward...
with the surface currents. With the spreading, its influence becomes smaller. Therefore, its cooling effect is small in the ACC region in summer.

Unlike the temperature change influenced by the sea ice cover, the general pattern of the salinity change is quite similar in different seasons (cf. Fig. 5.5c and d). The spatial distribution of the salinity change thus provides a better description of the areas directly affected by the increasing AISM. The larger change in summer is due to the higher AISM in summer.

5.3. Impact of increasing AISM on mixed layer depth

The surface mixed layer of the ocean is in constant contact with the atmosphere. High levels of vertical turbulence due to the combined action of wind and buoyancy lead to vertical quasi-homogeneous distribution of physical properties (temperature, salinity, and density) throughout the layer, accompanied by sharp gradients in properties at the base of the layer. The depth of this layer is the so-called MLD. To define the MLD, a number of criteria have been proposed in the literature; they can be classified in two groups: difference-based criteria and gradient-based criteria (Dong et al., 2008). The former defines the MLD as a depth where the oceanic property changes a certain amount from a reference surface value (e.g., Kara et al., 2000), while the latter defines MLD as a depth where the vertical gradient of the property equals or exceeds a threshold value (e.g. Lorbacher et al., 2006).

The potential density is one of the most widely used oceanic properties for determination of the MLD, particularly for the difference-based criteria. A variety of thresholds have been adopted for the potential density difference, ranging from 0.005 kg/m$^3$ to 0.03 kg/m$^3$ (e.g. Monterey and Levitus, 1997; de Boyer Montégut et al., 2004). In this study, we use potential density difference criteria with a threshold of 0.01 kg/m$^3$ to determine the MLD. It is noted that there might be noticeable differences in the resulting MLD when applying different criteria.

de Boyer Montégut et al. (2004) constructed a global MLD climatology from hydrographic profiles, but a detailed comparison is difficult because MLD observations in the Southern Ocean are relatively sparse. It is therefore encouraging that the simulated MLDs (Fig. 5.6) in the Southern Ocean are similar to those presented in de Boyer Montégut et al. (2004). As expected, the mixed layer is, in general, shallower in summer and deeper in winter. This is particular evident on the Antarctic continental shelves and in the ACC region. On the continental shelves, brine rejection during sea ice formation destabilizes the water column, thereby helping to deepen the mixed layer. A comparison with the monthly mean MLD maps in de Boyer Montégut et al. (2004, Fig. 5) shows an overall good agreement. The mixed layer becomes deep (more than 100 m) from March through October. The maximum MLD is up to 500 m in our simulation, again in good agreement with de Boyer Montégut et al. (2004). However, deep mixing is not well captured in November due to the faster decay of sea ice in our simulation than in observations (see chapter 6). In January and February, vigorous mixing in the western Weddell Sea is missing in our simulation due to the underestimated sea ice production in this region.

Another region of strong vertical mixing in the Southern Ocean is around the ACC (Fig. 5.6). Our simulated MLDs in this region are in good agreement with the MLDs of Dong et al. (2008; Sallée et al., 2010) derived from Argo float profiles. The spatial structures are similar in each month, with deep mixed layer just within and north of the ACC in the Indian and Pacific Oceans. The deepest mixed layer in the simulation is up to 600 m and found from June to September, largely matching the occurrence time of the derived MLDs of Dong et al. (2008). The simulated area of the deepest mixed layer is largest in August in the Pacific Ocean, similar to that derived by de Boyer Montégut et al. (2004). The simulated mixed layer is shallowest in summer, about 40 m, closer to those with less than 70 m from Rintoul and Trull (2001). However, the simulated MLD is only about 60 m in spring, much
shallower than the value of Dong et al. (2000). This big difference might be partly because we adopted a smaller (0.01 kg/m$^3$) threshold of potential density difference. Aside from the deep mixing regions, MLDs are generally not over 30 m in the Southern Ocean (de Boyer Montégut et al., 2004).

![Fig. 5.6. Simulated climatologic monthly MLDs (m) with AISM in February, May, August and November.](image)

Increasing AISM mainly influences the spatial distribution of deep mixed layers in the Southern Ocean (Fig. 5.7). It tends to deepen the mixed layer on the continental shelves, and to shoal them in the Atlantic and Pacific sectors of the ACC region (Wang and Beckmann, 2007). The changes in MLDs are small in summer, usually less than 10 m, while much larger in winter, more than 100 m. The different changes of MLDs on the continental shelves and in the ACC regions might be explained as follows. As already seen in section 5.1.3, inclusion of the ice shelf-ocean interaction generally results in density decrease in the upper ocean (above 1000 m). But the decrease is different on the continental shelves compared with that around the ACC region. On the continental shelves, it is smaller at the surface but larger at depth, while the signature is opposite in the ACC region. These changes tend to destabilize the water column on the continental shelves, but make the water column more stable in the ACC region. Therefore, increasing AISM makes the mixed layer become deeper on the continental shelves while shallower in the ACC regions (Fig. 5.7). As a result, the
simulated shallower MLDs in the ACC region agree better with the observations than the simulations without AISM when comparing with the MLDs derived by Dong et al., (2008).

**5.4. Impact on the circulation in the Southern Ocean**

**5.4.1. Ocean circulation in the Southern Ocean**

Figure 5.8 illustrates the modeled climatological annual mean of the large-scale ocean circulation in the Southern Ocean with AISM. We can clearly see the eastward ACC current and the westward Antarctic Coastal Current (ACoC) south of 65ºS at the surface (Fig. 5.8a), together with the Weddell Gyre, the Kerguelen Gyre, and the Ross Gyre. The ACC dominates the picture and is shown as a broad band, which is typical for models at this resolution. Flow becomes slower with depth. The velocity of the current is about 1 cm/s near the bottom about 4750 m deep (Fig. 5.8b). At this depth, flow is still eastward in the ACC region.

**5.4.2. Changes in ocean circulation due to increasing AISM**

Increasing AISM affects the ocean circulation in the Southern Ocean through changes in hydrography (Fig. 5.9). The impact is generally coherent in the vertical direction, although stronger at the surface and becoming weaker with depth. Near the bottom at the depth of 4750 m, the influence is quite small, generally one orders of magnitude smaller than at the surface (Fig. 5.9).

**5.4.2.1. The ACoC**

The most significant impact of increasing AISM is the enhancement of the ACoC. This is consistent with the study of Heywood et al. (1998), who suggested that the ACoC is likely to be stronger in summer, when wall melting is higher along the ice front. The enhancement of ACoC is most significant from the Amundsen Sea to the Ross Sea, where the surface density decrease is the largest (Fig. 5.3a) due to the addition of AISM. In addition, substantial enhancement is also found in the Ross Sea, the Weddell Sea and the Prydz Bay, where the density decrease is larger. In these regions the westward flow is sped up by 2–4 cm/s.

The enhancement of ACoC due to increasing AISM is related to its driving forcing. Although the ACoC is mainly barotropic and driven by zonal wind through Ekman transport (Sverdrup, 1953), more than 25% of the current is baroclinic (Heywood et al., 1998), generated by a horizontal density gradient between the Antarctic surface water and shelf
water (Fahrbach et al., 1992) and the corresponding thermal wind balance (e.g., Smedsrud et al., 2006). As shown in Fig. 5.4, density decreases along the Antarctic coast due to the addition of AISM. This decrease diminishes in the cross-shore direction. Thus, the northward decreasing density gives rise to a stronger baroclinic component in the westward flow along the coast (Smedsrud et al., 2006), and consequently, the ACoC is enhanced after inclusion of AISM.

The imposed ACoC due to increasing AISM flows westward along the Antarctic continent with the coast to its left (Fig. 5.9). This includes the northward flow in the western Weddell Sea. After passing the tip of the Antarctic Peninsula, one portion flows westward through DP. This feature is seen throughout the whole water column, although the magnitude decreases with depth (Fig. 5.9). Thus, the eastward transport of the ACC through DP is reduced after inclusion of AISM. On the other hand, south of the ACC, the enhancement of the ACoC leads to the strengthening of the cyclonic gyre circulation in the Weddell Sea and Ross Sea, a feature also found in the study of Hattermann and Levermann (2010).

![Fig. 5.8. The annual mean horizontal circulation (m/s) in the simulation with AISM: (a) at the surface (b) near the bottom (4750 m). Note the flow near the coast is westward.](image_url)
5.4.2.2. The ACC

Another consequence of increasing AISM for the ocean circulation is changes in the ACC (Fig. 5.9), although the changes are much smaller than those of ACoC. In general, the ACC becomes weaker after inclusion of AISM. This is because the strong freshening in the south due to AISM (Fig. 5.3) weakens the meridional density gradient and thereby slowing down the ACC. The weakening of the ACC is also found by Hattermann and Levermann (2010) in their first 150 years integration. Since the ACC is the major current connecting the world oceans, the ACC and its change has a strong impact on the global ocean circulation, which may significantly alter the global ocean hydrography as shown in chapter 7.

5.4.2.3. ACC transport through DP

The DP transport measures the zonal flow through the smallest latitudinal extent of the ACC. Observations indicate that it varies from 100 Sv (Orsi et al., 1995) to 134±(15-27) Sv (Cunningham et al., 2003). In contrast, the numerical models produced a wide range of DP transport from 40 Sv to 185 Sv for integration times less than 100 years (Griffies et al., 2009), due to the strong dependency of the DP transport on the model parameters and topography (Gent et al., 2001), and the dependency on the thermohaline circulation off the Antarctic shelf (Russell et al., 2006). Without taking into account the AISM, the simulated DP transport is about 180 Sv in the present study, close to the simulated value of 170 Sv in the study of T2005, who were using the same model (ORCA2-LIM).

Addition of AISM makes the ACC transport through DP more realistic. It effectively suppresses the increasing trend of ACC transport through DP to a more steady state in the first pass. In the second pass, the mean transport through DP is about 150 Sv in the simulation with AISM, approximately 30 Sv lower than that in the simulation without AISM. Thus, the simulated DP transport with AISM becomes much closer to the observed 134 Sv (Cunningham et al., 2003). It is also quite close to the simulated value of 148 Sv by Iudicone et al. (2008), who integrated the ORC2-LIM model (without AISM) for about 1500 years. The integration time in our simulation is about 86 model years, much shorter than that of Iudicone et al. (2008). It might imply that the addition of AISM can help the simulated DP...
transport reach a more realistic value in a much shorter integration period due to the more realistic hydrography (see chapter 4).

5.5. Freshwater flux from AISM: variability and connection with the oceanic temperature of the Southern Ocean

In previous chapter, we calculated the climatological annual mean of the circumpolar freshwater flux from the AISM. Here we identify the temporal variability of the circumpolar freshwater flux and investigate how the freshwater flux and the Southern Ocean temperature respond to each other.

5.5.1. Seasonal and interannual variability

Figure 5.10a illustrates the seasonal variations of the total circumpolar freshwater flux from the AISM. The basal melting rate of the Antarctic ice shelves changes with season, with maximum in summer and minimum in winter. The pattern is consistent with the result of a three-dimensional circulation model for the sub-ice shelf cavity (Nicolaus and Grosfeld, 2004). The maximum and minimum melting occurs in March and September, respectively, with a difference of 7 mSv in the freshwater flux. The occurrence time of maximum and minimum melting is generally consistent with the seasonal variation of the water temperature, although they might have a time lag.

Figure 5.10b shows the time series of the total circumpolar freshwater flux from the AISM during the integration period 1958–2000. We can notice substantial interannual fluctuations. The simulated freshwater flux increases since the mid-1960s, indicating a continuous increase in basal melting during the past decades (Rignot and Jacobs, 2002; Shepherd et al., 2004). An increase in basal melting has been suggested to be the cause of the thinning of Larsen C Ice Shelf (Shepherd et al., 2003) and ice shelves in the Amundsen Sea (Shepherd et al., 2004), owing to the ocean warming (Shepherd et al., 2003; 2004). Ocean temperature is a primary factor for basal attrition (Potter and Paren, 1985; Rignot and Jacobs, 2002). It increased by about 0.2°C over the recent decades at the mid-depth of the Southern Ocean (Gille, 2002; 2003), a depth where CDW, the heat source of melting (Jacobs et al., 1992), reaches the Antarctic continental margin.

Fig. 5.10. Variability of the simulated total circumpolar freshwater flux from the AISM: (a) climatological seasonal cycle, and (b) time series of the monthly (black) and annual (red) mean.
5.5.3. Correlation between the freshwater flux and the Southern Ocean water temperature

In section 5.1, we have seen the changes in temperature in the Southern Ocean due to the addition of AISM. Due to the density difference between the cold freshwater from the AISM and the seawater, the AISM water is mainly confined to the vicinity of the ice shelves. In addition, due to its small density, it rises to the surface quite rapidly (Fig. 7.15). This causes strong freshening and cooling of the Antarctic coastal waters. As a result of the rising of the freshened melt water (as will be seen in Fig. 7.13, there is noticeable vertical circulation in the vicinity of the Antarctic coast), warmer and more saline deep water is compensated to entering the ice shelf cavities for continuity reasons. This induces warming and salt enrichment in the deep layers in the vicinity of the ice shelves and at depth of the Southern Ocean as shown in section 5.1. This mechanism, as well as changes in hyrography in the Southern Ocean, is shown in Fig. 8.2, together with other changes on the global scale.

The increasing AISM induced changes in the water temperature (Section 5.1). The changed water temperature can in turn influence the AISM. In this section, the interaction between the AISM and the water temperature of the Southern Ocean is qualitatively investigated through analysis of the time lag correlation between the freshwater flux and the water temperature in the Southern Ocean.

5.5.3.1. Response of freshwater flux to oceanic temperature

Figure 5.11 shows the time lag correlation between the freshwater flux from the AISM and the Southern ocean water temperature at different levels. A positive time lag in the horizontal axis means that the phase of freshwater flux is in advance (earlier), while negative means the phase of water temperature in advance.

![Time lag correlation](image)

**Fig. 5.11.** Time lag correlations of the freshwater flux with the water temperature south of 40°S at different levels: (a) surface water (0–200 m), (b) intermediate water (200–1500 m), (c) deep water (1500–4000 m), and (d) bottom water (below 4000 m). The positive time lag in the horizontal axis means the phase of the freshwater is in advance (earlier), while negative means the phase of water temperature is in advance.
When the time lag is negative, the correlation coefficient is positive for all depths (Fig. 5.11). This means that the water temperature has positive impact on the freshwater flux, the higher the water temperature, the larger the freshwater flux. This positive response of the freshwater flux to the oceanic temperature is consistent with earlier studies (e.g., MacAyeal, 1984; Rignot and Jacobs, 2002; Holland et al., 2008), since basal melting has a positive correlation with oceanic thermal forcing (Rignot and Jacobs, 2002).

The response of the freshwater flux to the water temperature varies from the surface to the bottom. It is weak for the surface water, where the time lag correlation is generally less than 0.5. On the contrary, it is strong for the intermediate, deep and bottom waters, where the time lag correlation is largely over 0.5. This different response is likely due to the fact that the surface warming tends to stabilize the water column, while the intermediate to bottom warming tends to destabilize the water column.

High correlation is generally found when the time lag is less than 0, i.e. when the water temperature increase/decrease is several years before the freshwater flux increase/decrease (Fig. 5.11). When the time lag is larger than –4 years, the lag correlation between the freshwater flux and the surface water is very low (Fig. 5.11a), suggesting little impact of the surface water temperature on the freshwater flux. The higher lag correlation (over 0.5) occurs between the freshwater flux and the waters from the intermediate to the bottom, with the time lag generally between –5 and –10 years. This suggests that warming of the intermediate, deep and bottom waters would increase the freshwater flux after 5 to 10 years. The largest lag correlation is between the freshwater flux and the intermediate water when the time lag is –5 years, where the lag correlation is more than 0.95 (Fig. 5.11b). Such high correlation is clearly related to the oceanic processes, since the ISM is dependent upon the warm water transported beneath the ice shelf and the tidal process near the ice front (Jacobs et al., 1992). The results suggest that, after the Southern Ocean warms, it may take about 5–10 years for the warm water to be transported onto the continental shelf and to melt the ice shelf.

5.5.3.2. Response of the water temperature to the freshwater flux

When the time lag is positive, the correlation coefficients vary significantly through the water column. For surface waters, the correlation changes from positive to negative at zero lag, indicating that increasing AISM will has no instant influence on the surface water temperature. With the increase of AISM and the decrease of surface water temperature, the largest negative lag correlation occurs when the time lag is 9–12 years. This means that the impact of increasing AISM on the Southern Ocean surface temperature approaches the largest after about 10 years. The negative lag correlation between the freshwater flux and intermediate water takes place when the time at 5 years, with the lowest lag correlation appearing after about 13 years (Fig. 5.11b), indicating that increasing AISM begins to cool the Southern Ocean intermediate water after 5 years. The freshwater flux generally has a positive lag correlation with the deep and bottom waters. This seems to indicate that increasing of the AISM would warm up the deep and bottom water, since the increase upwelling of deep water causes more warm NADW intrusion the Southern Ocean, and thereby increasing the water temperature.

5.6. Discussion and conclusions

In this chapter, we have investigated the impact of increasing AISM on the Southern Ocean hydrography, MLD and ocean circulation, explored the seasonal and interannual variability of the freshwater flux, and examined the correlation between the AISM freshwater flux and the Southern Ocean temperature. The impact is investigated through the difference between the simulations with and without AISM, since the simulation without AISM can be regarded as an extreme case of small (vanishing) AISM. This difference
provides a description of the hydrographic changes in the Southern Ocean due to increasing AISM, and can be qualitatively compared with the observed trend.

Increasing AISM tends to cause freshening of the surface water, warming in the intermediate and deep waters, and warming and freshening in the bottom water. Therefore, the water in the Southern Ocean largely becomes lighter with the increasing AISM. These simulated hydrographic changes in the Southern Ocean due to increasing AISM are highly consistent with the observed trends in the recent studies. This implies that the increasing AISM is highly likely a major contributor to the recent changes in the Southern Ocean.

Increasing AISM leads to changes in the MLDs at the deep convection sites in the Southern Ocean. It leads to deepening of the mixed layer on the Antarctic continental shelves while shoaling in the ACC region. As a result, the simulated MLDs with AISM are in much better agreement with the observations.

Increasing AISM influences the current system in the Southern Ocean. It tends to weaken the ACC, but strengthen the ACoC as well as the Weddell Gyre and the Ross Gyre. The weakening of the ACC results in decrease of DP transport and a more realistic value during the short integration time. Under global warming, the strengthening of the ACoC would transport more heat to the ice shelves, thus producing even larger freshwater flux. Then the additional freshening of the Antarctic coastal water further accelerates the ACoC and enhances the warming of the deep Southern Ocean. Therefore, it forms a positive feedback with the AISM (Hettermann and Levermann, 2009). On the other hand, the strong freshening in the south due to the AISM leads to a reduced meridional density gradient, thus weakening the ACC and therefore transporting less heat to the ice shelves. This forms a negative feedback with the AISM (Hattermann and Levermann, 2009). Thus, the weakening of ACC and the enhancement of the ACoC has a competitive effect on the ISM under global warming.

The freshwater flux due to AISM shows a clear seasonal cycle, with the maximum in summer and minimum in winter. This is consistent with the seasonal variability of the Southern Ocean water temperature. There is large interannual variability in the freshwater flux, with a clear increasing trend of the freshwater flux in the simulations since the mid-1960s, consistent with the observed Southern Ocean warming in the past decades.

Under ocean warming, more heat is available to melt the ice shelves. Consequently, more freshwater flux is produced when the warm water is transported onto the continental shelves. This tends to increase the absorbed heat from the ocean, enhance the upwelling and cool the intermediate and surface water. The enhanced upwelling then attracts more warm water from deep, which is compensated by the warm deep water from north, and thereby warming the deep and bottom water.

The freshwater flux and the water temperature in the Southern Ocean influence each other. The Southern Ocean water temperature has a positive impact on the freshwater flux. The higher the water temperature, the larger the freshwater flux. On the other hand, the freshwater flux has varying impact on the water temperature at different depths. It tends to decrease the temperature of surface and intermediate waters, but to increase the temperature of deep and bottom waters. Thus, the surface and intermediate water temperature forms a negative feedback loop with the AISM, while the deep and bottom water forms a positive feedback loop with the AISM.

The simulated feedback between the freshwater flux and the Southern Ocean water temperature has a time lag of approximately 5–10 years. This range of values has to be regarded as the lower limit because the sub-ice shelf cavities are not explicitly included in the model and the flushing of these might prolong the time for feedback. For example, the flushing time of the Filchner-Ronner Ice Shelf is estimated to be of the order of 24–30 months (Nicholls and Østerhus, 2004), which could make the time lag 2-3 years longer.
Chapter 6 Response of the Antarctic Sea Ice to Antarctic Ice Shelf Melting

In Chapter 5, we have seen that inclusion of AISM introduced significant changes in the ocean system of the Southern Ocean. In this chapter, we focus on investigating the effect of AISM on the Antarctic sea ice. This is accomplished by a detailed analysis of the simulated Antarctic sea ice evolutions with and without AISM, together with comparisons with observations.

6.1. Sea ice extent and area

Here we use two variables, sea ice extent and area, to describe the sea ice coverage. The former is the integrated area of all grid cells having ice concentration at least 15%, as used in Chapter 3, and the latter is the summed areal coverage of ice, calculated as the sum of the products of ice concentration by the grid cell area. In the following, we will compare the time series and seasonal cycle between the observations and the simulations with and without AISM.

6.1.1 Seasonal cycle

Figure 6.1a compares the modeled and observed mean seasonal cycles of the Antarctic sea ice extent and area from October 1978 to December 2000. The observed cycles are computed from the ice concentration derived from SSMR-SSM/I satellite using Bootstrap algorithm (Comiso, 1999, updated 2003). The observed Antarctic sea ice cover shows a strong seasonality, with minima in February and maxima in September. Therefore, the ice growth season is longer than the ice decay season. During the period of 1978–2000, the minimum monthly mean ice extent is 3.37×10^6 km², in February, and the maximum monthly mean ice extent is 19.23×10^6 km², in September, with an annual mean of 12.36×10^6 km² (see Table 6.1). In contrast, the minimum and maximum ice areas are 2.14×10^6 km² and 15.89×10^6 km², respectively, about 1.23×10^6 km² and 3.34×10^6 km² smaller than the ice extent. The ratio of ice area to extent is 63.5% in February and 82% in September, implying higher sea ice concentration in September.

![Seasonal cycle](image)

**Fig. 6.1.** Seasonal cycle (a) and time series (b) of sea ice extent (top) and area (bottom) in observations (black) derived from SSMR-SSM/I and in the simulations with (red) and without (blue) AISM.
Table 6.1. Minimum, maximum and annual mean of the monthly climatology for sea ice extent and area from observations (SSMR-SSMI/I) and from simulations with and without AISM.

<table>
<thead>
<tr>
<th></th>
<th>Sea ice extent ($\times 10^6$ km$^2$)</th>
<th>Sea ice area ($\times 10^6$ km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Min</td>
<td>Max</td>
</tr>
<tr>
<td>Observations</td>
<td>3.37</td>
<td>19.23</td>
</tr>
<tr>
<td>with AISM</td>
<td>0.71</td>
<td>19.46</td>
</tr>
<tr>
<td>without AISM</td>
<td>0.39</td>
<td>18.57</td>
</tr>
<tr>
<td>(with-without) AISM</td>
<td>0.32</td>
<td>0.89</td>
</tr>
</tbody>
</table>

Compared with the observations, the modeled sea ice cover captures well the strong seasonality of the Antarctic sea ice extent and area, the occurrence time of the minimum ice coverage, as well as the magnitude of the maximum ice coverage. However, there are obvious shortcomings. First, the minimum sea ice extent (area) is 0.71 $\times 10^6$ km$^2$ (0.4 $\times 10^6$ km$^2$) and 0.39 $\times 10^6$ km$^2$ (0.22 $\times 10^6$ km$^2$) for the simulations with and without AISM. Thus, although the occurrence times of the minima of sea ice cover are well reproduced, the magnitude of the minimum is substantially underestimated. Second, although the maximum sea ice extent (area) is 19.46 $\times 10^6$ km$^2$ (16.1 $\times 10^6$ km$^2$) and 18.39 $\times 10^6$ km$^2$ (15.1 $\times 10^6$ km$^2$) for the simulations with and without AISM, both close to the observations, the occurrence time appears too early, about one month in advance, due to the rapid thermodynamic growth in the model (Timmermann et al., 2004). On the whole, the modeled sea ice cover expands too fast and shrinks too early compared with the observation.

Inclusion of AISM has minor effect on the occurrence times of the minima and maxima of the sea ice cover, but it improves the area of sea ice cover (Fig. 6.1). Due to the cooling of the mixed layer after including AISM (see Chapter 5), the simulated sea ice extent and area both increase. The increase in minimum sea ice coverage is small, about 0.32 $\times 10^6$ km$^2$ in sea ice extent and 0.18 $\times 10^6$ km$^2$ in sea ice area. Therefore, the minimum sea ice coverage is still substantially underestimated. This appears mainly due to the absence of summer sea ice in the east Antarctic, as discussed below. The increase in maxima is large, about 0.89 $\times 10^6$ km$^2$ in sea ice extent, and 0.69 $\times 10^6$ km$^2$ in sea ice area. As a result, the maximum ice extent and area are much closer to the observations after inclusion of AISM (Table 6.1).

The addition of AISM has a larger impact on the sea ice cover during the period of sea ice decay than during the period of sea ice growth. During the decay period, the surface water becomes warm with more incoming solar radiation. However, cooling of the mixed layer due to AISM retards this process and keeps the surface water closer to the freezing point. Therefore, sea ice melts later and more ice is left in the simulation with AISM. On the contrary, the surface water is already near the freezing point during the period of sea ice growth. The addition of AISM therefore has little effect on the sea ice production, resulting in less increase in sea ice cover when sea ice grows.

On the whole, inclusion of AISM leads to an increase in the annual mean sea extent from 10.41 $\times 10^6$ km$^2$ to 11.1 $\times 10^6$ km$^2$ and the sea ice area from 7.99 $\times 10^6$ km$^2$ to 8.65 $\times 10^6$ km$^2$ (Table 6.1), although the sea ice extent (area) is still underestimated by 10.9% (9.3%) due to the substantial underestimation of minima in February.

6.1.2. Time series

6.1.2.1. Monthly mean

Time series of sea ice extent and area (Fig. 6.1b) show interannual fluctuations during the period of 1978–2000 when satellite data is available. The interannual variability of the minimum sea ice extent and area is much smaller compared with that of maximum sea ice extent and area. The model basically reproduces these features, except the minima in
February are substantially underestimated. The correlation coefficient between the simulated and observed times series is high for integrated sea ice coverage whether with or without AISM, being 0.88 and 0.90 for the sea ice extent and area before inclusion of the AISM, and 0.90 and 0.92 after inclusion of the AISM.

### 6.1.2.2. Monthly mean anomalies

The modeled and observed time series of monthly sea ice extent and area anomalies over the time interval from October 1978 to December 2000 are illustrated in Fig. 6.2a. The anomalies are obtained by subtracting the monthly values for each individual month from the average of that month over the 22-year satellite period. The marked interannual variability seen in the observations is well captured by the model. In particular, the model successfully reproduces the abnormally low ice areal coverage during the first half of 1980s as observed, as well as the positive anomalies in 1985/86 and the positive and negative anomalies in 1996/97. However, the model fails to capture the anomalies in the autumn and winter of 2000, which is largely positive in the observations, while negative in the simulation. The reason is so far unclear.

Figure 6.2b shows the times series of the modeled sea ice extent and area anomalies during 1958–1978. It is seen that the anomalies are more positive before the period when satellite data became available (Fig. 6.2b) compared with those thereafter (Fig. 6.2a). Cavalier et al. (2003) showed that anomaly decreased from 1973 to 1976 through bridging the gap between the Nimbus 7 data and the earlier Nimbus 5 data. This feature is basically captured by our simulation. However, the model fails to reproduce the Weddell Polynya, an area of $0.1–0.3 \times 10^6 \text{ km}^2$ in the sea ice cover of the Weddell Sea only occurring in 1974–1976 (e.g., Zwally et al., 1983). This is mainly because the coarse resolution of the model cannot resolve the oceanic processes (Fichefet et al., 2003) which are thought at least partly to contribute to the formation of the Weddell Polynya. Such potential processes are, e.g., the eddy shed by the seamount (Holland, 2001), and the specific flow regime developed at the seamount as shown by Beckmann et al. (2001).

![Fig. 6.2. Monthly ice extent (top) and area (bottom) anomalies (a) during the period of satellite data available October 1978-December 2000, and (b) during 1958-1978 before satellite data available. Black lines denote observations, red lines for simulations with AISM, and blue line for simulations without AISM.](image-url)
Large positive anomalies of the sea ice extent are found in 1965 in our simulation. Compared with the climatological mean, the winter sea ice cover generally increases near the ice edge in the Southern Ocean. This is mainly due to the decrease of air temperature in this year. Other factors may also play a role, such as the Southern Annular Mode (SAM), which has a significant impact on the strength of the westerlies and the circulation variability of the lower atmosphere in the Southern Hemisphere (e.g., Connolley, 1997; Thompson and Wallance, 2000). When the SAM is positive, the associated pressure pattern induces meridional winds, which advects warmer air into the Weddell Sea, and vice versa (Lefebvre et al., 2004). Therefore, the negative SAM in 1965 tends to induce more southerly winds and drive the sea ice northward, thereby increasing the ice cover.

Similar to the seasonal cycle of sea ice extent, the time series of sea ice extent (Fig. 6.1b) show again that inclusion of ISM leads to generally higher values. However, inclusion of AISM generally has little influence on the anomalies of monthly sea ice extent and area (Fig. 6.2).

6.1.2.3. Trend of sea ice extent anomalies

Figure 6.3 depicts the trend of the monthly mean sea ice extent anomalies during 1987–2000 and 1958–2000. These trends are quite different during different intervals. Antarctic ice pack increased during 1978–2000 according to the satellite observations. This result was also found in the simulations. But during 1958–2000 the simulations showed an opposite trend. The increase after 1978 was recognized by Zwally et al. (2002), Cavalieri et al. (2003) and Cavalieri and Parkinson (2008) using satellite data over intervals of 1979-1998, 1972-2002, and 1979-2006, respectively. This increase is opposite to the decreasing trend in the Arctic sea ice (e.g., Parkinson et al., 1999; Parkinson and Cavalieri, 2008).

With the climate warming, the increase during 1978-2000 in the Antarctic sea ice is surprising. But, if the Antarctic ice pack appears to be in shrinking phase when extending the time interval back to 1958 (Fig. 6.3b). This is true even only dating back to 1972 (Cavalieri et al., 2003), mainly due to the regime shift during mid 1970s (Fig. 6.3). A marked decline in sea ice was found during the second half of the 1970s and at the beginning of 1980s, because of the rather dramatic weakening of the semiannual oscillation (SAO) since the mid-to-late 1970s (Fichefet et al., 2003).

![Fig. 6.3. Trend of the monthly mean ice extent anomalies during the interval (a) between 1978-2000, and (b) 1958-2000. Black dots: observation, red: with AISM, and blue: without AISM.](image-url)
6.2. Spatial distribution of sea ice concentration

6.2.1. In austral summer

As described above, sea ice cover shrinks to its minimum coverage in February. In this month, sea ice is still circumpolar around the Antarctic continent with a very small amount in the east Antarctic as shown in satellite observations (Fig. 6.4a). Sea ice is usually within 1–2º latitudes away from the east Antarctic coast. More sea ice is found in the west Antarctic. Higher sea ice concentration is found in the southern and western Weddell Sea, and in the Amundsen and Bellingshausen Sea. In the Ross Sea, sea ice concentration is lower, especially in the eastern Ross Sea (<10%), where is the location of Ross Sea Polynya occurrence.

Fig. 6.4. Average sea ice concentration in (a, b) summer (February) (c, d) winter (September for observation; August for simulation) over the years October 1978-December 2000 in observation (a, c) derived from SSMR-SSM/I and in the model (b, d) with AISM.
The model with AISM (Fig. 6.4b) captures the basic features of the spatial distribution of sea ice concentration in February, for example, the Ross Sea Polynya, and more sea ice in the west Antarctic. Higher sea ice concentrations in the Weddell Sea and in the Amundsen and Bellingshausen Seas are also reproduced, although the simulated sea ice concentration in these regions is relatively low. However, there is an obvious inconsistency. Sea ice along the Antarctic coast is missing in the simulation. This was attributed to the relatively coarse resolution of the model by T2005. On the other hand, computation of ice concentration from the remote sensing data is subject to large uncertainties with an inaccuracy of 15–20% in summer (Comiso et al., 1997; Cavalieri et al., 1999). Thus, sea ice concentration might be overestimated in the observations along the east Antarctic. Sea ice near the tip of the Antarctic Peninsula is also missing in the model. This is mainly attributed to a poor representation of the Antarctic Peninsula in the atmospheric forcing (Windmüller, 1997). On the whole, the underestimated sea ice cover in these regions contributes to the overall lower summer sea ice extent in the simulation.

6.2.2. In austral winter

In austral winter, sea ice attains its maximum areal coverage in September in the observations. In this month, the spatial distribution of sea ice concentration (Fig. 6.4c) displays several features. (1) Sea ice concentration is generally high (>80%) in the Southern Ocean, with abrupt changes near the ice edge due to influence of the warm ACC. (2) Sea ice concentration is relatively low along the Antarctic coast, along with formation of the so-called coastal polynyas. These are ubiquitous around Antarctica during winter (Massom et al., 1998), often caused by offshore blowing katabatic winds pushing the sea ice away from the coast. They were estimated to cover an area of \((0.2\pm0.01)\times10^6\) km\(^2\) during wintertime (June to September) from 1992–2008 by Kern (2009), playing an important role in the transformation of intermediate and deep waters in the global ocean due to the ocean surface cooling and brine rejection during sea ice growth (Morales Maqueda et al., 2004). (3) Sea ice protrudes northward corresponding to the Weddell Gyre, the Ross Gyre and the Kerguelen gyre. And (4) Sea ice concentration is relatively low (about 80%) between 45ºE–135ºE compared with other regions.

![Fig. 6.5 Monthly mean differences (with-without) of sea ice concentration in a) February and b) August.](image-url)
Compared with the summer simulation, the winter simulation comes out much better. The spatial features of sea ice concentration in the observations are generally well reproduced by the model with AISM (Fig. 6.4d). Nevertheless, there are slight differences. (1) Although we capture the feature of winter coastal polynyas, the areas of the polynyas in the simulation are smaller than in the observations due to the coarse resolution of the model. (2) Coastal polynyas in the east Antarctic seem to be overestimated in the simulation compared with observations. The regions with ice fraction lower than 50% in our simulation were identified as persistent polynyas according to Kern (2009) using an iterative classification technique based on SSM/I data.

### 6.2.3. Impacts of AISM on sea ice concentration

The presence of AISM generally results in increase in sea ice concentration in February (Fig. 6.5a). The increase is generally small in most regions. Large increase is found in the inner Southern Weddell Sea, where increase in sea ice concentration is up to 30% (Fig. 6.5a), making the sea ice concentration much closer to the observations due to addition of AISM.

In August (Fig. 6.5b), AISM has little influences on the inner Weddell Sea, Ross Sea, and the Amundsen and Bellingshausen Seas, where sea ice concentration is high. Major differences are mainly found near the ice edge toward the equator and in the east Antarctic where the polynyas account for about 40% of the wintertime coastal polynyas (Kern, 2009). Increase in sea ice concentration is mainly found in the Prydz Bay, near the ice edge of the Weddell Sea and the Ross Sea, while decrease is mainly found near the ice edge between 45°E–90°E in the east Antarctic.

Changes in sea ice concentration mainly result from warming and cooling of sea water (Fig. 5.5a, b). Cooling decreases the oceanic heat flux to the ice, and therefore more ice forms in open water areas, which leads to increase in sea ice concentration. In contrast, warming of the sea water increases the oceanic heat flux, and thus less ice forms in open water areas. Since the sea ice concentration is 100% or nearly 100% in the inner Weddell Sea, Ross Sea, and the Amundsen and Bellingshausen Sea in August, the cooling of the sea surface water has little influences on the sea ice concentration.

![Sea ice velocities in September during 1958-2000 from the simulation with AISM. The scale vector corresponds to an ice velocity of 0.1 m/s.](image)
Fig. 6.7. Simulated sea ice velocity differences (with-without) in (a) vector, and (b) speed mapping in September during 1958-2000. The scale vector corresponds to an ice velocity of 0.1 m/s.

6.3. Sea ice drift

Figure 6.6 displays the long-term mean sea ice velocities in September during 1958–2000 from the simulation with AISM. Sea ice motion generally follows the general circulation of the surface current (Fig. 5.8a) and the atmosphere. In the north, there is a cyclonic circulation, the Antarctic circumpolar drift, transporting sea ice from west to east all around the Antarctica. In the south, the East Wind drives the coastal flow westward. Therefore, sea ice drifts from east to west circumscribing the Antarctic continent. Between them are pronounced cyclonic ice circulations in Weddell Sea, Ross Sea, and in the vicinity of Amery Ice Shelf, together with the rapid offshore ice drift in the western Ross Sea. These features are all well reproduced in the simulation, similar to the reconstructed sea ice velocities (Fichefet et al., 2003; Fig. 4b) from animations of SSM/I brightness temperature (Drinkwater and Liu, 1999). The ice speed is about 5–8 cm/s along the Antarctic continent, 5–10 cm/s for the Antarctic circumpolar drift, and up to 11 cm/s in the western Ross Sea, generally consistent with the observations (e.g., Emery et al., 1997). The rapid offshore ice drift in the western Ross Sea quickly exports the newly formed ice and thin ice out of the southern Ross Sea into the Ross Gyre in the east or the coastal current in the west. This divergent flow of sea ice contributes to the formation of spring Ross Polynya (Fichefet and Goosse, 1999).

The presence of AISM has a notable influence on the sea ice motions (Fig. 6.7). The similar directions of the arrows in Fig. 6.6 and Fig. 6.7a imply that the sea ice drift velocities increases due to the addition of AISM. The increase is remarkable along the Antarctic coast, especially in the Weddell Sea, Ross Sea, Amundsen Sea and Bellingshausen Sea, and most pronounced in the region of the Antarctic circumpolar drift, particularly in the northern Weddell Sea. In the other regions, the different directions of the arrows in Figs. 6.6 and 6.7a indicate changes of velocity in both zonal and meridional components.

Figure 6.7b shows the corresponding speed of the velocity difference. It is apparent that the largest increase in the magnitude is near the ice edge, particularly in the northern Weddell Sea, where the increase is up to 15 cm/s. In the northern Ross Sea and in the Kerguelen Gyre the increase is also notable, about 10 cm/s. In the other regions, the difference is generally very small, less than 2 cm/s. When comparing with the increase in sea ice concentration (Fig. 6.5b), it is seen that the regions of large changes in sea ice speed are
overlapping with those in sea ice concentration (Fig. 6.5b). This indicates that the internal stress plays a very important role in altering the velocity field.

### 6.4. Spatial distribution of the sea ice thickness

The simulated sea ice volume grows to its maximum in September and shrinks to its minimum in February (Fig. 6.10). Therefore, in the following, we show the sea ice thickness in September to represent the sea ice conditions for austral winter and in February for austral summer. The geographic distribution of the mean sea ice thickness simulated with AISM is shown in Fig. 6.8.

![Spatial distribution of mean sea ice thickness](image)

**Fig. 6.8.** Spatial distribution of mean sea ice thickness in (a) February, and (b) September simulated with AISM.

#### 6.4.1. In austral winter

In austral winter (Fig. 6.8a), the simulated spatial pattern of the sea ice thickness is generally consistent with the mean ice thickness mapped by Worby et al. (2008) from ship-based observations. The simulated sea ice thickness is generally below 1.5 m in most regions of the Southern Ocean, which is also in good agreement with the observations (Wadhams et al., 1987). Thicker ice (>2 m) mainly concentrates in the southern Weddell Sea, the western Weddell Sea, and the Amundsen and Bellingshausen Seas. In these regions multi-year ice is found (Fig. 6.8b). The thickness of the multi-year ice in Antarctic is significantly thinner than its Arctic counterpart, where multi-year ice can be more than 3.5 m (Bourke and McLaren, 1992; Rothrock et al., 2003). This can be attributed to the larger oceanic heat flux (McPhee et al., 1999; Maykut and McPhee, 1995), and to the thicker snow cover in Antarctic. The latter weakens the conductive heat flux through ice (Eicken et al., 1995).

In the Weddell Sea, our simulation features two regions of thicker ice. One is in the southern Weddell Sea, and another is in the western Weddell Sea, both multi-year ice regions. In the southern Weddell Sea, multi-year ice is generally well reproduced by models (e.g. Zhang, 2007; Stössel et al., 2007), because of the good representation of the convergent motion of sea ice in this area. This builds-up ice along the coast. In the western Weddell Sea, sea ice is transported northward along the coast, and therefore ice is thicker in the north than in the south, consistent with the results of Worby et al. (2008). However, the thickness along the eastern side of the Antarctic Peninsula is underestimated by up to 1 m in austral winter compared with the observations (Harms et al., 2001). This is partly because the model tends to underestimate the impact of dynamics, for example, rafting and ridging (Timmermann et
al., 2004). Improvement is expected when the two-category ice thickness distribution is replaced by a multi-category ice thickness distribution (e.g. Haapala, 2000).

In the Amundsen and Bellingshausen Seas, sea ice drifts westward along the continent (Fig. 6.6). It becomes thicker at some regions along the western side of the Antarctic Peninsula, due to changeable coastline (Fig. 6.8a), with thickness up to 2 m. This sea ice distribution feature is consistent with the observations of Worby et al. (2008), although the thickness is slightly larger.

On the whole, the thick ice is well simulated, particularly in the western Weddell Sea (Worby et al., 2008) and in the area from the Amundsen Sea to the east Ross Sea (Jacobs and Comiso, 1989).

6.4.2. In austral summer

In austral summer (Fig. 6.8b), sea ice is much thinner than in austral winter (Fig. 6.8a). Multi-year ice is located in the southern Weddell Sea and the Amundsen Sea. The underestimation of the sea ice on both sides of the Antarctic Peninsula is due to the warm bias in surface air temperature in this region (Windmüller, 1997).

There is a notable deficiency in the simulation. Sea ice is totally absent off East Antarctica (Fig. 6.8b). In contrast, sea ice with high concentration is clearly visible in the sheltered embayment of the SSMR-SSM/I sea ice concentration map (Fig. 6.4a). This difference has been attributed to the coarse resolution of the model (T2005). There might be two other reasons. One reason is that the model only takes into account congelation ice and snow ice formation (Fichefet and Morales Maqueda, 1997) but not superimposed ice formation. Superimposed ice, as defined here, forms by refreezing of water-soaked snow resulting from the infiltration of rain or meltwater above the snow-ice interface. It contributes to the summer growth of sea ice (Kawamura et al., 1997; Jeffries et al., 1994, 1997; Massom et al., 2001). Ignoring of the superimposed ice formation in the model might contribute to the substantially underestimated sea ice extent in summer. Another, more important reason is that the simulated sea surface water is too warm in summer, commonly above 0ºC. Therefore, it is impossible for sea ice to survive throughout summer, even for landfast ice, which makes up as much as 14% of the area of the east Antarctic sea ice at its maximum (Fedotov et al., 1998). Fast ice freezes up in February or March (Lei et al., 2010) or survives through summer when a bay or fjord can provide protection from ocean swell and wind fetch (Crocker and Wadhams, 1989; Tang et al., 2007)

![Fig. 6.9. Spatial distribution of average sea ice thickness difference (with-without) in (a) September and (b) February.](image-url)
6.4.3. Impact of the AISM

The addition of AISM has a considerable influence on the simulated Antarctic sea ice thickness (Fig. 6.9). Sea ice thickness increases significantly in the vicinity of the ice shelves and downstream from them. This is because the cooling of the sea surface water (Fig. 5.2a) reduces the oceanic heat flux available for ice. Ice thickening is especially large in the regions of thick ice, where the oceanic heat flux is generally reduced by more than 15 Wm$^{-2}$.

The changing pattern of sea ice thickness in September (Fig. 6.9a) is largely consistent with the simulations of Hellmer (2004) who considered the sub-ice shelf freshwater input in his model with ice shelf cavities included explicitly. Nevertheless, the changes in sea ice thickness are relatively large in our simulation due to the larger glacial freshwater input. For instance, in the southern and western Weddell Sea, one region of significant ice thickening, the winter sea ice thickness increases by 40–50 cm in this study, while it was around 20 cm in the study of Hellmer (2004).

There are two other regions of large increase in the winter sea ice thickness in this study. One is close to Amery ice shelf. The thickness increases by 30–40 cm due to the cooling of surface water, consistent with the results of BG03. Another region is in front of EW Ice Shelves, a crucial region for water mass preconditioning and formation in the Weddell Sea (Thoma et al., 2006). Winter sea ice thickness in this region increases by up to 70 cm, being the most significant increase in the Southern Ocean. This is because the large freshwater from Fimbul Ice Shelf is transported westward by the coastal current, which induces large cooling of the surface water (Fig. 5.2a). The thickened sea ice then drifts to cover the adjacent ocean, thus resulting in thicker ice downstream. Such ice thickening is also clearly seen in the northern Weddell Sea (Fig. 6.9a).

Thickening of sea ice is not fully dependent on the adjacent AISM. For example, although the fresh water flux from EW Ice Shelves is small (Table 4.1), ice thickening is the most pronounced. Similarly, while the fresh water fluxes from Fimbul Ice Shelf and Getz Ice Shelf are the largest, the ice thickening is very small. These phenomena suggest that the melt water may be quickly transported to the west due to the close proximity to the coastal current in these regions (Rodehacke et al., 2007).

![Fig. 6.10. Seasonal cycle of Antarctic sea ice volume (top) in total and (bottom) in the Weddell Sea (line) and in the Ross Sea (line with symbols) in the simulations with (red) and without (blue) AISM.](image-url)
In addition to the thickening, thinning of sea ice (Fig. 6.9) is also seen in part of the Southern Ocean due to the presence of AISM. In austral winter, sea ice thinning is mainly found in the east Antarctic, around 45°E and between 90°E–135°E (Fig. 6.9a). Sea ice thinning is also found west of Antarctic Peninsula in austral winter (Fig. 6.9a) and summer (Fig. 6.9b). These regions are consistent with upper-ocean warming due to the presence of AISM (see Chapter 5), which increases the oceanic heat flux to the ice bottom. In general, the ice thinning is not more than 0.3 m.

6.5. Ice Volume

Sea ice volume here is defined as the mean ice thickness multiplied by ice area. Figure 6.10 illustrates the seasonal cycle of Antarctic sea ice volume in the simulations with and without AISM. Antarctic sea ice attains its maximum volume in September and minimum in February (Fig. 6.10, top). The simulated maxima are generally comparable with the results of other studies (e.g., Fichefet and Morales Maqueda, 1999). However, the minima seem too low due to the underestimation of sea ice thickness in austral summer as described in section 6.4. Therefore, the simulated annual mean volume is much lower than in other studies (e.g., Zhang, 2007). Due to the presence of multi-year ice in the Weddell Sea, sea ice volume is relatively high in this region in comparison with in the Ross Sea (Fig. 6.10, bottom).

Introduction of AISM leads to an overall increase in the sea ice volume (Fig. 6.10) due to the general increase in sea ice thickness (Fig. 6.9). Before inclusion of the AISM, the minimum and maximum sea ice volumes are 0.48×10^3 km^3 and 16.15×10^3 km^3, respectively. They increase to 1.08×10^3 km^3 and 18.84×10^3 km^3, respectively, after addition of AISM. The increase in maximum volume (2.69×10^3 km^3) is larger than in the minimum (0.6×10^3 km^3) due to the larger increase in ice thickness in winter. On average, the total ice volume increases by about 18% (annual mean) due to the addition of AISM. In the marginal seas, the increase in volume due to the AISM is more substantial, about 24% in the Weddell Sea, and about 29% in the Ross Sea.

6.6. Summary

In this chapter, we investigated the Antarctic sea ice cover and its response to the AISM. The simulations with and without AISM generally reproduced the observed features in the Antarctic sea ice cover, with the simulations with AISM much closer to the observations.

During the integrated period 1958–2000, the anomaly of sea ice extent (area) is more positive in the period before SMMR-SSM/I data are available. The trend of the sea ice extent anomaly has a large dependence on the analysis time interval. For example, the sea ice extent has a weak increasing trend during 1978–2000, but has a strong decreasing trend during 1958–2000. Inclusion of AISM does not change the anomaly and trend of the sea ice extent (area).

The presence of AISM has a notable impact on the sea ice concentration and thickness. These quantities generally become higher with the addition of AISM due to the cooling of seawater. Large increase in sea ice concentration mainly takes place near the ice edge in austral winter, since the inner sea ice already holds a high concentration (over 90%). Large increase in sea ice thickness is mainly found along the coast of the Weddell Sea, the Bellingshausen and Amundsen Sea, and the Ross Sea and downstream there from. The overall increase in sea ice thickness leads to the increase in sea ice volume.

Inclusion of AISM also influences the sea ice drift. The westward sea ice drift along the continent is generally increasing due to the presence of AISM, particularly in the Weddell Sea, the Bellingshausen and Amundsen Sea and the Ross Sea. Strengthening of the westward circulation along the coast is also seen in the surface ocean circulation, indicating the large
effect of current on sea ice drift. In addition, sea ice drift velocities are also found to increase
near the ice edge. This is because large change in sea ice concentration in this region greatly
modifies the internal stress, resulting in significant changes in the velocity fields.

Inclusion of AISM improves the sea ice extent and area, however, the minima in February
are still notably underestimated and the occurrences of the maxima are still one month earlier
than in the observations. The former is mainly attributed to the absence of sea ice in the east
Antarctic, which results from the coarse resolution of the model, the too warm water and the
missing superimposed ice formation. The latter is attributed to the too fast thermodynamic
growth in the sea ice model.
Chapter 7 Remote Impact of the Antarctic Ice Shelf Melting

In the previous two chapters, we have demonstrated that the addition of AISM noticeably alters the Southern Ocean hydrography and the Antarctic sea ice cover. Due to the connection of the Southern Ocean with the other three major oceanic basins, it is expected that the signal of AISM would not be confined to the Southern Ocean, but spread into the global oceans. Toggweiler and Samules (1995) suggested that up to 75% of the deep ocean’s water might retain the signature of the Antarctic ice shelf meltwater input. Losch’s (2008) study indicated that the signal from the two largest Antarctic ice shelves could reach as far as 15ºS. Wang and Beckmann (2007) revealed significant changes in the MLDs in the North Atlantic due to the addition of AISM. In this chapter, we will perform a thorough investigation of the impact of AISM on the temperature, salinity, MLDs and sea ice in the northern hemisphere. Special attention is laid on the North Atlantic Ocean and the Arctic Ocean, in which the most significant changes in the northern hemisphere are detected. Detailed comparisons between the simulations with AISM and observations are performed to investigate the overall skill of the model in the North Atlantic and the Arctic oceans; then the impact of AISM is identified through comparisons between the simulations with and without this process. Finally, a physical mechanism is proposed to explain the inter-hemispheric teleconnection.

7.1. Temperature
7.1.1. Comparison between simulated and observed temperature

In this section, we compare the observed and simulated annual mean temperature in the North Atlantic Ocean and the Arctic Ocean. The corresponding comparisons for salinity will be performed in section 7.2. For these comparisons, the simulated results are obtained from the simulation with AISM, and the observations are taken from the 1° latitude \( \times 1° \) longitude grid data of the WOCE 98 Atlas (http://www.ersl.noaa.gov/psd/data/grided) at the Earth System Research Laboratory. We choose four layers for detailed comparisons, namely surface, the depth of 200 m, 1000 m and 3000 m, respectively. The corresponding model depths are 5 m, 217 m, 1033 m and 3257 m, respectively, which differ only slightly from the standard depths due to the spatial arrangement of the model grid. It is expected that such small depth deviation should not cause significant difference in the results.

Figure 7.1 compares the observed and simulated annual mean potential temperature for the selected layers in the North Atlantic Ocean and the Arctic Ocean. It is seen that the main features of the SST (sea surface temperature) are well captured by the model (Fig. 7.1a, b). In the observation, the SST is high in the tropics, up to 30°C, and decreases northward, down to below 0°C in the Arctic Ocean. This feature is very well reproduced by the model. In addition, the SST gradients corresponding to the meandering water mass boundary across the North Atlantic Ocean into the Nordic Seas are represented well, considering the coarse resolution of the model.

There are, however, also some differences between the simulation and the observations. First, the simulated SST is slightly underestimated in the Barents Sea and in the Greenland Sea close to Svalbard. This implies too little warm water flowing into the Arctic Basin (the Eurasian Basin and the western Basins, namely the Makarov and the Canadian Basin), which would result in too cold water in the Arctic Basin. As a result, the SST in the Arctic Basin is
a little bit too low (down to –2ºC) in the simulation. In addition, the colder water in the Arctic Basin is also partly due to the overestimation of the SSS (sea surface salinity) in the simulation (see Fig. 7.3b), which leads to a decrease in the freezing point of the sea water. Secondly, the SST is underestimated in the Hudson Bay, Baffin Bay and Davis Strait. This is probably due to the artificial enlargement of the Nares Strait between Greenland and the Canadian Archipelago in the model, which allows more cold Arctic water flowing southward into the northwestern North Atlantic Ocean.

Compared with the surface water (Fig. 7.1a, b), the North Atlantic subsurface water at the depth of 200 m (Fig. 7.1c, d) becomes colder, with the temperature down to about 5ºC in the southern North Atlantic. In contrast, the Arctic subsurface water at this depth becomes warmer than the surface water, due to the intrusion of the warm Atlantic water. These features are found in both the observation (Fig. 7.1a, c) and in the simulation (Fig. 7.1b, d). As at the surface, the model well reproduces the sharp gradient of the temperature near Newfoundland and the meandering course of the Atlantic water across the North Atlantic into the Nordic seas, considering the coarse resolution of the model. However, the 6ºC isotherm outcrop is located too far eastward, owing mainly to the artificial enlargement of the Nares Strait as already seen at the surface. This leads to an overestimated fresh water flux from the Arctic Ocean and an eastward extension of the cold water pool.

![Fig. 7.2.](image-url) The movement of the Atlantic water in the Arctic Ocean, the Nordic Seas and the northern North Atlantic Ocean (from Karcher et al., 2007). Arrows indicate the cyclonic movement of the Atlantic water.

The observed water temperature in the Arctic Ocean is higher in the Barents Sea and the Eurasian Basin, while lower in the western Basins and the Greenland Sea close to the east coast of Greenland. This spatial pattern is largely consistent with the pathway of the Atlantic water into the Arctic Ocean (Fig. 7.2). The warm and saline Atlantic water enters the Arctic Ocean through two branches. One branch is via the Barents Sea. In this region, the water of Atlantic origin is subject to large negative buoyancy flux due to heat loss and sea ice freezing/melting, and becomes dense. Then this dense water leaves the Barents Seas via the St. Anna Trough, a deep trench connecting the Barents Sea to the Eurasian Basin, into the Arctic Ocean. Another branch of Atlantic origin water enters into the Arctic Ocean with the West Spitsbergen Current (WSC) via the Fram Strait. On the whole, the inflow of Atlantic
water feeds into different density layers of the Arctic Ocean, and joins the cyclonic gyres in the Eurasian Basin and Canadian Basin (Rudels et al., 1994; Schauer et al., 1997). After recirculating in different basins (Schlosser et al., 1995; Smethie et al., 2000), the mid-depth Atlantic water exits the Arctic Ocean via Fram Strait along the eastern Greenland coast.

However, due to the coarse resolution of the model, the St. Anna Trough is not adequately represented in the model, which leads to reduced Atlantic water flow into the Eurasian Basin through the trough. Therefore, the simulated temperature is higher in the Barents Sea, while lower in the Eurasian Basin compared with the observations. Consequently, less cold water exits the Arctic Ocean, resulting in a larger region of higher temperature and a smaller region of low temperature (< 1.2°C) in the western Greenland Sea (Fig. 7.1 c and d).

The simulated temperature pattern at the depth of 1000 m (Fig. 7.1f) also reproduces the major observed features (Fig. 7.1e). The warm tongue along the eastern boundary of the Atlantic Ocean, referred to as the Mediterranean Water tongue, extends northward along the eastern boundary, westward all the way to the western boundary in the interior of the Atlantic Ocean, and vertically over several hundred meters (Lozier et al., 1995; Jia et al., 2007). The spatial extent is well captured by the model. However, the tongue is colder (about 2°C lower for the core) in the simulation than in the observation, possibly due to the model’s inadequate representation of the outflow through the Strait of Gibraltar. Due to the coarse resolution of the model, the sharp temperature gradient around Newfoundland is not well reproduced, which results in the overestimate of temperature in that region and consequently the overestimate (about 1°C) in the subpolar gyre and Labrador Sea. Further north, the temperature is overestimated by about 0.8°C in the Arctic Proper, and by about more than 1°C in the Greenland Sea compared with the observations (cf. Fig. 7.1e and f). The overestimated temperature at the depth of 1000 m together with the underestimated water temperature at the surface as well as the depth of 200 m implies that the mixing in the upper ocean of the Arctic is too weak in the simulation, especially during the summertime.

At the depth of 3000 m (Fig. 7.1g, h), the water temperature is more spatially uniform whether in the Arctic Ocean or in the North Atlantic Ocean. It is about 2–4°C in the North Atlantic Ocean, and below 0°C in the Arctic Ocean with relatively warm water in the Canadian Basin and cold water in the Eurasian Basin. These features are well captured by the model, except the water bordering Newfoundland is still 1–2°C higher than the observation.

7.1.2. The influence of AISM on temperature

Although the ISM is applied in Antarctica, it causes changes in the hydrography of the North Atlantic and Arctic Oceans. This impact is investigated through the temperature difference at the same depth levels as above, that is, at the depths of 5 m, 217 m, 1033 m and 3275 m in the model.
Figure 7.3 displays the influence of including AISM on the potential temperature at different levels in the Arctic Ocean and North Atlantic Ocean. In general, inclusion of AISM has little influence on the temperature in the Arctic Ocean. Prominent changes are only found in the deep waters of the Greenland Sea and Nordic Seas (Fig. 7.3c), where the temperature becomes systematically higher. The temperature increase in these regions is up to 0.8°C, even larger than that in the North Atlantic Ocean at the same depth. In the other layers, the changes in temperature are generally much larger in the North Atlantic Ocean than in the Arctic Ocean.

In the North Atlantic Ocean, the water generally becomes warmer after inclusion of AISM (Fig. 7.3), with the most pronounced change in the subsurface and intermediate depth water in northern North Atlantic Ocean (Fig. 7.3b & c). The warming is relatively small (<0.1°C) at the surface. Large warming (>0.4°C) is only found in the region between 50–60°N (Fig. 7.3a), where the temperature increases up to 0.6°C after the inclusion of AISM. Going deeper to the depth of 217 m (Fig. 7.3b), the warming is much more significant. Not only the warming area becomes larger (extending southward from the region of 50–60°N), but also the magnitude increases, with the warming core increased up to more than 1.0°C. Going to the depth of 1033 m (Fig. 7.3c), the warming becomes weaker compared with that at the depth of 217 m, only up to about 0.4°C. Further down to the depth of 3275 m (Fig. 7.3d), the warming is quite small, generally less than 0.1°C.

The warming of the North Atlantic Ocean was found in observations and other simulations. Levitus et al. (2000, 2005) reported a net warming in the Atlantic Ocean since 1950s derived from a computation of the oceanic heat content. Levitus et al. (2000) also found substantial warming at the depth of 300–1000 m in the world ocean and in depth greater than 1000 m in the North Atlantic. Parrilla et al. (1994) observed that waters in the subtropical North Atlantic had warmed since 1950s. Using a simple atmosphere-ocean coupled model, Weaver et al. (2003) found warming in the North Atlantic through adding freshwater near Antarctica and just north of the ACC. These features are generally captured by our study.

The warming in the North Atlantic (Fig. 7.3a) together with the cooling in the Southern Ocean (Fig. 5.2a) forms the so-called “bipolar seesaw” in temperature anomaly (Broeker, 1998), i.e. warming in one hemisphere while cooling in the other. Such a bipolar seesaw has been found by a number of authors using proxy temperature estimates derived from ice cores in Antarctica and Greenland (e.g. Blunier and Brook, 2001). The warming in the North Atlantic could heat the overlying atmosphere, and hence possibly increases the surface air temperature and precipitation (Stouffer et al., 2007), and make the inter-tropical convergence zone (ITCZ) shift northward (Broccoli et al., 2006; Ma and Wu, 2011).
7.2. Salinity

In this section, we compare the observed and simulated annual mean salinities in the North Atlantic and Arctic Oceans. As for the temperature comparison in section 7.1, the simulated results are obtained from the simulation with AISM, and the observations are taken from the 1° latitude \( \times 1° \) longitude grid data of the WOCE 98 Atlas (http://www.ersl.noaa.gov/psd/data/grided) at the Earth System Research Laboratory. We also choose the same four layers for salinity comparisons.

Figure 7.4 compares the simulated and observed annual mean salinity for the selected four layers in the Arctic and North Atlantic Oceans. It is seen that the main features of SSS are fairly well reproduced by the model (cf. Fig. 7.4a and b). The south-north SSS contrast is both prominent in the observation and in the simulation. The SSS is over 37 psu in the subtropics, but is less than 30 psu in the polar ocean. In the south, the salinity maximum in the northern subtropical Atlantic is due to the excess of evaporation over precipitation. Its position is in good agreement in the simulation and in the observation, although its magnitude is a little bit larger (by 0.5 psu) in the simulation.

The simulated surface water in the Arctic Ocean is too saline in most regions compared with the observations. The higher SSS can be partly due to the inadequate representation of the freshwater discharge in the model. For example, the observed salinity is quite low near the coast of the East Siberia Sea, as a result of the freshwater outflow from the Kolyma River, which is not well reproduced possibly due to the inadequate representation of the river discharge. On the other hand, the observed SSS might be underestimated, since the WOCE observation is biased toward summer observation in the polar region (Tomczak and Godfrey, 2003).
Due to the coarse resolution, the boundary currents, such as the East Greenland Current and the Labrador Current, are too broad and too weak in the model. The low salinity in these regions might result from the melting of the overestimated sea ice cover (see Fig. 7.10). In addition, the artificial enlargement of the Nares Strait in the model allows more fresh water to enter from the Arctic Ocean, resulting in a much lower simulated salinity in the Labrador Sea and the subpolar gyre, and eastward extension of the low salinity pool (cf. Fig. 7.4a and b). This would in turn affect the SSS in the Greenland Sea and Barents Sea, where we find a considerable underestimation of SSS.

The overall salinity pattern at the depth of 200 m is well reproduced by the model (cf. Fig. 7.4c, d). Fed by the saline water from the Gulf Stream, the maximum salinity in the northern subtropical Atlantic shifts to the west at this depth. In the northern North Atlantic, both spatial pattern and magnitude of the salinity are well featured by the model. In the Arctic Ocean, the observed salinity is higher in the west while lower in the east. This spatial difference is well captured by the simulation, although the simulated salinity is slightly overestimated in the Canadian Basin, and slightly underestimated in the Eurasian Basin. The underestimate is likely again due to the inadequate representation of the St. Anna Trough connecting the Barents Sea and the Eurasian Basin because of the coarse resolution of the model.

At the depth of 1000 m and 3000 m (Fig. 7.3e–h), the simulated salinity is in rather good agreement with the observation. In 1000 m (Fig. 7.3e, f), the maximum salinity is found within the Mediterranean Water tongue extending from the Strait of Gibraltar to the European coast to the west. Near the source of this highly saline water, the core has a maximum salinity of 36.2 psu in the simulation, very close to the observed 36.4 psu. Good agreement between simulation and observation is also found in the Arctic Ocean, with the maximum difference of 0.1 psu. In 3000 m (Fig. 7.3g, h), the salinity is quite similar in both
the North Atlantic Ocean and the Arctic Ocean, in the range of 34.8–35 psu. This is rather well reproduced by the model, with the difference generally less than 0.03 psu.

### 7.2.2. The influence of AISM on salinity

Figure 7.5 shows the influence of the addition of AISM on the salinity in the North Atlantic Ocean and Arctic Ocean. In contrast to the temperature field which was not much affected by AISM in the Arctic Ocean (Fig. 7.3), the salinity field is considerably changed (Fig. 7.5). The water becomes systematically more saline below 217 m, whereas both freshening and salt enrichment are found above this depth.

The salinity decrease is mainly located in the region from the North Pole to the New Siberian Islands. The freshening is the most pronounced at the surface, up to 0.08 psu (Fig. 7.5a). It then decreases with depth to about 0–0.02 psu at the depth of 217 m (Fig. 7.5b). Below 217 m, the water becomes more saline in almost the whole region. At the depth of 1033 m, the salt enrichment generally becomes small, less than 0.02 psu in the Arctic Ocean, while up to 0.04–0.06 psu in the Greenland Sea. Further down to the depth of 3275 m, the change in salinity becomes even smaller, generally less than 0.02 psu, both in the North Atlantic Ocean and the Arctic Ocean.

**Fig. 7.5.** Climatological annual mean salinity differences (with–without AISM) at the depth of (a) 5 m, (b) 217 m, (c) 1033 m, and (d) 3275 m.

In the North Atlantic Ocean, inclusion of AISM has a more pronounced effect (>0.02 psu) in the upper ocean, with saline enrichment in the northern part and freshening in the tropical region (Fig. 7.5). The salt enrichment is most significant at the surface layer (Fig. 7.5a), up to 0.3 psu in the northern North Atlantic. It becomes weaker with depth. For example, the maximum salinity increase is up to 0.1 psu at the depth of 217 m (Fig. 7.5b), up to 0.08 psu
at the depth of 1033 m (Fig. 7.5c), and less than 0.02 psu at the depth of 3275 m (Fig. 7.5d). In contrast, the freshening in the tropical region is more significant at the depth of 217 m (>0.8 psu). It decreases upward and downward of this level.

7.2.3. The influence of AISM On density

In cold environments, density is mainly determined by salinity. Therefore, the change of density generally follows that of salinity, as already seen in the Southern Ocean (see Chapter 5). Is this the same on the northern hemisphere? In the following, we investigate the impact of the AISM on the density in the North Atlantic and Arctic Oceans. In the Arctic Ocean, the water becomes denser, except for the surface layers (0-217 m) between the North Pole to the New Siberian Island (Fig. 7.6), following the changes of salinity described in section 7.1.3.

In the North Atlantic Ocean, the change in density also follows much of the salinity (cf. Figs. 7.5 and 7.6). For example, the water generally becomes denser in the northern North Atlantic Ocean and lighter in the tropical region, with the density increase most significant at the surface in the former region and the density decrease most pronounced at the depth of 217 m in the latter region. However, there are exceptions in some regions, where the density change follows the temperature change. For example, the water at the depth of 217 m between 30–60ºN becomes saltier (Fig. 7.5b) and lighter (Fig. 7.6b), due to the large warming (Fig. 7.3b). Another example is the density decrease in the Mediterranean Water tongue at the depth of 1033 m (Fig. 7.6c), where the warming is also pronounced (Fig. 7.3c). Further down in 3275 m, the water becomes lighter in the whole North Atlantic (Fig. 7.6d), again as a result of the warming (Fig. 7.3d).

7.3. Mixed layer depth

7.3.1. Comparison with the observation

Figure 7.7 compares the observed and modeled climatological MLDs in the North Atlantic Ocean and Arctic Ocean. The modeled MLD is calculated from the simulation with AISM, and the observed MLD is obtained from http://www.lodyc.juessieu.fi/~cdblod/mld.html, determined by de Boyer Montégut et al. (2004) based on individual profiles. As in the Southern Ocean, the MLDs show strong seasonality in the North Atlantic and Arctic Oceans, with the deepest mixed layers in winter (Fig. 7.7). The mixed layer shallows rapidly in spring, stays thin during summer and gradually deepens in autumn to reach the maximum again in winter. This cycle is well captured by the model (Fig. 7.7).
Fig. 7.6. Climatological annual mean potential density differences (with-without AISM) at the depth of (a) 5 m, (b) 217 m, (c) 1033 m, and (d) 3275 m.

Fig. 7.7. Climatological MLD (unit: m) reconstructed from observations (upper; data available at http://www.lodyc.juissieu.fr/~cdblod/mld.html) and computed from the simulation with AISM (bottom) in winter (February), spring (May), summer (August) and autumn (November).

The observed winter MLDs are mostly larger than 100 m in the North Atlantic Ocean, with the deepest mixed layers occurring in the regions of deep water formation poleward of 50°N (Fig. 7.7a). These features are basically reproduced in the simulation, although the spatial extent of deep mixed layers (>100 m) is relatively smaller. The simulated MLD is 850 m in the Greenland-Iceland-Norway (GIN) Seas, which is 740 m in the observation of de Boyer Montégut et al. (2004). Considering a standard deviation of 60 m in the observed winter MLD (Sallée et al., 2010), the simulated value might be acceptable. The location of the maximum MLD in the GIN Seas is also in good agreement with the observations from Dickson et al. (1996).

The Labrador Sea is another deep convection region in the winter North Atlantic Ocean. The climatology MLD is usually more than 300 m (de Boyer Montégut et al., 2004), but it can also be up to more than 1000 m in some years (Dickson et al., 1996; Lavender et al., 2002). In contrast, our simulated MLD is much smaller, largely less than 100 m, owing to the substantially overestimated winter sea ice in this region. As shown below (section 7.4.2),
in the observation (Fig. 7.10b), the Labrador Sea is not fully covered by sea ice in winter, and the sea ice concentration is high only in a narrow band along the Labrador coast. However, in the simulation, it is totally covered by sea ice (Fig. 7.10a), due to the artificial enlargement of the Nares Strait. The presence of sea ice in the simulation isolates the ocean from the cold air in winter, and consequently prevents the vigorous mixing in this region.

In the Arctic Ocean, observations of MLD are sparse as shown in Fig. 7.7 (top). However, it is seen that the mixed layer is deeper in winter and shallower in summer as in the other regions, e.g. the North Atlantic Ocean and the Southern Ocean. The model basically reproduces the observed regions of large MLDs. However, the MLDs are underestimated, especially in summer, due to the weak mixing as shown in section 7.1.

### 7.3.2. Impact of AISM on MLD

In general, inclusion of AISM has a very limited impact on the MLD in most areas of the North Atlantic and Arctic Oceans (Fig. 7.8). Systematic changes in the MLD are mainly found in the northern North Atlantic, along the Greenland-Iceland-Scotland (GIS) ridge, a crucial region for European climate. In this region, the mixed layer becomes systematically deeper after inclusion of AISM, with an annual mean up to several tens meters (Wang and Beckmann, 2007). The intra-seasonal variability of the deepening of the mixed layer is substantial. It is small in spring and summer, but large in autumn and winter, with a maximum deepening of 40 m in autumn, and up to 200 m in winter (Fig. 7.8). The increase of the MLD in the northern North Atlantic Ocean is due to the changes in hydrography. As shown above, the upper ocean becomes more saline and denser (Fig. 7.5-7.6) after inclusion of AISM. This destabilizes the water column and leads to deep ocean convection (Hu et al., 2004).

![Fig. 7.8. The MLD differences (with-without) in February (winter) and November (autumn).]

### 7.4. Arctic sea ice

#### 7.4.1. Sea ice extent

Figure 7.9 shows the seasonal evolution of the Arctic sea ice extent from the simulations with and without AISM and from the observations derived from the SSMR-SSM/I satellite data. The evolution of Arctic sea ice is very well captured by the model, with the annual mean relative error less than 2%. Sea ice reaches its maximum extent in March, then retreats and eventually reaches its minimum extent in September. The simulated maximum and minimum sea ice extents are overestimated by 2% and 12%, respectively. Compared with the simulation in the Southern Ocean, amplitude and phase are in much better agreement with the observations.
Fig. 7.9. Climatological annual cycle of sea ice extent in Arctic: red – ISP, blue – REF, and black – observations. Note the sea ice extents in the simulation with and without AISM in the Arctic are nearly identical.

Fig. 7.10. Climatological monthly mean Arctic sea ice concentration: a) March sea ice concentration simulated with AISM, b) March sea ice concentration derived from SSMR-SSM/I, c) September sea ice concentration simulated with AISM, and d) September sea ice concentration derived from SSMR-SSM/I.
7.4.2. Sea ice concentration

The features of the winter sea ice concentration (Fig. 7.10, top) are well reproduced by the model, for example, the very high concentration in the whole Arctic Ocean, the Hudson Bay and the Baffin Bay, and the large gradient near the ice edge. Nevertheless, there are some deficiencies. Firstly, sea ice extent is noticeably overestimated in the Labrador Sea and Davis Strait, due primarily to the artificial enlargement of the Nares Strait. Secondly, the Odden Ice Tongue in the central Greenland Sea is not reproduced. Thirdly, the open water area in the west of Spitsbergen is missing from the simulation. The last two deficiencies are partly due to the coarse resolution of the model (T2005), but are also possibly owing to the underestimated SSS in these regions as shown in section 7.2, which results in more ice production. Finally, the sea ice concentration is markedly overestimated near the ice edge of the Barents Sea, due to the underestimated mixing and SSS in this region.

The main features of sea ice concentration in summer (Fig. 7.10, bottom) are also fairly well reproduced by the model. However, although the ice edge is very well simulated, the modelled sea ice concentration is noticeably lower than the observation in the Arctic. Since the derivation of sea ice concentration from satellite raw data leads to large uncertainties (inaccuracy of 15–20%) in summer (Comiso et al., 1997; Cavalieri et al., 1999), the quality of the simulation remains unclear. In the Baffin Bay, sea ice extent is greatly overestimated, again possibly due to the artificial enlargement of the Nares Strait.

7.4.3. Sea ice thickness and ice drift

Figure 7.11a shows the annual mean sea ice thickness averaged over the integration period. The pattern of the ice thickness field agrees reasonably well with that observed by Bourke and McLaren (1992). Thicker ice is situated north of Greenland and the Canadian Archipelago and thinner ice off the Siberian coast. The simulated maximum ice thickness is 5 m, consistent with the maximum sea ice thickness derived from ICESat by Kwok and Cunningham (2008), and in the range of other studies (e.g., Gerdes and Köberle, 2007). In the Beaufort Sea, sea ice thickness decreases following the anti-cyclonic sea ice motion (Fig. 7.11b), agreeing well with the deduced ice thickness from satellite altimeter (Laxon et al., 2003). Large ice thickness is also found along the east coast of Greenland following the Transpolar Drift (Fig. 7.11b), again in agreement with the deduced ice thickness by Laxon et al. (2003). As discussed above, sea ice concentration is overestimated in the Barents Sea due to the underestimated vertical mixing in the model. This also affects the simulated sea ice thickness, resulting in overestimate of ice thickness in this region.

The Arctic sea ice motion (Fig. 7.11b) features an anticyclonic gyre in the Beaufort Sea and a Transpolar Drift Stream (TDS). This pattern agrees with the observed drift of Arctic buoys (Rigor, 1992; Pfrirman et al., 1997) and the observation from the SSM/I data (Emery et al., 1997). Sea ice is exported from the Arctic Ocean into the North Atlantic through the Fram Strait and the Nares Strait. The ice transport via Fram Strait accounts for a large fraction of total freshwater transport into the Nordic Seas (Aagaard and Carmack, 1989). Its variability affects the deep ocean circulation in the Atlantic Ocean (Komuro and Hasumi, 2007). The ice flux through this strait is estimated to be 3,209 km$^3$ year$^{-1}$ during 1956–2000 in our simulation, comparable to the 1,520 to 3,600 km$^3$ year$^{-1}$ obtained by other studies (e.g., Aaggard and Carmack, 1989; Vinje et al., 1998; Martin and Wadhams, 1999; Vinje, 2001; Kwok et al., 2004). Ice flux through Nares Strait is simulated to be 2,169 km$^3$ year$^{-1}$ in this study, which is tremendously larger than the other export estimates of 110 ± 50 km$^3$ year$^{-1}$ (Sadler, 1976), 135 km$^3$ year$^{-1}$ (Agnew, 1998), 130 km$^3$ year$^{-1}$ Kwok (2005), and the large ice outflow of 254 km$^3$ in 2007 (Kwok et al., 2010), due to the artificial enlargement of the Nares Strait in the model.
7.4.4. Impact of AISM on the Arctic sea ice

Inclusion of the AISM has a very limited influence on the Arctic sea ice, consistent with the small changes in the Arctic SST (Fig. 7.3a). As seen in Fig. 7.9, the simulated annual cycles of sea ice extent with and without AISM are almost identical. Inclusion of AISM decreases the sea ice volume by $2.7 \times 10^4 \text{ km}^3$, which only accounts for about 0.2% of the annual mean volume, much smaller than the 11% increase in the Antarctic. Addition of AISM also has little effect on the sea ice concentration and sea ice thickness. Relatively large differences (up to 5%) in winter (Fig. 7.12) are found along the ice margins, where thin ice reacts faster to changes (Walsh and Johnson, 1979). Sea ice concentration decreases south of Svalbard and in the Labrador Sea, while increases north of Iceland. These changes generally correspond well to those of temperature (Fig. 7.3a), the warmer the water, the smaller the sea ice concentration. The dipole change character of sea ice cover in the Greenland Sea and the Labrador Sea has been found by Brauch and Gerdes (2005), who attributed the change to the North Atlantic Oscillation (NAO). Although much smaller, the addition of AISM also induces dipole characteristics in the sea ice cover, both in concentration and thickness. On the whole, the decrease and increase in sea ice concentration tend to balance each other, resulting in the quite small changes in the sea ice extent (Fig. 7.9).

The thinning of the Arctic sea ice has been documented from observations, for example, by Rothrock et al. (1999, 2008), and Kwok and Rothrock (2009). It has been attributed to the North Atlantic Oscillation (Hilmer and Jung, 2000), the Arctic Oscillation (Rigor and Wallace, 2004), the Pacific inflow (Shimada et al., 2006), and sea ice dynamic and thermodynamic effects (Nghiem et al., 2007). Inclusion of AISM generally leads to thinner Arctic sea ice. However, the thinning is negligibly small, for example only up to 4 cm in the region from the North Pole to the Barents Sea and in the Fram Strait. This might imply that addition of AISM has little influence on the Arctic sea ice thinning during our integration period.

7.5. Explanations for remote responses to the AISM

The addition of AISM not only induces local changes in the Southern Ocean (see Chapter 4-6) but also stimulates remote changes in the southern subtropical and tropical oceans and the North Atlantic (see above). In the following, we give detailed explanations for the remote changes.
7.5.1 **In the southern subtropical and tropical oceans**

Figure 7.13 shows the latitude-depth changes of zonally averaged annual mean salinity and temperature (°C) between the simulations with and without AISM (86th year). It can be seen that addition of the AISM causes cooling and substantial freshening in the southern subtropical and tropical oceans. This can be simply explained as the spreading of the SSS anomaly with the surface current, since the overall surface current in the southern hemisphere is northward (Fig. 7.14). Due to its lower density, positive subsurface freshwater anomalies quickly rise to the surface of the Southern Ocean (within two years, Fig. 7.14). This low salinity waters cannot be contained in the source region due to the geometry of the Southern Ocean (Seidov et al., 2005). They leave the Antarctic Marginal Seas because of the divergence of the surface currents, caused by the surface westerly wind surrounding the Antarctic continent and the anomalous surface currents resulting from the intense local freshwater perturbation (Stouffer et al., 2007). Then the surface gyres transport the freshwater anomaly northward into the southern subtropical and tropical oceans, resulting in cooling and freshening there, with more pronounced signal in the Atlantic Ocean (Fig. 7.14b). Due to the subduction (Fig. 7.15) along the isopycnals, the surface cooling and freshening penetrate to depth (around 1000 m) in the mid-latitude of the Southern hemisphere (Fig. 7.13).
7.5.2 In the North Atlantic Ocean

In the North Atlantic Ocean, we have shown that warming and salt enrichment dominate the top layers, especially in the high latitude (see section 7.1 & 7.2, and Fig. 7.13), whereas cooling and freshening are dominant in the Southern Ocean surface layers (Chapter 5, and Fig. 7.13), the so-called bipolar seesaw in SST (e.g., Crowley, 1992; Stocker et al., 1992, 2007) and in SSS.

Two hypotheses have been proposed to explain such anti-phase temperature relationship between the two hemispheres. One is related to changes in the strength of the thermohaline circulation (THC; Crowley, 1992; Stocker et al., 1992), and the other is concerned with changes in deep ocean ventilation (Broecker, 1998, 2000). The THC, a global conveyor belt, involves relatively warm surface water flowing northward and relatively cold deep water flowing southward in the Atlantic Ocean. A large amount of heat is thus transported toward the northern high latitudes. Consequently, the northern hemisphere warms by taking heat from the southern hemisphere (e.g., Crowley, 1992). Any changes in the strength of the flow would lead to warming of one hemisphere while cooling the other. Thus, a strengthening of the flow would tend to warm the northern hemisphere while cool the southern hemisphere. A weakening of the flow would lead to the opposite effect. The change in the THC strength is related to the deep ocean ventilation in the two hemispheres, where a weakening of the THC would increase the AABW formation and vice versa (Broecker, 1998, 2000). Therefore, changes in the deep ocean ventilation would also lead to changes in the amount of heat released to the upper ocean and overlying atmosphere. A weakening of NADW formation would cool the northern hemisphere, while the increase of AABW would warm the southern hemisphere, such as, during the Little Ice Age (Broecker et al., 1999), a modest cooling in the northern hemisphere about 500 years ago. Since the heat transported northward by the Atlantic THC is mainly ventilated upward through deep-water formation events, these two hypotheses are closely related (Manabe and Stouffer, 1988, 1999; Rind et al., 2001), and both could lead to bipolar seesaw in temperature anomalies (Rahmstorf, 2002).

7.5.2.1 Changes in the overturning

Figure 7.15 shows the meridional overturning streamfunction for the global ocean with and without AISM. There are two shallow cells that lead to upwelling at the equator, driven by the divergence of the Ekman transport due to the easterly trade winds. Most of this water sinks within ±30° of the equator. Ekman divergence in the subpolar zones produces two
adjoined counter-rotating cells. The cell on the northern hemisphere is quite weak because of being opposed by the thermohaline cell intruding from the north. On the southern hemisphere it is called the Deacon Cell, connected at intermediate depths to the thermohaline cell intruding from the North. The southernmost cell is caused by the formation and sinking of AABW. It extends to the sea floor and spreads far north across the equator to 50°N. The northernmost cell is associated with the strong Atlantic cell with a maximum centered at about 1000 m depth. These features are basically in agreement with the other studies (e.g. Danabasoglu and Mcwilliams, 1995; Meissner and Gerdes, 2002; Brix and Gerdes, 2003).

The additional freshwater anomaly in the high southern latitudes noticeably reduces the formation and export of AABW, and slightly increases the formation of NADW (Fig. 7.15). As a result, there would be several effects. First, the reduction of AABW production would retain more cold water at the surface, and more warm water at depth, resulting in cooling of surface waters (Stouffer et al., 2007) and noticeable warming at depth (Goosse and Filchfet, 1999) in the Southern Ocean. Second, the reduction of AABW export would lead to less cold water or more warm water flowing northward, resulting in warm water penetrating northward at the bottom (Fig. 7.13b). Third, the increase of NADW would warm the northern hemisphere (Broecker et al., 1999), especially the North Atlantic.

![Fig. 7.15. Meridional overturning streamfunction in Sv for the global ocean (a) without, and (b) with AISM.](image)

![Fig. 7.16. Northward heat transport in the global ocean with (blue) and without (red) AISM.](image)
7.5.2.2 The northward heat transport

The northward heat transport in the global ocean is shown in Fig. 7.16. It is clearly seen that in the northern hemisphere poleward heat transport increases with the addition of AISM, similar to the result of Weaver et al. (2003) and Stouffer et al., (2007) who added surface freshwater anomalies in the Southern Ocean in their models. The increase in the northward heat transport implies that more warm and saline waters in the tropical region are transported northward into the northern hemisphere, resulting in warming and salt enrichment there. However, the change of the northward heat transport is negligibly small north of 60°N. Therefore, the Arctic Ocean experiences little change in temperature.

7.5.2.3. Inhomogeneous transport

Due to the complex circulation system, the transport of fresh water may be non-uniform. This is revealed by the changes in ocean circulation (Fig. 7.17). There are notable additional northward flows along the western side of the basins in the northern hemisphere, indicating intensification of the western boundary currents. Of the three basins, the intensification is the most significant in the North Atlantic Ocean. The freshwater anomaly in the Atlantic Ocean is mainly across the equator along the western side, most apparent at the depth of 106 m (Fig. 7.17).

The northward transported anomalies of SST and SSS evolve differently. Since the water after the tropical region begins to lose more and more heat due to the strong interaction with the atmosphere, the SST anomaly becomes less and less significant, until finally negligibly small north of 60ºN, as illustrated by the northward heat transport figure (Fig. 7.16). This is consistent with the large warming in the North Atlantic and small temperature changes in the Arctic Ocean (section 7.1.2). On the contrary, the salt remains relatively coherent during the northward transport, thus yielding an increasing large change northward (Fig. 7.5).

![Fig. 7.17. Differences (with-without AISM) of ocean circulation and salinity at the depth of 106 m.](image)

7.5.2.4. The role of meridional heat and salt transport

As shown above, the water mainly becomes warmer and more saline in the northern North Atlantic, but cooler and fresher in the tropical region. Why does the northern North Atlantic
not become cooler and fresher with the spreading of freshwater anomaly like in the tropical region? To answer this question, we note that the change in the local SST and SSS is controlled by a number of factors. For example, the change in SST is controlled by solar radiation, longwave radiation, sensible and latent heat flux, horizontal and vertical heat transport and so on. Among these heat fluxes, the most important change between the case with and without AISM is the meridional heat transport, which may be expressed as

$$\frac{\partial T}{\partial t} = \frac{\partial T_1}{\partial t} - \frac{\partial T_0}{\partial t} \approx -\nu \frac{\partial T_1}{\partial y} + v_0 \frac{\partial T_0}{\partial y}$$  \hspace{1cm} (7.1)$$

where the subscripts 1 and 0 denote the cases with and without AISM, $T$ is zonal mean SST, $\nu$ is zonal mean surface velocity, and $y$ is the coordinate along the meridional direction. It is noted that the mean zonal velocity has to be zero due to the constraint by land. In general in the Atlantic Ocean, $v_1$ and $v_0$ are positive, both $\frac{\partial T_1}{\partial y}$ and $\frac{\partial T_0}{\partial y}$ are positive in the southern hemisphere and negative in the northern hemisphere. Such hemispheric distribution of SST is mainly controlled by solar radiation, which remains constant in the current simulations.

Inclusion of AISM causes surface cooling in the Southern Ocean in the Atlantic sector (Fig. 5.2a), and enhances northward surface flow in the Atlantic Ocean (Fig. 7.16). In such a case, we have $\frac{\partial T_1}{\partial y} > \frac{\partial T_0}{\partial y} > 0$ and $v_1 > v_0 > 0$ on the southern hemisphere. This leads to $\frac{\partial T}{\partial t} < 0$ according to Eq. (7.1), suggesting a cooling effect in the southern hemisphere due to the addition of AISM. On the contrary, in the northern hemisphere, while it remains $v_1 > v_0 > 0$, it is now $0 > \frac{\partial T_1}{\partial y} > \frac{\partial T_0}{\partial y}$ due to the relatively lower SST in the tropical region after inclusion of AISM. Such variations are unlikely to significantly change the overall pattern of SST distribution, for instance, changing the gradient of SST from negative to positive in the northern hemisphere. Under these circumstances, $\frac{\partial T}{\partial t}$ may be positive or negative according to Eq. (7.1). The increase in $v_1$ becomes more significant in the northern North Atlantic Ocean due to the much narrower width. In particular $v_1$ becomes significantly larger around the sites of forming NADW, resulting in a much larger northward heat transport $-\nu \frac{\partial T_1}{\partial y}$. This causes $\frac{\partial T}{\partial t}$ mostly larger than zero, leading to a warming effect in the northern North Atlantic Ocean due to the addition of AISM.

The above analysis shows that addition of AISM causes more warm water transported to the northern North Atlantic, although the temperature of the transported warm water is slightly lower than that without AISM. Such a mechanism is also valid for SSS. Thus, addition of AISM would transport more saline water to the northern North Atlantic, with the salinity of the transported water slightly higher than that without AISM.

### 7.5.3. Time lag of the inter-hemisphere teleconnection

With the addition of AISM, the water becomes warmer and more saline, and the mixed layer becomes deeper in the North Atlantic. It is physically obvious that such teleconnection would have a time lag, since the signals are transported by the current. The time lag can be seen in the temporal evolution of the MLD difference between the southern and northern hemispheres. Figure 7.18 shows the MLD difference at two points. One is in the Southern Ocean close to the continent (Fig. 7.18a), and the other in the northern North Atlantic at the latitude of 66ºN (Fig. 7.18b). To remove the relatively large interannual variations, a five-year low-pass filter is applied. It is seen that the change in the MLD is almost instantaneous along the Antarctic coast after introduction of AISM (Fig. 7.18a), while it is only visible after more than 20 years in the northern North Atlantic at the latitude of 66ºN (Fig. 7.18b), where the mixed layer deepening is no more than 5 meters. Substantial deepening (more than 10 m) is seen after more than 40 years.
Fig. 7.18. Time series of the five-year running mean MLD difference (unit: m) between simulations with-without AISM in the Atlantic Ocean at points (a) 44ºW, 77ºS, and (b) 7ºE, 66ºN. Two passes of the period of 1958–2000 are displayed and separated by the red dashed line. The time series 1–43 represents the first pass running while 44–86 represents the second pass.

The time lag can also be found in the temporal evolution of the differences of temperature and salinity between the simulations with and without AISM in the North Atlantic (Fig. 7.19). The differences of temperature and salinity are small in the first pass, especially in the first 30 years. They are gradually accumulated in the integration period (86 years), particularly during the second pass in the northern North Atlantic. For example, it takes about 30 years to develop a 0.2°C temperature difference at the beginning of the integration period, but it takes less than 20 years to yield another increase of this magnitude. This accelerated change is a general feature of the simulation and also applies to the development of salinity differences (Fig. 7.19b). These results suggest that it would take 35–40 years for the AISM signal to be noticeable in the North Atlantic, close to the appearance time (more than 40 years) of noticeable mixed layer deepening in the northern North Atlantic (Fig. 7.19b).

Fig. 7.19. Zonal mean of (a) temperature difference and (b) salinity difference in the northern North Atlantic Ocean with time. The ordinate combines the two 43-year passes of 1958–2000, separated by the black dashed line.
7.6. Discussion

7.6.1 SSS contrast in the North Atlantic Ocean and North Pacific Ocean

The inter-ocean salinity contrast in the North Atlantic and North Pacific has been shown to play a key role in maintaining the global ocean THC (e.g., Broecker, 1991; Schmitz, 1995; Gordon, 1996, 2001; Haug and Tiedemann, 1998; Seidov and Haupt, 2002, 2003, 2005). Its development is believed to be attributed to the formation of the Isthmus of Panama (Zaucker and Broecker, 1992; Haug and Tiedemann, 1998; Haug et al., 2004), a narrow strip of land lying between the Caribbean Sea and the Pacific Ocean and linking North and South American. Before the closure of the Isthmus of Panama, Pacific surface waters flowed into the Atlantic Ocean, and the salinity in the two oceans were roughly in balance (Haug et al., 2004). The gradual closure of the Isthmus of Panama from 13 to 1.9 million years ago (Keller et al., 1989; Duque-Caro, 1990; Collins et al., 1996) restricts the water mass exchange between the Pacific and the Atlantic oceans.

The AISM might be a contributor to the inter-ocean salinity contrast in the two basins. As shown in section 7.2.2, the SSS generally becomes higher in the North Atlantic Ocean after inclusion of AISM (Fig. 7.5a). Salt enrichment is also found in most of the North Pacific (Fig. 7.15b, 7.16), but its magnitude is much smaller because a larger portion of the freshwater anomaly is leaking into the Atlantic Ocean (Stouffer et al., 2007).

7.6.2 Surface and subsurface freshwater anomaly

Northern freshwater anomalies are widely discussed with numerical models (e.g., Manabe and Stouffer, 1995; Rind et al., 2001; Stouffer et al., 2006) due to the importance of NADW for maintaining the THC. The southern freshwater anomaly also attracts more and more attention in recent years (e.g., Goosse and Fichefet, 1999; Seidov et al., 2001; 2005; Stouffer et al., 2007; Ma and Wu, 2011), due to the unique unbounded geometry of the Southern Ocean. The freshwater anomalies in these studies are usually applied at the ocean surface whether in the high north or south. In contrast, in this study they are introduced at the subsurface (at 217 m or greater depth) as a lateral boundary. In the framework of freshwater anomaly experiments, what is the difference between this study and the others? In the following, we elucidate this issue (hereafter surface freshwater anomalies referred to SFA, while subsurface freshwater anomalies referred to SubFA).

The main difference between these two kinds of anomalies seems to be the depth (surface vs. subsurface) of the freshwater injection. But at the same time, the SubFA is coupled to a temperature anomaly, resulting in significant changes in the hydrography and circulation in the Southern Ocean (see Chapter 4 and 5). In the SFA experiment, an external source of freshwater is added at the surface of a designed domain through an additional freshwater flux (e.g., Stouffer et al., 2007; Ma and Wu, 2011), or a low salinity signal (e.g., Seidov et al., 2001). The relatively less dense water cannot sink to depth, and thereby remains at the surface as a cap in the SFA study. In this study (SubFA), the less dense water introduced at the subsurface layer tends to rise to the surface due to its lower density and then remains near the surface. Unlike in the experiment of SFA, e.g., Stouffer et al. (2007), where the Southern Ocean vertical overturning has fully ceased, there remains noticeable vertical circulation in the vicinity of the Antarctic coast in this study (Fig. 7.15), maintained by the sinking of salt water during sea ice formation and rising of the melted ice shelf water. To support this process, more saline deep water is required to flow southward and upward to compensate (Fig. 7.15), thus resulting in higher salinity at depth in the Southern Ocean.

In terms of temperature, both cooling near the surface and warming at depth are also found in the Southern Ocean in the study of SFA, e.g., Stouffer et al. (2007) and Ma and Wu (2011), and in our study (SubFA). This is because the added surface freshwater inhibits the local ocean convection and slows the overturning, which results in unanimously warming in
the deeper waters. In our study, the changes in the water temperature are also caused by the ice shelf-ocean interaction. This interaction needs continuous absorption of heat from the adjacent ocean, thus causing cooling of the surrounding waters (Fig. 5.2). Due to the rising of the melted ice shelf water, more WDW flows southward, resulting in warming at depth in the Southern Ocean. Therefore, the interaction between the ice shelf and ocean is an engine, stimulating changes in the global ocean.

7.7. Summary

In this chapter, we examined the model skill by comparing the simulated and observed temperature, salinity, MLDs and sea ice in the North Atlantic and Arctic Oceans. On the whole, the model successfully reproduces the main features of the sea ice-ocean system. It is noteworthy, however, that the artificial enlargement of the Nares Strait in the model causes an underestimation of temperature and salinity in the Hudson Bay, Baffin Bay and Davis Strait. This allows for a much larger than observed ice flux through this strait, thereby leading to a significant overestimation of the sea ice cover and underestimating the MLDs in the Labrador Sea.

We also investigated the impact of adding AISM on temperature, salinity, the MLD, and the ocean circulation for the northern hemisphere, with the focus on the North Atlantic Ocean and the Arctic Ocean. Although the freshwater from the ISM is locally added in Antarctica, its influence tends to propagate into the whole northern hemisphere, with the most significant impact on the North Atlantic and a relatively weaker impact on the Arctic. The water generally becomes warmer, more saline and denser in the North Atlantic Ocean with the addition of AISM. The most significant warming occurs in the subsurface of about 217 m deep, up to 1°C (Fig. 7.3b). The maximum salinity increase is at the surface, up to 0.2 psu (Fig. 7.5a). The mixed layer can be deepened up to several tens meters in the annual mean, and over 100 m or even 200 m along the Greenland-Scotland-Iceland ridge in winter.

The signal of AISM is likely to be transported with the surface current, similar to that of Stouffer et al. (2007). The additional freshwater from AISM tends to enhance the northward spreading of the surface water. As a result, more warm and saline water is transported from the tropical region to the North Atlantic, resulting in warming and salt enrichment there. However, there are big differences between the surface freshwater anomalies studies (e.g., Stouffer et al., 2007) and ours, due to the different depths for adding the freshwater anomaly, and in particular, the production of freshwater in our study needs to absorb heat from the adjacent ocean. These differences lead to noticeable differences in the circulation and hydrographic changes in the Southern Ocean. Therefore, the simple surface freshwater anomaly experiments should not be taken as the case of freshwater anomaly experiments for the ice shelf-ocean interaction.

It would be a natural consequence for the inter-hemispheric teleconnection between the AISM and the northern hemisphere oceans to have a decadal time lag, if the signals are transported by currents. Our results indicate that it would take about 30–40 years to establish a systematic noticeable change in temperature, salinity and MLD in the North Atlantic Ocean (Figs. 7.17 and 7.18). However, teleconnections may also depend on processes in the atmosphere, the atmosphere-ocean coupling, and wave propagation. The results analyzed here therefore must be treated with caution. Although there are a variety of important features worth of further study, our attention mainly focused on the impact on the North Atlantic, which seems to be the region on the northern hemisphere most affected by AISM.
Chapter 8 Discussion and Conclusions

8.1. Discussion

8.1.1. Bipolar seesaw in sea ice

In Chapter 7, we have seen that the additional freshwater anomaly induces the so-called bipolar seesaw in temperature anomaly. In fact, this bipolar seesaw is also found in sea ice. With the increase of AISM, the sea ice extent and volume increase in the Antarctic, while they decrease in the Arctic. As far as I know, this feature has not been revealed in the freshwater anomaly experiments before, since those studies have not touched the changes in sea ice (e.g., Seidov et al., 2005; Stouffer et al., 2007) or the two polar regions (e.g., Hellmer, 2004). Opposite changes in sea ice were also earlier found by Wu et al. (1999), who showed that the changes in snowfall (no-snow or doubling snow) under global warming would result in ice thickening in the Antarctic, but thinning in the Arctic, due to the relative difference of snow and ice thickness in the two polar regions. It is possible that the increasing trend in the Antarctic (Cavalieri and Parkinson, 2008) and decreasing trend in the Arctic (Parkinson and Cavalieri, 2008) are, at least partly, results of such bipolar seesaw anomaly caused by the increasing AISM.

8.1.2. Mass balance of the Antarctic ice sheet

Mass balance of an ice sheet can be estimated between the mass gain and loss. In Antarctica, there is little surface melting, even near the coast during summer due to the cold conditions (Rignot and Thomas, 2002). Therefore, mass gain is primarily by snow accumulation, and mass loss is mainly by basal melting and iceberg calving. Knowing these three quantities, the mass balance of the Antarctic ice sheet can be well evaluated. However, the large uncertainty in these three quantities makes it very difficult to accurately estimate the mass balance. It is therefore not surprising to find conflicting results in the references. For example, the recent estimate of Wingham et al. (1998; 2006) and Rignot and Thomas (2002) place the Antarctica ice sheet as a source of ocean mass, while the study of Davis et al. (2005) puts the Antarctica as a sink of ocean mass.

Snow accumulation has been estimated by a number of authors. Most of the estimates are in the range of 1749 and 2200 Gt year\(^{-1}\) as shown by Jacobs et al. (1992). On the other hand, a quite low value of 746 Gt year\(^{-1}\) was obtained by Rignot and Thomas (2002), possibly because they did not include the Antarctic Peninsula. In addition, a slightly higher value of 2918 Gt year\(^{-1}\) was estimated by Wingham et al. (2006) from satellite radar altimetry data. With more existing observations, Monaghan et al. (2006) found changes in the Antarctic snowfall had been insignificant since the International Geophysical Year, although there is a strong tropospheric warming signal over Antarctica during winters since the early 1970s (Turner et al., 2006). They estimated the mean snow accumulation to be about 2548 Gt year\(^{-1}\) for the grounded ice sheets, in the range of most estimates for the whole ice sheet. Since they include more existing observations than others, their estimate tends to be more reliable. In their study, snowfall increases from the interior of the continent, and most of the snowfall is near the periphery of the continent (Fig. 2 of Monaghan et al., 2006). This implies that there might be more snowfall near the coast, including on the ice shelves. In the periphery, snowfall is 600 mm year\(^{-1}\) in East Antarctica, and the maximum is up to 2000 mm year\(^{-1}\). Taking 600 mm year\(^{-1}\) as a conservative estimate produces a 900 Gt year\(^{-1}\) for the snowfall on the ice shelves. Thus, snow accumulation on the whole Antarctic ice sheet is at least 3448 Gt year\(^{-1}\).

Iceberg calving is a large factor in the mass balance of the Antarctic ice sheets. Its estimates have been mainly based on the observations of large icebergs. However, observations have large uncertainty due to calving size, surveying ocean area, calving frequency, and survival time of icebergs (Jacobs et al., 1992). During 1979-2003, Silva et al.
(2006) demonstrated that the mean production of giant icebergs was 1089 Gt year\(^{-1}\), slightly larger than the estimate of 1008 Gt year\(^{-1}\) for the period of 1979-90 by Jacobs et al. (1992). This might be because giant iceberg size is taken as larger than 18.5 km by Silva et al. (2006), instead of larger than 28 km by Jacobs et al. (1992). If taking the value of 1008 Gt year\(^{-1}\) from Jacobs et al. (1992) for small icebergs, the total iceberg calving is estimated to be 2097 Gt year\(^{-1}\), close to the estimate of 2016 Gt year\(^{-1}\) by Jacobs et al. (1992).

The simulated mean basal melting is 106 mSv in this study, which is an overestimate as discussed in Section 4.4. We suggest an amount of 60–70 mSv (see Chapter 4). Assuming a mean ice density of 917 kg m\(^{-3}\) (Jacobs et al., 1996), the freshwater volume due to basal melting is equivalent to 1735–2024 Gt year\(^{-1}\). Thus, we can derive a mass balance of \(-384 \sim -790\) Gt year\(^{-1}\) (-13 \sim -27 mSv) for the Antarctic ice sheet. This means that the Antarctic ice sheet is a source of the ocean mass, in particular the west Antarctic ice sheet, where rapid ice shelf thinning has been observed (Shepherd et al., 2004).

8.1.3. Freshwater budget in the Southern Ocean

The freshwater budget of the Weddell Sea has been analyzed by Timmermann et al. (2001). Here we analyze the freshwater budget for the whole Southern Ocean, taking the latitude 60°S as the northern boundary, and the Antarctica as the southern boundary, as shown in Fig. 8.1.

Water vapor flux across 60°S is southward, being a freshwater input for the Southern Ocean. This can be estimated through different approaches (Giovinetto et al., 1997) knowing the meridional wind and specific humidity; for example, from the moisture flux using numerical reanalysis products (e.g., Masuda, 1990), or estimate the transport divergence using surface and near-surface data (e.g., Baumgartner and Reichel, 1975). Using these two approaches, Giovinetto et al. (1997) got a consistent net atmospheric transport of water vapor across 60°S, about 17.8 kg m\(^{-1}\)s\(^{-1}\), corresponding to 356 mSv. Using the NCEP-NCAR reanalysis products, Cohen et al. (2000) calculated the meridional water vapor transport. From their results (Fig. 1 of Cohen et al., 2000), we can estimate that the water vapor transport is about \(3 \times 10^8\) kg s\(^{-1}\), equivalent to 300 mSv. Using the ERA-40 reanalysis product, Tietäväinen and Vihma (2008) produced a value of 17.7 kg m\(^{-1}\)s\(^{-1}\), equal to about 354 mSv, quite close to the study of Giovinetto et al. (1997). In this study we take the value of 354 mSv for the water vapor transport across 60°S.

![Fig. 8.1. Freshwater budget in the Southern Ocean.](image)
Part of the water vapor, after crossing 60°S, will be deposited on Antarctica as snowfall. As shown in section 8.1.2, snowfall over Antarctica is at least 3448 Gt year\(^{-1}\), equivalent to about 109 mSv freshwater. Thus, the net precipitation \((P-E)\) in the Southern Ocean is about 245 mSv (=354-109).

Iceberg calving is 2097 Gt year\(^{-1}\), equal to about 72.5 mSv freshwater. This flux has large spatial and temporal variability (Silva et al., 2006). Its magnitude is comparable to net precipitation in areas close to much of the Antarctic coastline (Gladstone et al., 2001), and approximately 30% of the total net precipitation (245 mSv) into the Southern Ocean south of 60°S. In addition, its magnitude is comparable with the freshwater due to basal melting of ice shelves (60-70 mSv).

Freshwater is exported out of the Southern Ocean through sea ice transport and ocean advection. Across 60°S, sea ice exports about 22.8 mSv of freshwater and ocean advection transports about 283 mSv freshwater out according to this study. Thus, the Southern Ocean gains freshwater of 377.5 mSv (=245+72.5+60), while loses 305.8 mSv (=283+22.8). As a consequence, the freshwater input is 66.7 mSv larger than the freshwater output across 60°S. This indicates that the Southern Ocean is in a state of freshening. This is supported by the recent observations in the intermediate water (Wong et al., 1999), in the surface water in the Ross Sea (Jacobs et al., 2002), and in the bottom water (Rintoul, 2007; Johnson et al., 2008; Ozaki et al., 2009).

**8.2. Conclusions**

Ice shelves are seaward extension of ice sheets. Their basal melting removes heat from and inputs freshwater into the adjacent ocean, affecting the stability of ice sheets (e.g., Dupont and Alley, 2006) and ocean processes (e.g., Beckmann et al., 1999; Timmermann et al., 2001; Hellmer, 2004; Thoma et al., 2006), especially in Antarctica where the ice shelves cover about 44% of the Antarctic coastline (Drewry, 1983).

In this study, we introduced Antarctic Ice Shelf Melting (AISM) into a global sea ice-ocean model ORCA2-LIM following the approach of BG03 for ISM, which forms our model ORCA2-LIM-ISP. We investigated the sensitivity of ORCA2-LIM to atmospheric forcing data through comparing with the results of Timmermann et al. (2005), conducted sensitivity experiments for three ice parameters, compared the capability of ORCA2-LIM and ORCA2-LIM-ISP models in simulating the hydrography of the Southern Ocean, and assessed the ORCA2-LIM-ISP model in simulating the sea ice-ocean system in the North Atlantic Ocean and in the Arctic Ocean, and investigated the impact of AISM on the global sea ice-ocean system. As an experiment of ISM, this study is the first one following Wang and Beckmann (2007) to include more numbers of Antarctic ice shelves, and to access the ISM impact on a global scale. As an experiment of freshwater anomaly, it is the first one to investigate the changes in global sea ice and ocean system in addition to Ma and Wu (2011).

Our study demonstrated that AISM has significant impact on the sea ice and ocean system. The consistency in the simulations and observations highly indicates that AISM might be a major contributor to the observed trends in temperature, salinity and density in the Southern Ocean. It contributes to the salinity contrast between the North Atlantic Ocean and the North Pacific Ocean, which plays a key role in the global ocean circulation (e.g. Seidov and Haupt, 2002, 2003). It also induces bipolar seesaw patterns in temperature, salinity and sea ice. The other findings are:

- ORCA2-LIM model can reproduce reasonable ocean circulation patterns and realistic large-scale sea ice features in both hemispheres, although has some deficiencies. Our study confirmed that the sea ice coverage in the model is sensitive to the atmospheric forcing data as shown by Timmermann et al. (2005), especially the summer sea ice extent.
- The three ice parameters, thickness of newly formed ice in leads \((h_0)\), compressive strength of ice \((P^*)\), and oceanic boundary layer turning angle \((\theta)\), have different impact
on the investigated properties as summarized in Table 3.2. \( h_0 \) and \( \theta \) are important for the global circulation due to changes in Drake Passage (DP) transport resulting from their variations, but \( P^* \) is not. \( P^* \) is a key parameter for the multi-year ice thickness due to its effect on sea ice dynamics, and therefore also for the seasonal sea ice volume, but it is unimportant for the other properties. \( h_0 \) is an important parameter for the Antarctic sea ice thickness and ice volume, but not for the sea ice extent. \( \theta \) is an important parameter for the Antarctic sea ice, but not for the Arctic.

- The comparisons between the simulations and observations showed that simulations with AISM were generally in better agreement in the Southern Ocean. First, the additional AISM significantly improves the simulated hydrography of the Southern Ocean, in particular the bottom water and the WDW. It makes the topographic effect on the hydrographic structure more evident, especially for salinity. Second, the simulated MLDs, especially on the deep convection sites, are much closer to the observations with the additional AISM. Third, the simulated Antarctic sea ice extent with AISM is in better agreement with observations. These commonly better agreements suggest that the applied AISM parameterization is a promising way to include the ice shelf-ocean interaction in a global sea ice-ocean model.

- There are still obvious deficiencies in the simulation with AISM. First, the simulated occurrence time of the maximum sea ice extent is still one month earlier than observed. Second, the simulated minimum sea ice extent is still underestimated, although better than without AISM. Finally, sea ice in East Antarctica is still lacked. The first shortcoming is due to faster thermodynamic growth than reality in the simulation (Timmermann et al., 2005). The second shortcoming is attributed to the missed sea ice in the East Antarctica and the ignoring the superimposed ice formation in the model. The third shortcoming is owing to the coarse resolution of the model and the too warm surface water in summer.

- Representation of the bathymetry is important for the simulation of water masses in the Southern Ocean. The highly smoothed topography in coarse climate models may cause more NADW intruding the area (Fig. 7.13), resulting in overestimate of temperature and salinity. This is particularly evident in the Weddell Sea region, where warm bias is found in both simulations with and without AISM, accompanied by underestimated ISM freshwater flux from Fimbul Ice Shelf. Such warm bias is commonly found in ocean models of this resolution and time scales, implying that bathymetry at some regions needs special treatment if one wants a better representation of the hydrography in the Southern Ocean.

- The simulated circumpolar ISM flux in this study was about 106 mSv. This amount is much larger than in previous studies (e.g., Jacobs et al., 1996; Hellmer, 2004), since we included more ice shelves and some of the ISM was highly overestimated primarily due to the over-smoothed bathymetry. Through comparing with the observations and simulations, we estimated the circumpolar flux to be about 60-70 mSv. This amount could be further verified with more observations and finer resolution of measurements available.

- The mass balance of the Antarctic ice sheet and the freshwater budget of the Southern Ocean were analyzed. We showed that the Antarctic ice sheet is a source of the ocean water mass as in some other studies (e.g., Rignot and Thomas, 2002), discharging 384 - 790 Gt year\(^{-1}\) ice into the ocean. Freshwater from AISM is a major contributor to the freshwater budget in the Southern Ocean. The Southern Ocean is freshening, as observed in the intermediate water by Wong et al. (1997) and in the bottom water by Rintoul (2007), Johnson et al. (2008) and Ozaki et al. (2009).

- ORCA2-LIM-ISP basically can reproduce the main features of temperature, salinity, MLDs and sea ice in the North Atlantic Ocean and the Arctic Ocean. However, there are notable shortcomings. First, the surface and subsurface waters are too cold in the Hudson Bay, the Baffin Bay and the Davis Strait. Second, the deep convection in the Labrador
Sea is missing. Finally, the sea ice cover is overestimated in the Labrador Sea. All these deficiencies are attributed to the artificial enlargement of the Nares Strait in the model, which allows more cold water and sea ice flow out of the Arctic Ocean.

- The addition of AISM reduces the AABW production and export, enhances the northward heat transport, and causes global changes in the sea ice and ocean system. The changes are most pronounced in the Southern Ocean. In the Southern Ocean:
  1. With the addition of AISM, the continental shelf waters become colder, fresher and lighter.
  2. The added freshwater below the surface tends to rise to the surface due to its small density, and appears at the surface not far from the ice shelf. It is then further transported with the surface current and finally remains at the surface in most regions of the Southern Ocean. Therefore, the surface waters become colder and fresher in most of the Southern Ocean.
  3. The cooling of surface water leads to thicker and more compact Antarctic sea ice cover, thickened ice mainly found near the coast, and more compact ice visible at the ice edge.
  4. The freshened surface water is less dense and cannot sink deep. Therefore, the AABW production and export are reduced. As a consequence, deep and bottom waters become warmer, and deep water becomes saltier.
  5. Addition of AISM deepens the mixed layer on the Antarctic continental shelf, and shallows it in the ACC region.
  6. The production of AISM needs to absorb heat from the adjacent ocean. The locally lost heat needs more warm and saline NADW for compensation (Fig. 7.13). This also increases the temperature and salinity of the deep water in the Southern Ocean.
  7. The density decrease is larger close to the coast and becomes smaller away from the coast. The northward decrease of density produces a westward flow along the coast according to the thermal wind relation. Thus, the Antarctic Coastal Current (ACoC) is enhanced after inclusion of AISM.
  8. The strong freshening in the south due to AISM (Fig. 5.3) weakens the meridional density gradient and thereby slows down the ACC. The ACC transport through DP is correspondingly reduced, becoming more agreement with the observation (Cunningham et al., 2003) during the short integration period.
  9. On the whole, in the Southern Ocean, deep water becomes warmer and more saline, bottom water becomes warmer and fresher, and intermediate water becomes fresher and warmer. Correspondingly, the water density becomes smaller, since density changes are mainly determined by salinity changes in a cold environment. These features are in good agreement with the observed trends in the Southern Ocean, strongly suggesting that AISM might be a major contributor to the recent changes in the Southern Ocean.
  10. ISM and Southern Ocean affect each other as shown by our time lag correlation analysis. The feedback is negative between ISM and the temperature of surface and intermediate waters, while positive between the ISM and the temperature of deep and bottom waters, with a time lag of approximately 5-10 years. However, the time lag should be taken qualitatively. This is because, first of all, the model used in this study is an ocean-only model. The time lag could be smaller with the atmosphere model included, since the atmosphere has a much faster response than the ocean. Second, the sub-ice shelf cavity was only implicitly included. If it were explicitly included, the time lag could be longer.
  11. The impact of AISM is not only found in the Southern Ocean but also reaches the northern hemisphere. With the reduction of AABW and the enhancement of the northward surface spreading of AISM, more cold and fresh water is first transported
into the tropical regions, resulting cooling and freshening over there, and then more warm and saline water is transported from the tropical regions to the North Atlantic Ocean.

(12) Due to the strong interaction with the overlying atmosphere, strengthening of the northward heat transport is large south of 60°N, but quite small northward. This results in large warming in the North Atlantic Ocean, less warming in the Arctic Ocean. In contrast, salinity remains relatively stable during the northward transport, and therefore, salt enrichment is found in both the North Atlantic Ocean and Arctic Ocean.

(13) In the northern hemisphere, changes in temperature, salinity and density are mainly found in the upper layers. Significant changes in MLDs are found in the northern North Atlantic Ocean, where the MLDs are deepened more than several tens meters, even more than hundreds of meters in winter, after inclusion of AISM.

(14) The minor warming in the Arctic Ocean results in a small decrease of the Arctic sea ice. A bipolar seesaw forms for changes in the polar sea ice: large increase in Antarctic sea ice while quite small decrease in the Arctic sea ice. Sea ice extent and volume increase by 11% and 18% in the Antarctic but only by 0.21% and 0.37% in the Arctic.

- The freshwater signal can reach the northern North Atlantic within several years but it would take 35-40 years to build up noticeable temperature and salinity anomalies.
- Finally, we can use the schematic map below to end our findings in this study.

![Schematic map for the impact of AISM on a global ice-ocean system.](image)

**Fig. 8.2.** A schematic map for the impact of AISM on a global ice-ocean system. AIS: Antarctic ice shelf; SO: Southern Ocean; EQ: Equator; NH: Northern hemisphere; ISM: ice shelf melting; AABW: Antarctic bottom water; NADW: North Atlantic deep water; S: salinity; T: temperature; θ: decrease; ⊕: increase.

### 8.3. Perspectives for future work

- As a first step to include AISM in a global climate ocean model, the implementation of the parameterization of BG03 needs further improvement. For example, instead of a uniform $L_{eff}$, different effective length scale shall be used for different ice shelves when more information about the ice shelves has been collected.
- The overlying atmosphere could be influenced and influence the underlying ocean in the freshwater anomaly experiment. AISM induces local cooling (Fig. 5.2) in this
study. This cooling can result in an intensification of the westerly wind between 40-60° S (Ma and Wu, 2010), which may accelerate ACC and then enhance AISM (Hattermann and Levermann, 2009). In addition, local wind forcing may affect the distribution of HSSW (Assmann et al., 2003) and further ISM (Timmermann et al., 2002). These issues only can be clarified with coupled atmosphere-ocean models.

- Many changes in ice shelves have occurred in recent years, especially after 2000. For example, the collapse of Larsen B Ice Shelf in 2002 along the eastern coast of the Antarctic Peninsula, rapid bottom melting widespread near Antarctic ice sheet grounding lines (Rignot and Jacobs, 2002), and high melt rates of ice shelves in response to climate change, especially those in the Amundsen Sea and Bellingshausen Sea (Rignot et al., 2008). These will influence the present-day real ocean, as already seen in this study during the period 1958-2000 (ISM has already increased during this period (Fig. 5.10b)). Therefore, more numerical experiments need to be performed with more recent atmospheric forcing data, such as, 2000-2010.
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