

Interannual variability and future changes of the Southern Ocean sea ice cover

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Word of Thanks

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List of Abbreviations

The abbreviations explaining the names of the different model runs with idealised forcings of Chapter 5 can be found in Table 5.1 on page 131. The explanation of the abbreviations denoting the different models used in Chapter 7 can be found in Table 7.1 on page 197.

20C3M	Climate of the 20th century experiment
4AR	4th Assessment report
A1B	A1B-scenario as defined in Nakicenovic and Swart (2000)
AABW	Antarctic bottom water
AAO	Antarctic oscillation
ACC	Antarctic circumpolar current
ACW	Antarctic circumpolar wave
ADP	Antarctic dipole
AOGCM	Atmosphere-ocean general circulation model
ASPeCt	Antarctic sea ice processes and climate
CGCM	Coupled general circulation model
CIRES	Cooperative institute for research in environmental sciences
CLIO	Coupled large-scale ice ocean
CLIVAR	Climate variability and predictability
CMIP	Coupled model intercomparison project
ECMWF	European centre for medium-range weather forecasts
EMIC	Earth model of intermediate complexity
ENSEMBLE	Ensemble-based predictions of climate changes and their impacts
ENSO	El niño southern oscillation
EOF	Empirical orthogonal function
ETOPO	Earth topographic project

GCM	General circulation model
GRDC	Global runoff data centre
HadISST	Met office Hadley centre's sea ice and sea surface temperature data set
HLM	High latitude mode
IHO	International hydrographic organization
IPCC	Intergovernmental panel on climate change
JJAS	June-July-August-September
JSC	Johnson space centre
LG05	Lefebvre and Goosse (2005)
LG07a	Lefebvre and Goosse (2007a)
LG07b	Lefebvre and Goosse (2007b)
LGTF04	Lefebvre <i>et al.</i> (2004)
LIM	Louvain-la-Neuve ice model
MEI	Multivariate ENSO index
NAM	Northern annular mode
NEMO	Nucleus for European modelling of the ocean
NCAR	National center for atmospheric research
NCEP	National centers for environmental prediction
NOAA	National oceanic and atmospheric administration
OPA	Océan parallélisé
ORCALIM	Coupled model that consists out of OPA and LIM
PC	Principal component
PCA	Principal component analysis
PCM	Parallel climate model
PgC	Petagramme de carbon ($= 10^{12}kg$ de carbon)
PHC	Polar science center hydrographic climatology
PCMDI	Program for climate model diagnosis and intercomparison
psu	practical salinity unit ($\approx g/kg$)
SAM	Southern annular mode
SAO	Semiannual oscillation
SHAM	Southern hemisphere annular mode
SLIM	Second generation Louvain-la-Neuve ice-ocean model
SLP	Sea level pressure
SO	Southern oscillation
SOI	Southern oscillation index

SRES	Special report on emission scenarios (Nakicenovic and Swart, 2000)
SSMI	Special sensor microwave imager
SSMR	Scanning multichannel microwave radiometer
SSS	Sea surface salinity
SST	Sea surface temperature
SWAMCO	Sea water microbial community model
TSU	Technical support unit
ULS	Upward looking sonar
WGCM	Working group on coupled modelling
WGI	Working group 1
ZW1	Zonal wave 1
ZW3	Zonal wave 3

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*Mijn hemel blauw met gouden harp,
mijn wolkentorens, ijskristallen,
kometen, manen en planeten,
aah, alles draait om mij.
En door de witte wolkenpoort
tot diep onder de golven boort
mijn vuur, mijn liefde, zich in de aarde.
En bij het water speelt een kind
en alle schelpen die het vindt
gaan blinken als ik lach.*

**'k Hou van je warmte op mijn gezicht.
Ik hou van de koperen kleur van je licht.
Ik geef je water in mijn hand
en schelpen uit het zoute zand.
Ik heb je lief, zo lief.**

*Ik scheur de rotsen met mijn stralen
verhoog de meren in de dalen en
onweersluchten doe ik vluchten,
aah, als de regen valt.
Verberg je ogen in een hand,
voordat m'n glimlach ze verbrandt,
m'n vuur, m'n liefde, mijn gouden ogen.
't Is beter als je nog wat wacht,
want even later komt de nacht
en schijnt de koele maan.*

**De nacht is te koud, de maan te grijs.
Toe neem me toch mee naar je hemelpaleis.
Daar wil ik zijn alleen met jou
en stralen in het hemelblauw.
Ik heb je lief, zo lief.**

*Als ik de aarde ga verwarmen,
laat ik haar leven in m'n armen.
Van sterren weefde ik het verre,*

aah, het noorderlicht.
Maar soms ben ik als kolkend lood.
Ik ben het leven en de dood,
in vuur, in liefde, in alle tijden.
M'n kind ik troost je, kijk omhoog,
vandaag span ik mijn regenboog.
Die is alleen voor jou.

Nee nooit sta ik een seconde stil.
'k Wil liever branden neem me mee.
Geen mens kan mij dwingen wanneer ik niet wil.
Wanneer je vanavond gaat slapen in zee.
Geen leven dat ik niet begon.
En vliegen langs jouw hemelbaan.
Je kunt niet houden van de zon.
Ik wil niet meer bij jou vandaan.
Ik heb je lief, zo lief.

Ik heb je lief, zo lief.
Ik heb je lief, zo lief.
Ik heb je lief, zo lief.
Ik heb je lief, zo lief.

Liesbeth List & Ramses Shaffy - Pastorale

Chapter 1

Introduction

1.1 The Southern Ocean and Climate Change

The Southern Ocean covers all longitudes. It is connected to three other large oceanic basins, the Atlantic, the Indian and the Pacific Oceans. As a result, the Southern Ocean is very important for the global ocean circulation and it influences the water mass characteristics over a large portion of the world. The Southern Ocean is also an area in which there are important exchanges between the atmosphere, the ocean, the sea ice and the ice-covered land mass of the Antarctic continent, at its southern border. Thanks to all these phenomena, the Southern Ocean can be seen as an enormous laboratory in which transfers of heat and mass between these different components can be studied. As the Southern Ocean and adjacent oceanic basins cover most of the Southern Hemisphere, their combined mass of water tends to damp changes in the climate due to its enormous heat capacity. Furthermore, numerous animal species live in or around the Southern Ocean and need it as an important part of their habitat. In addition, the ocean south of 30°S is one of the primary CO_2 sinks in the world (Denman *et al.*, 2007). However, due to the relative absence of land areas, the Southern Hemisphere has a considerable smaller human population than the Northern Hemisphere leading to less observations and less interest in the study of this hemisphere than its northern counterpart. **As a result, the Southern Hemisphere climate, and in particular the climate of the Southern Ocean, is**

much less known than the climate of the Northern Hemisphere, despite its importance for the global climate.

Sea ice affects, for instance in the Southern Ocean, different transfers with other components of the climate system, specifically the atmosphere and the ocean. As sea ice is white, its albedo¹ is high. So it reflects most of the shortwave solar radiation. As a result, it has an important cooling effect on the climate. Furthermore, by providing a cover between the atmosphere and the ocean, sea ice inhibits important heat exchanges between the two, thereby disabling a heat flux between the ocean and the atmosphere that would otherwise be very large, especially in winter. When seawater freezes, it drops an important part of its saline content into the ocean, increasing the salinity of the upper layer and thus decreasing the stability. As a result, sea ice formation in the Southern Hemisphere is important in the process of forming high saline bottom water that afterwards is exported to the north as Antarctic Bottom Water (AABW). Finally, gas exchanges between the atmosphere and the ocean, especially exchange of CO_2 (a major greenhouse gas) and O_2 (important for animal life) are strongly reduced in the presence of sea ice. **So, it is important to know the effect of the sea ice on the climate.**

At the end of the sixties, there may have been an increase in the sea ice extent² in the Southern Ocean as evidenced by early visible and infrared satellite images (Cavalieri *et al.*, 2003). Thereafter, a large decrease of almost two million squared kilometers took place in a period of about 5 years. During the last two decades of the 20th century and the beginning of the 21st century, the annual mean sea ice extent stayed mostly stable, although a small positive trend can be seen in some seasons. However, these changes show a large interdecadal and intradecadal variability. As shown by Zwally *et al.* (2002); Cavalieri *et al.* (2003), the interannual sea ice variability in the Antarctic is fairly large, certainly on a sectoral scale. However, some anomalies persist for periods of 3-5 years, like the Weddell Polynya (Carsey, 1980). Sea ice extent is not only determined by temperature but also by snowfall, salinity of the ocean, the stability of the ocean (due to its link to the ocean temperature), horizontal and vertical ocean currents, etc. So, interannual variability of the

¹Albedo is the ratio between the reflected and the incoming solar power.

²The sea ice extent is the sum of the surface of all the grid cells with a mean ice concentration of more than 15%, see later.

sea ice will be influenced by all these factors. Due to complex feedback mechanisms, some of these factors are also determined by the sea ice. As a result, it is sometimes difficult to find out what is the cause and what is the effect. **In any case, it is important to understand the interannual, intradecadal and interdecadal variability of the sea ice in the Southern Ocean.**

There is a large scientific consensus that the Earth is warming due to the emission of greenhouse gases and that this will continue in the near future. Some important effects include changes in the atmospheric circulation, oceanic circulation, sea level rise... Due to its vast body of water, the Southern Ocean has a dampening effect on the climate change. Finally, as said above, the Southern Ocean is a major sink for the CO_2 . Any changes influencing the uptake capacity for CO_2 of the Southern Ocean will induce changes in the global climate. **It is thus important to investigate the impacts of the climate change on the Southern Ocean and vice versa.**

Sea ice is generally seen as one of the harbingers of climate change. Next to that, it is also responsible for one of the most important feedbacks in climate. As the sea ice disappears, the albedo of the world decreases and the temperature continues to increase, which accelerates the melt of the sea ice. Despite global warming, the trend in the sea ice extent in the Southern Ocean is slightly upward in the last years. **It is important to know why the Southern Ocean sea ice does not seem to be directly affected by the climate change.**

The projections for the future show clearly a strong decline in the sea ice extent, in all seasons, on both hemispheres (Houghton *et al.*, 2001; Flato and Boer, 2001; Goosse and Renssen, 2001; Flato, 2004; Vinnikov *et al.*, 2006; Parkinson *et al.*, 2006b; Arzel *et al.*, 2006; Meehl *et al.*, 2007). There is a large variability between the projected decreases in the different scenarios and in the different models (see later). Furthermore, most models project atmospheric circulation changes due to the increase in greenhouse gas concentrations. Some of these changes are already being seen (Hegerl *et al.*, 2007). Further anthropogenic³ induced changes can have complex interactions with the existing climate. Furthermore, the natural variability can regionally mask or amplify during a certain

³Anthropogenic effects, processes, objects, or materials are those that are derived from human activities, as opposed to those occurring in natural environments without human influences. Source of the definition: Wikipedia.

period the effects of the climate change. **It is therefore important to assess the current climate variability in order to project the future climate.**

It is only since the end of the seventies that a large amount of interesting and reliable data is available, thanks to the satellites. So, finally, we have enough information to try to address some of the topics outlined above. Furthermore, during the last decades, the results from computer models have improved a lot due to three reasons. First of all, models developed to use actual observations as their boundary conditions can be more accurately forced using the increased amount of observations and the improved quality of them. Secondly, the calculation power of the computers has rocketed. As a result, more runs with a less coarse grid over longer periods can be done. Finally, thanks to advances in the modelling science, the models have become better. **So, some necessary tools are available to investigate in detail the interannual variability of the sea ice, using both reliable observational data and model outputs.**

1.2 Structure of this study

The goal of this thesis is to analyse the major processes responsible for the variability of the Southern Ocean regional and total sea ice extent variability. Furthermore, the current and future response of the Southern Ocean and the surrounding ocean basins to climate change, which is currently not well understood, is investigated. In order to better understand the mechanisms that govern this variability, model results have been used. The ORCALIM-model was developed by several people and was ready to be used at the moment I arrived at the institute. In order to investigate the sea ice variability, we have used reanalysis for the near-surface atmospheric temperature and winds, together with existing climatologies for precipitation, cloud fraction and surface humidity to force the model and we have then analysed its results. If necessary, an idealised forcing of the model has been constructed and applied. In order to project the future, runs that have been made in the framework of the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC-AR4) have been analysed. The results have been published in different peer-reviewed journals, which are reprinted

in Chapters 4 to 7. In addition, the same model (ORCALIM) was, for instance, also used in a detailed case-study of the Western Antarctic Peninsula region over the spring and summer 2005-2006 which has been submitted in Massom *et al.* (2007).

In the following chapter, some aspects of climate modelling are discussed. First of all, an overview of the structure of the ORCALIM-model is given. This is done, by first discussing the ocean model, then the ice model, the coupling followed by the configuration and forcing and finally the validation. The last section of Chapter 2 gives a short discussion on the models that have been used for the study detailed in subsection 3.2.6 and in Chapter 7.

Chapter 3 discusses the different physical factors that influence the interannual climate variability in the Southern Ocean, especially the sea ice. After having discussed some mathematical methods that will be used further on, the different modes of variability are discussed together with their influence on sea ice. Each time, we try to give a review of the current knowledge. Furthermore, we show how our work is linked to the other investigations about the same subject.

Chapters 4 to 7 are reprints of the papers that have been published during these last three years, for which I was first author. In order to prevent unnecessary repetition, the references are stripped off these papers and have been put together with the references of the other chapters at the end of this thesis (Page 229). The structure, as well as the figures and tables have been copied from the original, except for some small layout changes. In some cases, small typographic errors in the papers have been corrected. In addition, some sentences have been slightly adapted to take into account some new points of view acquired since the publication of the papers. At these locations, footnotes have been inserted to discuss this. Chapter 4 reprints a paper, published in the *Journal of Geophysical Research* in 2004, discussing the main response of the ocean and the sea ice to the interannual variability of the Southern Annular Mode (Lefebvre *et al.*, 2004). In order to better understand in detail the response of the sea ice to the Southern Annular Mode, a following paper investigating this response has been published in *Ocean Science* in 2005. This paper can be found in Chapter 5 (Lefebvre and Goosse, 2005). These two papers show that the influence of the SAM on the sea ice extent integrated over the Southern Ocean is small. Therefore, we wanted to find if there exists a mode of variability which has an important and robust impact

on the sea ice extent, and if not, which atmospheric configuration is the most favourable for a large September Southern Hemisphere sea ice extent. This paper was accepted by the *Journal of Geophysical Research* in 2007 (Lefebvre and Goosse, 2007a) and is reprinted here in Chapter 6. Finally, we have conducted an investigation to assess if the current variability of the sea ice will remain similar in the future. This has been discussed in a paper that was published in 2007 in *Climate Dynamics*. It can be found in Chapter 7 (Lefebvre and Goosse, 2007b). This paper also discusses the regional characteristics of the projected future response of the sea ice in a well defined scenario (SRESA1B).

Finally, conclusions and perspectives can be found in Chapter 8.

Chapter 2

Modelling of the Southern Ocean climate

2.1 Different classes of models

It is possible to classify the existing climate models in categories. We will discuss here some of the classifications that can be found in the literature.

2.1.1 Complexity

Claussen *et al.* (2002) have introduced a classification of the models depending on their complexity. The same classification has been used in the latest IPCC-report (Randall *et al.*, 2007). They distinguished three main categories, although the bounds between these categories are not well defined, which makes the classification of certain models unclear.

- Conceptual models.
These models do not attempt to resolve Earth's climate system in full detail. The equations resolved in the model are very strongly reduced in terms of complexity, the grid is very crude, no detailed bathymetry is available, and so on. Instead, the processes in these models are easily understood, so they can be used to better understand the basic physics behind some phenomena.
- General circulation models.
These models try to resolve the most processes and details possible

in order to get the best possible simulation of reality. In general, they are rather CPU-time-consuming and thus cannot be used for simulations over a very long period. In this work, only models of this type will be used. Their time-consuming behaviour increases strongly when working on a finer grid.

- Earth-system Models of Intermediate Complexity (EMIC's).
These models try to make a bridge between both previous categories. They have, as said by their name, an intermediate complexity, which enables them to simulate long periods. However, it is not possible to simulate all the details of all the processes by using this kind of models.

2.1.2 Regional versus Global Models

The regions for which the models simulate the climate can differ from one model to another. For instance, global models will mimic the climate on the complete Earth. As a result, due to the limited calculation capacity, the resolution of these models is in general quite coarse. However, unlike regional models, they do not need a forcing at their boundaries. All the models used in this work are global models. The advantage of the use of global models is the possibility to take into account influences over long distances.

2.1.3 The simulated aspects

Some models only simulate some aspects of the climate, such as the oceans, or the sea ice, or the atmosphere, but not all of them. If a model does not simulate one of the important components of the climate, a forcing has to be applied instead, in order to simulate the effect of this component on the simulated ones. For instance, an ocean-sea ice model has to be forced by an atmosphere, in order to have the effect of the atmosphere on the ocean and the sea ice. In general, a dataset based on observations is used to force the model, although the forcing can also be derived from another model. In this work, we will use two types of models:

- Ocean-sea ice models.

The model consists of two parts, an ocean model and a sea ice model which are coupled. In addition, the model uses forcing from actual atmospheric observations in order to mimic the interaction between the ocean/sea ice system and the changes in the atmosphere. The advantage of this kind of models is the greater accuracy of the modelled ocean and sea ice (Gerdes and Köberle, 2007), because the atmosphere is as realistic as it can be, i.e. directly derived from observations. However, this kind of models can only be used when enough and good quality information about the atmosphere is available, thus only for hindcast of the not so distant past simulations. Other simulations can be forced with the output of atmospheric models, but the problems in the simulations of the coupled models will automatically be incorporated in the results of the forced model. The ORCALIM-model, which will be discussed in section 2.2, is an example of a ocean-sea ice model used in hindcast experiments.

- Ocean-atmosphere-sea ice-land models.

As these models have a model for each part of the climate system which is important for the near future, they can be used to project the future, although they need information about the projected solar forcing, concentration of CO_2 , ... Further forcing data are needed for the non-simulated aspects as the ice sheets, the carbon cycle, ... If the interaction between these systems and the modelled systems is supposed to be negligible, no forcing is needed. However, if used to simulate the present state of the ocean and the sea ice, it becomes clear that the results, compared to the ones obtained with an ocean-sea ice model with a forced atmosphere, are less good. First of all, it is impossible to compare model results and observations year by year as the model will not reproduce what happened during a certain year, but only one of the possibilities that could have happened. Furthermore, the mean state of the climate and the variability around it is better reproduced in models with an observation-derived atmospheric forcing than in models with a simulated atmosphere (Gerdes and Köberle, 2007).

2.2 ORCALIM

ORCA2-LIM (or ORCALIM) results from the coupling of the Louvain-la-Neuve sea ice model (LIM) (Fichefet and Morales Maqueda, 1997) with the hydrostatic, primitive equation ocean model OPA (Océan Parallélisé) (Madec *et al.*, 1999). OPA is a finite difference ocean general circulation model with a free surface and a nonlinear equation of state (see below for more details). A structure scheme of the ORCALIM model can be found in Figure 2.1. ORCALIM is the model that will mostly be used in this PhD-thesis.

2.2.1 Ocean Model: OPA

A longer description of the model and its physics can be found in Madec *et al.* (1999); Timmermann *et al.* (2005). Here, we give a brief overview.

Primitive equations and approximations

First of all, let us take a look at the equations that have been used in the model. These are simplifications of the laws of nature. In order to solve them more easily, some approximations have been applied:

- Spherical Earth approximation: gravity is parallel to the earth's radius (and thus parallel to the $\vec{k}^1$).
- Thin-shell approximation: the ocean depth is neglected compared to the earth's radius. As a result, g is equal at the top and at the bottom of the ocean².
- Boussinesq approximation and the incompressibility hypothesis: density variations are neglected except for the gravity term (right hand side of equation 2.3). As a result, the continuity equation, taking into account the density variations,

$$\frac{1}{\rho} \frac{D\rho}{Dt} + \vec{\nabla} \cdot \vec{U} = 0 \quad (2.1)$$

simplifies to equation (2.4). It is shown (Kundu and Cohen, 2002) that this approximation can only be used for water speeds that

¹ $\vec{k}$ is the local upward unit vector.

²The values and definitions of the constants like g can be found in Table 2.1.

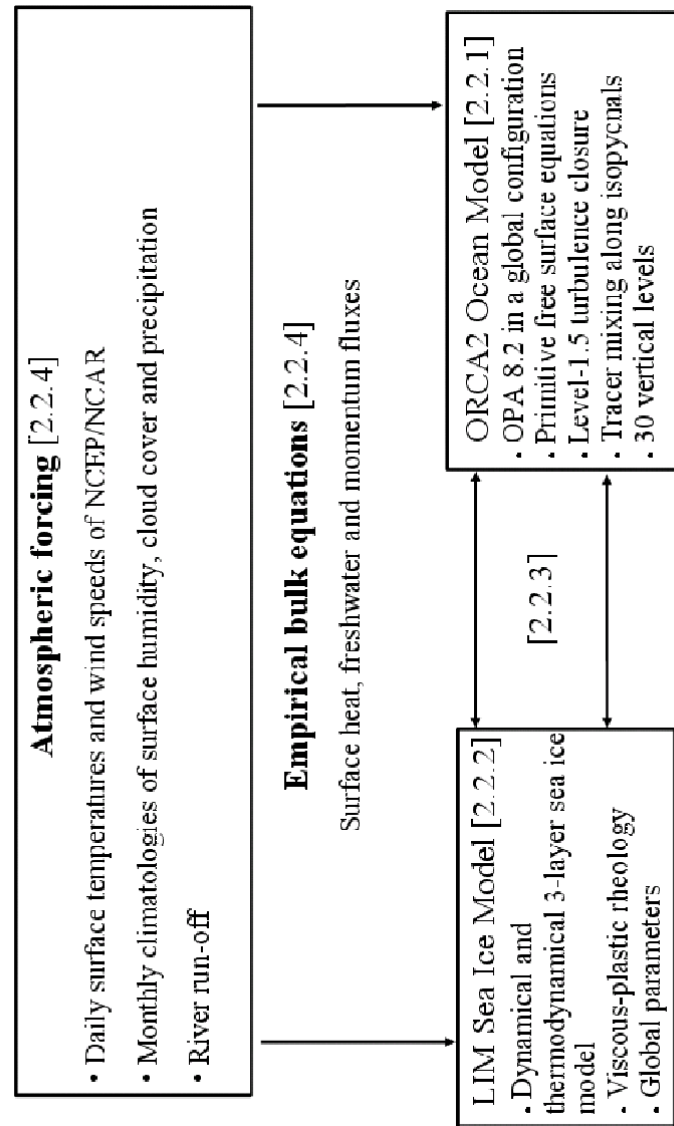


Figure 2.1: Structure of the ORCALIM model. The numbers in brackets denote the sections in which this specific part is explained.

are not too large (much lower than $1500 \frac{m}{s}$.) and in water volumes that are not too deep (under $200km$). The Boussinesq approximation changes also the Navier-Stokes equations (2.2) and the heat diffusion equation (2.5). The interested reader can find more information about this approximation in Kundu and Cohen (2002, section 4.18).

- The hydrostatic hypothesis: the vertical momentum equation is reduced to a balance between the vertical pressure gradient and the buoyancy force. This results in the need to parametrise the convective processes in the Navier-Stokes equations (2.2). An extended description of the effects of the hydrostatic hypothesis and the simplifications entrained by it can be found in Marshall *et al.* (1997). In short, there is no effect from the hydrostatic hypothesis on the resolution of processes with a horizontal length scale much larger than the vertical scale, like the processes we will discuss here (for instance, the global ocean circulation).

Applying these approximations leads to the following equations:

$$\frac{\partial \vec{U}_h}{\partial t} = -(\vec{U}_h \cdot \vec{\nabla}_h) \vec{U}_h - w \frac{\partial \vec{U}_h}{\partial z} - f \vec{k} \times \vec{U}_h - \frac{1}{\rho_0} \vec{\nabla}_h p + \vec{D}^U \quad (2.2)$$

$$\frac{\partial p}{\partial z} = -\rho g \quad (2.3)$$

$$\vec{\nabla} \cdot \vec{U} = 0 \quad (2.4)$$

$$\frac{\partial T}{\partial t} = -\vec{\nabla} \cdot (T \vec{U}) + D^T \quad (2.5)$$

$$\frac{\partial S}{\partial t} = -\vec{\nabla} \cdot (S \vec{U}) + D^S \quad (2.6)$$

$$\rho = \rho(T, S, p) \quad (2.7)$$

$(\vec{i}, \vec{j}, \vec{k})$ is an orthogonal set of unit vectors so that $\vec{k}$ is the local upward vector, $\vec{i}$ is horizontal and directed to the north and $\vec{j}$ is horizontal and directed to the east. $\vec{U}$ is the velocity which is decomposed in a horizontal vector ($\vec{U}_h$) and a vertical component (w) as follows: $\vec{U} = \vec{U}_h + w \vec{k}$. T is the potential temperature, S the salinity and ρ the local density of the ocean water. The coordinates following respectively $\vec{i}$, $\vec{j}$ and $\vec{k}$ are denoted as x , y and z , while t is time. ρ_0 is a reference density, p is the

pressure and f is the Coriolis acceleration (with $f = 2\vec{\Omega} \cdot \vec{k}$ in which $\vec{\Omega}$ is the Earth angular velocity vector). g is the gravitational acceleration. Last but not least, $\vec{D}^U$, D^T and D^S are the parametrisations of small-scale physics and the source and sink terms at the boundaries (surface fluxes). The equation of state (2.7) consists out of an equation with 41 terms. It can be found in Jackett and McDougall (1995).

The first equation (2.2) assumes that the water moves following the second law of Newton. The left hand side of this equation is the acceleration, while the first two terms of the right-hand side describe the advection. Thereafter, respectively, the Coriolis force, the force due to the horizontal pressure difference and the parametrisation term can be found. The second equation (2.3) says that the vertical pressure gradient is balanced by the buoyancy force. The following equation (2.4) is the continuity equation which imposes that no mass can be lost or created. Equations (2.5) and (2.6) describe the temporal evolution of the temperature and the salinity, respectively. Finally, the equation of state (2.7) discusses how the density changes in function of the temperature, the salinity and the pressure.

Boundary Conditions

As the ocean is bounded by coastlines, bottom topography, atmosphere and sea ice, the model has to take into account several fluxes from and to the ocean. Let us take a more precise look at these fluxes:

- Land-ocean fluxes.

The only flux taken into account is the run-off flux (from land to ocean) due to the discharge of fresh water into the oceans by rivers and the ice sheets.

- Bottom-ocean fluxes.

No heat³ and salt fluxes between the bottom of the ocean and the ocean itself are taken into account. For momentum, there is no flow across solid boundaries. As a result, the bottom velocity is parallel to the solid boundaries. This can be noted as

$$w = -\vec{U}_h \cdot \vec{\nabla}_h(H) \quad (2.8)$$

³As a result, geothermal factors are not taken into account.

with H the depth of the ocean. In addition, there is an exchange of momentum between the ocean and the bottom topography (see also Figure 2.2 for the Southern Hemisphere). This factor needs to be parameterised. The parametrisation is done as follows:

$$\tau_{bot} = \rho_0 C_{db} |\vec{u}_{bot}| \vec{u}_{bot} \quad (2.9)$$

with $\vec{u}_{bot}$ the ocean velocity at the bottom of the ocean and C_{db} a drag coefficient.

- Atmosphere-ocean fluxes.
The atmosphere and the ocean exchange heat, freshwater and momentum, thus affecting the salinity (via the precipitation-evaporation term (P-E)), temperature and momentum.
- Sea ice-ocean fluxes.
There is an exchange of salt, freshwater, heat and momentum. This will be explained further later on.

Grid and Numerical Methods

There are several reasons to make use of a so-called curvilinear geopotential coordinate system as a grid.

- In some cases, we want a more important resolution in some regions (here, the poles). Indeed, an improved resolution is needed to simulate some phenomena. Some examples are surface layers, western boundary currents, equatorial currents and ocean fronts. In this model, a finer grid has been chosen around the equator, close to the poles and in some seas (see later for more details).
- The common geographical coordinate system (latitude, longitude, height) has a singular point at the North Pole (in the ocean). Indeed, in a latitude-longitude grid, a model point has normally 4 neighbours, whereas the North Pole has as many neighbours as there are divisions in longitude (in this case 180). The South Pole has the same problem. However, the South Pole is a land point and, therefore, there is no problem for an ocean model to have a singular point there. For the North Pole, the model has changed from one singular ocean point towards a new grid with two singular

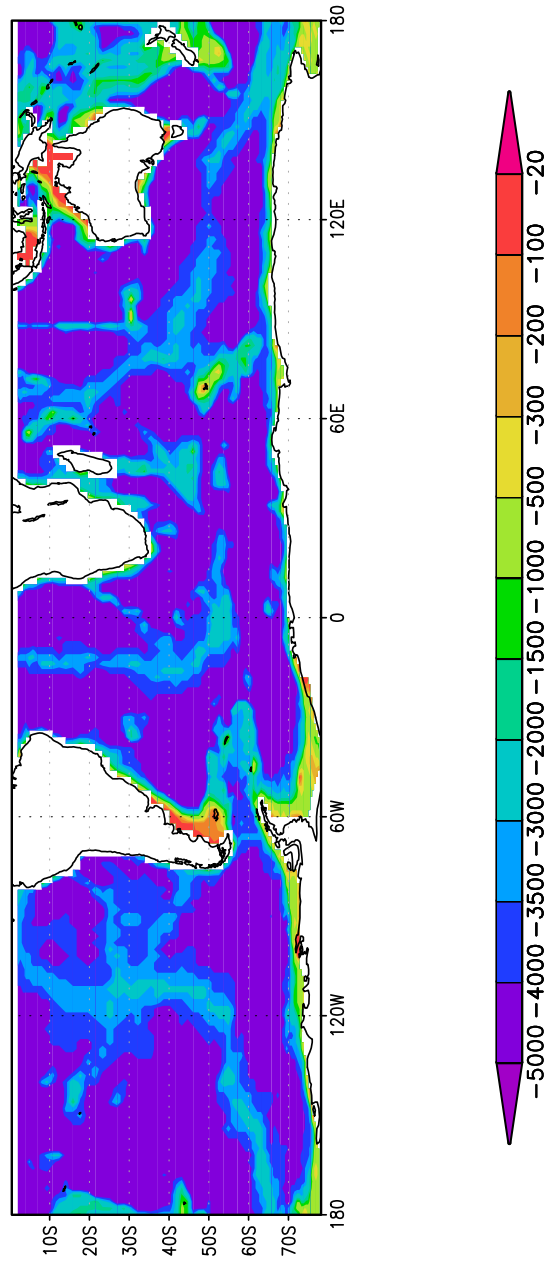


Figure 2.2: The bathymetry of ORCALIM in the Southern Hemisphere (in m).

points, which are both on land, one in Canada and the other one in Siberia (see Figure 2.3). This eliminates the problems that would occur in the common geographical coordinate system.

As a result, a three dimensional orthogonal grid on the sphere has been chosen on which the calculations have been performed. Let us transform the equations seen above into this new grid system (Figure 2.3).

Let (i, j, k) be a set of orthogonal curvilinear coordinates on the sphere linked to a orthogonal set of vectors $(\vec{i}, \vec{j}, \vec{k})$ defined above. Let (λ, φ, z) be the geographical coordinates system defined by the latitude φ , the longitude λ and the altitude above a reference sea level z . Then, the distance to the centre of the Earth is $a + z \approx a$ (thin-shell approximation, see above). Herein, a is the Earth's radius. Due to the change in coordinates, scale factors have to be introduced into the equations.

$$e_1 = a \sqrt{\left(\frac{\partial \lambda}{\partial i} \cos \varphi\right)^2 + \left(\frac{\partial \varphi}{\partial i}\right)^2} \quad (2.10)$$

$$e_2 = a \sqrt{\left(\frac{\partial \lambda}{\partial j} \cos \varphi\right)^2 + \left(\frac{\partial \varphi}{\partial j}\right)^2} \quad (2.11)$$

$$e_3 = \frac{\partial z}{\partial k} \quad (2.12)$$

Then the equations that have been solved on the grid are (details can be found in Madec et al. (1999)):

$$\begin{aligned} \frac{\partial u}{\partial t} = & +(\zeta + f)v - \frac{1}{e_3} w \frac{\partial u}{\partial k} - \frac{1}{2e_1} \frac{\partial}{\partial i} (u^2 + v^2) - \frac{1}{\rho_0 e_1} \frac{\partial p_h}{\partial i} \\ & - \frac{1}{\rho_0 e_1} \frac{\partial p_s}{\partial i} + D_u^W + \frac{1}{e_3} \frac{\partial}{\partial k} \left(\frac{A^{vm}}{e_3} \frac{\partial u}{\partial k} \right) \end{aligned} \quad (2.13)$$

$$\begin{aligned} \frac{\partial v}{\partial t} = & -(\zeta + f)u - \frac{1}{e_3} w \frac{\partial v}{\partial k} - \frac{1}{2e_2} \frac{\partial}{\partial j} (u^2 + v^2) - \frac{1}{\rho_0 e_2} \frac{\partial p_h}{\partial j} \\ & - \frac{1}{\rho_0 e_2} \frac{\partial p_s}{\partial j} + D_v^W + \frac{1}{e_3} \frac{\partial}{\partial k} \left(\frac{A^{vm}}{e_3} \frac{\partial v}{\partial k} \right) \end{aligned} \quad (2.14)$$

$$\frac{\partial p_h}{\partial k} = -\rho g e_3 \quad (2.15)$$

$$\frac{\partial w}{\partial k} = -e_3 \chi \quad (2.16)$$

ORCA R2 : mesh indexation

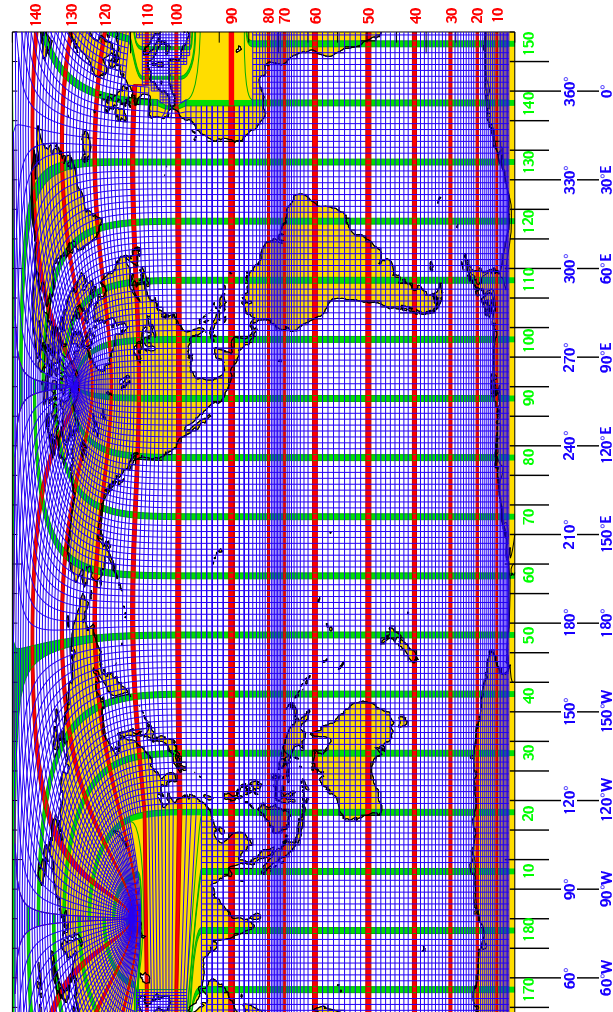


Figure 2.3: The horizontal grid on which ORCALIM works.

$$\frac{1}{\rho_0 e_1} \frac{\partial p_s}{\partial i} = \overline{M_u} + \frac{1}{H e_2} \frac{\partial}{\partial j} \left(\frac{\partial \psi}{\partial t} \right) \quad (2.17)$$

$$\frac{1}{\rho_0 e_2} \frac{\partial p_s}{\partial i} = \overline{M_v} - \frac{1}{H e_1} \frac{\partial}{\partial i} \left(\frac{\partial \psi}{\partial t} \right) \quad (2.18)$$

$$\begin{aligned} \frac{\partial T}{\partial t} = & -\frac{1}{e_1 e_2} \left[\frac{\partial}{\partial i} (e_2 T u) + \frac{\partial}{\partial j} (e_1 T v) \right] \\ & -\frac{1}{e_3} \frac{\partial}{\partial k} (T w) + D^{lT} + \frac{1}{e_3} \frac{\partial}{\partial k} \left(\frac{A^{vT}}{e_3} \frac{\partial T}{\partial k} \right) \end{aligned} \quad (2.19)$$

$$\begin{aligned} \frac{\partial S}{\partial t} = & -\frac{1}{e_1 e_2} \left[\frac{\partial}{\partial i} (e_2 S u) + \frac{\partial}{\partial j} (e_1 S v) \right] \\ & -\frac{1}{e_3} \frac{\partial}{\partial k} (S w) + D^{lS} + \frac{1}{e_3} \frac{\partial}{\partial k} \left(\frac{A^{vS}}{e_3} \frac{\partial S}{\partial k} \right) \end{aligned} \quad (2.20)$$

$$\zeta = \frac{1}{e_1 e_2} \left(\frac{\partial}{\partial i} [e_2 v] - \frac{\partial}{\partial j} [e_1 u] \right) \quad (2.21)$$

$$\chi = \frac{1}{e_1 e_2} \left(\frac{\partial}{\partial i} [e_2 u] + \frac{\partial}{\partial j} [e_1 v] \right) \quad (2.22)$$

$$\overline{M_u} = \frac{1}{H} \int_{-H}^0 \left[\frac{\partial u}{\partial t} + \frac{1}{\rho_0 e_1} \frac{\partial p_s}{\partial i} \right] e_3 dk \quad (2.23)$$

$$\overline{M_v} = \frac{1}{H} \int_{-H}^0 \left[\frac{\partial v}{\partial t} + \frac{1}{\rho_0 e_2} \frac{\partial p_s}{\partial j} \right] e_3 dk \quad (2.24)$$

$$\begin{aligned} \frac{\partial}{\partial i} (e_2 \overline{M_v}) - \frac{\partial}{\partial j} (e_1 \overline{M_u}) \\ = \frac{\partial}{\partial i} \left[\frac{e_2}{H e_1} \frac{\partial}{\partial i} \left(\frac{\partial \psi}{\partial t} \right) \right] + \frac{\partial}{\partial j} \left[\frac{e_1}{H e_2} \frac{\partial}{\partial j} \left(\frac{\partial \psi}{\partial t} \right) \right] \end{aligned} \quad (2.25)$$

$$\rho = \rho(T, S, p) \quad (2.26)$$

with $\vec{M} = (M_u, M_v)$ the collected contributions of non-linear, viscous and hydrostatic pressure gradient terms. The overbar indicates a vertical average over the whole water column. ψ is the volume transport streamfunction, ζ is the relative vorticity and χ is the divergence of the horizontal velocity field. p_h is the hydrostatic pressure and p_s the surface pressure so that $p = p_h + p_s$. Finally, A^{vm} is the vertical eddy viscosity and A^{vT} and A^{vS} are the vertical diffusivity coefficients.

The ocean model uses a C grid arrangement of the variables. In the centre of each cell, the calculations for the scalar variables, such as potential temperature, pressure and salinity are made. The velocities

are calculated in the centre of each face of the grid cell. The equations discussed above are solved using a finite difference scheme. This scheme consists of three parts:

- A leapfrog scheme for the non-diffusive part with an Asselin filter to reduce divergence between the odd and the even time steps.
- A forward time differencing scheme for the horizontal diffusive part.
- A backward time differencing scheme for the vertical diffusive part.

For a discussion on the exact implementation of these schemes, their advantages and disadvantages, we refer the reader to Madec *et al.* (1999).

Subgrid-Scale Physics

The terms describing the subgrid-scale physics in equations (2.2) to (2.6), $\vec{D}^U$, D^T and D^S have been replaced in equations (2.13) to (2.25) by both a lateral part ($\vec{D}^{lU}$, D^{lT} and D^{lS}) and a vertical part ($\vec{D}^{vU}$, D^{vT} and D^{vS} , by means of A^{vm} , A^{vT} and A^{vS} , see below). This has been done in order to make clear the strong anisotropy between the lateral and the vertical motions. We will discuss both parts of this subgrid-scale physics. Following the laws of Fourier and Fick, the turbulent fluxes have been chosen to depend linearly on the gradients of large-scale quantities. This formulation uses a method similar to that of molecular diffusion and dissipation. For example, the turbulent vertical heat flux is given by $\overline{T'w'} = -A^{vT} \partial_z \overline{T}$. As a result, the resulting diffusive operators are:

$$\vec{D}^{vU} = \frac{\partial}{\partial z} \left(A^{vm} \frac{\partial \vec{U}_h}{\partial z} \right) \quad (2.27)$$

$$D^{vT} = \frac{\partial}{\partial z} \left(A^{vT} \frac{\partial T}{\partial z} \right) \quad (2.28)$$

$$D^{vS} = \frac{\partial}{\partial z} \left(A^{vS} \frac{\partial S}{\partial z} \right) \quad (2.29)$$

This has already been applied in equations (2.13) to (2.25). A level-1.5 turbulence closure scheme, based on a prognostic equation for the turbulent kinetic energy developed by Bougeault and Lacarrère (1989) and Gaspar *et al.* (1990), has been used to calculate the vertical eddy

diffusivity and the viscosity coefficients (Blanke and Delecluse, 1993). Thanks to this parametrisation, momentum is mixed along model level surfaces using coefficients varying with latitude, longitude and depth. The turbulent kinetic energy has been made dependent of the vertical shear, stratification, vertical diffusion and dissipation. In locations with a statically unstable stratification, a value of $100m^2s^{-1}$ is assigned to the vertical eddy coefficients for momentum and tracers. The lateral tracer mixing has been made dependent on the large-scale gradient of the tracer along isopycnals. Furthermore, an eddy-induced advective part has been added following Gent and McWilliams (1990) with decreased coefficients in the tropics (between $20^\circ S$ and $20^\circ N$). Double diffusive mixing (i.e. salt fingering⁴ and diffusive layering⁵) is computed following Merryfield *et al.* (1999) by adjusting the vertical mixing coefficients. The Beckmann and Döscher (1997) bottom boundary layer scheme ensures an improved representation of dense water spreading over topography in this model.

2.2.2 Ice Model: LIM

Having discussed the ocean in the subsection above, we still need to discuss the other important part of the model, i.e. the sea ice model. We have based our discussion on the works of Fichefet and Morales Maqueda (1997); Goosse *et al.* (2000); Timmermann *et al.* (2005) and split it up in a part discussing the vertical growth and decay of sea ice, the lateral growth and decay, the dynamics of the sea ice and the horizontal transport of it. In order to be able to do this, we have split up each grid cell into a part with ice and one without ice.

Vertical growth and decay of sea ice

The vertical growth and decay of the sea ice due to thermodynamic processes is modelled by an improved version of the Semtner (1976) three-layer model, which will be explained here. The ice in the ice-covered part of the grid cell is a homogeneous slab of ice on which snow may accumulate if the surface temperature is below the melting point. The temperature in the snow and ice layers are governed by a one-dimensional

⁴Convection that occurs when warm, salty water overlies cold freshwater.

⁵This happens in the case when cold freshwater overlies warm, salty water.

heat diffusion equation

$$\begin{cases} \rho c_p \frac{\partial T}{\partial t} = \underbrace{\left[1 + \frac{1}{2} \ln \left(\frac{2h_e}{e\epsilon} \right) \right]}_I k \frac{\partial^2 T}{\partial z^2} & \text{if } h_e > \frac{e\epsilon}{2} \\ \rho c_p \frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2} & \text{else} \end{cases} \quad (2.30)$$

where ρ , c_p and k are the density, specific heat, and thermal conductivity, respectively. T is the temperature, t is the time and z is the vertical coordinate. A threshold thickness is introduced in this equation as ϵ and e is Euler's number. The term between square brackets (term I) is a correction factor that accounts for the unresolved ice floe thickness distribution. This is necessary due to the different contribution to the average heat conduction of varying ice floe thicknesses (Mellor and Kantha, 1989). In this equation h_e is an effective thickness for heat conduction defined as

$$h_e = \frac{k_s k_i}{k_s + k_i} \left(\frac{h_s}{k_s} + \frac{h_i}{k_i} \right) \quad (2.31)$$

where subscripts s and i stand for snow and ice and h is the mean thickness. At the surface of the snow-ice system, a balance of heat fluxes is supposed to exist:

$$\underbrace{(1 - i_0)(1 - \alpha_{su})F_{sw}}_I + \underbrace{F_{lw}}_{II} - \underbrace{\epsilon_{su}\sigma T_{su}^4}_{III} + \underbrace{F_h}_{IV} + \underbrace{F_{le}}_V + \underbrace{F_{cs}}_VI = h_{su} (\rho c_p)_{su} \frac{\partial T_{su}}{\partial t} \quad (2.32)$$

with the subscript su denoting the property of a thin surface layer of thickness $h_{su} = 0.1m$. The left hand term denotes the energy change of this surface layer, in which the $h_{su}(\rho c_p)_{su}$ -factor accounts for the thermal inertia of this layer. We will discuss the right-hand terms one by one:

- Term I denotes the part of the incoming shortwave radiation (solar) energy that is absorbed at the surface of the snow/ice. F_{sw} is the shortwave radiation that reaches the surface of the snow/ice. A certain fraction is reflected directly into the atmosphere. This fraction is known as the albedo (α). Another part penetrates into the ice. Following Maykut and Untersteiner (1971) the fraction

that penetrates is defined as

$$i_0 = \begin{cases} 0.18(1 - c) + 0.35c, & \text{ice is free of snow} \\ 0, & \text{snow-covered ice} \end{cases} \quad (2.33)$$

where c denotes the fractional cloud amount.

- Term II is the absorbed longwave radiation.
- Term III is the emitted longwave radiation. ϵ_{su} denotes the emissivity of the surface and σ is the Stefan-Boltzmann constant.
- Term IV is the incoming sensible heat flux.
- Term V is the incoming latent heat flux.
- Term VI is the conductive heat flux coming from below the surface.

Sometimes this balance equation results in a temperature above the freezing point. If there is still snow/ice present in the system, such temperatures will not occur. Instead, the excess energy (B_{su}) will be used to melt the snow/ice. This melting process follows the equation:

$$\left(\frac{\partial h_{s/i}}{\partial t} \right)_{su,abl} = - \frac{B_{su}}{L_{s/i}} \quad (2.34)$$

in which $L_{s/i}$ denotes the volumetric latent heat of fusion of snow if ice is snow-covered and of ice if not.

If it snows on the ice and the load of snow becomes large enough to depress the lower boundary of the snow beneath sea level, a snow-ice cap is formed. This formation is given by

$$-(\Delta h_s)_{si} = (\Delta h_i)_{si} = \frac{\rho_s h_s - (\rho_0 - \rho_i) h_i}{\rho_s + \rho_0 - \rho_i} \quad (2.35)$$

with ρ_0 , ρ_s and ρ_i the reference densities of seawater, snow and sea ice, respectively. The total mass of newly formed ice is supposed to consist of the mass of the soaked snow and the mass of the infiltrated seawater. The latent heat and brine is discharged into the mixed layer. Afterwards the internal temperatures of the ice cover are recomputed. To incorporate

the mechanism of “brine damping”⁶, the approach of Semtner (1976) has been chosen, wherein the solar energy absorbed inside the ice is stored in a heat reservoir that represents internal meltwater. This energy is used to keep the temperature of the upper layer from dropping below the freezing point in autumn, thereby simulating the release of latent heat through refreezing of internal brine pockets. Following Grenfell and Tucker (1977), we assume an exponential absorption of the shortwave radiation within the ice. As a result, Q_{st} , the energy stored in this reservoir, evolves according to

$$\frac{\partial Q_{st}}{\partial t} = i_0 (1 - \alpha_{su}) F_{sw} \left[1 - e^{-1.5(h_i - \epsilon)} \right] - F_{lbp} \quad (2.36)$$

where F_{lbp} is the heat flux necessary to prevent the upward ice layer from cooling below the freezing point of the ice. This heat reservoir is restricted to 50% of the energy required to melt the whole ice sheet. Once this limit is reached, the shortwave radiation is used to directly melt the ice. At the bottom of the ice, any imbalance in the fluxes is compensated by a creation or melt of ice:

$$\left(\frac{\partial h_i}{\partial t} \right)_b = \frac{F_{cb} - F_{wi}}{L_i} \quad (2.37)$$

The exact calculation of the oceanic heat flux (F_{wi}) will be given later.

Lateral growth and decay of sea ice

Next to the vertical growth and decay, the ice can also grow or decay laterally, by increasing/decreasing its fraction of the grid cell covered. Therefore, we define the ice concentration (A), which is defined as the fraction of the grid cell covered by ice. The thermodynamic changes in A depend on B_l (the heat budget of the open water area), for which the calculation will be discussed later on. If B_l is negative, the energy imbalance is covered by an increase in ice concentration at the one hand, and an increase in ice thickness at the other hand. The change in ice concentration is as follows:

$$\left(\frac{\partial A}{\partial t} \right)_{acc} = \underbrace{\sqrt{1 - A^2}}_I \frac{(1 - A)B_l}{L_i h_0} \quad (2.38)$$

⁶Brine pockets are small pockets of very salty water trapped in the ice. Heat is trapped into this pockets, which dampens the decay of sea ice. This phenomenon is called “brine damping”.

Factor I is the term that distributes the energy over the lateral and the vertical growths. The remaining energy imbalance after the lateral growth is used for changing the thickness of the previously existing ice. Then, in order to obtain a single value for the ice thickness, the ice thickness of the new ice and the preexisting is averaged. Afterwards, the snow is distributed over the new covered ice area. Finally, the internal ice temperatures are modified accordingly.

The melting of ice is done by assuming that only vertical melting of thin ice takes place. Assuming that the ice is uniformly distributed in thickness between 0 and $2h_i$ over the ice-covered portion of the grid cell and that the melting rate does not depend on the local sea ice thickness, vertical melting decreases the ice concentration (as the ice with thickness close to 0 is completely melted away)⁷:

$$\left(\frac{\partial A}{\partial t}\right)_{abl} = \begin{cases} \frac{A}{2h_i} \left(\frac{\partial h_i}{\partial t}\right)_{abl-acc}, & \text{if ice ablation} \geq \text{accumulation} \\ 0, & \text{otherwise} \end{cases} \quad (2.39)$$

When the ice concentration decreases, the snow present at the surface of the removed ice is piled up onto the remaining ice.

Ice dynamics

It has been shown (Thorndike, 1986) that inertial forces due to ice acceleration are small compared with other forces acting on the pack. Then, the momentum equation reduces to

$$m \frac{\partial \vec{u}}{\partial t} = \underbrace{-mf\vec{k} \times \vec{u}}_I + \underbrace{\vec{\tau}_{ai}}_II + \underbrace{\vec{\tau}_{wi}}_III - \underbrace{mg\vec{\nabla}\eta}_{IV} + \underbrace{\vec{F}}_V \quad (2.40)$$

where m is the mass of the snow-ice system per unit area and $\vec{u}$ is the sea ice velocity. The different right-hand terms take into account different forces.

- Term I is the Coriolis factor. Here f is the Coriolis parameter and $\vec{k}$ the upward unit vector.
- Term II is the drag imposed by the atmosphere on the sea ice, in units of force per area.

⁷The formation of snow-ice is taken into account in the accumulation part.

- Term III is the drag imposed by the ocean on the sea ice, in units of force per area.
- Term IV is the gravity force due to the horizontal gradient in the ocean surface height. g is the gravity acceleration and η is the surface dynamic height.
- Term V is the internal force of the sea ice, respresented by $\vec{F}$.

Following Hibler (1979), sea ice is assumed to have a viscous-plastic constitutive law, where $\vec{F}$ is the divergence of the two-dimensional stress tensor σ which is given by

$$\sigma = 2\mu\dot{\epsilon} + \left[(\zeta - \mu)T(\dot{\epsilon}) - \frac{P}{2} \right] \mathbf{I} \quad (2.41)$$

where μ and ζ are respectively bulk and shear viscosities. $\dot{\epsilon}$ is the strain rate tensor, $T(\dot{\epsilon})$ is the trace of $\dot{\epsilon}$, P is the ice strength and $\mathbf{I}$ is the two-dimensional unity tensor. The viscosities μ and ζ depend on two invariants of $\dot{\epsilon}$ and on P . The details can be found in Hibler (1979). This rheology allows sea ice to flow plastically for high enough strain rates and to deform in a linear viscous manner for very small rates. The ice strength is parameterised as

$$P = P^* A h_i e^{-C(1-A)} \quad (2.42)$$

in which P^* and C are empirical constants. The contribution of snow to P is ignored.

Horizontal Transport

Changes in different variables are governed by a general conservation law:

$$\frac{\partial \Psi}{\partial t} = -\vec{\nabla} \cdot (\vec{u}\Psi) + S_\Psi + D\nabla^2 \Psi \quad (2.43)$$

where Ψ is one of the variables (ice concentration, snow volume per unit area, ice volume per unit area, snow enthalpy per unit area, ice enthalpy per unit area, latent heat contained in the brine reservoir per unit area), S_Ψ is the rate of change of Ψ due to thermodynamics and D is a horizontal diffusivity to avoid nonlinear instabilities arising from the coupling between ice dynamics and transport. D is constant in the ice pack and zero at the ice edge.

Numerical Methods

The sea ice model uses a B grid arrangement of the variables. In the centre of each cell, the calculations for the scalar variables are made. The velocities are calculated at the corners of each grid cell. The equations are solved, like in the ocean model, by a finite difference scheme. This scheme consists of three parts:

- A fully implicit scheme to solve equation (2.30).
- A Newton-Raphson scheme for equation (2.32).
- A underrelaxed staggered predictor-corrector scheme for the ice dynamics.

2.2.3 ORCALIM, the coupling of OPA and LIM

The coupling is done as in Goosse and Fichefet (1999) and Goosse *et al.* (2000), so this section is based on these works.

Exchange of momentum

To calculate the exchange of momentum between ice and ocean, the ice-ocean stress is taken to be a quadratic function of the difference between the ice velocity $\vec{u}_i$ and the ocean surface velocity $\vec{u}_w$:

$$\tau_{iw} = \rho_0 \frac{k^2}{\underbrace{\left(\ln \frac{|z|}{z_0}\right)^2}_{c_{iw}}} |\vec{u}_i - \vec{u}_w| (\vec{u}_i - \vec{u}_w) \quad (2.44)$$

with τ_{iw} the stress of the ice on the water and k the von Karman constant.

Exchange of heat

Following McPhee (1992), the heat flux from the ocean to the ice (F_{wi}) is taken to be proportional to the difference of temperatures between the surface layer, T_w , and the freezing point T_{fw} . It is also proportional to the friction velocity u_{*io} , the difference of velocity between the ice and the surface layer of the ocean.

$$F_{wi} = \rho_0 c_{pw} c_h u_{*io} (T_w - T_{fw}) \quad (2.45)$$

where c_{pw} is the specific heat of seawater and c_h is a coefficient. T_{fw} is calculated as follows:

$$T_{fw} = 273.15 - (3.0 + 52.75 S_m + 0.04 S_m^2 + 0.0004 S_m^3)10^{-3} \quad (2.46)$$

with S_m the surface layer salinity.

Exchange of Mass

The salt exchanges between ice and ocean are represented as follows:

$$F_{salt} = \frac{(S_w - S_i)}{\rho_0 z_s} \left(\frac{\partial m_i}{\partial t} \right) \quad (2.47)$$

with S_i the sea ice salinity, assumed to be constant. S_w is a reference seawater salinity and m_i is the mass of ice per unit area. The thickness of the upper ocean layer is given by z_s . In this equation, the effects of the ablation of sea ice and the rejection of salt by the seawater when forming sea ice are combined. The ablation of snow and the salt-flux needed to bring the snow salinity to the ice salinity in case of snow-ice formation, is represented by a freshwater flux:

$$F_{fresh} = -\frac{1}{\rho_0} \left(\frac{\partial m_s}{\partial t} \right). \quad (2.48)$$

2.2.4 Configuration, Initialisation and Forcing

Configuration and Numerical Methods

The model is used here in a global configuration on a domain that extends from 78°S to 90°N. The model uses a tripolar grid (one pole in Canada, one in Siberia, one on the South Pole) with a zonal resolution of 2° (Figure 2.3). The meridional resolution varies from 0.5° at the equator to 2° cos φ , south of 20°S. Local mesh refinements are applied to the Mediterranean, the Red, the Black and the Caspian Seas. Minimum horizontal resolution around Antarctica is 50km. Bottom topography and coastlines are derived from the study of Smith and Sandwell (1997), complemented by the ETOPO5 dataset⁸. The vertical discretisation uses 30 levels in the water column with 10 levels in the top 100m. The ocean model is run with a time step of 5760s (which converts to 15 time steps a day) and coupled bidirectionally to the ice model every five steps.

⁸<http://www.ngdc.noaa.gov/mgg/global/etopo5.HTML>

Symbol	Definition	Value	Unit
a	Earth's radius	6.371×10^6	m
A_M	Diffusive Coefficient for unstable stratification	100	$m^2 s^{-1}$
$atfp$	Assalin filter parameter	0.1	
c_{pi}	Specific heat of sea ice	2093	$Jkg^{-1}K^{-1}$
c_{ps}	Specific heat of snow	2093	$Jkg^{-1}K^{-1}$
c_{pw}	Specific heat of seawater	4000	$Jkg^{-1}K^{-1}$
C	Ice strength parameter	20	
C_{iw}	Water drag coefficient	5×10^{-3}	
C_{aw}	Atmosphere-Ocean drag coefficient	2.25×10^{-3}	
C_{ai}	Atmosphere-ice drag coefficient	2.25×10^{-3}	
C_{db}	Bottom of the ocean drag coefficient	4×10^{-4}	
c_h	Coefficient in heat exchange	0.006	
D	Horizontal Diffusivity	350	$m^2 s^{-1}$
ϵ	Threshold thickness for the ice temp. eq.	0.1	m
ϵ_{su}	Emissivity of snow or ice	0.97	
ϵ_w	Emissivity of seawater	0.97	
g	Acceleration due to gravity	9.80665	ms^{-2}
h_0	Thickness of newly formed ice in leads	1	m
k	Von Karman Constant	0.4	
k_i	Thermal conductivity of sea ice	2.0344	$Wm^{-1}K^{-1}$
k_s	Thermal conductivity of snow	0.22	$Wm^{-1}K^{-1}$
L_i	Volumetric latent heat of fusion of sea ice	300.330×10^6	Jm^{-3}
L_s	Volumetric latent heat of fusion of snow	110.121×10^6	Jm^{-3}
P^*	Ice strength parameter	20000	Nm^{-2}
om	Underrelaxation param. ice dyn. num. method	0.5	
ρ_i	Density of sea ice	900	kgm^{-3}
ρ_s	Density of snow	330	kgm^{-3}
ρ_0	Reference density of seawater	1029	kgm^{-3}
S_i	Salinity of sea ice	6	psu
S_w	Reference salinity of seawater	34.7	psu
σ	Stefan-Boltzmann Constant	5.67×10^{-8}	$Wm^{-2}K^{-4}$
T_{fi}	Melting point of sea ice	273.05	K
T_{fs}	Melting point of snow	273.15	K
Ω	Earth's angular velocity	7.2722×10^{-5}	s^{-1}
z_0	Roughness length in ocean model	0.017	m
z_s	Thickness of the upper layer of the ocean	10	m

Table 2.1: Numerical Values of the Model Constants and Parameters.

Initialisation

Model runs are initialised using the January data from the Polar Science Center Hydrographic Climatology (PHC, Steele *et al.*, 2001). An initial sea ice thickness of $3m$ in the Arctic and $1m$ in the Antarctic is prescribed in regions with a sea surface temperature (SST) below 0°C . An initial snow cover with a thickness of $0.5m$ in the Arctic and $0.1m$ in the Antarctic is assumed. The model is then run through the period 1977-1999. Afterwards, the model has been run again from 1970 on. The data used in this work are taken from 1979 out of the second run. Further description of the initialisation can be found in Timmermann *et al.* (2005).

Forcing

Daily 2-m air temperatures and 10-m winds from the NCEP/NCAR reanalysis project for the simulated period (Kalnay *et al.*, 1996), and monthly climatologies of surface relative humidity (Trenberth *et al.*, 1989), cloud fraction (Berliand and Strokina, 1980), and precipitation rate (Xie and Arkin, 1996) are used to drive the model. River runoff rates are prescribed using the climatological Baumgartner and Reichel (1975) dataset combined with a mean seasonal cycle derived from the Global Runoff Data Center (GRDC, 2000) data. The surface fluxes of heat and moisture are determined from these data by using empirical parameterisations described by Goosse (1997). Evaporation/sublimation is derived from the turbulent flux of latent heat. In some simulations, a correction of ocean surface freshwater fluxes has been derived from the time-mean restoring salinity flux diagnosed from the experiment described by Timmermann *et al.* (2005). This has been done in order to circumvent the surface salinity drift that occurs through the lack of any stabilising feedback if the model is forced with slightly incorrect evaporation, precipitation and/or river runoff fields. Disabling this salinity restoration showed the onset of important convection along the East Antarctic Coast, reducing the Southern Ocean ice volume significantly. Furthermore, after some years of integration, the water mass in the Southern Ocean tends to become homogeneous resulting in a stronger ACC (220 Sv instead of 170 Sv through the Drake Passage). These simulations are useful to investigate the interannual variability of the salinity. In the other simulations, the salinity of the ocean is restored towards the observed values.

In general, these simulations are more stable. The momentum fluxes at the various interfaces are obtained from standard bulk formulas. These will not be discussed in here, but can be found in Chapter 5 of the CLIO manual (Goosse *et al.*, 2000).

One could pose oneself the question why we use the reanalysis data and not directly the observations. During the second half of the 20th century, there has been an important increase in the number of meteorological observations. Sometimes, strange jumps in some variables can be found and it is not clear if these jumps are real or just due to changes in the observation methods (Kalnay *et al.*, 1996). In addition, the observations cannot be used to directly force a model, as they are not gridded. In other words, the observations have not been made on a regular grid and are, in some regions (like Western Europe), far more abundant than in others (for instance, above the oceans). Therefore, several centres have been making reanalyses. Their goal is to provide a gridded dataset of the important variables which are not biased by changes towards newer versions of the model, and with a smaller bias due to changes in observational methods. This can be done for different mediums, like ocean and sea ice. However, the most known and most used reanalyses have been made for the atmosphere. The one used in this work is the NCEP/NCAR-reanalysis (Kalnay *et al.*, 1996). We will try to discuss how a reanalysis is made and how it can be used.

First of all, the data are prepared for the reanalysis. They are submitted to a quality control, hopefully eliminating all the problems that possibly occur. The observations that have been used are land surface observations, ship observations, rawinsonde, aircraft, satellite, to name just a few. These data are then preprocessed to have an uniform format. Again, quality control is enforced, comparing the different datasets. The resulting data are fed into a state-of-the-art atmospheric model, which will not change during the reanalysis, in order to have homogeneous results. During the model runs the observations constrain the model results. For more reliable observations like wind, temperature and pressure the model is strongly constrained. For other variables like heat fluxes, snow depth, precipitation and cloud fractions, the model is not constrained directly as no reliable observations exist. The results of the model are put again in an uniform format and are subjected to a final quality control. Afterwards, they are released to the public.

The NCEP/NCAR reanalyses have shown to be reliable in most cases and very useful to force a model. Furthermore, in many cases, it can be used as a replacement of observational data when comparing model results to the observations. However, it is known that there are some problems with the reanalyses data, especially before 1979. In this period, the absence of reliable satellite data has its influence on the reanalysis. This is especially true in polar regions, where other data are sparse. More information on the problems of the NCEP/NCAR-reanalyses can be found for instance in Hines *et al.* (2000); Marshall (2002); Struraro (2003).

In the ORCALIM model, the parameterisation of Large and Pond (1981) is utilised to compute the drag coefficient between air and water. For the drag coefficients between air and snow/ice and between ice and water, we use constant values of $2.25 \cdot 10^{-3}$ and $5 \cdot 10^{-3}$, respectively (see also Table 2.1). This leads to a ratio of these two coefficients of 0.45, which is consistent with values adopted in other modelling studies of the Antarctic sea ice (e.g. Stössel, 1992; Fischer and Lemke, 1994; Harder and Fischer, 1999). Further details of the coupled model are given by Timmermann *et al.* (2004, 2005). The quality of the NCEP/NCAR dataset is questionable in polar regions because of the small number of observations there, in particular for the period prior to the start of the assimilation of satellite data in the late 1970s (e.g. Hines *et al.*, 2000; Marshall, 2002; Struraro, 2003). As a consequence, the analysis presented here will be performed on the last 20 years of the experiment (i.e. the second run). Even during this period, some uncertainties remain. Sensitivity experiments (Timmermann *et al.*, 2005) have shown that the use of the complete set of NCEP/NCAR daily forcing data (so, in addition to the air temperature and the wind speed, also using the daily specific humidity, total cloudiness and precipitation derived from the reanalysis), strongly diminishes the sea ice thickness in both the Arctic and the Southern Ocean. As a result, in summer, the ice almost disappears in the Southern Ocean, while unrealistic open water polynya's form in the Arctic Ocean. Further sensitivity experiments (Timmermann *et al.*, 2005) have shown that this comes from the use of the NCEP/NCAR daily specific humidity. However, taking into account a daily varying cloudiness and precipitation coupled to a climatological humidity does not amount to significantly better results than the control run. Therefore, the control

run (daily winds and temperatures, climatological humidity, cloudiness and precipitation) has been chosen as the run to analyse.

2.2.5 The geography and climate of the Southern Ocean

Before being able to validate the model, some information about the geography and the climate of the Southern Ocean is needed. This information can be found in this subsection.

Geography of the Southern Ocean

As the world oceans are all connected, it remains rather arbitrary how many oceans cover our world. Before 2000, four oceans were defined by the International Hydrographic Organization⁹ (IHO): the Arctic Ocean, the Pacific Ocean, the Indian Ocean and the Atlantic Ocean. A fifth one was added to these four: the Southern Ocean. The Southern Ocean is defined as the water body south of 60°S. This definition is very convenient as the border of the Southern Ocean never crosses land. Furthermore, the 60°S latitude line coincides with the Antarctic Treaty Limit. The covered surface of the Southern Ocean is 20.327 million km^2 , slightly more than twice the size of the United States and larger than the Arctic Ocean. However, this definition of the Southern Ocean is not convenient for the purposes of this work as, for instance, the winter sea ice crosses this border at many places. Furthermore, it does not correspond to a separation of different water masses, nor to a physical barrier. As a result, we will discuss not only the Southern Ocean, but also the southern parts of the Pacific, Atlantic and Indian Oceans. For convenience, we will not follow the IHO compromise and we will call henceforth the ensemble of all these ocean parts the Southern Ocean. In general, when analysing the sea ice, the border of the Southern Ocean can be put at 50°S, as sea ice normally does not cross this line.

The Southern Ocean is the only ocean of the world which covers every longitude. As a result, it plays a very important role in exchanging water masses between the different oceanic basins. The Southern Ocean is deep, 3000 to 5000 meters over most of its extent with limited areas of shallow water. The Antarctic continental shelf is in general narrow and unusually deep, the edge lying at 400 to 800m depth (Figure 2.4). The Southern

⁹<http://www.iho.shom.fr/>

Ocean can be split up in five sectors as, for instance, in Zwally *et al.* (2002) and Lefebvre *et al.* (2004): the Weddell Sea (60°W-20°E), the Indian sector (20°E-90°E), the Pacific sector (90°E-160°E), the Ross Sea (160°E-140°W) and the Amundsen-Bellingshausen Seas (140°W-60°W) (Figure 4.6). The major chokepoint of the Southern Ocean is the Drake Passage between South America and the Antarctic continent.

The mean climate of the Southern Ocean

The temporal mean atmospheric pressure above the polar Southern Hemisphere is characterised by a high sea level pressure area on the Antarctic continent¹⁰, in general centred above East Antarctica¹¹ and a band of low pressure areas surrounding the continent (Figure 2.5). Further north, the sea level pressure rises again towards the subtropical high pressure ridge (Simmonds, 2003). Although there is large uncertainty, it is clear from the observations that the mean pressure on the Antarctic continent is higher in the local winter than in local summer. Except for this high pressure area, which is centered on East Antarctica, the pressure distribution is rather zonal with a slightly higher pressure on the Antarctic Peninsula. This pressure distribution induces strong westerlies in most of the Southern Ocean, and easterlies close to the coast of the Antarctic continent.

The temperature distribution over the Southern Ocean is rather zonal, with warmer air more to the north and colder air to the south. At the borders of the Antarctic continent where the slopes are large, mostly in east Antarctica, the air is cooled by strong katabatic winds (Parish, 1988). These winds are due to cold air rushing down the slopes of the Antarctic ice sheet as it is denser than the warmer air around. These winds are prominent over most of the Antarctic ice sheet during most of the time.

In general, the ocean currents follow the winds in the Southern Ocean. This means a strong and large eastward current away from the coast and a westward current close to the coasts of the Antarctic Continent. However, the currents are also deviated by the bathymetry. For instance, the

¹⁰Due to the height of the Antarctic continent, the pressure over the central Antarctic is low, despite a high sea level pressure.

¹¹The Transantarctic Mountains divide the Antarctic continent in two parts, the smaller part being West Antarctica and the larger one being East Antarctica.

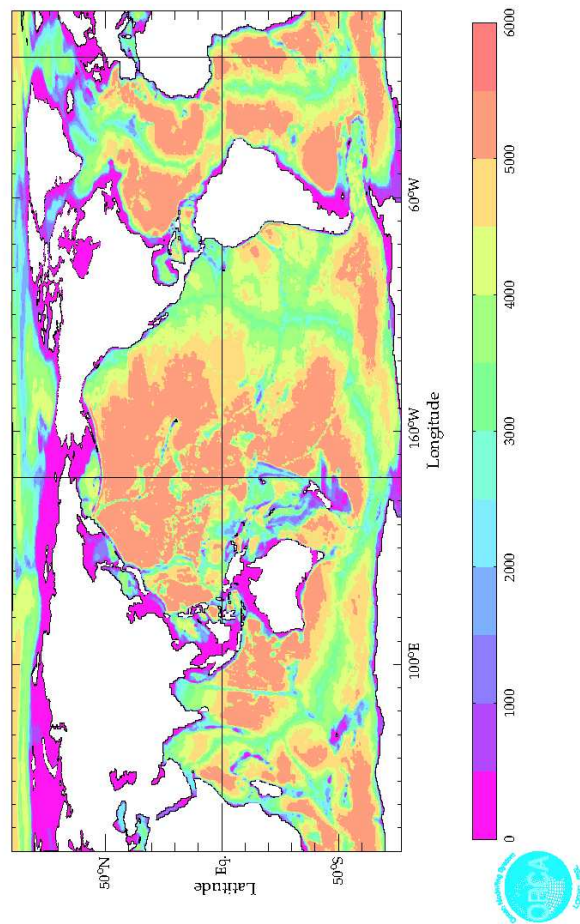


Figure 2.4: The bathymetry of the ORCALIM model at a resolution of half a degree. This bathymetry can be seen as rather precise thanks to the high resolution.

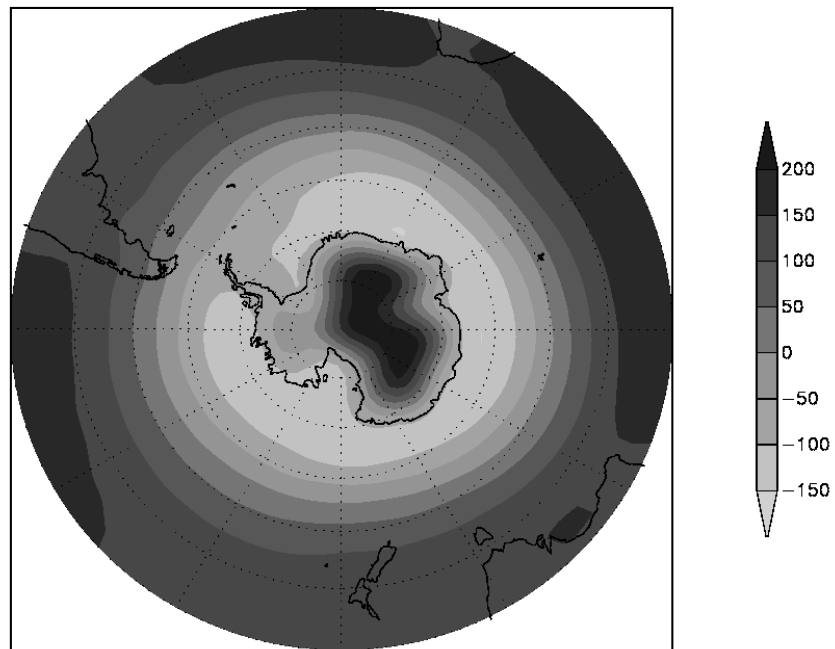


Figure 2.5: The annual mean geopotential height of the 1000hPa-level, calculated over the period 1979-2006, in the NCEP/NCAR reanalysis.

strong eastward current mentioned above, known as the Antarctic Circumpolar Current (ACC), is deviated to the north after it passes Drake Passage¹². This is due to the Scotia Ridge¹³, disabling the possibility for the current to go straight through. There are three gyres¹⁴ visible in the currents of the Southern Ocean. These are the Kerguelen¹⁵, Ross and Weddell Gyres. For instance, the Weddell Gyre has a northward branch just off the eastern coast of the Antarctic Peninsula, turning eastward with the ACC. The southward branch can be found in the eastern Weddell and the western Indian sectors and the loop is closed by the westward Coastal Current.

The sea ice cover in winter has a mostly annular form, with a more northward extended ice edge in the Weddell sector, just east of Prydz Bay and in the eastern Ross sector (Figure 2.9). This is due to the gyres discussed above. The concentrations of the winter sea ice are in general high throughout the complete pack (Figure 2.9). In summer, sea ice is present in the western Weddell sector, in the southern Amundsen-Bellingshausen Seas and in a small coastal band in the Indian and the Pacific sector (Figure 2.8). Finally, part of the eastern Ross sector remains covered with some sea ice during summer. The average sea ice extent is about 17.1 million km^2 in winter and 3.9 million km^2 in summer (1979-1998, Zwally *et al.*, 2002).

The interannual variability of the sea ice in the Southern Ocean

In addition to the mean sea ice concentration and extent, it is important to talk about the interannual variability of the sea ice. In comparison to the Arctic sea ice, the sea ice extent in the Southern Ocean does not show a large trend since the beginning of the satellite era (Cavalieri *et al.*, 2003;

¹²As said above, this is the passage between South America and the Antarctic Peninsula.

¹³A ridge consisting out of two parts, the Northern Scotia Ridge and the Southern Scotia Ridge. The Northern Scotia Ridge extends from the tip of South America (approximate coordinates 54°S, 65°W) over the Falkland Islands (52°S, 60°W) and South Georgia (54°S, 37°W) to the Southern Sandwich Islands (58°S, 25°W). The Southern Scotia Ridge starts at the top of the Antarctic Peninsula (63–64°S, 58°W) and the Southern Shetland Islands (62°S, 60°W) over the Southern Orkney Islands (61°S, 45°W) towards the Southern Sandwich Islands where it connects to the Northern Scotia Ridge.

¹⁴A gyre is a large-scale ocean circular current.

¹⁵Named after the Kerguelen Islands which can be found at 49°S, 70°E.

Lefebvre and Goosse, 2007a). Indeed, the trend in the Southern Ocean sea ice extent over the period 1979-2004 is slightly positive, amounting to 0.21 million km^2 per decade. This trend can be split in regional ones giving two sectors with almost no trend (Weddell and Indian sectors), one sector with a small positive trend (Pacific sector), one with a small negative trend (Amundsen-Bellingshausen sector) and finally one with a larger positive trend (Ross sector). Trend maps (Figure 7.5a) show a dipole, with a negative trend around the Antarctic Peninsula and a positive trend in the Ross and Pacific sectors.

The total variability of the sea ice extent over the same period amounts only to 0.21 million km^2 , which is small compared to the mean extent (12.44 million km^2 , 1,68%). The sectoral variability is of the same order as the total variability, amounting to a more important relative variability. The sector with the largest variability is the Weddell sector (with an absolute variability higher than the absolute variability of the complete Southern Ocean).

More information about the interannual variability of the sea ice in the Southern Hemisphere can be found in Zwally *et al.* (2002); Cavalieri *et al.* (2003); Lefebvre and Goosse (2007b).

2.2.6 Comparison of the model results with the observations

The model has been validated for the polar regions by Ralph Timmermann and co-authors in two different papers (Timmermann *et al.*, 2004, 2005).

Sea ice

The summer sea ice extent in the Southern Ocean is underestimated by about 40% (Figures 2.6 and 2.8). The winter sea ice extent corresponds well to the observations (Figures 2.6 and 2.9). The yearly maximum of the sea ice extent is a little early in the model (Figure 2.6). Time series of monthly anomalies indicate that the model captures only part of the interannual ice extent variability, despite representation of the most pronounced sea ice extent anomalies. Specifically, the decline in the sea ice extent in the seventies and the low summer ice extents in the austral summers 1979/1980 and 1986/1987 are well reproduced (Figure

2.7). However, the model produces also a low sea ice extent anomaly in the beginning of the nineties, which has not been observed. Furthermore, at a regional scale, the low ice extent in the Bellingshausen Sea at the end of 2005 and the beginning of 2006 is simulated (Massom *et al.*, 2007). However, a similar phenomenon at the end of 2001 (Massom *et al.*, 2006) is not well reproduced.

The simulated geographic distribution of the sea ice in the Antarctic features a lack of sea ice near the tip of the Antarctic Peninsula at the moment of the annual minimum sea ice distribution (Figure 2.8). This is probably due to the poor representation of the Antarctic Peninsula in the forcing. Furthermore, there is a lack of sea ice along the coasts of the Indian and the Pacific sector, which is a common feature in models. This is due to the resolution which is too coarse to represent correctly the small band that is ice-covered in summer and to the too weak katabatic winds in the forcing. During the austral winter, an ice cover is present in large parts of the Southern Ocean (Figure 2.9). Ice is led outward due to the three important gyres, the Kerguelen, Ross and Weddell Gyres. These gyres are well represented in the model. In the Amundsen-Bellingshausen Seas, there is too much ice in the model in winter. Finally, there is a significant underestimation of multi-year ice in the Weddell sector. For a more detailed validation of the response of the sea ice to the atmospheric modes of variability we refer the reader to Chapter 3.

The ice thickness in the Southern Ocean which has been investigated in Timmermann *et al.* (2004, 2005) (Figure 2.10), seems in general in reasonable agreement with the observations, except along the coasts. When comparing specific points to the measurements made by Upward Looking Sonars (ULSs) in the Weddell Sea, we find that ice thickness is seriously underestimated at both the eastern Weddell sector coast and off the eastern coast of the Antarctic Peninsula. However, comparison to other ULSs shows a very good agreement. The problems in the model ice thickness are probably connected to scale problems as the relevant scales for the important processes (for instance, deformation of sea ice in case of onshore winds) in these regions are smaller than the grid scale. The problems in the Western Weddell Sea could be linked to the warm bias in the forcing data for that region, disabling the formation of thick, multi-year ice. However, the model generally tends to underestimate the dynamical thickening of the Antarctic sea ice.

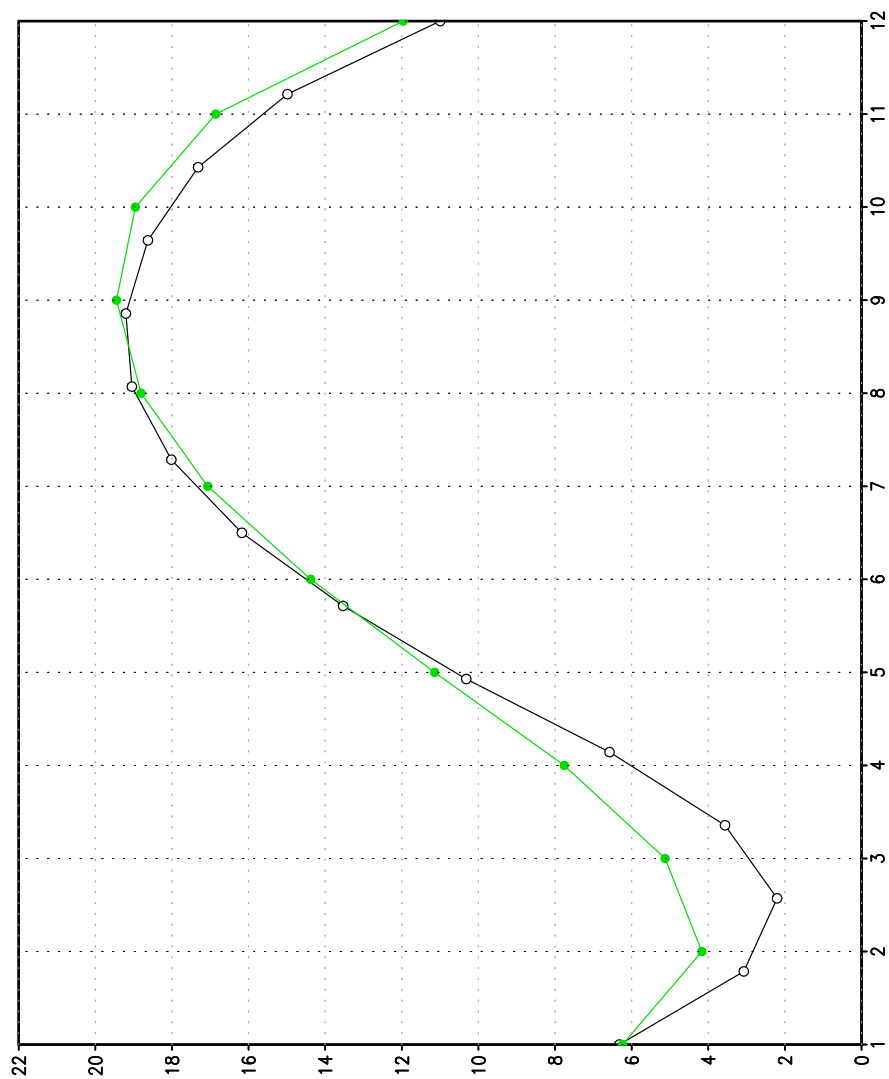


Figure 2.6: The mean annual cycle of the Southern Ocean ice extent (in 10^6 km^2) in the ORCALIM-model (empty circles) and in the HADISST-observations (full circles). The numbers on the x-axis denote the months. The average has been taken over the 30-year period 1970-1999.

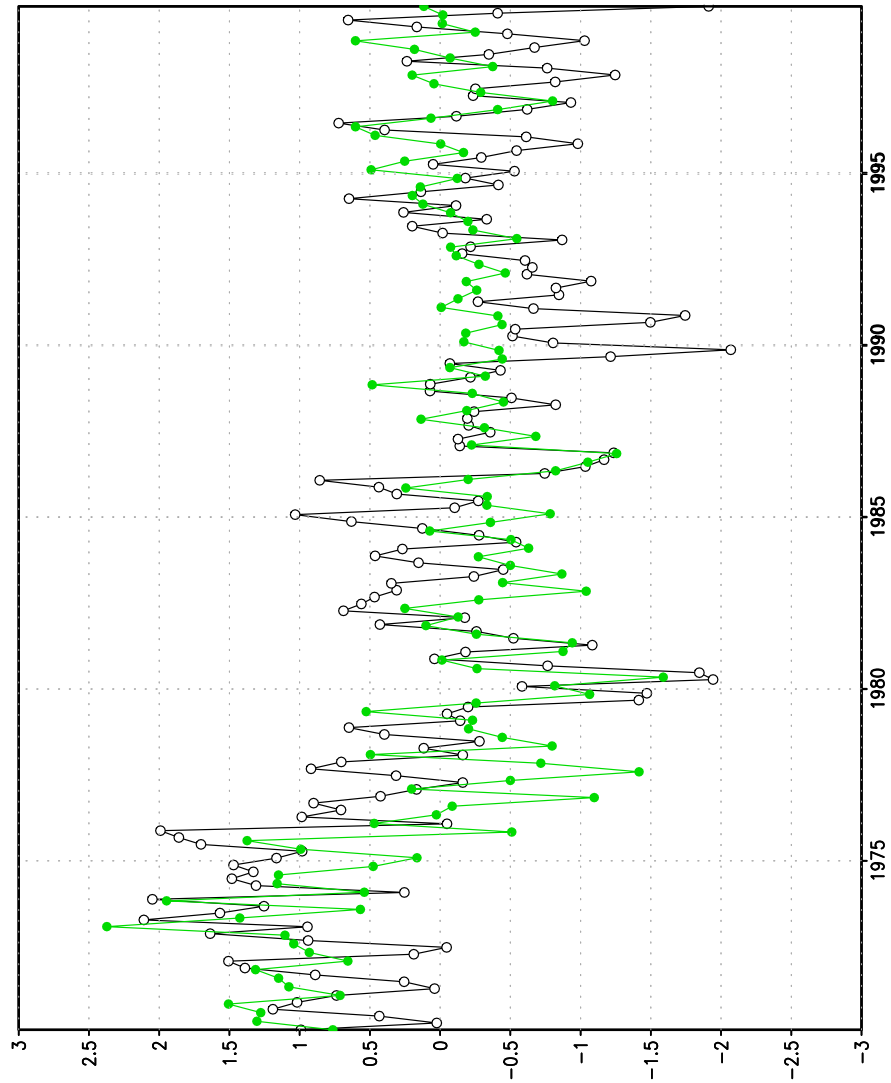


Figure 2.7: The Southern Ocean sea ice extent minus the climatological one (in 10^6 km^2) in the ORCALIM-model (empty circles) and in the HADISST-observations (full circles). The climatology has been calculated on the same 30-year period 1970-1999.

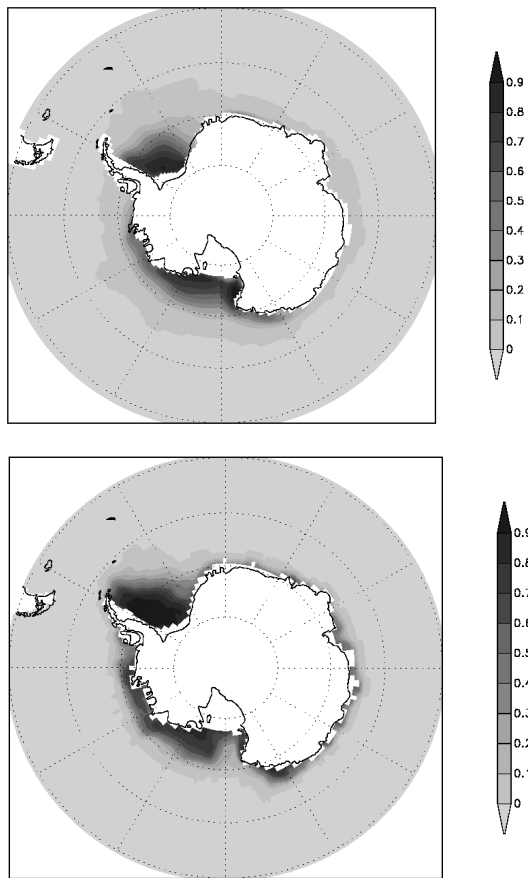


Figure 2.8: The March mean sea ice concentration in the ORCALIM-model (above) and in the observations (HADISST, below) (Rayner *et al.*, 2003). The mean has been taken over the period 1979-2004.

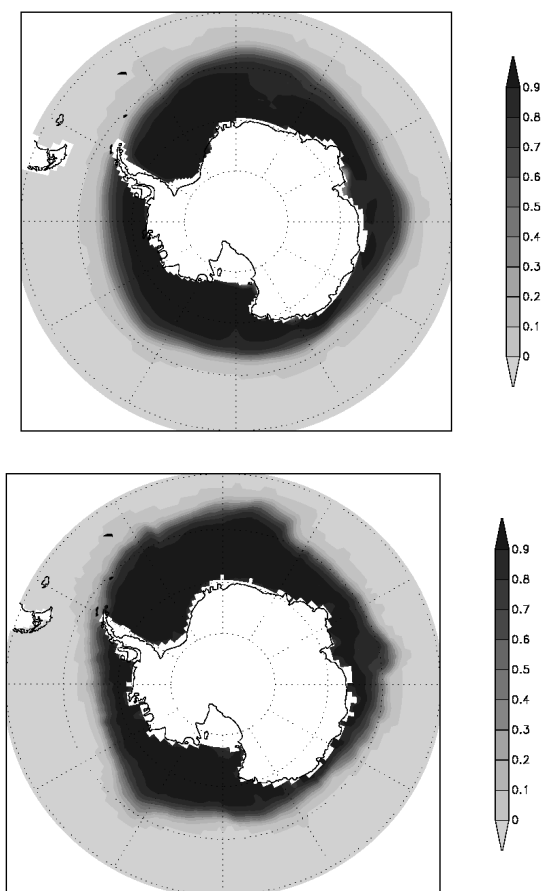


Figure 2.9: The September mean sea ice concentration in the ORCALIM-model (above) and in the observations (HADISST, below) (Rayner *et al.*, 2003). The mean has been made over the period 1979-2004.

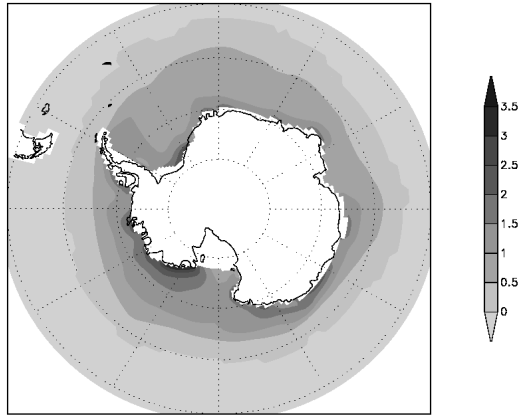


Figure 2.10: The September mean sea ice thickness in the ORCALIM-model. The mean has been taken over the period 1979-2004.

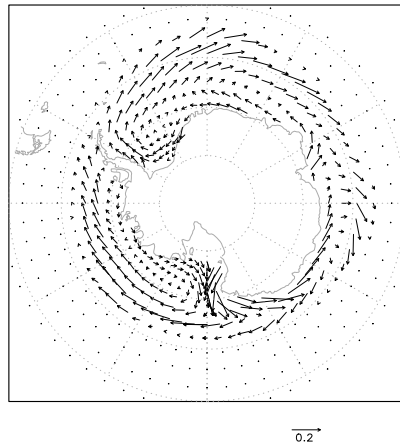


Figure 2.11: The September mean sea ice velocity (in $\frac{m}{s}$) in the ORCALIM-model. The mean has been taken over the period 1979-2004. For clarity, only one ninth of the vectors has been shown.

The ice thickness in the Southern Ocean has also been validated using ASPeCt (Antarctic Sea Ice Processes and Climate)-data (Timmermann *et al.*, 2004). Their results show, in agreement with the discussion above, a good correspondence between the model and the observation data. However, a negative bias in the northwestern Weddell sector and a positive one in the eastern Bellingshausen Sea seem to be present. This is probably due to a poor representation of the Antarctic Peninsula in both the model and the reanalysis model. This last problem leads also to an overestimation of the wind speeds on the Peninsula. However, there is little agreement between the individual observations and the corresponding model thicknesses due to the size of the model grid cell.

The large-scale sea ice drift on both hemispheres is closely linked to the winds and the ocean currents. This leads to a generally good representation of the sea ice drifts. There is a mean eastward drift in the northern part of the ice pack in the Southern Ocean, and a coastal westward drift further south (Figure 2.11). The three gyres (Weddell, Ross and Kerguelen) are well represented. Both qualitatively and quantitatively, the model simulated ice drift concurs with the remote sensing data.

In the Arctic, the annual cycle of sea ice growth and decay is realistically captured. Ice extent, ice thickness and ice drift are in close agreement with the observations. However, there is a slight overestimation of summer ice extent in the Arctic Ocean. The sea ice trend in the Arctic is not well covered, as it seems to be unrealistic insignificant for the period 1977-1999 (the period on which the validation was done). However, that could be due to the method of validation in which the investigated period is run just after an initial cycle of the same period. As a result, the ice conditions in the beginning of the period are the ones of the end of the nineties instead of the ones at the end of the seventies. The model has too much ice in the Kara and Baffin Seas in summer, while in winter the Baffin and Greenland Seas have slightly too much ice. The ice thickness in the Arctic is overestimated along the Siberian coast and in the Canadian Basin, but is good elsewhere. As a result, the thickness gradient between them is well reproduced. The model reproduces rapid thinning in the ice since 1988, which is in good agreement with the observations.

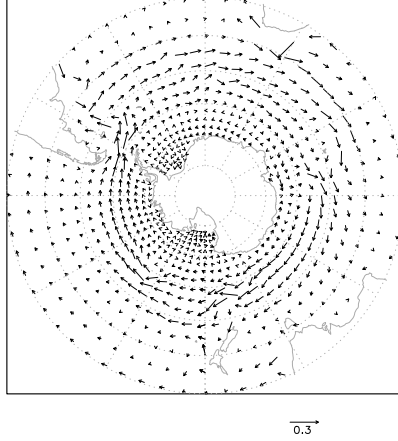


Figure 2.12: The annual mean surface ocean velocity in the ORCALIM-model. The mean has been made over the period 1979-2004. For clarity, only one ninth of the vectors has been shown.

Ocean

The location of the large-scale ocean currents is well represented in the model, both in the Arctic and the Antarctic region. We will focus here on the validation of the latter. The surface velocities in the Southern Ocean (Figure 2.12) depict clearly the westward Antarctic Coastal Current, the eastward Antarctic Circumpolar Current (ACC) and the Weddell, Ross and Kerguelen Gyres. The vertically integrated transport through Drake Passage in the model is too high, amounting to a $170Sv$ average, which is significantly higher than the net observed transport of $(137 \pm 8)Sv$ (Cunningham *et al.*, 2003). However, the simulated strength of the Weddell Gyre ($20Sv$, Timmermann *et al.*, 2004) is below the observations ($29.5Sv$, Fährbach *et al.*, 1994). Furthermore, the ACC is too wide and extends too far southwards. As a result, the development of a deep westward coastal current is inhibited in the Bellingshausen Sea and parts of the East Antarctic.

If the mixed layer depth¹⁶ is big, this can be an indication for convection sites. In the Southern Hemisphere, these sites are found on the

¹⁶which is defined in the model as the depth at which the potential density exceeds surface density by more than $0.01 \frac{km}{m^3}$

continental shelves, in the Weddell Basin and around Maud Rise. The mixed layer depths are the largest between late March and the beginning of August, except for the Maud Rise region.

2.3 Global Coupled Atmosphere-Ocean-Sea Ice-Land Models

Several institutes have been using coupled atmosphere-ocean-sea ice-land models (further on denoted with GCM, General Circulation Model) in order to project the future in the framework of the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC). In this work, these models have been used in Chapter 7. An extensive description and validation of the models can be found in Randall *et al.* (2007) and the references therein. The models used here are the ones that provided the data necessary for the investigation as explained in Chapter 7 and can be found in Table 7.1. Further information on a specific model can be found in Randall *et al.* (2007, Table 8.1 and references therein).

Chapter 3

The most important modes of atmospheric variability and their influence on the climate of the Southern Ocean

3.1 Explanation of some key mathematical methods

Before we are able to discuss thoroughly the climate variability of the Southern Ocean, it is important to take a look at some mathematical methods used in this work. The goal of this section is not to give a complete mathematical vision of the methodology, but to show how these methods can be used in this work. Therefore, we will also discuss the advantages and disadvantages of the respective methods. This section is based on the works of Snedecor and Cochran (1971) and Wilks (1995).

3.1.1 Linear Trend

The linear trend allows us to see if there is a systematic change in a set of data. Let us take a time series of data, for instance, the temperatures at one specific point. These data, named A , consist of n values A_i . These

data are taken at times t consisting of n values t_i . Let us define:

$$\bar{A} = \frac{\sum_{i=1}^n A_i}{n} \quad (3.1)$$

$$\bar{t} = \frac{\sum_{i=1}^n t_i}{n} \quad (3.2)$$

as the means of both time series and

$$A'_i = A_i - \bar{A} \quad (3.3)$$

$$t'_i = t_i - \bar{t} \quad (3.4)$$

as the anomalies of the data. Then the linear trend, following the least square method, can be defined as:

$$b = \frac{\sum_{i=1}^n A'_i t'_i}{\sum_{i=1}^n (t'_i)^2} = \frac{n \sum_{i=1}^n (A_i t_i) - \sum_{i=1}^n A_i \sum_{i=1}^n t_i}{n \sum_{i=1}^n (t_i^2) - (\sum_{i=1}^n t_i)^2} \quad (3.5)$$

b is called the linear trend. Then, the data A can be estimated by $\hat{A}$, while A' can be estimated by $\hat{A}'$ using:

$$\hat{A}' = b t' \quad (3.6)$$

$$\hat{A} - \bar{A} = b(t - \bar{t}) \quad (3.7)$$

$$\hat{A} = b t + (\bar{A} - b \bar{t}) \quad (3.8)$$

Let us now call d_i the difference between A_i and $\hat{A}_i$:

$$d_i = A_i - \hat{A}_i \quad (3.9)$$

So d_i represents the discrepancy between the estimated and the real values. Then, due to our method, $\sum_{i=1}^n d_i = 0$. The standard deviation of the linear trend (s_b) is then given by (Snedecor and Cochran, 1971, section 6.2)

$$s_b = \sqrt{\frac{\sum_{i=1}^n d_i^2}{(n-2) \sum_{i=1}^n (t'_i)^2}} \quad (3.10)$$

However, care has to be taken when using the linear trend method. The trend and its standard deviation are only valid for the period of the analysed sample. This can pose problems in some situations. For example, suppose that our real values follow a long period oscillation and the

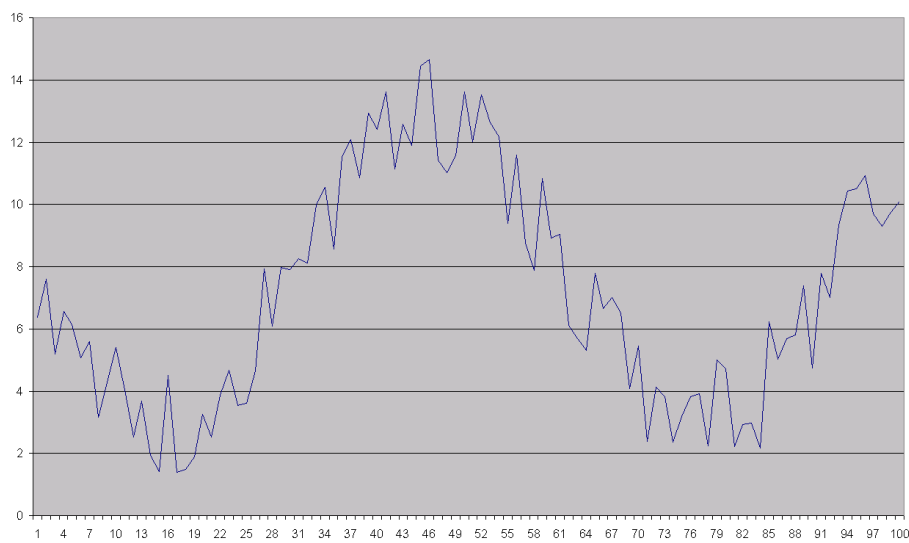


Figure 3.1: Suppose that the complete dataset has been shown on the figure above. We can see that the data represent a oscillating function with noise superposed on it. However, if only part of the dataset would be available, wrong conclusions could be taken. Suppose that only the data between point 20 and 40 are available, the graph could be seen as rising and we could by extrapolation predict a further increase. If points 50 to 70 were available, a decreasing extrapolation would have been made.

values on which the trend is calculated are all in the ascending branch. Then, the trend will have a strong positive value and a small s_b . However, concluding that the values will continue to rise with the established trend would be a mistake (see Figure 3.1).

Another problem that can affect the results is that this is only a first-order approximation. As a result, for instance, data that have in reality an exponential curve will be wrongly represented by a linear fit.

3.1.2 Correlations and regressions

Correlation and regression are two quantities which allow to express how strong two datasets are related. Suppose that we have two data series A and B each containing n values A_i and B_i . Then, we can define $\bar{A}$, $\bar{B}$, A'_i and B'_i as above (equations 3.1 and 3.3). The correlation (C_{AB}) between the two data series is then defined by

$$C_{AB} = \frac{\sum_{i=1}^n A'_i B'_i}{\sqrt{\sum_{i=1}^n (A'_i)^2 \sum_{i=1}^n (B'_i)^2}} = C_{BA} \quad (3.11)$$

The correlation always ranges between -1 and 1 . If the correlation is equal to -1 the data series are perfectly anticorrelated, i.e., $\forall i \Rightarrow B'_i = aA'_i$, with $a < 0$. On the contrary, if the same equation applies, but with $a > 0$, the correlation between A and B will be 1 . If the correlation is 0 , it appears that the time series are not linked. However, care has to be taken in the same way as with linear trends. Indeed, this method examines only for linear relationships (see Figure 3.2).

Furthermore, a correlation does not prove a causal link between the two parameters. Even with high correlations between A and B , we cannot distinct whether it is A which causes B , or B which causes A , or A and B caused by a series C and so on. The regression of B onto A is calculated as follows

$$R_{BA} = C_{BA} \sqrt{\sum_{i=1}^n (B'_i)^2} = \frac{\sum_{i=1}^n A'_i B'_i}{\sqrt{\sum_{i=1}^n (A'_i)^2}} \neq R_{AB} \quad (3.12)$$

The value R_{BA} describes how much B will rise if A rises with 1 standard deviation. The same limitations as for the correlations are valid for the regressions.

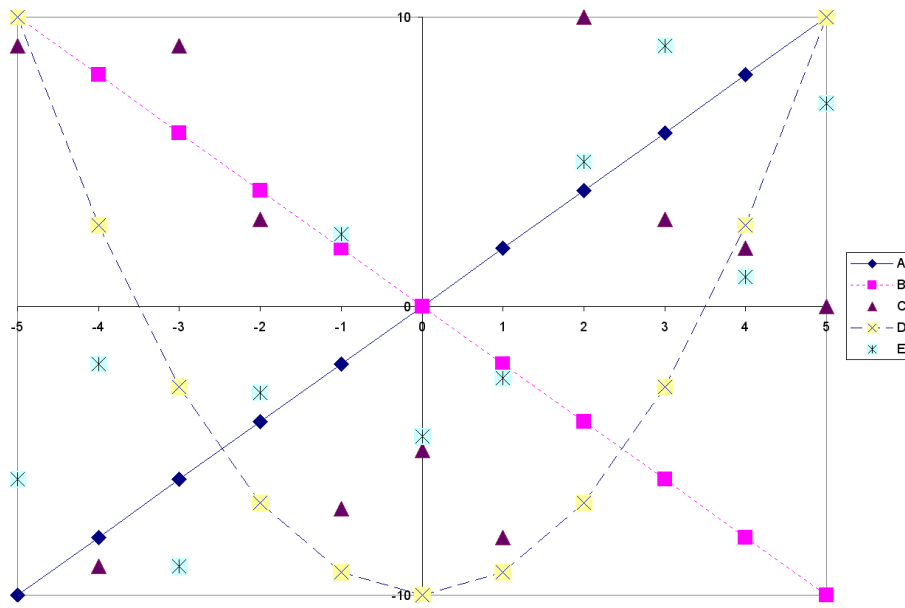


Figure 3.2: Some examples of the use of correlations. The plot shows for 5 series a scatter plot for which correlations have been calculated. A: perfectly positive correlation ($C = 1$). B: perfectly negative correlation ($C = -1$). C: no correlation ($C = -0.03$). D: non-linear coupling. Linear correlation inexistent ($C = 0$). E: high but not perfect correlation ($C = 0.76$).

3.1.3 Empirical Orthogonal Functions

The goal of EOF analysis is to reduce a data set containing a large number of variables to a data set containing many fewer new variables, but nevertheless representing a large fraction of the variability contained in the original data. This commonly used statistical technique is also called the Principal Component Analysis (PCA). However, in this work, this technique will sometimes be referred to as the EOF (Empirical Orthogonal Functions) analysis and EOF as the result.

Suppose that we want to know the daily atmospheric pressure at every place in Europe. The original data is represented in a matrix with all the pressure data for Europe at the different places and at different times. However, there is not enough place to store the daily data at every place in Europe. Therefore, we want to reduce our dataset to a smaller dataset without losing a too large part of the pressure variability. Suppose that we put in our new dataset the pressure at Brussels at every timestep. Then, putting in the pressure at Louvain-la-Neuve would not increase that much our information on the atmospheric pressure at every place in Europe as the pressure at Louvain-la-Neuve is closely linked to the one at Brussels. So, we can imagine that knowing the atmospheric pressure at ten well chosen places would enable a relative complete view of the atmospheric pressure at every place in Europe. As a result, we have reduced our original dataset (atmospheric pressure for each space-time point in Europe) to a more restrained set (atmospheric pressure for ten stations only for all time-points) containing almost the same amount of information. How can we determine what would be the most efficient dataset to restrain our original dataset to? Suppose we want to use a restrained set of only one data point at each time interval. We can choose the atmospheric pressure in Brussels. But maybe the atmospheric pressure in Athens tells us more about the pressure in Europe than the one in Brussels? Or maybe the difference between the two pressures contains yet more information? One can imagine that there is a unique linear combination of pressure data out of the initial dataset which contains the most information that could possibly be stored in one data point. This unique combination is the first EOF. We can continue this in order to get the 10 linear combinations. These 10 EOFs are the best possibility we have to reduce our dataset to only 10 datapoints (per timestep).

A mathematical algorithm has been proposed to reduce the datasets which we will discuss shortly. Suppose that the data that have to be analysed can be put into a matrix $\mathbf{Y}$ ($N \times K$) consisting of N time values of each time K locations. Then $[\mathbf{Y}]$ can be written as follows:

$$[\mathbf{Y}] = \begin{bmatrix} y_{1,1} & y_{1,2} & \cdots & y_{1,K} \\ y_{2,1} & y_{2,2} & \cdots & y_{2,K} \\ \vdots & \vdots & \ddots & \vdots \\ y_{N,1} & y_{N,2} & \cdots & y_{N,K} \end{bmatrix} \quad (3.13)$$

with $y_{l,m}$ the original data at place m on time l . First of all, each data point is replaced by its anomaly, i.e.:

$$y'_{l,m} = y_{l,m} - \frac{1}{N} \sum_{i=1}^N y_{i,m} \quad (3.14)$$

Let us define $[\mathbf{Y}']$ as the matrix with these anomalies. In other words, if $[\mathbf{1}]$ is a $(N \times N)$ matrix whose elements are all equal to one then:

$$[\mathbf{Y}'] = [\mathbf{Y}] - \frac{1}{N} [\mathbf{1}][\mathbf{Y}] \quad (3.15)$$

Now, we can calculate the so-called variance-covariance matrix $[\mathbf{S}]$ which is a symmetric $(K \times K)$ matrix:

$$[\mathbf{S}] = \frac{1}{N-1} [\mathbf{Y}']^T [\mathbf{Y}'] \quad (3.16)$$

The diagonal elements of $[\mathbf{S}]$ are the variances of the variables, while the non-diagonal elements are the covariances of the variables. These covariances are linked to the numerator of the correlation equation (3.11). The only difference is the factor $\frac{1}{N-1}$. Now, we calculate the eigenvectors of $[\mathbf{S}]$. In general the eigenvectors are constrained so that $||\vec{e}|| = 1$. In other words, their length is taken as 1. As $[\mathbf{S}]$ is a real, symmetric matrix, the resulting eigenvectors $\vec{e}$ are orthonormal and the corresponding eigenvalues are real. Then the EOF patterns are these eigenvectors ($\vec{e}_1, \vec{e}_2, \dots, \vec{e}_K$). We can write these eigenvectors into a $(K \times K)$ -matrix, with in each column an eigenvector. Each row represents a place.

$$[\mathbf{E}] = \begin{bmatrix} e_{1,1} & e_{1,2} & \cdots & e_{1,K} \\ e_{2,1} & e_{2,2} & \cdots & e_{2,K} \\ \vdots & \vdots & \ddots & \vdots \\ e_{K,1} & e_{K,2} & \cdots & e_{K,K} \end{bmatrix} \quad (3.17)$$

In other words, the element $e_{l,m}$ represents the component of the m -th eigenvector at place l . To each of these eigenvectors corresponds one eigenvalue. These eigenvalues can be written in a $(1 \times K)$ -matrix:

$$[\lambda_1, \lambda_2, \dots, \lambda_K] \text{ with } |\lambda_1| \geq |\lambda_2| \geq \dots \geq |\lambda_K| \quad (3.18)$$

The first EOF, i.e. the eigenvector with the largest eigenvalue (in absolute value), points to the direction in which the data exhibit the most variability. In other words, the eigenvectors define a new coordinate system which is oriented such that each new axis is aligned along the direction of the maximum joint variability of the data and is orthogonal to all other axes. The indices of the different EOFs can be found by calculating

$$\mathbf{Z} = \mathbf{Y}'(\mathbf{E}^T)^{-1}. \quad (3.19)$$

This results into a $(N \times K)$ -matrix, giving a index for time t on the t -th row and for EOF i on the i -th column. So, we can represent our complete matrix as a linear combination of the K eigenvectors using the indices found in matrix $\mathbf{Z}$. However, if we want to reduce the dataset, we can truncate this series after some eigenvectors. As a result, if we take now the L first eigenvectors, we have a new data set containing fewer variables (the L first EOFs and their corresponding indices), but nevertheless representing a large fraction of the variability contained in the original data. The original values are then approximated with the following equation

$$\hat{\mathbf{Y}}' = \mathbf{Z}_t \mathbf{E}_t^T \quad (3.20)$$

in which the matrix $\hat{\mathbf{Y}}'$ forms a $(N \times K)$ -matrix with the approximated values for $\mathbf{Y}'$, $\mathbf{Z}_t$ forms the matrix of the indices $(N \times L)$, truncated to have only the first L columns of $\mathbf{Z}$. Finally, $\mathbf{E}_t$ is the matrix $\mathbf{E}$ which is now truncated to have only L columns (a $(K \times L)$ -matrix).

Although the EOF procedure was originally used to reduce datasets, it has been used more often in geophysics to determine spacial patterns in data. If a certain eigenvector contains a large part of the variability and the values are covarying following this EOF, there could be a physical explanation that explains this behaviour. Finally, if, thanks to the understanding of this physical process, we can estimate future values for the EOF, we know a lot about the future variability of the dataset, by predicting only one value. An important disadvantage of this technique is its inadaptability to recognise non-static patterns.

3.2 SAM

3.2.1 SAM as the principal EOF of the atmosphere in the non-tropical Southern Hemisphere

Following Thompson and Wallace (2000), the Southern Annular Mode (SAM, also known as Antarctic Oscillation (AAO), as the High Latitude Mode (HLM) and as the Southern Hemisphere Annular Mode (SHAM)) is defined as follows:

Definition SAM is the leading principal component (PC, see subsection 3.1.3) of the 850-hPa geopotential height anomalies south of 20°S.

Taking the principal component of the sea level pressure instead of the 850-hPa geopotential height anomaly is very similar. Their corresponding time series¹ display a correlation of 0.99 (Thompson and Wallace, 2000). A positive SAM index is associated with lower sea level pressures at high latitudes and higher sea level pressure at low latitudes. Therefore, Gong and Wang (1999) proposed another definition which gives very similar results.

Definition The SAM index is defined as the normalised difference of the zonal mean sea level pressure between 40°S and 65°S.

Both definitions will be used in this work as the results do not differ much depending on which one has been used. However, it is of course necessary to say exactly what definition is used for what part. If no specification has been given, the results are valid for both definitions.

The leading EOFs for other atmospheric fields have also been calculated, especially for the geopotential heights of different pressure levels. However, the SAM pattern is not always clearly separated from the other modes of variability, depending on methodology and studied domain. For instance, Kidson (1999) calculated the EOFs of the 300-hPa streamfunction on different timescales in the complete Southern Hemisphere. On the interannual timescale, the leading EOF (26 %) expresses fluctuations in the strength of the subtropical jet, which does not interest us as this jet can only be found far away from the polar regions. The following two modes show a combination of the ENSO (see section 3.3) and SAM patterns. On the interdecadal timescale, the first EOF is part of the ENSO

¹The time series can be found at <http://www.jisao.washington.edu/ao/>.

variability, the second is due to variations before the 1970s (for which the data accuracy can be questioned). The third intradecadal EOF has a SAM pattern. There are several possible explanations why the first EOF is not the SAM on some timescales in this study. First of all, in the definition, the tropical regions are excluded, leading to a decrease in strength of the ENSO signal and of the signal strength of the changes in the subtropical jet. In the analysis of Kidson (1999), the ENSO signal and changes in the subtropical jet are found to be very important when looking at the complete Southern Hemisphere. However, the pressure pattern of the ENSO is not that important close to the regions with sea ice (see subsection 3.3.2). Secondly, the EOFs in Kidson (1999) have been done on the 300-hPa streamfunction, where the definition states the 850-hPa geopotential height.

The EOFs of the complete Southern Hemisphere 500-hPa geopotential heights have been done by Mo (2000) in both the NCEP/NCAR and the ECMWF reanalyses for different seasons. The first EOF depicts the SAM pattern, while the second EOF has very high correlations with the SOI-index². This corresponds with the results discussed above. Another possible method to find out the connections of the SAM to the upper atmosphere over the non-tropical Southern Hemisphere has been done by Thompson and Wallace (2000). They took the index following their definition of the SAM (leading EOF of the 850hPa-height anomaly south of 20°S) and regressed the atmospheric pressure data onto it. They have shown that the SAM pattern extends high up into the atmosphere, at least to the 50-hPa pressure level. As a result, Kidson (1999); Mo (2000); Thompson and Wallace (2000); Sen Gupta and England (2006) all come to the same conclusion that SAM is the most important pattern of variability in the non-tropical Southern Hemisphere on the interannual timescale or longer throughout the complete troposphere.

Why is the annular mode the principal component of the extratropical atmospheric variability? Is this just coincidence, or is it linked to the inherent characteristics of the annular modes? This has been discussed by Wittman *et al.* (2005). Suppose a world with only a zonal jet, from which the variability depends on three parameters: the mean latitudinal position, the strength and the width. Now, a stochastic model can be put in place to change these parameters, in which their variation are

²An index for ENSO (see section 3.3).

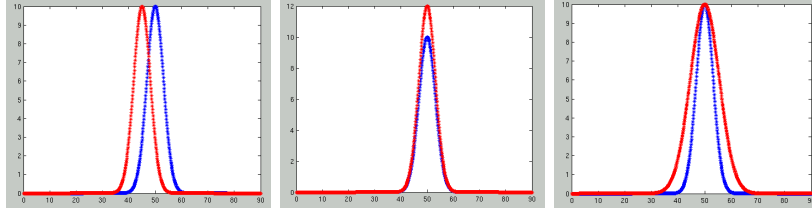


Figure 3.3: Wobbling, pulsing and bulging. The x-axis denotes the latitude while the y-axis shows the strength of the jet in arbitrary units. The blue line denotes the idealised initial situation with a jet centered around 50°S . The red line denotes respectively what happens in the case of wobbling, pulsing and bulging. The situations are all longitudinal symmetric. This figure is based on figure 1a-c of Wittman *et al.* (2005).

described by a combination of pure noise and relaxation to a mean value. Variations of the mean position are called wobbling, variations of the strength are called pulsing and variations of the width are called bulging (Figure 3.3). In their model, the first EOFs of these three processes are then calculated. The wobbling process creates a first EOF which is dipolar and antisymmetric with respect to the jet axis, with higher sea level pressure north of the mean position of the jet and lower pressure south of it. This EOF corresponds to the SAM pattern as discussed above. Following EOFs for the wobbling are alternately symmetric and antisymmetric about the jet centre and are each time higher order standing waves. The pulsing process creates only one EOF, a symmetric standing wave centered on the jet centre. The EOFs due to the bulging are all symmetric and more complex than a simple dipole. As a result, we can conclude that only wobbling is able to create EOFs that are dipolar and antisymmetric about the mean jet position, i.e. EOFs that are similar to SAM. Indeed, the longitudinal mean pressure pattern is bipolar and antisymmetric, as the pressure is higher than usual northward of the jet and lower than usual southward of it. As a result, the SAM is likely due to variations in the mean position of the midlatitude jet.

In a second part of their paper, Wittman *et al.* (2005) have used the NCEP/NCAR reanalysis to obtain realistic model coefficients. Not surprisingly, the first EOF of their model becomes a dipole, antisymmetric about the mean position of the jet. This EOF corresponds to the first

EOF due to wobbling and to the SAM. As a result, even a low resolution model should be able to reproduce the main atmospheric characteristic of the SAM, i.e. the low pressure anomaly centered on the South Pole and the high pressure anomaly at mid latitudes. However, it remains unclear why wobbling is more important than bulging and pulsing in both the real world and in models.

The sea level pressure pattern of the SAM can be seen in Figure 4.1. It consists of a more or less annular pattern with a large low pressure anomaly centered on the South Pole and a ring of high pressure anomalies in mid-latitudes (see also Figure 3.6). This leads to an important zonal wind anomaly in a broad band around 55°S with more important westerlies in the case of a high SAM index. As seen above, this pattern is extended into the high atmosphere. Furthermore, due to the southward shift of the storm track, a high SAM index is associated with anomalously dry conditions over southern South America, New Zealand and Tasmania (all around 40°S) and wet conditions over much of Australia and South Africa (between 20°S and 30°S) (Gillett *et al.*, 2006). This is in agreement with the findings of Sen Gupta and England (2006) who show a decrease of the rain in a band around 45°S, while showing wetter conditions around 30°S and around 60°S in high SAM-index years. This happens in both the NCAR-model and the ECMWF-reanalysis. Over the Antarctic continent and the southern part of the Southern Ocean, the variability in the precipitation due to the SAM is small. Furthermore, it is difficult to analyse this because the data are sparse and not always reliable as well as the effects of the other modes of variability, especially ENSO, are not easy to separate (Genthon *et al.*, 2003).

The increased westerlies above the Southern Ocean also increases the insulation of the Antarctic Continent (Thompson and Solomon, 2002). As a result, there is less heat exchange between the tropics and the poles, leading to a cooling of the Antarctic Continent and the surrounding seas. In addition, the Antarctic Peninsula warms due to a western wind anomaly bringing maritime air onto the Peninsula. Indeed, ocean surrounding the Peninsula is in general warmer than the Peninsula itself. As a result, increased westerlies induce an enhanced heat transport onto the Peninsula.

For most of the atmospheric characteristics of the SAM, its annular form (Figures 4.1 and 3.6) is the most important. However, the pattern is not completely annular, having a low pressure anomaly in the Amund-

sen Sea in high index years. As a result, in years with a high SAM-index, southerly wind anomalies are induced in the Ross Sea, increasing the cold character of this region. On the other hand, the Antarctic Peninsula is warming due to an increased northern component of the wind. This effect is reinforced by an increased transport of maritime air onto the Peninsula during a high SAM-index year, which leads to a temperature of the Antarctic Peninsula closely linked to the SAM index (Marshall *et al.*, 2006). Sen Gupta and England (2006) have shown that both the zonal and meridional advection of heat on the Antarctic Peninsula are important in the NCAR model. In their model, the zonal advection is stronger than the meridional one but this could be due to the underestimation of the Amundsen low pressure anomaly in their model in high SAM-index years.

This non-annular component is probably due to the non-axisymmetric location of the Antarctic continent (Lachlan-Cope *et al.*, 2001). This asymmetry implies a higher variability of the sea level pressure in the Amundsen-Bellingshausen Seas and projected on the Southern Annular Mode this can be the atmospheric source of the low pressure anomaly in the regression onto the SAM index (Thompson, personal communication).

3.2.2 The trend in the SAM index in the last 40 years

Several papers have reported a trend in the SAM-index towards positive values, with corresponding low pressure anomalies above the Antarctic and high pressure anomalies at mid-latitudes. Most of the papers investigating this trend (Kidson, 1999; Gong and Wang, 1999; Mo, 2000; Thompson *et al.*, 2000; Raphael, 2003) have used the NCEP/NCAR-reanalysis (Kalnay *et al.*, 1996) (Figure 3.4). They showed an important increase from 1950 to mid 1960s, a decrease in the beginning of the sixties, a slow increase until the seventies, followed by a sharp increase between the 1970s and the 1980s. Afterwards, the SAM index has stayed positive and more or less stable, except for a brief change of sign at the beginning of the 1990s. There is a small decrease seen during the last years (not shown here).

However, it is known that there exists serious problems with the pressure fields at high southern latitudes in the NCEP/NCAR reanalysis (e.g. Hines *et al.*, 2000; Marshall, 2003). For instance, Hines *et al.* (2000)

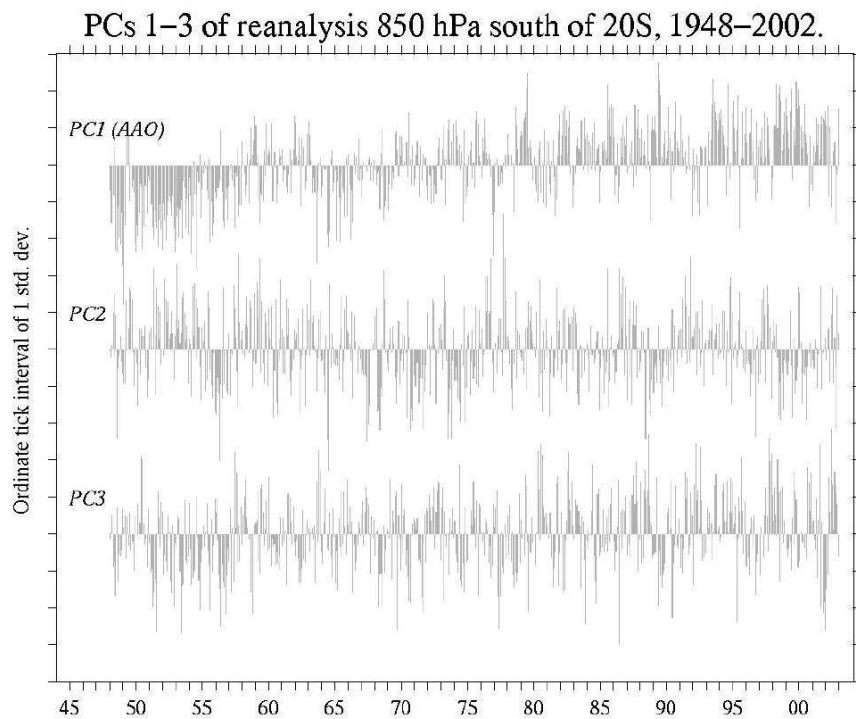


Figure 3.4: The indexes of the 3 most important non-tropical Southern Hemisphere principal components at the 850hPa-level. The uppermost graph corresponds to the SAM-index. Source: <http://www.jisao.washington.edu/aao/pcs.jpg>

showed that some pressure trends in the reanalysis are spurious and occur due to the increased accuracy of the reanalysis during the discussed period in this region. Therefore, Marshall (2003) used observation data and the definition of the SAM-index proposed by Gong and Wang (1999) to assess the real trend in the SAM-index. Using this method, Marshall (2003) has proven that the trend in the SAM-index using NCEP/NCAR-reanalysis is overestimated by a factor 2. This overestimation is more important in winter than in summer, resulting in a maximum trend in winter in the reanalysis while the observations show a maximum trend in summer. In general, the resulting index (Figure 3.5) is qualitatively very similar to the figure derived from the reanalysis. It also shows the increase at the end of the 1950s, the decrease at the beginning of the 1960s and the strong positive SAM index in the 1990s and at the beginning of the 2000s. The sharp increase between the 1970s and the 1980s is mostly absent although a slow increase in the SAM-index in these decades can be found. Although the trend is much smaller in the observations, it is still significant. We will investigate the origin of this trend in the next section.

3.2.3 The influences of climate change and stratospheric ozone loss on the SAM index

In the Third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (Houghton *et al.*, 2001) it is shown that the state-of-the-art models are well capable of reproducing the main characteristics of the surface temperatures during the 20th century. The change in atmospheric circulation, more specifically, in sea level pressure (SLP) is much less known. However, one feature seems to be consistent between the models. Due to the increase in greenhouse gas concentrations, the pressure in both hemispheres (but more clearly in the Southern than in the Northern Hemisphere) is projected to decrease above the poles and to increase at mid-latitudes. For the Southern Hemisphere, this means that the SAM-index increases. As the observed SAM-trend is consistent in sign with these projected changes, the increase in the SAM-index could be explained by the increase in greenhouse gas concentrations (Kushner *et al.*, 2001; Marshall *et al.*, 2004).

Thompson and Solomon (2002); Gillett and Thompson (2003); Shindell and Schmidt (2004) have indicated that the same trend can also be

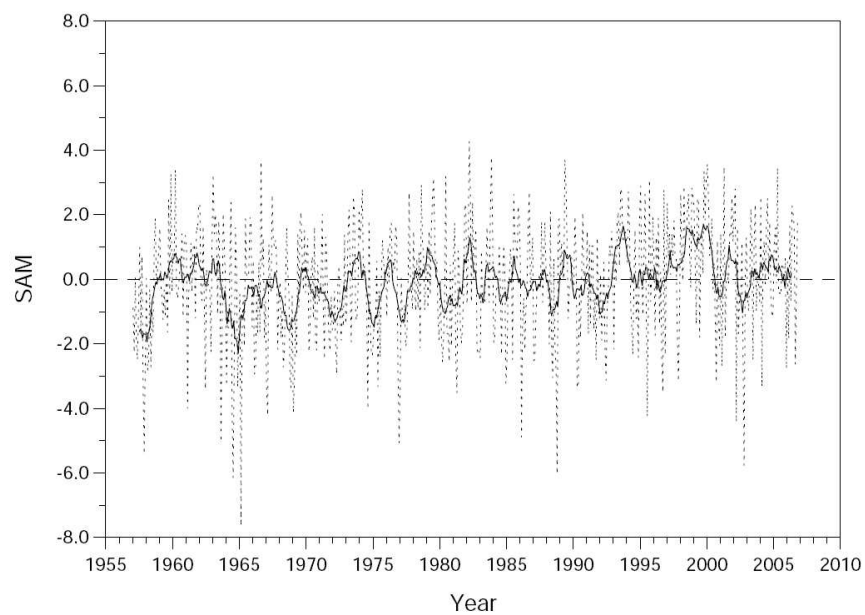


Figure 3.5: The station-based SAM index following Marshall (2003). More information on the method can be found in Marshall (2003). The fine line depicts a monthly value of the SAM-index, while the thick line is a running 12-month mean. Source figure: <http://www.nerc-bas.ac.uk/icd/gjma/sam.html>

explained by the decrease in stratospheric ozone. There are even indications that ozone loss has contributed more than greenhouse gases to the increase in the SAM index at the end of the 20th century. For instance, the trend in the SAM index is most important in summer and autumn. This is consistent with the model results, when they are forced with decreasing stratospheric ozone concentrations (Gillett and Thompson, 2003; Arblaster and Meehl, 2006). As both the amount of greenhouse gases and the ozone loss became more important at the end of the 20th century, it is difficult to distinguish the exact cause of the rise in the SAM-index purely from observations.

Therefore, Arblaster and Meehl (2006) have used the Parallel Climate Model (PCM) to test the response of the atmospheric variability to the forcing of five different factors, i.e., volcanic eruptions, changes in solar activity, greenhouse gas concentration changes, sulfate aerosol concentration changes and stratospheric ozone concentration changes. This was done by forcing the model with each of the different possible factors only. Furthermore, a control run has been done using all these variable forcings together. The authors came to the conclusion that ozone loss has played the largest role in the change in the SAM-index, although greenhouse gas forcing is also important. Furthermore, the effect of the increase in greenhouse gases was found to be towards a positive SAM-index in all seasons, whereas the effect due to an ozone loss was more seasonally dependent. For the future, the greenhouse gas effect is projected to become more important. At the other hand, the ozone loss effect should diminish thanks to regulations protecting the ozone layer but other phenomena could hamper the recovery.

An increase in the SAM-index tends to cool the Antarctic Continent (see above). As a result, the Antarctic Continent is one of the regions in the world where the temperature trend over the last 30 years is not determined as significantly positive. The only exception is the Antarctic Peninsula. Due to the non-annular form of the SAM pattern, a rise in the SAM-index increases the northern component of the wind around the Antarctic Peninsula. As said above, the advection of maritime air over the Peninsula is increased in high SAM-index situations. Combining these factors with the warming due to the increase in greenhouse gas concentrations makes the Antarctic Peninsula one of the regions of the world with very rapid regional warming (Vaughan *et al.*, 2003; Marshall *et al.*, 2006). The last reference even hints that the recent trend in the

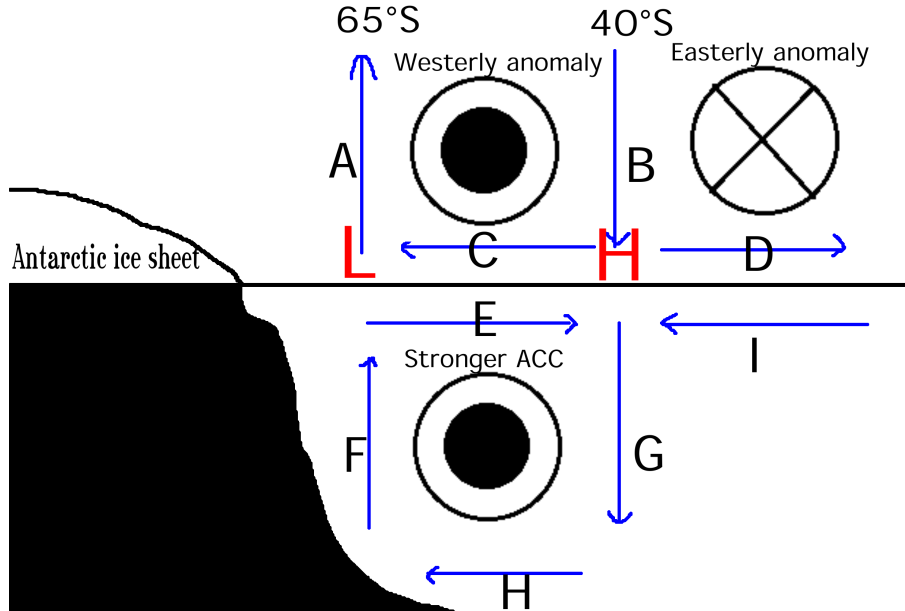


Figure 3.6: Schematic drawing of the changes in the atmosphere and ocean that occur when the SAM index is positive. The drawing has a left-right north-going latitude axis. It is based on Hall and Visbeck (2002, Figure 12), but the part about the sea ice in the figure of Hall and Visbeck (2002) is omitted.

SAM-index is responsible for the collapse of the Larsen B ice shelf³. However, it is possible that other effects, such as a changed mesoscale oceanic circulation, than those just mentioned play also an important role in the explanation for the very large upwards temperature trend around the Antarctic Peninsula.

3.2.4 The oceanic response to SAM

For the oceanic component, the response to SAM is mainly the response to the annular part, whereas the non-annular part has little effect (Hall and Visbeck, 2002; Lefebvre *et al.*, 2004). The increased westerlies in

³An ice shelf previously located (around 65°S, 60°W) at the east coast of the Antarctic Peninsula which collapsed in 2002.

a high SAM-index phase increase the Ekman drift towards the north (Figure 3.6, blue arrow labeled E). By continuity, a downwelling further north (Figure 3.6, blue arrow labeled G) and a (Ekman) pumping further south (Figure 3.6, blue arrow labeled F) are needed. A southward deep current anomaly closes the cycle (Figure 3.6, blue arrow labeled H). Due to the increased pumping close to the Antarctic continent, there is an increase in sea surface temperature and salinity there (Lefebvre *et al.*, 2004; Sen Gupta and England, 2006). Furthermore, cold water is driven northward at the surface south of 40°S, cooling the surface there, while more to the north southward surface currents bring warm water (Figure 3.6, blue arrow labeled I). Combining these factors, an increase in sea surface temperature (SST) is found in a small band close to the continent, a cooling further north and a warming north of about 50°S (Lefebvre *et al.*, 2004) (Figure 4.4). These last two features have also been found in the NCAR model and the NOAA SST reanalysis (Sen Gupta and England, 2006), although the border between them is more to the north in these datasets (around 45°S). In the reanalysis dataset, the SST response to the SAM is also much less annular with a large negative SST anomaly in the southern Pacific Ocean (90°W-150°W 35°S-70°S) and a positive SST anomaly in the New Zealand-Tasmania region. The highest correlation between the SAM-index and the SST is when the SST lags 1 month behind the index. The absence of the warm water band close to the continent in the NCAR-model has been attributed to the small vertical heat flux (Sen Gupta and England, 2006) in the ocean compared to the ORCALIM-model of Lefebvre *et al.* (2004). In addition to the reaction of the SST to the SAM index, the SAM index reacts on the SST. This has been shown by Zhou and Yu (2004) who did a hindcast of the SAM-index using a atmospheric model forced by observed SSTs. Although the correlation between the observed historical SAM-index and the modelled SAM-index is significant, the correlation is low, showing that a large part of the variability is not linked to the SST.

In addition, the western winds increase the mixed layer depth between 40°S and 65°S. This effect amounts to around 5m in summer and up to a tenfold of that in winter (Lefebvre *et al.*, 2004). This is probably due to increased mixing of the upper water column induced by the increased wind velocity. In the NCAR-model (Sen Gupta and England, 2006), the increase is between 45°S and 60°S and there is a decrease further north. Furthermore, there are also clues that a high-SAM index

increases the strength of the ACC (Gnanadesikan and Hallberg, 2000; Hall and Visbeck, 2002; Hughes *et al.*, 2003; Sen Gupta and England, 2006; Meredith *et al.*, 2004; Meredith and Hogg, 2006) but these changes are relatively small. The sea level elevation is also coupled to the SAM index (Aoki, 2002). When the SAM-index is high, the sea level is low around the Antarctic continent. This is due to the horizontal divergence there during high SAM-index years.

There are some indications (Sen Gupta and England, 2006) that SAM also changes the pathways of the intermediate waters. During high SAM-index years, the northwestward flow to the south and the west of Australia and South Africa is enhanced, while the equivalent flow in the Pacific is weakened.

3.2.5 The response of sea ice to SAM

In a model study, Hall and Visbeck (2002) proposed a significant rise in the sea ice extent due to an increase in the SAM-index. However, their model sea ice component is very crude and the sea ice moves freely with the surface ocean currents provided the ice thickness is less than $4m$ ⁴. However, Hibler (1979) has described how the sea ice reacts to both the wind stress and the ocean forcing as the ice does not move freely with the surface layer of the ocean but is also driven by the wind. Kwok and Comiso (2002b); Liu *et al.* (2004) did not find the sea ice response as described by Hall and Visbeck (2002) in the observations. They both found that a high SAM-index corresponds to more ice in the Amundsen and Ross Seas and less ice in the Bellingshausen Sea and in the Weddell sector. A smaller effect of SAM can be found in the Pacific sector and in the Indian sector, although these changes give slightly more ice with a high SAM-index.

Lefebvre *et al.* (2004) and Lefebvre and Goosse (2005) have discussed the effect of SAM on the winter sea ice using a more performant sea ice model than the previous studies. They came to the following conclusions about the effect of SAM on the sea ice, which are in line with the observations (Kwok and Comiso, 2002b; Liu *et al.*, 2004). First of all, the low pressure anomaly in the Amundsen Sea during the high SAM-index years induces cold winds in the Ross Sea and the Amundsen Sea, and warm

⁴Ice thickness is less than $4m$ almost anytime and almost everywhere in the Southern Ocean.

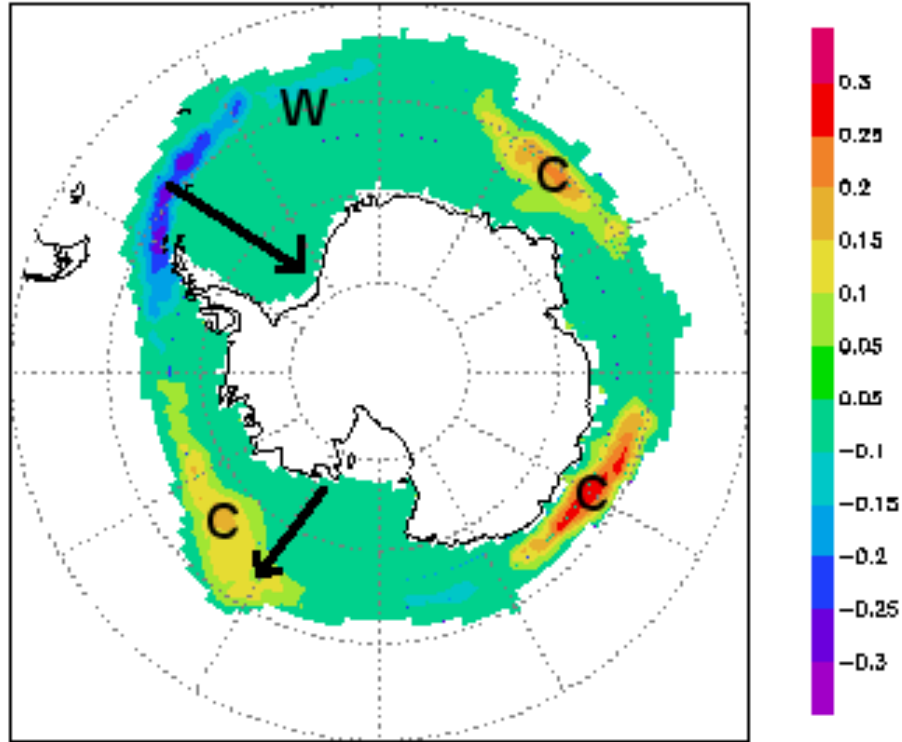


Figure 3.7: Regression between the July-August-September mean SAM index and the July-August-September mean ice concentration for the period 1980-1999 in the HadISST1 observations (Rayner *et al.*, 2003) (colour). On top of that, a schematic drawing is superimposed to highlight the origin of the regressions. In letters, the thermal part is shown, with C and W denoting colder air and warmer air during high SAM-index winters, respectively. The arrows show the direction of the mean sea ice velocity anomaly during high SAM-index winters. Letters and arrows are only shown at places where their respective effects create important anomalies in the sea ice concentration. This figure is based on Figures 4.7, 5.8 and 5.12.

winds around the Peninsula. As a result, the ice extent in the Ross Sea increases whereas the ice extent around the Antarctic Peninsula decreases (Figures 3.7, 4.7 and 4.8). Furthermore, these wind anomalies account for northward ice velocity anomalies in the Ross sector and southward anomalies around the Antarctic Peninsula, respectively (Figure 3.7, arrows). Both wind-induced effects reinforce each other, giving more ice in both the Ross and Amundsen sectors and less sea ice around the Antarctic Peninsula. Further important factors for the sea ice are its northward drift (due to Ekman Drift) in the Pacific sector and the increased heat flux due to the Ekman pumping close to the continent. The combined effect of all these factors is a positive sea ice concentration anomaly in the Ross, Amundsen and Pacific sector and a negative anomaly in the Weddell and the eastern Bellingshausen sectors (Figures 3.7, 4.7 and 4.8, Table 4.1). In the Indian sector, the situation is less clear as the sea ice response to SAM is not well represented in the ORCALIM model. These results are confirmed by the NCAR model (Sen Gupta and England, 2006) that shows also that the isolation effect is most important in summer, increasing the sea ice extent in high SAM-index years. Furthermore, these results are also seen in the advance and retreat days of the ice edge as shown by Stammerjohn *et al.* (2007).

Lefebvre *et al.* (2004) discusses extensively the ocean and sea ice responses to SAM in the Southern Ocean. Lefebvre and Goosse (2005) tries to distinguish the wind and temperature driven effects. These results can be found respectively in Chapters 4 and 5.

A link has been proposed between the SAM-index, the SOI-index (see section 3.3) and the Weddell Polynya (Gordon *et al.*, 2006). Following this paper, a Weddell Polynya and similar features can only occur after a long period with negative SAM indexes. They decrease the precipitation and would so destabilise the water column around Maud Rise. Further work is needed to investigate this link.

3.2.6 The projected response of the sea ice to the further rise of SAM

As the SAM-index is expected to rise into the future (see section 3.2.3), we could be tempted to extrapolate the sea ice response to the SAM (see section 3.2.5) into the future. In order to find out whether this strategy is acceptable, we have used a set of simulations performed with

atmosphere-ocean general circulation models for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR4). Averaged over the models used, it is shown that the late 20th century SAM-pattern and the pressure trend over the same period are well reproduced by the models (see Figure 7.6). Furthermore, the late 20th century ice concentration trends are too negative in the models although their distribution is well reproduced. Indeed, the multimodel mean shows a large region with positive ice concentration trends in the Ross sector and a stronger negative signal around the Antarctic Peninsula for the late 20th century.

Furthermore, the sea ice response to the SAM is well reproduced in the multimodel mean (see Figure 7.8). Indeed, the seesaw between positive regression of sea ice concentration to SAM in the Ross and Amundsen Seas and the negative regressions around the Antarctic Peninsula are clearly visible. In addition, the generally positive regressions in both the Indian and the Pacific sectors are also well reproduced.

Despite the projected rise in the SAM pattern (see Figure 7.7), the projected model pressure trends are not exactly similar to the SAM pattern (see Figure 7.6). The major difference is the absence of the low pressure area in the Amundsen Sea in the projected pressure changes. As seen above, this low pressure area plays a important role in the response of the sea ice to SAM. Therefore, the projected response of the sea ice in the 21st century does not show a clear SAM-like response and is much more zonal than the trends recently seen.

Other possible factors for possible differences in regional sea ice concentration trends in the 21st century Southern Ocean have been investigated. These encompassed both atmospheric and oceanic responses. The results of this work can be found in Chapter 7.

3.2.7 The role of SAM on the Southern Ocean biology and the Southern Ocean air-sea- CO_2 fluxes

Lovenduski and Gruber (2005) have investigated the role of the SAM onto the satellite observed chlorophyll concentrations. According to their results, the chlorophyll concentration increases close to the Antarctic continent during high SAM-index years. On the contrary, the biological activity decreases further north in these years. They hypothesized that the increase is closely linked to the increase in iron availability due to

Ekman pumping. As a result, the phytoplankton is more abundant, leading to a higher chlorophyll concentration. On the other hand, further north, the increase in the mixed layer depth mixes the phytoplankton too deep into zones where no light is available. As a result, the phytoplankton dies and the chlorophyll concentrations decrease.

Using hindcast simulations of coupled ocean-sea ice models including biological components, Verdy *et al.* (2007); Lenton and Matear (2007); Lovenduski *et al.* (2007) have shown the effect of the SAM on the air-sea- CO_2 fluxes in the Southern Ocean. Three effects seem to be important:

- In a situation with a positive SAM-index, the increased westerlies in the Southern Ocean create a positive outgassing⁵ anomaly.
- Due to the effect of SAM on the ocean circulation (see subsection 3.2.4), water with a high DIC (Dissolved Inorganic Carbon) is pumped up close to the Antarctic Continent and then exported at the surface to the north. There, it is warmed, decreasing the ability to hold the CO_2 , which leads to a positive outgassing anomaly.
- The rise in the SAM-index decreases the ocean temperature close to the surface (see also Figure 4.4), inducing a negative outgassing anomaly.

This last effect is shown to be significantly smaller than the sum of the previous two. So, an increase in the SAM-index leads to decreased air to ocean carbon flux in the Southern Ocean with a rate of $0.1PgC/yr$ per standard deviation of the SAM-index.

Using the results of the previous paragraph, we could expect a smaller role of the Southern Ocean oceanic CO_2 sink due to the recent positive trend in the SAM-index (see subsection 3.2.2). However, during the same period, the atmospheric CO_2 concentrations have risen sharply, which has also an influence on the CO_2 fluxes. Combining both factors, the trend in the fluxes is estimated by Lovenduski *et al.* (2007) as $0.005PgC/yr/yr$ towards a stronger sink over the period 1958-2004. This trend would have been still higher if the SAM-index had not shown such a dramatic increase. The part of the trend that is purely due to the rise in the SAM-index has been estimated as $0.006PgC/yr/yr$ over the period 1958-2004 (towards a stronger outgassing) in the model of Lovenduski

⁵This is a CO_2 flux from the ocean to the atmosphere.

et al. (2007) and as $0.008 PgC/yr/yr$ (towards a stronger outgassing) over the period 1981-2004 in the observations (Le Quéré *et al.*, 2007).

3.3 ENSO

3.3.1 The tropical ENSO

The text in this section is based on Glantz *et al.* (1991) and on a website summarising their results⁶. The goal of this section is to explain the basic principles behind ENSO (El Niño-Southern Oscillation).

Each year, in the first few months of the year, the waters at the coast of Peru warm up. However, in some years, this warming begins earlier, is stronger and lasts longer. This is combined with much rain in arid regions, making a garden out of (semi-)deserts. On the other hand, the fishes stay away from the nets of the fishermen as the warm current is much less nutrient-rich. The coastal residents called this phenomenon El Niño, the Christ Child. However, El Niño is not limited to the coastal South-American region, it extends far outside of it and is part of ENSO, the El Niño-Southern Oscillation.

In the 1930s, Walker discovered the inverse relation between surface pressures at Darwin in Australia and Tahiti in the tropical Southern Pacific Ocean. When the pressure rises in Darwin, the pressure decreases in Tahiti and vice versa. He called the corresponding pattern the Southern Oscillation. At that point, it was not clear that this pattern was part of the leading pattern of atmospheric variability in the world (Turner, 2004), but the Southern Oscillation has become an important name for the ENSO-phenomenon (and part of the ENSO acronym). A further extension to this picture was given by Bjerknes (1969). He reasoned that, as the trade winds blow from east to west in the tropical Pacific, the loop must be closed in the upper atmosphere by westerly winds. Furthermore, rising air must be apparent in the West Pacific and descending air in the East Pacific. He called this circulation the Walker circulation. The persistent easterly winds push the water towards the West Pacific, enabling a sea level elevation difference between the West and the East Pacific of about 60cm. Furthermore, while crossing the Pacific, the surface waters gain heat, forming at their destination the Pacific warm pool. As a result,

⁶http://www.ucar.edu/communications/factsheets/el_nino/

the warm waters pool in the Western Pacific, pushing down the colder waters underneath. As a result, the thermocline⁷ descends. The inverse effect happens at the Pacific coasts of South America enabling the ascent of the thermocline close to the surface. The pooling of the warm water in the western Pacific and the ascending branch of the Walker circulation increase the precipitation there, due to frequent thunderstorms. So, the atmospheric component (trade winds) enhance the warm pool which in turn enhances the atmospheric component (thunderstorms). A coupled behaviour is thus the result.

The ENSO variability is due to the variability around the climate pattern discussed above. In the El Niño phase of the cycle, the trade winds weaken, the Central and Eastern Pacific Ocean warm, the precipitation shifts eastward, the western part of the thermocline ascends while the eastern part of the thermocline descends. Finally, the pressure in the Western part of the Pacific Ocean increases, while it decreases in the central part of it. As a result, the difference between the normalised pressure at Darwin and Tahiti, called the Southern Oscillation Index (SOI) drops. In the La Niña phase of the cycle, the trade winds strengthen, the precipitation is even more to the west than normal and the SOI rises.

3.3.2 ENSO and the Southern Ocean

Next to the tropical changes, ENSO has influences almost everywhere in the world. It has an influence too in the region that is important for us, the Southern Ocean. The mechanisms linking the tropical part of ENSO to the effects seen in the Southern Ocean are discussed in Liu *et al.* (2002b) and Yuan (2004).

There is a distinct influence of the ENSO phenomenon on the Bellinghausen Sea and the Antarctic Peninsula. Thanks to the many stations on the last one, the ENSO-due variability is well represented in this region. As the SOI is high, which is equivalent with a low MEI-index⁸, i.e. in a La Niña situation, there is a large low pressure anomaly west of the Antarctic Peninsula, especially in winter (Kwok and Comiso, 2002a;

⁷The place where the vertical temperature gradient in the ocean reaches its maximum.

⁸The MEI-index is the Multivariate ENSO Index and is positive in a El-Niño phase and thus anticorrelates with the SOI-index. It can be found at <http://www.cdc.noaa.gov/people/klaus.wolter/MEI/mei.html>.

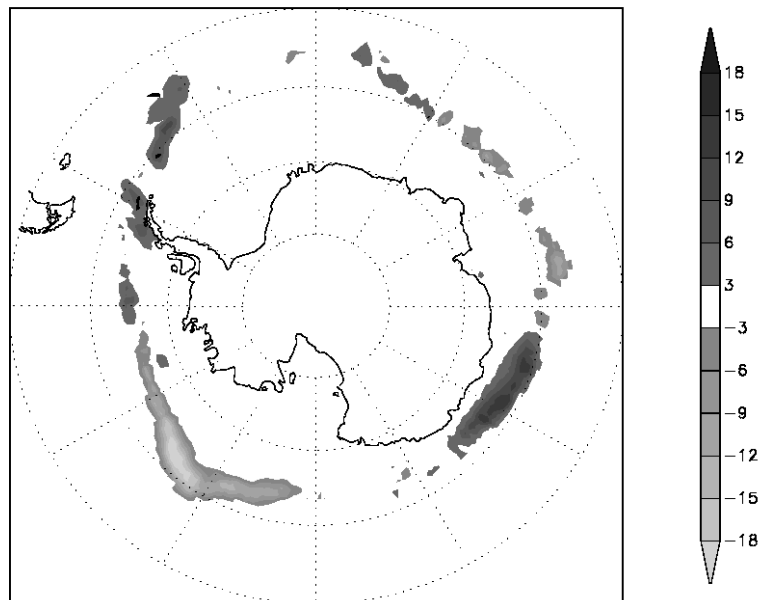


Figure 3.8: Regression between the annual mean MEI-index and the annual mean ice concentration for the period 1979-2004 in the HadISST observations (Rayner *et al.*, 2003).

Turner, 2004; Yuan, 2004). On the contrary, in SOI-low cases, low pressure areas are relatively absent in this region. The net result is a warmer Antarctic Peninsula in SOI-high cases, although the correlation between the SOI and the temperature at Faraday (west coast of the Antarctic Peninsula) is not significant. This is due to the large amount of noise in the temperatures at the Antarctic Peninsula created by various sources of natural variability (like SAM). However, in the Eastern Bellingshausen Sea, the signal/noise ratio of the SOI-influence of the temperature is larger, showing a significant warming at high SOI-indexes (Kwok and Comiso, 2002a). Although this effect is quite robust, one of the most important low-SOI-periods (1982-1983) did not show those effects (Turner, 2004). The SSTs change accordingly as could be expected from the pressure anomalies with higher eastern Bellingshausen Sea and lower Ross Sea SSTs in high SOI-years. The Antarctic continent close to the Ross Sea also cools (Bertler *et al.*, 2004). Finally, the sea ice in Ross Sea increases in high SOI-years, while the sea ice off the coast of the Antarctic Peninsula decreases (Figure 3.8) (Yuan and Martinson, 2000). It is important to note that the surface pressure pattern in high SOI-years is very similar to the one in high SAM-years, despite no correlation between the respective indices. As a result, part of the warming of the Antarctic Peninsula is maybe due to the small downward trend in the SOI-index (Kwok and Comiso, 2002a; Stammerjohn *et al.*, 2003).

Finally, Genton and Cosme (2003) have shown that an important part of the precipitation variability is probably ENSO-driven.

3.4 Other atmospheric modes of variability

3.4.1 ACW

In 1996, White and Peterson (1996) discussed the existence of an Antarctic Circumpolar Wave (ACW). This wave was found in the sea level pressure, in the meridional wind stresses, in the sea surface temperatures and in the sea ice concentration. On these data, White and Peterson applied a 3-7yr bandpass filter, which filtered out most of the interannual variability. The remaining variability appeared in sort of a wave, with rather high amplitudes. This wave has a period of 4 to 5 years and takes 8 to 10 years to encircle the Antarctic continent. As a result, two consequent waves form a complete circle. They further showed that warm SSTs

trail high sea level pressures by 1 year, and are in phase with poleward anomalies in meridional wind stress and sea ice edge. The variability explained by the Antarctic Circumpolar Wave seemed to be locally large. However, due to the wave nature, the influence of the ACW on the sea ice extent integrated over the whole Southern Ocean is probably small, as the positive and the negative patches compensate each other.

The signal of the ACW is only clear in the period 1985-1994 (Connolley, 2003; Carleton, 2003), which is just one cycle of the wave. There are indications that the ACW is only present when the Southern Ocean is in a certain mode. The other mode consists merely of a non-propagating static wave 3 structure (Connolley, 2003, see also further down). The absence of the wave before 1985 and after 1994 has caused to decrease the interest in the ACW.

3.4.2 SAO

The seasonal temperature variation of the Southern Hemisphere south of 40°S is mainly determined by the following factors (Van Loon, 1967):

- The incoming solar radiation around the winter solsticium is very small over the Antarctic continent, as the Sun is mostly beneath or just above the horizon. The absence of a clear minimum in insolation hampers the creation a real temperature minimum.
- The outward longwave radiation is small but continues to slowly cool the winter night as long as this outgoing heat flux is not compensated by an incoming one.
- The open ocean is the prime heat source for the Antarctic continent in winter. Due to the prevalence of sea ice in the southern winter, this heat source is at its maximum distance of the continent when the sea ice extent is the largest, i.e. in August and September.
- Furthermore, this open ocean itself continues to cool until September. Thus, on the average, warm air advected poleward from a given parallel will leave this region with increasingly lower temperatures during winter.
- Around the summer solstice, the insolation is reasonably high, enabling a temperature maximum at the end of December and the

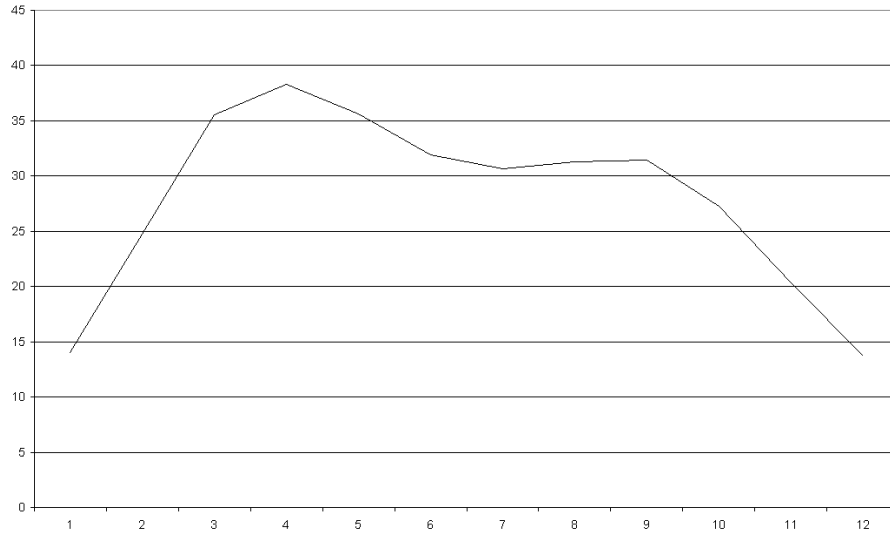


Figure 3.9: Zonal mean annual cycle of difference (in K) of NCEP/NCAR surface temperature between 65°S and 80°S . The average has been made over the period 1979-2004. The months are shown on the x-axis, with 1 being January and 12 being December.

beginning of January for the Antarctic continent (Van Loon, 1967, Fig 13) (Van Loon and Jenne, 1972, Fig 3.32).

- In the Southern Ocean, due to the thermal inertia of the ocean, the maximum temperature is found in February-March, while the minimum is found in August-September.

As a result, the temperature difference between the mid-latitudes and the high latitudes is the most important around the equinoxes (Figure 3.9). As the temperature contrast is higher between the Southern Ocean and the Antarctic Continent around the equinoxes, the cyclonic activity increases over the polar sea during these months (Van Loon, 1967) (Figure 3.10). As a result, the wind speed in the Drake Passage and at Kerguelen increases (Van Loon, 1967). Due to this changes, a second harmonic is present in the pressure data above the Southern Ocean. The phase in which the cyclonic activity increases over the polar sea is called the contracted phase, while the other is the expanded phases. These

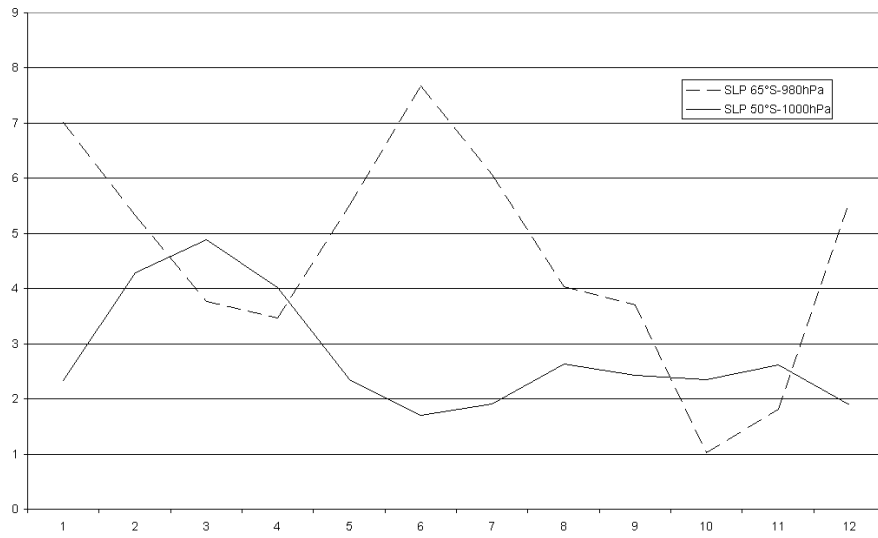


Figure 3.10: Zonal mean annual cycle of NCEP/NCAR sea level pressures (in hPa) at 50°S (solid line) and at 65°S (long dashed line). In order to improve the readability of the graph, a constant value of 1000hPa has been subtracted from the sea level pressures at 50°S, and a value of 980hPa from the sea level pressures at 65°S. The months are shown on the x-axis, with 1 being January and 12 being December.

names point towards the form of the circumpolar through which is more compacted around the equinoxes.

The cycle discussed above is a climatological cycle, which happens each year. However, the strength of the cycle differs from year to year. This is measured by the SAO-index, which is defined as the zonally and meridionally (40°S - 70°S) averaged amplitude of the second harmonic in the annual march of the surface pressure ($H_2(p)$) (van den Broeke, 2000b). If the SAO-index is high, the cycle that has been discussed above is more prominent. In the second expanded phase of the cycle, the sea level pressure is higher on the border of the Antarctic continent and lower around 50°S in high SAO-index years. However, this is not exactly uniform. Three negative anomalies can be found: one just east of South America, one in the centre of the southern Indian Ocean and one southeast of New Zealand. These low pressure anomalies bring warm air to the Weddell and Ross sectors, decreasing the sea ice extent there. These important anomalies persist until September creating a significant negative correlation between the SAO-index and the September sea ice extent (Lefebvre and Goosse, 2007a). However, if the SAO-index is not calculated over the months January-December but over a slightly different period, there is a large difference in the simulated and observed response of the sea ice to the SAO. Furthermore, the correlation between the total sea ice extent and the SAO-index which was significant if the index was calculated over the period January-December becomes insignificant (both for the period November-October as for the period January-December). If we calculate the response of the September sea ice to the SAO-index for the period 1958-1999, the result (Fichefet *et al.*, 2003b) is also different than the one obtained here. More information about the effect of the SAO on the sea ice is given in Lefebvre and Goosse (2007a), see section 6.3. More investigation towards the precise role of the SAO in the ice variability is needed.

3.4.3 ZW1

We have discussed before the low pressure area west of the Antarctic Peninsula as part of the variability of both SAM (section 3.2.1) and ENSO (section 3.3.2). However, it can also be seen as a stand-alone mode which is correlated to the SAM and ENSO (Van Loon and Jenne, 1972; Hobbs and Raphael, 2007). It is then seen as the Zonal Wave 1

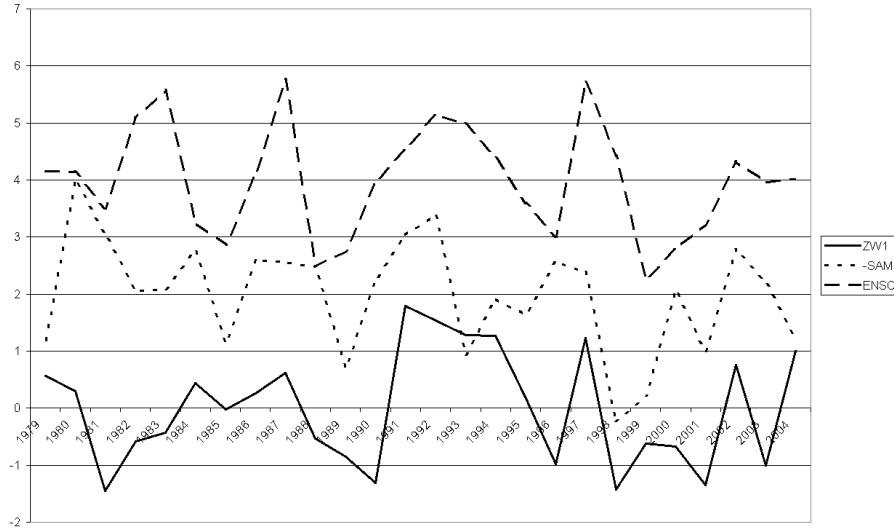


Figure 3.11: Annual mean time series of ZW1 (continuous line), -SAM (dotted line) and ENSO (dashed line). All the time series have been normalised and have a standard deviation of 1. The SAM-index is multiplied by -1 and offset by 2. The original SAM-index has been based on the first definition given in section 3.2.1. The ENSO-index is based on the Multivariate ENSO Index (MEI) and is offset by 4. The ZW1-index has been based on the index for the amplitude of the ZW1 made by Hobbs and Raphael (2007).

(ZW1), being the first harmonic of the zonal geopotential height. Hobbs and Raphael (2007) have made some time series for it. It can be shown that the ZW1 time series are correlated with both the SAM-indices (negatively) and the ENSO-indices (positively) (Figure 3.11). Indeed, the high pressure area in the Amundsen Sea occurring with a negative SAM-index or with a positive ENSO-index⁹ is the same as the high pressure area described by the ZW1-index. The sea ice response to the ZW1 is induced directly by the sea level pressure, i.e. more ice in the Bellingshausen Sea and less ice in the eastern Ross and Amundsen Seas (Figure 3.12). However, as we have discussed this high pressure area as part of SAM and ENSO, we will not discuss it further here.

⁹The Multivariate ENSO Index (MEI) can be found at <http://www.cdc.noaa.gov/people/klaus.wolter/MEI/mei.html>.

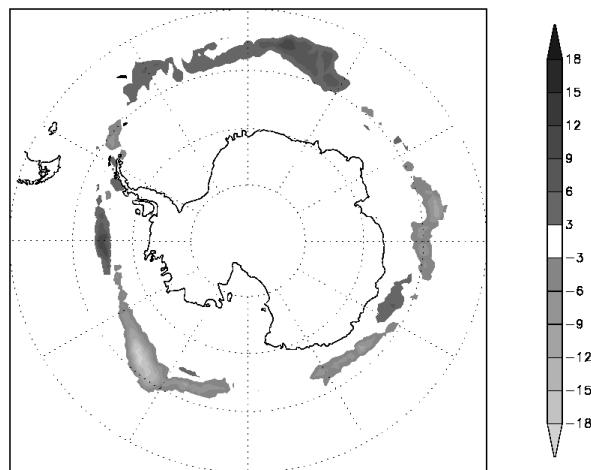


Figure 3.12: The regression of the September mean observed (HADISST) sea ice concentration on the annual mean ZW1-index. The ZW1-index has been based on the index for the amplitude of the ZW1 made by Hobbs and Raphael (2007).

3.4.4 ZW3

Van Loon and Jenne (1972); Mo and White (1985); Kidson (1988) have shown that there is at least one mode that we have not yet discussed here, the zonal wave 3. This zonal wave is quasi-stationary and is a dominant feature on daily, seasonal and interannual timescales (Raphael, 2004). Up to today, little investigation has been done to the origin of this ZW3 and it is only since some years that the ZW3 receives some attention. An index for the ZW3 has been proposed by Raphael (2004) as the normalised average of values at three selected grid points. The locations are chosen at the centre of the ridges in the 500 hPa geopotential height field. Instead of the 500hPa geopotential height, the sea level pressure or the zonal anomaly of the 500 hPa geopotential height can be used. The effect of the ZW3 on the sea ice extent integrated over the Southern Ocean is small, as the positive and negative regressions of the sea ice concentration onto the ZW3-index tend to cancel each other out (Lefebvre and Goosse, 2007a). Indeed, induced by the wave 3 in the atmosphere, the sea ice changes. In general, the changes concur with what could be expected: more ice with southern wind anomalies, less ice with northern wind anomalies. Figure 3.13 shows the response of the September mean sea ice extent to the ZW3.

3.5 Sea ice variability and atmospheric modes

As discussed above, each of the atmospheric modes of variability has an influence on regional sea ice conditions in the Southern Ocean. However, the response of the sea ice to the modes of variability mostly occurs in the form of a mixture of positive and negative patches, which tend to cancel each other out, when integrated over the Southern Ocean. This has been shown and is discussed in the first part of the Lefebvre and Goosse (2007a) paper (see Chapter 6).

One could also look the other way around. Which are the most important patterns of variability (EOF) of the sea ice concentration and what are they due to? This has been done by Yuan and Martinson (2000, 2001); Holland and Raphael (2005a,b). They have shown that the first EOF of the winter (JJAS) sea ice concentration is a dipole pattern, with a positive patch in the Amundsen and Ross Seas and a negative patch extending from the Antarctic Peninsula eastward into the

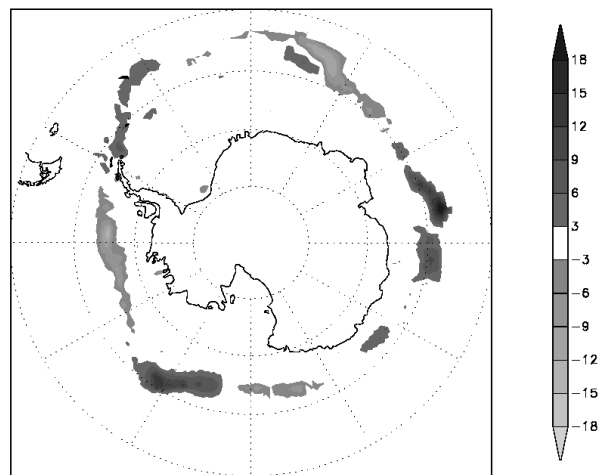


Figure 3.13: The regression of the September mean observed (HADISST) sea ice concentration on the ZW3-index, averaged over the growth season (April-September). The ZW3-index has been calculated as in Raphael (2003).

Weddell Sea. This dipole is called the Antarctic Dipole (ADP). Yuan and Martinson (2000) connected this dipole to the ENSO-phenomenon, but as seen above and in several papers (Lefebvre *et al.*, 2004; Holland and Raphael, 2005b), both ENSO and SAM contribute to the ADP. Although the ADP is the first EOF in all of the important sea ice concentration datasets, it does not explain much of the variance of the sea ice (24% following Holland and Raphael (2005b)). When the sea level pressure is regressed onto the index of the first EOF of the sea ice concentration, a mixture of a SAM and an ENSO pattern is found (Holland and Raphael, 2005b). However, due to its dipole nature, the ADP is not capable of explaining much of the variance in the sea ice extent integrated over the Southern Ocean. However, as the first EOFs of the sea ice concentration all consist of positive and negative patches that tend to cancel each other out, the total sea ice extent variability is relatively small compared to the sectoral variabilities (Table 7.2) (Zwally *et al.*, 2002; Cavalieri *et al.*, 2003).

To find out what the most important atmospheric influences are on the integrated sea ice extent, several fields have been regressed onto the winter Southern Ocean sea ice extent (Lefebvre and Goosse, 2007a). This has shown that the sea ice extent integrated over the Southern Ocean is high when a specific static wave 3 pattern is found in the sea surface pressure. This wave 3 is associated with a low pressure anomaly above the southern Atlantic Ocean, southwest of Australia and in the Ross Sea and a high pressure anomaly above the Antarctic continent, the western Indian sector, the eastern Pacific sector and the Bellingshausen Sea. This pattern is not a mode of variability as the different centres of action are barely related. Furthermore, it does not have high correlations with any of the previous discussed atmospheric modes of variability. So, we must consider that the sea ice extent in the Southern Ocean over the last 30 years does not behave as a single entity. In other words, the variability of the sea ice in the Southern Ocean is the result of the combination of the variability in the different sectors. More information about this can be found in Lefebvre and Goosse (2007a) (see Chapter 6).

Chapter 4

Influence of the Southern Annular Mode on the sea ice–ocean system

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Abstract

The global sea ice–ocean model ORCA2-LIM, driven by the NCEP/–NCAR (National Centers for Environmental Prediction–National Center for Atmospheric Research) reanalysis daily 2-m air temperatures and 10-m winds and by monthly climatologies for precipitation, cloud cover and relative humidity, is used to investigate the impact of the Southern Annular Mode (SAM) on the Antarctic sea ice–ocean system. Our results suggest that the response of the circumpolar Southern Ocean consists of an annular and a non-annular component. For the sea ice cover, the non-annular component seems to be the most important. The annular component strongly affects the overall patterns of the upper ocean circulation. When the SAM is in its positive phase, a northward surface Ekman drift, a downwelling at about 45°S and an upwelling in the vicinity of the Antarctic continent are simulated. The non-annular component has a significant impact at the regional scale, especially in the Weddell, Ross, Amundsen and Bellingshausen Seas. In those regions, the pressure pattern associated with the SAM induces meridional winds which advect warmer air in the Weddell Sea and around the Antarctic Peninsula and colder air in the Amundsen and Ross Seas. This implies a dipole response of sea ice to the SAM, with on average a decrease in ice area in the Weddell Sea and around the Antarctic Peninsula and an increase in the Ross and Amundsen Seas during years with a high SAM index. The long-term trend in the observed sea ice area does not appear to be related to the trend in the SAM index.

4.1 Introduction

The circulation variability of the lower atmosphere in the Southern Hemisphere is dominated by an annular mode that significantly influences the strength of the westerlies (e.g. Connolley, 1997; Gong and Wang, 1999; Thompson and Wallace, 2000; Simmonds, 2003). This mode has been referred to as the High Latitude Mode (HLM), the Antarctic Oscillation (AAO), or the Southern Annular Mode (SAM). We will use the latter name and follow Thompson and Wallace (2000) in defining the SAM as the leading principal component (PC) of the 850-hPa geopotential height anomalies south of 20°S (or of the sea level pressure; the corresponding time series displaying a correlation of 0.99). The index time series can be

found at <http://www.jisao.washington.edu/aao/slp/>. A positive (negative) SAM index is associated with lower (higher) sea level pressures at high latitudes and higher (lower) sea level pressures at low latitudes (Figure 4.1). As a consequence, during years with a positive SAM index, the westerlies are intensified in the region of the Antarctic Circumpolar Current (ACC) and weakened north of 40°S. These characteristics are not confined to the lower levels of the atmosphere since the variability of the circulation at 500 hPa, 300 hPa, and in the stratosphere is also dominated by an annular mode (e.g. Kidson, 1999; Thompson and Wallace, 2000; Mo, 2000; Genthon *et al.*, 2003). Although the variability associated with SAM has a mostly zonal structure, a non-annular component can be found in the NCEP/NCAR and ECMWF reanalyses as well as in coupled model simulations (e.g. Mo, 2000; Limpasuvan and Hartmann, 2000; Cai and Watterson, 2002; Hall and Visbeck, 2002; Genthon *et al.*, 2003). In particular, a low pressure anomaly west of the Antarctic Peninsula is found for high values of the SAM index (Figure 4.1).

In a recent study, Thompson and Solomon (2002) have suggested that a high SAM index favors cooling over Antarctica, due to an increased isolation of the Antarctic continent. An important exception is the Antarctic Peninsula, where the enhanced westerlies associated with a high SAM index increase the advection of relatively warm oceanic air resulting in a significant warming of the area.

Atmospheric variability associated with SAM has an significant impact on the sea ice cover and the oceanic circulation. Because of the lack of observations, this effect has been mainly studied with Coupled General Circulation Models (CGCMs) (e.g. Watterson, 2000, 2001; Cai and Watterson, 2002; Hall and Visbeck, 2002). As those models are able to reproduce the observed characteristics of the SAM in the atmosphere reasonably well, a realistic response of the ocean and of the sea ice pack to the SAM can be expected. In particular, Hall and Visbeck (2002) summarized the influence of SAM in their model as follows. Stronger westerlies in the region of the ACC induce an intensified eastward surface ocean current and, as a result of the Ekman drift, a stronger northward surface current south of 45°S. North of 45°S, the pattern is reversed due to weaker westerlies. This causes an upwelling in the ocean around 65°S and a downwelling around 45°S. Positive phases of SAM are also associated with an oceanic heat divergence at about 60°S and a heat convergence at about 40°S, leading to a modest decrease in sea surface

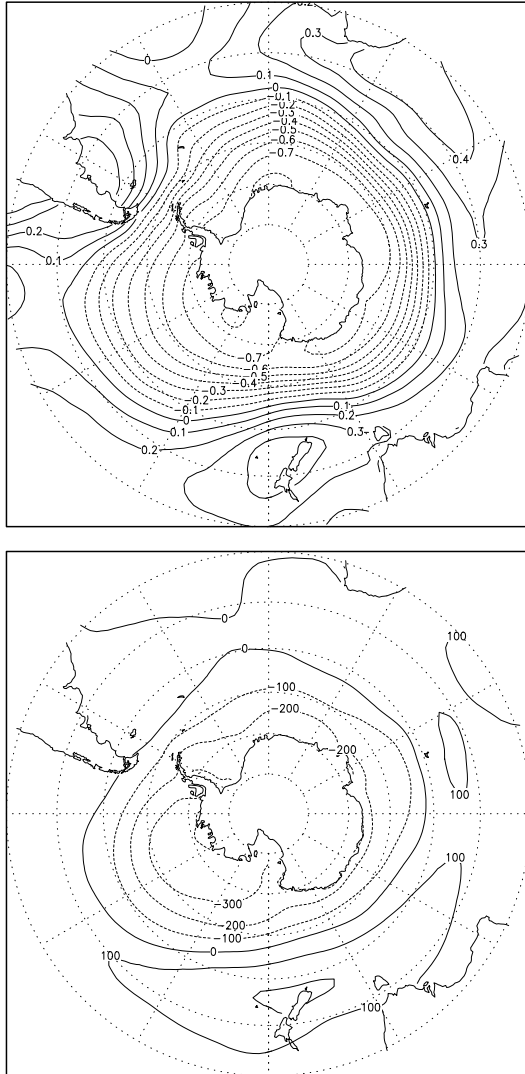


Figure 4.1: (Top) Annually averaged correlation between the seasonal mean (JFM, AMJ, JAS, OND) atmospheric surface pressure (NCEP/NCAR) and the seasonal mean SAM index for the period 1960-1999. (Bottom) Annually averaged regression [Pa] between the seasonal mean atmospheric surface pressure and the seasonal mean SAM index.

temperature (SST) at high latitudes and a weak warming¹ at mid latitudes. In addition to the role of wind stress variations discussed above, Watterson (2000) showed that surface heat fluxes are very important in generating anomalies of SST.

In the model of Hall and Visbeck (2002), the sea ice velocity is set equal to the velocity of the first oceanic level. As a consequence, because of the stronger northward oceanic current south of 45°S, the model ice is driven northward when the SAM index is positive. This yields a decrease in ice thickness close to the Antarctic continent, an increase further north, and a more extensive ice cover. However, using satellite observations, Kwok and Comiso (2002b) and Liu *et al.* (2004) showed that the response of sea ice to SAM variability over the last 20 years is different from the one proposed by Hall and Visbeck (2002) as it appears to be mainly characterized by a decrease in the Weddell and Bellingshausen sectors and an increase in the Ross and Amundsen sectors, i.e. by a non-annular response. Based on the SSMR/SSMI satellite data, Liu *et al.* (2004) also showed that the recent trends in the sea ice area cannot be attributed to SAM nor to the El Niño-Southern Oscillation (ENSO).

The goal of the present study is to provide further information on the response of the ice-ocean system to the variability associated with SAM. The study is based on a numerical experiment performed with the coupled sea ice-ocean model ORCA2-LIM (Timmermann *et al.*, 2004, 2005). Similar simulations with different models have been carried out to investigate the variability of the sea ice cover and of deep water formation in the Southern Ocean (e.g. Beckmann and Timmermann, 2001; Timmermann *et al.*, 2002a,b; Fichefet *et al.*, 2003a,b). Nevertheless, in contrast to the Northern Hemisphere where such a type of analysis is common and fruitful (e.g. Proshutinsky and Johnson, 1997; Zhang *et al.*, 2000; Krahmann and Visbeck, 2003), those experiments have not been used to systematically investigate the influence of a particular atmospheric mode of variability on sea ice and ocean. We will take profit of the complementarity of the approach proposed here to the previous studies conducted with CGCMs. On the one hand, CGCMs provide consistent fields from the top of the atmosphere to the bottom of the ocean. On the other hand, we employ a relatively sophisticated ice-ocean model with a higher resolution and more complete physics than in the CGCM stud-

¹This is a mistake in the original paper which states cooling.

ies. Furthermore, the use of atmospheric reanalysis data avoids possible model biases resulting in a better representation of the Southern Ocean in ice-ocean models than in CGCMs.

The model and the forcing are briefly described in Section 4.2. Section 4.3 presents the response of the model to the SAM in the zonal mean and compares our results to the previous investigations. This is followed by a discussion of the non-zonal component of the response, which appears quite important in our model and has not been thoroughly analyzed before with an OGCM. Conclusions are presented in Section 4.4.

4.2 Model description and experimental design

The model used here, called ORCA2-LIM, results from the coupling of the Louvain-la-Neuve sea ice model (LIM) (Fichefet and Morales Maqueda, 1997) with the hydrostatic, primitive equation ocean model OPA (Océan PARallélisé) (Madec *et al.*, 1999). OPA is a finite difference ocean general circulation model with a free surface and a non-linear equation of state in the Jackett and McDougall (1995) formulation. In the ORCA2-LIM configuration, lateral tracer mixing is done along isopycnals. Eddy-induced tracer advection is parameterized following Gent and McWilliams (1990) with the coefficients decreased between 20°N and 20°S. Momentum is mixed along model level surfaces using coefficients varying with latitude, longitude and depth. Vertical eddy diffusivity and viscosity coefficients are computed from a level-1.5 turbulence closure scheme based on a prognostic equation for the turbulent kinetic energy (Blanke and Delecluse, 1993). Double diffusive mixing (i.e. salt fingering and diffusive layering) is computed following Merryfield *et al.* (1999). In locations with a statically unstable stratification, a value of $100 \frac{m^2}{s}$ is assigned to the vertical eddy coefficients for momentum and tracers. The Beckmann and Döscher (1997) bottom boundary layer scheme ensures an improved representation of dense water spreading over topography in this geopotential-coordinate model.

The thermodynamic part of LIM uses a three-layer model (one layer for snow and two layers for ice) for sensible heat storage and vertical heat conduction within snow and ice. Vertical and lateral sea ice growth/decay rates are obtained from energy budgets at the upper and lower surfaces of the snow-ice cover, and at the surface of leads present

within the ice pack. At the upper surface, a prognostic treatment of the temperature in a very thin surface layer is used. The effect of the subgrid-scale snow and ice thickness distributions is accounted for through an effective thermal conductivity, which is computed by assuming that the snow and ice thicknesses are uniformly distributed between zero and twice their mean value over the ice-covered portion of the grid cell. Storage of latent heat inside the ice resulting from the trapping of shortwave radiation by brine pockets is taken into account. When the load of snow is large enough to depress the snow-ice interface under the water level, seawater is assumed to infiltrate the entirety of the submerged snow and to freeze there, forming a snow ice cap. The ice concentration, the snow volume per unit area, the ice volume per unit area, the snow enthalpy per unit area, the ice enthalpy per unit area, and the brine reservoir per unit area are advected with the ice drift velocity. The sea ice velocity field is determined from a momentum balance considering sea ice as a two-dimensional continuum in dynamical interaction with atmosphere and ocean. The viscous-plastic constitutive law proposed by Hibler (1979) is used for computing internal stress within the ice for different states of deformation. The model also includes a parameterization for the formation of leads due to shearing deformation following Stern *et al.* (1995). The coupled model is run on a global grid with 2° nominal resolution and a mesh refinement in high latitudes and near the equator. As a consequence, the meridional resolution increases to 0.5° close to the Antarctic continent, corresponding to a horizontal grid width of 50 km. Vertical discretization uses 30 z-levels, with 10 levels in the top 100 m. Model runs are initialized using data from the end of the 46-yr experiment described by Timmermann *et al.* (2005).

Daily 2-m air temperatures and 10-m winds from the NCEP/NCAR reanalysis project for the period 1948-1999 (Kalnay *et al.*, 1996), and monthly climatologies of surface relative humidity (Trenberth *et al.*, 1989), cloud fraction (Berliand and Strokina, 1980), and precipitation rate (Xie and Arkin, 1996) are utilized to drive the model. The surface fluxes of heat and moisture are determined from these data by using empirical parameterizations described by Goosse (1997). Evaporation/sublimation is derived from the turbulent flux of latent heat. In addition, a correction of ocean surface freshwater fluxes has been derived from the time-mean restoring salinity flux diagnosed from the experiment described by Timmermann *et al.* (2005) in order to circumvent the sur-

face salinity drift that occurs through the lack of any stabilizing feedback if the model is forced with slightly incorrect evaporation, precipitation and/or river runoff fields. The momentum fluxes at the various interfaces are obtained from standard bulk formulas. The parameterization of Large and Pond (1981) is utilized to compute the drag coefficient between air and water. For the drag coefficients between air and snow/ice and between ice and water, we use constant values of $2.25 \cdot 10^{-3}$ and $5 \cdot 10^{-3}$, respectively. This leads to a ratio of these two coefficients of 0.45, which is consistent with values adopted in other modelling studies of the Antarctic sea ice (e.g. Stössel, 1992; Fischer and Lemke, 1994; Harder and Fischer, 1999). Further details of the coupled model are given by Timmermann *et al.* (2004, 2005). The quality of the NCEP/NCAR dataset is questionable in polar regions because of the small number of observations there, in particular for the period prior to the start of the assimilation of satellite data in the late 1970s (e.g. Hines *et al.*, 2000; Marshall, 2002; Struraro, 2003). As a consequence, the analysis presented here will be performed on the last 20 years of the experiment. Even during this period, some uncertainties remain. An attempt to use the cloud cover, precipitation, and humidity from the NCEP/NCAR re-analysis data induced large errors in polar regions² (Timmermann *et al.*, 2005), which forced us to use climatologies for these variables. This implies that interannual variability of latent and solar heat fluxes and of the precipitation rate is not properly taken into account in our simulations. However, Timmermann *et al.* (2005) have shown that this forcing strategy still yields a good representation of the interannual variability of the sea ice cover. Furthermore, the conclusions derived from model results are compared to available observations (see end of Section 4.3) to estimate, at least qualitatively, the impact of these shortcomings.

4.3 Results

4.3.1 Zonal mean response

For zonal mean fields, the SAM index is highly correlated with the zonal and meridional wind stresses as well as with the zonal, meridional and vertical currents simulated by the model, with peak correlations between

²See also subsection 4.4.2, 5th paragraph under the heading Forcing.

0.6 and 0.8 (Figure 4.2). This suggests that a substantial part of the large-scale current variability in the Southern Ocean is due to the SAM. The strong positive correlation with the zonal wind stress between 45 and 65°S (Figure 4.2, top) is straightforward from the pressure pattern associated with the SAM (Figure 4.1), while the negative correlation with the meridional wind stress in the same latitude band mainly arises from the rotation of the wind vector in the surface boundary layer. The stronger westerlies found for positive SAM index induce enhanced zonal ocean surface currents and, due to the Ekman drift, also stronger meridional currents (Figure 4.2, middle). This results in a horizontal divergence at about 65°S, near the Antarctic continent, which leads to oceanic upwelling there. Besides, a horizontal convergence is found at about 65°S and, consequently, downwelling occurs there. The northward transport in the upper ocean is balanced by a southward return flow below 1500 m (Figure 4.2, middle). This indicates that the SAM has a large influence on the ocean currents over the whole water column as also shown by the correlation with vertical motion, which remains high even in the deep ocean (Figure 4.2, bottom).

The correlation between the simulated zonal mean mixed layer depth with the SAM index has a broad peak between 40°S and 65°S, with a maximum value exceeding 0.6 (Figure 4.3). This corresponds to a regression between the SAM index and the mixed layer depth of 5 to 7 m in summer and up to 50 m in winter. Since the latitudinal band 40–65°S is also the region with the maximum increase in the magnitude of the wind stress, the deeper mixed layers are likely due to a higher wind stirring of the water column. Variations of the density flux at the ocean surface are only weakly correlated with the simulated mixed layer depths. The density flux in our simulation even tends to stabilize the water column at latitudes with an increased mixed layer depth, which shows that the effect of the density flux on the simulated mixed layer depth is negligible. However, we must recall here that the forcing used in our simulation does not include any interannual variation in precipitation and features only a reduced variability of surface heat fluxes. Taking into account this additional variability might alter the conclusion about the effect of the SAM on the density flux (e.g. Genthon *et al.*, 2003).

We can consider all correlations above 0.4 as significant with a con-

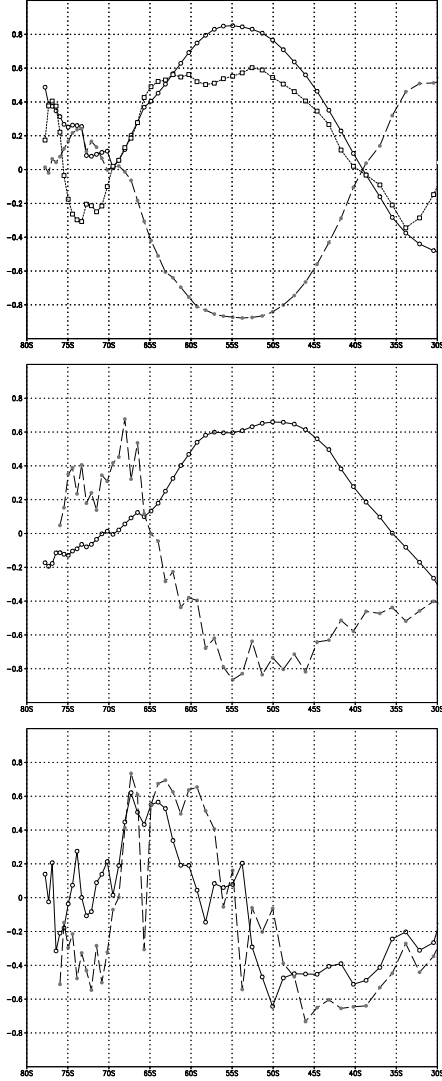


Figure 4.2: Correlation between the annual means of the SAM index and (top) the zonal means of the zonal wind stress (solid line with open circles), the meridional wind stress (long dashed line with circles), and the zonal current at the surface (dotted line with squares), (middle) the meridional current at the surface (solid line with circles) and at 2290 m depth (long dashed line with circles), (bottom) the vertical current at 45 m depth (solid line with circles) and at 2290 m depth (long dashed line with circles).

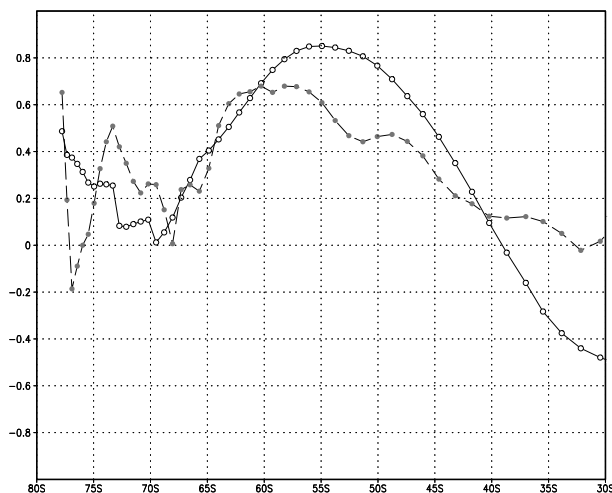


Figure 4.3: Correlation between the annual mean SAM index and the annual mean zonally averaged zonal wind stress (solid line with open circles), and the mixed layer depth (long dashed line with circles).

fidence of more than 95%³. All the SAM-related changes in the ocean currents (Figure 4.2) and also the increased mixed layer depth (Figure 4.3) between 65°S and 45°S are therefore significant. In contrast to that, Figure 4.4 (top) shows that the SAM explains only a very small fraction of the variations of the zonal mean surface air temperature over the ocean and of the modeled sea surface temperature (SST) in our simulation. The correlation between the SAM and the atmospheric temperature is smaller than 0.1 at all latitudes and thus not significant. The correlation with the SST has a slightly higher magnitude, but it barely exceeds -0.2 between 55°S and 65°S. This result is in qualitative agreement with the one obtained by Hall and Visbeck (2002), but the magnitude of the response is smaller here. Southward of 65°S, a complex pattern is found with cooling between 70°S and 74°S and warming in the latitudinal bands 68 – 70°S and 74 – 77°S, a feature not present in the Hall and Visbeck (2002) study. This results from different responses in the various sectors of the Southern Ocean as discussed in the next section. At depth, the temperature in the Southern Ocean tends to decrease when the SAM

³This has been tested using a Monte-Carlo method with 1.000.000 simulations of auto-correlated data series.

index is positive and to do so over the whole range of latitudes (south of 50°S). This can be explained by the intensified overturning (Figure 4.2) that brings cold deep water upward.

The correlation between the SAM index and the ice thickness is higher than the one with SST, but it still has a modest magnitude (Figure 4.4, middle) and is only marginally significant. Northward of 70°S , a negative correlation is found with a maximum value of about -0.4, and a positive correlation is found southward of this latitude with a peak value of 0.6. The correlation with the ice concentration features a similar pattern. These results are in contrast to the ones of Hall and Visbeck (2002) who found negative correlations with the ice thickness close to the coast and positive ones offshore. This difference is due to the different techniques used to compute the ice velocity in the two models. Hall and Visbeck (2002) assume that the ice is transported at the surface velocity of the ocean. This velocity has a strong diverging component resulting in a decrease in ice thickness close to the coast and a larger extension of the pack when the SAM index is positive. In our model, the response of the ocean surface velocity is similar to the one described by Hall and Visbeck (2002) (Figure 4.5, middle). Nevertheless, when solving the full momentum balance for sea ice, the angle between the wind stress and the ice velocity is much smaller than between the wind stress and the ocean velocity. Thorndike and Colony (1982); Hibler (1986) estimated that this angle is of the order of 20° . As a consequence, the divergence of the ice is much weaker than the one of the ocean at surface and the regression between ice velocity and the SAM index provides mainly a zonal transport that only has a little effect on the ice distribution. Hence the correlation between the zonal mean meridional sea ice velocity and the SAM index is always smaller than 0.3 in absolute value. The details of the sea ice response in our model will be discussed in Section 4.3.2.

Finally, the results of our model simulation allow studying the impact of the SAM on the salinity distribution in the ocean (Figure 4.4, bottom). In the zonal mean, we can easily explain the salinity variations by looking at the ocean current anomalies. The enhanced overturning and the increased northward currents cause a rise in sea surface salinity (SSS). At depth (around 2000 m), the increased southward currents are responsible for a decrease in salinity. Due to the inflow of saltier water from the Atlantic Ocean, the salinity around 3700 m rises again with increasing SAM south of 60°S . However, the magnitude of variability

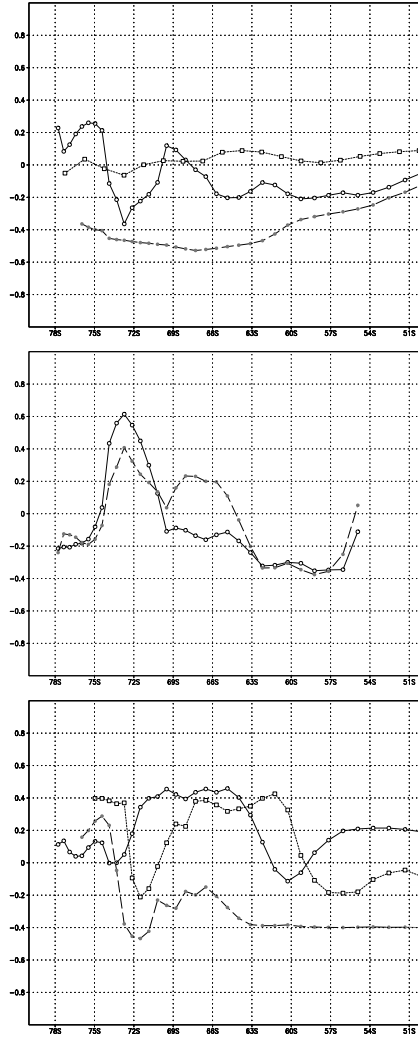


Figure 4.4: Correlation between the annual mean SAM index and the annual mean zonally averaged (top) 2-m air temperature over the ocean [NCEP/NCAR] (dotted line with squares), the sea surface temperature (solid line with open circles), and the ocean temperature at 2290 m depth (long dashed line with circles), (middle) ice thickness (solid line with open circles), and ice concentration (long dashed line with circles), (bottom) sea surface salinity (solid line with open circles), salinity at 2290 m depth (long dashed line with circles), and salinity at 3752 m depth (dotted line with squares).

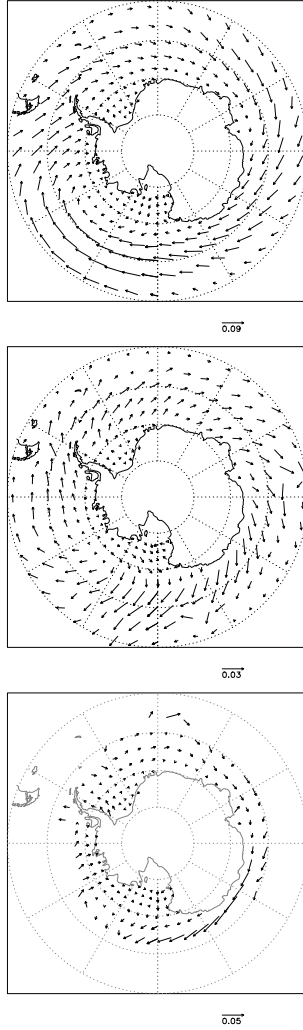


Figure 4.5: (Top) Annual mean of the regression of the horizontal wind stress [N/m^2] with the annual mean SAM index for the period 1980-1999. (Middle) Annual mean of the regression of horizontal ocean surface currents [m/s] with the annual mean SAM index for the period 1980-1999. (Bottom) Winter (July, August, September) regression of the ice velocity [m/s] with the seasonal mean SAM index for the period 1980-1999. Only points in which the ice concentration is higher than 0.15 for at least 10 years are taken into account. For clarity, only one sixteenth of the vectors are shown.

strongly decreases with depth. Variations of the zonal mean SSS exceed 0.1 psu, while salinity signals in the deep ocean are about 0.01 psu at 2290 m, and only 0.001 psu at 3700 m depth.

All results discussed above have been obtained with a zero time lag, since this has been found to yield the maximum correlations between the SAM index and the ocean currents. However, the correlation between SAM and the oceanic temperature increases by about 10% for a time lag of 4 months, with the SAM index leading, due to the huge heat capacity of water. Since this increase appears rather small and does not affect our conclusions, all further analysis continues using a zero time lag.

4.3.2 Geographical response

The results discussed in the previous section have revealed that the SAM explains only a small fraction of the variability of the simulated zonal mean SST, ice thickness and ice concentration. This leads to two contrasting hypotheses. The first one is that the SAM is very useful for understanding the oceanic current structure but not for understanding the SST and ice cover patterns. The second one assumes that the SAM has an influence on the SST and the ice cover in various regions but, as this impact may have a different sign at different longitudes along one latitudinal band, it is not seen clearly in the zonal mean. The goal of this section is to decide which of these two hypotheses is valid.

As mentioned in Sections 4.1 and 4.3.1, the correlation between the SAM index and the atmospheric surface pressure (Figure 4.1) has a clear annular component. Nevertheless, this pressure pattern is not perfectly annular. Specifically, a low pressure anomaly is found in the Amundsen and Ross sectors when the SAM index is high (the division in sectors can be seen in Figure 4.6). This yields a significant meridional component of the surface wind stress in these regions (Figure 4.5, top), which is directed towards the south in the Weddell, Amundsen and Bellingshausen Seas and towards the north in the Ross Sea. A similar pattern can be found in the ECMWF (European Centre for Medium-range Weather Forecasts) reanalysis data and also in results from CGCMs (e.g. Hall and Visbeck, 2002; Mo, 2000; Limpasuvan and Hartmann, 2000; Cai and Watterson, 2002; Genthon *et al.*, 2003).

In the zonal mean, the anomalies of ocean surface currents are eastward and northward south of 45°S for a high SAM index (Figure 4.2 and 4.5

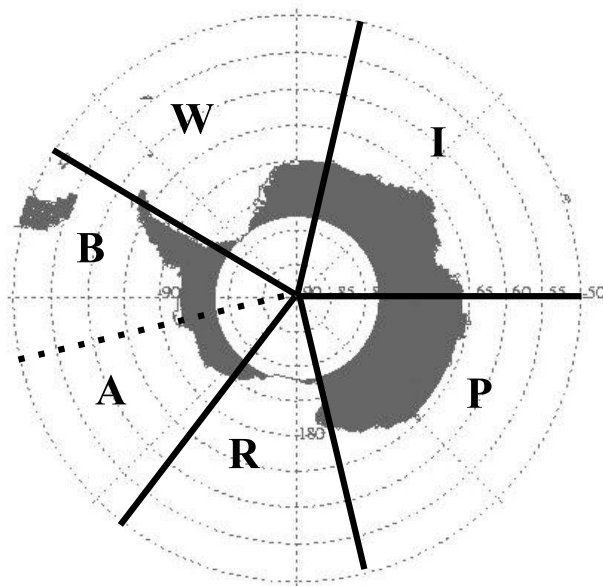


Figure 4.6: Sectors of the Southern Ocean used for analysis of results. These are the Weddell Sea (W; 60°W - 20°E), the Indian Ocean (I; 20°E - 90°E), the Pacific Ocean (P; 90°E - 160°E), the Ross Sea (R; 160°E - 140°W) and the Amundsen-Bellinghousen Seas (AB; 140°W - 60°W).

	Weddell	Indian	Pacific	Ross	A-B
W(M)	<u>0.50</u>	<u>0.56</u>	<u>0.50</u>	<u>0.67</u>	<u>0.46</u>
SSS(M)	<u>0.35</u>	<u>0.46</u>	<u>0.39</u>	0.23	0.23
SST (M)	<u>0.45</u>	-0.10	-0.25	<u>-0.69</u>	-0.17
Ice area (M)	<u>-0.29</u>	-0.16	0.05	<u>0.27</u>	0.18
Ice area (O)	<u>-0.30</u>	<u>0.33</u>	<u>0.25</u>	0.20	-0.02

Table 4.1: Correlation of the SAM index with the vertical current anomaly at 45 m depth (W), the sea surface salinity (SSS) anomaly, the sea surface temperature (SST) anomaly, and the ice covered area anomaly (ice concentration > 15%) for the period 1980-1999 for all seasons. For the vertical currents, the SST and the SSS, only the region south of 55°S is taken into account. M: Model results. O: Observations according to the HadISST dataset (Rayner *et al.*, 2003). Each time series consists of 80 values. Correlations exceeding the 90% significance thresholds are underlined. Correlations of more than 95% significance are double underlined. The significance has been tested using a Monte-Carlo method with 1.000.000 simulations of autocorrelated data series with the same auto-correlation as the time series presented in the table. The A-B denotes the Amundsen-Bellingshausen Seas

middle), but the non-annular component of the pressure pattern associated with the SAM has a clear local impact. In the Bellingshausen Sea, the correlation between the SAM index and the meridional surface currents is only slightly positive, with a very small divergence of the surface currents and a weak upwelling there. In the Ross Sea, currents associated with a high SAM index are westward close to the coast and have a very strong northward component offshore. sector by sector, there do not seem to be huge differences between the various regions (Table 4.1). Nevertheless, in the Amundsen-Bellingshausen Seas, high values of the SAM index are correlated with an intensified upwelling in the western sector that compensates for the very small vertical motion in the eastern Bellingshausen Sea. In the Ross Sea, there is a slightly more pronounced Ekman pumping due to the southerly winds in that region.

Although the meridional component of the wind associated with the SAM has a much weaker magnitude than the zonal one, it can have a significant influence on the atmospheric temperature because of the meridional atmospheric temperature gradient. In particular, the south-

ward transport in the vicinity of the Antarctic Peninsula is likely to play a role in the warming observed there during years with positive SAM index. This explanation is complementary to the hypothesis of Thompson and Solomon (2002) that attributes the local warming for a positive SAM index to an increase in the zonal flow. The increase in the southward component of near-surface winds (Figure 4.5) could also explain the air temperature rise at Orcada station (61°S , 45°W) that was not discussed by Thompson and Wallace (2000). Furthermore, a strong temperature signal in the NCEP/NCAR reanalysis is found in the eastern Ross Sea, the Amundsen Sea and the western Bellingshausen Sea, which is consistent with a response of the temperature to the anomalous meridional transport in those areas. Unfortunately, the temperature changes in this area were not analyzed by Thompson and Solomon (2002) because of the lack of station data in those regions.

The non-annular component of the wind pattern associated with the SAM also affects the SST through its influence on oceanic currents and on air temperature. In general, there is a cooling in all sectors during years with a high SAM index, which is consistent with the zonal mean response (see Figure 4.4). Nevertheless, because of the strong northward wind associated with a high SAM index in the Ross Sea, the cooling is much larger there (Table 4.1). The SST in the Weddell Sea, in turn, rises when the SAM index is positive, due to the enhanced advection of warm air and stronger oceanic upwelling that brings warmer water to the surface. In the Pacific and Indian sectors, the response is less clear. Northward of 60°S , the SST decreases as expected but, during some seasons, it increases close to the continent because of the large increase in upwelling there. When averaged over the Amundsen-Bellingshausen sector, only a small response remains. This is the result of a warming on the eastern part of this sector because of a stronger advection of warm air in this part and a cooling in the western part due to southerly winds in that part. In contrast to the SST, surface salinity changes do not show large differences between the various regions and the response is well represented by the zonal mean as discussed in Section 3.1 (see also Table 4.1 and Figure 4.4, bottom). The response of the simulated ice concentration generally follows the wind anomaly pattern and the changes in SST, with a general increase in the Ross Sea and a decrease in the Weddell Sea for positive phases of the SAM (Table 4.1). When analyzing the regional patterns season by season, the correspondence between the

observations (Figure 4.7) and the model (Figure 4.8) remains quite good. In particular, both model and data feature a large decrease of ice coverage in the Weddell Sea and an increase in the Ross Sea in winter during years with a high SAM index. Likewise, following the anomalies of near-surface winds (Figure 4.5, top) and ice drift (Figure 4.5, bottom), the Amundsen-Bellingshausen sector features a decrease of ice area around the Antarctic Peninsula and an increase in the Amundsen Sea during years with a high SAM index. This is in close agreement with the results of Liu *et al.* (2004) and Kwok and Comiso (2002b) derived from satellite data. In the western Pacific area between 100°E and 150°E, the signal appears more clear. Both observations and the model (Figures 4.7 and 4.8) feature an increase of winter ice concentration near the ice edge. This is due to a large northward sea ice transport in that region for positive phases of SAM (Figure 4.5, bottom). This can also be clearly seen in the winter ice thickness distribution (not shown), which features an increase offshore and a sharp decrease close to the continent. The response in this region could thus be explained by a mechanism similar to the one described by Hall and Visbeck (2002).

The only major disagreement between model and observations occurs in the Indian sector. In this region, the model response shows mainly a decrease in ice concentration during years with positive SAM index, except close to 50°E. In the observations, however, a positive anomaly of ice coverage centered around 50°E covers the majority of the Indian sector, with a decrease in ice concentration only at the boundaries of this sector. Furthermore, in the model the response in the Amundsen Sea appears to be larger than that close to the Antarctic Peninsula, while it seems to be the opposite in the observations (Table 4.1).

So, our results indicate that the SAM explains only a small fraction of the large-scale variations of sea ice coverage. This is in close agreement with the observational results of Liu *et al.* (2004) and Kwok and Comiso (2002b). Nevertheless, some model results are robust and significant, in particular the dipole Weddell Sea - Ross Sea (Table 4.1).

Note that, except in some regions (like the Ross Sea), the correlation between the SAM index and the ice-covered area is quite low and in particular lower than in Hall and Visbeck (2002), who obtained a correlation of 0.51 between ice coverage and the SAM index. The effect of internal ice stress, which is neglected in the Hall and Visbeck (2002) model but considered in LIM, obviously tends to blur the response of the ice pack to

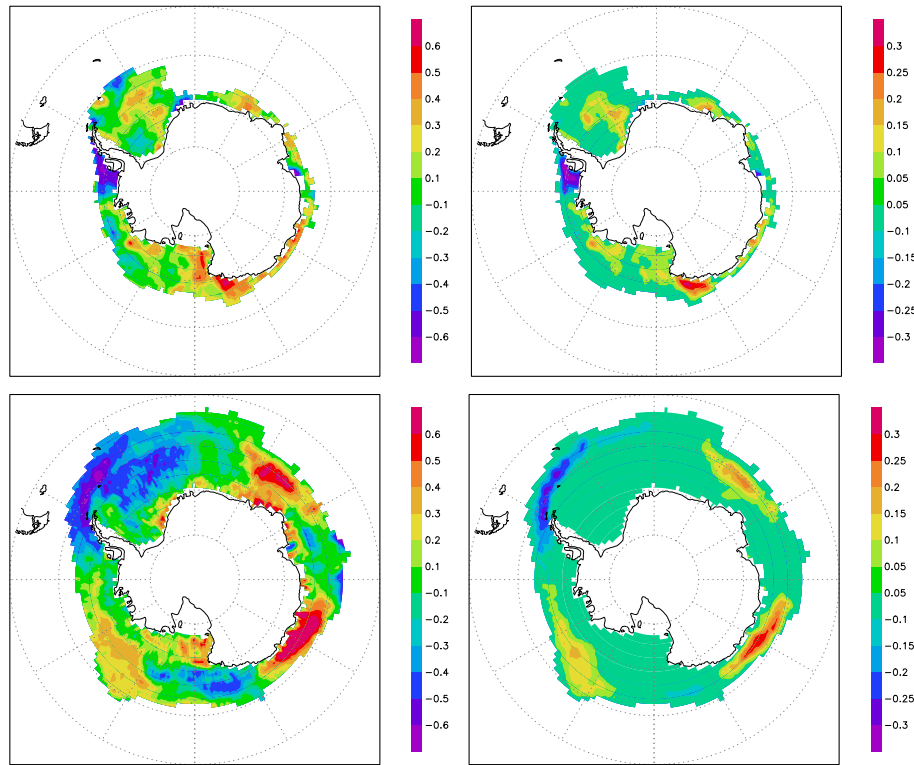


Figure 4.7: Correlation (left) and regression (right) between the seasonal mean SAM index and the seasonal mean ice concentration for the period 1980-1999 in the HadISST1 observations (Rayner *et al.*, 2003). Top: January, February, March; bottom: July, August, September.

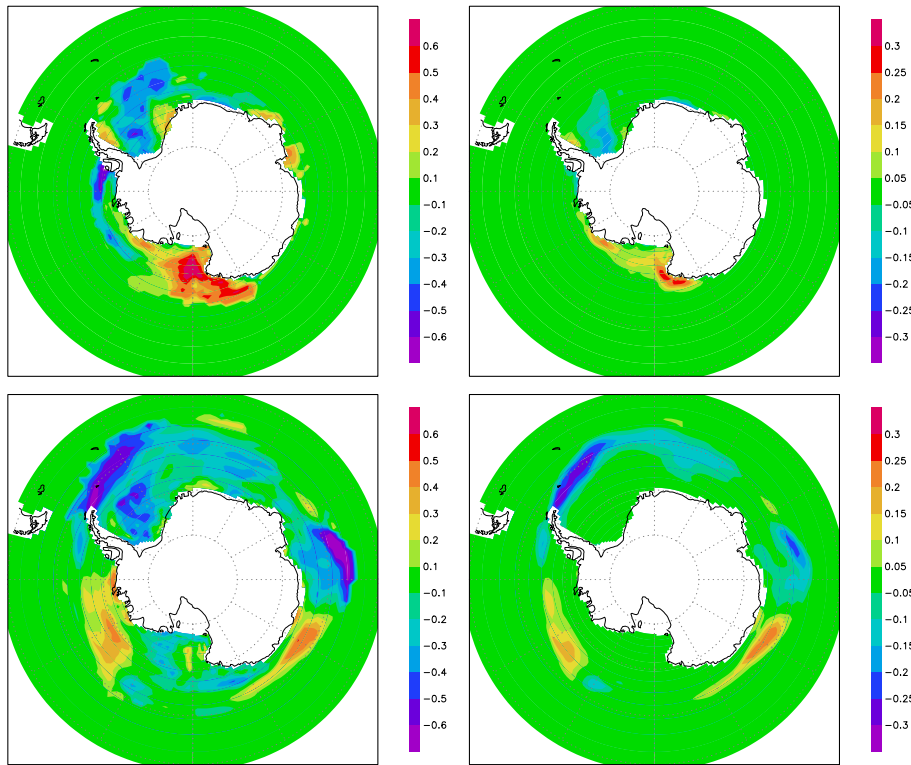


Figure 4.8: Correlation (left) and regression (right) between the seasonal mean SAM index and the seasonal mean ice concentration for the period 1980-1999 in ORCA2-LIM. Top: January, February, March; bottom: July, August, September.

anomalies of wind and ocean currents and therefore the close correlation between these fields.

In the evolution of the ice cover in the different sectors of the Southern Ocean described by Zwally *et al.* (2002), a clear influence of the SAM is found only in the Weddell Sea. For instance, the years 1980, 1981, 1991 and 1992 all feature a positive anomaly of sea ice extent in the Weddell Sea when the SAM index is low. The years 1989, 1993 and 1998, which are characterized by a high SAM index, all show an anomalously small ice area in the Weddell Sea. In the other sectors, the influence of the SAM on the ice-covered area is largely overwhelmed by other factors.

In our simulations there are no substantial salinity (or temperature) changes on the continental shelf associated with the SAM. Furthermore, the response of the mixed layer depth in regions where deep mixing occurs is weak. As a consequence, we can conclude that, in our simulation, the SAM does not have a significant impact on the formation of deep waters in the Southern Ocean and, in particular, on the Antarctic Bottom Water (AABW) production.

Although the quality of the forcing data is reduced during the first half of the NCEP/NCAR reanalysis, we have extended our analysis into the period 1960–1979. All the major characteristics of the response of the sea ice–ocean system to SAM are well reproduced for this early period, increasing our confidence in the robustness of our conclusions. For instance, during the period 1960–1979, we still find a decrease of ice extent in the Weddell sector when the SAM index is positive, along with an increase of ice coverage in the Amundsen sector and an increase of ice between 90°E and 140°E.

4.4 Conclusions

Using a coupled sea ice–ocean model driven by daily NCEP/NCAR reanalysis data for near-surface air temperatures and winds, and monthly climatologies for humidity, cloudiness and precipitation, we have confirmed earlier results obtained with CGCMs showing that the Southern Annular Mode (SAM) has a clear and strong impact on the structure of ocean currents over large parts of the water column in the Southern Ocean. At the surface, transport anomalies associated with a positive phase of SAM are directed towards the northeast at high latitudes (south

of 45°S) inducing an upwelling that features a maximum around 65°S and a downwelling at about 45°S. Due to continuity, this is balanced by a southward return flow below 1500 m. Furthermore, the enhanced wind during years with positive SAM index implies a stronger stirring of the water column and a deeper mixed layer during all seasons. The latter and the modification in the intensity of the upwelling, which have both an impact on the availability of nutrients in the upper ocean, might well affect the biological production in the Southern Ocean.

The impact of the SAM on the ocean surface temperature or on the ice-covered area varies strongly for the different regions of the Southern Ocean. Because of a low pressure anomaly in the Amundsen-Bellinghaussen sector during positive SAM years, the Weddell and Bellingshausen Seas are subject to more northerly winds, while the Ross Sea tends to have more southerly winds. This induces a cooling at the surface and an increase in the ice-covered area in the Ross and Amundsen Seas during years with positive SAM index while, at the same time, the SST increases and the ice-covered area decreases in the Weddell Sea. In summary, our results indicate that the main response of the ice concentration to the SAM is a dipole between the Weddell and the Ross Seas with a decrease of ice in the Weddell sector, and an increase in the Amundsen sector and parts of the Pacific sector for years with a positive SAM index. Integrated over the Southern Ocean, these regional differences tend to cancel out, so that signals in the zonally integrated SST and ice concentration are very small.

Those conclusions are statistically significant and in good agreement with observations, and with previous investigations (Kwok and Comiso, 2002b; Liu *et al.*, 2004). This suggests that despite the errors in the NCEP/NCAR data set and although the forcing used to drive our simulation does not include interannual variability of cloud cover, precipitation and relative humidity, we are able to capture, at least qualitatively, the main characteristics of the response of the sea ice-ocean system to the SAM.

Our results also indicate that the recent changes in ice-covered area in the Southern Ocean cannot be attributed to the general increase in the SAM index during the last 20 years. Indeed, an increase in SAM index would imply a less extensive ice cover in the Weddell Sea and a larger area of ice in the Ross Sea. For the Ross Sea this is consistent with observation of the trend performed by Zwally *et al.* (2002), but not for

the Weddell Sea. We conclude that the trend observed in the sea ice coverage by Zwally *et al.* (2002) cannot be attributed to the long-term trend of the SAM index. This clearly illustrates that the behavior of the ice cover in the Southern Ocean is a complex phenomenon and understanding its evolution needs to take into account all the factors that could affect it (e.g. Yuan and Martinson, 2000; Kwok and Comiso, 2002b; van den Broeke, 2000a).

The large regional differences in the response of the Southern Ocean to the SAM may appear strange in the context of an annular mode. Nevertheless, a zonal asymmetry of the temperature pattern can easily be attributed to the non-symmetric land distribution, even with purely zonal winds as argued by Thompson and Solomon (2002) to explain the temperature changes over the Antarctic Peninsula. Furthermore, despite its name, the definition of the SAM does not impose that it is strictly an annular phenomenon; the meridional components of the wind appear to play an important role in controlling the ice cover, in particular the low pressure anomaly over the Bellingshausen-Amundsen Seas during years with positive SAM index. A similar feature occurs in the Northern Hemisphere where the non zonally symmetric component of the Northern Annular Mode (or NAM, which is closely related to the North Atlantic Oscillation) is responsible for the good correlation between NAM and the ice export through Fram Strait (Kwok and Rothrock, 1999; Hilmer and Jung, 2000; Vinje, 2001). However, the response of the meridional winds on the NAM appears as a second order characteristic of the annular mode, not as a first order one like the change in zonal mean oceanic currents. For instance, the correlation between NAM and the ice export at Fram strait does not hold anymore for long time scales (Hilmer and Jung, 2000; Vinje, 2001).

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Chapter 5

Influence of the Southern Annular Mode on the sea ice-ocean system: the role of the thermal and mechanical forcing

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Abstract

The global sea ice-ocean model ORCA2-LIM is used to investigate the impact of the thermal and mechanical forcings associated with the Southern Annular Mode (SAM) on the Antarctic sea ice-ocean system. The model is driven by idealized forcings based on regressions between the wind stress and the air temperature at one hand and the SAM index the other hand. The wind-stress component strongly affects the overall patterns of the ocean circulation with a northward surface drift, a downwelling at about 45° S and an upwelling in the vicinity of the Antarctic continent when the SAM is positive. On the other hand, the thermal forcing has a negligible effect on the ocean currents. For sea ice, both the wind-stress (mechanical) and the air temperature (thermal) components have a significant impact. The mechanical part induces a decrease of the sea ice thickness close to the continent and a sharp decrease of the mean sea ice thickness in the Weddell sector. In general, the sea ice area also diminishes, with a maximum decrease in the Weddell Sea. On the contrary, the thermal part tends to increase the ice concentration in all sectors except in the Weddell Sea, where the ice area shrinks. This thermal effect is the strongest in autumn and in winter due to the larger temperature differences associated with the SAM during these seasons. The sum of the thermal and mechanical effects gives a dipole response of sea ice to the SAM, with a decrease of the ice area in the Weddell Sea and around the Antarctic Peninsula and an increase in the Ross and Amundsen Seas during high SAM years. This is in good agreement with the observed response of the ice cover to the SAM.

5.1 Introduction

The Southern Annular Mode (SAM, also referred as the Antarctic Oscillation (AAO) or High Latitude Mode (HLM)) is the dominant mode of variability of the atmospheric circulation southwards of 20° S. It is associated with changes in the position and the strength of the westerlies (e.g. Connolley, 1997; Gong and Wang, 1999; Thompson and Wallace, 2000; Simmonds, 2003) that induce a stronger Ekman Drift towards the north in the Southern Ocean during years with high SAM index as well as an increased upwelling close to the continent and a downwelling around

45° S (Mo, 2000; Watterson, 2000, 2001; Cai and Watterson, 2002; Hall and Visbeck, 2002; Lefebvre *et al.*, 2004).

The influence of the SAM on the ice cover appears more complex (Kwok and Comiso, 2002b; Liu *et al.*, 2004; Lefebvre *et al.*, 2004) than this annular oceanic response, whose main characteristics are similar at all the longitudes. Indeed, the correlation between the SAM index and the ice extent displays a dipole, with more ice in the Ross Sea and less in the Weddell Sea. This clearly non-annular pattern appears driven by a decrease in sea level pressure in the Amundsen-Bellingshausen Seas when the SAM is high. This induces a deflection of the westerlies towards the south-east around the Peninsula and towards the north-east in the Ross sector. These winds change the sea ice velocities, the advection of heat by ocean currents, and also advect warmer air in the Weddell Sea and colder air in the Ross Sea, resulting in the dipole described above (Lefebvre *et al.*, 2004). Those studies also underline that, in some regions, the SAM explains only a small fraction of the internal variability of the system and a lot of other factors must be invoked in order to understand the variability of the Southern Ocean.

The SAM has a distinct effect on the location and strength of the Antarctic Circumpolar Current (ACC), although the lack of coherence between the SAM and the circumpolar current on the multidecadal timescale shows that the circumpolar current is primarily controlled by another mode on long timescales (Hall and Visbeck, 2002). Furthermore, anthropogenic factors could influence both the ACC and the SAM, as discussed for instance by Fyfe and Saenko (2005).

The goal of this study is to provide further information about the origin of the response of sea ice and ocean to the SAM. In particular, we try to disentangle the effect of the wind stress change from those resulting from the changes in air temperature. To do so, several numerical experiments have been performed with the coupled sea ice-ocean model ORCA2-LIM (Timmermann *et al.*, 2005). In those experiments, we force the model with idealized forcings derived from the regressions between the SAM and the air temperature and wind stress, respectively. The model and the forcings are briefly described in Section 5.2. Section 5.3 provides a justification of our methodology. In Sections 5.4 and 5.5, the responses to the thermal and mechanical forcings are analyzed. Finally, the conclusions are presented.

5.2 Model description and experimental design

5.2.1 Model description

The model used here, called ORCA2-LIM, results from the coupling of the Louvain-la-Neuve sea ice model (LIM) (Fichefet and Morales Maqueda, 1997) with the hydrostatic, primitive equation ocean model OPA (Océan PARallélisé) (Madec *et al.*, 1999). The coupling technique between OPA and LIM is identical to the one described in Goosse and Fichefet (1999). The model version used here is exactly the same as the one that has been used in Lefebvre *et al.* (2004), which includes a more extensive model description.

OPA is a finite difference ocean general circulation model with a free surface and a non-linear equation of state in the Jackett and McDougall (1995) formulation. The thermodynamic part of LIM uses a three-layer model (one layer for snow and two layers for ice) for sensible heat storage and vertical heat conduction within snow and ice while the sea ice velocity field is determined from a momentum balance considering sea ice as a two-dimensional continuum in dynamical interaction with atmosphere and ocean. The viscous-plastic constitutive law proposed by Hibler (1979) is used for computing internal stress within the ice for different states of deformation.

The coupled model is run on a global grid with 2° nominal resolution and a mesh refinement in high latitudes and near the equator. As a consequence, the meridional resolution increases to 0.5° close to the Antarctic continent, corresponding to a horizontal grid width of about 50 km. Vertical discretization uses 30 z-levels, with 10 levels in the top 100 m. Model runs are initialized using data from the end of the 46-yr experiment described by Timmermann *et al.* (2005). Daily 2-m air temperatures and 10-m winds from the NCEP/NCAR reanalysis project for the period 1980–1999 (Kalnay *et al.*, 1996), and monthly climatologies of surface relative humidity (Trenberth *et al.*, 1989), cloud fraction (Berliand and Strokina, 1980), and precipitation rate (Xie and Arkin, 1996) are utilized to drive the model. The surface fluxes of heat and moisture are determined from these data by using empirical parameterizations described by Goosse (1997).

A correction of ocean surface freshwater fluxes has been derived from the time-mean restoring salinity flux diagnosed from the experiment de-

scribed by Timmermann *et al.* (2005) in order to circumvent the surface salinity drift that occurs through the lack of any stabilizing feedback if the model is forced with slightly incorrect evaporation, precipitation and/or river runoff fields.

5.2.2 Experimental design

In addition to the control run (CTRL) analyzed in Lefebvre *et al.* (2004), a series of experiments has been performed in which a perturbation of the forcing is applied on top of the normal forcing. Those idealized perturbations are derived from the regression between the SAM-index (defined as the leading principal component and the monthly NCEP/NCAR reanalysis (Kalnay *et al.*, 1996) sea level pressure anomalies south of 20°S ¹) with the wind stress and the air temperature for the period 1980–1999. In a first group of experiments (mechanical forcing), the perturbation is applied only to the wind stress: in M+, an annual anomaly of the wind stress corresponding to one standard deviation of the SAM (regression pattern of Figure 5.1) is added, while this anomaly is subtracted in M– (Table 5.1). In a second group of experiments, the perturbation is only applied to the air temperature: in T+, a seasonal anomaly of the air temperature corresponding to one standard deviation of the SAM (regression patterns found in Figure 2) is added, while this anomaly is subtracted in T–. In a third group of experiments, both the air temperature and the wind stress forcing have been perturbed. Adding to the standard forcing one regression between the wind stress and the air temperature and the SAM results in S+. Subtracting them gives S–. All these forcings have been imposed during the period 1980–1999 (see also Section 5.3). Nevertheless, as discussed in Section 5.3, we will use below only the first seven years of the simulation. This provides a robust description of the model response to the idealized forcing while avoiding some long-term effects that are not the subject of the present study.

The mechanical forcing (Figure 5.1) displays an almost annular signal, with wind stress anomalies coming from the northwest. However, in the Weddell Sea and around the Antarctic Peninsula, the northern component is more intense. In Ross Sea, the wind stress anomaly is in general eastwards between 50°S and 60°S and directed northwards close to the continent. Those regional features are due to the SAM-related

¹ which can be found at <http://www.jisao.washington.edu/aao/slp/>

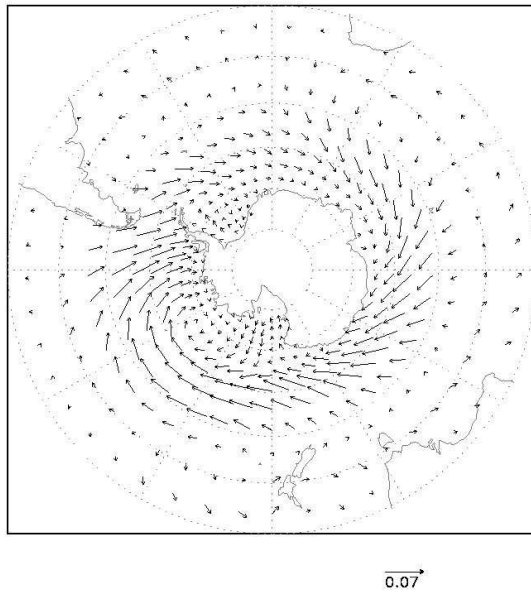


Figure 5.1: Annual mean of the regression between the wind stress ($\frac{N}{m^2}$) and the annual mean SAM index for the period 1980–1999. This pattern has been used in the idealized forcings M+, S+ (NCEP/NCAR wind stress forcing + regression), M– and S– (NCEP/NCAR wind stress forcing – regression).

Table 5.1: Description of the experiments.

Name	Wind stress perturbation	Air temp. perturbation
CTRL	No Change	No Change
M+	+1 standard deviation	No Change
M-	-1 standard deviation	No Change
T+	No Change	+1 standard deviation
T-	No Change	-1 standard deviation
S+	+1 standard deviation	+1 standard deviation
S-	-1 standard deviation	-1 standard deviation
Ann+	+1 annular part of the regr.	No Change
Ann-	-1 annular part of the regr.	No Change
NA+	+1 non-annular part of the regr.	No Change
NA-	-1 non-annular part of the regr.	No Change

low-pressure area in the Amundsen-Bellingshausen Seas (Limpasuvan and Hartmann, 2000). Furthermore, the increase in the magnitude of the westerlies associated with a high SAM index is slightly stronger in the Ross and Amundsen sectors than in the Weddell and Indian sectors.

In order to compare the impact of the annular component of the wind stress anomalies with the non-annular part, due to the low pressure in the Amundsen-Bellingshausen Seas, an additional series of experiments has been performed (Table 5.1). In a first group of experiments, only the zonal mean of the wind stress regression is applied (annular part of the forcing): in Ann+, the zonal mean is added to the standard forcing for the years 1980–1999, while the anomaly is subtracted in Ann-. In a last series of experiments, the same methodology has been used for the non-annular anomalies (NA+ and NA-) obtained by subtracting the zonal mean from the wind stress pattern of Figure 5.1.

In all seasons, the air temperature above the Weddell Sea and around the Antarctic Peninsula is higher when the SAM is positive, while everywhere else, the air temperature is lower, especially close to the continent (Figure 5.2). This is in good agreement with the regression data from Thompson and Solomon (2002), who saw the same pattern in the air temperature data series on the Antarctic continent. The largest changes are observed in autumn and winter. Although these air temperature

changes appear quite large, the correlation is low (at maximum around 0.4) (Lefebvre *et al.*, 2004). This is related to big natural variations in air temperature in this region and thus a large standard deviation of the air temperature there that leads to large regression values despite those low correlations. This implies that, even if we find a robust response of the ocean to the air temperature forcing, the low correlations of the SAM with the air temperature warns us that a lot of other factors will be at least as important for temperature changes as the SAM. As the air temperature regression pattern to the SAM is clearly non-annular, analyzing the influence of the zonal mean changes in a way similar to experiments Ann and NA for the air temperature forcing does not appear very instructive and such experiments are not discussed here.

The method presented here has the advantage that we keep the interannual variability of the forcing (as we keep on having the variability in the NCEP/NCAR dataset) and that we can identify clearly the role of each component of the atmospheric forcing (wind stress or air temperature). However, this method can only be successful if the linearization provides a good approximation of the response of the real system and if there are no important, either positive or negative, feedbacks between the oceanic response on the wind stress on the one hand and the air temperature on the other hand. Furthermore, the method could pose problems because the same perturbation is applied every year. After a few years, the long-term response of the system to a constant perturbation could thus be different than the one observed in the real world where the polarity of the SAM changes nearly each year. These limitations will be discussed in Section 5.3.

5.3 Justification of the experimental design

5.3.1 Checking the linearity and comparing with the observations

The approach of this paper is only valid if the southern sea ice-ocean system responds approximately in a linear way to the SAM. This assumption could be decomposed in two parts. Firstly, it is necessary to check that the sum of the response to the thermal forcing only and to the mechanical forcing only is equal to the response to the total forcing and that this response is comparable to the results of the control ex-

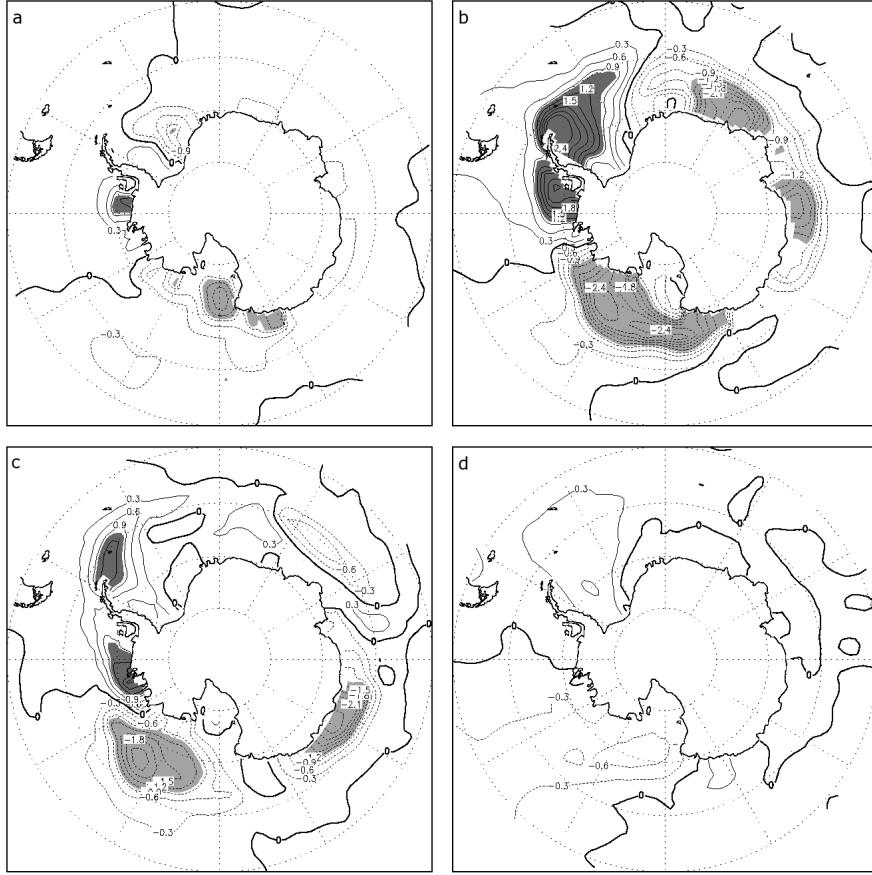


Figure 5.2: Seasonal mean of the regression between the air temperature ($^{\circ}\text{C}$) (NCEP/NCAR reanalysis (Kalnay *et al.*, 1996)) and the seasonal mean SAM index (NCEP/NCAR reanalysis) for the period 1980-1999. a) summer. b) autumn. c) winter. d) spring. Contour interval is 0.3°C . Solid lines correspond to positive values, while dotted lines show negative values. Shading in dark gray means a regression of more than 1°C , whereas shading in light gray shows a regression of less than -1°C . This pattern has been used in the idealized experiments T+, S+ (NCEP/NCAR air temperature forcing + regression), T- and S- (NCEP/NCAR air temperature forcing - regression).

periment (Lefebvre *et al.*, 2004) and to the observations. To illustrate that, we have displayed on Figure 5.3 the sum of the response of the total sea ice area to the thermal and mechanical forcings taken separately (y-axis) as a function of the sea ice area response to the total (thermal + mechanical) forcing (x-axis). For each dot, we have selected the same time during the first seven years of the experiments in all experiments in order to have a reasonable sampling of simulated variability. As can be seen, the correlation is very good, so we can state that if we are able to understand the response to the individual forcings, we also understand the response to the total idealized forcing associated to the SAM. Furthermore, it is necessary to check if the response to this idealized forcing (S) is similar to the observed response of the ice cover to the SAM. The validity of this hypothesis is shown clearly on Figure 5.4. We see that the geographical distribution of the winter sea ice response to the idealized SAM forcing in the model is very similar to the observed response in the HadISST-dataset (Rayner *et al.*, 2003). As discussed in Lefebvre *et al.* (2004), this observed response is also very close to the one obtained in the control experiment. Nevertheless, there are some discrepancies that must be kept in mind when interpreting the model response, in particular, the model tends to underestimate the decrease in sea ice concentration in the Amundsen-Bellingshausen sectors associated with a positive SAM index, while the decrease is overestimated in the Weddell sector.

As a second aspect of the linearity issue, one has to determine if the response to an upward change in the SAM is the opposite of the response to a similar downward change in the SAM. As can be seen in Figure 5.5, this is not totally valid. There is a clear linear relationship between the responses to positive and negative anomalies. But, the changes induced by a decrease in the SAM are smaller than the responses due to an increase in the SAM. More precisely, in Figure 5.6, we observe that the spatial distribution of the anomalies is similar in experiments S+ and S- but the magnitude of the changes is larger in S+.

5.3.2 Checking the dependency of the results to initial conditions and the long-term effect of the SAM

The response to the SAM could be influenced by the choice of initial conditions and by the years selected to start the experiments. For example, if the SAM is already high during the first year of the experiment, does

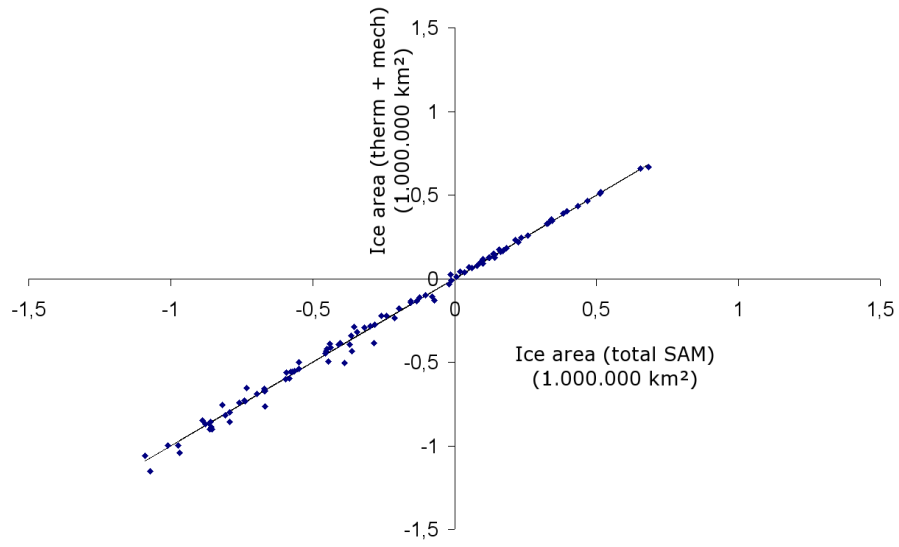


Figure 5.3: For the first seven years of the simulations (15 points per year), we represent the following sums: on the X-axis; the ice area (in $1.000.000 \text{ km}^2$) of $(S^+ - S^-)/2$ and on the Y-axis; the ice area (in $1.000.000 \text{ km}^2$) of $(M^+ - M^- + T^+ - T^-)/2$. The line ($X=Y$) shows a perfect correspondence between the different sets of experiments.

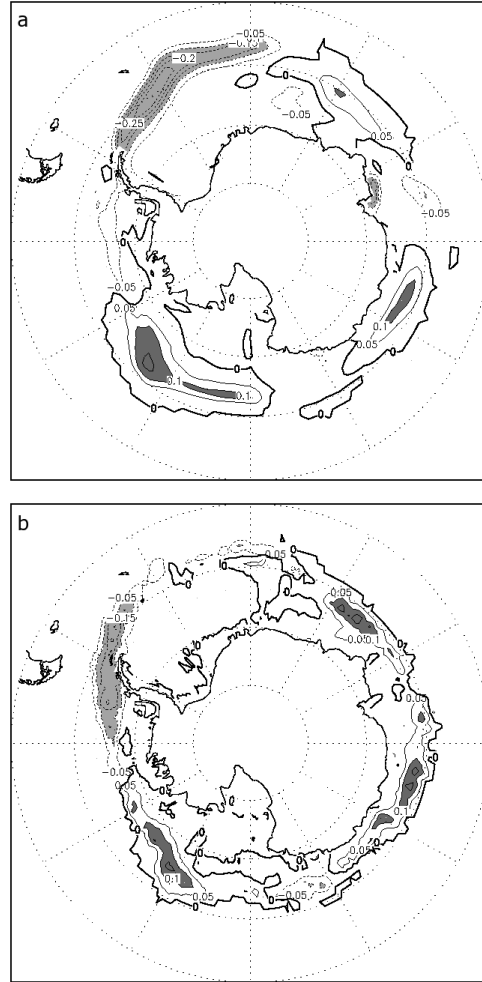


Figure 5.4: a) Half of the differences in sea ice concentration at the end of July between S+ and S− for the first seven years of the idealized simulations. b) Regression between the seasonal mean SAM index and the July mean ice concentration for the period 1980–1999 in the HadISST1 (Rayner et al., 2003) observations. Contour interval is 0.05. Solid lines are positive values, while dotted lines show negative values. Shading in dark gray means a value of more than 0.1, whereas shading in light gray shows a value of less than 0.1.

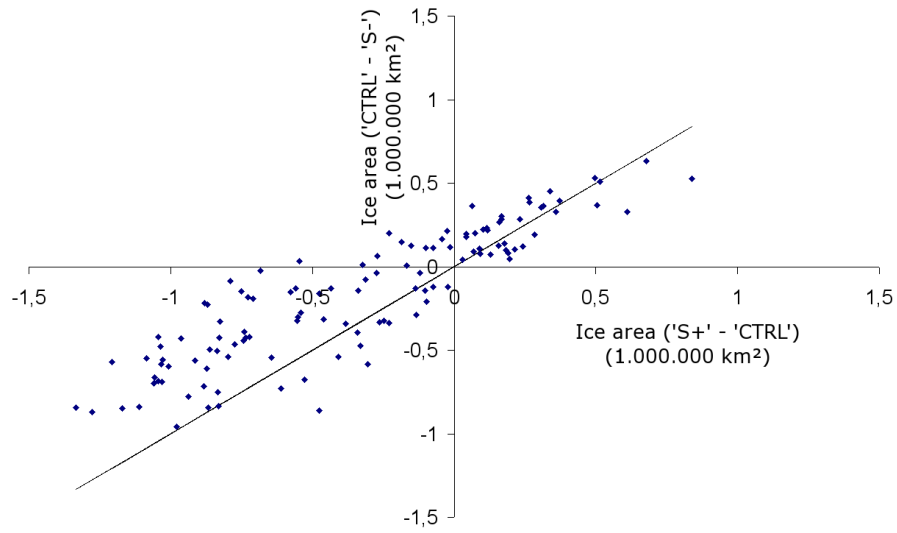


Figure 5.5: For the first seven years of the simulations (15 points per year), we represent the following sums: on the X-axis; the ice area (in $1.000.000 \text{ km}^2$) of ('S+' - 'CTRL') and on the Y-axis, the ice area (in $1.000.000 \text{ km}^2$) of ('CTRL' - 'S-'). The line (X=Y) shows a perfect correspondence between the different sets of experiments.

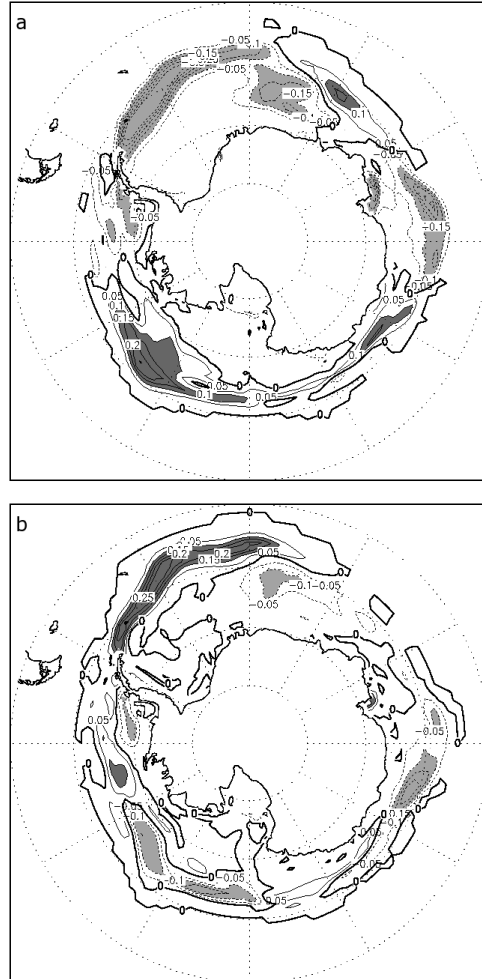


Figure 5.6: a) The difference in sea ice concentration at the end of July between S+ and CTRL for the first seven years of the idealized simulations. b) The difference in sea ice concentration at the end of July between S- and CTRL for the first seven years of the idealized simulations. Contour interval is 0.05. Solid lines are positive values, while dotted lines show negative values. Shading in dark gray means a value of more than 0.1, whereas shading in light gray shows a value of less than 0.1.

an increase of the SAM has the same effect as if the SAM was initially low? If the initial state was a cold one with a lot of sea ice, will this influence the response of the sea ice to the SAM? To answer these questions, we have started the idealized simulations at different years. In the standard experiment, the simulations are started in 1980. In addition, we have carried out experiments similar to S+ but starting in 1981, 1982 and 1983. For each of those experiments, the model is initialised using the corresponding year of the control run. The anomalies due to the modified forcings have been compared to the perturbations obtained for the experiments starting in the year 1980. We can see that the results of the various experiments (Figure 5.7) are, in general, similar, although all the simulations starting in 1981, 1982 and 1983 seem to have slightly less ice than the standard experiment. More precisely, they show a stronger response of the sea ice area if the sea ice area anomaly is negative and a smaller response if this anomaly is positive. If we split up the results in two parts (on the one hand, the first two years after the beginning of the perturbation (empty symbols) and, on the other hand, the other years (full symbols)), we see that the initial conditions only play a role in the first years of the idealized simulations. Even in these years, the differences (memory effect) in the changes of the sea ice area due to the SAM, in general, account for about 200.000 km^2 only. This shows that the influence of initial conditions and starting years is weak. In addition, as the difference in the response of the runs starting in different years decreases with time, the observed trend in the SAM has only a negligible effect on our results. Furthermore, the natural trend of the SAM shows a sharp jump upwards at the end of the seventies. Afterwards, the index shows a sharp decline and a slight upwards trend (Marshall, 2003). In our results of the runs starting in different years, we do not see a difference that could be attributed to these natural changes in the trend of the SAM index.

Finally, we have to take into account another possible problem. We are using, in the different runs, a constant perturbation year after year. As a consequence, long-term effects could kick in after a few years. The first signal of such a long-term effect can be seen after eight years, when an increased convection around Maud Rise can be found in the M+ and the S+ experiments. To eliminate these unwanted effects, we will work in the rest of this paper with means over the first seven years of the simulation. The sea ice response to the SAM in these years does not

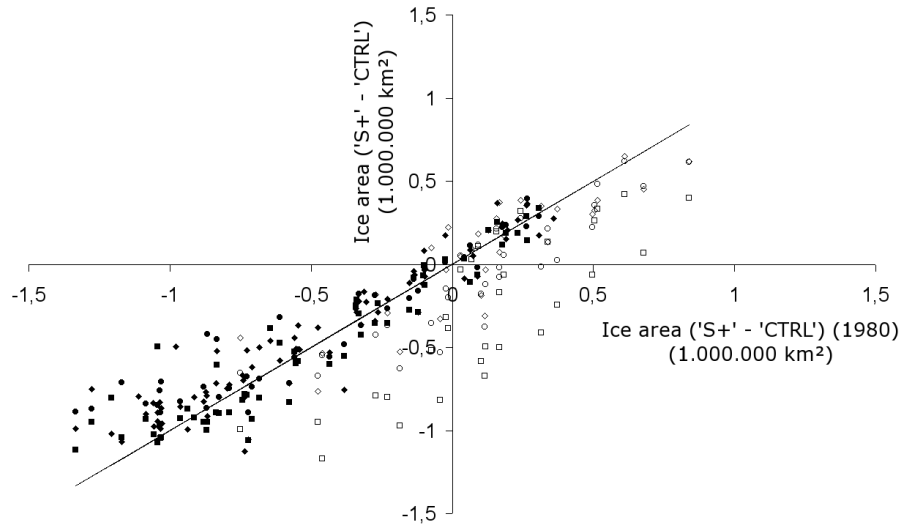


Figure 5.7: The dependency of the sea ice response to the SAM to the initial conditions. Each point represents a comparison between two different initial conditions at a certain time after starting the experiment with S+ forcing. On the x-axis: the sea ice area anomaly (in 1.000.000 km²) (difference with the CTRL-run) after a certain time after the starting of the simulation in 1980. On the y-axis: the sea ice area anomaly (in 1.000.000 km²) (difference with the CTRL-run) after a certain amount of time after the starting of the simulation in 1981 (diamonds), in 1982 (squares) and in 1983 (circles). If the time after the beginning of the simulation is smaller than two years empty symbols are used, otherwise filled symbols are used. The line shows the perfect correspondence, i.e. the case where we get exactly the same differences whatever year we take.

differ a lot from one year to another, so by averaging over these years, a reasonable SAM-like sea ice response is obtained (see above). The period 1980-86 of the simulation provides thus a good representation of the subdecadal model response associated to the idealized perturbation. As a consequence, only results for this period are presented here. Nevertheless, analysing a period of similar length but starting at a different date would give similar conclusions.

The discussion presented in this section illustrates that the linear hypothesis used to construct our experimental design is reasonable. Our numerical experiments could thus be used to better understand the response of the ocean and sea ice as discussed hereafter.

5.4 Role of the thermal forcing

The response of the Southern Ocean to the thermal part of the SAM is relatively straightforward. The colder air in the Indian, Pacific, Ross and Amundsen sectors induces a larger sea ice cover in these regions (Figs.5.2 and 5.8). The effect is the strongest during the seasons with the largest changes in temperature: autumn and winter. However, in the the Weddell and Bellingshausen sectors, an increase in air temperature results only in a marginal decrease in winter sea ice concentration (see Figure 5.8). The autumn ice production is lower in these regions due to the large temperature differences in this season (Figure 5.2) inducing a lower ice extent (not shown). The difference in temperature decreases strongly in winter and this induces only a small effect on the winter sea ice cover, as shown in Figure 5.8. We have also to take into account the discrepancies between the observations and the model in this region, as mentioned in Section5.3.1.

The net effect of the thermal forcing is an increase in sea ice area associated with a positive SAM index (Table5.2), which reaches its maximum in June with a value of 538.000 km². The change in air temperature does not have, as could be expected, much effect on the currents. Furthermore, there are no big changes in the mixed layer depth in response to the changes in air temperature, even when taking into account the sea ice changes. However, we must recall here that the forcing used in our simulation does not include any interannual variation in precipitation and features only a reduced variability of surface heat fluxes (only

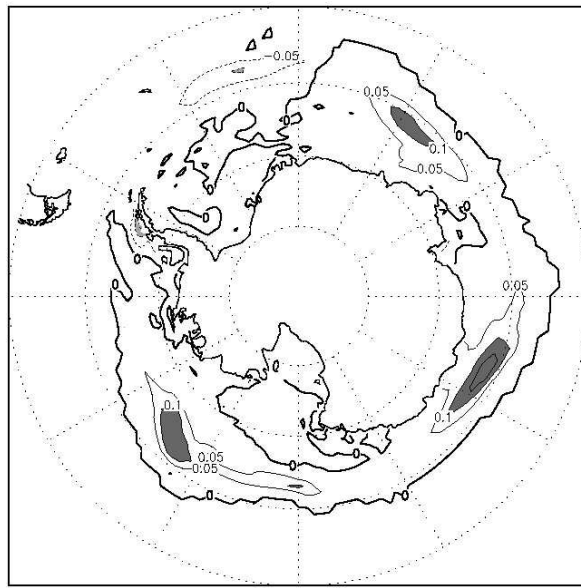


Figure 5.8: Half of the differences in sea ice concentration at the end of July between $T+$ and $T-$ for the first seven years of the idealized simulations. Contour interval is 0.05. Solid lines are positive values, while dotted lines show negative values. Shading in dark gray means a value of more than 0.1, whereas shading in light gray shows a value of less than 0.1.

Table 5.2: Ice area winter (JAS) differences between the different simulations in $1.000.000 \text{ km}^2$. T: $(T+ T-)/2$. M: $(M+ M-)/2$. Ann: $(\text{Ann}+ \text{Ann-})/2$. NA: $(\text{NA}+ \text{NA-})/2$. S: $(S+ S-)/2$. The division into sectors is as follows: Weddell Sea (60° W – 20° E), Indian Ocean (20° E – 90° E), Pacific Ocean (90° E – 160° E), Ross Sea (160° E – 140° W) and the Amundsen-Bellingshausen Seas (140° W – 60° W).

	Weddell	Indian	Pacific	Ross	Amundsen- Bellingshausen	Total
T	−0.035	0.136	0.117	0.091	0.030	0.339
M	−0.326	−0.118	−0.071	0.035	−0.051	−0.530
Ann	−0.219	−0.089	−0.029	−0.015	−0.052	−0.403
NA	0.058	0.098	0.029	0.078	0.015	0.279
S	−0.382	−0.020	0.034	0.110	−0.010	−0.269

temperature is varying). Taking into account this additional variability might alter the conclusion about the effect of the SAM on the density flux (e.g. Genthon *et al.*, 2003).

5.5 Role of the mechanical forcing

In contrast to the thermal part, the mechanical forcing implies complex mechanisms. When the SAM index is high, the stronger westerlies induce enhanced zonal ocean surface currents and, due to the Ekman drift, also stronger meridional currents (Figure 5.9). This is the source of a horizontal divergence close to the Antarctic continent (around 65° S), which leads to oceanic upwelling there. A horizontal convergence can be found at 45° S and, as a consequence, a downwelling occurs there. The anomalous vertical velocities have the same sign over the whole depth of the Southern Ocean. On the other hand, the northward surface current is balanced by a southward flow around 2300 m depth (not shown). All these results are in good agreement with the results of Hall and Visbeck (2002) and Oke and England (2004). In the latter paper, the oceanic response to a change in the latitude of the Southern Hemisphere Subpolar Westerly Winds was investigated, which is closely related to the SAM. In our simulations, the enhanced zonal currents give a relatively small increase in the Antarctic Circumpolar Current, but its location does

not change. These results are also in agreement with Hall and Visbeck (2002).

The annual mean of the zonal mean response of the mixed layer depth is displayed in Figure 5.10. We can see an increase of the mixed layer depth in the Southern Ocean at almost all latitudes with a peak between 45° S and 60° S. This increase is likely due to wind stirring as the region between 45° S and 60° S is also the region where the largest changes in wind stress occur. Because of the time average, Figure 5.10 hides also the differences between the seasons. In particular, during the local summer, the changes are generally of the order of a few meters, but in winter, they reach locally 80 m. Although the wind stirring plays the dominant role on the zonal mean, the changes in the vertical velocity could also affect the stability of the water column locally and thus have an influence on the mixed layer depth, as discussed below.

To understand the changes in sea ice, it is necessary to look at both the annular and the non-annular components of the SAM. The non-annular component of the wind is most important in the Ross, the Bellingshausen and the Weddell sectors, with a southern wind component in Ross Sea and a northern wind component in the Weddell Sea and in the Bellingshausen sector. This yields a significant meridional component of the surface wind stress in these regions (Figure 5.1), which is directed to the south in the Weddell and Bellingshausen Seas and towards the east-north-east in the Ross Sea. These changes in wind stress also lead to local changes in the currents.

Our sensitivity experiments (Table 5.1) help to determine precisely the influence of those regional anomalies compared to the annular ones. The winter annular response could be separated into two parts (Figure 5.11). First of all, the annular component induces larger mixing and so induces an upward transfer of warm and salty water. Secondly, the enhanced upwelling due to the increased westerlies also pumps up warm and salty water. These two effects lead to a decrease of the sea ice concentration. In addition, this enhanced pumping of salty water could contribute to the destabilization of the water column by increasing the surface density. In this experiment, the changes in meridional ice velocities are weak and have no significant effect on the ice cover. This contrasts with the results of Hall and Visbeck (2002), as discussed in details in Lefebvre *et al.* (2004). This could be due to the simple ice drift scheme in the model used by Hall and Visbeck (2002), which assumes

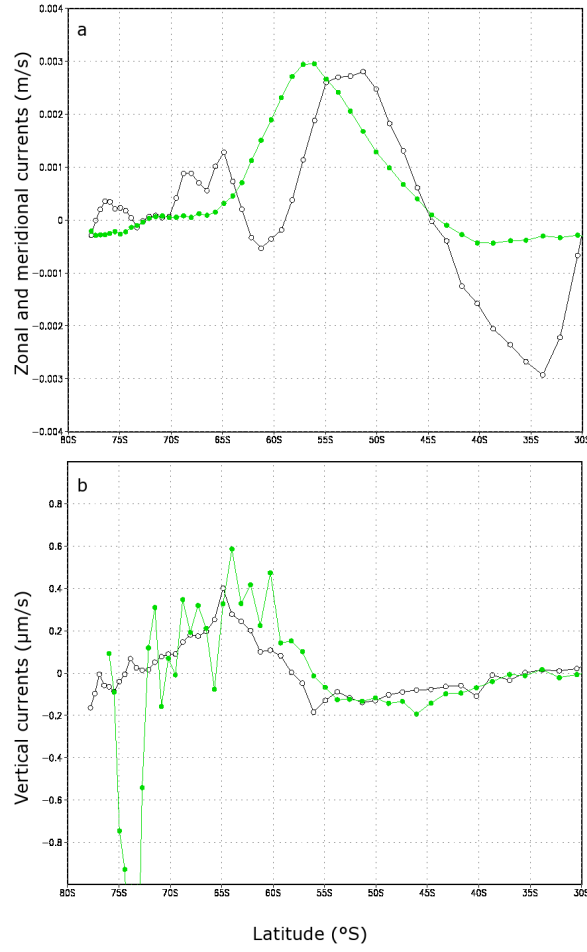


Figure 5.9: Half of the annual mean differences of the zonally averaged ocean currents between M+ and M− for the first seven years of the simulation. a) the zonal surface currents (in $\frac{m}{s}$) (black, empty circles) and the meridional surface currents (in $\frac{m}{s}$) (green, full circles). b) the vertical currents (in $\frac{\mu m}{s}$) at 45 m depth (black, empty circles) and at 2290 m depth (green, full circles).

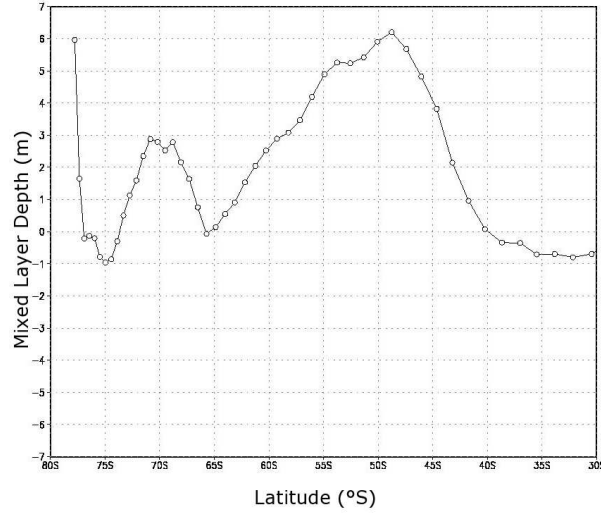


Figure 5.10: Half of the annual mean differences of the zonal means of the mixed layer depth (in m) between M+ and M- for the first seven years of the simulation.

that the sea ice velocity is the same as the oceanic velocity in the top ocean level.

The response to the non-annular wind stress forcing (Figure 5.11, b) shows a large change in the Ross Sea. This is due to a more northward ice drift induced by the more southerly winds. The small spots in the Bellingshausen Sea are a result of the opposite effect, i.e. a southward ice drift. The relatively small effects in the Indian and the Pacific sectors are also results of northward ice drift, induced by a slight increase in southerly wind. In the center of the Weddell Sea, the ice is pushed towards the continent. This cannot be seen clearly on the ice concentration, as the winter sea ice concentration in the Southern Weddell Sea is close to 100%, nevertheless, the sea ice thickness increases in the southern Weddell Sea (around 70° S; 40° W). Besides, the wind stress and currents close to the Antarctic Peninsula induce an increase in the eastward velocity of the sea ice. This leads to a sharp decrease in sea ice thickness at the eastern coast of the Peninsula.

The sum (not shown) of the responses to the annular and the non-annular wind stress forcings (Figure 5.11) is close to the response to

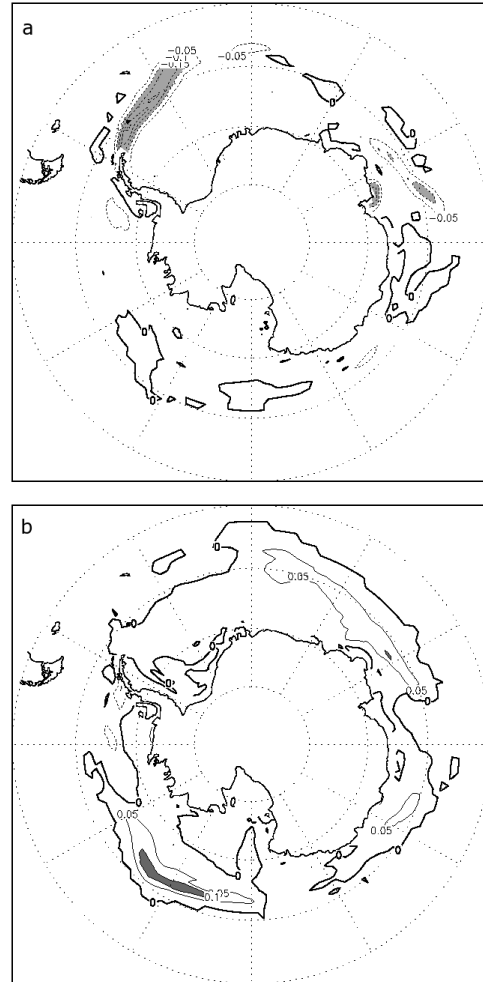


Figure 5.11: a) Half of the differences in sea ice concentration at the end of July between Ann+ and Ann- for the first seven years of the idealized simulations. b) Half of the difference in sea ice concentration at the end of July between NA+ and NA- for the first seven years of the idealized simulations. Contour interval is 0.05. Solid lines are positive values, while dotted lines show negative values. Shading in dark gray means a value of more than 0.1, whereas shading in light gray shows a value of less than 0.1.

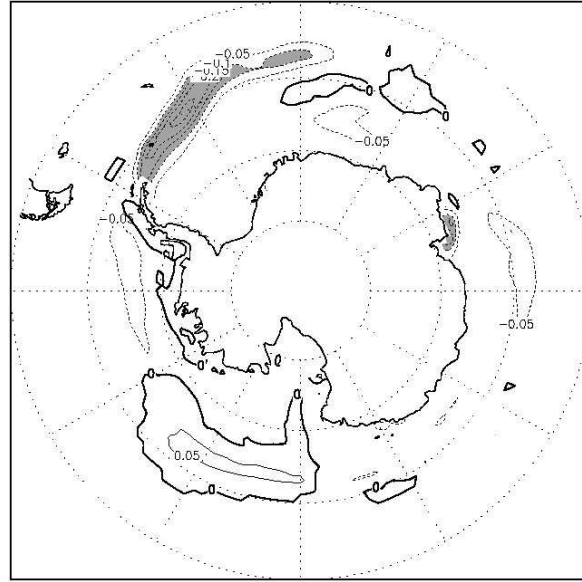


Figure 5.12: Half of the differences in sea ice concentration at the end of July between M+ and M- for the first seven years of the idealized simulations. Contour interval is 0.05. Solid lines are positive values, while dotted lines show negative values. Shading in dark gray means a value of more than 0.1, whereas shading in light gray shows a value of less than 0.1.

the total wind stress forcing (Figure 5.12) with only some minor differences, mainly located in the Indian and Pacific sectors. Integrated over the whole Southern Ocean, the net effect of the mechanical (wind-stress driven) part of the SAM is a decrease in ice area that reaches a September value of about 530.000 km^2 (Table 5.2).

As the winter sea ice thickness is in the Weddell Sea everywhere lower in the M+ run than in the CTRL run (due to the large response to the annular part of the forcing), the summer sea ice area decreases sharply there because the thinner sea ice melts more quickly. The total differences in sea ice area between the M+ and CTRL runs in summer (around 577.000 km^2 at the beginning of March) can be accounted for the most important part to these changes in the Weddell Sea (around 499.000 km^2 at the beginning of March). In the Ross Sea, the differences

in summer sea ice area are small (around 7.000 km^2). In the other sectors (Amundsen-Bellingshausen, Indian and Pacific), the summer ice area anomaly is small in the M+ run, as almost all the ice is melted in both runs.

5.6 Conclusions

In order to extract the role of the wind stress and the air temperature forcing in the subdecadal response associated with the SAM, we have made a series of runs with a coupled sea ice-ocean model. In those experiments, an idealized anomaly based on the regression between the NCEP/NCAR reanalysis data and the SAM index has been added to the standard forcing, some of the runs with only the mechanical part of the anomaly of the forcing, others with only the thermal part of the idealized anomaly. We have seen that the thermal part (air temperature) resulted in more sea ice when the SAM index is positive, because of the lower air temperature, mainly in autumn and winter except in the Weddell Sea and around the Peninsula where the warmer air temperature there is responsible for a decrease in the ice concentration.

The response to the mechanical forcing (wind stress) can be decomposed into several different terms. The first one is related to changes in the ice velocity, which is largely due to the non-annular part of the mechanical forcing. This is responsible for an increase in ice area in the Ross Sea as it pushes ice away of the continent. The annular part induces higher upward oceanic velocities close to the Antarctic continent due to an increased Ekman pumping when the SAM is high. As a result, warmer water rises and the ice melts. The latter effect can be seen in the Indian, Pacific, Ross and Amundsen-Bellingshausen sectors. Another effect of this pumping is the rise in surface salinity in the Weddell Sea, which destabilises the water column. This results in an increased mixed layer depth and a lesser ice area. In addition, the mechanical forcing has an effect on the currents (northward surface currents and southward deeper in the ocean, pumping close to the continent, downwelling around 40° S) and on the mixed layer depth (increased mixed layer depth due to increased wind stirring and locally to the increased upwelling) and the salinity.

When analysing the ocean currents, the results of the total forcing are, in general, very close to those from the runs with only wind-stress forcing. This is logical, as it is the wind stress and not the air temperature that forces the Ekman drift.

For the ice cover, the response to the SAM forcing can be explained as the sum of the responses to the distinct mechanical and thermal forcings. The combination of the two gives a dipole in winter, with less ice in the Weddell sector and more ice in the Ross sector (Figure 5.4). The response in the eastern Bellingshausen sector follows the Weddell sector, whereas the western Amundsen sector has a behaviour similar to the Ross sector. In the other sectors, the decreases in sea ice concentration due to the enhanced pumping in the winter sea ice concentration are compensated by the increases due to the lower air temperatures. These changes are all in good agreement with the observations and with our previous experiments (Lefebvre *et al.*, 2004) (Figure 5.4). When integrated over the different sectors (Table 5.2), the decrease in ice extent in the Weddell Sea and the increase in the Ross Sea are noticed in nearly all the runs. For those two areas, the thermal and mechanical effects thus tend to reinforce each other. On the other hand, mechanical and thermal forcings tend to have opposite responses in the other sectors, resulting in weaker total changes. The responses to the thermal and mechanical forcings have more or less the same amplitude. This shows that, despite the low correlation between the SAM index and the air temperature, the thermal forcing could by all means not be neglected. This is due, on the one hand, to the large variability of the air temperature in the polar areas (low correlations are thus not associated with small temperature changes) and, on the other hand, to the strong impact of temperature changes on the ice cover.

Table 5.2 also illustrates that, integrated over the whole Southern Ocean, the thermal forcing induces an increase in ice extent, while the mechanical forcing is responsible for a decrease in ice extent. The latter is mainly due to the annular part of the forcing since the non-annular part is associated with an increase of the ice extent. Overall, the SAM is associated with a decrease of the ice extent in winter in our simulations. Nevertheless, the complexity of the geographical distribution of the response and of the mechanisms involved with compensating effects of thermal and mechanical forcings imply that any long-term influence of the SAM on the ice extent integrated over the Southern Hemisphere

is very difficult to estimate from the available information. As a consequence, this integrated response could not be considered as a robust model result at this stage.

Furthermore, we must keep in mind that our distinction between the thermal and mechanical forcings is a bit artificial. First, in the real world, changes in wind direction (and thus in wind stress) are generally associated with changes in air temperature. In this case, both wind stress and temperature changes could be related to a common cause. Secondly, our experimental design does not allow to take into account any feedback between the atmosphere and the ice-ocean system. For instance, a change in the wind stress could induce a modification of the location of the ice edge and thus a change in air temperature that could amplify the initial response. By using our experimental design, those changes in wind stress and air temperature would appear unrelated while, in this example, they are strongly linked. As a consequence, our experiments show clearly that both changes in mechanical and thermal forcings are important to understand the response of the ice-ocean system, but we are not able in the present framework to determine the links between those two forcings nor to analyse the way the atmospheric anomalies are produced.

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Chapter 6

An analysis of the atmospheric processes driving the large-scale winter sea-ice variability in the Southern Ocean

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Abstract

The response of the sea-ice in the Southern Ocean to the variability of the atmosphere on the period 1979-2004 is investigated using both model and observational data. On the one hand, our results show that, in line with previous investigations, the classical modes of atmospheric variability do not explain a large part of the winter sea-ice extent variability integrated over the entire Southern Ocean. On the other hand, the regression between the ice extent and the atmospheric pressure displays a pattern with low pressure areas over the South Atlantic, over the Indian Ocean and on the southwestern part of the Pacific Ocean and a high pressure area over the Antarctic region, extending towards both the southeastern Pacific and the region south of Australia. This pattern, which does not correlates significantly with any of the known modes of atmospheric variability, induces changes in the sea-ice extent by influencing the air temperature, the ice production and both the meridional and the zonal ice velocity. However, there is no clear link between the different centers of action of this pattern. Furthermore, the correlation of the sea-ice extent between the different sectors is generally low. As a conclusion, we must consider that the sea-ice extent in the Southern Ocean over the last 30 years, does not behave as a single entity: its variability is the result of the combination of regional sea-ice changes. Each sector should thus be examined separately.

6.1 Introduction

Thanks to satellite data, we now have nearly 30 years of high quality information on the sea-ice concentration in the Southern Ocean (Zwally *et al.*, 2002; Cavalieri *et al.*, 2003). Over this period, the sea-ice extent integrated over the whole Southern Ocean does not show a significant trend and the Antarctic continent (except the Antarctic Peninsula) is one of the only regions of the Earth to defy the upward surface temperature trend due to the greenhouse effect (Houghton *et al.*, 2001; Vaughan *et al.*, 2003; Trenberth *et al.*, 2007). For some periods (1979-1998, monthly data), there is even a small significant rising trend in the sea-ice extent (Zwally *et al.*, 2002). However, regional trends in sea-ice extent have been significant (e.g. Stammerjohn and Smith (1997), Zwally *et al.* (2002) and Cavalieri *et al.* (2003)). This is in clear contrast with the rapidly

declining sea-ice extent in the high latitudes of the Northern Hemisphere (Cavalieri *et al.*, 2003).

On the one hand, this absence of a downward trend in the sea-ice extent in the Southern Hemisphere has classically been attributed to the slow response of the Southern Ocean to radiative forcing (Manabe *et al.*, 1991, 1992). On the other hand, this absence could be related to changes in the modes of variability of the climate system. In particular, it has been suggested that the rising trend in the Southern Annular Mode (SAM) (Figure 6.1) during the last 25 years has induced a cooling over the continent (Thompson and Solomon, 2002) and an increase in the ice cover that could counterbalance the warming due to the greenhouse effect. However, as Kwok and Comiso (2002b), Liu *et al.* (2004), Lefebvre *et al.* (2004), Lefebvre and Goosse (2005) and Stammerjohn *et al.* (2007) have shown, the correlation between the SAM index (defined as the index of the leading principal component (PC) of the surface pressure anomalies south of 20°S which can be found at <http://www.jisao.washington.edu/aao/slp> (Thompson and Wallace, 2000)) and the sea-ice extent integrated over the Southern Ocean is slightly positive but not significant, even if we only take a 90% significance threshold. Indeed, the correlation is 0.25 (see Table 6.1).

The El Niño-Southern Oscillation (ENSO) induces a dipole response in the sea-ice (Kwok and Comiso, 2002a; Yuan, 2004; Stammerjohn *et al.*, 2007), with more ice in the Amundsen sector and less around the Antarctic Peninsula in a year with a high Southern Oscillation Index, i.e. in a La Niña year. Possible mechanisms for this response have been discussed by Liu *et al.* (2002a), Turner (2004) and Yuan (2004). However, the correlation between the Multivariate ENSO Index (MEI) (which can be found at <http://www.cdc.noaa.gov/people/klaus.wolter/MEI/mei.html>) and the total sea-ice extent is very weak with a value of -0.01. The correlation between the total sea-ice covered area and another well-known index of variability, the ZW3-index, is also very weak (Table 6.1). This ZW3 corresponds to the Zonal Wave 3, the most important quasi-stationary wave 3 pattern in the non-tropical Southern Hemisphere of the total variance in the annual mean 500mb pattern as discussed by Van Loon and Jenne (1972) and Raphael (2004). The ZW3-index is calculated here using the method discussed in Raphael (2004) using the zonal anomaly of the sea level pressure.

	W(O)	I(O)	P(O)	R(O)	AB(O)	T(O)
SAM (Ann)	<u>-0.53</u>	<u>0.47</u>	0.20	0.27	0.30	0.25
ENSO (Ann)	0.21	-0.17	0.31	-0.32	-0.14	-0.01
ZW3 (Ann)	0.07	0.05	0.09	0.09	-0.22	0.08
SAO (J-D)	-0.22	-0.21	0.26	-0.28	-0.06	<u>-0.44</u>
SAO (N-O)	-0.26	-0.22	0.09	0.17	0.05	-0.18
SAO (M-F)	-0.17	-0.23	0.19	-0.26	0.09	-0.31

Table 6.1: The correlation between the observed (1979-2004) (O; HADISST (Rayner *et al.*, 2003) and updates) September sea-ice extent in the different sectors (columns) and several modes of variability over the Southern Ocean (rows). W, I, P, R, AB and T denote the Weddell sector, the Indian sector, the Pacific sector, the Ross sector, the Amundsen-Bellingshausen Seas and the total Southern Ocean, respectively.

For the SAO, J-D denotes that the amplitude is calculated on the period January to December. N-O corresponds to a period starting two months earlier (November to October), whereas M-F corresponds to a period starting two months later (March to February).

Significances are shown in **bold** when they exceed the 90% threshold, while significances exceeding the 95% threshold are shown underlined. Numbers that are both **underlined and bold** are significant above the 99% threshold. The significance was tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the table.

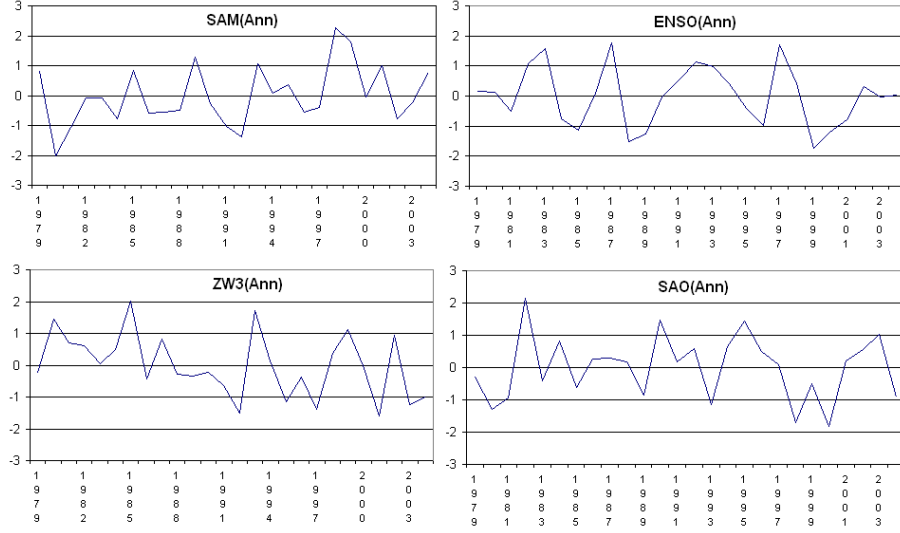


Figure 6.1: The interannual variability of the different indices. All the indices have been normalised to a mean of 0 and a standard deviation of 1.

Following Van Loon (1967) and van den Broeke (2000b), the Semi-annual Oscillation (SAO) is the twice-yearly contraction and expansion of the circumpolar pressure trough in the middle and high latitudes of the Southern Hemisphere. An SAO index can be calculated by taking the zonally and meridionally (40°S - 70°S) averaged amplitude of $H_2(p)$, the second harmonic of the annual surface pressure evolution (van den Broeke, 2000b) derived from the NCEP/NCAR re-analysis (Kalnay *et al.* (1996) and updates) over the period January to December. As shown in Table 6.1, this SAO index is significantly correlated with the total area covered with sea ice (correlation of -0.44). However, the correlations between the sea-ice extent in the different sectors and the SAO are relatively small. Furthermore, if the SAO index is calculated over a slightly different period (November to October or March to February) the significance of the correlations between the sea-ice and the SAO index decreases. This shows that we have to be careful when we interpret the mechanisms explaining the links between the sea-ice extent and the SAO as discussed in Section 6.3.

Other modes also describe part of the variability in the Southern Ocean. In particular, the ACW (Antarctic Circumpolar Wave) (White and Peterson, 1996) seems to explain a significant part of the regional sea-ice concentration and extent variability in the Southern Ocean in the 1980s and in the beginning of the 1990s, while the ACW could not be clearly detected either before or after this period. As a result, no index has been proposed for the ACW so far. This makes the analysis of its impact on the integrated sea-ice extent during the last few decades a difficult if not impossible task. Furthermore, the response of the sea-ice to the ACW exhibits both positive and negative anomalies. The response of the total sea-ice extent in the Southern Ocean to the ACW is thus likely to be very small and is not discussed further in this study.

This overview of the influence of the various modes of variability shows that none of these standard modes has a clear impact on the total sea-ice extent in the Southern Ocean. In this paper, we thus investigate the variability of the Antarctic sea-ice cover (integrated over the whole Southern Ocean) by looking directly at the atmospheric pressure patterns leading to a net change in the September sea-ice extent, integrated over the whole Southern Ocean. This is done using observations as well as model results. More precisely, the goal of this paper is to find out if there is a large-scale mode of atmospheric variability which governs the sea-ice extent variability integrated over the Southern Ocean or if this variability is dominated by a combination of independent processes in the different sectors of the Southern Ocean.

The model, the forcing and the methodology are briefly discussed in Section 6.2. Section 6.3 discusses in more detail the influence of the SAO on the sea-ice cover. Section 6.4 presents the results for the sea-ice extent integrated over the whole Southern Ocean, while Section 6.5 tries to explain the response of the sea-ice to the different atmospheric patterns in the various sectors of the Southern Ocean. A synthesis of the most important processes and conclusions are presented in Sections 6.6 and 6.7.

6.2 Model and method description

6.2.1 Model description

The model used here, called ORCA2-LIM, results from the coupling of the Louvain-la-Neuve sea-ice model (LIM) (Fichefet and Morales Maqueda, 1997) with the hydrostatic, primitive equation ocean model OPA (Océan PARallélisé) (Madec *et al.*, 1999). The coupling technique used to link OPA and LIM is identical to that described by Goosse and Fichefet (1999).

OPA is a finite difference ocean general circulation model with a free surface and a non-linear equation of state in the Jackett and McDougall (1995) formulation. The thermodynamic part of LIM uses a three-layer model (one layer for snow and two layers for ice) for sensible heat storage and vertical heat conduction within snow and ice. The sea-ice velocity field is determined from a momentum balance considering sea-ice as a two-dimensional continuum in dynamical interaction with the atmosphere and the ocean. The viscous-plastic constitutive law proposed by Hibler (1979) is used for computing internal stress within the ice for different states of deformation.

The coupled model is run on a global grid with 2° nominal resolution and a mesh refinement in high latitudes and near the equator. As a consequence, the meridional resolution increases to 0.5° close to the Antarctic continent, corresponding to a horizontal grid width of about 50 km. Vertical discretization uses 30 z-levels, with 10 levels in the top 100 m. Model runs are initialized using data from the end of the 46-year experiment described by Timmermann *et al.* (2005). The model is driven by daily 2-m air temperatures and 10-m winds from the NCEP/NCAR (National Centers for Environmental Prediction-National Center for Atmospheric Research) re-analysis project for the period 1979-2004 (Kalnay *et al.* (1996) and updates), and monthly climatologies of surface relative humidity (Trenberth *et al.*, 1989), cloud fraction (Berliand and Strokina, 1980), and precipitation rate (Xie and Arkin, 1996). The surface fluxes of heat and moisture are determined from these data by using the empirical parameterizations described by Goosse (1997).

The model configuration used in this paper is exactly the same as the one presented by Timmermann *et al.* (2004, 2005). By comparing model results with satellite data, Timmermann *et al.* (2004, 2005) have shown

that the model reproduces the observed evolution of the sea-ice concentration well. The comparisons between the simulated ice thickness and values derived from ULSs (Upward Looking Sonars) and the ASPeCT (Antarctic Sea Ice Processes and Climate) dataset are also very satisfactory. Furthermore, the response of the model to the most important mode of atmospheric variability is similar to the observed response as shown in Lefebvre *et al.* (2004) and in Lefebvre and Goosse (2005).

6.2.2 Methodology of the analysis and limitations

The methodology used here to analyse the variability of the ice cover consists of computing correlations and regressions of different atmospheric and sea-ice fields (ice concentration, ice production, ice velocity, ice thickness, sea level pressure, surface air temperature) on the sea-ice extent integrated over the whole Southern Ocean and over the different sectors.

In order to limit the space needed, in Table 6.3 only the regression will be given without the correlation. An idea of the correlation can be obtained by looking at the significance of the various regressions. The regressions have the advantage of showing how large the effect is, while a large correlation with a small variability can be misleading. A regression can, however, give a false view of the cause-effect relation as it does not proof a causal link between two parameters.

We do not present here results from oceanic variables because of the smaller amount of data available and because no clear and interesting additional information compared to those discussed here was found from the analysis of ocean variables in the model. This suggests that, on the time-scale investigated here, the atmosphere is the main driver of the simulated sea-ice extent variability, not the ocean. It is possible that the observed temperature or pressure changes in the atmosphere are induced by changes in the ocean but the model used in this paper is not capable of analysing this atmosphere-ocean coupling due to the absence of an interactive atmospheric component in the model. On the other hand, the response of the ocean to the atmospheric forcing is of course included in our model.

The observation data for the sea-ice concentration have been taken from the HadISST1-dataset (Rayner *et al.* (2003) and updates) which is a compilation of various sea-ice concentration datasets from 1870. Only the period between 1979 and 2004 onwards is used in this study, as this

is the period for which satellite data enables the most accurate observations of the sea-ice extent. The data for the atmosphere (i.e. 2-m air temperature, 10-m wind and surface pressure) are derived from the NCEP-NCAR re-analysis project (Kalnay *et al.* (1996) and updates). For the analysis performed here, as in Zwally *et al.* (2002) and in Lefebvre *et al.* (2004) for instance, we have divided the Southern Ocean into five sectors: the Weddell Sea (60°W-20°E), the Indian sector (20°E-90°E), the Pacific sector (90°E-160°E), the Ross Sea (160°E-140°W) and the Amundsen-Bellingshausen Seas (140°W-60°W) (Figure 6.2). This division is classical but relatively arbitrary. Fortunately, our conclusions are not dependant on the location of the boundaries between the various sectors.

The advantage of the method applied here is its simplicity; it is easy to understand the method and to interpret its results. However, we would like to draw attention to its limitations. Firstly, due to the use of static correlations, the method is not well designed to find propagating patterns such as the Antarctic Circumpolar Wave (ACW). Secondly, although our model describes the mean and the temporal variability of the sea-ice cover reasonably well, both on seasonal and longer timescales (Timmermann *et al.*, 2004, 2005), we have to take into account the fact that no model is perfect. Moreover, the forcing is mainly composed of monthly climatologies except for the near-surface temperatures and 10m-winds, which are taken daily. It was necessary to use this kind of forcing given the sometimes poor description of cloudiness and humidity in the re-analyses, due to the lack of measurements above both the Southern Ocean and the Antarctic continent. This limited data constraint has an influence of the other variables too, but to a smaller extent (Bromwich and Fogt, 2004; Timmermann *et al.*, 2005). Unfortunately, there are no better datasets available than those provided by the re-analysis agencies. Despite these limitations, model results provide complementary information to observations. They can give us a perfectly consistent data set, while observations coming from different products derived from different observations (NCEP/NCAR re-analysis and the HadISST dataset) could display some incompatibilities. Furthermore, model results include variables such as ice production that are not available in the observed data. In any case, whenever possible, we performed an identical analysis on model results and observed data to check whether the model was capable of reproducing the observed correlation well. If the observed and the

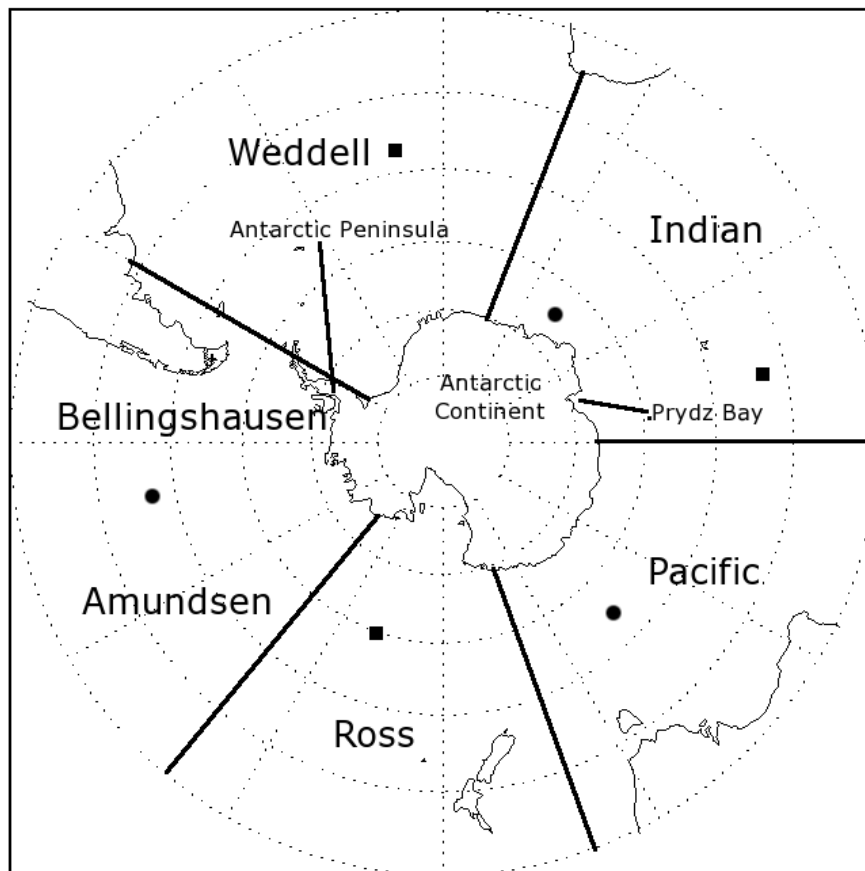


Figure 6.2: sectors of the Southern Ocean used for analysis of results plus some landmarks. The black dots and squares show the location of the ridges and troughs respectively used in Subsection 4.2.

modelled results show large differences, due to both the shortcomings in the model and in the datasets used, this will provide strong caution on the robustness of the analysis.

Because the model is forced by air temperatures and winds derived from the NCEP/NCAR re-analysis, some feedbacks existing in the real world between the sea-ice or the ocean and the atmosphere cannot be assessed. For instance, an increase in the atmospheric heat transport to one particular location and thus in air temperature there generates a decrease in the sea-ice extent which then induces a change in surface albedo amplifying the initial temperature change. The lower insulation between the ocean and the atmosphere, caused by the decrease in sea-ice concentration, could provide an additional positive feedback. Both feedbacks, which are partly responsible for the observed evolution of the system recorded in the observations, are implicitly taken into account but not explicitly in ORCALIM, as the model is forced by observed atmospheric temperatures. Using a coupled general circulation model (GCM) with an interactive atmospheric component would solve this problem but would mean that it was no longer possible to do a hindcast simulation over the last 20 years. Using direct heat and moisture fluxes derived from the re-analysis instead of the 2m temperature and the 10m wind speeds would provide several advantages for the analysis of some feedbacks. However, it is well known that these fluxes contain large errors that hamper a realistic simulation of the ice cover in the Southern Ocean.

Additionally, the period discussed in this paper covers only 26 years, i.e., the period between 1979 and 2004. We have chosen this period because of the good quality of the observations of both sea-ice and atmospheric variables, thanks to satellite data. This means that variability with timescales larger than a decade cannot be investigated here. Furthermore, this period did not show any significant trends in the sea-ice extent in the Southern Ocean. Consequently, our results cannot be extrapolated for periods with a clear trend in the sea-ice extent, as projected for the future (Houghton *et al.*, 2001; Flato and Boer, 2001; Goosse and Renssen, 2001; Flato, 2004; Arzel *et al.*, 2006; Parkinson *et al.*, 2006b; Vinnikov *et al.*, 2006; Meehl *et al.*, 2007; Lefebvre and Goosse, 2007b). It is possible, for example, that a projected mode of variability has an influence on the ice cover which depends on the time scale or that this influence would be different if there were a clear trend in the sea-ice extent. In addition, the dominant patterns of variability can exhibit a

	W(M)	I(M)	P(M)	R(M)	AB(M)	T(M)
SAM (Ann)	<u>-0.41</u>	0.03	0.16	0.16	0.34	0.14
ENSO (Ann)	0.17	-0.37	<u>0.42</u>	-0.34	-0.12	-0.05
ZW3 (Ann)	-0.34	-0.05	0.02	0.35	-0.29	0.26
SAO (J-D)	-0.28	0.21	<u>0.54</u>	-0.34	0.02	0.01
SAO (N-O)	-0.23	-0.02	0.33	0.02	0.13	0.12
SAO (M-F)	-0.31	0.07	<u>0.51</u>	-0.31	0.10	-0.01

Table 6.2: Same as Table 6.1 but now for the model (M; ORCALIM) results.

trend due to changes in the external forcing such as that related to the increase in greenhouse gas concentrations in the atmosphere (Fyfe *et al.*, 1999; Brandefelt and Källén, 2004; Marshall *et al.*, 2004).

Our study is devoted to the sea-ice extent in September, i.e. the month in which the sea ice is at its maximum. It would be possible to study sea-ice cover during other months or seasons too, but this would significantly increase the size of this paper. We have thus preferred to focus on a single month.

The variables correlated with the ice extent are generally averaged over the period for which the correlation is the highest, i.e. September, August and September combined, the combination of July, August and September or the complete ice growth season (April-September). On the other hand, the correlations in Tables 6.1 and 6.2, except for the SAO, are given for the annual mean indexes of the different modes of atmospheric variability, implicitly assuming an integrated effect of these modes on the winter ice extent. Their correlations have also been calculated for other periods. The results obtained from these analyses are similar to the ones discussed here, giving no significant correlation between the total sea-ice covered area and any of the major modes of atmospheric variability and are thus not shown here in order to improve the readability of the tables and the paper itself.

6.3 The Semiannual Oscillation

The SAO (Semiannual Oscillation) index (Figure 6.1) represents the strength of the oscillation in both surface pressure and air temperature with a period of 6 months (Van Loon, 1967). In January, the positive

phase of the SAO is associated with a high pressure anomaly on the Antarctic continent and low pressure anomalies around 45°S . This is the end of the so-called first expanded phase. In March and April, the pressure anomalies have the opposite sign to January, leading to the first contracted phase. In June and July, the anomaly pattern (Figure 6.3a) (second expanded phase) strongly resembles the one in January, whereas the September (Figure 6.3b) and October anomalies (second contracted phase) are very similar to those observed in March and April. Finally, in December, the first expanded phase comes back.

In the observations, the regression of the September sea-ice onto the SAO index is mostly negative, with positive patches only in the Pacific sector, the Bellingshausen Sea and around 30°E (Figure 6.4a). Those negative regressions explain the negative correlation between the September total sea-ice covered area and the SAO index (see Table 6.1, correlation of -0.44). However, the correlation between the sea-ice concentration and the SAO index is in any individual area not significant, and the regressions of the sea-ice concentration are rather small (almost everywhere $< 10\%$). In general, this is in agreement with the model (Figure 6.4b), with mainly non-significant negative regression values in the Weddell, Amundsen-Bellingshausen and Ross sectors. However, in the model, the regression in the Indian sector is positive, whereas in the observations, it is negative. This strongly influences the correlation between the integrated sea-ice extent and the SAO, which is very small in the model (see Table 6.2, correlation of 0.01). If we analyse the correlations by sector, there is a good agreement between the model and the observations, except for the Indian sector (Tables 6.1 and 6.2).

The mechanisms explaining how the sea-ice concentration responds to the SAO appear relatively straightforward when analysing the pressure changes. The negative values in the Weddell sector in July are due to northerly wind anomalies in June and July (Figure 6.3a) bringing warmer air to this sector and pushing the sea-ice to the south. The same effect occurs in the Ross sector with a northeasterly wind anomaly. The June-July easterly anomaly pushes the ice out of the Ross sector into the eastern Pacific sector. Furthermore, an inflow of colder air in June and July in the Bellingshausen sector increases the sea-ice concentration there in July. These sea-ice concentration anomalies all persist until September, inducing the pattern displayed in Figure 6.4. In the western part of the Pacific sector cold southerly wind anomalies in August and

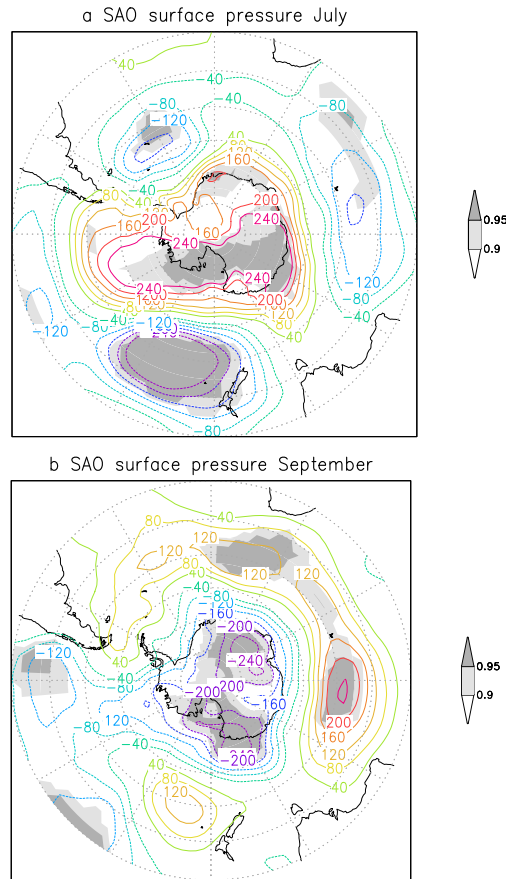


Figure 6.3: The regression of the observed sea surface pressure (NCEP-NCAR, in Pa, 1979-2004) onto the SAO index. The surface pressure has been taken for the mean of (a) July (end of second expanded phase) and (b) September (start of second contracted phase). The significance level is shown in gray shading. The significance has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

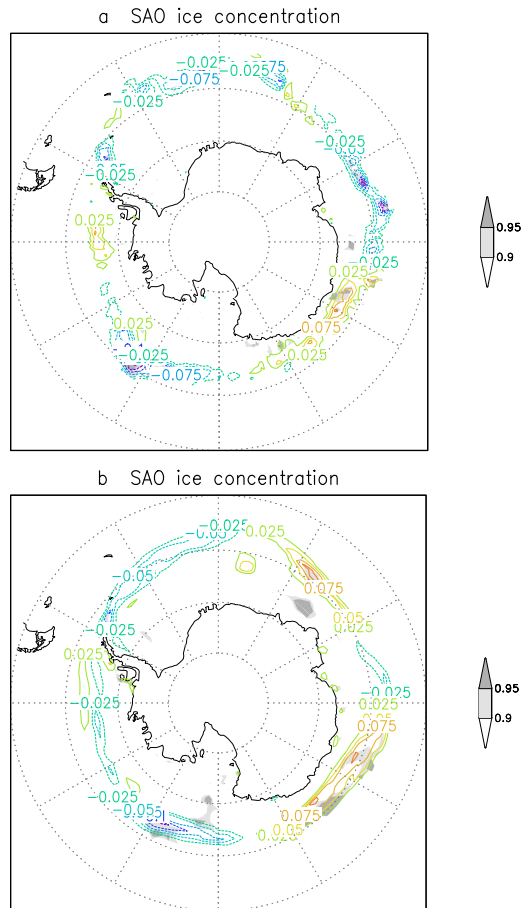


Figure 6.4: The regression of (a) the observed (HadISST, 1979-2004) and (b) the modeled (ORCALIM, 1979-2004) September sea-ice concentration onto the SAO index (NCEP-NCAR). The significance level is shown in gray shading. The significance has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

September (Figure 6.3b) push the ice to the north creating a larger sea-ice extent. The observed sea-ice concentration in the Indian sector seems to be a result of warm northerly wind anomalies in August, whereas in the model the import of colder air in September seems to be predominant.

In a previous modeling study using a different ice-ocean model (called CLIO), Fichefet *et al.* (2003b) linked the large decrease in the sea-ice extent integrated over the Southern Ocean in the seventies obtained in their simulation to the decrease in the SAO index occurring at the same time. This would imply a positive correlation between the SAO index and the ice extent during this event while our results for the last 25 years suggest a negative correlation. Therefore, we have done the same analysis as in Figure 9 of Fichefet *et al.* (2003b), using ORCALIM results, obtaining a very similar pattern. Nevertheless, this pattern is clearly different than the one in our Figure 6.4. In addition, the regression of the sea-ice concentration to the SAO for the period 1948-1978 (not shown), shows a large difference in the Pacific and Ross sectors compared to the pattern in ORCALIM (Figure 6.4). This indicates that the effect of the SAO on the total sea-ice extent in the model is different when analysing various periods. It is impossible to do the same analysis over the period 1948-1978 using the observed sea-ice extent due to the lack of reliable observational sea-ice data. However, it would be interesting to further analyse the stationarity of the response of the sea ice to the SAO over different periods and in different models. However, this lies outside the scope of this paper.

6.4 September sea-ice extent variability integrated over the Southern Ocean

6.4.1 Sea-ice concentration

The regression of the observed September sea-ice concentration onto the observed total sea-ice extent in September for the period 1979-2004 (Figure 6.5a, Table 6.3) shows a ring-like pattern located close to the climatological September sea-ice edge. As expected, most of the regressions are positive, because a larger sea-ice concentration around the sea-ice edge results in a larger sea-ice extent. The only exceptions are slight negative values in the Amundsen-Bellingshausen Seas (75°W - 130°W) and in the western Indian sector (20°E - 40°E). The regression of the modeled

	W	I	P	R	AB
W	<u>0.39</u>	-0.06	0.07	-0.05	<u>-0.13</u>
I	-0.12	<u>0.20</u>	-0.05	-0.02	0.02
P	0.12	-0.04	<u>0.24</u>	0.05	-0.01
R	-0.07	-0.02	0.05	<u>0.27</u>	-0.05
AB	<u>-0.19</u>	0.02	-0.01	-0.05	<u>0.27</u>
T	0.09	0.02	0.08	<u>0.14</u>	0.03

Table 6.3: The winter sea-ice extent (in $10^6 km^2$) regressed onto the winter sea-ice extent sector per sector in the HadISST1-observations (Rayner *et al.* (2003) and updates). On each row, the data are given regressed onto the winter sea-ice extent of the sector at the beginning of the row. The sea-ice data are split up, sector per sector (columns). Each entry shows the regression of the sea-ice extent of its sector (column) onto the sea-ice extent of another sector (row). W, I, P, R, AB and T denote the Weddell sector, the Indian sector, the Pacific sector, the Ross sector, the Amundsen-Bellinghshausen Seas and the Total Southern Ocean respectively.

Significances are shown in **bold** when they exceed the 90% threshold, while significances exceeding the 95% threshold are shown underlined. Numbers that are both **underlined and bold**, are significant above the 99% threshold. The significance was tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the table.

September sea-ice concentration onto the total modeled sea-ice extent in September for the same period shows a similar ring-like pattern (Figure 6.6a).

In both model results and observations, the variability of the total sea-ice covered area seems to be dominated by changes in the Weddell, Pacific and Ross sectors. For instance, the regressions of the modeled September ice extent in these sectors onto the modeled integrated September sea-ice extent are 0.15, 0.13 and 0.14 million km^2 respectively.

6.4.2 Sea level pressure

The regression of the observed atmospheric pressure onto the observed total sea-ice extent in September displays three large areas of low pressure anomalies, one over the south Atlantic, one over the Indian Ocean

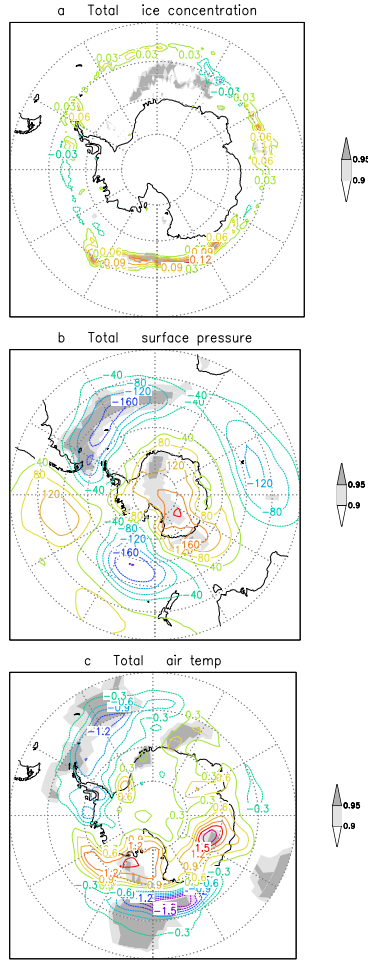


Figure 6.5: The regression of (a) the September sea-ice concentration (HadISST, 1979-2004), (b) the surface pressure (NCEP-NCAR, in Pa, 1979-2004, averaged over August and September) and (c) the air temperature (NCEP-NCAR, in $^{\circ}\text{C}$, 1979-2004, averaged over August and September) onto the September sea-ice extent integrated over the whole Southern Ocean (HadISST). The significance level is shown in gray shading. The significance has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

(around 45°S, 75°E), and one on the southwestern part of the Pacific Ocean (around 58°S, 160°W) (Figure 6.5b). Above the Antarctic continent, a high pressure anomaly can be found, extending towards the north to the southeastern Pacific Ocean (center of the extension around 48°S, 100°W). A smaller extension of this high pressure anomaly can be found south of Australia (around 140°E). This pattern is close to a wave 3 pattern but is not symmetric.

A similar pattern can be found in the regression of the NCEP/NCAR surface pressure onto the modeled sea-ice extent (Figure 6.6b), although some differences are apparent. First of all, the low pressure anomaly in the Atlantic is slightly further east (around 10°W). Secondly, the small high pressure anomaly that could be seen in the Figure 6.5b northeast of New Zealand (35°S, 160°W) is stronger in Figure 6.6b and is connected to the high pressure anomaly over the Pacific (two cores, one around 47°S, 140°E and the other around 60°S, 120°E). Consequently, the low pressure anomaly in the southwestern part of the Pacific Ocean is found in Figure 6.6b closer to the continent into the eastern Ross Sea (around 62°S, 160°E). Finally, the high pressure anomaly in the southeastern Pacific Ocean is similar using both the observed and modeled ice extents, although it is weaker in the latter case. Using the modeled sea-ice extent, the dominant atmospheric patterns are clearly significant while on Figure 6.5b only the low pressure anomaly in the Weddell Sea and the high pressure anomaly on the continent are significant. The modelled sea ice seems thus to be more responsive to the pressure pattern than the observed sea ice. This might be related to shortcomings in the model and in the experimental design. On the other hand, the model and the forcing provide a completely consistent dataset while the observations incorporate different noise, which are probably largely uncorrelated. This can lead to smaller correlations than in the model.

In order to compare this pattern with the other modes of variability considered in Tables 6.1 and 6.2, we calculated an index for our pattern named the PPI (Pressure Pattern associated with the Ice extent variation) index, as follows:

$$PPI\ index = \sum_{i=1}^3 P'_i - \sum_{j=1}^3 P'_j \quad (6.1)$$

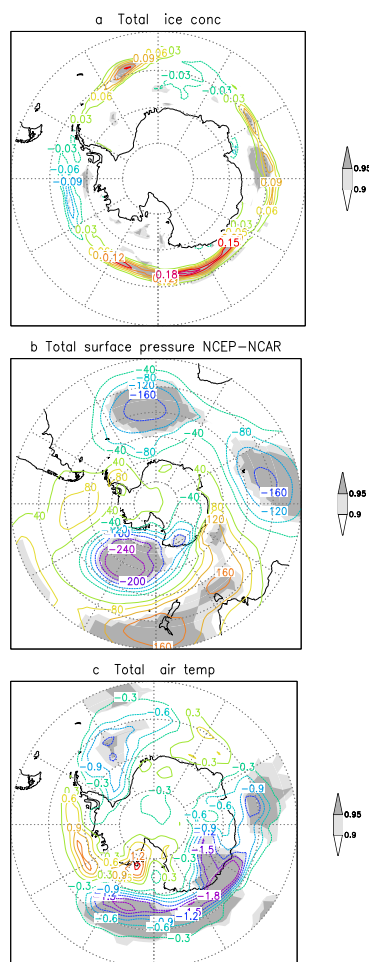


Figure 6.6: The regression of (a) the September sea-ice concentration (ORCALIM, 1979-2004), (b) the surface pressure (NCEP-NCAR, in Pa, 1979-2004, averaged over August and September) and (c) the air temperature (NCEP-NCAR, in $^{\circ}C$, 1979-2004, averaged over August and September) onto the September sea-ice extent integrated over the whole Southern Ocean (ORCALIM). The significance is shown in gray shading. The significance level has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

where $P'_{i,j}$ is the pressure anomaly (i.e. the surface pressure minus the mean surface pressure) averaged over July, August and September at point i or j . The points i are chosen so that they correspond to the ridges in Figures 6.5b and 6.6b. They are located in the Bellingshausen Sea (47.5°S , 102.5°W), in the western Indian sector (65°S , 40°E) and in the Pacific sector (55°S , 135°E). The points j , corresponding to the troughs, can be found in the Weddell sector (47.5°S , 10°W), in the eastern Indian sector (42.5°S , 77.5°E) and in the Ross Sea (60°S , 160°W) (Figure 6.2).

When computed over July, August and September, the resulting PPI index has a correlation of 0.50 with the observed and 0.68 with the modeled September sea-ice extent integrated over the Southern Ocean. The first is significant at the 95% level, the second one even at the 99% level. As expected, both values are higher than the correlations of the other indexes (SAM, ENSO, etc.) with the integrated sea-ice extent (Tables 6.1 and 6.2). However, even with these high correlations, our index still only explains 25% of the observed and 46% of the modeled sea-ice extent. Significant correlations between the PPI index and the local sea-ice extent can be found in the Weddell sector (significant at 90% for the model), the Indian sector (significant at 95% for the model) and the Ross sector (significant at 99% for both the model and the observations). The PPI-index taken over September alone gives similar results as the one averaged over July, August and September. Actually, the correlation between the September PPI-index and the observed sea-ice extent is barely higher (0.53), while the correlation with the model drops significantly (0.51). Further on, we will only use the PPI-index averaged over July, August and September.

The pattern described in Figures 6.5b and 6.6b should not be confused with the quasi-stationary wave 3 pattern first discussed by Van Loon and Jenne (1972), for which Raphael (2004) has constructed an index (ZW3). Their wave has its maximum amplitude between 25°S and 60°S in all months (Van Loon and Jenne, 1972), whereas our pattern has its maximum further south. Furthermore, there is no significant correlation between the winter sea-ice extent integrated over the whole Southern Ocean and the index proposed by Raphael (2004) (see Table 1) and there is no correlation (0.03) between the PPI index defined here and the ZW3 index of Raphael (2004). Despite this absence of a significant correlation between the ZW3-index and the sea-ice extent integrated over the South-

ern Ocean, Raphael (2007) has recently shown a significant influence of the ZW3 on the sea-ice extent in some regions.

On the contrary, the correlation between our index and the SAM index is significant at 90% (-0.36). This can be explained by the fact that the pressure pattern shown in Figure 6.5b exhibits a high pressure anomaly over the continent and a low pressure anomaly further north. However, this pattern is not a negative SAM pattern, because of the non-annularity and the high-pressure anomaly at about 40°S in the Pacific sector. Finally, the correlation between our index and the SAO index is not significant (-0.09).

6.4.3 Atmospheric temperature, ice production, ice thickness and ice transport

The pressure anomalies induce changes in the winds and thus in the atmospheric heat advection. This strongly influences the interannual variability of the air temperature. As a consequence, the regression of the observed air temperature (August-September) onto the observed sea-ice extent (Figure 6.5c) and the modeled sea-ice extent (Figure 6.6c) exhibits the following pattern. Around the Antarctic Peninsula, cold southerly wind anomalies induce a large negative temperature anomaly in the Weddell sector (west of 0°E) in both the model and observations which is later advected by the mean westerly winds prevailing in the Weddell sector around 60°S. Southeast of the large negative pressure anomaly in the Atlantic Ocean, a positive temperature anomaly is advected in the eastern Weddell sector (east of 0°E). On the contrary, the low pressure anomaly in the Ross Sea advects warmer air over the Amundsen-Bellingshausen Seas (except against the Peninsula) and the southern Ross Sea (around 70°S), whereas the northern Ross Sea (close to the ice edge) and the eastern Pacific sector (east of 135°E) exhibit a cold anomaly due to the same pressure anomaly. The only difference between the temperature regression on the modeled and observational sea-ice extent lies in the western Pacific sector (90°E-135°E), where there is a cold anomaly in the regression with the modeled sea-ice extent and a warm anomaly in the one with the observational sea-ice extent. This is probably due to the different form of the high pressure anomaly south of Australia. In Figure 6.5b, this high pressure anomaly extends towards the Indian sector,

keeping the heat away from the sea-ice, whereas in Figure 6.6b, the heat is transported from lower latitudes towards the western Pacific sector.

In addition, the pattern of sea level pressure discussed above induces rather straightforward changes in the ice production, ice thickness and ice transport. Southerly (northerly) wind anomalies induce a higher sea-ice divergence (convergence) rate close to the continent and a northward (southward) anomalous sea-ice transport (Figure 6.7c). This can be most clearly seen in the Weddell, Ross and Amundsen-Bellingshausen Seas. As a result, this leads to a larger (smaller) sea-ice production close to the continent (Figure 6.7a) and a larger (smaller) melting of sea-ice at the ice edge. As a consequence of this anomalous transport, the ice thickness increases (decreases) at the ice edge (Figure 6.7b).

6.4.4 The relationship between the different centers of action in the surface pressure pattern

Following these analyses, an important question that can be raised is: are there links between the different centers of action in the surface pressure pattern responsible for the variability in the ice extent? In other words, is this new pattern a new mode of variability that should be added to the big family of modes important for the Southern Ocean (SAM, ENSO, ZW3, SAO, etc.)? There are only two significant ($> 95\%$) correlations between the pressure in the different centers of action (defined in Subsection 6.4.2 and on Figure 6.2 as the points used to calculate the index). The first one is a positive correlation (0.59) between the pressure in the high pressure anomaly in the Bellingshausen Sea (47.5°S , 102.5°W) and that in the high pressure anomaly in the Pacific sector (55°S , 135°E). The other is a negative correlation (-0.50) between the pressure in the high pressure anomaly in the western Indian sector (65°S , 40°E) and the low pressure anomaly in the eastern Indian sector (42.5°S , 77.5°E). The other correlations between the different low and high pressure anomalies are not significant. As a consequence, we can conclude that the pattern shown here is not a mode of variability for the sea surface pressure. It rather seems to be a combination of different, more or less local, independent patterns, driving sea-ice changes in the various sectors of the Southern Ocean, but with no important teleconnections between them.

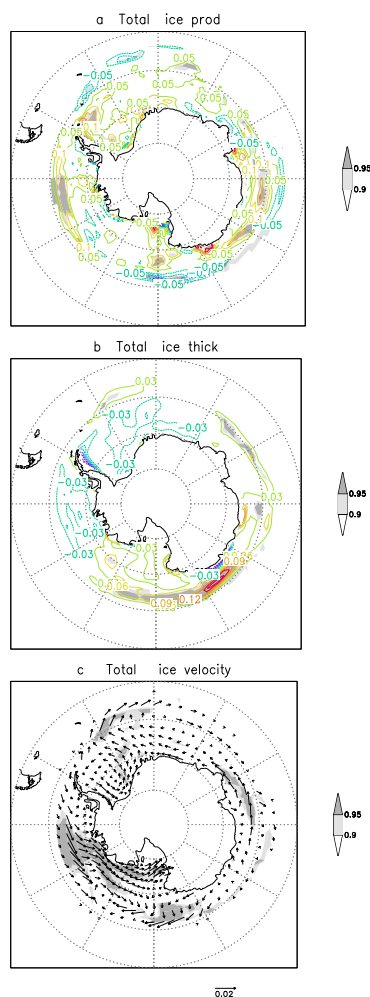


Figure 6.7: The regression of (a) the sea-ice production (ORCALIM, in m, 1979-2004, integrated over the growth season) (b) the ice thickness (ORCALIM, in m, 1979-2004, September) and (c) the ice velocity (ORCALIM, in $\frac{m}{s}$, 1979-2004, averaged over the growth season, i.e. April to September) onto the September sea-ice extent integrated over the whole Southern Ocean (ORCALIM). The significance is shown in gray shading. The significance level has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

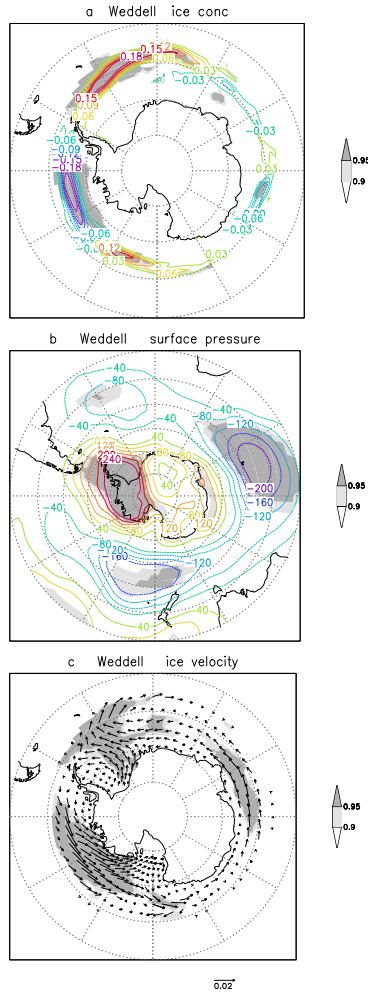
6.5 Regional variability

The discussion above suggests that it is not possible to find a single large-scale atmospheric mode that has a strong influence on the total sea-ice extent in the Southern Ocean. Atmospheric processes seem to have a clearer impact at a regional scale. This is the reason why the causes of regional sea-ice extent variability in the different sectors are analyzed in this section. We will focus here on three sectors, i.e., firstly, the two sectors with the largest sea-ice extent variability (Weddell and Ross Seas), because they are representative of the behavior of the majority of the sea-ice. In addition, the Amundsen-Bellingshausen Seas will be presented at the end of the section because it has a clearly different response from the other sectors.

6.5.1 The Weddell Sea

When making the same type of analysis as in Section 4 for the Weddell sector instead of the entire Southern Ocean, we can clearly see a negative correlation between the Weddell and the Amundsen-Bellingshausen Seas ice extent (Figure 6.8a). Furthermore, in the model, the Weddell and the Ross Sea ice extents are positively correlated, whereas this effect is not found in the observations. If we look at the pressure pattern that is most favorable for a maximum of sea-ice in the Weddell sector (Figure 6.8b), it shows roughly the same pattern as the pressure pattern deduced from the sea-ice extent integrated over the Southern Ocean (Figure 6.5b). The three cores of low pressure can be seen, with only a shift to the north for the one in the Ross sector. Furthermore, the high pressure anomaly above the Bellingshausen Sea and the Antarctic Peninsula is stronger, while the high pressure anomaly over the Pacific sector has almost disappeared. This pattern is associated with a weaker heat advection in the Weddell Sea and a stronger one in the Amundsen-Bellingshausen Seas contributing to the changing sea-ice concentrations in these seas.

The changes in the pressure pattern also have an influence on the wind stress and thus on the meridional sea-ice velocity. Southerly (northerly) wind anomalies induce a higher sea-ice divergence (convergence) rate close to the continent and a northward (southward) anomalous sea-ice transport. For instance, when the September sea-ice extent is high in the Weddell sector, the low pressure anomaly (Figure 6.8b) there induces



a southerly wind anomaly that pushes the ice to the north. The regression of the Weddell Sea northward sea-ice transport integrated over the ice growth season (April-September) at 65°S onto the September Weddell sea-ice extent amounts to 0.34 million km^2 (see also Figure 6.8c and Auxiliary Materials). This generates thicker ice away from the continent. On the contrary, a thinner ice layer is formed close to the continent. In addition, when the September Weddell Sea ice extent is higher than average, the Amundsen-Bellingshausen sea-ice is pushed against the continent due to northerly wind anomalies resulting in a lower ice extent there (regression of the northward transport onto the September Weddell sea-ice extent during the period April-September of -0.54 million km^2 at 65°S , see Figure 6.8c).

Finally, the positive anomalies of sea-ice extent in the Weddell Sea are highly correlated with both an increased northward transport (see above) and an import anomaly of sea ice out of the Indian sector into the Weddell sector. This import corresponds to a regression of 0.08 million km^2 during the period April to September.

6.5.2 The Ross Sea

The analysis for the Ross sector shows a decrease in ice concentration in the Amundsen-Bellingshausen Seas and an increase in both Ross sector and Weddell sector in the model (Figure 6.9a). The most important changes in the sea-ice cover are due to the large low pressure anomaly in the eastern Ross Sea (Figure 6.9b) and the high pressure anomaly extending from the Antarctic continent onto the Antarctic Peninsula and into the Pacific sector. This low pressure anomaly induces important transport processes (Figure 6.9c) with a regression of a northward sea-ice transport during the period April to September reaching 0.39 million km^2 at 65°S . In the Amundsen-Bellingshausen Seas, the meridional sea-ice transport anomaly is southward (southward transport anomaly during the period April to September of 0.54 million km^2 at 65°S), when the sea-ice extent is high in the Ross Sea. Furthermore, changes in the zonal sea-ice velocity induce an increase in the total sea-ice extent due to the existence of the Ross Gyre. Sea-ice brought by zonal transport out of both the Amundsen Sea and the Pacific sector is pushed northward by the Ross Gyre increasing the Ross sea-ice extent. As a result, the regression between the total sea-ice extent and the zonal import of sea-

ice to the Ross Sea amounts to 0.19 million km^2 during the period April to September.

6.5.3 The Amundsen-Bellingshausen Seas

For the Amundsen-Bellingshausen Seas, the pressure pattern is the opposite of the patterns discussed above: i.e. high pressure anomalies in the Indian and Ross sectors and a low pressure anomaly over the Antarctic Peninsula with a small extension over the Antarctic continent (Figure 6.10b). This pattern is also responsible for an increase in the Amundsen-Bellingshausen Seas sea-ice extent and a decrease of the sea-ice extent in both the Weddell and the Ross sectors (Figure 6.10a).

When the sea-ice extent in the Amundsen-Bellingshausen Seas reaches its maximum, the ice transport (Figure 6.10c) is the opposite of the one discussed above when both the Weddell and the Ross sectors have high position sea-ice anomalies (Figures 6.8c and 6.9c). An increased northward sea-ice transport (Figure 6.10c, regression of 0.54 million km^2 during the period April to September at $65^\circ S$) provides a larger sea-ice extent in the Amundsen-Bellingshausen Seas (Figure 6.10a). On the contrary, in the Weddell sector this pattern is associated with a southward ice transport anomaly of 0.35 million km^2 at $65^\circ S$ during the period April to September (Figure 6.10c) and, in the Ross sector, with a southward (0.18 million km^2 during the period April to September at $65^\circ S$, Figure 6.10c) ice transport anomaly which tends to diminish the local sea-ice extent. However, as the sea-ice extent variability in the Weddell and Ross sectors is far larger than the variability in the Amundsen-Bellingshausen Seas, the pattern associated with the variability in the ice extent integrated over the whole Southern Ocean (Figures 6.5b and 6.6b) resembles the one for the Weddell and the Ross sectors (Figures 6.8b and 6.9b) far more than the one for the Amundsen-Bellingshausen Seas (Figure 6.10b).

6.6 Synthesis of the dominant processes

In summary, we have shown that a static pattern is obtained when the surface pressure is regressed onto the total sea-ice extent. It is associated with a low pressure anomaly above the southern Atlantic Ocean and a high pressure anomaly above the Antarctic continent, the western Indian sector and the Bellingshausen Sea that pushes cold air towards

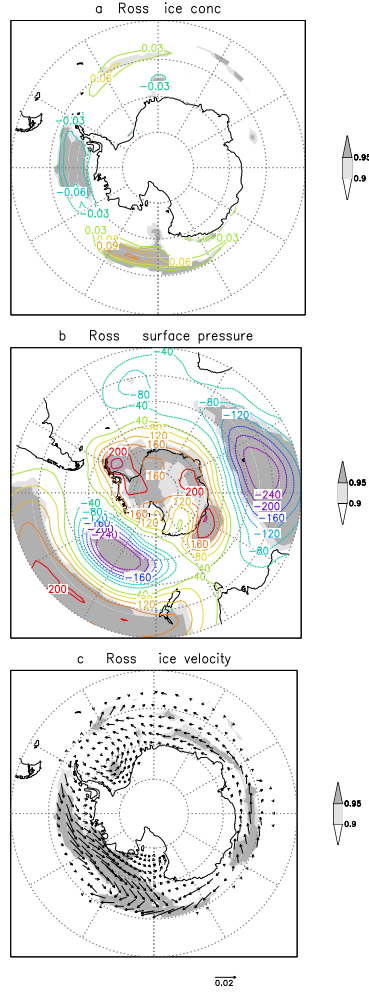


Figure 6.9: The regression of (a) the September sea-ice concentration (ORCALIM, 1979-2004) (b) the surface pressure (NCEP-NCAR, in Pa, 1979-2004, averaged over August and September) and (c) the ice velocity (ORCALIM, in $\frac{m}{s}$, 1979-2004, averaged over the growth season, i.e. April to September) onto the September sea-ice extent integrated over the Ross sector (ORCALIM). The significance is shown in gray shading. The significance level has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

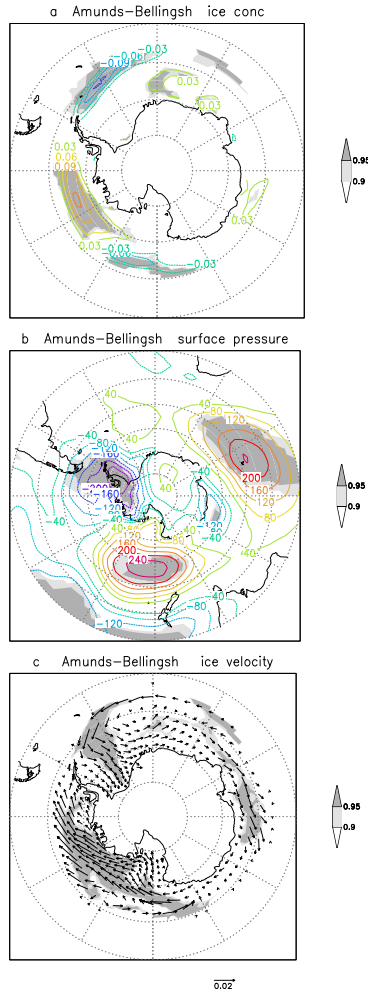


Figure 6.10: The regression of (a) the September sea-ice concentration (ORCALIM, 1979-2004) (b) the surface pressure (NCEP-NCAR, in Pa, 1979-2004, averaged over August and September) and (c) the ice velocity (ORCALIM, in $\frac{m}{s}$, 1979-2004, averaged over the growth season, i.e. April to September) onto the September sea-ice extent integrated over the Amundsen-Bellingshausen Seas (ORCALIM). The significance is shown in gray shading. The significance level has been tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the corresponding point.

the north in the Weddell sector resulting in more ice in this sector. By contrast, this pressure anomaly warms up the western Indian sector (between 20°E and 40°E) and the Amundsen-Bellingshausen Seas. The low pressure anomaly also induces a northward sea-ice transport anomaly in the central Weddell Sea and an import of sea-ice from the Indian sector towards the Weddell sector. All these processes result in more ice in most of the Weddell sector while the ice concentration tends to be lower at the boundary between the Weddell sector and the Indian sector.

The low pressure anomaly southwest of Australia, together with the high pressure anomaly above the Antarctic continent and the high pressure anomalies in the Indian and the eastern Pacific sectors, cool down the region around Prydz Bay (60°E - 90°E), and help warm up the western Indian sector.

The low pressure anomalies in the Ross Sea together with the high pressure anomalies in the Amundsen-Bellingshausen Seas, the western Pacific sector and around north New Zealand, push warm air into the Amundsen Sea and the southern Ross sector. On the other hand, the northern Ross sector, close to the winter ice edge, is cooled, as is the eastern Pacific sector. These pressure anomalies also induce some ice velocity anomalies: a southward velocity anomaly in the Amundsen-Bellingshausen Seas, then an export of sea-ice from the Amundsen-Bellingshausen Seas towards the Ross sector, an export out of the Pacific sector into the Ross Sea and an export towards the north in the Ross Sea. This leads to a smaller ice extent in the Amundsen-Bellingshausen Seas (with a higher production) and a greater ice extent in the Ross Sea.

We have described here the influence of the atmosphere on the sea-ice extent variability and described the mechanisms responsible for that influence. Besides, sea-ice changes could have feedbacks on the atmosphere and eventually reinforce, destroy or modify the atmosphere pattern that created them. Unfortunately, these feedbacks could not be analysed in the present framework.

6.7 Conclusions

From observations and a coupled sea-ice ocean model driven by daily NCEP-NCAR re-analysis, we have shown that the sea-ice extent integrated over the whole Southern Ocean is not significantly correlated

with the classical modes of surface pressure variability in the atmosphere (SAM, ENSO, etc.), except with the SAO. However, locally and in the different sectors of the Southern Ocean individually, the correlations between the SAO and the sea-ice extent are small. Furthermore, the link between the SAO and the sea-ice extent appears not stationary when investigating different periods. So, more data are required to investigate more precisely the robustness of the link between the SAO and the ice extent.

The atmospheric wave 3 pattern that is associated with the largest sea-ice extent is a non-symmetric pattern, that induces a shifted pattern in the air temperature. One dominant characteristic of this pattern is the occurrence of a low pressure anomaly at about 60°S in both the Weddell and the Ross Seas. These low pressure anomalies have a clear impact locally on the ice extent in these two seas. As these two sectors have the largest variance, the patterns that have a strong influence there should also be present in the regression with the ice extent integrated over all the sectors. However, using available observations, it is not possible to state that these regional variations are linked. The pressure of the two regions is only weakly correlated. Furthermore, the correlation between the sea-ice extent in the Ross and Weddell sectors is weak (even negative in the observations). We could thus conclude that our pattern is not a mode of variability on its own, but a combination of regional variations that are not strongly correlated.

Our results show that the variability of the sea-ice extent integrated over the Southern Ocean can not be considered at this stage as being caused by one single large-scale process. It must rather be analyzed as the sum of sea-ice extent variabilities in the different sectors, which are influenced by different mechanisms. This is in agreement with the findings presented by Zwally *et al.* (2002). They showed that the observed overall monthly variability of the sea-ice extent in the Southern Ocean is comparable in magnitude to the monthly variability in the individual sectors. However, the overall relative extent variability is smaller than the sector values. They attributed this to the spatial averaging over positive and negative anomalies that tend to balance each other out among the sectors. This concurs with our findings that the correlation of the sea-ice extent variability between the different sectors is small with some exceptions: the correlation of the September sea-ice extent in the Amundsen-Bellingshausen sector with both the Weddell (in both the

model and the observations) and the Ross sectors (only in the model) is significant (at more than 95%) and negative.

In this work, only the interannual variability has been covered and not the variability on longer (interdecadal) or shorter (intraannual) time-scales. Furthermore, the analysis has been performed for a period in which no significant trend has been found for the total sea-ice covered area. As a result, care has to be taken when extrapolating these results to either the past or the future, when a different behavior of the system might be present. In particular, our results show that reconstruction of the integrated sea-ice extent using only a few or even one regional proxies e.g. (Curran *et al.*, 2003) for the sea-ice extent calibrated using observations performed during the last few decades should be performed with great caution. Furthermore, the mechanisms presented here have only been tested on a short period. Assuming long-term stability of the links between the sea-ice extent variability and various regions in a system like the Southern Ocean where numerous modes of variability could play a role is a strong and probably unwarranted conjecture.

In the results described above, we have considered the atmosphere as the only driver of the variability in the sea-ice extent. However, in general, other factors should also be considered. For low frequency changes, the oceanic heat transport variations, even carrying heat from the Northern Hemisphere (e.g. Goosse *et al.*, 2004), could be the driving factor for the large scale sea-ice extent variability. Another example of a possible oceanic influence relates to the Antarctic Coastal Wave that has been described by Beckmann and Timmermann (2001). According to these authors, this wave is responsible for a significant part of the variability of the sea-ice thickness and the bottom water formation in the Weddell sector.

Despite these limitations, we have analyzed, using the available information, the physical processes governing the variability of the ice extent during the last 25 years. The atmospheric mechanisms that have been proposed have the advantage of being relatively clear. They could constitute a test for analyses of past and future changes, i.e. to determine if similar mechanisms could be useful in explaining the sea-ice changes on those time scales.

Acknowledgments

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Auxiliary Materials

See next page.

	W		I		P		R		AB	
W	10	<u>0.35</u>	-2	0.01	3	-0.04	-5	<u>0.12</u>	7	<u>-0.26</u>
	<u>0.34</u>	0.03	-0.10	0.06	0.03	-0.08	0.07	<u>0.21</u>	<u>-0.54</u>	<u>-0.22</u>
I	<u>-21</u>	0.01	7	<u>0.18</u>	5	0.00	5	0.05	-3	-0.07
	0.07	0.04	-0.07	0.01	<u>0.12</u>	-0.13	0.00	0.04	-0.15	0.04
P	8	-0.05	2	0.00	<u>10</u>	<u>0.26</u>	-1	-0.03	-3	0.03
	0.03	-0.05	0.02	-0.08	0.11	<u>0.12</u>	-0.06	0.01	-0.11	0.00
R	-8	<u>0.18</u>	0	0.04	<u>10</u>	-0.03	-1	<u>0.24</u>	4	<u>-0.18</u>
	0.11	<u>0.11</u>	-0.08	-0.03	-0.08	<u>-0.17</u>	<u>0.39</u>	<u>0.25</u>	<u>-0.54</u>	-0.16
AB	-9	<u>-0.28</u>	2	-0.04	-1	0.02	3	<u>-0.14</u>	-1	<u>0.33</u>
	<u>-0.35</u>	0.00	0.09	<u>-0.08</u>	-0.05	0.06	-0.18	<u>-0.22</u>	<u>0.54</u>	<u>0.24</u>
T	-7	<u>0.15</u>	4	<u>0.07</u>	<u>16</u>	<u>0.13</u>	0	<u>0.14</u>	4	-0.08
	0.13	<u>0.08</u>	-0.07	-0.07	0.06	<u>-0.12</u>	0.13	<u>0.19</u>	<u>-0.49</u>	-0.08
	ice production	ice extent								
	northward transport	ice import								

Table 6.4: The sea-ice extent, production and different ice fluxes regressed onto the winter sea-ice extent, in the ORCALIM model. On each row, the data are regressed onto the winter sea-ice extent of the sector at the beginning of the row. The regressions are given for each sector (columns). W, I, P, R, AB and T denote the Weddell sector, the Indian sector, the Pacific sector, the Ross sector, the Amundsen-Bellingshausen Seas and the total Southern Ocean, respectively.

In each case there is shown:

- top left: the regression with the ice production (in $10^{10}m^3$) integrated over the ice growth season (April to September).
- bottom left: the regression with the northward sea-ice transport integrated over the ice growth season (April to September) (in 10^6km^2) at $65^\circ S$.
- top right: the regression with the ice extent (in 10^6km^2) in September.
- bottom right: the regression with the net zonal ice transport at the boundaries of the sector (import of ice is positive) integrated over the ice growth season (April to September) (in 10^6km^2).

A sketch showing the location of the various terms is included at the bottom left of the table.

Significances are shown in **bold** when they exceed the 90% threshold, while significances exceeding the 95% threshold are shown underlined. Numbers that are both **underlined and bold**, are significant above the 99% threshold. The significance was tested using a Monte-Carlo method with 10,000 simulations of autocorrelated data series with the same autocorrelation as the time series presented in the table.

Chapter 7

Analysis of the projected regional sea ice changes in the Southern Ocean during the twenty-first century

Analysis of the projected regional sea ice changes in the Southern Ocean during the twenty-first century

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Abstract

Using the set of simulations performed with atmosphere-ocean general circulation models (AOGCMs) for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR4), the projected regional distribution of sea ice for the 21st century has been investigated. Averaged over all those model simulations, the current climate is reasonably well reproduced. However, this averaging procedure hides the errors from individual models. Over the 20th century, the multimodel average simulates a larger sea ice concentration decrease around the Antarctic Peninsula compared to other regions, which is in qualitative agreement with observations. This is likely related to the positive trend in the Southern Annular Mode (SAM) index over the 20th century, in both observations and in the multimodel average. Despite the simulated positive future trend in SAM, such a regional feature around the Antarctic Peninsula is absent in the projected sea ice change for the end of the 21st century. The maximum decrease is indeed located over the central Weddell Sea and the Amundsen-Bellingshausen Seas. In most models, changes in the oceanic currents could play a role in the regional distribution of the sea ice, especially in the Ross Sea, where stronger southward currents could be responsible for a smaller sea ice decrease during the 21st century. Finally, changes in the mixed layer depth can be found in some models, inducing locally strong changes in the sea ice concentration.

Keywords: Climate change, Sea ice, Southern Ocean, Mechanisms, Regional patterns

7.1 Introduction

During the end of the 20th century, the sea ice trend has shown a distinct asymmetry between the Northern and the Southern Hemispheres. In the Northern Hemisphere, the sea ice extent exhibits a large decrease. In contrast, there is no significant trend in the sea ice extent of the Southern Hemisphere (Cavalieri *et al.*, 2003). This absence of a downward trend in the sea ice in the Southern Hemisphere during the last 25 years has classically been attributed to the slow response of the Southern Ocean to the radiative forcing (Manabe *et al.*, 1991, 1992), caused by a large heat uptake by the ocean at high southern latitudes. However, recent studies

show that the response of the Southern Ocean is more complex than the one of a motionless system with a large heat capacity corresponding to a 3000m deep ocean covering nearly all longitudes. In particular, Flato and Boer (2001) found that, in their model, an improved parameterization of the effect of meso-scale eddies in the ocean results in a reduced heat penetration into the Southern Ocean. This diminishes the asymmetry of the warming between Northern and Southern Hemispheres compared to the one obtained by Manabe *et al.* (1991) and Manabe *et al.* (1992) but does not eliminate it. Goosse and Renssen (2001) proposed a detailed scenario for the response of the Southern Ocean to the changes in radiative forcing due to the increase in greenhouse gas concentrations. First of all, the ocean damps the surface warming because of its large heat capacity. Secondly, the warming in the Southern Ocean is amplified by an increased oceanic meridional heat transport. This occurs with a lag of about one century after the major increase of greenhouse gases. Recently, Bitz *et al.* (2006) have shown that in the CCSM3 model, the deep convection weakens, most importantly in the vicinity of 55°-70°S¹, in response to an increasing CO_2 concentration in the atmosphere. This is largely due to a surface freshening caused by modified sea ice and freshwater transports. As a result, the Southern Ocean warms at depth, influencing the ocean heat uptake in the model. Furthermore, the upward transport of heat in the ocean around Antarctica is reduced, which tends to damp the response of the sea ice extent to the radiative perturbation.

In any case, the projections for the future show clearly a strong decline in the sea ice extent, in all seasons, in both hemispheres (Houghton *et al.*, 2001; Flato and Boer, 2001; Goosse and Renssen, 2001; Flato, 2004; Vinnikov *et al.*, 2006; Parkinson *et al.*, 2006b; Arzel *et al.*, 2006). For instance, Arzel *et al.* (2006) documents a September sea ice extent decrease of about 20% in the Southern Hemisphere between 1981-2000 and 2081-2100 using a multimodel ensemble of different General Circulation Models (GCMs) simulations submitted for the IPCC AR4 (Fourth Assessment Report of the Intergovernmental Panel on Climate Change) in the SRES (Special Report on Emission Scenarios) A1B scenario. In their study, Arzel *et al.* (2006) also showed that, despite the fact that the multimodel average extent is in good agreement with the observa-

¹The text 'most importantly in the vicinity of 55°-70°S' is absent in the published version of this paper.

tions (Figure 7.1), only a few models were able to reproduce reasonably well the winter and summer sea ice extents in the Southern Hemisphere. On the other hand, Parkinson *et al.* (2006b,a) who studied the seasonal cycle of the modelled sea ice in eleven major models concluded that the phase of the annual cycle corresponds well to the observations. Vinnikov *et al.* (2006) compared also the different models and the observations in the period 1973-2004. They pointed out that the standard deviation of the detrended annual mean sea ice extent is much larger in the Southern Hemisphere than in the Northern Hemisphere, which is well reproduced by the investigated models. However, the observed ice extent variability is to the low end of the modelled variability. Furthermore, they proved that the sea ice trend during the investigated period is only reasonably well reproduced.

All those previous studies focused on the Southern Ocean as a whole, which is a natural first step. However, large variations have been observed in the various sectors of the Southern Ocean during the last decades. In particular, the trend in the sea ice extent integrated over the Southern Ocean for the period 1979-2004 is almost zero, while a decline of sea ice extent in the Amundsen-Bellingshausen Seas and an increase in the Ross Sea have been noticed (Zwally *et al.*, 2002). This suggests that a detailed analysis of the regional changes simulated by the GCMs during both the 20th and 21st centuries could bring complementary information compared to the one provided by studies devoted to the ice extent integrated over the whole Southern Ocean. Furthermore, it would allow determining if one could expect a different evolution in the various sectors of the Southern Ocean or a more or less annular response to increasing greenhouse gas concentrations.

Here, after a brief description of the datasets used (Section 7.2), we will analyse the regional characteristics of the sea ice change at the end of the 20th century and during the 21st century simulated by different GCMs (Table 7.1, Section 7.3). Section 7.4 discusses the potential mechanisms responsible for local differences in the projected sea ice changes. This is an important part of the present work, because previous studies analysing changes in the Southern Ocean ice extent (Arzel *et al.*, 2006; Vinnikov *et al.*, 2006; Parkinson *et al.*, 2006b, e.g.) were mainly descriptive. Finally, conclusions are presented in Section 7.5. In our analysis, the Southern Ocean is generally divided in 5 sectors as, for instance, in Zwally *et al.* (2002) and in Lefebvre *et al.* (2004): the Weddell Sea

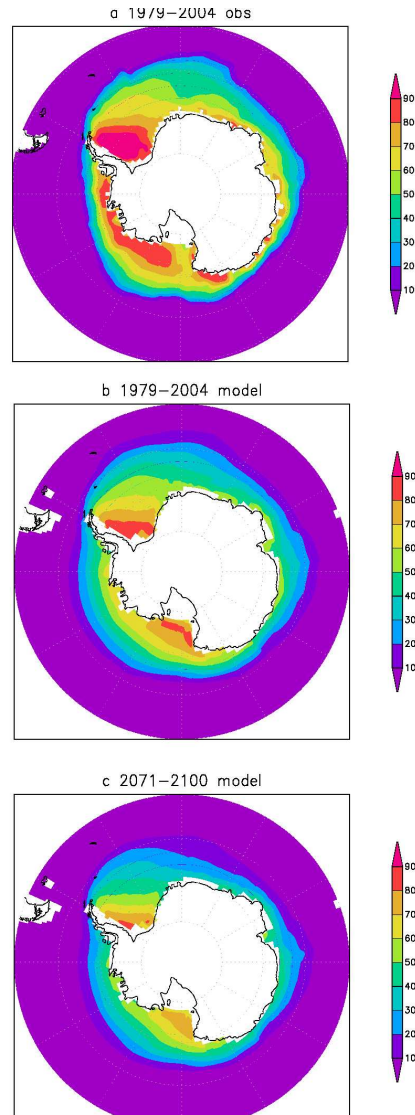


Figure 7.1: (a) The observed annual mean sea ice concentration (HadISST, Rayner *et al.* (2003) and updates) over the period 1979-2004 (in %). (b) The multimodel averaged annual mean sea ice concentration (in %) over the period 1979-2004 in the 20C3M-runs and (c) 2071-2100 in the SRES A1B-runs of the 16 discussed GCMs.

(60°W-20°E), the Indian sector (20°E-90°E), the Pacific sector (90°E-160°E), the Ross Sea (160°E-140°W) and the Amundsen-Bellingshausen Seas (140°W-60°W) (Figure 7.2).

7.2 Datasets

The data used here are derived from the set of simulations performed with atmosphere-ocean general circulation models (AOGCMs) for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR4). Our analysis is based on model outputs from the "Climate of the 20th Century Experiment" (20C3M) and from the scenario SRES (Special Report on Emission Scenarios) A1B defined by the Special Report on Emission Scenarios (Nakicenovic and Swart, 2000). The A1B scenario corresponds to a continuous increase of the CO_2 concentration over the 21st century until a level of 720 ppm by 2100. We have chosen this run because it presents a mid-range scenario which has been discussed previously in the literature (for instance in Arzel *et al.*, 2006). To be able to compare model results with the data observed at the beginning of the 21st century, we have added the first four years of the SRES A1B-runs to the 20C3M-runs.

When writing this paper, 17 AOGCMs (IPSL-CM4, CNRM-CM3, GISS-AOM, GISS-ER, FGOALS-g1.0, CSIRO-Mk3.0, INM-CM3.0, UKMO-HadGEM1, UKMO-HadCM3, MRI-CGCM2.3.2, MIROC3.2 (hires), MIROC3.2 (medres), CGCM3.1 (T47), CGCM3.1 (T63), CCSM3, ECHAM5/MPI-OM, ECHO-G) had complete sea ice outputs for the 20C3M and the SRES A1B (21st century) runs available on the data distribution center website (see Table 7.1). As in Arzel *et al.* (2006), the FGOALS-g1.0 model has been omitted because of an unrealistic sea ice extent in both hemispheres in both summer and winter. Detailed information about the different models can be found at the website <http://www-pcmdi.llnl.gov/>.

The observation data for the sea ice concentration have been taken from the HadISST1-dataset (Rayner *et al.* (2003) and updates) which is a compilation of various sea ice concentration datasets from 1870 until now. Only the period between 1979 and 2004 will be used in this study, as this is the period for which the satellite data enable the most accurate observations of the sea ice extent.

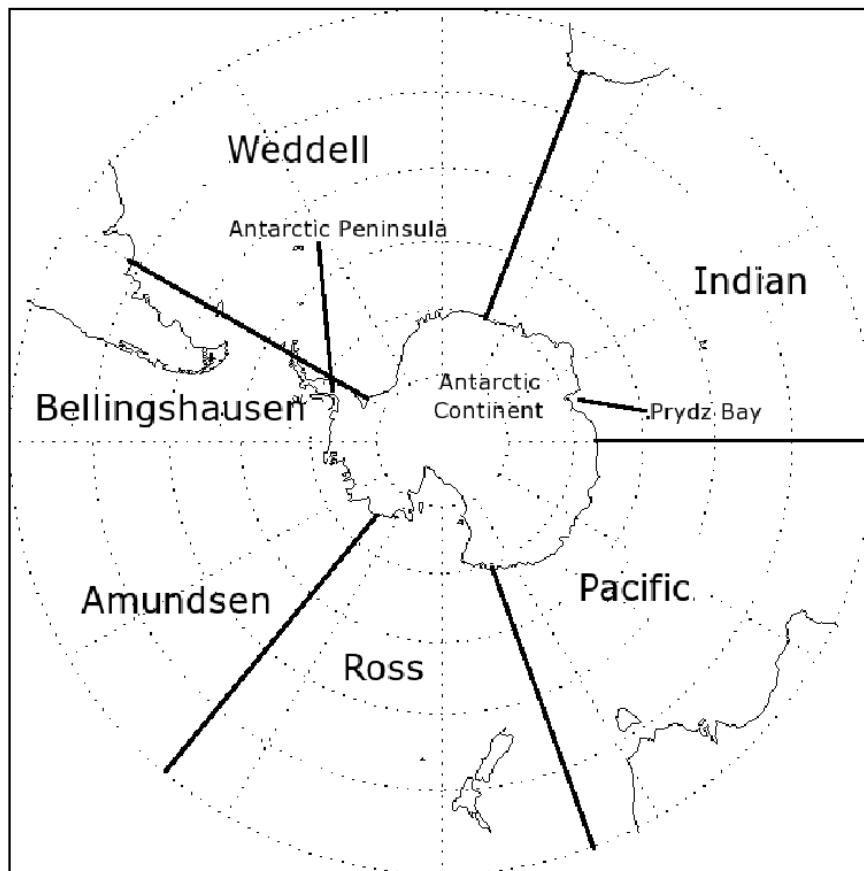


Figure 7.2: sectors of the Southern Ocean used for analysis of the results. Also, some landmarks have been shown.

We will investigate here primarily the annual mean sea ice concentration. However, caution has to be made when averaging the sea ice concentration, as the sea ice concentration is zero in summer at many grid points. As a consequence, changes of sea ice in regions where sea ice is only present in winter will appear to be small, while in reality, they can be quite large during a small part of the year. This has to be taken into account when analysing the results. Nevertheless, as the same averaging technique has been applied to the observations and the model results, the comparison between them should not pose any major problems for our discussion.

7.3 Sea ice patterns

7.3.1 The end of the 20th century

As discussed in previous studies (Arzel *et al.*, 2006; Vinnikov *et al.*, 2006; Parkinson *et al.*, 2006b), the annual mean sea ice extent varies a lot between the different models (Figure 7.3). The observed annual mean sea ice extent averaged over the period 1979-2004 (12.44 million km^2 , Table 7.2) is close to the mean of all the models which equals to 12.26 million km^2 (Figures 7.1 and 7.3, Table 7.2). However, the values for each individual model range from a minimum under the 8 million km^2 (MIROC3.2 (medres)) to a maximum above 18 million km^2 (CGCM3.1 (T63)) (Figure 7.3). The annual mean sea ice extent averaged over all the models is also in relatively good agreement with the observed one for the various sectors of the Southern Ocean (Figure 7.3, Table 7.2). This results in a reasonable position of the ice edge for the model ensemble mean (Figure 7.1). Nevertheless, this averaging procedure hides errors from individual models. In general, errors for the various sectors are larger than the errors on the ice extent integrated over the whole Southern Ocean. For instance, the sea ice extent in the Pacific sector (Figure 7.3c) is very small in both the GISS-ER model and the ECHO-G model. The latter model has the same problem in the Amundsen-Bellingshausen Seas (Figure 7.3e) as it is only reproducing a reasonable amount of sea ice in the Weddell, the Ross and the western Indian sectors. The MIROC3.2 (medres) model has its largest problems in the Weddell Sea (Figure 7.3a) simulating a too low sea ice extent there. In the Ross sector, both the IPSL-CM4 and the CNRM-CM3 model underestimate the sea ice ex-

Name	Country	Institution
GISS-ER	USA	NASA - Goddard Institute for Space Studies
CGCM3.1 (T47)	Canada	Canadian Centre for Climate Modelling and Analysis
CCSM3	USA	National Center for Atmospheric Research
ECHAM5/MPI-OM	Germany	Max Planck Institute for Meteorology
MIROC3.2 (medres)	Japan	{ Center for Climate System Research (The University of Tokyo), National
MIROC3.2 (hires)		{ Institute for Environmental Studies and Frontier Research for Global Change
UKMO-HadCM3	UK	Hadley Centre for Climate Prediction and Research / Met Office
UKMO-HadGEM1	UK	Hadley Centre for Climate Prediction and Research / Met Office
CSIRO-Mk3.0	Australia	CSIRO Atmospheric Research
CGCM3.1 (T63)	Canada	Canadian Centre for Climate Modelling and Analysis
IPSL-CM4	France	Institut Pierre Simon Laplace
CNRM-CM3	France	Météo-France / Centre National de Recherches Météorologiques
GISS-AOM	USA	NASA - Goddard Institute for Space Studies
MRI-CGCM2.3.2	Japan	Meteorological Research Institute
INM-CM3.0	Russia	Institute for Numerical Mathematics
ECHO-G	{ Germany Korea	Meteorological Institute (Univ of Bonn), Meteorological Res. Inst. of KMA, Model & Data Group

Table 7.1: The 16 AOGCMs from the IPCC AR4 used in this paper.

	Extent 20C3M ($10^6 km^2$)	Extent A1B ($10^6 km^2$)	Trend 20C3M ($\frac{10^6 km^2}{10yr}$)	Trend A1B ($\frac{10^6 km^2}{10yr}$)	Var 20C3M ($10^6 km^2$)	Var A1B ($10^6 km^2$)
W	4.17 (4.52)	3.05	-0.08 (0.01)	-0.12	0.29 (0.28)	0.22
I	1.92 (2.03)	1.34	-0.09 (0.04)	-0.04	0.16 (0.11)	0.12
P	1.41 (1.32)	1.11	-0.03 (0.07)	-0.04	0.13 (0.11)	0.12
R	2.53 (2.57)	2.16	0.00 (0.18)	-0.04	0.19 (0.16)	0.17
AB	2.24 (2.00)	1.59	-0.07 (-0.09)	-0.06	0.20 (0.15)	0.16
T	12.26 (12.44)	9.25	-0.27 (0.21)	-0.30	0.56 (0.21)	0.46

Table 7.2: Summary of some key values. Extent denotes the annual mean sea ice extent ($C > 15\%$) averaged over the 16 models while Trend denotes the trend of the annual mean sea ice extent averaged over the 16 models. In the Var column, the standard deviations of the detrended annual mean sea ice extent averaged over the 16 models are shown. The 20C3M columns give the results over the period 1979-2004 in the 20C3M-runs, while the A1B columns give the results over the last 30 years of the 21st century in the SRES A1B-runs. The rows correspond to the different sectors (W=Weddell; I=Indian; P=Pacific; R=Ross; AB=Amundsen-Bellingshausen; T=Total). In brackets, the observed values (HadISST) can be found, if applicable.

tent. Other problems are a too low ice concentration in the Indian sector (Figure 7.3b) in both the IPSL-CM4 and the GISS-AOM model and a large excess of sea ice in the INM-CM3.0 Amundsen-Bellingshausen Seas (Figure 7.3e).

The observed sea ice extent displays a small positive trend for the period 1979-2004 (0.21 ± 0.12 million km^2 per decade, see Figures 7.4 and 7.5a, Table 7.2). However, in several models, a decreasing Southern Ocean sea ice extent is observed for the period 1979-2004, with a mean trend of -0.27 ± 0.30 million km^2 per decade (Table 7.2). In Figure 7.4, error bars denote two standard deviations of the trend, which has been calculated in a standard way (Snedecor and Cochran, 1971, e.g.). However, this number accounts only for the internal variability within one run for the investigated period. If an ensemble of several runs of the same model were provided (5 complete runs were available from the GISS-ER, the CGCM3.1 (T47) and the MRI-CGCM2.3.2 model, 3 from the ECHO-G and the MIROC3.2 (medres) model and 2 from the GISS-AOM model), only one run has been taken here. For GISS-ER,

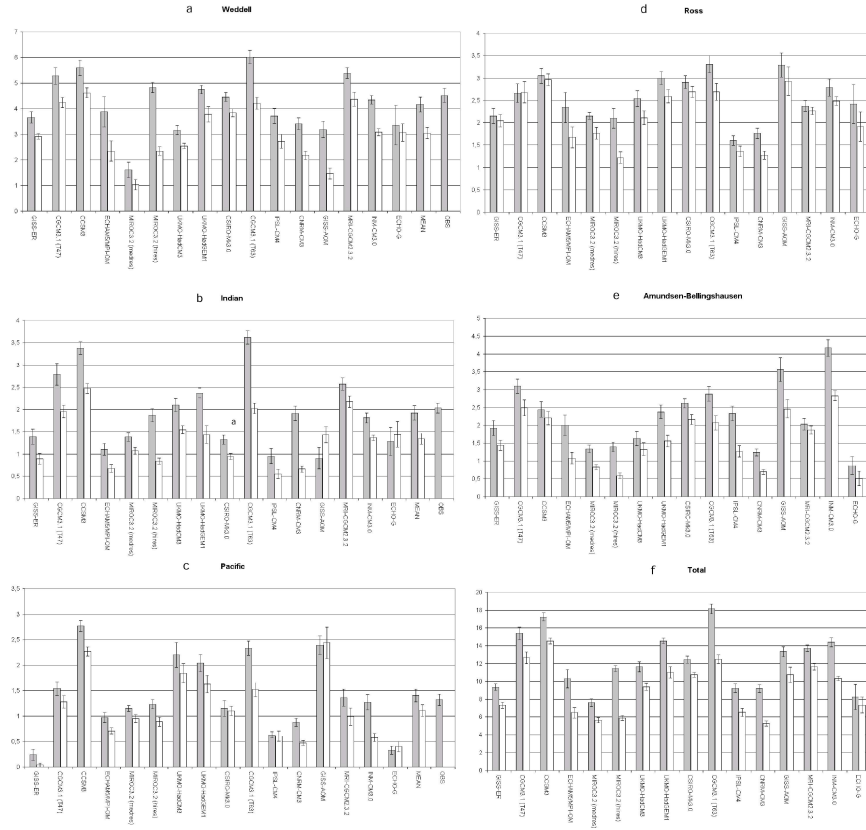


Figure 7.3: The annual mean sea ice extent (bars) in the Southern Hemisphere (SH) in the different GCMs (in $10^6 km^2$) in (a) the Weddell (top left), (b) Indian (middle left), (c) Pacific (bottom left), (d) Ross (top right), (e) Amundsen-Bellingshausen sectors (middle right) and (f) the entire Southern Ocean (bottom right). The mean has respectively been taken over the period 1979-2004 (20C3M run, gray) and on the last 30 years of the 21st century in the SRES A1B run (white). MEAN denotes the annual mean sea ice extent averaged over the 16 models. OBS gives the annual mean sea ice extent in the HadISST (Rayner *et al.*, 2003) dataset. The error bars denote the standard deviations of the detrended annual mean sea ice extent over the same period.

CGCM3.1 (T47), ECHO-G and the GISS-AOM model, the differences between the members of the ensemble are smaller or equal than the errors due to the internal variability, so, our estimates provide a reasonable value for the uncertainties in the trend. However, this is not valid for the modelled sea ice extent trend at the end of the 20th century for both the MIROC3.2 (medres) and MRI-CGCM2.3.2 models. It is important that we take this element into account when analyzing our results as we thus underestimate for those models the uncertainty of the trend by just using one standard deviation of one run. No similar test could be done for the other models.

The decreasing trend simulated by the models is probably² due to the radiative forcing induced by the increased amount of greenhouse gases. The model trend is particularly important in the ECHAM5/MPI-OM, the MIROC3.2 (hires) and the CGCM3.1 (T63) models with a loss in sea ice of more than a half million km^2 per decade. On the other hand, the trend is close to zero in seven other models. The largest positive sea ice extent trend is observed in the Ross sector (Figures 7.4 and 7.5a). For the multimodel mean, the most positive trend is also simulated in that region, which, in this case, means no negative trends (0.00 ± 0.10 million km^2 per decade, 0.0% per decade, Figures 7.4 and 7.5b, Table 7.2). In the Indian sector, the mean model sea ice extent trend is negative (-0.09 ± 0.08 million km^2 per decade, -4.8% per decade, Table 7.2), while the observed trend is slightly, but not significantly positive (Figure 7.4). In the Pacific sector, the observed trend is significantly positive, while the model trend is not significant (model trend of -0.03 ± 0.04 million km^2 per decade, -1.9% per decade, Table 7.2). In the Weddell sector, neither the observed nor the multimodel averaged trends are significant (multimodel averaged trend of -0.08 ± 0.16 million km^2 per decade, -1.9% per decade, Table 7.2). In the Amundsen-Bellinghshausen Seas both the observed and modelled trend are negative (albeit insignificant for the multimodel average: -0.07 ± 0.10 million km^2 per decade, -3.1% per decade, Table 7.2). In general, the model trends are too negative but their regional distribution is rather well reproduced (Figure 7.5). However, the trend shown in the individual models (Figure 7.4) is in general less accurate than the multimodel averaged trend.

²The word probably is absent in the published version of this paper.

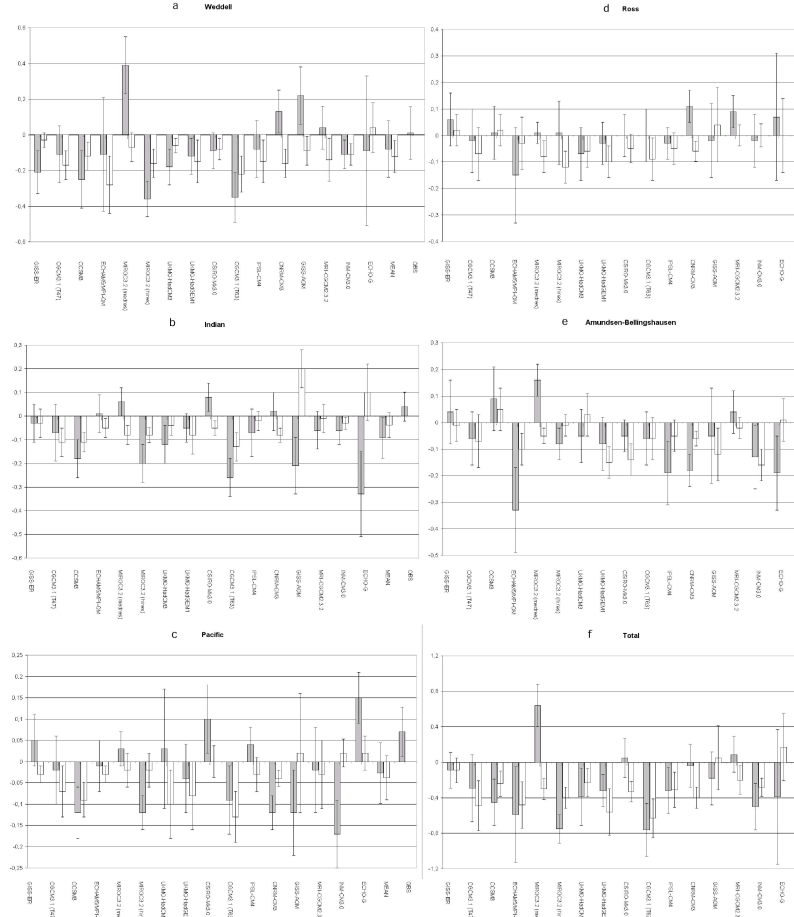


Figure 7.4: The trend of the annual mean sea ice extent in the different GCMs (in $\frac{10^6 \text{ km}^2}{10 \text{ yr}}$) in (a) the Weddell (top left), (b) Indian (middle left), (c) Pacific (bottom left), (d) Ross (top right), (e) Amundsen-Bellinghousen sectors (middle right) and (f) the entire Southern Ocean (bottom right). The trend has respectively been calculated over the period 1979-2004 (20C3M run, gray) and on the last 30 years of the 21st century in the SRES A1B run (white). MEAN denotes the annual mean sea ice extent trend averaged over the 16 models. OBS gives the annual mean sea ice extent trend in the HadISST (Rayner *et al.*, 2003) dataset. The error bars denote two standard deviations of this trend.

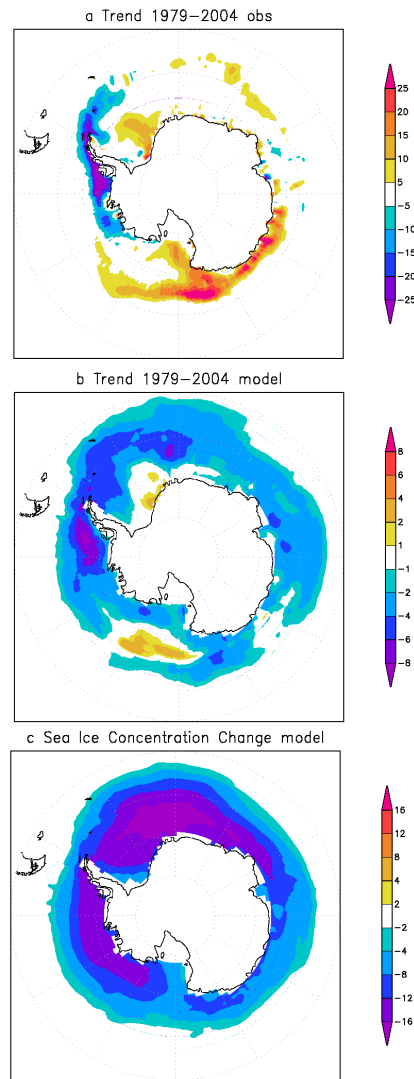


Figure 7.5: (a) The observed (HadISST, Rayner *et al.* (2003) and updates) annual mean sea ice concentration trend (in %) integrated over the period 1979-2004. (b) The multimodel averaged sea ice concentration trend (in %) integrated over the period 1979-2004 in the 20C3M-runs (c) The difference in sea ice concentration between the 2071-2100 period in the SRES A1B-runs and the 1979-2004 period in the 20C3M-runs averaged of the 16 discussed AOGCMs (in %). Be careful of the different scales being used in the three figures.

A third important parameter is the simulated interannual variability measured here by the standard deviation of the detrended Southern Ocean sea ice extent over the period 1979-2004 (error bars in Figure 7.3). The AOGCMs display a reasonable variability in all the sectors (albeit always a little bit too high, Table 7.2). However, the multimodel averaged Southern Hemisphere sea ice extent variability is almost three times higher than the observed variability (0.21 million km^2 , 1.68%, Table 7.2) in the same period (Figure 7.3). Only two models produce a reasonable total variability for the period 1979-2004 (MIROC3.2 (hires) and UKMO-HadGEM1) although their variability is still 1.5 times the observed one. The observed detrended interannual variability for the period 1979-2004 of the Southern Ocean sea ice extent is relatively small compared to the value obtained for the individual sectors (Zwally *et al.*, 2002). This is partly due to the negative correlation between the observed sea ice extents in the different sectors during the period 1979-2004. Indeed, positive anomalies in the Weddell or Ross sectors are generally associated with a negative anomaly in the Amundsen-Bellingshausen Seas (Yuan and Martinson, 2000, 2001; Liu *et al.*, 2002a; Lefebvre and Goosse, 2007a). This Antarctic Dipole (seesaw between the sea ice concentration in the Weddell and the Amundsen sectors) is well represented in some GCMs at least (Holland *et al.*, 2005, as described for CCSM3). As the variability for each sector is good in the CCSM3 model, but the total sea ice variability is too large, we can conclude that the negative correlation between the different sectors is still not strong enough in this model with respect to the observations. These results concur with the ones of Arzel *et al.* (2006), Vinnikov *et al.* (2006) and Parkinson *et al.* (2006b). However, we have extended those papers by looking at the individual sectors.

7.3.2 SRES A1B

The annual mean sea ice extent decreases between the end of the 20th and the end of the 21st (in the projections of the SRES A1B scenario) century in all the models (Figures 7.1 and 7.3). The smallest absolute decrease is in ECHO-G (0.93 million km^2 , 11%), the largest in CGCM3.1 (T63) (5.65 million km^2 , 31%). The smallest relative decrease is also in ECHO-G, but the largest can be found in the MIROC3.2 (hires) model (5.55 million km^2 , 49%). The average of all the models shows a decrease

of 3.02 million km^2 which accounts for a loss of 25% of the 20th century ice extent (Table 7.2). In general, the simulated trends in the annual mean sea ice extent for the period 1979-2004 are similar to the ones projected at the end of the 21st century in the SRES A1B run. This is consistent with the results of Arzel *et al.* (2006). The multimodel average of the decrease in sea ice extent between the end of the 20th and the end of the 21st century is quite annular (Figure 7.1). However, almost none of the individual models shows such an annular decrease.

In general, the largest absolute decrease is found in the Weddell sector (mean decrease of somewhat more than 1 million km^2 , 27%, Table 7.2), and the smallest in the Pacific sector (mean decrease of 0.30 million km^2 , 21%, Table 7.2). This is logical as these two sectors displayed respectively the largest and smallest mean sea ice extent to begin with (Figures 7.1 and 7.3, Table 7.2). Besides, the smallest relative decrease is in the Ross sector (mean decrease of 0.37 million km^2 , 14%, Table 7.2), while the largest can be found in the Indian sector (mean decrease of 0.58 million km^2 , 30%, Table 7.2) and the Amundsen-Bellingshausen Seas (mean decrease of 0.65 million km^2 , 29%, Table 7.2). The interannual variability of the sea ice extent in the 21st century is also projected to decrease as a result of a decreased annual sea ice extent. As could be expected, the variation decrease is the largest in the Weddell sector, where the mean ice extent diminishes the most. On the contrary, the Pacific sector exhibits almost no decrease of interannual variability in the AOGCMs (Figure 7.3).

7.4 Mechanisms

While the previous section contains the description of the changes in the sea ice extent in the different models, the goal of this section is to analyse the mechanisms responsible for those changes. Indeed, despite quantitative differences between the model responses, some processes which are common for the majority of the models could provide useful information about the physics governing sea ice. These processes can be divided into three categories. First of all, an important role is played by the local conditions at the end of the 20th century, which will be discussed in subsection 7.4.1. Secondly, the response of the sea ice to the major atmospheric changes will be addressed in subsection 7.4.2. These

changes are mainly captured by the ones related to a stronger Southern Annular Mode (SAM) (Fyfe *et al.*, 1999; Thompson *et al.*, 2000; Marshall *et al.*, 2004; Arblaster and Meehl, 2006; Fyfe and Saenko, 2006). Finally, subsections 7.4.3 and 7.4.4 will discuss the oceanic-induced response.

7.4.1 Different responses of the sea ice due to local conditions at the end of the 20th century

Even if the temperature would rise equally over all areas, the changes in sea ice concentration would be locally different. Obviously, no ice will melt during the 21st century in regions where no ice was present at the end of the 20th century in a particular model. Small changes in sea ice concentration are also expected in regions whose temperature was well under the melting point for most of the year, since the risen temperature will remain very low. The location where there is, at the end of the 20th century, sea ice at temperatures only slightly below the melting point will thus be affected the most. In general, these locations can be found in a latitude band around 65°S. Northward of this band, there is only sea ice in winter which limits the possible decrease of the annual mean sea ice concentration. Furthermore, northward of 65°S, some models have almost no sea ice, which induces a smaller decrease in the multimodel average. Southward, the temperatures are much colder throughout the year, limiting the possible sea ice melt. As a result, the difference between the annual mean ice concentration for the late 20th and the late 21st centuries, averaged over all the models, reaches its maximum at around 65°S (Figure 7.5c). Because of the shape of the Antarctic continent, this is associated with large changes in the Weddell and Bellingshausen Seas. Southward of 70°S, in the southern Weddell and Ross Seas, the simulated changes in ice concentration are small at the end of the 21st century for scenario A1B in most models. Those regions remain cold as the southerly winds tend to advect there cold air from the continent. Low changes are also simulated off the eastern coast of the Antarctic Peninsula, due to ice transported from the southern Weddell Sea and in the central Ross Sea because of the ice transported out of the southern Ross Sea.

7.4.2 The Southern Annular Mode

Here, the definition of Gong and Wang (1999) has been chosen for the index of the SAM (Southern Annular Mode): the difference between the zonal mean sea level pressure at 40°S and the zonal mean sea level pressure at 65°S. The multimodel averaged SAM pattern (calculated by projecting the sea surface pressure onto the SAM-index) is very close to the observational pattern (Figures 7.6a and 7.6b). Indeed, in addition to the annular component of SAM most of the models (CGCM3.1 (T47), CGCM3.1 (T63), ECHAM5/MPI-OM, MIROC3.2 (hires), UKMO-HadCM3, UKMO-HadGEM, CSIRO-Mk3.0, CNRM-CM3, MRI-CGCM2.2.2, INM-CM3.0 and PCM) have the largest decrease in sea level pressure in the Ross and Amundsen Seas, which is consistent with the observations (Mo, 2000; Lefebvre *et al.*, 2004; Liu *et al.*, 2004). Furthermore, the multimodel average of the trend in sea level pressure seen at the end of the 20th century experiment has a distinct SAM signature. Such signature is a merely zonal pattern with the largest change being a sharp decrease (> 1.4 hPa) of the pressure in the Amundsen-Bellingshausen Seas, creating southerly wind anomalies in the Ross Sea (Figure 7.6c). This is in good agreement with the observed positive trend of the SAM index over the last 26 years and with the observed pressure trend (Figure 7.6d) seen in this period, although the multimodel averaged pressure trend (Figure 7.6c) is weaker and more SAM-like than the observed trend. It is important to note that both (multimodel average and observations) give an increased southerly wind anomaly in the Ross Sea. However, some of the models do not show this trend. Especially, in the IPSL-CM4 model, the end of the 20th century trend is a negative SAM pattern, and quite a lot of models (e.g. GISS-ER and CCSM3) have the most important pressure changes outside of the Amundsen and Ross Seas.

At the end of the 21st century, the projected SAM pattern (Figure 7.6f) resembles very much the one of the 20th century (Figure 7.6a), which is a merely zonal pressure distribution disturbed by a more negative pressure regression in the Amundsen and Ross Seas. The most important changes in the pattern are a somewhat more contracted pattern and an even deeper low pressure regression in the Amundsen and Ross Seas. As the SAM-index increases between the ends of the 20th and the 21st century (Figure 7.7, see also Fyfe *et al.* (1999); Marshall *et al.* (2004); Arblaster and Meehl (2006); Fyfe and Saenko (2006)), we

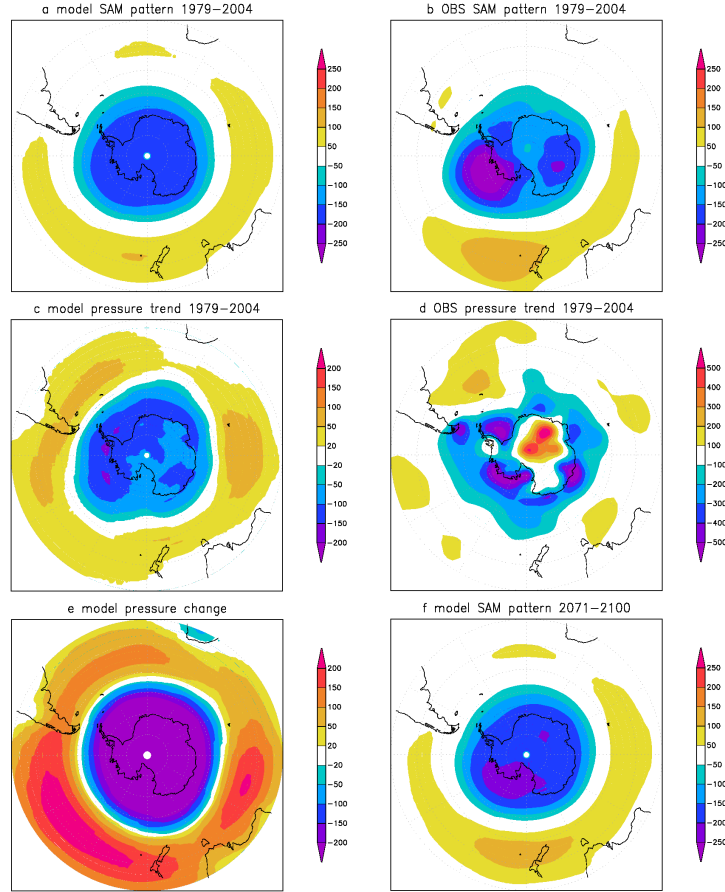


Figure 7.6: (a) The annual mean detrended SAM pattern (in Pa) calculated over the period 1979–2004 in the 20C3M-runs averaged over the 16 discussed GCMs. (b) The observed annual mean detrended SAM pattern (NCEP/NCAR, in Pa) calculated over the period 1979–2004. (c) The multimodel averaged annual mean sea level pressure trend (in Pa) integrated over the period 1979–2004 in the 20C3M-runs. (d) The observed annual mean sea level pressure trend (NCEP/NCAR, in Pa) integrated over the period 1979–2004. (e) The difference in annual mean sea level pressure between the 2071–2100 in the SRES A1B-runs period and the 1979–2004 in the 20C3M-runs period averaged over the 16 discussed AOGCMs (in Pa). (f) The annual mean detrended SAM pattern (in Pa) calculated over the period 2071–2100 in the A1B-runs averaged over the 16 discussed GCMs.

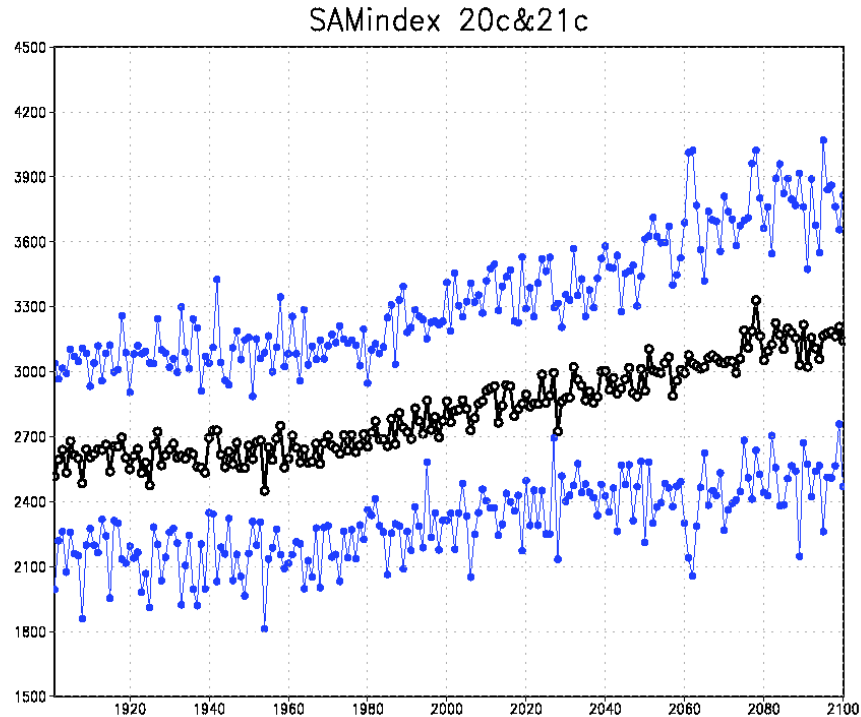


Figure 7.7: The multimodel averaged annual mean SAM index (in Pa) over the period 1900-2100 in the 20C3M-runs and SRES A1B-runs of the 16 discussed GCMs (middle line). The outer lines are the mean plus and minus two intermodel standard deviation.

could thus expect that the change in the multimodel averaged sea level pressure would show a SAM pattern. The pressure change (Figure 7.6e) looks indeed rather like both the modelled and the observed SAM patterns with a clear annular structure. However, the low pressure area in the Amundsen and Ross Seas present in the SAM pattern cannot be found. This small change in the pressure trend compared to the late 20th century has a significant impact on sea ice as discussed below.

The sea ice pattern associated with SAM at the end of the 20th century is in good agreement between the models and the observations (Figures 7.8a and 7.8b). In both cases, there is a clear regional structure with a large positive anomaly of sea ice concentration associated with the high index years in the Ross and Amundsen Seas. A decrease of the sea

ice concentration can be found in the waters adjacent to the Antarctic Peninsula. This is linked to the low pressure anomaly in the Amundsen Sea associated with SAM (Lefebvre *et al.*, 2004). Furthermore, the positive anomaly of the sea ice concentration in the eastern Indian and the western Pacific sectors can be seen in both the models and observations, together with a decrease in sea ice concentration in the Eastern Pacific sector. The major difference between the simulated and the observed sea ice variabilities due to SAM can be found in the western Indian sector where the models have difficulties to simulate the observed increase in sea ice concentration in high SAM-index years. Integrated over the whole Southern Ocean, there is in general a small increase in sea ice extent in high SAM-index years in both the models and observations linked to the increased isolation of the Antarctic continent (Thompson and Solomon, 2002; Lefebvre *et al.*, 2004; Liu *et al.*, 2004; Lefebvre and Goosse, 2007a).

The multimodel averaged regression of the ice concentration onto the SAM index at the end of the 21st century (Figure 7.8c) is quite similar to the one obtained for the end of the 20th century (Figure 7.8a) with positive anomalies in the Ross sector and negative anomalies close to the Antarctic Peninsula. This is logical, as the sea level patterns (Figures 7.6a and 7.6f) are very similar between the two periods. However, some differences arise. The main difference is a more general positive sea ice concentration regression onto the SAM index at the end of the 21st century. Furthermore, the positive anomaly found at the end of the 20th century in the Amundsen and Ross Seas has shifted westward into the western Ross and eastern Pacific sectors (compare Figures 7.8b and 7.8c).

The analysis presented above of both the observed and simulated sea ice concentrations at the end of the 20th century shows that during this period, the trend at regional scale can partly be linked to the increased SAM index (compare Figures 7.5a, 7.5b, 7.8a and 7.8b), due to colder southerly wind anomalies in the Ross sector and warmer northerly wind anomalies around the Antarctic Peninsula (Thompson and Solomon, 2002; Lefebvre *et al.*, 2004; Marshall *et al.*, 2006). Furthermore, many studies indicate that the SAM-index is projected to rise into the 21st century (Figure 7.7). However, the change in projected change in pressure is not exactly similar to the SAM pattern (Figure 7.6). The most important difference is the pronounced low pressure area in the Amundsen Sea in the SAM pattern, which is absent in the projected pressure change. This low pressure area is very important for the response of the sea ice

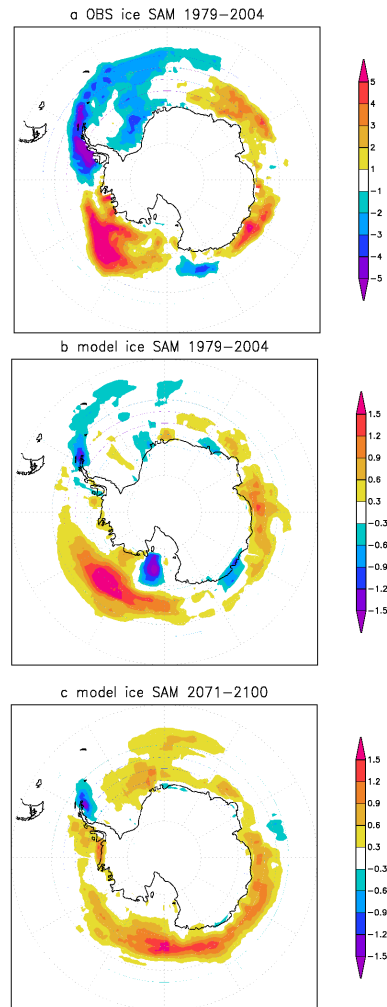


Figure 7.8: (a) The observed regression of the annual mean detrended sea ice extent (HADISST, in %) onto the detrended SAM index (NCEP/NCAR) calculated over the period 1979–2004. (b) The modelled regression of the annual mean detrended sea ice extent (in %) onto the detrended SAM index calculated over the period 1979–2004 averaged over the 16 discussed GCMs. (c) The modelled regression of the annual mean detrended sea ice extent (in %) onto the detrended SAM index calculated over the period 2071–2100 averaged over the 16 discussed GCMs.

to the SAM (Kwok and Comiso, 2002b; Liu *et al.*, 2004; Lefebvre *et al.*, 2004). Due to the absence of the low pressure anomaly in the Amundsen sector in the pressure trends (Figure 7.6e), the projected changes in sea ice concentration (Figure 7.5c) do not show a clear SAM-like response of the sea ice and are thus much more zonal in the 21st century than at the end of the 20th century. In particular, the stronger decrease of the sea ice in the region of the Antarctic Peninsula, seen in both sea ice responses to SAM (Figure 7.8a) and at the end of the 20th century (Figure 7.5a), is absent.

7.4.3 Deep convection

Only three of the models (CNRM-CM3, IPSL-CM4 and GISS-AOM) that are being discussed in this paper provide us with the mixed layer depth. Nevertheless, those three models point towards some effects on the sea ice due to changes in the mixed layer depth. Two of the models (CNRM-CM3 and IPSL-CM4) include the same ocean component. This results in very similar ocean ventilation and consequently, we will discuss only one of them (CNRM-CM3). In addition, we have to take into account the different definition of the mixed layer depth in the CNRM-CM3 and the GISS-AOM model³.

The mixed layer depth patterns for the end of the twentieth century in the GISS-AOM and the CNRM-CM3 model are qualitatively similar, with a larger mixed layer depth close to the coast of the Antarctic continent, a minimum between 55°S and 65°S and higher values again around 50°S (Figure 7.9). However, the ocean ventilation depth in the GISS-AOM model is much larger than the one in the CNRM-CM3 model, a difference that could not be explained only by the difference in definition of the mixed layer depth. In both models, the ocean ventilation along the coasts decreases during the 21st century (up to more than 100m in the GISS-AOM model and around 10m in the CNRM-model), although this decrease is mostly confined to the Pacific sector in the CNRM-CM3

³In the GISS-AOM the mixed layer depth is determined as the shallowest depth where the bulk Richardson number is equal to the critical value (0.3). Bulk Richardson numbers are evaluated by computing velocity and buoyancy differences between values at the investigated level and at the ocean surface. In the CNRM-CM3 model the mixed layer depth is defined as the depth of the highest level where the difference between the potential density of this level and the surface density exceeds 0.01 kg.m^{-3} .

model. In the GISS-AOM model, the mixed layer depth generally decreases also in other regions of the Southern Ocean. Bi *et al.* (2001) and Bitz *et al.* (2006) also found a decrease in convection in the CSIRO and the CCSM3.1 models, respectively, which they partly attributed to changes in the precipitation-evaporation balance due to increased precipitation. Furthermore, surface freshening through altered sea ice and ocean freshwater transport is proposed to be a key player in the weakening of the mixed layer depth. In addition to this general decrease of the mixed layer depth, an increase in mixed layer depth is projected around Maud Rise in the CNRM-CM3 model and in the northern Amundsen-Bellingshausen Seas ($70 - 130^\circ\text{W}$, $55 - 60^\circ\text{S}$) in the GISS-AOM model (Figure 7.9).

Locally, changes in ocean ventilation can have tremendous effect on the sea ice concentration (e.g. Gordon and Huber, 1990) because of the associated changes in the vertical oceanic heat flux. Both the coastal changes in mixed layer depth and the one at Maud Rise might thus contribute to sea ice concentration change patterns. On the one hand, this is not the case for the CNRM-CM3 model as no clear relationship between the mixed layer depth and the sea ice concentration could be noticed. On the other hand, the sharp decrease of the mixed layer depth in GISS-AOM around Prydz Bay is likely related to the increase in ice concentration there. However, it is out of the scope of the present study to demonstrate a causal link between the two changes. Furthermore, the GISS-AOM model is the only one among the 17 presented here that displays such a large increase in ice concentration in this region. As a consequence, from the limited evidence available, we must conclude that changes in mixed layer depth could play a role in the distribution of the sea ice changes during the 21st century in some areas, but it is unlikely that this corresponds to a robust and general result for all the models.

7.4.4 Changes in the oceanic currents

For this section, we discuss only the 12 models providing their barotropic streamfunction, i.e., GISS-ER, CGCM3.1 (T47), ECHAM5-MPI/OM, MIROC3.2 (medres), MIROC3.2 (hires), UKMO-HadCM3, CGCM3.1 (T63), CSIRO-Mk3.0, IPSL-CM4, MRI-CGCM2.3.2, INM-CM3.0 and ECHO-G. The transport values through Drake Passage have been calculated by making the difference at 68°W between the barotropic stream-

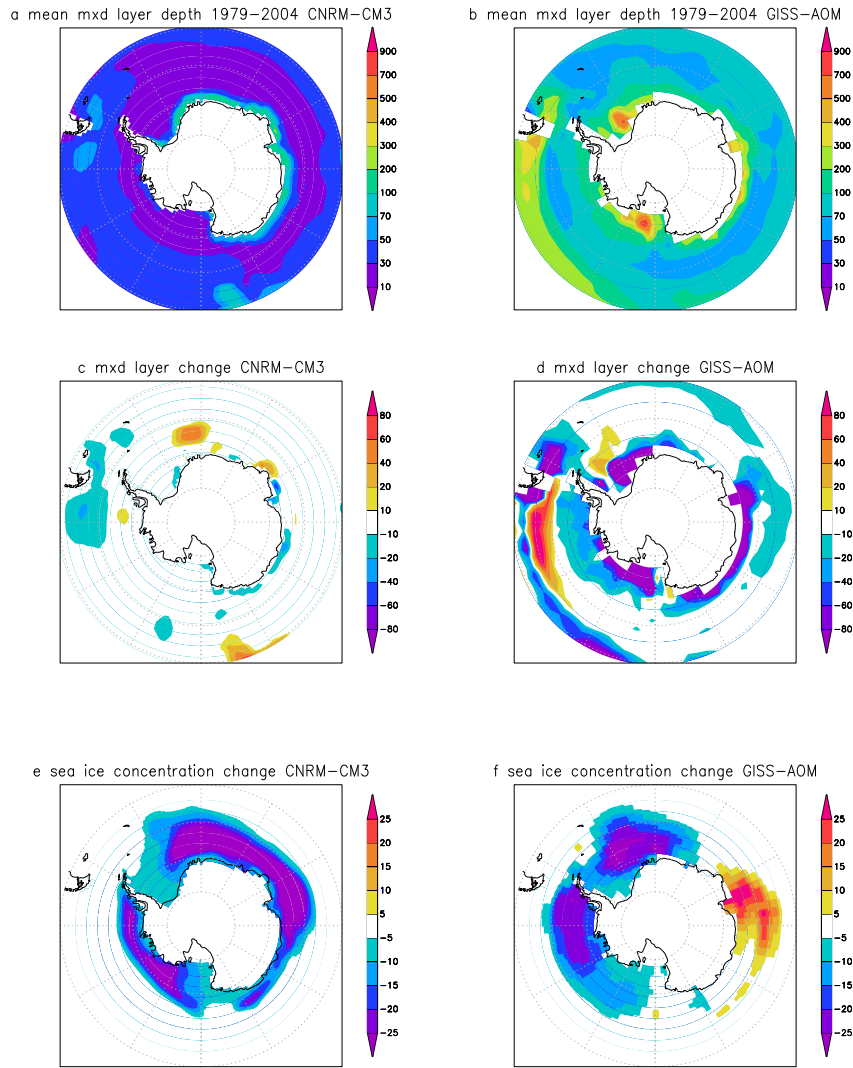


Figure 7.9: The annual mean mixed layer depth (in m) over the period 1979–2004 in the 20C3M-runs of the (a) CNRM-CM3 and the (b) GISS-AOM model and the differences between 2071–2100 in the SRES A1B-scenario and the period 1979–2004 in the 20C3M-scenario for the (c) CNRM-CM3 and the (d) GISS-AOM model. The annual mean sea ice concentration change between 2071–2100 in the SRES A1B-scenario and the period 1979–2004 in the 20C3M-scenario for the (e) CNRM-CM3 and the (f) GISS-AOM model.

function at the northernmost ocean point and the southernmost ocean point of the Drake Passage. In this subset of models, the range of the strength of the flow through Drake Passage during the last 30 years of the 20th century goes from around 50Sv until more than 300Sv, compared with an observed value of 137 ± 8 Sv (Cunningham *et al.*, 2003). Using other model runs and a slightly different method, Russell *et al.* (2006) have similar results for the model strength of the ACC. This illustrates that the ability to simulate ocean circulation in the Southern Ocean varies greatly between the different models. This can be considered as a warning and we must thus be careful in interpreting any change in the ocean currents.

The flow at Drake Passage increases in most of the AOGCMs discussed here, although the relative increase is small and some models exhibit a decrease in throughflow. For instance, the largest increase of the Drake Passage throughflow can be found in the CGCM3.1 (T63) model with a rise from 110.3Sv to 122.0Sv between period 1979-2004 and the end of the 21st century. On the contrary, the GISS-ER model gives the largest decrease, i.e. a drop from 230.5Sv to 220.8Sv. In others, the changes are negligible (e.g. a change of 1.4Sv in the CSIRO-Mk3.0 model). However, most of these changes are not significant or embedded in low frequency oscillations that are presumably not connected to the enhanced greenhouse effect.

A more robust change of the circulation can be found for the mean position of the Antarctic Circumpolar Current (ACC). Indeed, the location⁴ of the ACC exhibits a mean southward shift (Fyfe and Saenko, 2005, 2006) of about 0.7° between the end of the 21st century and the end of the 20th century. This southward change of the ACC can be found in almost all the discussed models, but is quite small as it is much smaller than the mean size of the grid. Furthermore, the change in the position of the ACC does not seem to have a large influence on the sea ice concentrations as the core of the ACC is located far north of the sea ice edge, at the longitudes where the changes in the ACC are the most important (not shown).

In addition to the changes in the ACC, there are some regional changes in the barotropic streamfunction for several models. The changes are present in locations differing between the models, but averaged over

⁴The location of the ACC has been calculated as in Fyfe and Saenko (2006).

all the models important changes occur in the northern Ross and Weddell Seas. The associated changes in oceanic heat transport likely play a role on the ice concentration. Specifically, the decreased heat transport due to more northward currents could account for the smaller simulated decrease in sea ice extent in the Ross Sea compared to other regions (Figures 7.3, 7.4 and 7.5) in the 21st century.

Looking at individual models (for instance, CGCM3.1 (T47), see Figure 7.10) the link between the changes in the streamfunction and the sea ice appears more clearly. In particular, the northward current change of at least 20 Sv in CGCM3.1 (T47) at the boundary of the Ross and Pacific sector results in a slight increase of sea ice concentration between the end of 20th and the 21st century (up to >5%) in the Ross Sea. On the contrary, the southward currents in the Bellingshausen Sea and the eastern Weddell Sea likely contribute to the locally larger decrease in sea ice extent in that region.

7.5 Conclusions

The goal of this study was to analyse the projected changes of the ice cover in the different sectors of the Southern Ocean using simulations performed by AOGCMs in the framework of the IPCC AR4. However, before presenting our final conclusions, it is important to remind the limitations of this type of analysis. First of all, the majority of the models have trouble to reproduce reasonably well the current state of the Southern Ocean. In particular, they generally tend to overestimate the total variability of the ice extent and display large errors in the regional distribution of the ice extent as well as of the trends in sea ice extent with an opposite sign compared to the observed ones during the last 26 years in most of the sectors. The period for which we have reliable observations is relatively short. It is thus not guaranteed that this period is representative of the climate variability of the Southern Ocean. However, it is unlikely that this short observation period explains much of these model-data discrepancies. The models also display large differences in the simulated magnitude of the ACC or in the vertical mixing in the Southern Ocean, making it difficult to obtain any robust conclusions about the role of the ocean in future changes of the ice cover.

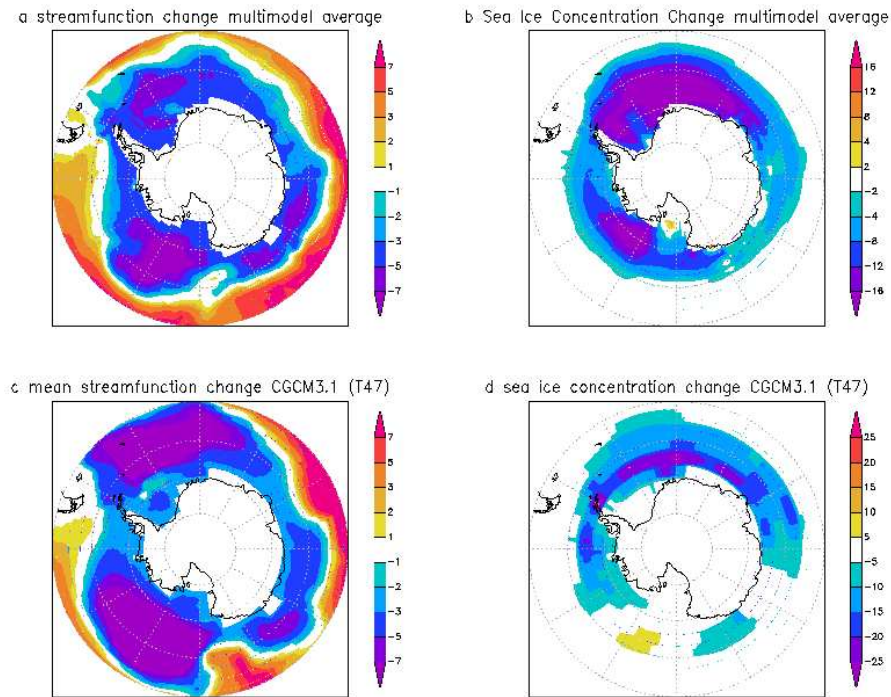


Figure 7.10: The difference in annual mean barotropic streamfunction (in Sv) between the 2071-2100 in the SRES A1B-runs period and the 1979-2004 in the 20C3M-runs period (a) averaged of the 12 discussed AOGCMs and (c) in the CGCM3.1 (T47) model. The sign of the streamfunction is defined as being positive if the currents turn counterclockwise around it. The difference in annual mean sea ice concentration (in %) between the 2071-2100 in the SRES A1B-runs period and the 1979-2004 in the 20C3M-runs period (b) averaged of the 12 discussed AOGCMs and (d) in the CGCM3.1 (T47) model.

On the other hand, the models capture quite well the SAM-pattern, and the influence of SAM on the sea ice cover.

Although individual models could have strong biases, the average over all the models provides a good estimate of the present state of the Southern Ocean. This result indicates that the model ensemble does not exhibit any systematic biases for the late 20th century and that a simple averaging procedure already cancels out many of the individual errors of the models for this period. We have thus considered that the multi-model average provides also the most reliable estimate of future changes. However, the model mean Southern Ocean sea ice extent trend at the end of the 20th century is estimated in the models to be significantly negative, on the contrary to the observations. The model mean sea ice extent variability is also not in correspondence with the observed values. As a result, we have to be very careful when analysing the future sea ice trends, in order to avoid overestimation of the model capacity⁵. A more sophisticated weighting than just taking the ensemble mean of the various models is also possible. However, designing such a technique and showing that it provides more reliable results than a single average is out of the scope of this study.

Taking into account these limitations, the most interesting conclusions that could be derived using the present day state of the art model simulations are, to our point of view, the following: the annual mean sea ice extent for the Southern Ocean in the GCMs in the SRES A1B-scenario is decreasing. Averaged over all models, a nearly annular decrease is found, with a slightly larger decrease in the Amundsen-Bellingshausen Seas and in the Weddell sector (without the region just east of the Antarctic Peninsula). The projected decrease is smaller in the Ross Sea and especially in the western part. The multimodel averaged annular regime is presently the most reasonable projection for the future. Nevertheless, nearly all the individual models (e.g. Figures 7.9e, 7.9f and 7.10d) present a non-annular decrease of the sea ice extent with even an increase in some sectors in some models (Figure 7.3). As a consequence, we can not exclude the possibility of strong regional disparities in the real evolution of the ice extent during the 21st century. In order to explain these local variations, we have discussed here four mechanisms.

⁵The sentences beginning with 'However, the model mean Southern Ocean ...' and finishing with '... avoid overestimation of the model capacity' do not appear in the published paper.

- Even if the temperature would rise equally at every location covered by sea ice at the end of the 20th century, an increase in a zone where the temperature is already close to the melting point will have a much larger impact on the sea ice concentration than an increase in a region where the temperature is always strongly below zero degrees Celsius. This could explain why the changes in sea ice in the southern Ross sector and off the southeastern coast of the Peninsula are much smaller than elsewhere. Furthermore, as the western Ross and Weddell Seas are characterized by an import from cold regions with a high concentration of sea ice (respectively from the southern Ross Sea and the southern Weddell sector), the projected changes are also generally smaller there.
- During the late 20th century, the increase of the SAM index is associated with regional changes in the ice cover, particularly with a decrease of the ice extent close to the Antarctic Peninsula and an increase in the Ross-Amundsen area. Despite the subsequent projected increase in the SAM during the 21st century, a similar response of the sea ice extent for this period is not simulated. This is due to the fact that the sea level pressure trend has indeed a strong SAM component but is more zonal than SAM, with a strong impact on the ice extent. This shows once more that it is very dangerous to simply extrapolate one connection obtained during a period to a different period.
- The mixed layer depth significantly changes in some models between the end of the 20th and 21st centuries. These variations can locally have large effects on the ice cover. However, as the variations are different in magnitude, sign and location between the various models, no conclusion can be made on the eventual location and magnitude of changes in sea ice associated to variations in the mixed layer depth.
- The Antarctic Circumpolar Current (ACC) slightly strengthens in most models and shifts somewhat polewards, although, in most of the models, these changes are small. The effects on the sea ice are projected to be quite small due to the relatively small change of the ACC and the distance between the main core of the ACC and the sea ice edge at most longitudes. On the contrary, regional

current changes can increase or decrease the heat advection by the ocean and as a result change the sea ice conditions. This happens in most of the models albeit not in all of them and seems to be most important in the Ross sector, contributing to the relatively small decrease of the ice extent in this area.

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Chapter 8

Conclusions and perspectives

Using a coupled sea ice–ocean model (Chapter 2) driven by daily NCEP /-NCAR reanalysis data for near-surface air temperatures and winds, and monthly climatologies for humidity, cloudiness and precipitation, we have investigated several aspects of the internannual variability of the sea ice. First of all, we looked at the effect of the Southern Annular Mode (SAM), which is the first EOF of the sea level pressure in the non-tropical Southern Hemisphere. For the oceanic part, our work confirmed the results obtained by Hall and Visbeck (2002). At the surface, stronger westerlies at high latitudes associated to high SAM-indexes push the surface water there to the northeast. This induces an upwelling close to the Antarctic continent, which is compensated by a downwelling around 45°S. Finally, the cycle is closed by a southward anomaly of the ocean currents in the deep sea, in years with a high SAM-index. In addition, due to increased stirring of the water column in a broad band, the depth of the mixing layer increases in all seasons.

On the contrary to the confirmation by our results of the oceanic part of the scheme of Hall and Visbeck (2002), the response of the sea ice to SAM they proposed was not confirmed by our results. The response of sea ice to the SAM displays large regional differences in our model, in agreement with the observations (Kwok and Comiso, 2002b; Liu *et al.*, 2004). Due to a non-annularity in the SAM-pattern, i.e. a stronger low pressure anomaly in the Amundsen-Bellingshausen Seas than at corresponding longitudes, colder southerly winds prevail more often in the Ross Sea, while warm maritime northerly winds occur more around the Antarctic Peninsula in years with a positive SAM-index. Furthermore,

this same pressure anomaly gives a northward velocity anomaly in the Ross Sea, increasing the sea ice extent even more. Close to the Antarctic Peninsula however, the ice is pushed against the coast, decreasing the sea ice extent there. In the Pacific sector, the sea ice extent increases due to an increased northward transport of the ice (Ekman drift due to increased westerlies) and to increased insulation of the region from the Pacific Ocean. These results suggest that the increase in the sea ice extent in the Ross sector and the decrease around the Antarctic Peninsula could be linked to an increase in the SAM-index during the last decades.

Despite its annular appearance, some non-annular aspects of the SAM-pattern can play an important role in the response of sea ice to it. Indeed, the regional response of sea ice to the SAM is not mainly determined by effects resulting from the annular pressure pattern of the SAM but also from the non-annularity. In order to find out which part is due to the annular part and which is not, a new series of experiments has been done. Furthermore, this series has as goal to find which of the effects are due to thermal or mechanical effects. We have done this by respectively adding only the mechanical part¹, only the thermal part², only the annular part, only the non-annular part and the combination of all the parts to the atmospheric forcing of the model. As the mechanical forcing has no large interseasonal variations, an annual mean forcing has been applied for it. On the contrary, the atmospheric temperature regressions to the SAM do show a large difference, which we have considered by taking a seasonal differing temperature forcing. This has been done each time for the same period and the results of the different runs have been compared.

However, it should be noted that the division we have made between the mechanical and thermal forcings is artificial. In reality, both changes occur simultaneously and influence each other. Furthermore, due to the experimental design, no feedback between the atmosphere and ocean is represented in the model. For instance, a change in wind stress can induce a shift in the ice edge. This would normally also change the temperature, probably amplifying the response.

We have found that, in line with previous investigations, the thermal forcing increases the sea ice extent, mainly in autumn and winter,

¹The part driven by the wind stress response to the SAM.

²The part driven by the atmospheric temperature response to the SAM.

by cooling the ocean close to the Antarctic continent. The only exception is the region around the Antarctic Peninsula and the Weddell Sea where warmer conditions decrease the sea ice extent. The response to the mechanical part of the SAM, i.e. the wind stress, is important in several domains. On the one hand, the wind stress forcing enables the same oceanic response as discussed above. As a result, relatively warm and saline water is pumped up close to the continent, melting the sea ice there. In our simulations this destabilises the water column in some regions of the Weddell sector, which increases the heat flux towards the surface, decreasing the sea ice there. On the other hand, the wind stress drives the changes in sea ice velocity. The southerly wind stress anomaly in the Ross Sea pushes the ice outward increasing the sea ice extent there. In addition, the westerly wind anomalies increase the Ekman Drift pushing the ice to the north when the SAM index is high.

In general, the mechanical and thermal components of the forcing reinforce each other in the Ross sector and in the Weddell sector, as seen above. This leads to a dipole reaction of the sea ice to the SAM. In the other sectors, the responses to the thermal and mechanical forcings are mostly opposite. In total, both forcings have more or less the same amplitude.

The variability of the winter sea ice extent integrated over the Southern Ocean cannot be explained by the SAM variability as the response of the sea ice to SAM is a dipole, as seen above. Indeed, integrating this dipole over the total Southern Ocean leads to only a very small response. Other modes of variability, like ENSO, have a similar dipole response, leading to very small correlations between the total sea ice extent and the ENSO-index. We have shown that these correlations are not significant for all the known modes of atmospheric variability, except for the SAO. However, this last correlation is not very robust. In addition, in the different sectors of the Southern Ocean, the correlations between the SAO and the sea ice extent are small. Finally, the link between the SAO and sea ice does not seem to be stationary when investigating different periods.

As none of the atmospheric modes of variability seem to be able to explain a large part of the total sea ice variability, what can explain this variability? The atmospheric pressure pattern associated with the largest sea ice extent is a non-symmetric wave 3 pattern. This induces a shifted pattern in temperature. An important feature of this pattern is

a low pressure anomaly at about 60°S in both the Weddell and the Ross sectors. These low pressure anomalies have a clear impact locally on the ice extent in those seas. This induces important changes in the total sea ice extent, as these two sectors have the largest variability. However, the pressure pattern most favourable for a large sea ice extent is no dominant mode of variability as most of the centres of action of the wave 3 pressure pattern are not related. Furthermore, there is no positive correlation between the sea ice extent in the Ross and the Weddell sectors. This shows that, when there is a large sea ice extent in one of them, this does not have to mean a large one in the other too. In conclusion, the wave 3 pattern does not seem to be a mode of variability on its own, but merely a combination of regional variations with no strong correlation between them. Finally, our results showed that the total sea ice extent variability cannot be seen as a single entity but as the sum of the sea ice variabilities in different sectors. The sectoral sea ice behaves without important influences on the other (even adjacent) sectors, except for the Amundsen-Bellingshausen Seas. In this last sector, the sea ice extent is strongly correlated (negatively) with the Weddell sea ice extent and the Ross sea ice extent, although this last correlation is only found in the model and not in the observations.

Having studied the sea ice variability in the late 20th century, it is important to look into the future. For instance, the knowledge of the current variability of the Southern Ocean sea ice can help us to interpret the changes projected for the future. Therefore, we analysed the projected changes of the ice cover in the different sectors of the Southern Ocean using simulations performed by AOGCMs in the framework of the IPCC AR4. First of all, we made an effort to assess if the models were capable of simulating the end of the 20th century situation. In general, the models tend to overestimate the variability of the sea ice extent. Furthermore, they display large errors in the regional distribution of the ice extent and showed trends with an opposite sign compared to the observed ones during the last 26 years, in most of the sectors. However, we could not exclude that these shortcomings could be due to the short period for which we have observations, as there is no guarantee that this period is representative for the natural climate variability in the Southern Ocean. The simulated ocean is, in some models, far off the reality, especially for the ACC and the mixing layer depth. On the other hand, the atmosphere, and in particular the SAM-pattern, is well

represented. Finally, the model average provides, in general, a much better estimate of the present state of the Southern Ocean than any of the individual models alone. Therefore, we can assume that there are no systematic biases in the models and a simple averaging procedure cancels out most of the important individual errors of the models for the late 20th century. For the 21st century, it is possible that the average is not any more the best representation of the climate. However, there is no way to assess this reasonable hypothesis.

The models project that the annual mean sea ice extent will decrease in the SRES A1B-scenario. Averaged over all the models, the decrease is almost annular, slightly more pronounced in the Amundsen-Bellingshausen Seas and in the Weddell sector. On the other hand, the decrease is somewhat smaller in the Ross Sea. However, nearly all the models provide a less annular response than the multimodel average. As a consequence, we cannot exclude the possibility that there will be significant differences in the sea ice extent decrease between the different sectors. In order to explain the regional differences, four mechanisms have been discussed.

First, a similar temperature increase in the different sectors could lead to a non-similar decrease in the sea ice extent. Specifically, an increase in a ice-covered zone with temperatures close to the melting point will have a larger effect on the sea ice extent than the same increase in a zone with temperatures far from the melting point. In addition, such an increase in a zone without sea ice cover would not have any effect on the sea ice extent. Due to their cold character, the southern Ross Sea and the region off the southeastern coast of the Antarctic Peninsula will probably have less influence of the rising temperatures.

Secondly, the rise in the SAM-index, documented in the second half of the 20th century is projected to continue. However, the sea level pressure trend projected for the 21st century has a strong SAM component but is more zonal than SAM, whereas the sea level pressure trend projected for the end of the 20th century resembles more the SAM pattern. As seen above, the non-zonal part of the SAM pressure pattern is very important for the sea ice response to the SAM. As a result, the changes in sea ice in the 21st century due to the changes in pressure cannot be extrapolated from the 20th century response of the sea ice to SAM.

Thirdly, changes in the mixed layer depth could play a role. However, we have shown that the current models are absolutely not capable of well

representing the current mixed layer depth. Furthermore, the projected changes into the 21st century are different in the different models. We have seen that these changes can induce important changes in the sea ice concentration, in some individual models. However, these changes in the mixed layer depth appear to occur at different locations in the models, and there are different responses of the sea ice associated to them. Therefore, no firm conclusion can be taken on whether these effects will be important, their eventual location and magnitude.

Fourthly, an important response of the sea ice to changes in the ocean currents is possible. Regionally, in some models, current changes induce changes in the sea ice extent. This effect seems to be the most important in the Ross sector, enabling a smaller projected decrease of sea ice there. The ACC increases in strength and also moves slightly poleward in the projections for the 21st century. However, this effect is projected to have no important effects on the sea ice.

Despite the increase in knowledge during the last years, a lot of unsolved questions remain. In my opinion, one of the major constraint on the current knowledge of the Southern Ocean is the amount of data available. Luckily, there is a continuous instream of new, high-quality information. Here, we will discuss shortly how these new information sources can help us in understanding the climate of the Southern Ocean.

First of all, a new version of the ORCALIM-model has been developed, namely the NEMO-LIM model. This model has been coded in order to be able to run in parallel on several processors. As a result, it can be used with a much higher resolution. This enables a more accurate description of some important factors, like coastal polynyas or water exchanges on the continental shelf and slope. These last two play an important role in the formation of bottom water and a correct representation of them will enable a better description of the Southern Ocean water mass characteristics. This better description is needed for a thorough study of these water mass characteristics at different timescales, which has not been done here. In addition, an improved version of the LIM-model has been developed by Martin Vancoppenolle. Some of the interesting changes in the new model are the subgrid-scale distribution of ice thickness with both deformed and undeformed ice types and the inclusion of new diagnostics, like the ice age. In contrast to the old version of the ORCALIM model, the ice is not assumed to have a constant salinity, taking into account the changes in the shape of the ice salinity

profile. This new model should provide a better simulation of the sea ice variability and the salinity of the upper ocean layer, while the new diagnostics can play an important role in understanding the mechanisms governing this variability.

Furthermore, the NEMO-LIM model has been recently coupled to the SWAMCO-4 (Sea Water Microbial COmmunity model) model. This last model is a biogeochemical model of the marine planktonic system, describing, for instance, the iron cycling in the ocean and the exchange of CO_2 between the ocean and the atmosphere. The results of this model could be used to investigate the influence of the modes of variability on the biology and the CO_2 flux in the Southern Ocean. The coupling of the NEMO-LIM model to an ice shelf model could also improve the representation of the water mass characteristics (especially the Antarctic Bottom Water and the properties of the water on the continental shelf).

On a somewhat longer term, SLIM (Second generation Louvain-la-Neuve Ice ocean Model) should become available. This model is based on a unstructured mesh offering much more flexibility than the structured mesh models, like ORCALIM or NEMO-LIM. More specifically, this approach leads to a more efficient use of computer time as it uses only a high resolution grid where needed. When available, this model could grant interesting possibilities in the research of phenomena occurring at a scale for which the current model resolution is insufficient, like the coastal polynyas or eddies in the ACC.

Improvements in the domain of global coupled atmosphere-ocean-sea ice-land models enable detailed projections for the future of climate. However, most of the studies using these data have been made on the Northern Hemisphere, as the impacts of climate change affect many more people in the Northern Hemisphere than in the Southern Hemisphere. In the last decade, the interest in the Southern Hemisphere has risen, leading to more studies devoted to this hemisphere. We can only hope that in the future, this will continue, amounting to an increased knowledge of the climate of the Southern Hemisphere.

In addition, further research also increases the quality of the reanalysis data, increasing their usefulness as forcing of a ocean-sea ice model. Most importantly around the Antarctic Peninsula, an improved accuracy of the atmospheric forcing would result in better model simulations. Therefore, a higher resolution of the model used to make the reanalysis is needed, in order to improve the geographical description of the

Antarctic continent and especially the Antarctic Peninsula. One way to do this would be using a regional model, focussing on the Southern Ocean and the Antarctic continent to make a detailed regional reanalysis, as proposed by David Bromwich and Hubert Gallée. Furthermore, an improved coverage of the Southern Hemisphere with data collection points, such as weather stations, would further increase the quality of the reanalyses. As seen in Chapter 2, an improved representation of the cloud cover, precipitation and humidity in the reanalysis would enable the possibility to force ocean-sea ice-models with daily changing values of these parameters. This would increase the consistency of the simulated variability.

An increasing set of observations is not only useful for the quality of the reanalyses. Both longer time series and new reliable datasets, like satellite ice thickness measurements, will become very helpful in the future to understand the climate variability in the Southern Ocean. Currently, in the framework of the International Polar Year, large observational programs are being carried out, on the Antarctic continent and in the Southern Ocean, both at the surface and in depth. Hopefully, these data will give us important information in understanding the climate of the Southern Ocean.

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