Marina Noro dos Santos

**Future Changes of the Antarctic
Coastal Current as seen from a
Downscaling Experiment**

São Paulo
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Master thesis submitted to *Instituto Oceanográfico da Universidade de São Paulo* in partial fulfillment of the requirements for the degree of Master of Science, Post-Graduation Oceanography Programme, with emphasis in Physical Oceanography.

Advisor: Prof^a Dra. Ilana Wainer

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UNIVERSIDADE DE SÃO PAULO - USP
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List of Acronyms

AABW Antarctic Bottom Water

AASW Antarctic Surface Water

ACC Antarctic Circumpolar Current

ACoC Antarctic Coastal Current

AR4 Assessment Report 4

ASF Antarctic Slope Front

AZ Antarctic Zone

CAM Community Atmosphere Model

CDW Circumpolar Deep Water

CLM Community Land Model

CCSM Community Climate System Model

CESM Community Earth System Model

CMIP3 Coupled Model Intercomparison Project Phase 3

GHG greenhouse gases

HSSW High-Salinity Shelf Water

ISW Ice Shelf Water

IPCC Intergovernmental Panel on Climate Change

LCDW Lower Circumpolar Deep Water

LSSW Low-Salinity Shelf Water

MWDW Modified Warm Deep Water

NCAR National Center for Atmospheric Research

NCEP National Centers for Environmental Prediction

NOAA National Oceanic and Atmospheric Administration

GCM General Circulation Model

PF Polar Front

PFZ Polar Frontal Zone

ROMS Regional Ocean Model System

SACCF South Antarctic Circumpolar Current Front

SAF Subantarctic Front

SAM Southern Annular Mode

SAMW Subantarctic Mode Water

SASW Subantarctic Surface Water

SAZ Subantarctic Zone

SB Southern Boundary

SH Southern Hemisphere

SLP Sea Level Pressure

SODA Simple Ocean Data Assimilation

SRES Special Report Emissions Scenarios

SSS Sea Surface Salinity

SST Sea Surface Temperature

STF Subtropical Front

SW Shelf Water

SZ Southern Zone

THC Termohaline Circulation

UCDW Upper Circumpolar Deep Water

WDW Warm Deep Water

WG Weddell Gyre

WSBW Weddell Sea Bottom Water

WSDW Weddell Sea Deep Water

WW Winter Water

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ABSTRACT

Climate global models have contributed to the understanding of climate changes impacts on several Earth system processes. It is known that impacts on the ocean large scale circulation are considerably relevant. However, these models are not suitable to represent regional scale processes. Therefore, to evaluate the climate change impacts on small scale processes downscaling is necessary. This study was divided in two parts. The first part consists of evaluation and understanding of the behavior of the Southern Ocean circulation from a large scale perspective. For this, we used the ocean component (Parallel Ocean Program version 2 – POP2) of the National Center for Atmospheric Research–Community Earth System Model version 1 (NCAR–CESM1.0). The second part aimed to evaluate the Antarctic Coastal Current (ACoC) in the Weddell Sea through high-resolution regional simulations with the Regional Ocean model System (ROMS) which were forced with results from NCAR–Community Climate System Model version 3.0 (NCAR–CCSM3), an earlier version of NCAR–CESM1, experiments of the Intergovernmental Panel on Climate Change (IPCC) Assessment Report 4 (AR4). The first simulation was forced with 20th century historical scenario (ROMS–20C3M) and comprises the 1980 to 1999 period, whereas the second run considers the 21st century scenario A2 (ROMS–SRESA2) which comprises the 2021 to 2047 period. Results from CESM1–POP2 represents satisfactorily the large scale mean patterns and is in agreement with available data. The results show the limitation of these models to reproduce important features of the coastal region, which makes downscaling necessary to understanding smaller scale processes. The ACoC and its transport (≈ 22 Sv) were satisfactorily represented by regional simulations. From the 20th to 21st century projections, weakening of the ACoC transport was observed and shown to be mainly related with changes in the termohaline forcing.

Key-words: Antarctic Coastal Current, Climate Changes, High-resolution regional model, Weddell Sea

RESUMO

A utilização de modelos climáticos globais tem contribuído para o entendimento dos efeitos da mudança no clima em diversos processos do sistema terrestre. Os impactos na circulação oceânica de larga escala são conhecidamente relevantes. Porém, esses modelos não são adequados para a representação de processos na escala regional. Dessa forma, para avaliar os impactos da mudança do clima nos processos de menor escala a regionalização numérica é necessária. Este trabalho foi dividido em duas partes. A primeira consiste na avaliação e entendimento da circulação no Oceano Austral do ponto de vista da larga escala. Para isso a componente oceânica (*Parallel Ocean Program version 2* – POP2) do *National Center for Atmospheric Research – Community Earth System Model version 1* (NCAR–CESM1.0) foi utilizada. A segunda parte teve como objetivo avaliar as mudanças na Corrente Costeira Antártica (CCoA) no Mar de Weddell através de simulações regionais de alta resolução com o *Regional Ocean Model System* (ROMS) forçados com dois experimentos do NCAR–*Community Climate System Model version 3.0* (NCAR–CCSM3) para os cenários do *Intergovernmental Panel on Climate Change (IPCC) Assessment Report 4* (AR4). A primeira simulação foi forçada com o cenário histórico do século XX (ROMS–20C3M) e compreende o período de 1980 a 1999, já a segunda considera o cenário pessimista A2 (ROMS–SRESA2) para o século XXI e compreende o período de 2021 a 2047. Os resultados obtidos do CESM1–POP2 mostraram que o modelo representa de forma satisfatória os padrões de larga escala e concordam com os dados disponíveis. Os resultados também evidenciam a limitação de modelos como esse em reproduzir características importantes da região costeira, o que torna necessário a regionalização para o entendimento dos processos de menor escala. Os resultados das simulações regionais reproduziram de forma consistente a CCoA e o seu transporte (≈ 22 Sv). Do cenário do século XX para o do século XXI houve um enfraquecimento do transporte da CCoA, que mostrou estar relacionado principalmente à mudanças na forçante termohalina.

Palavras-chave: Corrente Costeira Antártica, Mudanças Climáticas, Modelo regional de alta resolução, Mar de Weddell

1 Introduction

Coastal regions are sites of constant exchange of mass and energy with open ocean. Therefore, the understanding of coastal dynamics is important not only to evaluate the meso and small-scale processes response to changes in the large-scale processes, but also the reverse influence of these smaller scale processes on larger scale processes.

The Antarctic coastal region is particularly important to processes that affects the global climate. The Antarctic continental shelf includes source areas of Antarctic Bottom Water (AABW), which is a key element to Meridional Overturning Circulation (MOC) (Orsi et al., 1999). Therefore, changes in Antarctic coastal circulation could affect this water mass formation which can significantly alter the role of the ocean on the storage and distribution of heat.

Despite its importance, the Antarctic coastal dynamic is still poorly understood. Descriptions of hydrographic properties and the circulation of the extensive continental shelf surrounding Antarctic began to be stimulated in the late 1960s–early 1970s due the development of commercial fisheries in the Southern Ocean (Hoffmann and Klinck, 1996).

Although several programs have been developed in the region, few of them were designed specifically to study the hydrography and the circulation of the Antarctic continental shelf (Hoffmann and Klinck, 1996). Moreover, the majority of the available data are from observations made since the mid-1970s and are from austral summer (Hoffmann and Klinck, 1996).

Due to the lack of observational data, in time and space, climate global models have been applied to understand the response of several Earth system processes to climate changes. However, these models are not suitable to represent regional scale processes such as the Antarctic coastal circulation. Therefore, to evaluate the climate impacts on meso and small-scale processes the downscaling is necessary.

In this study we first evaluated the behavior of Southern Ocean circulation from a large-scale perspective using the ocean component (Parallel Ocean Program version 2 – POP2) of the National Center for Atmospheric Research – Community Earth System Model (NCAR–CESM1.0). The second part aimed to investigate the response of the Antarctic Coastal Current (ACoC) in the Weddell Sea to climate changes. For this, we evaluated the results of regional simulations performed with Regional Ocean Model System

(ROMS) developed by Ocean, Climate and Cryosphere Laboratory (OC2) group from the Oceanographic Institute of University of São Paulo (IOUSP).

1.1 Climate Changes in Antarctic

To understand the global climate impacts on Antarctic coastal circulation, we first need evaluate the climate changes impacts over Antarctica.

According to the Intergovernmental Panel for Climate Change (IPCC, 2014), each of the last three decades has been warmer at the Earth's surface than any preceding decade since 1850. The globally averaged combined land and ocean surface temperature data as calculated by a linear trend show a warming of $0.85 [0.65 \text{ to } 1.06]^{\circ}\text{C}$ over the period of 1880 to 2012 (IPCC, 2014).

Changes observed in high latitudes of the Southern Hemisphere (SH) are not characterized only by mean temperatures, but also by important changes in atmospheric circulation. Several studies, such as Thompson and Solomon (2002), Arblaster and Meehl (2006) and Thompson et al. (2011), showed that changes in the tropospheric circulation are associated with the development of the ozone hole, but greenhouse gases (GHG) also contribute significantly.

The mean atmospheric circulation of mid-high latitude SH is dominated by a westerly circumpolar vortex which extends from the surface to the stratosphere (Thompson and Solomon, 2002). The vortex is strongest during midwinter in the stratosphere, when polar temperatures are coldest, and weakest during summer months (Thompson and Solomon, 2002).

Several studies reported a significant trend toward the positive phase of the SAM since mid-1960s (Thompson and Wallace, 2000; Kwok and Comiso, 2002; Marshall, 2003). However, according to Hartmann et al. (2013), it is probable that SAM index has become more positive since the 1950s.

The trend toward the SAM high-index polarity is associated with polar vortex strengthening and the poleward shift of westerlies (Thompson and Solomon, 2002). Besides that, Kwok and Comiso (2002), Thompson and Solomon (2002) and Abram et al. (2014), show the SAM high-index polarity favors cooling over much of Antarctica, except over the Antarctic Peninsula, where a warming trend is observed.

This pattern of surface temperatures has been observed in recent decades. According to Hansen et al. (1999) the Antarctic Peninsula and Bellingshausen Sea are included in the areas where a particularly rapid regional warming between 1950 and 1998 was detected.

Although the annual mean Antarctic sea-ice extent increased from 1.2 to 1.8% per decade between 1979 and 2012, strong regional differences are observed in Antarctica (IPCC, 2014). Several studies showed that the most rapid mass losses are being along the Amundsen and Bellingshausen Seas coasts, with substantial thinning of adjoining ice shelves (Paolo et al., 2015; Pritchard et al., 2012), and along Antarctic Peninsula after ice-shelf disintegration events (Scambos et al., 2004). According to Paolo et al. (2015), the Amundsen and Bellingshausen Seas regions account for less than 20% of the total West Antarctic ice-shelf area but, combined, contribute more than 85% of the total ice-shelf volume loss from West Antarctica. In contrast, a large increase in sea ice extent has been observed in the Ross Sea (Vaughan et al., 2013).

The ice shelves buttress the ice sheet (Joughin and Alley, 2011). Therefore, loss of ice shelves would lead to instability and could accelerate the melting of the ice sheet. Shepherd et al. (2012) showed through an ensemble of satellite altimetry data that between 1992 and 2011 West Antarctica and Antarctic Peninsula had a loss of 65 ± 26 and 20 ± 14 Gt $\cdot$ yr $^{-1}$ in ice sheet mass. Rignot et al. (2008) showed a maximum loss in ice sheet mass in 2006 of 132 ± 60 and 69 ± 46 Gt $\cdot$ yr $^{-1}$ in these two regions. In contrast, the East Antarctica showed a gain of 14 ± 43 Gt $\cdot$ yr $^{-1}$ in the ice sheet loss (Shepherd et al., 2012).

According to Shepherd et al. (2012) since 1992 ice sheets contributed with approximately 0.59 ± 0.20 mm $\cdot$ yr $^{-1}$ to the rate of global sea level rise. Hock et al. (2009) showed that from 1961 to 2004 the Antarctic glacial flow contributed with 28% of the global sea level rise and its contribution is likely to increase.

Besides maintaining the stability of ice sheets, sea ice is an important component of the climate system. Sea ice cover on the ocean changes surface albedo, insulates the ocean from heat loss, and provides a barrier to the exchange of momentum and gases such as water vapor and CO₂ between ocean and atmosphere (Vaughan et al., 2013). Regional climate changes affect sea ice characteristics and these changes can feedback on the climate system, both regionally and globally (Vaughan et al., 2013).

Energetic interactions between the atmosphere, ocean and sea ice result in water mass formation, which plays an important role in the global overturning, and ventilates a substantial fraction of the global ocean volume (Rintoul et al., 2001). Furthermore, water masses receive their signatures from atmospheric processes and consequently are good indicators for changes in climatic conditions (Leffanue and Tomczak, 2004).

1.2 The Intergovernmental Panel on Climate Change – IPCC

According to [IPCC \(2014\)](#) climate change is unequivocal and human influence to climate system is certain. Anthropogenic greenhouse GHG emissions have increased since the pre-industrial era ([IPCC, 2014](#)).

Future projections of GHG emissions are determined by several factors such as demographic and socio-economic development, technological change, land use patterns, and climate policy ([IPCC, 2000, 2014](#)). The future evolution is uncertain therefore, to evaluate variations in these driven forces the Special Report on Emissions Scenarios (SRES) were developed for the IPCC Assessment Report 4 (AR4) ([IPCC, 2007](#)).

Four possible futures were developed to describe the relationship between emission driving forces and their evolution considering different demographic, social, economic, technological, and environmental developments ([IPCC, 2000](#)). Scenarios based on the same storyline constitute a scenario "family" ([IPCC, 2000](#)).

IPCC SRES projected an increase of 25 to 90% in GHG global emissions between 2000 and 2030, considering fossil fuels the main energy source ([IPCC, 2007](#)). Continuous emissions of these gases at current and above rates could cause more warming and leads to global climate changes in 21st century greater than observed in 20th century ([Figure 1.1](#)) ([IPCC, 2007](#)).

Figure 1.1: **Left Panel:** Global GHG emissions (GtCO₂-eq) in the absence of climate policy. **Right Panel:** solid lines represents multi-models global averages of surface warming for scenarios A2, A1B and B1, shown as continuations of 20th century simulations. the pink line is not scenario, but is for Atmosphere-Ocean General Circulation Model (AOGCM) simulations where atmospheric concentrations are held constant at year 2000 values (IPCC, 2007).

In this study, we considered regional simulations forced by the Community Climate System Model version 3.0 (CCSM3) experiments of the AR4 IPCC/CMIP3 (Coupled Model Intercomparison Project phase 3) for the 20th century historical scenario (Climate of 20th Century Experiment – 20C3M) and the pessimist SRES scenario A2 (Figure 1.2).

The 20C3M experiment considers temperature and salinity from *Levitus* climatology as initial conditions and atmospheric component uses a fixed concentration of GHG, which represents pre-industrial (1870) concentration of 348 ppm and reaches 380 ppm (Pereira and Wainer, 2015).

These simulations consists of five ensembles that uses the mean to eliminate internal variabilities (drifts) and initial conditions dependency (Pereira and Wainer, 2015). Each run comprises the period of 1850 to 2000 and includes historical records of GHG (CO₂, CH₄, N₂O e CFC_s), tropospheric sulfates, volcanic activity, and solar forcings (Yukimoto, 2005).

Figure 1.2: Results from CCSM3 experiments of IPCC AR4. Source: http://www.cgd.ucar.edu/ccr/strandwg/CCSM3_AR4_Experiments.html (accessed in November 2017).

The A2 scenario from SRES uses the end of 20C3M run as initial condition (present–2100). Its future storyline describes a very heterogeneous world and is based on self-reliance and preservation of local identities (IPCC, 2000). Fertility patterns converge very slowly, which results in continuously increasing global population. Economic development is primarily regionally oriented and per capita growth and technological change are more fragmented and slower than in other storylines (IPCC, 2000). The emissions in A2 scenarios families remain positives up to 2100 and reaches a GHG concentration of 870 ppm (Pereira and Wainer, 2015).

1.3 The Antarctic Environment

Different from the Northern Hemisphere (NH) polar region, which is a semi-enclosed ocean almost entirely surrounded by land, the SH polar region features a continental landmass surrounded by an ocean. As a consequence of this pronounced geography difference, the meridional temperature gradients become enhanced in SH, resulting in a semipermanent baroclinic zone around Antarctic (Bromwich and Parish, 1998). This zone has the effect of thermally isolating the Antarctic continent. As a result, Antarctica experiences the coldest and most harsh climate on the Earth (Bromwich and Parish, 1998).

Another important characteristic of the Antarctic continent is its orography, which has a mean elevation of nearly 2400 m, constituting the highest continent on earth (Bromwich

and Parish, 1998). The continental orography has important influences on the meteorological features.

Antarctica can be divided into three major areas: West Antarctica, East Antarctica and the Antarctica Peninsula (King and Turner, 1997). East Antarctica is the largest area and is dominated by the high Antarctic plateau (King and Turner, 1997). Eastern and western portions are unequal divided by the Transantarctic Mountains (Figure 1.3), which stretch from Victoria Land to the Ronne Ice Shelf and rise to a maximum elevation of 4528 m (King and Turner, 1997). This north-south mountain barrier has a major influence on the west-east oceanic and atmospheric circulation of the high southern latitudes (Ainley et al., 2009).

Antarctica is dominated by the Antarctic Ice Sheet, which covers almost the entirely continent. Antarctic Ice Sheet could exceed 4 km in thickness (King and Turner, 1997; Bromwich and Parish, 1998), and the continent as a whole contain around 90% Earth's fresh water ice (King and Turner, 1997).

Figure 1.3: Antarctic map. Source: <https://geology.com/world/antarctica-satellite-image.shtml> (accessed in November 2017).

Ice shelves constitute 11% of the total area of Antarctic, with the two largest being the Ronne-Filchner Ice Shelf in Weddell Sea and Ross Ice Shelf (King and Turner, 1997; Ainley et al., 2009). Ice shelves are several hundreds of meters thick and the ocean areas under them are among the most isolated and unvarying environments on the Earth. Nevertheless, ocean currents carry water masses into the cavities beneath the ice shelves (Ainley et al., 2009).

1.3.1 Antarctic Meteorology

Atmospheric general circulation is primarily driven by the unequal solar radiation distribution over the Earth's surface, that creates a large Equator-to-pole temperature gradient and promotes the heat transport between the tropics and the poles (King and Turner, 1997; Ainley et al., 2009). Besides being a region that receives much less solar radiation, the Antarctic also constitutes a heat sink and exerts considerable influence on circulation of SH (King and Turner, 1997).

Snow and ice surfaces have high albedo and thus enhance the cooling by reflecting more solar radiation (King and Turner, 1997). According to King and Turner (1997) and Bromwich and Parish (1998) the continental albedo is around 0.80 or higher, implying that only about 20% of the incident solar radiation is available to warm the continental ice slope. As a result, a strong radiative deficit is present for most of the year Bromwich and Parish (1998).

The almost continuous net radiative loss leads to development of the antarctic surface temperature inversion (Phillpot and Zillman, 1970; Bromwich and Parish, 1998). Temperature inversion over the Antarctic continent persists for almost the entire year except during the brief midsummer period (Phillpot and Zillman, 1970; Bromwich and Parish, 1998).

The strongest inversion occurs during winter over the highest portions of the continent and exceeds 25°C (Phillpot and Zillman, 1970; Connolley, 1996; Bromwich and Parish, 1998). During clear and calm conditions it is occasionally in excess of 35°C (Phillpot and Zillman, 1970). Although inversion is more pronounced in the interior of continent during winter, significant inversions also occur for much of the year even in the coastal region (King and Turner, 1997).

Temperature inversion has important dynamical constraints on the near-surface wind-field (Connolley, 1996; Bromwich and Parish, 1998). Radiative cooling establishes a density contrast, with most dense air near the surface. As a result, a horizontal pressure

gradient is developed in the lowest portion of the atmosphere over the sloping terrain, directed in a downward slope direction (Parish et al., 1993; Bromwich and Parish, 1998).

The lower branch consists primarily of the katabatic wind regime, the prevalent feature of the lower boundary layer over Antarctica (Parish et al., 1993; Bromwich and Parish, 1998). Katabatic winds constitute an important mechanism for the transport of mass and cold air northward (King and Turner, 1997; Bromwich and Parish, 1998). The mass flux in the katabatic layer implies in a mass divergence, which must be compensated by warmer subsidence air in the middle troposphere (Parish et al., 1993; King and Turner, 1997). Then, a thermally-direct circulation cell is established and helps to maintain the heat balance of the Antarctic, since cold air is exported at low levels while warmer air from mid-latitudes is advected towards Antarctica at higher levels, subsiding into katabatic layer over the continent (King and Turner, 1997). Convergence in higher levels leads to the development of a cyclonic flow and a circumpolar vortex is established (Parish et al., 1993; King and Turner, 1997).

The intensity of the katabatic winds is proportional to the steepness of the underlying terrain (Bromwich and Parish, 1998). Consequently, the most intense katabatic winds are found near the steep-sloped coast (Bromwich and Parish, 1998; Nylen et al., 2004).

Antarctic surface winds are among the most persistent flow regimes on the Earth (Parish and Bromwich, 1987; Parish and Cassano, 2003). Observations showed that these winds have directional constancy (King and Turner, 1997; Parish and Cassano, 2003), which is also an evidence that the flow is largely controlled by topography (King and Turner, 1997; Bromwich and Parish, 1998). Parish and Bromwich (1987) obtained a qualitative estimate of the near-surface time-averaged winter windfield over the Antarctic continent and showed the strong control of the underlying terrain and the associated directional constancy of the wind.

Parish and Bromwich (1987) shows that the preferred cold-air drainage channels pattern is not uniform, but has a number of confluence zones. These zones represent areas of cold-air supplies, which feed the katabatic winds downstream, and where it becomes anomalously intense and persistent (Bromwich and Parish, 1998).

During summer only weak or non-existent katabatic flow is expected since the temperature inversion is weak or does not exist. However, intense and persistent katabatic winds are observed in the near-coastal region of Adélie Land (Kodama et al., 1989; Parish and Cassano, 2003) and in Terra Nova Bay on the western coast of Ross Sea (Bromwich, 1989). Kodama et al. (1989) and Parish and Cassano (2003) showed that in the near-coastal region of Adélie Land, although the winds are weaker in the summer, they present a

pronounced diurnal cycle and still remain high constancy in wind direction with little seasonal variation.

Therefore, it implies that other factors besides katabatic forcing must be acting ([Parish and Cassano, 2003](#)). [Parish et al. \(1993\)](#) and [Parish and Cassano \(2003\)](#) showed that the ambient horizontal pressure field associated with the passage of cyclones also has an important effect on the Antarctic surface wind regime.

As seen previously, there is a semipermanent baroclinic zone around Antarctic due to the horizontal temperature gradients in the troposphere. The major meridional temperature change occurs at the polar front which separates the warmer mid-latitude air from the cold polar air masses ([King and Turner, 1997](#)). In the Antarctic coastal region, a significant thermal contrast is also present between the colder continental air and the relatively warm maritime air masses ([King and Turner, 1997](#)).

These baroclinic zones result in the development of mesoscale or small synoptic-scale cyclones which can track eastward to the coast or occasionally move northward over the Southern Ocean ([King and Turner, 1997](#)). When these cyclones reach the Antarctic coast they usually move toward the east coast ([King and Turner, 1997](#)) because the continental orography acts as a barrier to cyclones inland progression ([Bromwich and Parish, 1998](#)). As these cyclones move eastward parallel to the coast, the katabatic flow enhances considerably by synoptic-scale circulation ([Murphy and Simmonds, 1993](#); [Bromwich and Parish, 1998](#)).

Antarctic coastal margin is one of the most active cyclogenetic regions on the Earth ([Parish and Cassano, 2003](#)). The low-pressure belt surrounding Antarctic continent and extending to the Southern Ocean are reflective of this highest cyclones frequency ([Bromwich and Parish, 1998](#)).

Pressure gradient between the low-pressure belt, centered at 60–65°S, and the high pressure in subtropics, under influence of Earth's rotation, lead to development of the westerlies, the main forcing of Antarctic Circumpolar Current (ACC) ([Ainley et al., 2009](#)).

The ascending air at 60–65°S moves poleward at upper levels and, sinks over the poles forming a high-pressure system over the Antarctic continent ([Ainley et al., 2009](#)). In this way, as katabatic winds move northward the pressure gradient force decreases abruptly ([Bromwich and Parish, 1998](#)). The airflow adjustment to the offshore circulation is responsible for mass increase, which leads to the development of a northward-directed horizontal pressure gradient force in the lowest level of the atmosphere ([Bromwich and Parish, 1998](#)). This pressure field gives rise to the easterly winds along the Antarctic coast, which drives the Antarctic Coastal Current (ACoC) ([Bromwich and Parish, 1998](#); [Ainley](#)

et al., 2009).

1.3.2 Southern Ocean

Southern Ocean, surrounding Antarctic continent, is the only one with a circumglobal circulation. The absence of land barriers in the latitude band of Drake Passage, between South America and Antarctic Peninsula, has an important influence on dynamics currents in the Southern Ocean (Rintoul et al., 2001). As a result, the strong ACC flows continuously eastward, connecting each of the ocean basins within this band (Talley, 2011; Rintoul et al., 2001).

ACC dominates the Southern-Ocean large-scale circulation and extends over the latitude band of 40–65°S (King and Turner, 1997; Talley, 2011). This current is driven primary by the world’s strongest westerly winds and is also known as the West Wind Drift (Trenberth et al., 1990; Orsi et al., 1995; Talley, 2011).

Surface speed of the ACC varies according with the latitude between $0.4 \text{ m}\cdot\text{s}^{-1}$ at Polar Front (PF) and $0.15 \text{ m}\cdot\text{s}^{-1}$ north of this front (King and Turner, 1997). However, at Drake Passage a convergent flow is present and the current speed reaches up to $1 \text{ m}\cdot\text{s}^{-1}$ (King and Turner, 1997). This current has a volume transport of 130–150 Sv (King and Turner, 1997; Rintoul et al., 2001; Fyfe and Saenko, 2006).

Interbasin connection provided by this current allows the exchange of properties (e.g. heat, water masses) and oceanic teleconnections, which makes possible the transport of anomalies around the globe (Rintoul et al., 2001). Therefore, the Southern Ocean plays an important role in the global climate.

South of ACC, the two largest indentations are present, the Weddell and Ross Seas, which support regional wind-driven gyres with boundary currents (Talley, 2011).

Close to the coast the westward flow of the ACoC driven by easterly winds is present (Deacon, 1937; Núñez-Riboni and Fahrbach, 2009; Talley, 2011), which is the focus of this study and is described in more details below.

Southern Ocean is bounded to the south by the Antarctic continent and its northern boundary is not well defined. Antarctic Treaty Limit at 60°S could be taken as a political northern limit. However, the Southern Ocean oceanographic regime extends well north of 60°S (Talley, 2011). The northernmost excursion of the ACC is at about 38°S (Talley, 2011).

Due the pronounced meridional gradients in the surface properties, the northern limit is also generally determined by the fronts, separating the waters of the Southern Ocean

from the warmer and saltier waters of the subtropical regime (Orsi et al., 1995; King and Turner, 1997).

1.3.2.1 Fronts in the Southern Ocean

Hydrographic boundary between subtropical and subantarctic waters is the Subtropical Front (STF), which is marked by changes in surface temperature and salinity around $4\text{--}5^\circ\text{C}$ and 0.5, respectively (Orsi et al., 1995). This front was first described by Deacon (1937) and according to him the approximate location lies within a band where temperature increases northward from 10°C to 12°C and salinities from 34.6 to 35.0 at 100 m.

South of STF is the ACC eastward flow, which is nearly zonal (Orsi et al., 1995; Talley, 2011). Because of this zonal nature, the properties isopleths in the Southern Ocean are also nearly zonal (Talley, 2011).

Figure 1.4: Fronts and oceanographic zones in the Southern Ocean from Talley (2011).

Isopleths of properties are organized into three major fronts which account for most of the ACC transport (Orsi et al., 1995; Talley, 2011). Fronts encircling Antarctica as part of

the ACC are the Subantarctic Front (SAF), the Polar Front (PF), and the Southern ACC Front (SACCF) (Orsi et al., 1995; Talley, 2011). Southern boundary (SB) of the ACC is marked by abrupt end of the Upper Circumpolar Deep Water (UCDW) characteristic signal (Orsi et al., 1995). Further south are the weakly-sheared circulations of subpolar regimes (Orsi et al., 1995).

SAF is the northern edge of the ACC and is identified at many locations by the occurrence of the 4–5°C isotherm at 200 m depth. PF is within the ACC and is identified by northernmost location of 2°C isotherm surrounding the temperature minimum layer or as the location where the shallow temperature minimum begins to steeply descend toward the north. Finally, the SACCF includes minimum potential temperature less than 0°C at depths shallower than 150m and maximum potential temperature greater than 1.8°C at depths greater than 500 m (Orsi et al., 1995; Talley, 2011).

These fronts are separated by four broad zones: Subantarctic Zone (SAZ; north of SAF), Polar Frontal Zone (PFZ; between SAF and PF), Antarctic Zone (AZ; between PF and SACCF), the Southern Zone (SZ; between the SACC and the SB), and the Subpolar Region (or Subpolar Zone, SZ), which includes the Weddel and Ross gyres (Talley, 2011). The mean fronts location according to Orsi et al. (1995) are shown in Figure 1.4 from Talley (2011). Figure 1.5 also shows an scheme of meridional circulation in the Southern Ocean showing the main fronts and water masses.

South of the ACC we find the Antarctic Slope Front (ASF) at most locations along the continental slope, with a westward flow (Talley, 2011). This front is an important feature for Antarctic coastal circulation and it will also be describe in next sections.

Figure 1.5: Southern Ocean meridional circulation scheme showing the fronts position and water masses. Acronyms: Continental Shelf Water (CSW), Antarctic Surface Water (AASW), Subantarctic Mode Water (SAMW), Subantarctic Surface Water (SASW), and Subtropical Surface Water (STSW) from [Talley \(2011\)](#).

1.3.2.2 Sea Ice in the Southern Ocean

Sea ice is an important component of climate system. It has a major impact on the surface albedo and also plays a role in the exchange of heat, momentum, and gases ([King and Turner, 1997](#); [Talley, 2011](#); [Vaughan et al., 2013](#)).

Although the expansion and retraction of sea ice in the Southern Ocean are governed by the same energy fluxes as Arctic, there are considerable factors that differs them ([Gow and TuckerIII, 1990](#)). Differences are primarily because of their different geography.

In Arctic the sea ice is confined and remains in the cold Arctic waters ([NSIDCa, NSIDCa](#)). The semi-enclosed Arctic ocean makes the floes more prone to converge, which makes the Arctic ice thicker ([NSIDCa, NSIDCa](#)). In addition, these converging ice pile up into thick ridges, which has longer life cycle and leads to ice that stays frozen during summer ([NSIDCa, NSIDCa](#)).

In the 60° – 70° latitude band, where the majority of Antarctic sea ice is found, the energy fluxes are similar to those found in Arctic and there are small differences in shortwave radiation component due the latitude difference ([Gow and TuckerIII, 1990](#)). However, the turbulent exchanges are larger because of strong winds in the region ([Gow and TuckerIII, 1990](#)).

Moreover, Antarctic sea ice leads in an open sea, which provides conditions for a marginal ice zone to exist. Therefore, the sea ice is influenced by the warmer air and ocean temperatures and is more susceptible to the wave action, which tends to break up the ice cover into small floes (Gow and TuckerIII, 1990).

Close to the continent, there is also the influence of the cold katabatic wind and winds associated with cyclones (Gow and TuckerIII, 1990; King and Turner, 1997). When the surface pressure gradient increases as deep lows over the ocean, which moves southward and encounter high pressure associated with the semi-permanent anticyclone over the continent, it leads to the development of strong winds near the coast (King and Turner, 1997). These winds can rapidly thicken ice and affect the surface conditions over the ocean surface and sea ice (King and Turner, 1997).

As a result, sea ice in the Southern Ocean has a wide seasonal variation. It reaches a maximum area of approximately $2 \times 10^7 \text{ km}^2$ in late austral winter. In its maximum extent the edge of the sea ice is close to 60°S around most of the continent, and near 55°S north of the Weddell Sea (King and Turner, 1997; Ainley et al., 2009). During austral summer, unlike the Arctic Ocean, most of sea ice melts and reaches its minimum, which covers an area of about $3 \times 10^6 \text{ km}^2$ (Zwally et al., 1983; King and Turner, 1997; Ainley et al., 2009). Therefore, much of the sea ice in Southern Ocean is first-year-ice and relatively thin (King and Turner, 1997; Talley, 2011). The exceptions are the western Weddell Sea and along the Ross Sea Ice Shelf, where significant amounts of sea ice persist throughout the year (King and Turner, 1997; Talley, 2011).

This pronounced variations in the sea ice results in strong seasonal surface heat budget changes over the Southern Ocean (Bromwich and Parish, 1998). Furthermore, sea ice also has important effects on climate at coastal regions, since heat, fresh water (salt) and momentum fluxes between ocean and atmosphere over the Antarctic continental shelf regions potentially have strong seasonal components (King and Turner, 1997; Hoffmann and Klinck, 1996).

Polynyas

Another important feature, is the presence of polynyas, where waters remains open even when atmospheric conditions are such that the surface water should freeze (Hoffmann and Klinck, 1996). There are two types of polynyas, which are differentiated by formation mechanisms: sensible-heat or latent-heat flux. Both processes can occur at the same location, but one of them will always dominate. In general, on continental shelves latent-

heat polynyas are present, whereas those in the open ocean are sensible-heat polynyas (Hoffmann and Klinck, 1996).

In open ocean, polynyas are formed by sensible-heat transfer between upwelling warmer waters and ice (Hoffmann and Klinck, 1996; NSIDCb, NSIDCb). Formation of sensible-heat polynyas are related with topography of the ocean bottom or overturning of water, which causes warm water to rise to the surface providing heat to melt the ice (Hoffmann and Klinck, 1996; NSIDCb, NSIDCb).

Latent heat polynyas are commonly found around the coastline and at ice shelf edge (Talley, 2011). Latent-heat polynyas are formed as a result of mechanical removal of ice by winds blowing in a persistent direction (Hoffmann and Klinck, 1996; NSIDCb, NSIDCb). In this case the water is at the freezing point, but as new ice grows within polynyas, wind blows and removes it before consolidate into a solid sheet (Hoffmann and Klinck, 1996; NSIDCb, NSIDCb). Even though the ocean temperature does not change, latent heat is released as water freezes and considerable heat is liberated to the atmosphere as water evaporates (Hoffmann and Klinck, 1996; NSIDCb, NSIDCb).

When polynyas are formed, salt is expelled into the water and produces dense shelf waters, some of which is dense enough to create bottom water (Talley, 2011; NSIDCb, NSIDCb). Latent-heat polynyas in coastal regions of Antarctica are the major source of the Antarctic Bottom Water (AABW). Three polynya regions are most productive of AABW: southern Weddell Sea (68%), Ross Sea (8%), and Adélie Land (24%) (Talley, 2011). However, the relationship between sea ice production and dense water formation is not one-to-one (Talley, 2011). There are important characteristics of coastal circulation dynamics that influence the formation of bottom waters and will be discussed in the next section.

By exporting bottom waters the Southern Ocean ventilates the vast majority of water in the rest of the World Ocean (Orsi et al., 1999). Therefore, the formation of AABW and its equatorward flow are key elements since drives the Termohaline Circulation (THC) (Orsi et al., 1999).

1.3.3 The Weddell Sea

This study focused on the Weddell Sea region located in the Atlantic sector of Southern Ocean. This region has an important role on the stratification and abyssal properties of the World Ocean as the Weddell Sea is the main site of AABW production and export in the Southern Ocean (Deacon, 1937; Carmack, 1974; Fahrbach et al., 1995; Orsi et al.,

1999). According to previous studies the Weddell Sea contributes from 50% to 90% for the AABW production (Foldvik and Gammelsrød, 1988; Fahrbach et al., 1995; Orsi et al., 1999).

The major feature of large-scale circulation in Weddell Sea is the cyclonic Weddell Gyre (WG) (Robertson et al., 2002). WG is located south of the ACC, and is divided by the SACCF (Orsi et al., 1995). The WG extends from the Antarctic Peninsula to 20°–30°E (Orsi et al., 1993). Its northern boundary is the Scotia Ridge in the west and then it approximately follows the 4000 and 5000 m bottom contour (Talley, 2011).

The WG transport varies between 30 and 60 Sv (Schröder and Fahrbach, 1999; Fahrbach et al., 1994). This last value was found by Schröder and Fahrbach (1999) in the eastern edge of WG.

The AABW formation in the Weddell Sea is favored by the WG which carries relatively warm and saline water masses from ACC southward (Orsi et al., 1993; Fahrbach et al., 1995). The major inflow of ACC waters occurs in the eastern boundary of the WG (Schröder and Fahrbach, 1999), where topographic effect has important constraints on ocean currents. The Southwest Indian Ridge causes convergence and increase shear, which leads to instabilities which produce intense mesoscale eddy field (Schröder and Fahrbach, 1999). These eddies transport ACC waters into the gyre, and drives water masses transformation (described below) to those of Weddell regime (Schröder and Fahrbach, 1999). In turn, these water masses are carried by the WG to Antarctic continental shelves, where deep and bottom waters are produced (Orsi et al., 1993).

Further south, the ACoC flows. In the Weddell Sea, the V-shaped ASF is evident, and reflects the geostrophic shear with westward flow along shelf break (Talley, 2011). ACoC in the Weddell Sea leaves the coast near Halley Bay (75°S and 27°W) and flows along the edge of the continental shelf as a cold fresh water ribbon (Gill, 1973). In the southwest Weddell Sea, ACoC flows across the shelf toward Filchner-Ronne Ice Shelf and may flow beneath it (Heywood et al., 2004). According to Heywood et al. (1998) in the southeastern Weddell Sea, ACoC accounts for 47% of the total transport (barotropic and baroclinic) of the Weddell Gyre.

1.3.3.1 Water Masses in the Weddell Sea

When water masses are exported to the WG, they undergoes transformations and acquires characteristics of Weddell regime. Moreover, several factors (e.g. temperature,

sea ice) influences water masses properties which differs between regions. For those reasons, in this section we provided a description of Weddell Sea water masses.

Weddell Sea water masses are divided into surface (0–200 m), intermediate (200–1500 m), and deep (>1500 m) layers (Schröder and Fahrbach, 1999; Kerr et al., 2009). This last one is separated into deep and bottom waters by their thermohaline characteristics and age (Kerr et al., 2009). The main water masses are summarized in Table 1.1 and values were obtained from Whitworth et al. (1998); Orsi et al. (1993); Talley (2011); Robertson et al. (2002).

Surface waters consist of Antarctic Surface Water (AASW), Winter Water (WW) and Shelf Waters (SW). In the intermediate, deep and bottom layers the water masses found are Warm Deep Water (WDW), Weddell Sea Deep Water (WSDW), and Weddell Sea Bottom Water (WSBW) (Schröder and Fahrbach, 1999; Kerr et al., 2009).

Surface Waters

South of SAF the cold and fresh AASW is found. Due its location at the air/sea/ice interface, AASW is the most variable in properties of all Antarctic water masses (Whitworth et al., 1998). During summer, AASW has a characteristic near-freezing minimum temperature layer, at depths between 50 m and 200m, remnant of the isothermal winter mixed layer. For this reason this minimum temperature layer is referred as Winter Water (WW) (Carmack, 1990; Whitworth et al., 1998; Talley, 2011). The WW is a common feature observed in open ocean, because it is underlain by warmer CDW (Whitworth et al., 1998).

Over the continental shelf CDW is not present to produce the relative minimum layer. However, south of PF, the SACCF, SB and ASF are also subsurface features, and do not retard exchange of surface waters (Whitworth et al., 1998). For this reason, on shelf waters that are less dense than the regional CDW should be considered as part of AASW (Whitworth et al., 1998).

SW are commonly distinguished between High Salinity Shelf Water (HSSW) and Low Salinity Shelf Water (LSSW) (Carmack, 1990). In the Weddell Sea there a westward salinity increase is detected (Gill, 1973; Carmack, 1974). For this reason, HSSW is found above the broad continental shelf in southwestern Weddell Sea and is also known as Western Shelf Water (WSW). Whereas, LSSW is found in the narrow eastern continental shelf and is also called Eastern Shelf Water (ESW) (Carmack, 1974; Fahrbach et al., 1994). LSSW presents salinities lower than WW which prevents bottom water formation (Gill,

1973).

Gill (1973) attributed high salinity to the presence of coastal polynyas. However, coastal polynyas are a common feature of western and eastern continental shelves. Differences between these two regions remains in bottom topography and currents beneath polynyas (Fahrback et al., 1994). Broad continental shelf in southwestern Weddell Sea restrict polynyas to areas over the shelf and the brine rejection accumulates in shallow layers having small velocities (Fahrback et al., 1994). In contrast, on the extremely narrow eastern continental shelf, polynyas can extend out over deep water. Low salinities of the LSSW are also usually attributed to inflow of relatively-fresh waters from ACoC (Fahrback et al., 1994).

In the continental shelf, Ice Shelf Water (ISW) is also found. ISW is unique to the Antarctic, and is most common in Weddell and Ross Seas (Carmack, 1990). ISW is situated below ice shelves and reaches the greatest mean depths of continental shelf (300–600 m) (Carmack, 1974).

ISW is mainly characterized by temperatures below freezing point at sea level pressure (Carmack, 1974; Carmack and Foster, 1975a; Carmack, 1990). This unusual properties are acquires through cooling at the base of ice shelf because freezing temperature is depressed by increasing pressure and temperatures as low as -2.4° have been observed (Carmack, 1974, 1990).

HSSW and ISW, present high salinities, and are the main contributors to dense water formation in the southwestern Weddell Sea (Kerr et al., 2009).

Intermediate Waters

WDW forms a temperature and salinity maximum at intermediate depths in Weddell Sea . This water mass originates from CDW advected into WG from the ACC (Kerr et al., 2009). Within the WG, CDW mixes with surface waters which alters CDW initial termohaline characteristics and results in the cooler and fresher WDW (Kerr et al., 2009).

CDW is the most voluminous water mass of ACC, and it is derived from deep waters masses from Pacific (Pacific Deep Water; PDW), Atlantic (North Atlantic Deep Water; NADW) and Indian (Indian Deep Water; IDW) Oceans that enter in the ACC where they are mixed to become CDW (Orsi et al., 1999; Fahrback et al., 2004; Talley, 2011).

Table 1.1: Temperature and salinity ranges of the main Weddell Sea water masses.

Water Mass	Acronym	Temperature (°C)	Salinity
Antarctic Surface Water	AASW	-1.9°–4°C	33 – 34.5
Winter Water	WW	<-1.7°C	34.3–34.45
Shelf Water	SW		
Ice Shelf Water	ISW	> -1.9°C	34.2–34.7
High-Salinity Shelf Water	HSSW	< -1.7°C	34.56–34.84
Low-Salinity Shelf Water	LSSW	< -1.7°C	34.3–34.4
Modified Warm Deep Water	MWDW	-1.7°– -0.7°C	34.4–34.6
Warm Deep Water	WDW	0°–1°C	34.6–34.75
Weddell Sea Deep Water	WSDW	-0.7° – 0°C	34.62–34.68
Weddell Sea Bottom Water	WSBW	<-0.7°C	34.62–34.68

CDW is usually divided in two layers: Upper CDW, characterized by an oxygen minimum and nutrients maxima, and the Lower CDW identified as a salinity maximum layer (Orsi et al., 1995, 1999; Whitworth et al., 1998). As seen previous, the end of UCDW marks the ACC southernmost boundary (Orsi et al., 1995). Only LCDW is found south of the ACC (Whitworth et al., 1998).

MWDW (Table 1.1) is the transitional water mass that results from mixing of WDW with surface waters (Fahrbach et al., 1994; Robertson et al., 2002). This water mass is cooler than WDW and at shelf break it mixes with WSBW to form the WSDW (Robertson et al., 2002; Kerr et al., 2009).

Deep Waters

In layers bellow WDW, WSDW and WSBW are found. WSDW presents the same

characteristics of AABW, however it does not fill the bottom layer of Weddell Sea. For this reason, the term WSDW is more appropriate (Fahrback et al., 1995; Schröder and Fahrback, 1999).

WDSW is the mixing product of WDW and/or MWDW with WSBW, formed by deep convection at shelf break (Carmack and Foster, 1975b; Orsi et al., 1993; Fahrback et al., 1995; Kerr et al., 2009). WSBW, in turn, occupies the layer below WSDW and is formed by mixing between WDW and shelf waters (Orsi et al., 1993; Fahrback et al., 1995). WSBW is formed in the southwestern continental shelf, where high salinity shelf waters are found (Foster and Carmack, 1976).

WSBW is topographically constrained and remains at abyssal depths (Kerr et al., 2009). This water mass circulates cyclonically within western portion of WG (Orsi et al., 1993). WSBW's outflows occurs only through by mixing with WSDW (Fahrback et al., 1995), or via deep trenches which allows this water mass fill the bottom layer of the South Sandwich Trench and Georgia Basin (Orsi et al., 1999). Therefore, WSDW is the main source for AABW world ocean.

1.3.4 Antarctic Coastal Region and its Circulation

Continental shelves surrounding the Southern Ocean are generally quite narrow, the exceptions being the two broad (about 400 km) shelves of the Weddell and Ross Seas (Fahrback et al., 1994; Carmack, 1990). Shelf regions are also characterized by irregular depressions, submarine canyons, and by glacial ice shelves which extend seaward from the continent (Carmack, 1990).

Antarctic continental shelf has unique morphological characteristics due the presence of ice in the region. First, the shelf region has a mean depth varying between 400–700 m, about two times deeper than midlatitudes shelves and second the deeper inshore, which potentially restrict exchanges between nearshore and offshore regions (Hoffmann and Klinck, 1996). These characteristics are due the large mass of ice on Antarctic that depresses the entire continent and its shelves (Talley, 2011).

In the midlatitudes, where continental shelves are shallower and has an average depth of about 200 m, the seasonal effects of winds and heat are more pronounced and results in strong seasonal cycle of coastal waters. However, in Antarctic continental shelf besides the presence of the sea ice which has important dynamical constraints, as discussed above, the greater depth in this region reduces those sea seasonal effects to a smaller portion of the water column (Hoffmann and Klinck, 1996). Therefore, deep water is not affected by

these processes and the oceanic character of the bottom water remain intact ([Hoffmann and Klinck, 1996](#)).

Antarctic Coastal Circulation

Prevailing easterly winds along the Antarctic coast induce the coastal downwelling of the cold and salty shelf waters. As a result, an associated subsurface meridional density gradient is present between cold, relatively fresh waters on the continental shelf and warmer, saltier waters farther offshore ([Jacobs, 1991](#); [Spence et al., 2014](#)). This gradient is marked by a strong front known as the ASF. In broad continental shelves ASF is marked by a strong V-shaped front, while in narrow shelves only the northern half 'V' appears ([Gill, 1973](#); [Jacobs, 1991](#)). In some locations, the cold and salty shelf waters at the base of this front meet with the warmer fresher Circumpolar Deep Water (CDW) and mixes to form bottom water ([Gill, 1973](#); [Foster and Carmack, 1976](#); [Whitworth et al., 1998](#)).

The significance of the ASF stems from its prime location, since its the primary site for renewal of Antarctic surface and the deepest layers of the World Ocean ([Jacobs, 1991](#)). Moreover, ASF regulates the properties exchange (e.g. heat, salt, nutrients, gases) on and off continental shelf ([Jacobs, 1991](#)).

The ASF and its associated subsurface density gradient support the cold, low-salinity ACoC westward flow along the Antarctic margin ([Gill, 1973](#); [Hoffmann and Klinck, 1996](#); [Heywood et al., 1998, 2004](#); [Spence et al., 2014](#)) ([Figure 1.6](#)). Where the continental shelf is broad, ACoC separates from ASF and travels over the shelf. However, in some areas, where continental shelf is very narrow, like in the southeast Weddell Sea, is difficult to distinguish the locations of ACoC and ASF ([Heywood et al., 1998, 2004](#)).

The ACoC flows with velocities between 0.1 ms^{-1} and 0.3 ms^{-1} ([Sverdrup, 1953](#); [Gill, 1973](#); [Fahrbach et al., 1992](#); [Núñez-Riboni and Fahrbach, 2009](#)), south of the belt of low atmospheric pressure that extends around the continent at roughly 65°S ([Carmack, 1990](#)).

The ACoC was first described by [Deacon \(1937\)](#) and is also called as the East Wind Drift. Although it is largely observed in the Antarctic coast, this current is not completely circumpolar ([Sverdrup, 1953](#)). In the Weddell and Ross Seas, this current becomes part of the cyclonic gyres ([Carmack, 1990](#)).

Figure 1.6: Schematic of Antarctic continental shelf cross section at Weddell Sea, illustrating the Antarctic Slope Front (ASF) and its associated westward flow driven by the easterlies. The illustration also shows the water masses and its interaction in Antarctic coastal region: Ice Shelf Water (ISW), High-Salinity Water (HSSW), Antarctic Surface Water (AASW), Warm Deep Water (WDW), Weddell Sea Deep Water (WSDW) and, Weddell Sea Bottom Water (WSBW).

According to [Sverdrup \(1953\)](#), the coastal current is mainly barotropic and its westward geostrophic flow is a result of the downsloping isotherms towards the coast forced by wind-induced Ekman transport. [Fahrbach et al. \(1992\)](#) showed through the correlation between the seasonal cycle of wind and current speeds that winds are indeed an important driving force of the coastal current. However, stratification seasonal variations were also observed, which imply in a seasonal change of the baroclinic current field.

[Núñez-Riboni and Fahrbach \(2009\)](#) quantified the contribution of ACoC driving mechanisms to the seasonal variability of its barotropic and baroclinic components and concluded that coastal current has mainly a barotropic transport. However, there is still a significant contribution to the variability of baroclinic field. Therefore, changes that have been observed in Antarctic (see section 1.1) can affect the dynamics of coastal circulation.

[Spence et al. \(2014\)](#) showed through numerical simulations that the strengthening and shift poleward of westerlies associated with the SAM positive trend, observed since 1950s, reduce the meridional extent and strength of the polar easterlies. As a result, a decrease on the onshore Ekman Transport and in its associated downward Ekman pumping near coast are observed, as well as the weakening of the coastal currents and ASF. Hence, an

intense warming of subsurface coastal waters (between 200 and 700 m depth) occur due to the CDW intrusion and exceeds 2°C (Spence et al., 2014).

The relatively warm, mid-depth Southern Ocean Waters are the largest oceanic heat source for reduce the extent and thickness of ice shelves, driving basal melting which reduces buttressing and accelerates glacier flow (Pritchard et al., 2012; Hattermann et al., 2014; Schmidtko et al., 2014). Consequently, interactions between ocean and ice shelves have a potential impact on Antarctic ice sheet loss and hence, on sea level rise (Rignot et al., 2011; Shepherd et al., 2012; Schmidtko et al., 2014; Spence et al., 2014).

The ACoC and the associated ASF, have an important role on the oceanographic regime around Antarctic continent (Fahrback et al., 1992). Therefore, changes in the coastal dynamics can have a significant impact on processes that affects the global climate.

1.4 Hypothesis

The scientific hypothesis of this study is that climate change can impact the circulation of the Weddell Sea from the end of the 20th century to the early 21st century.

1.5 Objectives

This study aims to characterize the circulation of the Weddell Sea at the end of the 20th century (1980 to 1999) and for a period of the 21st century (2021 to 2047) using high resolution simulation results.

The specific objectives were:

- Assess the behavior of wind stress, temperature and salinity fields, which are important for circulation dynamics, from a large scale perspective through the use of oceanic component of the Community Earth System Model version 1.0 (CESM1);
- Evaluate if 20th century regional simulations with ROMS were capable to reproduce coastal circulation in the Weddell Sea;
- Characterize the coastal circulation in the Weddell Sea at the end of 20th century;
- Investigate whether significant changes occurred in the coastal circulation from late 20th century (1980 to 1999) to 21st century (2021 to 2047);
- Investigate the contribution of wind versus thermohaline forcing for changes in the coastal circulation.

2 Material and Methods

2.1 Models and Data

In this study we evaluated results from models and reanalysis, and also used an observational data, which are described in this section.

2.1.1 Regional Ocean Model System (ROMS)

To understand the global climate impacts on a regional-to-local scale, downscaling is necessary. Downscaling allows the tracking and understanding of particular ocean features and their intrinsic variability related to climate changes introduced as system forcings (Pereira and Wainer, 2015). The complementarity between global-regional-local scales and the comparison (when possible) with observations becomes essential for the understanding of regional processes (Pereira and Wainer, 2015).

Therefore, to characterize the coastal circulation in the Weddell Sea we used numerical simulations based on Regional Ocean Model System (ROMS) for the South Atlantic and Southern Oceans, with implementation of version 19c, developed by Ocean, Climate and Cryosphere Laboratory (OC2) group from the Oceanographic Institute of University of São Paulo.

ROMS is a state-of-the-art hydrostatic, free-surface ocean general circulation model that can be configured for any region of the world ocean ranging from local to basin scale (Moore et al., 2004, 2009). The model numerics is described in by Shchepetkin (2003); Shchepetkin and McWilliams (2005).

Simulations with ROMS have a curvilinear horizontal grid with horizontal resolution varying between 7 and 20 km, favoring the Atlantic Southwest (10 km) and Weddell Sea (7 km) regions. Spatial and temporal resolutions of the domain used was calculated with a compromise between the meso scale (eddy resolving) resolution and numerical performance for long climate (secular) integrations (Pereira and Wainer, 2015).

For the model physical consistency and climatic stability, the numerical formulation used includes the coupling with a sea ice model (described by Budgell (2005)), which ensures an accurate physical representation of high latitude ocean (Pereira and Wainer,

2015).

ROMS outputs were interpolated to a regular horizontal resolution grid of $0.1417^\circ \times 0.1436^\circ$ and 18 vertical levels with a frequency of 2 days (except ROMS-20C3M, described below, which has a 4-day temporal resolution). The vertical resolution is not uniform and varies from 2.5 m at the surface to 500 m in the bottom.

In this study three experiments with ROMS have been analysed in the Weddell Sea region, that comprises the area between 80°S and 60°S and 63°S and 40°E (Figure 2.1).

Figure 2.1: Study area map with AWI233 mooring location.

The first simulation was forced with National Center for Environmental Prediction/ National Centers for Atmospheric Research (NCEP/NCAR) Reanalysis I and used Simple Ocean Data Assimilation version 2.4.4 (SODA2.2.4) described in Carton et al. (2005); Carton and Giese (2008)), as initial conditions and lateral boundary. This simulation is considered as the control run centered in the 20th century and will be referred to as ROMS-NCEP.

The other two were forced with CCSM3 experiments of IPCC AR4/CMIP3. The first will be named ROMS-20C3M and refers to the simulation forced with the CCSM3 AR4 historical scenario, the Climate of the 20th Century Experiment (20C3M). The last simulation used in this study is forced with an experiment of CCSM3 which considers the scenario A2 from SRES, and will be referred to as ROMS-SRESA2.

In this study we used monthly mean data for the period of 1980 to 1999 referring to 20th century simulations and 2021 to 2047 to 21st projection.

2.1.2 Auxiliary Data

2.1.2.1 Ocean Component of the Community Earth System Model (CESM) – POP2

The Community Earth System Model (CESM) is a coupled climate model for simulating Earth’s climate system developed by National Center for Atmospheric Research (NCAR). composed of five separate components that simulate the Earth’s atmosphere, ocean, land, land-ice, and sea-ice ([Vertenstein et al., 2010](#)).

The components include the Community Atmosphere Model version 5 (CAM5; [Neale et al. \(2012\)](#)), the Community Land Model version 4 (CLM4; [Lawrence et al. \(2011\)](#); [Oleson et al. \(2010\)](#)), the Parallel Ocean Program, version 2 (POP2; [Smith et al. \(2010\)](#)), the Community Ice Sheet Model (CISM; [Lipscomb et al. \(2013\)](#)) and the Community Ice code, version4 (CICE4; [Bailey et al. \(2011\)](#)).

Ocean Component:

Part of the results of this study were obtained from the CESM1 ocean component, POP2 of Los Alamos National Laboratory. The POP2 is a level-coordinate ocean general circulation model that solves the three-dimensional primitive equations with hydrostatic and Boussinesq approximations for ocean dynamics ([Smith et al., 2010](#); [Danabasoglu et al., 2012](#)).

CESM1–POP2 has a horizontal resolution of $1^\circ \times 1^\circ$ and 60 levels. The vertical resolution is uniform at 10 m in the upper 160 m, and vertical grid spacing increases to 250 m by a depth of about 1500 m ([Danabasoglu et al., 2012](#)). This integration (OCN) is forced with the Coordinated Ocean-Ice Reference Experiments version 2 (CORE–II; [Griffies et al. \(2009\)](#)).

CORE proposes a protocol to performing global ocean-ice coupled simulations forced with a common atmospheric data set ([Griffies et al., 2009](#); [Danabasoglu et al., 2014](#)). CESM1–POP2 was performed in the second phase of COREs, which uses an interannual variability atmospheric forcing from [Large and Yeager \(2009\)](#) ([Griffies et al., 2009](#); [Danabasoglu et al., 2014](#)). This dataset combines reanalysis and remote sensing products and is considered more reliable than the corresponding NCEP reanalysis data ([Griffies et al., 2009](#); [Downes et al., 2015](#)). CORE–II simulations are usually referred to as hindcast experiments ([Danabasoglu et al., 2014](#)).

Dataset available covers the period of 1948–2015. However, in the present study we used the monthly mean variables for the same period of the regional simulations (1980 to 1999).

2.1.2.2 Community Climate System Model (CCSM)

The Community Climate System Model (CCSM) is a coupled climate model developed by National Center for Atmospheric Research (NCAR). The CCSM consists of four components for the atmosphere, ocean, sea ice, and land surface linked through a coupler that exchanges fluxes and state information among these components (Collins et al., 2006).

Here, we used the CCSM3 version 3. The CCSM3 includes the Community Atmosphere Model, version 3 (CAM3; Collins et al. (2006)), the Community Land Surface Model version 3 (CLM3; Oleson et al. (2004); Dickinson et al. (2006)), the Community Sea Ice Model version 5 (CSIM5, Briegleb et al. (2004)), and the ocean is based on the Parallel Ocean Program version 1.4.3 (POP; Smith and Gent (2002)).

Two of the regional simulations were forced with CCSM3 IPCC AR4 experiments. Therefore, considering wind as one of the main coastal circulation forcing, the 10 m winds of atmospheric component of CCSM3 (CAM3) were also evaluated.

CAM3 represents the sixth generation of the atmospheric general circulation models (AGCMs) developed by the climate community in collaboration with the NCAR (Collins et al., 2006). CAM3 is integrated along with the CLM, the CSIM5, and the POP (Collins et al., 2006).

CCSM3 (CAM3) outputs have a horizontal grid of $1.41^\circ \times 1.41^\circ$ and a daily temporal resolution. CCSM3 (20C3M), the historical simulation, comprises the period from 1870 to 1999, and the CCSM3 (SRESA2) covers the period from 2000 to 2099. However, we used the monthly mean 10 m winds for periods in common with regional experiments (1980–1999 and 2021–2047, respectively).

2.1.2.3 National Centers for Environmental Prediction (NCEP) – Reanalysis

1

The NCEP/NCAR Reanalysis 1 was also used to force one of the regional simulations. Therefore, the NCEP 10 m winds were also evaluated.

The NCEP/NCAR Reanalysis 1 project is using a state-of-the-art analysis/forecast system to perform data assimilation with past data from 1948 to the present (Kalnay et al., 1996; Kistler et al., 2001). The project development has been supported by the National

Oceanic and Atmospheric Administration (NOAA) and involved an effort to recover land surface, ship, rawinsonde, pibal, aircraft, satellite, and other data (Kalnay et al., 1996; Kistler et al., 2001). The reanalysis assimilation system is described in detail by Kalnay et al. (1996).

The reanalysis has a horizontal grid of $2.5^\circ \times 2.5^\circ$ and the products are available in the original 4 times daily format and as daily averages. Here, we used monthly means for the period from 1980 to 1999.

2.1.2.4 Ocean Reanalysis – SODA3.4.1

In this study, we used the ocean reanalysis, Simple Ocean Data Assimilation version 3.4.1 (SODA3.4.1) to compare and evaluate the consistency of CESM1 (POP2).

SODA is an ocean reanalysis data set consisting of gridded variables for the global ocean. Its goal is to provide an improved estimate of ocean state compared to other reanalysis based solely on observations or numerical simulations (Carton and Giese, 2008).

SODA version 3 represents a major upgrade of SODA. Previous versions of SODA have used an ocean general circulation model based on POP numerics (Carton et al., 2005; Carton and Giese, 2008) and in the version 3 the model has been switched to the Modular Ocean Model (MOM) coupled with the Sea Ice Simulator (SIS) of Geophysical Fluid Dynamics (GFDL) with eddy permitting $1/4^\circ \times 1/4^\circ \times 50\text{lev}$ resolution (28 km at the Equator down to < 10 km at polar latitudes) (Carton et al., 2018).

MOM is a hydrostatic generalized level coordinate with mass conserving non-Boussinesq or volume conserving Boussinesq kinematics (Griffies, 2012). Possible vertical coordinates are described in more details by Griffies et al. (2000); Griffies (2012).

The main datasets assimilated by SODA are the World Ocean Database of historical hydrographic profiles, and in-situ and remotely sensed SST from the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) (Carton et al., 2018).

In this study we used the SODA3.4.1 which is forced by ERA-Interim atmospheric reanalysis from the European Centre for Medium-Range Weather Forecasts (ECMWF). The monthly ocean state variables are mapped onto the regular $1/2^\circ \times 1/2^\circ$ Mercator horizontal grid 50 z^* vertical levels. The dataset is available for the 1980 to 2015 period, but in this study the analyses were performed over the 1980 to 1999 period.

2.1.2.5 Observational Data

To determine whether the simulations with ROMS reproduce the ACoC satisfactorily, we used the mooring AWI-233 dataset from Alfred Wegener Institute ([Fahrbach and Rohardt, 2012](#)) to compare with the simulations centered in 20th century.

The mooring is located at the Antarctic continental slope, near 0° and 69.4°S ([Figure 2.1](#)), and provides current data with a frequency of one measurement each 2 h from February 1997 to March 2008. The moorings had to be changed occasionally during this period and, because of the rough topography, ice conditions and strong currents, they could not be redeployed exactly at the same locations and depths. Although the locations and depths of the instruments changed slightly, they can be considered as negligible ([Núñez-Riboni and Fahrbach, 2009](#)). The mean instrument depths at mooring AWI-233 in a water depth of 1975 m are 247, 744 and 1925.

For the comparison with regional simulations we used the period of 1998–1999 with a four day frequency.

2.2 Data Analysis

2.2.1 Volume Transport

Volume transport was derived from the horizontally and vertically integrated zonal and meridional velocities, expressed by [Equation 2.1](#):

$$\begin{aligned}\Gamma_{Sv_y}(x,t) &= \int_{-H}^0 \int v(x,t) \cdot dz dx \\ \Gamma_{Sv_x}(y,t) &= \int_{-H}^0 \int v(x,t) \cdot dz dy \\ \Gamma_{Sv}(x,y,t) &= \sqrt{\Gamma_{Sv_y}^2 + \Gamma_{Sv_x}^2}\end{aligned}\tag{2.1}$$

where dx and dy are the model horizontal grid resolution, and dz the vertical resolution; u and v are the horizontal velocities, and $-H$ is the maximum depth.

2.2.2 Barotropic and Baroclinic components

[Fahrbach et al. \(1992\)](#) and [Núñez-Riboni and Fahrbach \(2009\)](#) showed that although ACoC is mostly barotropic, there is still a variability in the baroclinic field. In order to

investigate changes in coastal circulation, we evaluated the variability of the barotropic and baroclinic components.

According to [Núñez-Riboni and Fahrbach \(2009\)](#), the barotropic component can be defined as the contribution of the sea-surface slope to the geostrophic velocity field ([Equation 2.2](#)):

$$\begin{aligned} u_g &= -\frac{g}{f_0} \frac{\partial \eta}{\partial y} \\ v_g &= \frac{g}{f_0} \frac{\partial \eta}{\partial x} \end{aligned} \tag{2.2}$$

where, g is the gravitational acceleration ($g=9.8\text{m}\cdot\text{s}^{-2}$), f_0 the Coriolis acceleration, and $\partial\eta/\partial x$ and $\partial\eta/\partial y$ are the zonal and meridional Sea Surface Height (SSH) variation, respectively.

To determine the baroclinic component we subtracted the barotropic component from the total velocity field. Alternatively, the baroclinic component can be defined as the contribution of the density shear ([Núñez-Riboni and Fahrbach, 2009](#)).

2.2.3 Climatological Analysis

To describe climate patterns and determine whether there will be a significant change in the mean fields of analyzed variables (wind stress, temperature, salinity and volume transport), we conducted a climatological analysis.

Climatological analysis consists of averaging a variable over a period of time. Climatologies were calculated for the entire period and for seasonal fields (summer and winter). Further, in the case of volume transport we computed the climatological year. This allows us to describe the annual cycle of the variable.

Results are presented in climatological maps and in cross sections for selected subregions in the Weddell Sea ([Figure 2.2](#)).

Figure 2.2: Selected subregions in the Weddell Sea for analysis.

Table 2.1: Subregions Areas

Areas	Latitude	Longitude
A1	76°S – 66°S	58°W – 28°W
A2	70°S – 66°S	6°W – 12°E
A3	73°S – 62°S	58°W – 50°W

For CESM1 (POP2), we only analyzed the subregions A1 and A2.

2.2.4 Differences

Climatological fields were initially compared through simple differences. This analysis was used to compare:

- The model CESM1 (POP2) with SODA3.4.1 reanalysis, to evaluate how different the numerical model is from the reanalysis and to show, in a simple way, regions with higher correlations;
- The ROMS (20C3M) with ROMS (NCEP), with the same objective as the previous item;
- The ROMS (SRESA2) with ROMS (20C3M), to highlight changes from the 20th to the 21st century.

2.2.4.1 Student's T-test

To evaluate how significant the differences are, we applied a student's t-test for means. This is a two-sided test for the null hypothesis that the 2 independent samples have identical average (expected) values.

The t-test is given by [Equation 2.3](#):

$$t = \frac{\bar{x} - \mu_0}{\frac{s}{\sqrt{n}}} \quad (2.3)$$

where, $\bar{x}$ is the sample mean, μ_0 hypothesized population mean, s is the sample standard deviation, and n the sample size.

The t distribution is used to find the p -value. In this study, we checked whether the differences were statistically significant, using a threshold value (α) of 0.05.

2.2.5 T-S Diagram

To characterize the water masses in the Weddell Sea, we used the temperature-salinity (TS) diagrams. This also allowed us to determine whether the models are capable of reproducing the water mass structure in the region.

2.2.6 Standard Deviation

Standard deviation is a measure of the dispersion of a set from its mean. Low standard deviations indicate that data points are close to the mean, while high values indicate that data points are spread out.

Standard deviation is calculated as the square root of variance ([Equation 2.4](#)):

$$\sigma = \sqrt{\frac{1}{n-1} \sum_{i=1}^n (X_i - \bar{X})^2} \quad (2.4)$$

Standard deviations were computed to evaluate uncertainties in the anomaly time series.

2.2.7 Anomaly

To investigate the dispersion of the variables not related with seasons, we calculated the anomaly. Anomaly measures the deviation from the climatological mean and is computed as [Equation 2.5](#):

$$\alpha_{i,m}(k) = X_{i,m} - \bar{X}_m(k) \quad (2.5)$$

where i , m and k indexes indicate the year, month, and location, respectively. Positive anomalies indicate deviations above the mean, and negative anomalies indicate deviations below.

2.2.8 Trends

Trend describes the behaviour of a variable, reflecting its increase, decrease or stability over time, which allows to determine whether long-term changes will occur ([Chatifield, 2013](#)).

The trend was calculated with the least squares method, which explains the behaviour of the variable through a linear function ([Equation 2.6](#)). This method is similar to linear regression, with time as the independent variable for time series.

$$\hat{y} = a + bt \quad (2.6)$$

$\hat{y}$: trend;

a : intercept (linear coefficient);

b : angular coefficient (indicates if the trend is increasing or decreasing);

t : time value.

Angular coefficient (b) is obtained with the least square method, given by [Equation 2.7](#), and the linear coefficient (a) is calculated with [Equation 2.8](#):

$$b = \frac{\sum_{i=1}^n (t_i \times y_i) - \frac{1}{n} \sum_{i=1}^n t_i \times \sum_{i=1}^n y_i}{\left(\sum_{i=1}^n t_i^2 \right) - \frac{1}{n} \left(\sum_{i=1}^n t_i \right)^2} \quad (2.7)$$

$$a = \frac{\sum_{i=1}^n y_i - b \times \sum_{i=1}^n t_i}{n} \quad (2.8)$$

where y_i is the variable value in time t_i .

2.2.8.1 Mann-Kendall Test

We applied the non-parametric Mann-Kendall (MK) test ([Mann, 1945](#); [Kendall, 1975](#)) to determine whether there is a monotonic upward or downward trend in the variable of interest.

MK tests the null hypothesis that the data are independent and identically distributed (no monotonic trend), and is based on Kendall rank correlation to test for monotonic trend.

Kendall's rank correlation coefficient (τ) measures the strength of monotonic association between vectors x and y and is expressed by [Equation 2.9](#):

$$\tau = \frac{S}{D} \quad (2.9)$$

where S , is the Kendall score given by:

$$S = \sum_{i < j} (\text{sign}(x[j] - x[i]) \text{sign}(y[j] - y[i])) \quad (2.10)$$

where sign denotes the sign function and D is the maximum possible value of S .

$$D = \frac{n(n-1)}{2} \quad (2.11)$$

In MK test the Kendall's rank is used to autocorrelate a time series, where every value is compared to every preceding value. In this case, S is given by:

$$S = \sum_{i=1}^{n-1} \sum_{j=i+1}^n \text{sign}(x[j] - x[i]) \quad (2.12)$$

where,

$$\text{sign}(x) = \begin{cases} 1, & \text{if } x > 0; \\ 0, & \text{if } x = 0; \\ -1, & \text{if } x < 0 \end{cases}$$

MK test statistic is given by:

$$Z = \begin{cases} \frac{S-1}{\sigma} & \text{if } S > 0; \\ 0, & \text{if } S = 0; \\ \frac{S+1}{\sigma}, & \text{if } S < 0 \end{cases}$$

where σ is the square root of S variance. A positive (negative) value of Z indicates that the data tend to increase (decrease) with time.

To determine whether the trends were statistically significant the p -value was calculated using a threshold value (α) of 0.05.

2.2.9 Correlation

To assess the dependency between the ACoC and its forcings, we used a correlation analysis. Correlation can be interpreted as the existing linear regression between two variables.

We used the Pearson's correlation coefficient (r), that measures the degree of linear relation between two quantitative variables and the shared variance between them. Pearson's correlation coefficient is given by [Equation 2.13](#):

$$r = \frac{\sum_{i=1}^n (x_i - \bar{x})(y_i - \bar{y})}{\sqrt{\left(\sum_{i=1}^n (x_i - \bar{x})^2\right) \left(\sum_{i=1}^n (y_i - \bar{y})^2\right)}} \quad (2.13)$$

Coefficient r varies between -1 and 1. A value of 0 indicates that the relation between variables is diffuse or not linear, a correlation value of 1 indicates a perfect linear relation, and a value of -1 corresponds to a perfect inverted linear relation, i.e., one of the variables increase while the other decreases.

Correlation coefficients were tested using the Student's t-test, considering the null hypothesis that there is no correlation. In this case, t is given by [Equation 2.14](#):

$$t = \frac{r\sqrt{n-2}}{\sqrt{1-r^2}} \quad (2.14)$$

with $n - 2$ degrees of freedom. In this case we used a threshold value (α) of 0.05.

3 Results and Discussion

3.1 The Ocean General Circulation Model

In order to understand circulation changes in a regional scenario, we first evaluate and characterize the Southern Ocean circulation using an Ocean General Circulation Model (GCM). The behavior of the wind stress, temperature and salinity fields, which are important for the circulation dynamics, from a large-scale perspective are evaluated. In this study we use the monthly mean data fields between 1980 to 1999 from the ocean component (POP2) of the NCAR–CESM1, hereafter referred to as CESM1–POP2. The results obtained from CESM1–POP2 are compared with the SODA3.4.1 ocean reanalysis.

3.1.1 Wind Stress

The wind stress magnitude climatology obtained from CESM1–POP2 and SODA3.4.1 is shown in [Figure 3.1](#). In agreement with the description in [Talley \(2011\)](#), the mean wind forcing is dominated by westerlies in the latitude band 40°–60°S and by easterlies closer to Antarctica, south of 60°S.

Greater magnitudes associated with the westerlies occurs in the east Antarctic from 0° to 180°W in both climatologies. The maximum values are found in the Indian Ocean sector, near 90°E, where the westerlies reaches about $0.22\text{N} \cdot \text{m}^{-2}$. This locally maximum over the southern Indian Ocean was also found by [Trenberth et al. \(1990\)](#), who found values above $0.3\text{N} \cdot \text{m}^{-2}$.

Figure 3.1: **Wind stress magnitude $[\text{N}\cdot\text{m}^{-2}]$ climatologies calculated for the 1980–1999 period. (a) CESM1–POP2 climatology, (b) SODA3.4.1 climatology and (c) significant differences, at 95% level, between CESM1–POP2 and SODA3.4.1 climatologies.**

This pattern is more pronounced during the winter months, as we can see in [Figure 3.2](#). In this season the maximum wind stress magnitude exceeds $0.25 \text{ N}\cdot\text{m}^{-2}$. This result is in agreement with [Risien and Chelton \(2008\)](#).

In the southern regions of Africa ($\approx 10^\circ\text{--}30^\circ\text{E}$) and Australia ($\approx 120^\circ\text{--}150^\circ\text{E}$) the wind stress magnitudes reaches more than $0.15 \text{ N}\cdot\text{m}^{-2}$ which corroborate with [Trenberth et al. \(1990\)](#). According to them, the meridional wind stress exceeds $0.1 \text{ N}\cdot\text{m}^{-2}$ in these two regions, which could be sufficient to affect Southern Ocean dynamics.

In the summer climatologies ([Figure 3.2a](#) and [Figure 3.2b](#)) a secondary maximum of about $0.15 \text{ N}\cdot\text{m}^{-2}$ is found in the Pacific sector, and extends from 120°W to 70°W reaching South America. This pattern could be attributed to the influence of Southeast Pacific anticyclone which reaches its southernmost position with its center at $\approx 35^\circ\text{S}$ during summer ([Rahn and Garreaud, 2014](#)).

Greater wind stress magnitudes in East Antarctic coastal region are also observed, and are more pronounced during winter. In the CESM1–POP2 winter climatology ([Figure 3.2d](#)) it reaches magnitudes that exceed $0.2 \text{ N}\cdot\text{m}^{-2}$ in Adélie Land coast ($\approx 145^\circ\text{E}$). Magnitudes of about $0.15 \text{ N}\cdot\text{m}^{-2}$ are found between Queen Mary Land ($\approx 110^\circ\text{E}$) and Adélie Land ($\approx 145^\circ\text{E}$), and in $\approx 60^\circ\text{E}$.

Figure 3.2: **Seasonal wind stress magnitude $[\text{N}\cdot\text{m}^{-2}]$ climatologies calculated for the 1980–1999 period.** CESM1–POP2 summer (DJF) (a) and winter (JJA) (d) climatologies; SODA3.4.1 summer (DJF) (b) and winter (JJA) (e) climatologies; differences between CESM1–POP2 and SODA3.4.1 summer (c) and winter (f) climatologies.

These higher magnitudes near the Antarctic continent could be due to katabatic winds. these winds reach their maximum during wintertime and are proportional to the steepness of the underlying terrain (Bromwich and Parish, 1998; Talley, 2011). Therefore, the most intense surface wind conditions are expected to be found in East Antarctica (Bromwich and Parish, 1998), which explains partially the difference in the wind stress magnitude of East and West coasts of Antarctica.

Figures 3.1c, 3.2c, and 3.2f shows differences between the CESM1–POP2 and SODA3.4.1 climatologies. Greater values in the wind stress magnitude are detected in the CESM1–POP2 climatologies in the westerlies region, around $\approx 64^\circ\text{S}$, reaching a difference of about $0.05 \text{ N}\cdot\text{m}^{-2}$ relative to SODA3.4.1.

The maximum difference (about $0.1 \text{ N}\cdot\text{m}^{-2}$ bigger than SODA3.4.1) during winter is observed in the coastal region between the Queen Mary Land ($\approx 110^\circ\text{E}$), where the Shackleton Ice Shelf and Oates Land ($\approx 170^\circ\text{E}$) coasts are located. This difference is clear in the winter climatologies (Figures 3.2d and 3.2e), where the model CESM1–POP2 shows greater magnitudes.

The coastal region between $\approx 10^\circ\text{W}$ and 20°E , that encompasses the Fimbul Ice Shelf, is the region where the CESM1–POP2 showed smaller wind stress magnitudes compared to SODA3.4.1. During winter the magnitudes are almost $0.1 \text{ N}\cdot\text{m}^{-2}$ lower than the SODA3.4.1 magnitude.

3.1.1.1 Trends and Variability in the Wind Stress

Figure 3.3 shows the linear trends calculated for the wind stress magnitude (τ) and for the zonal wind stress (τ_x). As seen before, in this region the wind stress is nearly zonal. Therefore, trends in the wind stress magnitude are directly proportional to zonal wind stress magnitudes.

Significant positive trends, at 95% confidence level, are mainly observed in the latitude band associated with the westerlies, between approximately 50°S – 60° , in both the model and the reanalysis. In CESM1–POP2 (Figures 3.3a and 3.3c) the trend is slightly higher and more pronounced in the Indian sector ($\approx 30^\circ$ – 60°E) and south of Australia in the Pacific sector ($\approx 140^\circ\text{W}$ – 170°E), where we observe the maximum positive trend. We observe a trend of $3.25 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$ in wind stress magnitude in this region, whereas for zonal wind stress, the trend reaches $4.73 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$.

In SODA3.4.1 (Figures 3.3b and 3.3d) the most pronounced trends are observed in South America ($2.88 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$) and in the Indian sector, in 120°E and 57°S , with trends of $2.6 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$ in the wind stress magnitude, and $2.82 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$ in the zonal component.

Figure 3.3: **Linear trend maps in the Southern Ocean for the 1980–1999 period** [$\text{N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$]. CESM1-POP2 wind stress magnitude (τ) (a) and zonal wind stress (τ_x) (c) linear trends; SODA3.4.1 wind stress magnitude (τ) (b) and zonal wind stress (τ_x) (d) linear trends. Significant linear trends, at 95% level, are marked with black dots.

From the Antarctic Peninsula to further south we observe a decreasing trend in wind stress magnitude mainly in CESM1-POP2 (Figure 3.3a). In the region close to the continent, wind stress is dominated by the easterlies (negative values of wind stress), therefore the decrease in magnitude is an indication for positive trends. The maximum decrease in wind stress magnitude is found in the East Antarctic with a significant negative trend, at 95% level, of $-2.7 \times 10^{-3} \text{N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$. For the zonal wind stress (Figure 3.3c) the positive trend has the same magnitude.

A significant increasing trend, at 95% level, near the coast is observed only at the Amundsen Sea region, with a trend of $1.54 \times 10^{-3} \text{N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$ in the wind stress magnitude and $-1.75 \times 10^{-3} \text{N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$ in the zonal wind stress.

In SODA3.4.1 (Figures 3.3b and 3.3d) decreasing trend south of Antarctic Peninsula is also observed. However, trends are smaller ($\approx 1.4 \times 10^{-3} \text{N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$) and significant (at 95% level) in few regions. Near approximately 67°S and 70°E significant increasing trend of $2.19 \times 10^{-3} \text{N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$ in wind stress magnitude (Figure 3.3b) and $-1.43 \times 10^{-3} \text{N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$ in zonal wind stress (Figure 3.3d) are observed.

To determine whether there is evidence of poleward shifting, we evaluate the latitudinal profile for the zonal mean in the region of the Weddell Sea, between 63°W and 40°E . The profiles were computed every 5 years from 1980 to 1999 and the results are shown in Figure 3.4a for CESM1–POP2, and in Figure 3.5a for SODA3.4.1. The annual mean position of maximum magnitude associated with the westerlies is also shown in Figure 3.4b (CESM1–POP2) and Figure 3.5b (SODA3.4.1).

In CESM1–POP2 (Figure 3.4) we also observe a gradual increase in the maximum magnitude associated with the westerlies from 1980 to 1999. With respect to the annual mean position (Figure 3.4b), small difference from the beginning to the end of the analyzed period is observed. In 1980 the maximum wind stress is centered at 47.5°S , while in 1999 it is located 1° south. However, over the analyzed period there were some years where the position of maximum is centered south of 52°S , with the southernmost position reaching 57.5°S in 1982.

Figure 3.4: **(a)** Latitudinal zonal wind stress profiles in the zonal mean between 63°W and 20°E obtained from CESM1–POP2 every 5 years over the 1980–1999 period; **(b)** CESM1–POP2 annual latitudinal mean position of the zonal wind stress maximum magnitude .

Figure 3.5 shows the results obtained for the SODA3.4.1. Results shows an increase in zonal wind stress magnitude (Figure 3.5a). The zonal wind stress mean between 1990–1994 is slightly higher than in 1995–1984. Regarding its mean position (Figure 3.5b), the southernmost latitude reached was 52°S in 1981 and 1992.

Figure 3.5: **(a)** Latitudinal zonal wind stress profiles in the zonal mean between 63°W and 20°E obtained from SODA3.4.1 every 5 years over the 1980–1999 period; **(b)** SODA3.4.1 annual latitudinal mean position of the zonal wind stress maximum magnitude.

The zonal wind stress anomaly was analyzed, i.e. we investigated the wind stress variability without seasonal oscillations relative to the 1980 to 1999 period. The analyses were performed in the Weddell Sea region for four latitudes: 50°S, 55°S, 65°S, and 70°S (Figure 3.6).

In the four latitudes where the zonal wind stress anomaly was analyzed, CESM1–POP2 shows greater linear trends than SODA3.4.1 (up to two orders of difference). We applied the Mann-Kendall test to evaluate the significance of linear trends. Although linear trends are greater in CESM1–POP2, they are not significant at 95% level, in any of the analyzed latitudes whereas in SODA3.4.1, linear trends are significant at 95% level in almost all latitudes, except in 70°S.

Figure 3.6: Zonal wind stress linear trend map in the Weddell Sea region from CESM1-POP2 over the 1980 to 1999 period. The dashed lines marks the latitudes where the anomalies were calculated.

Figure 3.7 shows the zonal wind stress anomaly at 50°S . Anomalies up to $\pm 0.06 \text{ N}\cdot\text{m}^{-2}$ in CESM1-POP2 (Figure 3.7a) and between -0.051 and $0.07 \text{ N}\cdot\text{m}^{-2}$ in SODA3.4.1 (Figure 3.7b) are detected. These most pronounced anomalies were observed during winter and spring months in both model reanalysis.

In both zonal wind stress anomaly is observed a small increasing trend. The anomaly linear trend obtained from CESM1-POP2 was $1.03 \times 10^{-3} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$, while in SODA3.4.1 (Figure 3.7b) it shows a linear trend of $4.62 \times 10^{-5} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$.

Figure 3.7: **Zonal wind stress anomalies over the 1980–1999 period in the Weddell Sea region at 50°S.** (a) zonal wind stress anomaly relative to CESM1–POP2 and (b) zonal wind stress anomaly relative to SODA3.4.1. Uncertainties are indicated by coloured shading.

At 55°S anomalies do not exceed $\pm 0.1 \text{ N} \cdot \text{m}^{-2}$ in SODA3.4.1 (Figure 3.8b), while in CESM1–POP2 (Figure 3.8a) anomalies slightly above this value are found. At this latitude, the anomalies peaks were observed in the spring months.

A increase trend in both model and reanalysis is also observed. In CESM1–POP2 the linear trend obtained was $1.03 \times 10^{-3} \text{ N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$, and in SODA3.4.1 it was $7.25 \times 10^{-5} \text{ N} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$.

Figure 3.8: **Zonal wind stress anomalies over the 1980–1999 period in the Weddell Sea region at 55°S.** (a) zonal wind stress anomaly relative to CESM1–POP2 and (b) zonal wind stress anomaly relative to SODA3.4.1. Uncertainties are indicated by coloured shading.

At 65° anomalies are slightly higher in CESM1–POP2 (Figure 3.9a) than in SODA3.4.1 (Figure 3.9b). Anomalies reach up to $0.08 \text{ N}\cdot\text{m}^{-2}$ in CESM1–POP2, while in SODA3.4.1 the maximum observed was $0.06 \text{ N}\cdot\text{m}^{-2}$. These positive maximum anomalies were observed in June 1989 in CESM1–POP2 and in August 1989 in SODA3.4.1. Negative anomalies do not exceeds $-0.05 \text{ N}\cdot\text{m}^{-2}$ in both model and reanalysis.

A small increasing linear trend is also observed in both model and reanalysis. In CESM1–POP2 (Figure 3.9a) it is one order smaller than in further north latitudes ($9.8 \times 10^{-4} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$). In SODA3.4.1 (Figure 3.9b) trend is smaller compared to the latitude of 65°S, but is the second greatest linear trend observed in reanalysis ($6.05 \times 10^{-5} \text{ N}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$).

Figure 3.9: **Zonal wind stress anomalies over the 1980–1999 period in the Weddell Sea region at 65°S.** (a) zonal wind stress anomaly relative to CESM1–POP2 and (b) zonal wind stress anomaly relative to SODA3.4.1. Uncertainties are indicated by coloured shading.

The last latitude analyzed was at 70°S, where easterlies are predominant. The positive anomalies do not exceed $0.05 \text{ N}\cdot\text{m}^{-2}$, but negative anomalies are around $-0.07 \text{ N}\cdot\text{m}^{-2}$ in both model and reanalysis. The most pronounced negative anomalies were detected in the spring and summer months in both model and reanalysis. At this latitude trends are not detected.

Figure 3.10: **Zonal wind stress anomalies over the 1980–1999 period in the Weddell Sea region at 70°S. (a)** zonal wind stress anomaly relative to CESM1–POP2 and **(b)** zonal wind stress anomaly relative to SODA3.4.1. Uncertainties are indicated by coloured shading.

The zonal wind stress presented greater positive anomalies in almost all latitudes, except at 70° where negative anomalies are more pronounced. Overall, CESM1–POP2 anomalies showed a satisfactory agreement relative to SODA3.4.1.

Anomalies are overall, around $\pm 0.05 \text{ N}\cdot\text{m}^{-2}$, except at 55°S where anomalies up to $\pm 0.1 \text{ N}\cdot\text{m}^{-2}$ are found. Greater anomalies were found in the winter and spring months, except at 70°S where negative anomalies peaks were also observed in the summer months.

3.1.2 Sea Surface Temperature and Salinity

The SST climatologies are shown in Figure 3.11. South of approximately 60°S, the SST around Antarctica reaches values near the freezing point. The minimum temperature found in CESM1–POP2 (Figure 3.11a) was -1.6°C , while in the reanalysis SODA3.4.1 the minimum value was -1.8°C .

Although SST climatologies do not look very different around Antarctica, [Figure 3.11c](#) shows that the model is up to 2°C higher than the reanalysis north of Antarctic Peninsula and in the Pacific sector ($\approx 145^{\circ}\text{S}$). For the other regions around Antarctic, SST differs up to 1°C .

Comparing with the World Ocean Circulation Experiment (WOCE) climatology ([Figure 3.13a](#)), the CESM1–POP2 and SODA3.4.1 SST climatologies show agreement in the expected mean patterns.

Figure 3.11: SST [$^{\circ}\text{C}$] climatologies calculated for the 1980–1999 period. (a) CESM1–POP2 climatology; (b) SODA3.4.1 climatology; (c) significant differences, at 95% level, between CESM1–POP2 and SODA3.4.1 climatologies.

Figure 3.12: SSS climatologies calculated for the 1980–1999 period. (a) CESM1–POP2 climatology; (b) SODA3.4.1 climatology; (c) significant differences, at 95% level, between CESM1–POP2 and SODA3.4.1 climatologies.

The SSS climatologies are shown in [Figure 3.12](#). In the SODA3.4.1 result ([Figure 3.12b](#)) a circumpolar pattern of the minimum salinity is observed in the region of ACC, with salinities between 33.6–34. The minimum values, around 33.6, are detected in the Pacific sector, around 145°W, until Drake Passage from approximately 60°S to south. This lower salinity extends near the continent in the Amundsen and Bellingshausen seas. In these regions salinities around ≈ 33.2 can be seen very close to the coast.

In the CESM1–POP2 ([Figure 3.12a](#)) the circumpolar pattern is not so evident. The region of minimum salinity is in agreement with SODA3.4.1. However, in the CESM1–POP2, the salinity band around 33.6 is found further south, and extends to the Ross Sea region. These differences are also shown in [Figure 3.12c](#). In the region of the Pacific sector, in the latitude band of the Drake Passage, positive differences of up to 0.4 are observed, which implies that higher salinities are derived from CESM1–POP2. South of this latitude band, negative differences around -0.6 are found.

According to [Talley \(2011\)](#), because of the open zonal passage and the nearly zonal ACC, isopleths of all properties in the Southern Ocean are nearly zonal to great depth. South of the ACC, the surface properties and circulation are organized by the cyclonic gyres in the Weddell and Ross Seas, and are not as zonal as in the ACC latitude band ([Talley, 2011](#)).

Figure 3.13: **WOCE climatologies at 50 m depth.**(a) Potential temperature [°C], (b) salinity. Source: WOCE Southern Ocean Atlas from [Orsi and Whitworth \(2005\)](#).

This difference in the distribution of ocean properties is more evident in the SST climatologies ([Figure 3.11](#)) and in SODA3.4.1 SSS climatology ([Figure 3.12b](#)).

In the SST and SSS climatologies (Figure 3.11 and Figure 3.12, respectively), of both the model and the reanalysis, a clear transition between subantarctic and subtropical regimes is observed. In the SST climatologies a band of temperatures of about 10°C is found, and in the SSS climatologies an increase in the salinity of up to 34.5 is observed, which is also consistent with WOCE climatologies (Figure 3.13).

The largest differences between CESM1–POP2 and SODA3.4.1 are found in SSS climatologies in the Weddell and Ross Seas regions. The reanalysis (Figure 3.12b) shows salinities around 34.3 in both regions. In the Weddell Sea, this higher salinity is observed in the full extension of the WG.

In CESM1–POP2 climatology (Figure 3.12a) salinity values do not exceed 34 in the Weddell Sea, except in a small region in the northern Antarctic Peninsula. In the east of Weddell Sea, salinities reach values of ≈ 33.5 and differs by up to 0.5 (Figure 3.12c) from the SODA3.4.1 climatology.

In the Ross Sea, lower salinities occurs in the western portion with values around 33.6 and reaching even lower values near the coast (≈ 33.2). In a small area east of Ross Sea, salinities around 34.3 are found and extend westward to Adélie Land coast. This is the region where CESM1–POP2 showed higher salinities than SODA3.4.1 close to the continent, with a difference around 0.4.

The pattern observed in the SODA3.4.1 SSS climatology is in agreement with the WOCE climatology (Figure 3.13b), which shows higher salinities, around 34.2, in Weddell and Ross Seas.

The seasonal variations of surface properties are shown in Figure 3.14 (SST) and Figure 3.15 (SSS). CESM1–POP2 SST summer climatology (Figure 3.14a) shows higher temperatures than SODA3.4.1 (Figure 3.14b). In SODA3.4.1 temperatures near the freezing point occurs in almost the entire Antarctic. Temperatures reach values slightly above zero only in East Antarctic. The SODA3.4.1 minimum temperature in summer was -1.7°C, and is observed up to $\approx 63^\circ\text{S}$.

In the CESM1–POP2, the Weddell and Ross Seas are the only regions where temperatures reach values below zero during summer with minimum values of -1.3°C. Figure 3.14c shows that CESM1–POP2 temperature is more than 4°C higher than SODA3.4.1. SODA3.4.1 has a higher temperature only in a small area in the east Weddell Sea, where the CESM1–POP2 reaches minimum temperature values.

Figure 3.14: **Seasonal SST [°C] climatologies for the 1980–1999 period.** CESM1–POP2 summer (DJF) **(a)** and winter (JJA) **(d)** climatologies; SODA3.4.1 summer (DJF) **(b)** and winter (JJA) **(e)** climatologies; differences between CESM1–POP2 and SODA3.4.1 summer **(c)** and winter **(f)** climatologies.

The winter climatologies (Figures 3.14d and 3.14e) reflect the maximum sea ice extent. In both climatologies, temperatures near freezing point extend almost to 50°S. Figure 3.14f shows that around Antarctic the difference between the model and the reanalysis is null.

Downes et al. (2015) investigated sea ice in the Southern Ocean in CORE-II ocean-ice models simulations. In their study, it is shown that CESM1–POP2 has greater differences during summer. The minimum sea ice extent (SIE) in the model is $\approx 1.8 \times 10^{-6} \text{ km}^2$, while observation of SIE reaches $\approx 4.4 \times 10^{-6} \text{ km}^2$. Therefore, warmer temperatures observed in the CESM1–POP2 during summer are probably related to a higher melting rate.

During winter CESM1–POP2 sea-ice extend shows a good representation with ≈ 18.3

$\times 10^{-6} \text{ km}^2$, whereas observation of SIE is $\approx 18.8 \times 10^{-6} \text{ km}^2$.

Figure 3.15: **Seasonal SSS climatologies for the 1980–1999 period.** CESM1–POP2 summer (DJF) (a) and winter (JJA) (d) climatologies; SODA3.4.1 summer (DJF) (b) and winter (JJA) (e) climatologies; differences between CESM1–POP2 and SODA3.4.1 summer (c) and winter (f) climatologies.

In the SSS seasonal climatologies (Figure 3.15), we notice that significant differences between the model and the reanalysis occurs mainly during the summer. The CESM1–POP2 summer climatology (Figure 3.15a) reaches minimum salinity around 33.2 in the west coast of the Antarctic, in the Weddell Sea, and east of the Weddell Sea. In the other regions, the salinity does not exceed 33.6, except in the Oates Land coast ($\approx 140^\circ\text{E}$) and in a small region of Ross Sea.

In the SODA3.4.1 summer climatology (Figure 3.15b), we observe values around 33.2 only in the Bellingshausen Sea, mainly near the west Antarctic Peninsula coast. The

salinities around Antarctic remain around 33.6, but even in the summer marked differences are found in the Weddell and Ross Seas, where salinities exceed 34. In the region of northern Antarctic Peninsula salinities around 34.2 are found in a small region. Further south in the east coast, a region where ice shelf collapse events have been reported, we observe a decrease to 33.6.

In winter, CESM1–POP2 (Figure 3.15d) shows a pattern similar to SODA3.4.1 (Figure 3.15e). In the Weddell Sea, SODA3.4.1 reaches a salinity of almost 34.6, while in the CESM1–POP2 salinity do not exceed 34.2. In CESM1–POP2, salinities around 34.6 are found only in the region near the east coast of Ross Sea.

In the region of Bellingshausen Sea, we note that the salinity band around 33.6 is closer to the coast in the CESM1–POP2 climatology. Moreover, even during winter, salinity around 33.2 is observed near the coast of the west Antarctic Peninsula only in the SODA3.4.1 climatology (Figure 3.15e). The differences between winter climatologies (Figure 3.15f) do not exceed 0.4 around Antarctic.

3.1.3 Water Masses in the Weddell Sea

The results of the previous section show that the Weddell Sea is the region with the greatest differences in sea surface properties, mainly in the salinity. The Weddell Sea is the main site of AABW formation and constitutes our greatest region of interest.

Figure 3.16: **TS Diagram in the Atlantic Ocean sector of the Southern Ocean:** Subantarctic Zone (SAZ; red), Polar Front Zone (PFZ; blue), and Weddell Sea Zone (magenta) (a) TS diagram obtained from WOCE data in October 1993 by Talley (2011), (b) TS diagram reproduced using CESM1–POP2 for the 1980–1999 period.

Initially, we compared [Talley \(2011\)](#) TS diagram ([Figure 3.16a](#)) with CESM1–POP2. [Talley \(2011\)](#) used data from WOCE to produce the TS diagram, which takes into account three stations in the Atlantic Ocean sector of the Southern Ocean in October 1993: the Subantarctic Zone, the Polar Front Zone, and the Antarctic Zone (Weddell Sea).

Overall, CESM1–POP2 TS diagram ([Figure 3.16b](#)) shows a good representation of water mass structure. To summarize the results, the comparison between CESM1–POP2 TS diagram ([Figure 3.16b](#)) and [Talley \(2011\)](#) TS diagram ([Figure 3.16a](#)) are given from [Table 3.1](#) to [Table 3.3](#), where we show the approximate values of temperature, salinity, and density associated with each water mass represented.

In SAZ, all water masses in CESM1–POP2 are less dense than those observed in the [Talley \(2011\)](#) TS diagram (see [Table 3.1](#)). Greater differences are found in surface waters (Subantarctic Surface Water – SASW and Subantarctic Mode Water– SAMW) and intermediate water (Antarctic Intermediate Water – AAIW).

In CESM1–POP2 ([Figure 3.16b](#)), SASW covers a lower range of temperature and salinity. It shows warmer waters in both minimum and maximum temperatures, and reaches smaller maximum values of salinity. SASW is found up to 500 m of upper water column north of the SAF ([Talley, 2011](#)). According to [Talley \(2011\)](#) its temperature ranges from 4°–10°C in winter to 14°C in summer. Its salinity varies between 33.9–34 during winter and <33 in summer. Salinity values in both [Talley \(2011\)](#) ([Figure 3.16a](#)) and CESM1–POP2 ([Figure 3.16b](#)) TS diagrams exceed the maximum value of salinity.

SAMW is found within the SASW in a very thick mixed layer. In the South Atlantic, this layer is of the order of 200 m ([Talley, 2011](#)). In the CESM1–POP2 TS diagram ([Figure 3.16b](#)) SAMW shows warmer temperatures and lower values of salinity than those observed in [Figure 3.16a](#).

Table 3.1: Temperature, salinity and density obtained from [Talley \(2011\)](#) and from CESM1–POP2 in SAZ

Water Mass	Talley (2011)			CESM1–POP2		
	Temperature (°C)	Salinity	Density (kg·m ⁻³)	Temperature	Salinity	Density (kg·m ⁻³)
SAMW	9.7°C	34.72	26.78	10.8°C	34.68	26.41
SASW	5.5°–11°C	34.3–34.88	26.8–27.08	7.8°–11.5°C	34.45 – 34.72	26.3–26.8
AAIW	3.2°C	34.24	27.25	4.2°C	34.42	27.15
UCDW	2.6°C	34.54	27.55	2.5°C	34.6	27.51
LCDW	1.8°C	34.84	27.82	2.3°C	34.86	27.71

AAIW lies in a layer between 500 and 1500 m of depth north of the SAF ([Talley, 2011](#))

and constitutes a low salinity water mass. AAIW is known to be formed by the sinking of Antarctic Surface Water (AASW) across the SAF as a result of northward Ekman transport (Talley, 2011).

According to Emery (2001), AAIW has temperatures between 2°–6°C and salinities ranging from 33.8 to 34.8. Therefore, although AAIW has a higher salinity in CESM1–POP2, and is 1°C warmer than that observed in Talley (2011), the values representing AAIW are in agreement with temperature and salinity values expected for this water mass.

Regarding UCDW and LCDW, the values obtained from CESM1–POP2 are similar to those observed by (Talley, 2011).

In the PFZ (Table 3.2), water masses are less dense in CESM1–POP2. UCDW and LCDW constitute the water masses that show smaller density differences. However, the temperature indicates great differences of 0.6°C in UCDW and 0.5°C in LCDW. In LCDW, salinities values are close, while in UCDW they differ by almost 0.1.

Table 3.2: Temperature, salinity and density obtained from Talley (2011) and from CESM1–POP2 in PFZ

Water Mass	Talley (2011)			CESM1–POP2		
	Temperature (°C)	Salinity	Density (kg·m ⁻³)	Temperature	Salinity	Density (kg·m ⁻³)
AASW	2.7°–6.2°C	33.92–34.14	26.68–27.15	2.6°–5°C	33.85–33.93	26.62–26.92
WW	2.51°C	34.15	27.25	2.4°C	34.96	26.95
UCDW	2.4°C	34.62	27.65	1.8°C	34.53	27.5
LCDW	2°C	34.78	27.8	1.5°C	34.75	27.68

Comparison between CESM1–POP2 and Talley (2011) for AZ (in the Weddell Sea region) are shown in Table 3.3.

AASW in CESM1–POP2 is not well represented. This water mass has a wide range of temperatures and can reach 4°C (see Table 1.1). Furthermore, minimum temperatures do not reach the freezing point as observed by Whitworth et al. (1998) and Talley (2011).

The WW also shows higher temperatures than that observed by Talley (2011) and differs by 0.4°, while its salinity differs by 0.2. Moreover, the temperature and salinity are respectively above and below the values reported in Table 1.1.

WDW shows the same value of salinity as observed by Talley (2011), but a higher temperature. The corresponding temperature is within the range reported in Table 1.1.

WSDW shows values close to those found by Talley (2011), and in agreement with those found by other authors (Table 1.1).

The temperature value estimated for WSBW in the model is close to the maximum temperature limit found in the literature (see [Table 1.1](#)). Furthermore, the model shows a difference of $0.14 \text{ kg}\cdot\text{m}^{-3}$ from [Talley \(2011\)](#) density. Therefore, WSBW is not well represented in the model.

Table 3.3: Temperature, salinity and density obtained from [Talley \(2011\)](#) and from CESM1–POP2 in AZ (Weddell Sea)

Water Mass	Talley (2011)			CESM1–POP2		
	Temperature ($^{\circ}\text{C}$)	Salinity	Density ($\text{kg}\cdot\text{m}^{-3}$)	Temperature	Salinity	Density ($\text{kg}\cdot\text{m}^{-3}$)
AASW	$-2^{\circ}\text{--}3^{\circ}\text{C}$	33.75–34.25	27.05–27.56	$-1.6^{\circ}\text{--} -1.2^{\circ}\text{C}$	33.88–34.08	27.13–27.3
WW	-2°C	34.3	27.62	-1.6°C	34.1	27.32
WDW	0.5°C	34.68	27.82	0.8°C	34.68	27.68
WSDW	-0.25°C	34.68	27.85	-0.2°C	34.65	27.7
WSBW	-1°C	34.65	27.86	-0.7°C	34.62	27.72

It's important to highlight that the results obtained by [Talley \(2011\)](#) correspond to only one month (October 1983) while the results obtained from CESM1–POP2 covers the entire period (1980–1999). The LCDW and WDW correspond to the water masses best represented in the model.

To provide a better evaluation of CESM1–POP2 water masses structure in the Weddell Sea, we performed an analysis in two subregions named A1 and A2 in [Figure 2.2](#), and the results are shown below.

3.1.3.1 Area – A1

Area A1 comprises the Weddell basin and is located between 66°S – 76°S and 28°W – 58°W ([Figure 2.2](#)).

Temperature and salinity cross sections are shown in [Figure 3.17](#) and [Figure 3.18](#), respectively. Following the water masses described in section [1.3.3.1](#), we observe higher temperatures at the surface in CESM1–POP2 ([Figure 3.17a](#)) associated with AASW. This water mass is limited to the minimum temperature layer of WW with temperatures around -1.65°C observed in depths around 100 m. In SODA3.4.1 ([Figure 3.17b](#)), these surface waters are found in a thin layer that extends from the surface down to 50 m, with temperatures below -1.5°C .

With respect to salinity, CESM1–POP2 (Figure 3.18a) has values varying from 34 to 34.43 in the upper 100 m. SODA3.4.1(Figure 3.18b) has a smaller salinity gradient at the surface layer, ranging from 34.2 to 34.43. In both the model and the reanalysis, the boundary of WW is marked by the 34.43 isohaline. However, in CESM1–POP2 the layer relative to surface WW is thicker and we observe the salinity gradient between 34.36 and 34.43. In SODA3.4.1, although salinity values are the same, this gradient occurs in a thinner layer (up to 50 m) and the temperature and salinity associated with surface waters are found in smaller depths. As a result, the subsequent isotherms and isohalines also occur in shallower regions.

Transition between surface waters and WDW occurs in the layer bellow surface. In both the model and the reanalysis, temperatures range from approximately -1.65°C to 0°C and salinities range between 34.43 and 34.57 within this layer. According to the literature (Table 1.1) the MWDW, which is the mixing product between these water masses, is marked by temperatures of up to -0.7°C .

These transitional waters reach approximately 210 m further north and the corresponding isolines slope down towards the continental shelf. However, it's important to highlight that in SODA3.4.1 the southern boundary is located at 74.7°S and the continental shelf topography is not represented in this region. This is why these waters are observed until 800 m depth.

Figure 3.17: **Longitudinal mean temperature [$^{\circ}\text{C}$] cross sections in subregion A1 for the 1980–1999 period. (a) cross section obtained from CESM1–POP2; (b) cross section obtained from SODA3.4.1.**

The most pronounced difference between CESM1–POP2 and SODA3.4.1 lies in the temperature associated with WDW. In both the model and the reanalysis, the temperature associated with this water mass reaches depths bellow 2000 m (Figure 3.17).

The results obtained from CESM1–POP2 (Figure 3.17a) show higher temperatures, reaching up to 1°C . Whereas in SODA3.4.1 (Figure 3.17b), temperatures do not exceed 0.5°C . With respect to salinity, the reanalysis (Figure 3.18b) shows higher values around 34.7 in the deeper layers, while in CESM1–POP2 (Figure 3.18a) values are around 34.65.

Temperatures associated with WSDW and derived from CESM1–POP2 (Figure 3.17a) and SODA3.4.1 (Figure 3.17b) show similar values varying between -0.5°C and 0°C . Salinity values are also similar, however SODA3.4.1 (Figure 3.18b) reaches higher salinities in deeper layers. The corresponding values range from 34.6 to 34.69, while in CESM1–POP2 (Figure 3.18a) they vary between 34.6 and 34.64.

Temperatures associated with WSDW shows close values between CESM1–POP2 (Figure 3.17a) and SODA3.4.1 (Figure 3.17b) varying between -0.5°C and 0°C . The salinity also showed similar values, however SODA3.4.1 (Figure 3.18b) reaches higher salinities in deeper layer. It ranges from 34.6 to 34.69, while in CESM1–POP2 (Figure 3.18a) it varies between 34.6–34.64.

Figure 3.18: **Longitudinal mean salinity cross sections in subregion A1 for the 1980–1999 period.** (a) cross section obtained from CESM1–POP2; (b) cross section obtained from SODA3.4.1.

The seasonal temperature and salinity cross sections are shown from Figure 3.19 to

[Figure 3.22](#). Differences between the model and the reanalysis are found in the surface layer.

In the summer climatology, AASW reaches temperatures above zero in CESM1–POP2 ([Figure 3.19a](#)) at the surface, while in SODA3.4.1 ([Figure 3.20a](#)) temperatures do not exceed -1°C .

As described in section [1.3.3.1](#), the characteristic AASW minimum temperature layer associated with WW is found during summer. This layer is clearly observed in CESM1–POP2 between 50 and 100 m depth. However, it is absent in SODA3.4.1. In the reanalysis, waters below -1.5°C are observed in a thin and small feature in the upper 50 m, that extends to 70°S ([Figure 3.20a](#)).

In the winter climatologies, temperatures near freezing point are found up to 100 m in CESM1–POP2 ([Figure 3.19b](#)) and in the first 50 m in SODA3.4.1 ([Figure 3.20b](#)), which implies a thicker ice layer in the model.

With respect to salinity, values range from 34 to 34.43 at the surface during summer in both the model ([Figure 3.21a](#)) and the reanalysis ([Figure 3.22a](#)). However, in CESM1–POP2 a thicker layer between the 34.36 and 34.43 isohalines is observed. In SODA3.4.1 the salinity gradient occurs within 50 m.

In winter, the differences in surface salinities between CESM1–POP2 and SODA3.4.1 are more pronounced. In CESM1–POP2 ([Figure 3.21b](#)) values range from 34.29 to 34.4 in the upper 100 m, whereas in SODA3.4.1 ([Figure 3.22b](#)) salinity values are of the order of 34.43 in the entire layer.

Figure 3.19: **Seasonal longitudinal mean temperature cross sections in subregion A1 obtained from CESM1–POP2 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference (DJF - JJA).**

In the layers below the surface, the differences and similarities between the reanalysis and the model are similar to those described above for the climatologies obtained for the entire period (Figure 3.17 and Figure 3.18).

Figure 3.20: **Seasonal longitudinal mean temperature cross sections in subregion A1 obtained from SODA3.4.1 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference (DJF - JJA).**

In Figure 3.19c temperatures up to 1°C higher can be seen in the summer at the surface. At 100 m depth positive difference of about 0.14°C is observed. In the layers below, there is a negative difference associated with the transitional layer, which implies higher temperatures during winter. However, the differences are very small, i.e. of the order of 10^{-2} °C.

In SODA3.4.1 (Figure 3.20c) the maximum difference observed is 0.3°C at the surface.

Below that, there are very small differences of the order of 10^{-2}°C .

Figure 3.21: Seasonal longitudinal mean salinity cross sections in subregion A1 obtained from CESM1–POP2 for the 1980–1999 period: **(a)** summer (DJF); **(b)** winter (JJA); **(c)** seasonal difference (DJF - JJA).

Salinity in both CESM1–POP2 (Figure 3.21c) and SODA3.4.1 (Figure 3.22c), show a layer of negative differences in the upper level which indicates freshening during summer. Salinities at the surface differ by up to 0.1. Below that, a positive difference that extends down to 100 m is found in CESM1–POP2 and reaches differences of about 0.1. In SODA3.4.1 a positive difference layer is also observed. Its maximum difference is about

0.05 and reaches a depth of 200 m. However, in SODA3.4.1 (Figure 3.22c) this layer does not extend to all latitudes. In the southern boundary (74.8°S), negative differences of about 0.03 are found.

Figure 3.22: Seasonal longitudinal mean salinity cross sections in subregion A1 obtained from SODA3.4.1 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference (DJF - JJA).

Figure 3.23 shows the TS diagrams obtained from CESM1–POP2 (3.23a) and SODA3.4.1 (3.23b) results over the 1980 to 1999 period. The scatter diagram represents the temperature and salinity points over time for each depth point. The solid black line represents

the mean.

In order to evaluate the water mass structure, the range of temperature and salinity values associated with each one reported in the literature (Table 1.1) were delimited in the graphs.

Figure 3.23: Scatter TS diagram in subregion A1 for the 1980–1999 period. (a) TS diagram obtained from CESM1 (POP2); (b) TS diagram obtained from SODA3.4.1. The solid black line represents the mean temperature and salinity in each depth point over the 1980–1999 period.

As seen in Figure 3.17, the temperatures associated with WDW are higher in CESM1–POP2 than in SODA3.4.1. This is also evident in the TS diagrams, which shows temperatures up to 1°C in CESM1–POP2 (Figure 3.23a) and 0.5°C in SODA3.4.1 (Figure 3.23b).

The TS diagrams also show variations of the surface waters, where the points are scatter. For the other water masses they are close of the mean values. This characteristic is more evident in CESM1–POP2 (Figure 3.23a).

3.1.3.2 Area – A2

The subregion A2 is located in the eastern Weddell Sea, between 66°S–70°S and 6°W–12°E (Figure 2.2). Cross sections climatologies for the period from 1980 to 1999 are

shown in Figure 3.24 (temperature) and Figure 3.25 (salinity). At the surface, CESM1–POP2 (Figure 3.24a and Figure 3.25a) show colder and fresher waters than SODA3.4.1 (Figure 3.24b and Figure 3.25b). In CESM1–POP2 the temperature associated with surface waters ranges from -1.25°C to -1.65°C , and salinities range from <34 to 34.17 . In SODA 3.4.1, these values vary between -0.5° – -1.25°C and 34.25 – 34.33 .

In the area A2 the surface waters occupies a thicker layer than in area A1, but this layer is still shallower in the reanalysis. In the model it extends up to 150 m depth and in the reanalysis up to 100 m.

Figure 3.24: **Longitudinal mean temperature cross sections in subregion A2 of Weddell Sea for the 1980–1999 period.** (a) cross section obtained from CESM1 (POP2); (b) cross section obtained from SODA3.4.1.

The transitional layer shows larger differences between the model and the reanalysis than in area A1. It covers a thinner layer in CESM1–POP2 which is marked by temperatures varying between -1.65° and 0° (Figure 3.24a) and salinities varying between 34.17 and 34.4 (Figure 3.25a). In SODA3.4.1, these values vary between -1.25°C – 0°C and 34.33 – 34.5 (Figure 3.24b and Figure 3.25b).

In region A2, WDW also extends to depths below 2000 m and its temperature are even higher than in area A1 since it comprises an oceanic region. In CESM1–POP2 (Figure 3.24a) the maximum temperature reached is close to 1.25°C and in SODA3.4.1 (Figure 3.24b) it is up to 0.7°C . In both the model and the reanalysis, salinities range from 34.57 to 34.66 (Figure 3.25).

Figure 3.25: **Longitudinal mean salinity cross sections in subregion A2 of Weddell Sea for the 1980–1999 period.** (a) cross section obtained from CESM1 (POP2); (b) cross section obtained from SODA3.4.1.

As in subregion A1, the WSDW temperatures are similar in the model and the reanalysis and vary between 0°C and -0.65°C (Figure 3.24). The salinity is slightly higher in the reanalysis, i.e. around 34.7, whereas in the model it is about 34.66 (Figure 3.25).

The seasonal variability is shown from Figure 3.26 to Figure 3.29. Unlike region A1, the warmer surface waters during summer are found in SODA3.4.1 (Figure 3.27a) with temperature values between 0.1°C and approximately -0.85°C . Waters below that are found in a small core near 69°S . In region A2, the layer of WW is also absent in SODA3.4.1 but is represented in CESM1–POP2 (Figure 3.26a) and can be found between approximately 50 and 150 m.

The absence of WW layer in SODA3.4.1 results in a salinity gradient in the upper 100 m (Figure 3.29a). For this reason, as in the subregion A1, the isohalines are found in shallower depths than in CESM1–POP2 (Figure 3.28a). In CESM1–POP2 a thicker layer of salinities associated with WW that ranges from 34.17 to 34.25 is observed.

During winter, CESM1–POP2 also shows a thicker layer of temperatures near freezing point (Figure 3.26b) than SODA3.4.1 (Figure 3.27b). However, the greatest differences are found in the surface salinities. In CESM1–POP2 (Figure 3.25a) salinity values are around 34 while in SODA3.4.1 (Figure 3.25b) they are close to 34.4.

Below surface waters, the differences between the model and the reanalysis are the same

as described above for the climatologies of the entire period (Figure 3.24 and Figure 3.25).

Figure 3.26: Seasonal longitudinal mean temperature cross sections in subregion A2 obtained from CESM1–POP2 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference.

With respect to seasonal variability, the greatest differences occur at the surface and, below the surface, are smaller than in region A1. In CESM1–POP2, summer temperatures are higher at the upper layer (up to 50 m) and differ by up to 0.94°C (Figure 3.26c). In the layer below approximately 150 m, winter temperatures are warmer and we observe a negative difference around -0.66°C .

Figure 3.27: **Seasonal longitudinal mean temperature cross sections in subregion A2 obtained from SODA3.4.1 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference.**

With respect to salinity, differences are also concentrated in the first 200 m (Figure 3.28c) with values around 0.1. Higher salinities are observed in the summer in the layer between 50 and 120 m. Above and below that, salinities are higher during winter.

In SODA3.4.1, seasonal differences also occur down to 200 m. Temperatures are approximately 0.96°C warmer in summer (Figure 3.26c). Summer salinities are lower in the first 50 m, where they differ by ≈ 0.1 , and are ≈ 0.07 higher below that.

Figure 3.28: Seasonal longitudinal mean salinity cross sections in subregion A2 obtained from CESM1–POP2 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference (DJF - JJA).

Figure 3.29: Seasonal longitudinal mean salinity cross sections in subregion A2 obtained from SODA3.4.1 for the 1980–1999 period. (a) summer (DJF); (b) winter (JJA); (c) seasonal difference (DJF - JJA).

TS diagrams in subregion A2 are shown in Figure 3.30. The results shows a satisfactory representation of the water masses structure according to the literature (Table 1.1), and the main biases are the same discussed above in the cross sections description.

Figure 3.30: **Scatter TS diagram in subregion A2 for the 1980–1999 period.** (a) TS diagram obtained from CESM1 (POP2); (b) TS diagram obtained from SODA3.4.1. The solid black line represents the mean temperature and salinity in each depth point for the 1980–1999 period.

A remarkable difference between CESM1-POP2 and SODA3.4.1 is the thickness of surface layer. In CESM1-POP2 it is thicker and it's possible to distinguish the WW layer from the AASW during summer. In SODA3.4.1 this layer is thinner and there is not a striking transition between AASW and WW.

WDW representation in CESM1-POP2 and SODA3.4.1 covers a layer that reaches depths below 2000 m, which according to [Kerr et al. \(2009\)](#) is unrealistic for this water mass. Further, this water mass shows warmer waters in CESM1-POP2 in both regions.

Although WSDW has shown values of temperature and salinity consistent with the literature, the WSBW and SW, which are important in its formation process, are absent in both CESM1-POP2 and SODA3.4.1.

The absence of WSBW and SW could be related to physical processes that are not accounted for in large-scale models. As described in section [1.3.3.1](#) there are some regions with high salinity in SW.

The results reported in this section only provide a simple characterization of the water mass representation in CESM1-POP2 and SODA3.4.1. Although water masses have shown values of temperature and salinities consistent with the index adopted in this study

(Table 1.1), important coastal dynamics constraints, related to the formation of dense waters, are not accounted for in the model and in the reanalysis.

3.2 Downscaling

It was shown that the Ocean GCM results reproduces satisfactorily the large scale patterns in the Southern Ocean. The results also show that important coastal dynamics features are neglected by those models which shows the necessity of downscaling to capture small scale physical processes. In this section we start to analyze the results from the downscaling experiments to understand the climate impacts on the ACoC.

3.2.1 Winds

The regional simulations were forced with 10 m winds from NCEP Reanalysis 1, which is considered the reference for the 20th century, and two CCSM3 experiments of IPCC AR4. The first considers the historical 20th century scenario (CCSM3–20C3M) and the second for the A2 scenario from SRES (CCSM3–SRESA2). Monthly mean data fields between 1980 to 1999 were used for the NCEP and CCSM3–20C3M, and for 2021 to 2047 period for CCSM3–SRESA2.

3.2.1.1 Mean Patterns

Wind magnitude climatological fields of the three datasets are shown in Figure 3.31. First we evaluated climatologies performed in the end of 20th century (1980–1999) to compare NCAR climate model winds with NCEP. Although CCSM3–20C3M climatology (Figure 3.31b) shows the same pattern (Figure 3.31a), with greater magnitudes found in the westerlies region, between 45°S and 65°S, and close to the continent in the East of Weddell Sea (10°W to 40°E), the modeled winds magnitude are substantially stronger than in reanalysis. The difference between CCSM3–20C3M and NCEP (Figure 3.32a) showed to be significant at 95% level in all domain. It differs in up 3.6 m/s in the westerlies region and 5.11 m/s where easterly winds are dominant.

Figure 3.31: **Winds climatologies** [m/s]. **(a)** NCEP climatology for the 1980–1999 period; **(b)** CCSM3–20C3M climatology for the 1980–1999 period; **(c)** CCSM3–SRESA2 climatology for the 2021–2047 period.

Figure 3.32: Differences between **(a)** CCSM3–20C3M and NCEP climatologies and **(b)** CCSM3–SRESA2 and CCSM3–20C3M climatologies. Significant differences at 95% level are marked with black dots.

Figure 3.33: **Seasonal winds climatologies [m/s]**. NCEP summer (DJF) **(a)** and winter (JJA) **(b)** climatologies for the 1980–1999 period; CCSM3–20C3M summer (DJF) **(c)** and winter (JJA) **(d)** climatologies for the 1980–1999 period; CCSM3–SRESA2 summer (DJF) **(e)** and winter (JJA) **(f)** climatologies for the 2021–2047 period.

Comparing the 20th century climatology (Figure 3.31b) and 21st century projection (Figure 3.31c) smaller differences are found. Significant differences are found between 50°S and 65°S (Figure 3.32b). In this latitude band an increase in wind magnitude up to 0.45 m/s is observed in the 21st century. Significant differences are also found in the Weddell

basin from 70°S to south, with magnitude up to 0.21 m/s higher.

In the East Weddell Sea, where higher magnitudes of the easterlies were observed, the magnitudes showed to be smaller in 21st century and negative differences up to -0.25 m/s are found.

Seasonal climatologies are shown in [Figure 3.33](#). During summer maximum values associated with the westerlies reaching positions further south are observed. South of 65°S weakening is observed during summer months.

Summer and winter climatologies differences are shown in [Figures 3.34a](#) and [3.34b](#), respectively. During summer model winds magnitude were 4.15 m/s higher than reanalysis, and it differs up to 7.27 m/s in the westerlies region in winter. According to [Hurrell et al. \(2006\)](#) SLP errors in high SH latitudes are larger during winter.

Figure 3.34: **Seasonal differences.** Differences between CCSM3–20C3M and NCEP summer (DJF) [\(a\)](#) and winter [\(b\)](#) climatologies; and CCSM3–SRESA2 and CCSM3–20C3M summer (DJF) [\(c\)](#) and winter (JJA) [\(d\)](#) climatologies.

Differences between CCSM3–20C3M and CCSM3–SRESA2 summer and winter climatologies are shown in [Figures 3.34c](#) and [3.34d](#), respectively. CCSM3–SRESA2 showed

greater values during summer in almost the entire domain. During this season magnitudes are up to 0.77 m/s higher than in CCSM3–20C3M. Negative differences around -0.44 m/s, were observed only in two small regions. The first is located in the northern boundary between 45°–50°S and 20°–60°W, and the second between 65°–70°S and 0°–55°W.

During winter (Figure 3.34d) the differences are also predominantly positives, but smaller than during summer. The maximum positive difference is found in the eastern boundary between 55°–60°S and is around 0.59 m/s. However, in the East Weddell Sea more pronounced negative difference is observed in the easterlies which showed to be up 0.66 m/s weaker in CESM1–SRESA2.

3.2.1.2 Zonal wind trends and variability

In this section we analyze only the zonal wind (UWND) trend and variability, since winds are nearly zonal in our study area. Trends results for NCEP can be seen in Figure 3.35a. Significant trends occurs mainly in the latitude band between 63°S and 70°S, where an increase up to $0.1 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$ is observed. Negative trends in the East Weddell Sea, near the continent, also denotes significant increase (up to $0.12 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$) in the easterlies magnitude.

In CCSM3–20C3M (Figure 3.35b) the westerlies shows a decrease of $0.05 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$ between 45°S and 53°S, and increase up to $0.06 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$ south. Negative trends, associated with increasing of easterlies magnitudes, are up to $0.052 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$. Trends in CCSM3–20C3M showed to be significant only in the region where its observed the maximum positive trend.

Figure 3.35: **Zonal wind linear trend maps** [$\text{m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$]. Linear trend maps obtained from (a) NCEP over 1980–1999 period; (b) CCSM3–20C3M over the 1980–1999 period; (c) CCSM3–SRESA2 over the 2021–2047 period. Significant linear trends at 95% level are marked with black dots.

Results for the CCSM3–SRESA2 (Figure 3.35c) shows decreasing trends between 45°S and 60°S, and increasing southward. However, trends are small ($\approx 0.02 \text{ m}\cdot\text{s}^{-1}\text{yr}^{-1}$) and not significant.

Figure 3.36: **(a)** Latitudinal zonal wind profiles in the longitudinal mean between 63°W and 20°E obtained from NCEP every 5 years during the 1980–1999 period; **(b)** NCEP annual latitudinal mean position of maximum zonal wind magnitude in each year during the 1980–1999 period.

Winds latitudinal profiles were evaluated for each dataset and [Figure 3.36](#) presents NCEP reanalysis results. Small differences are observed in zonal wind stress magnitude ([Figure 3.36a](#)), values below the mean are only detected in the first 5 years. In relation to the annual mean position ([Figure 3.37b](#)) eleven of the twenty years analyzed showed greater variation relative to the initial position (48.5°S). The southernmost latitude reached was 52.5°S which is consistent with [Swart and Fyfe \(2012\)](#).

Figure 3.37: **(a)** Latitudinal zonal wind profiles in the longitudinal mean between 63°W and 20°E obtained from CCSM3–20C3M every 5 years during the 1980–1999 period; **(b)** CCSM3–20C3M annual latitudinal mean position of the maximum zonal wind magnitude.

Whereas in NCEP (Figure 3.36a) the zonal-wind maximum is around 6.5 m/s, in CCSM3–20C3M (Figure 3.37a) it exceeds 10 m/s. Over the 5 years means great differences are not observed in the maximum zonal wind magnitude, which remains around 10.5 m/s.

Figure 3.38: **(a)** Latitudinal zonal wind profiles in the longitudinal mean between 63°W and 20°E obtained from CCSM3-SRESA2 every 5 years during the 2021–2047 period; **(b)** CCSM3-SRESA2 annual latitudinal mean position of the maximum zonal wind magnitude.

Annual latitudinal mean positions in CCSM3–20C3M (Figure 3.37b) remains almost constant throughout the years ($\approx 51^\circ\text{S}$). Only in 1995 it reaches a position further south (52.5°S). In some years positions northward are detected in approximately 49.5°S .

Figure 3.38 shows the results relative to CCSM3-SRESA2. The 21st scenario showed small difference in zonal-wind maximum relative to CCSM3–20C3M. CCSM3–20C3M (Figure 3.37a) show a maximum of 10.74 m/s between 1985 and 1989, and the mean over the entire period (1980–1999) was 10.57 m/s. Whereas in CCSM3-SRESA2 (Figure 3.38a) zonal-wind maximum was 10.84 m/s during the 2031–2035 period, and the mean for the entire period (2021–2047) was 10.66 m/s.

Although CCSM3-SRESA2 (Figure 3.38b) shows greater number of events where

zonal flow displaces poleward, the results are not sufficient to evaluate the poleward shift significance of the westerlies.

Zonal wind anomalies relative to the 1980 to 1999 climatology for NCEP and CCSM3–20C3M, and 2021 to 2047 climatology for CCSM3–SRESA2, were also evaluated at 50°S, 55°S, 65°S, and 70°S. NCEP showed small but significant (at 95% level) linear trends in all latitudes. In both CCSM3 scenarios linear trends are not significant.

At 50°S (Figure 3.39) anomalies remains around ± 2 m/s for all data. For CCSM3–20C3M (Figure 3.39b) a negative anomaly peak of 5.5 m/s is detected in August 1997. CCSM3–SRESA2 (Figure 3.39c) also shows some negative anomaly peaks which reaches almost 4 m/s in June 2031.

NCEP (Figure 3.39a) showed a significant increasing trend at 50°S of $1.8 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$. Linear trends in CCSM3 scenarios were not significant, but in contrast to NCEP reanalysis they showed a decreasing trend of $-9.3 \times 10^{-4} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$ for CCSM3–20C3M and $-4.1 \times 10^{-4} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$ for CCSM3–SRESA2.

Zonal wind anomalies results at 55°S are shown in Figure 3.40. Relative to other latitudes analyzed, greater anomalies are observed at this latitude mainly in CCSM3 scenarios. In the NCEP (Figure 3.40a) anomalies do not exceed ± 4 m/s. Whereas in CCSM3 scenarios some negative anomalies above that are observed in October 1983 and August 1997 in CCSM3–20C3M (Figure 3.40b), and in August 2026 and September 2027 in CCSM3–SRESA2 (Figure 3.40c). In the NCEP the negative maximum peaks occurs in summer months, while the maximum positive anomalies are observed in autumn and winter.

NCEP (Figure 3.40a) has a significant trend (at 95% level) of $2.4 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$, and in CCSM3–20C3M (Figure 3.40b) the trend is of $1.3 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$ but it is not significant. In CCSM3–SRESA2 (Figure 3.40c) a decreasing trend of $-2.9 \times 10^{-4} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$ is noticed, but it is not significant.

At 65°S NCEP (Figure 3.41a) a significant (at 95% level) increasing of $3.4 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$. CCSM3–20C3M (Figure 3.41b) a positive trend of the same magnitude ($1.3 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}\cdot\text{yr}^{-1}$) is observed, however it is not significant.

Results at 70°S (Figure 3.42) show a negative trend in all datasets. In CCSM3–20C3M (Figure 3.42b) and CCSM3–SRESA2 (Figure 3.42c) anomalies are almost two times greater than NCEP.

Figure 3.39: **Zonal wind anomalies at 50°S.** (a) NCEP zonal wind anomaly over the 1980–1999 period; (b) CCSM3–20C3M zonal wind anomaly over the 1980–1999 period; and (c) CCSM3–SRESA2 zonal wind anomaly over the 2021–2047 period. Uncertainties are indicated by coloured shading.

(a)

(b)

(c)

Figure 3.40: **Zonal wind anomalies at 55°S.** (a) NCEP zonal wind anomaly over the 1980–1999 period; (b) CCSM3–20C3M zonal wind anomaly over the 1980–1999 period; and (c) CCSM3–SRESA2 zonal wind anomaly over the 2021–2047 period. Uncertainties are indicated by coloured shading.

Figure 3.41: **Zonal wind anomalies at 65°S.** (a) NCEP zonal wind anomaly over the 1980–1999 period; (b) CCSM3–20C3M zonal wind anomaly over the 1980–1999 period; and (c) CCSM3–SRESA2 zonal wind anomaly over the 2021–2047 period. Uncertainties are indicated by coloured shading.

Figure 3.42: **Zonal wind anomalies at 70°S.** (a) NCEP zonal wind anomaly over the 1980–1999 period; (b) CCSM3–20C3M zonal wind anomaly over the 1980–1999 period; and (c) CCSM3–SRESA2 zonal wind anomaly over the 2021–2047 period. Uncertainties are indicated by coloured shading.

Although the results performed in this section showed some evidence of changes in the westerlies in CCSM3 scenarios, they were not statistically significant. In the present study the winds were evaluated with the aim to understand its behavior during the period that

regional simulations were performed.

3.2.2 Regional Simulations with ROMS in the Weddell Sea

In this section we evaluate the results from the regional simulations with the aim to characterize the ACoC and investigate its changes in a climate scenario. Results were analysed for three different simulations with ROMS. Two of them were performed from 1980 to 1999 period. The first was forced with NCEP and SODA2.4.4 reanalyses (referred to as ROMS–NCEP), and the second was forced with the CCSM3 AR4 20th century historical scenario (referred to as ROMS–20C3M). The third simulation was forced with CCSM3 experiment for the 21st century pessimist scenarios SRESA2, and will be referred to as ROMS–SRESA2. This last simulation comprises the 2021 to 2047 period.

3.2.2.1 ACoC Transport

The ACoC transport was calculated using [Equation 2.1](#) for three simulations. Initially, we compared ROMS–NCEP ([Figure 3.43a](#)) and ROMS–20C3M ([Figure 3.43b](#)) transport climatological fields, both for the 1980–1999 period. The ROMS–NCEP simulation was used as reference for the 20th century. Therefore, this first comparison aimed to highlight the differences between simulations forced with reanalysis and with a global circulation model to evaluate whether this last one consistently reproduces ACoC’s dynamic features.

[Figure 3.43](#) shows that maximum transport related to the ACoC occurs between 10°W e 0° in both climatologies. ROMS–20C3M climatology ([Figure 3.43b](#)) presents higher transport values throughout its domain. ACoC maximum transport ROMS–20C3M simulation reaches the value of 20.46 Sv, while in ROMS–NCEP ([Figure 3.43a](#)) the maximum is 10.8 Sv. Furthermore, the ACoC seen to be narrower in ROMS–NCEP. The maximum difference between these two climatologies was 16.8 Sv at 0°W and 68.9°S . The differences are illustrated in [Figure 3.44a](#).

The ACoC is mainly driven by wind-induced Ekman transport. Therefore, differences between ROMS–NCEP and ROMS–20C3M transport could be related to stronger winds detected in CCSM3–20C3M (see section [3.2.1](#)).

Figure 3.43: **Transport magnitude climatologies [Sv].** (a) ROMS–NCEP climatology over the 1980–1999 period; (b) ROMS–20C3M climatology over the 1980–1999 period; (c) ROMS–SRESA2 climatology over the 2021–2047 period.

Figure 3.44: Significant differences, at 95% level, between (a) ROMS–20C3M and ROMS–NCEP, and (b) ROMS–SRESA2 and ROMS–20C3M transport climatologies.

In order to investigate the differences between the two simulations we also used the AWI233 mooring data (Fahrback and Rohardt, 2012) to evaluate which of them better

represents the ACoC. This mooring was chosen because it has a longer time series (1997–2008). Moreover, the AWI233 is located in the region where ACoC is stronger (Figure 2.1).

Figure 3.45: Zonal velocities time series from AWI233 mooring (blue), ROMS–NCEP simulation (orange), and ROMS–20C3M (red) at 247, 744 and 1925 m depths.

Simulated zonal velocities were compared with mooring data (Figure 3.45). The analysis was performed in the three mean instrument depth (247, 744 and 1925 m) and the comparison was made from February 1998 to September 1999, except for the 247 m depth where observational data are limited to January 1999. We considered a 4-day frequency for all time series, since ROMS–20C3M has this temporal resolution, ROMS–NCEP a 2-day frequency and AWI233 mooring a 2-hour temporal resolution.

The results in Figure 3.45 shows that ROMS–NCEP simulations underestimate current velocities. At the analyzed depths, ROMS–NCEP presents velocities of $10^{-2} \text{ m}\cdot\text{s}^{-1}$ order, while for AWI233 and ROMS–20C3M the velocities are of $10^{-1} \text{ m}\cdot\text{s}^{-1}$ order.

For all depths ROMS–NCEP shows a weak correlation with mooring data ($r < 0.5$). Only at 744 m depth higher correlation is detected with AWI233 ($r=0.49$) than ROMS–20C3M ($r=0.41$). At 247 and 1925 m ROMS–NCEP and mooring data correlation was too small ($r=0.23$ and 0.18 respectively).

ROMS–20C3M shows a moderate correlation with mooring data at 247 m depth

($r=0.51$). At 744 m and 1925 m depths a weak correlation is detected with AWI233 ($r=0.41$ and 0.28 , respectively).

For the three mean depths time series its possible to notice that ROMS-20C3M overestimates velocity mainly in the winter months. However, ROMS-20C3M simulated velocities are of the same order of magnitude of mooring data ($10^{-1} \text{ m}\cdot\text{s}^{-1}$) and reproduces the seasonal variability of currents, which is stronger during winter relative to summer.

ROMS-20C3M transport show a better agreement with literature values. [Schröder and Fahrbach \(1999\)](#) also estimated volume transport across the Greenwich Meridian, where coastal current is stronger. They found a 22 Sv transport associated with ACoC. This value is consistent with ROMS-20C3M value (20.46 Sv), and also reinforce ROMS-NCEP transport is underestimated (10.8 Sv).

Next, comparisons between the 20th century (ROMS-20C3M) and projection (ROMS-SRESA2) results with ROMS are presented:

[Figure 3.43](#) shows a decrease in the 21st century modeled transport ([Figure 3.43c](#)) relative to 20th century ([Figure 3.43b](#)). In the east Weddell Sea region (between 10°W and 10°E) where the ACoC is stronger, the transport reaches values of 20.46 Sv in ROMS-20C3M, while In ROMS-SRESA2 it is less than 15.95 Sv. Volume transport are overall smaller in ROMS-SRESA2 ([Figure 3.44b](#)). In the region where ACoC is stronger, the maximum difference found was 15.78 Sv.

[Figure 3.46](#) shows the seasonal climatologies. During winter ACoC reaches a volume transport of 27.3 Sv in ROMS-20C3M ([Figure 3.46b](#)), while in ROMS-SRESA2 ([Figure 3.46d](#)) the maximum observed is 22.09 Sv. In the summer the transport is much weaker, in ROMS-20C3M ([Figure 3.46a](#)) the maximum value associated with ACoC was 11.81 Sv and in ROMS-SRESA2 ([Figure 3.46c](#)) 9.28 Sv.

Previous studies such as [Núñez-Riboni and Fahrbach \(2009\)](#), showed ACoC is stronger during autumn. In this study, the climatologies for other seasons were also calculated (not show). However, in ROMS simulations the values during autumn do not exceed the values in winter. The maximum transport in this season was 19.7 Sv and 14.47 Sv for ROMS-20C3M and ROMS-SRESA2, respectively.

[Heywood et al. \(1998\)](#) measured transport at 72°S and 17°W in the early autumn, and found a value of 14 Sv . In ROMS-20C3M the transport in this region was 9.51 Sv while in ROMS-NCEP it was 7.47 Sv in this region.

Figure 3.46: **Seasonal Transport Climatologies [Sv]**. ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) climatologies for 1980–1999 period; ROMS-SRESA2 summer (DJF) (c) and winter (JJA) (d) climatologies for 2021–2047 period.

Figure 3.47: Seasonal differences between ROMS-SRESA2 and ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) transport climatologies.

Differences between ROMS-20C3M and ROMS-SRESA2 seasonal climatologies are shown in Figure 3.47. Summer climatologies (Figure 3.47a) of ROMS-20C3M show transport magnitudes around 6.67 Sv higher than ROMS-SRESA2 in the eastern Weddell Sea,

where ACoC is stronger. The difference between winter climatologies are even higher and differs up to 8.79 Sv at the same region.

3.2.2.2 Trends in ACoC transport

Linear trends were calculated and are shown in Figure 3.48. Trends showed to be significant at 95% over the entire domain, and in CCSM3–20C3M (Figure 3.48a) trends are overall positive, whereas in ROMS–SRESA2 (Figure 3.48b) it is predominantly negative. Positive trends in ROMS–SRESA2 are found between 64°S and 66°S and are around 0.31 $\text{Sv}\cdot\text{yr}^{-1}$.

Figure 3.48: **Volume transport linear trend maps** [$\text{Sv}\cdot\text{yr}^{-1}$]. Significant linear trend, at 95% level, obtained from (a) ROMS–20C3M over 1980–1999 period and (b) ROMS–SRESA2 over 2021–2047 period.

In the East Weddell Sea, where ACoC is stronger an increasing trend in its magnitude of 0.92 $\text{Sv}\cdot\text{yr}^{-1}$ in ROMS–20C3M (Figure 3.48a) is observed. In ROMS–SRESA2 (Figure 3.48b) a trend decrease of 0.26 0.92 $\text{Sv}\cdot\text{yr}^{-1}$ is noticed.

3.2.2.3 Subregions in the Weddell Sea

In order to investigate changes in ACoC we selected three subregions in the Weddell Sea according to the ACoC transport (Figure 3.49).

Figure 3.49: Selected subregions in the Weddell Sea according to the transport climatologies. Background colors represent ROMS–20C3M climatology.

3.2.2.3.1 Subregion A1

The subregion A1 comprises the Weddell basin between 66° – 76° S and 28° – 58° W. [Figure 3.50](#) shows the Hovmöller diagrams obtained from the monthly means over the entire period of each simulation. In ROMS–20C3M ([Figure 3.50a](#)), the maximum magnitudes (≈ 4 Sv) at 72° S are observed from May to September. Although magnitudes transport are smaller in summer, values around 3.5 Sv are still observed from 67° S to 72° S. In the Hovmöller diagrams we can notice an increasing northward in the transport magnitude.

Figure 3.50: **Hovmöller diagrams of the volume transport in subregion A1.** (a) ROMS-20C3M transport obtained from the monthly means over the 1980–1999 period; (b) ROMS-SRESA2 transport obtained from the monthly means over the 2021–2047 period.

Figure 3.51: Volume transport annual cycle in subregion A1 from ROMS-20C3M (blue) and ROMS-SRESA2 (red)

Greater decrease in the transport magnitudes are observed from ROMS-20C3M to ROMS-SRESA2 (Figure 3.50b), mainly in the summer and early autumn. During these seasons, the transport magnitudes do not exceed 2.5 Sv. Maximum magnitudes are also observed from April to October and are around 3.5 Sv. At 72°S and 67°S, magnitudes around 3 Sv are observed up to early summer. Decrease in autumn and winter are observed between 68°S and 71°S where magnitudes do not exceed 2.5 Sv.

Those differences between simulations, described above, are also seen in the annual cycle (Figure 3.51), which shows smaller transport magnitudes in ROMS–SRESA2 throughout the year.

Figure 3.52: **Longitudinal mean zonal transport cross section in subregion A1.**(a) ROMS–20C3M zonal transport cross section over the 1980–1999 period; (b) ROMS–SRESA2 zonal transport cross section over the 2021–2047 period.

Since subregion A1 comprises a large area, the results obtained from Hovmöller diagrams and annual cycles also includes the transport associated with the WG. Therefore, to evaluate ACoC transport in subregion A1, the zonal transport was analyzed in depth in longitudinal mean cross sections. The meridional transport was analyzed separately in a third subregion as marked in Figure 3.49.

Results of the zonal transports are shown in Figure 3.52. According to Carmack and Foster (1975a), Foster and Carmack (1976) and Fahrbach et al. (1994), in this region where the continental shelf is wide, the ACoC splits into two branches, one following the continental slope and another penetrating onto the shelf. In both scenarios it is observed two westwards cores, one found at shelf break, between 74°S and 75°S, while the second occurs further north, between 71°S and 72°S.

The multibanded structure reproduced in ROMS simulations is similar to those described by Heywood et al. (1998) and Núñez-Riboni and Fahrbach (2010). However, in both studies the ACoC was evaluated in the eastern Weddell Sea where the continental shelf is narrow. In this region, the current is displaced offshore by the ice shelf and is closer to the ASF (Núñez-Riboni and Fahrbach, 2010).

Fahrbach et al. (1994, 1995) attributed the multibanded structure to the bottom topography. However, Núñez-Riboni and Fahrbach (2010) showed that the sea ice is the more likely cause of these cores.

The observed westward flow (Figure 3.52) extends up to 1000 m depth. Greater differences between both scenarios are found in the northern core, where a weakening in ROMS-SRESA2 (Figure 3.52b) transport is observed. In ROMS-20C3M (Figure 3.52a), transports around 0.23 Sv are found up to 1000 m depth, whereas in ROMS-SRESA2 a transport of 0.11 Sv is observed.

At the shelf break however, an increasing from surface to approximately 350 m depth is observed. In ROMS-20C3M (Figure 3.52a), the zonal transport do not exceed 0.2 Sv, while in ROMS-SRESA2 (Figure 3.52b), it varies from 0.35 Sv at surface to 0.23 Sv in depths bellow.

The seasonal patterns of zonal transport are shown in Figure 3.53. The climatologies show the current weakening in summer. In both seasons, the flow is stronger in ROMS-20C3M (Figure 3.53a and Figure 3.53b) than in ROMS-SRESA2 (Figure 3.53c and Figure 3.53d).

During summer months we observed three weaker cores. In ROMS-20C3M (Figure 3.53a) a small core occurs at shelf break with 0.17 Sv zonal transport. The second core is found between 73°S and 74°S and also do not exceed 0.2 Sv. In the northern core, between 71°S and 72°S, the maximum transport of about 0.31 Sv is detected. In ROMS-SRESA2 (Figure 3.53d) these cores are too weak and the transport do not exceed 0.17 Sv. The northern core shows greater transports only in the first 100 m depth.

Compared to ROMS-20C3M (Figure 3.64b), during winter ROMS-SRESA2 (Figure 3.64d) shows a weakening in the northern core and a strengthening at the shelf break core, where zonal transport exceeds 0.6 Sv.

In both scenarios, we observed a counter-current over the continental shelf from 400 to 600 m depth, with an associated transport of about 0.05 Sv in ROMS-20C3M (Figure 3.52a) and 0.1 Sv in ROMS-SRESA2 (Figure 3.52b). The presence of counter-currents were also evidenced in previous studies such as Fahrbach et al. (1994); Heywood et al. (1998); Núñez-Riboni and Fahrbach (2009) either at continental slope or under the ice shelf. However, all of them were performed in the eastern Weddell Sea where the continental shelf is narrow.

Figure 3.53: **Seasonal longitudinal mean zonal transport cross sections in subregion A1.** ROMS–20C3M summer (DJF) (a) and winter (JJA) (b) zonal transport cross sections over the 1980–1999 period; ROMS–SRESA2 summer (DJF) (c) and winter (JJA) (d) zonal transport cross sections over 2021–2047 period.

In both seasons, the counter-current eastward flow is also present, and it is weaker during summer. In this season, ROMS–20C3M (Figure 3.53a) shows a counter-current at the surface and in the slope between 1000 and 2000m. Whereas in ROMS–SRESA2 (Figure 3.53c), the counter-current occurs between 400 and 600 m depth. In both scenarios the transport associated with the counter-current is around 0.05 Sv. During winter however, the counter-current showed to be slightly stronger in ROMS–SRESA2 (Figure 3.53d) than

in ROMS–20C3M (Figure 3.53b). In ROMS–SRESA2 the counter-current reaches almost 0.2 Sv, while in ROMS–20C3M its transport is around 0.12 Sv.

According to Núñez-Riboni and Fahrbach (2009) and Smedsrud et al. (2006), the thermal wind is the main mechanism responsible for the development of the counter-current. The decrease of the southward density gradient gives rise to the eastward flow, which increases with depth and overcomes the barotropic component leading to the development of the counter-current (Smedsrud et al., 2006; Núñez-Riboni and Fahrbach, 2009). The variability of the counter-current is a response to the mutual mechanisms interacting on the coastal region, which are important to the ACoC variability.

In order to investigate the termohaline forcings, we also evaluated the temperature and salinity. Temperature cross sections are shown in Figure 3.54 and salinity in Figure 3.57. The salinity cross sections show that regional simulations are capable to reproduce the ASF V-shaped structure. In the temperature cross sections this structure is not so marked and the isopycnals are more close to the half 'V' pattern observed in narrow shelves. However, there is a clear front that separates cold fresher waters from the shelf to the warmer and saltier waters offshore.

Temperatures are overall higher in ROMS–SRESA2 (Figure 3.54b) than in ROMS–20C3M (Figure 3.54a), except over the continental shelf where temperatures do not show greater variability and remains below -1.64°C in both scenarios. However, at the surface this minimum temperature extends until 71.5°S in ROMS–SRESA2 (Figure 3.54b), while in ROMS–20C3M (Figure 3.54a) this isotherm is observed until 73.5°S .

Figure 3.54: **Longitudinal mean temperature cross sections in subregion A1.** (a) ROMS–20C3M temperature cross section over the 1980–1999 period; (b) ROMS–SRESA2 temperature cross section over the 2021–2047 period.

Greater differences in the SW are found in the salinity. In ROMS–20C3M (Figure 3.57a) salinity ranges from 34.35 to 34.7 over the continental shelf. Whereas in ROMS–SRESA2 (Figure 3.57b) smaller salinity are found at the surface, reaching salinities below 34.3. The 34.5 isohaline intrusion onto the shelf is also observed. In the 20th century scenario this isohaline downward slope was found below 600 m at the continental slope.

Further north, offshore waters overall did not show a good representation on the temperature field of the surface and intermediate waters, mainly in ROMS–20C3M (Figure 3.54a). Temperatures presented negative values in almost the entire column in ROMS–20C3M. The presence of WDW associated with temperatures above zero, occurs only at 1000 m depth and reaches a maximum of 0.04°C .

ROMS–SRESA2 (Figure 3.54b) shows an increasing in temperatures relative to 20th century. Surface and intermediate waters, however, are still poorly represented. The WDW reaches temperatures up to 0.28°C in the 21st run, but it is also found to deep, at 600 m depth. Furthermore, this water mass is found further north (70°S) and is far from the coastal region.

Temperatures associated with the WSDW and the WSBW (see Table 1.1) are observed in both scenarios, and in ROMS–SRESA2 an increase of the WSBW temperature is also observed.

With respect to the salinity field, ROMS–SRESA2 (Figure 3.57b) shows smaller

salinities at the surface (up to 34.3) and an increase in the salinities of the layers below. In ROMS–20C3M (Figure 3.57a), depths from 200 to 800 m have salinities varying from 34.51 to 34.66, whereas in ROMS–SRESA2 the salinity values reach up to 34.71.

Figure 3.55: **Seasonal longitudinal mean temperature cross sections in subregion A1.** ROMS–20C3M summer (DJF) (a) and winter (JJA) (b) temperature cross sections over the 1980–1999 period; ROMS–SRESA2 summer (DJF) (c) and winter (JJA) (d) temperatures cross sections over the 2021–2047 period.

Below 1000 m, where it was observed the presence of WSDW and WSBW, differences between both scenarios are not significant.

In the temperature seasonal fields Figure 3.55, SW and surface waters do not show

great differences between both scenarios. In the summer climatologies, the SW remains below -1.64°S . Further north, at the surface, temperatures varies between -1.32°C and -1°C . However, ROMS-20C3M (Figure 3.55a) shows a thicker layer of waters with temperatures around -1.32° . In ROMS-SRESA2 (Figure 3.55c) temperatures rises at 50 m depth.

Figure 3.56: **Seasonal differences in the temperature field in subregion A1.** Differences between summer and winter temperatures in ROMS-20C3M (1980–1999) (a) and ROMS-SRESA2 (2021–2047) (b).

In the winter months, smaller temperatures at the surface are observed in ROMS-SRESA2 (Figure 3.55d), where the minimum temperature extends further north. In ROMS-20C3M (Figure 3.55b) the -1.64°C isotherm is found up to 72.5°S and further north at 69.8°S , but not in a continuous way, as in ROMS-SRESA2. In the winter climatologies, at 50 m depth, the temperatures in ROMS-SRESA2 also become higher than in ROMS-20C3M.

Relative to the seasonal differences (Figure 3.56), temperatures are up to 0.7°C higher during summer in the first 200 m depth in ROMS-20C3M than in ROMS-SRESA2 (Figure 3.56a). In the layer between 200 and 1000 m, the temperatures are lower and vary up to 0.18°C . Over the continental shelf the differences are too small and are around $\pm 0.02^{\circ}\text{C}$.

In ROMS-SRESA2 (Figure 3.56b), the seasonal differences are smaller. At the northernmost boundary, the maximum differences at surface, are of 0.3°C . Negative differences are around 0.02° and are observed in the region of ASF. Over the continental shelf, small positive differences ($\approx 0.02^{\circ}\text{C}$) are detected.

In both seasons, great differences in temperature are found in the WDW. Although it is still poorly represented, ROMS–SRESA2 show a warming of this water mass in the 21st century. Robertson et al. (2002) shows a warming of $0.012 \pm 0.007^\circ \text{yr}^{-1}$ in this water mass from 1970 to 1990. However, this warming was not compensated by an increase in the salinity, and the WDW showed a displacement of 200 m towards the surface. The results here presented also show a displacement of about 200 m, although the salinities associated with this water mass showed an increase from the 20th to the 21st century (Figure 3.58).

In the seasonal fields, a warming in the WSDW and WSBW (Figure 3.55) is also observed, but changes in their salinities (Figure 3.58) are not observed. Azaneu et al. (2013) showed a warming of the AABW in almost the entire Southern Ocean. The Weddell Sea was one of the regions where the warming was more pronounced. With respect to the salinity field, Azaneu et al. (2013) found a non-significant freshening trend.

According to Azaneu et al. (2013), changes in Southern Ocean dense waters are probably related with a freshening of SW and a warming of the AABW at the open ocean, due to interactions with WDW above, which shows a recent warming.

Figure 3.57: **Longitudinal mean salinity cross sections in subregion A1.** (a) ROMS–20C3M salinity cross section over the 1980–1999 period; (b) ROMS–SRESA2 salinity cross section over the 2021–2047 period.

With respect to the salinity seasonal fields (Figure 3.58), smaller values at the surface layer are observed in ROMS–SRESA2, reaching up to approximately 34.3. However, as already mentioned, below 50 m depth an increase in the salinity is observed up to 800 m depth.

As to the seasonal differences, higher salinities are observed during winter at the surface in both scenarios (Figure 3.59). In ROMS-20C3M (Figure 3.59a) higher salinities are observed during summer in the layer between 50 and 600 m, including the continental shelf. Positive differences are up to 0.01. In ROMS-SRESA2 (Figure 3.59b), positive salinities are found only in the coastal region and differs up to 0.02 at the shelf break.

Figure 3.58: **Seasonal longitudinal mean salinity cross sections in subregion A1.** ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) salinity cross sections over the 1980–1999 period; ROMS-SRESA2 summer (DJF) (c) and winter (JJA) (d) salinity cross sections over the 2021–2047 period.

Figure 3.59: **Seasonal differences in the salinity field in subregion A1.** Differences between summer and winter salinities in ROMS–20C3M (1980–1999) **(a)** and ROMS–SRESA2 (2021–2047) **(b)**.

We also evaluated the Weddell Sea water masses structure in the regional simulations through the TS diagrams (Figure 3.60). The regional simulations provide a better representation of SW, but the temperature and salinity are closer to the LSSW, according to the existing literature (Table 1.1).

As already mentioned, WDW occurs in a deeper layer and the temperature values are underestimated. However, an increase in its temperature in the 21st century scenario is observed (Figure 3.60b).

Both the WSDW and the WSBW are well represented with respect to the temperature field. Nevertheless, the salinities found are higher than the values obtained from the available literature (Table 1.1). As a consequence, the water masses are denser. From the 20th to the 21st century scenario, the increase in WSBW temperatures is also observed in the TS diagrams.

Figure 3.60: **Scatter TS diagram in subregion A1.** (a) ROMS-20C3M TS diagram over the 1980–1999 period; (b) ROMS-SRESA2 TS diagram over the 2021–2047 period. The solid black line represents the mean temperature and salinity in each depth point.

Although the regional simulations showed a better representation of the SW and the deep waters, as discussed in section 1.3.3.1, the WDW and/or MWDW have an important role in the formation processes of the WSDW and WSBW. Furthermore, these intermediate waters are the main source of heat for the continental shelf and drives basal melting.

In the region marked by ASF the results showed increased temperature in the layer between 50 and 300 m. This might be related with the response of the MWDW to the WDW warming. However, pronounced temperature changes in the SW were not observed from the 20th into the 21st century scenario.

3.2.2.3.2 Subregion A2

The subregion A2 is located in the East Weddell Sea, between 66°S–70°S and 6°W–12°E. In this region the strongest transport associated with ACoC is observed due to the WG influence.

Hovmöller diagrams are shown in Figure 3.61. Greater transport magnitudes associated with the ACoC are found between 68°S and 70°S. In ROMS-20C3M (Figure 3.61a), the maximum transport magnitude exceeds 10Sv from May to October. During summer, the transport magnitude ranges from 5 Sv to 7 Sv.

ROMS–SRESA2 (Figure 3.61) hovmöller also show a weakening in the ACoC transport at this region. The maximum transport reached is of 8.5 Sv at 68°S in August, and it remains around 7 Sv from April to October. During summer the transport is around 4 Sv.

The annual cycle (Figure 3.62), shows that the bigger differences between both scenarios occurs during winter months, when the transport reaches its maximum. In ROMS–20C3M the maximum transport magnitude was of 6.5 Sv in June and July, while in ROMS–SRESA2 it was of ≈ 4.6 Sv in August.

Figure 3.61: **Hovmöller digrams of the volume transport in subregion A2.** (a) ROMS–20C3M transport obtained from the monthly means over the 1980–1999 period; (b) ROMS–SRESA2 transport hovmöller diagram obtained from the monthly means over the 2021–2047 period.

Figure 3.62: Volume transport annual cycle in subregion A2 from ROMS–20C3M (blue) and ROMS–SRESA2 (red)

The cross sections analysis in subregion A2 were extended further north, up to 64°S . The results of the zonal transport cross sections (Figure 3.63) show the eastward flow associated with the northern branch of the WG and the ACC. However, the analysis of the present study we focuses only in the transport associated with the ACoC.

Figure 3.63: **Longitudinal mean zonal transport cross section in subregion A2.**(a) ROMS–20C3M zonal transport cross section for the 1980–1999 period; (b) ROMS–SRESA2 zonal transport for the 2021–2047 period.

The subregion A2 includes the study area of previous studies (e.g. Núñez-Riboni and Fahrbach (2010)) where the multibanded structure of the ACoC and the presence of counter-currents were observed. However, zonal transport results (Figure 3.63) the multibanded structure is not so evident like in subregion A1, but it is still present. In ROMS–20C3M (Figure 3.63a) the ACoC westward flow is observed between 70°S and 68°S , over the continental slope. The first core occurs at approximately 69.5°S and the second at approximately 68°S . The southern core shows a transport that exceeds 0.8 Sv within the first 150 m depth, while the northern branch has an associated transport of 0.59 Sv .

In ROMS–SRESA2 (Figure 3.63b), both cores are observed at the same latitude and they are weaker than in ROMS–20C3M. In the 21th century scenario, the southern core shows a transport of 0.39 Sv , while the northern core has an associated transport of 0.19 Sv .

In subregion A2 the presence of the counter-current is observed only in ROMS–20C3M (Figure 3.63a) over the continental slope from 600 m to 2500 m depth, and it has a transport of 0.01 Sv .

Seasonal zonal transports are shown in Figure 3.64. During summer the transport associated with the ACoC is weaker and the multibanded structure is absent in both scenarios. In ROMS-20C3M it is observed one core with a westward transport that ranges from 0.59 Sv at surface to 0.19 Sv at 1000 m during summer (Figure 3.64a). In ROMS-SRESA2 (Figure 3.64c) the ACoC transport is weaker and it varies from 0.59 Sv at surface to 0.19 Sv up to 600 m. Furthermore, the ACoC showed to be narrower in ROMS-SRESA2.

Figure 3.64: **Seasonal longitudinal mean zonal transport cross sections in subregion A2.** ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) zonal transport cross sections over the 1980–1999 period; ROMS-SRESA2 summer (DJF) (c) and winter (JJA) (d) zonal transport cross sections over the 2021–2047 period.

During winter, the ACoC transport is stronger. In ROMS-20C3M (Figure 3.64b) it exceeds 0.8 Sv from the surface up to 350 m in the southern core. The northern core shows a transport of 0.59 Sv in the 20st century scenario. In ROMS-SRESA2 (Figure 3.64d), transports above 0.8 Sv are found only at the surface, in the southern core. Throughout the water column the transport is around 0.39 Sv up to 800 m. The northern core also presents a transport of 0.39 Sv up to 600 m.

In contrast to subregion A1, the seasonal cross sections in subregion A2, show the presence of a counter-current only during summer. In both scenarios this current occurs over the continental slope approximately from 400 m to 2500 m depth and it has a transport of 0.01 Sv (Figure 3.64). In the winter months this counter-current occurs only in ROMS-20C3M at 1000 m and is weaker (Figure 3.64b).

In subregion A2 the temperatures were higher in the 21th century scenario, almost at the entire column Figure 3.65. At the surface, the temperatures were around -1.16°C in ROMS-20C3M (Figure 3.90c). In ROMS-SRESA2 (Figure 3.91c) the temperatures were slightly higher, reaching -1°C in the northern boundary, except in the southernmost location where temperatures below -1.32°C are observed.

Figure 3.65: **Longitudinal mean temperature cross sections in subregion A2.** (c) ROMS-20C3M temperature cross section over the 1980–1999 period; (c) ROMS-SRESA2 temperature cross section over the 2021–2047 period.

Figure 3.65 also shows a better representation of the WDW which are found from 200 m to 1500 m, further north. Near the ASF, this water mass occurs from approximately 400 m to 1000 m depth in both scenarios. However, in ROMS-20C3M (Figure 3.90c) this

water mass reaches positions further south, close to the ASF front at 800–1000 m depth. In ROMS–SRESA2 (Figure 3.91c) the southernmost position that this water mass reaches is around 68.5°S.

Figure 3.66: **Seasonal longitudinal mean temperature cross sections in subregion A2.** ROMS–20C3M summer (DJF) (a) and winter (JJA) (b) cross sections over the 1980–1999 period; ROMS–SRESA2 summer (DJF) (c) and winter (JJA) (d) temperatures cross sections over the 2021–2047 period.

In subregion A2 the WDW showed higher temperature values, which are around 0.4°C in ROMS–20C3M (Figure 3.90c), and exceed 0.5°C in ROMS–SRESA2 (Figure 3.91c). Relative to the salinity (Figure 3.68), there is no evidence of great changes from the 20th

to 21st century scenario in the water masses properties.

Seasonal temperatures are shown in Figure 3.66. Temperatures in ROMS–SRESA2 are overall slightly higher than in ROMS–20C3M. However, some exceptions are observed in both seasons. During summer, colder waters are found at 70°S in ROMS–SRESA2 (Figure 3.66c). Whereas in the winter months (Figure 3.66d), smaller temperatures are observed at surface and over the continental slope, between 68°–70°S.

Although the WDW is found in northern position in ROMS–SRESA2, this water mass reaches warmer temperatures in ROMS–SRESA2 that exceeds 0.5°C. In both seasons higher temperatures associated with the WSDW and WSBW are also observed.

Figure 3.67: **Seasonal differences in the temperature field in subregion A2.** (a) differences between summer and winter temperatures in ROMS–20C3M (1980–1999); (b) differences between summer and winter temperatures in ROMS–SRESA2 (2021–2047).

Relative to seasonal differences, both scenarios show higher temperatures during summer at the surface and over the continental slope (Figure 3.67). The temperature differs from 1°C to 0.02°C. In approximately 150 m depth, negative differences of up to 0.2°C are found.

Figure 3.68: **Longitudinal mean salinity cross sections in subregion A2.** (b) ROMS–20C3M salinity cross section over the 1980–1999 period; (b) ROMS–SRESA2 salinity cross section over the 2021–2047 period.

Seasonal salinities (Figure 3.69) show that minimum salinity layer (< 34.3) in ROMS–20C3M reaches deep layers over the continental slope. During summer (Figure 3.69a) this salinity is observed up to 300 m depth, while during winter (Figure 3.69b) it reaches 500 m depth.

In ROMS–SRESA2 this minimum salinity layer is found up to 250 m during summer months (Figure 3.69c) and 400 m in the winter (Figure 3.69d).

Below the minimum salinity layer until 800 m depth, the salinity values also showed to be slightly smaller in ROMS–SRESA2, in both seasons. Compared to ROMS–20C3M, the isohalines are found displaced further to the bottom.

Relative to the seasonal differences, in ROMS–20C3M salinity values are up to 0.1 higher over the continental slope, and 0.1 smaller at the surface during summer (Figure 3.70a). ROMS–SRESA2 also shows a negative difference up to 0.1 at the surface (Figure 3.70b). Over the continental slope, a pronounced positive difference is also found. The salinity in this regions differs up to 0.06.

Figure 3.69: Seasonal longitudinal mean salinity cross sections in subregion A2. ROMS–20C3M summer (DJF) (a) and winter (JJA) (b) cross sections over the 1980–1999 period; ROMS–SRESA2 summer (DJF) (c) and winter (JJA) (d) salinity cross sections over the period 2021–2047.

Figure 3.70: **Seasonal differences in the salinity field in subregion A2.** (a) differences between summer and winter salinities in ROMS-20C3M (1980–1999); (b) differences between summer and winter salinities in ROMS-SRESA2 (2021–2047).

Figure 3.71 shows the TS diagrams obtained for the subregion A2. As shown in the cross sections, the WDW is better represented in this region. In ROMS-20C3M (Figure 3.71a) this water mass has a mean temperature of 0.4°C , but it reaches values of up to 0.7°C . In ROMS-SRESA2 (Figure 3.71b), although the mean temperature for the 2021–2047 period was of 0.5°C , temperatures above 1°C are observed between these years.

Both the WSDW and the WSBW showed salinities higher than the observed (Table 1.1) in both scenarios. In ROMS-20C3M, the WSBW reaches temperatures up to 1.4°C whereas in ROMS-SRESA2 the minimum temperature is of -1°C .

Figure 3.71: **Scatter TS diagram in subregion A2.** (a) ROMS-20C3M TS diagram for the 1980–1999 period; (b) ROMS-SRESA2 TS diagram for the 2021–2047 period. The solid black line represents the mean temperature and salinity in each depth point.

In subregion A2 the SW is absent. This is because this region has a narrow shelf, and continental shelf is not accessed by regional model. Due to its narrow shelf, the ACoC occurs over the continental slope in the region of the ASF as shown by previous studies (e.g. Heywood et al. (1998)). Therefore, this fact does not limitate the study of the ACoC in this region.

3.2.2.3.3 Subregion A3

The last subregion analyzed comprises the region along the Antarctic Peninsula between 62°S–73°S and 58°W–50°W. In this subregion ACoC transport is nearly meridional and, as shown by Hovmöller diagrams (Figure 3.72), the transport magnitude increases to the north, as it approaches the ACC.

Figure 3.72: **Hovmöller digrams of the volume transport in subregion A3.** (a) ROMS-20C3M transport obtained from the monthly means over the 1980–1999 period; (b) ROMS-SRESA2 transport obtained from the monthly means over the 2021–2047 period.

Figure 3.73: Volume transport annual cycle in subregion A3 from ROMS-20C3M (blue) and ROMS-SRESA2 (red)

Differences from the 20th to the 21st century scenario in subregion A3 are not so pronounced as in the subregion A2, but the ACoC transport magnitude also shows a decrease in this region. In ROMS-20C3M (Figure 3.72a) five cores are found where the ACoC is stronger: at 63°S, 64°S, 66°S, 68°S, and 72°S. The maximum magnitude is of 7 Sv at 63° and the minimum is of 3.8 Sv at 68°S. The transport is intensified from May to September, as also shown in the annual cycle (Figure 3.73).

Four of the five cores represented by ROMS–20C3M simulation are observed in ROMS–SRESA2 (Figure 3.72b). The weaker core at 68°S is absent in ROMS–SRESA2. In ROMS–SRESA2 the maximum magnitudes are around 5.6 Sv.

The annual cycle (Figure 3.73) show a small decrease in the ROMS–SRESA2 transport magnitude from June to July followed by an increase in August, when it reaches a maximum of 3.4 Sv.

The longitudinal mean of the meridional transport is shown in Figure 3.74. As shown in Hovmöller diagrams, the transport increases to the north. The northern core occurs at about 64°S and shows a transport of about 0.31 Sv in both scenarios. However, in ROMS–SRESA2 (Figure 3.74b) this core is narrower than in ROMS–20C3M (Figure 3.74a).

Figure 3.74: **Longitudinal mean meridional transport cross section in subregion A3.**(a) ROMS–20C3M zonal transport cross section for the 1980–1999 period; (b) ROMS–SRESA2 zonal transport for the 2021–2047 period.

The seasonal fields (Figure 3.75) show the weakening of the transport during summer. In ROMS–20C3M (Figure 3.75a) the maximum transport observed in this season is of 0.31 Sv on the northern core. During winter months (Figure 3.75b) it reaches up to 0.46 Sv at the same location. In ROMS–SRESA2 the maximum transport during summer is of 0.21 Sv (Figure 3.75c), and in the winter months (Figure 3.74b) it increases to 0.41 Sv.

Figure 3.75: **Seasonal longitudinal mean meridional transport cross sections in subregion A3.** ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) zonal transport cross sections over the 1980–1999 period; ROMS-SRESA2 summer (DJF) (c) and winter (JJA) (d) zonal transport cross sections over the period 2021–2047.

In the subregion A3, the temperature pattern in the water column is similar to the observed in the subregion A1 (Figure 3.76). ROMS-SRESA2 (Figure 3.94c) also shows higher temperatures than ROMS-20C3M (Figure 3.93c).

In ROMS-20C3M, temperatures over the shelf are below -1.64°C and the isotherm of -1.32°C is found up to 410 m. Whereas in ROMS-SRESA2 an increase in the intermediate waters temperatures displaced the isotherms toward the surface. Over the continental

shelf, temperatures below -1.64°C are found only at the surface, and below that layer, the temperatures remains up to -1.32°C .

In this region, the WDW is also not well represented. In both scenarios temperatures above zero were not observed, and in ROMS-20C3M the highest temperatures occurs between 800 and 2000 m depth.

Figure 3.76: **Longitudinal mean temperature cross sections in subregion A3.** (c) ROMS-20C3M temperature cross section over the 1980–1999 period; (c) ROMS-SRESA2 temperature cross section over the 2021–2047 period.

Figure 3.77 shows the seasonal temperature fields. Differences between two scenarios are the same as described above.

Figure 3.77: **Seasonal temperature cross sections in subregion A3.** ROMS-20C3M summer (DJF) (a) and winter (JJA) (b) temperature cross sections over the 1980–1999 period; ROMS-SRESA2 summer (DJF) (c) and winter (JJA) (d) temperature cross sections over the 2021–2047 period.

Relative to seasonal differences, the temperature is up to 1°C higher in summer than winter for both scenarios (Figure 3.78). In ROMS-20C3M (Figure 3.78a) negative differences are observed from the continental slope to 64°S . However, this negative difference is more pronounced over the continental slope where the difference is up to 0.2°C .

In ROMS-SRESA2 (Figure 3.78b) a small negative difference, of about 0.02°C , is observed over the continental slope. Further north, in the intermediate layer, differences

are positive and the temperatures is 0.2°C higher during summer.

Figure 3.78: **Seasonal differences in the temperature field in subregion A3.** (a) differences between summer and winter temperatures in ROMS-20C3M (1980–1999); (b) differences between summer and winter temperatures in ROMS-SRESA2 (2021–2047).

Figure 3.79: **Longitudinal mean salinity cross sections in subregion A1.** (a) ROMS-20C3M salinity cross section over the 1980–1999 period; (b) ROMS-SRESA2 salinity cross section over the 2021–2047 period.

Salinity fields (Figure 3.79) show that from 20th to 21st century salinity decreased. However, an increase in the salinity is observed in the intermediate layer. As well as the

temperature, this increase displaces the isohalines upward. Therefore, below surface, in ROMS–SRESA2 (Figure 3.94b) salinity is higher than in ROMS–20C3M (Figure 3.93b). The same differences between both scenarios are also observed in the seasonal climatologies (Figure 3.80).

Figure 3.80: **Seasonal longitudinal mean salinity cross sections in subregion A3.** ROMS–20C3M summer (DJF) (a) and winter (JJA) (b) salinity cross sections over the 1980–1999 period; ROMS–SRESA2 summer (DJF) (c) and winter (JJA) (d) salinity cross sections over the 2021–2047 period.

In both scenarios, great differences between summer and winter salinities are found at the surface layer, which shows higher salinity during winter. The differences were

quantified and are shown in Figure 3.81. In both scenarios, the salinity difference is up to 0.1 at the surface. However, in ROMS–SRESA2 (Figure 3.81b) negative differences (up to 0.02) are also observed over continental shelf and slope. Further north differences are positive and salinity is 0.03 higher during summer.

In ROMS–20C3M (Figure 3.81a) differences below the surface are overall positives and differs up to 0.04. Small negative differences are found over the continental slope and are around 0.02.

Figure 3.81: **Seasonal differences in the salinity field subregion A3.** (a) differences between summer and winter salinities in ROMS–20C3M (1980–1999); (b) differences between summer and winter salinities in ROMS–SRESA2 (2021–2047).

Figure 3.82: **Scatter TS diagram in subregion A3.** (a) ROMS-20C3M TS diagram for the 1980–1999 period; (b) ROMS-SRESA2 TS diagram for the 2021–2047 period. The solid black line represents the mean temperature and salinity in each depth point.

TS diagrams are shown in Figure 3.82. As observed in cross sections, the WDW mean temperatures are below zero in both scenarios. In ROMS-20C3M (Figure 3.82a) the mean temperature is -0.3°C , while in ROMS-SRESA2 (Figure 3.82b) it is around 0.4°C . However, in ROMS-SRESA2 is observed the occurrence of temperature up to 0.4°C associated with this water mass.

In the subregion A3 it is also observed the presence of SW, which shows values more consistent with LSSW. The WSDW and the WSBW also show higher salinities than observed (Table 1.1) in subregion A3. The minimum WSBW temperature in ROMS-20C3M is -1.2°C , whereas in ROMS-SRESA2 the minimum temperature is -0.8°C .

3.2.2.4 ACoC variability

In the previous analysis ACoC showed a weakening from the 20th to the 21st century scenario. Núñez-Riboni and Fahrbach (2009) showed that the variability of this current is controlled by the interaction between mutual mechanisms. Therefore, in this section we investigated the ACoC variability and its relation with some of the main forcings of this current.

The analyses in this section were performed in two transects (Figure 3.83), the first is defined in the East Weddell Sea (Figure 3.83a), where the ACoC is stronger, and

the second in the Antarctic Peninsula coast (Figure 3.83b), where this current show a meridional transport. As mentioned before, subregion A1 comprises a huge area and overlaps subregion A3. Therefore, in the analyses of this section we decided to investigate the ACoC variability only in these two regions.

Figure 3.83: Transects in the subregion A2 (a) and in subregion A3 (b) are marked with the dotted lines.

As described by Núñez-Riboni and Fahrbach (2009) the ACoC has two components, the barotropic and baroclinic. Therefore, before analyze the current variability we first evaluated the SSH (Figure 3.84), temperature and salinity gradients (Figure 3.85 to Figure 3.88).

Figure 3.84 shows that both zonal and meridional SSH gradients decreased from the 20th to the 21st century scenario. In the East Weddell Sea, ACoC is nearly zonal. Therefore, its barotropic component is proportional to the meridional SSH gradient. In this region, ROMS-20C3M (Figure 3.84a) show a negative gradient of 0.41×10^{-5} , whereas in ROMS-SRESA2 (Figure 3.84c) a decrease of this gradient magnitude to $0.33 \text{ m} \times 10^{-5}$, is observed.

In the Antarctic Peninsula coast, where the transport is nearly meridional, a decrease in the magnitude of the zonal SSH gradient is also observed. In ROMS-20C3M (Figure 3.84b) it reaches a negative gradient of 0.57×10^{-5} , whereas in ROMS-SRESA2 (Figure 3.84d) it decreases to 0.43×10^{-5} .

Figure 3.84: **SSH gradient maps.** ROMS-20C3M meridional (a) and zonal (b) SSH gradients for the 1980–1999 period; ROMS-SRESA2 meridional (c) and zonal (d) SSH gradients for the 2021–2047 period.

In the previous sections we observed changes in the termohaline. In all subregions analyzed we observe a warming of the intermediate waters from the 20th to the 21st scenario. Relative to the salinity, in subregion A2 a small decrease in the intermediate waters was observed in ROMS-SRESA2, whereas in subregion A3 it was observed an increase in this scenario.

As discussed before, the ASF has an important role to the dynamics of the coastal region. Changes in the hydrographic properties could influence the ASF and the baroclinic component of the ACoC. Therefore, we also investigated the behavior of the temperature and salinity gradients in the transects of subregions A2 and A3.

In subregion A2 an increase in the temperature gradient is observed over the continental slope from the 20th to the 21st scenario (Figure 3.85). In ROMS-20C3M (Figure 3.85b) there is a gradient of $0.4 \times 10^{-5} \text{ } ^\circ\text{C} \cdot \text{m}^{-1}$, whereas in ROMS-SRESA2 (Figure 3.85d) it reaches $1.5 \times 10^{-5} \text{ } ^\circ\text{C} \cdot \text{m}^{-1}$. The strongest gradient in ROMS-SRESA2 does not allow the WDW intrusion on the continental slope, which explains why this water mass is found

further north.

The increase in temperature gradient is followed by a decrease in the salinity gradient in subregion A2 (Figure 3.86). In ROMS-20C3M (Figure 3.86b) the salinity gradient reaches up to $0.4 \times 10^{-5} \text{ m}^{-1}$ while in ROMS-SRESA2 (Figure 3.86d) it not exceed $0.35 \times 10^{-5} \text{ m}^{-1}$.

Figure 3.85: ROMS-20C3M temperature (a) and temperature gradient (b) in the subregion A2 transect over the 1980–1999 period; ROMS-SRESA2 temperature (c) and temperature gradient (d) in the subregion A2 transect over the 2021–2047 period.

Figure 3.86: ROMS–20C3M salinity (a) and salinity gradient (b) in the subregion A2 transect over the 1980–1999 period; ROMS–SRESA2 salinity (c) and salinity gradient (d) in the subregion A2 transect over the 2021–2047 period.

In the subregion A3 transect, it is observed an increasing further east in both temperature (Figure 3.87) and salinity (Figure 3.88) gradients towards the surface from the 20th to the 21st scenario. In the temperature the gradient is nearly zero at 54°W in ROMS–20C3M (Figure 3.87b), while in ROMS–SRESA2 (Figure 3.87d) it exceeds $1.5 \times 10^{-5} \text{ }^\circ\text{C} \cdot \text{m}^{-1}$. At this same location, the salinity gradient increased from $0.15 \times 10^{-5} \text{ m}^{-1}$ (Figure 3.88b) to $0.3 \times 10^{-5} \text{ m}^{-1}$ (Figure 3.88d).

Figure 3.87: ROMS–20C3M temperature (a) and temperature gradient (b) in the subregion A3 transect over the 1980–1999 period; ROMS–SRESA2 temperature (c) and temperature gradient (d) in the subregion A3 transect over the 2021–2047 period.

Figure 3.88: ROMS–20C3M salinity (a) and salinity gradient (b) in the subregion A3 transect over the 1980–1999 period; ROMS–SRESA2 salinity (c) and salinity gradient (d) in the subregion A3 transect over the 2021–2047 period.

After analyze the behavior of the gradients we evaluated the ACoC response to those changes observed. For that we analyzed the barotropic and baroclinic components the two transects at 200 m depth.

Figure 3.89: Barotropic (orange) and baroclinic (green) components of the zonal current in subregion A2 transect at 200 m depth. **(a)** ROMS–20C3M (1980–1999) and **(b)** ROMS–SRESA2 (2021–2047).

Figure 3.89 shows the barotropic and baroclinic components of zonal velocity in the transect in subregion A2. In ROMS–20C3M (Figure 3.89a) the current showed to be almost entire barotropic, as showed by previous studies (Sverdrup, 1953; Fahrbach et al., 1992; Núñez-Riboni and Fahrbach, 2009). The baroclinic component has a eastward flow and is near zero.

Relative to ROMS–20C3M, ROMS–SRESA2 (Figure 3.89b) shows a strengthening of the baroclinic counter-current, which reaches magnitudes near of those observed for the barotropic component, but in the opposite direction. As seen before, the thermal wind is the mechanism responsible for the development of the counter current. According to Núñez-Riboni and Fahrbach (2009) the thermal wind is the common vehicle to the ACoC driving mechanisms that induce variability to both components.

Therefore, at subregion A2 the observed ACoC weakening is explained by the increase of the baroclinic component which acts as a counter-current to the barotropic field. From 2038 to 2048 a decrease in both components is observed, but the magnitude of the baroclinic component remains near barotropic component.

Figure 3.90: (a) CCSM3–20C3M zonal wind in subregion A2 transect; ROMS–20C3M salinity (b) and temperature (c) at 200 m depth in the subregion A2 transect.

Table 3.4: Pearson’s correlation coefficient (r) between forcings and barotropic and baroclinic components in subregion A2 obtained from ROMS–20C3M

	Winds	Salinity	Temperature
Barotropic	0.38	0.41	-0.10
Baroclinic	-0.23	-0.20	0.32

To evaluate the causes of these changes we compared the results with winds, salinity and temperature over the same period and depth (in the case of salinity and temperature).

As suggested by previous studies (Núñez-Riboni and Fahrbach, 2009; Fahrbach et al., 1992; Sverdrup, 1953), the prevailing easterly winds are the main driver of ACoC. In the results obtained in section 3.2.1 we showed that there are small differences between the 20th and 21st century wind fields. The time series obtained from CCSM3–CAM3 in subregion A2 transect (Figure 3.90a and Figure 3.91a), used to force regional simulations, also shows that from 20th to 21st century scenario there is no significant changes in the easterly magnitudes.

The thermohaline fields results (Figure 3.85 to Figure 3.88) also suggests that changes observed in the region are not due to the winds. As showed by Núñez-Riboni and Fahrbach (2009), the winds also have an important contribution to baroclinic field by depressing the isopycnals due to Ekman transport. In the subregion A2 we observed a decrease in the easterlies magnitude from 20th to 21st century (Figure 3.32b). With the weakening of the winds, it was expected a reduction in the associated Ekman pumping, and a consequent warming in the water column since surface cold waters would not to be pumped to the bottom. Therefore, it would reduce the gradient southward and would allow the intrusion of WDW on the continental slope. However, it was observed the opposite with a decrease of temperature in the 21st century scenario (Figure 3.85c).

The winds also showed the weaker and non-significant (at 95% level) correlation with both barotropic and baroclinic components (Table 3.4 and Table 3.5). Therefore, our results suggests that the wind is not the main forcing contributing to the changes in the modeled ACoC.

In the case of the winds, it is important to highlight that the winds used are not output of regional model, but the winds used to force it. Therefore, the difference in the resolution could affect the comparison between the results obtained from regional model with the dataset used to force it.

Relative to temperature and salinity in ROMS–20C3M a decrease in the temperature (Figure 3.90c) and salinity (Figure 3.90b) are observed in the end of the period analyzed between 1994–1999. However, it is not observed any pronounced variation in both barotropic and baroclinic fields in the 20th century scenario (Figure 3.89a).

In the 21st century scenario the salinity (Figure 3.91b) is generally higher than in the 20th century scenario (Figure 3.90b) and the temperature (Figure 3.91c) is lower, except between 2030 and 2038 when a warming pulse occurs accompanied by a decrease in the salinity. After this period, between 2038 and 2047, same period which is observed a

decrease in the barotropic and baroclinic current in ROMS–SRESA2 (Figure 3.89b), there is a significant increase in the salinity accompanied by a decrease in temperature (Figures 3.91b and 3.91c, respectively).

As seen in previous results the decrease in temperature resulted in the strengthening of the temperature gradient (Figure 3.85), while the salinity gradient (Figure 3.86) became weaker in the subregion A2 from the 20th to the 21st century. However, the results obtained in this study suggests that the more likely cause to the increase in the baroclinic component is the increase in the density gradient due to the decrease in the temperature.

The correlation results (Table 3.4 and Table 3.5) showed that the ROMS–SRESA2 temperature (Figure 3.91c) have a moderate and significant (at 95% level) correlation with both components, while although significant the correlation between salinity and current is weak.

Figure 3.91: **(a)** CCSM3–SRESA2 zonal wind in subregion A2 transect; ROMS–SRESA2 salinity **(b)** and temperature **(c)** at 200 m depth in subregion A2 transect.

Table 3.5: Pearson’s correlation coefficient (r) between forcings and barotropic and baroclinic components in subregion A2 obtained from ROMS–SRESA2

	Winds	Salinity	Temperature
Barotropic	0.3	0.27	-0.50
Baroclinic	-0.31	-0.30	0.50

Figure 3.92: Barotropic (orange) and baroclinic (green) components of the meridional current in subregion A3 transect at 200 m depth. **(a)** ROMS–20C3M (1980–1999) and **(b)** ROMS–SRESA2 (2021–2047).

Figure 3.92 shows the barotropic and baroclinic currents in the subregion A3. In both scenarios meridional velocity is mainly barotropic and the contribution of the baroclinic counter current is small. In ROMS–20C3M (Figure 3.92a) an increase in the barotropic component is observed in the end of the period analyzed, from 1997 to 1999. Whereas in ROMS–SRESA2 (Figure 3.92b) a pronounced decrease in barotropic current is observed from 2040 to 2047, followed by a small decrease in baroclinic component.

Relative to the forcings, as well as in subregion A2, the meridional winds also not show significant differences from 20th to 21st scenario (Figure 3.93a and Figure 3.94a, respectively) in subregion A3 transect. Moreover, weaker and non-significant, at 95% level, correlations were found in the comparison between winds and currents (Table 3.6 and Table 3.7). Therefore, changes in barotropic component are not related with the winds.

From 2038 to 2047, when we observed their decrease in ACoC components, is also observed an increasing in the temperature (Figure 3.94c) and salinity (Figure 3.94b). The results also showed that in ROMS–SRESA2 (Table 3.7) there is a strong and significant (at 95% level) with temperature in both components, which implies that at this region ACoC variability is also influenced by changes in the temperature field.

Therefore, the results presented in this section showed that changes in the modeled ACoC are not due to the winds. Although there are regional differences, changes from the 20th to the 21st century scenario are mainly due to changes in the termohaline forcing which are probably related with other mechanisms not analyzed in this study (e.g. sea ice).

Figure 3.93: **(a)** CCSM3-20C3M meridional wind in subregion A3 transect; ROMS-20C3M salinity **(b)** and temperature **(c)** at 200 m depth in subregion A3 transect.

Table 3.6: Pearson's correlation coefficient (r) between forcings and barotropic and baroclinic components in subregion A3 obtained from ROMS-20C3M

	Winds	Salinity	Temperature
Barotropic	0.08	0.21	0.07
Baroclinic	-0.06	-0.23	-0.18

(a)

(b)

(c)

Figure 3.94: (a) CCSM3-SRESA2 meridional wind subregion A3 transect; ROMS-SRESA2 salinity (b) and temperature (c) at 200 m depth in subregion A3 transect.

Table 3.7: Pearson’s correlation coefficient (r) between forcings and barotropic and baroclinic components in subregion A3 obtained from ROMS–SRESA2

	Winds	Salinity	Temperature
Barotropic	0.20	-0.32	-0.75
Baroclinic	-0.10	0.17	0.79

4 Conclusions

This study was divided in two parts. Since regional simulations were forced with an Ocean General Circulation Model, the first part of this study aimed to characterize and understand the behavior of the wind stress, temperature and salinity from a large-scale perspective. We evaluate the results from the ocean component of the NCAR–CESM1 which were compared to SODA3.4.1 reanalysis.

The second part consists in the main goal of the present study which was to characterize the ACoC and investigate its changes from the 20th to the 21st century through high resolution downscaling experiments.

Conclusions Part I:

CESM1–POP2 overall represented satisfactorily the mean patterns in the Southern Ocean. The asymmetries found in the Southern Ocean wind stress were reproduced with the intensification of the westerlies in the Indian sector by both model and reanalysis. Compared to SODA3.4.1, CESM1–POP2 has shown greater wind stress magnitudes associated with the westerlies and at some locations near the coastal region.

Significant increase in the trends associated with the westerlies were found, evidencing the strengthening reported in the last decades (e.g. [Thompson and Solomon \(2002\)](#)). The poleward shift of these winds usually associated with strengthening was not observed.

Biases in the salinity and temperature were found in the Weddell Sea mainly during summer.

Relative to the water masses in the Weddell Sea, surface waters were better represented in CESM1–POP2 than in SODA3.4.1. The WDW was the best represented water mass in both model and reanalysis, although it reached greater depths than observed in the available literature (e.g. [Kerr et al. \(2009\)](#)).

The simulated WSDW presented consistent temperature and salinity values. The SW and the WSBW, which are important for this water mass formation, are absent in CESM1–POP2 and SODA3.4.1. The absence of these water masses and the ASF, points to the limitation of climate models to reproduce important features of the coastal dynamics.

Conclusions Part II:

To investigate coastal dynamics it is important to capture physical processes over the continental shelf. In this study we evaluate downscaling simulations performed using ROMS which is more appropriate for this purpose.

Results for the 20th century scenarios (ROMS–NCEP and ROMS–20C3M) were capable to reproduce the ACoC structure in the Weddell Sea. Results showed that the simulation forced with NCAR CCSM3 historical experiment (20C3M) provided a better representation of the ACoC than the run forced by reanalysis which underestimated the current velocities.

Regional simulations were capable to reproduce the ACoC multibanded structure and the counter current, described in previous studies which used observations to analyze this current.

From the 20th to the 21st century scenario weakening of the ACoC current was observed. Therefore, we evaluated the behavior of the winds, temperature and salinity to investigate the possible causes of the ACoC changes. The winds used were the same used to force regional simulations because we did not have access to regional outputs.

Significant differences, at 95% level, in the magnitudes of the winds between 20th and 21st century scenario, were found. The differences suggested an increase in the westerlies further south and in the meridional wind along Antarctic Peninsula coast, and a decrease in the easterlies magnitudes. Differences in the westerlies showed to be more pronounced during summer which can be an evidence of the more pronounced strengthening during this season.

Relative to temperature and salinity, regional simulations provided a better representation of ASF which separates the continental colder and fresher waters from the warmer and saltier waters offshore.

In the Weddell basin region, WDW temperature was underestimated by the regional model. This water mass was better represented in the subregion located at the East Weddell Sea, where the cross sections were extended further north.

In the selected Weddell basin regions, where the continental shelf is wide, regional simulations results indicate the presence of SW. These water masses temperatures and salinities show values closer to those of the LSSW, but not of the HSSW which is the major source for WSDW formation. Despite this, the WSDW and the WSBW were well represented in all subregions analyzed, although they are saltier and denser than observed in the available literature.

Overall, we observed an increase in the intermediate and deep water temperatures

from the 20th to the 21st projection, which is probably related with the WDW warming. With respect to salinity regional differences were detected. While in the regions located in the western Weddell Sea an increase in the intermediate waters salinity is observed, in the East Weddell the salinity decreased.

In the East Weddell Sea, an increase in the temperature gradient was accompanied by a decrease in the salinity gradient from the 20th into the 21st century. There, the contribution of the baroclinic component increased and reached almost the same magnitude of the barotropic component but in the opposite direction. Therefore, the decrease in the ACoC transport is explained by the destructive interference between barotropic and the baroclinic counter-current in this region.

The results suggest that the baroclinic component increase is mainly due to the increase in the temperature gradient, which showed a significant and moderate correlation with the barotropic and baroclinic currents. The correlation of temperature with the barotropic component was negative which indicates temperature changes are out of phase with the barotropic velocity.

Along the Antarctic Peninsula coast, an increase in the temperature and salinity gradients in the ASF region were observed. From the 20th to 21st projection baroclinic current is slightly stronger, but differences are not significant. However, at the end of the analyzed period for ROMS-SREAS2, from 2040 to 2047, a pronounced decrease in the barotropic component, followed by a decrease in the baroclinic component, was observed. Both barotropic and baroclinic components showed a strong correlation with salinity suggesting that in this region, changes in ACoC are more likely caused by the increase in the salinity.

The wind results were poorly correlated to barotropic and baroclinic currents. Although, differences due to resolution could affect comparison between the results obtained from regional model and the dataset used to force it, the thermohaline forcings and SSH changes are not consistent with changes observed in the winds.

Therefore, we concluded that changes from the 20th to the 21st century, in the modeled ACoC, are not influenced by the wind forcing, but by the thermohaline changes which is probably related with other mechanisms not evaluated in this study (e.g. sea ice).

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