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The Weddell Gyre, Southern Ocean: Present Knowledge and Future Challenges

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Key Points:

- Major research priorities to advance understanding of the Weddell Gyre are identified and justified against current knowledge
- Interdisciplinary approaches are needed to support system science research of the Weddell Gyre and promote collaborative projects
- Winter conditions, connectivity to lower latitudes, and intensification of the gyre are the main interdisciplinary priorities

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Abstract The Weddell Gyre (WG) is one of the main oceanographic features of the Southern Ocean south of the Antarctic Circumpolar Current which plays an influential role in global ocean circulation as well as gas exchange with the atmosphere. We review the state-of-the art knowledge concerning the WG from an interdisciplinary perspective, uncovering critical aspects needed to understand this system's role in shaping the future evolution of oceanic heat and carbon uptake over the next decades. The main limitations in our knowledge are related to the conditions in this extreme and remote environment, where the polar night, very low air temperatures, and presence of sea ice year-round hamper field and remotely sensed measurements. We highlight the importance of winter and under-ice conditions in the southern WG, the role that new technology will play to overcome present-day sampling limitations, the importance of the WG connectivity to the low-latitude oceans and atmosphere, and the expected intensification of the WG circulation as the westerly winds intensify. Greater international cooperation is needed to define key sampling locations that can be visited by any research vessel in the region. Existing transects sampled since the 1980s along the Prime Meridian and along an East-West section at ~62°S should be maintained with regularity to provide answers to the relevant questions. This approach will provide long-term data to determine trends and will improve representation of processes for regional, Antarctic-wide, and global modeling efforts—thereby enhancing predictions of the WG in global ocean circulation and climate.

Plain Language Summary The Weddell Gyre is one of the main oceanographic features in the ocean surrounding Antarctica, the Southern Ocean. Although located far from other continents, this polar region affects the planet through the exchange of gases between frigid ocean waters and the atmosphere, regulating oxygen and carbon dioxide farther north. Studying the Weddell Gyre is challenging, as sea ice covers the ocean surface year around, restricting access by research ships and sensing of ocean surface from satellites. New technology is now available to avoid past limitations, autonomous underwater vehicles, instruments flown by planes, and floats instrumented with sea-ice detection. Only through international collaboration can we obtain adequate data to populate environmental models and study key areas in the gyre or hot spots. In this review we identify the missing links in our knowledge of the gyre, proposing research to address those questions. Three aspects are critical to understanding the processes that drive the gyre's oceanography, ice, geology, chemistry, and biology: winter and under-ice conditions that set the stage for the evolution of physics, ice, and biogeochemistry; exchange of water, material, and energy (or heat) with lower latitudes; and intensification of the clockwise circulation of the gyre with changes in winds.

1. Introduction

The Weddell Gyre (WG) is in many respects an unusual and extreme region, directly controlling the properties of substantial parts of the global deep (200–3,000 m) and abyssal (>3,000 m) ocean and indirectly shaping global climate. South of the Antarctic Circumpolar Current (ACC)—the strongest current system on Earth—the WG is the largest of several coherent oceanic regions (Figure 1), spanning about 5.6×10^6 km^2. Together with the Ross Gyre, the WG comprises the southernmost open ocean reaches on the planet (Figure 1). Extremely low sea surface temperatures and extensive sea-ice formation are characteristics of this region. Consequently, it is a foremost location of deep and bottom water formation on the Southern Hemisphere. This region also plays an important role in regulating air-sea exchanges and heat flow as a result of a delicate balance between upwelling of relatively warm subsurface waters and via its sea-ice cover (Fahrbach et al., 1994; Gordon & Huber, 1990). As outlined in more detail below, this strong connection between the atmosphere, surface ocean, and deep waters makes the WG arguably a most influential oceanic region with the capability of influencing global climate on time scales of hundreds to thousands of years.

The WG is situated in the Atlantic sector of the Southern Ocean, south of 55–60°S and roughly between 60°W and 30°E (Deacon, 1979). It stretches over the Weddell abyssal plain, where the Weddell Sea is situated, and extends east into the Enderby abyssal plain (10–20°E; Figure 2). The gyre is, however, a dynamic feature, with its eastern boundary not fixed to a location; it has been suggested that it might extend as far as 70°E (Mackintosh, 1972; Y.H. Park et al., 2001). The WG is principally driven by wind forcing (see section 3.1) and steered by topography, including continental boundaries and seabed structures. In the south and west (50°W), the gyre is bordered by the Antarctic mainland and its northward extension, the Antarctic Peninsula. In the north, the boundary is formed by the frontal structure of the southern ACC, which in turn is topographically guided by the South Scotia Ridge in the west and the North Weddell Ridge in the east. In this way, the northern boundary of the gyre in the west is at about 60°S, whereas further east, it moves to about 55°S, locally extending further north into the South Sandwich Trench at about 30°W (Figure 2). East of about 20°E, the gyre boundary takes a southward direction following the course of the ACC, steered east of the Gunnerus Ridge. Along its southern and western boundaries, the WG is flanked by floating ice shelves, the seaward extension of the Antarctic ice sheet (depicted by the red contour in Figure 2; see section 4). They form as a result of the gravity-driven flow of inland ice toward the coast. Some of these ice shelves extend more than a hundred kilometers offshore, with the Filchner-Ronne Ice Shelf, which covers a portion of the southern Weddell Sea continental shelf over an area extent roughly the same size of Sweden, being the largest body of floating ice on Earth. The ocean cavities beneath ice shelves give rise to a unique environment where seawater interacts with glacial ice at depths up to over 1 km below sea level; numerous canyons run from the continental shelf break into the abyssal plains, connecting coastal and continental shelf processes to the Weddell basin. Waters deeper than 3,000 m are connected to the north through passages in the South Scotia Ridge. The WG can be subdivided into smaller regions. The small Powell and Jane basins are found extending off the Antarctic Peninsula. East of the Prime Meridian, the Lazarev and Riiser-Larsen Seas are found along the coast, separated by the (extension of the) Astrid Ridge (Figure 2). A major topographic structure significantly influencing ocean circulation, the timing of sea-ice retreat, and the subsequent productivity in this region is Maud Rise, an undersea promontory centered just east of the Prime Meridian at 65°S. Maud Rise also played a prominent role in the formation of the great Weddell Polynya, which occurred for three consecutive years in the 1970s (Carsey, 1980; Gordon, 1978), and recently reemerged during some short transient events, the latest in the austral winter of 2017. The formation of the great Weddell Polynya had a far-reaching and long-lasting impact on the water mass structure of the gyre by cooling the Warm Deep Water (WDW) and moving the vertical temperature distribution in the WG (Robertson et al., 2002).

From a global perspective, the Weddell Sea, located in the western part of the WG, plays a crucial role in driving the global overturning circulation (the global “ocean conveyor belt”) as a major site of Antarctic Bottom Water (AABW) formation (Orsi et al., 2002), a component of the ventilation of the global abyssal ocean. Formation of this water mass occurs off the southern and western ice shelves, where highly saline and oxygen-enriched shelf waters are produced as a result of extensive sea-ice formation and by brine rejection in coastal polynyas (Figure 3). The formation of these dense water masses in the Weddell Sea is facilitated by the large-scale cyclonic circulation of the WG, which is driven by westerlies in the north and easterlies along the southern coast (see section 3). The cyclonic circulation leads to divergence in the

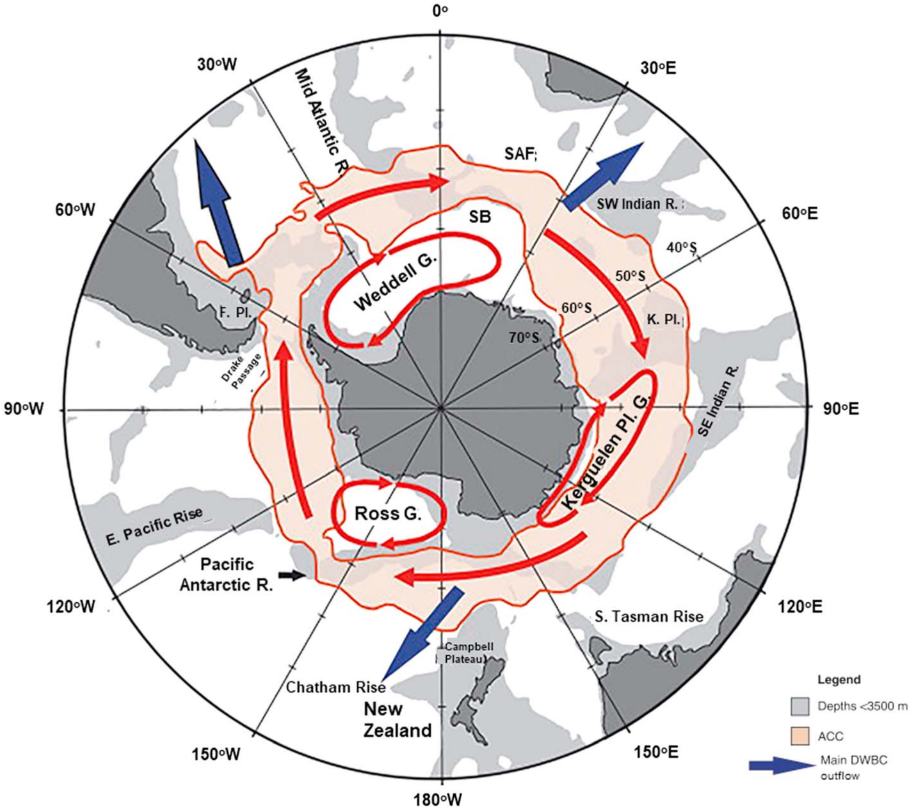

Figure 1. The main oceanographic features of the Southern Ocean: (i) the Antarctic Circumpolar Current (ACC) south of the Subantarctic Front (SAF) and southern limit of Upper Circumpolar Deep Water or southern boundary (SB); (ii) the Ross, Weddell, and Kerguelen Plateau gyres; and (iii) the main exit locations of deep western boundary currents (DWBC) from the Southern Ocean (blue arrows). Bathymetric elevations are annotated as R = ridge; K. Pl. = Kerguelen Plateau; F. Pl. = Falkland/Malvinas Plateau; and G. = gyre (Carter et al., 2008, reprinted with permission from Elsevier). Y. H. Park et al. (2009) called the gyre over the Kerguelen Plateau the Australian-Antarctic gyre.

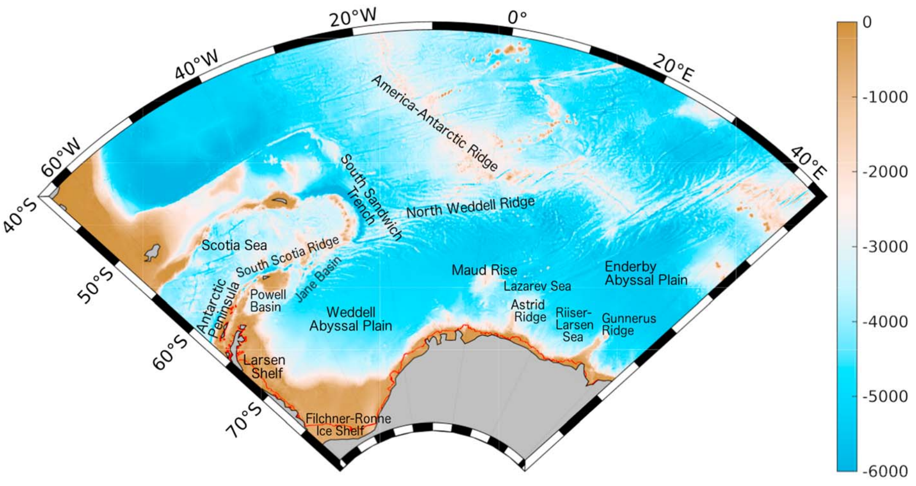

Figure 2. Topography map of the Weddell Gyre and its environs with the features outlined in the text, as in D. E. Hayes (1991; sose.ucsd.edu).

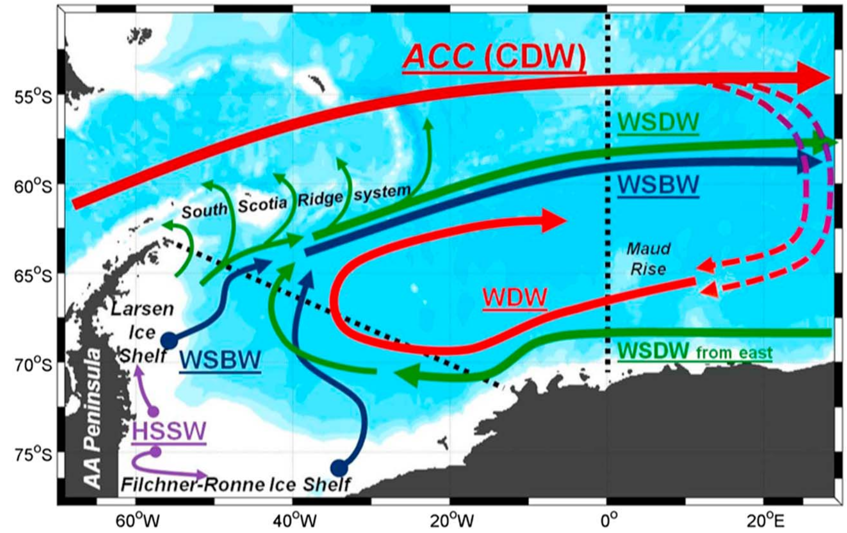

Figure 3. Map of the Atlantic sector of the Southern Ocean south of the Polar Front with main circulation pattern. Arrows indicate the circulation of deep water masses. Colors represent temperature from above 0 °C (red) to −0.7–0 °C (green) and below −0.7°C (dark blue). ACC = Antarctic Circumpolar Current; CDW = Circumpolar Deep Water; HSSW = High Salinity Shelf Water; WDW = Warm Deep Water; WSDW = Weddell Sea Deep Water; WSBW = Weddell Sea Bottom Water. Black stippled lines mark two repeat sections along the Prime Meridian and across the Weddell Sea. (Hellmer et al., 2016, Open Access).

interior area of the gyre, which induces deep upwelling, integrating the WG within the largest contiguous region of upwelling in the world (Marshall & Speer, 2012; Talley, 2013). Processes in the gyre margins are complex and locally include both upwelling (Foster & Carmack, 1976) and downwelling (Graham et al., 2013; Naveira Garabato et al., 2016; Nøst et al., 2011). Within the WG, different water masses interact and eventually form AABW, which is exported northward from the gyre. The Weddell Sea Deep Water (WSDW), another water mass formed in the region, has characteristics similar to those of the AABW (Figure 4). It may be generated from complex mixing sequences between surface water masses and Circumpolar Deep Water (CDW; also called WDW within the gyre). WSDW is also formed by uplifting of the densest water mass, Weddell Sea Bottom Water (WSBW, Figure 4), and its mixing with the overlying WDW (further details in section 3).

Geological, hydrographical, and biological processes in the WG are heavily influenced by sea-ice formation, melting, mixing, and advection of water and sea ice, each having effects on ocean-atmosphere-solid Earth interactions. Changes in sea-ice-ocean interaction affecting water masses occur on time scales of decades (Behrendt et al., 2011), making the WG a natural laboratory to understand polar processes and their response to climate variability. The sensitivity of the heat balance in the WG, as well as the potentially global effects of relatively small local changes, makes it a key area for understanding the implications of global heat and freshwater changes.

Beyond its importance for global climate, the WG is unique in many other respects. From a geological perspective, its weak sedimentation regime characterized by very low sedimentation rates (Honjo et al., 2008) stands in sharp contrast to the ACC to the north for reasons that are not yet understood (see section 2). The WG is home to some of the largest populations of large marine mammals, which serves as an indirect metric for the magnitude and distribution of primary production (see sections 8 and 9). Its unique conditions provide habitats to highly specialized Antarctic species, such as seals and Emperor penguins but also blue whales. Arguably, the most conspicuous characteristic of the present Weddell Sea is the abundant and persistent sea-ice cover and its extreme seasonal variability (see section 5), shrinking to a third of its September maximum area by the following April. Sea-ice formation and melting, the presence of ice shelves, and the production of icebergs affect (i) the freshwater balance within the gyre and (ii) its role in the formation and export of deep water to the world's oceans (see section 4) with concomitant sequestration of atmospheric gases, nutrients, and carbon (Brown et al., 2014; Grant et al., 2006; see sections 6 and 7). Sea ice and icebergs,

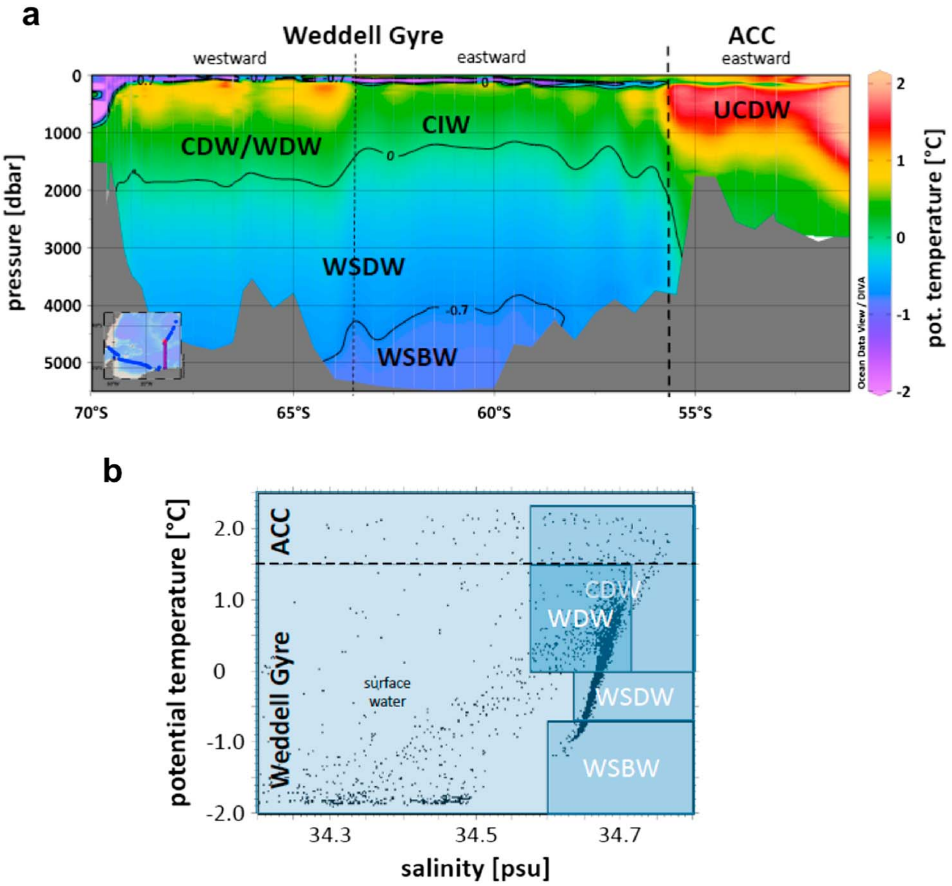

Figure 4. (a) North-south section of potential temperature (°C) across the Weddell Gyre along the Prime Meridian, illustrating characteristic water masses and topography (see section 3 for details). (U)CDW = (Upper)Circumpolar Deep Water; WDW = Warm Deep Water; WSDW = Weddell Sea Deep Water; WSBW = Weddell Sea Bottom Water. The Central Intermediate Water (CIW; see section 4 for details) is a modified form of CDW. Data from Polarstern expedition ANT-XXIV/3 (data from G. Rohardt et al., from the Alfred Wegener Institut, extracted via Webodv). Left inset figure shows the cruise track. (b) The water mass characterization via a potential temperature and salinity diagram.

originating from local or external ice shelves, are important transport means for essential trace elements, which have a potential to respond drastically to climatic changes by surging glaciers and changes in sea-ice distribution. Sea ice is critical in explaining the locations of high biological productivity in the WG, both through sea-ice edge dynamics and through the formation of coastal polynyas (Arrigo et al., 2008; Cape et al., 2014). The close interlinking of physical, biological, geological, and chemical processes makes modeling the WG a particularly promising, albeit challenging, task (see section 10). Furthermore, if heavy sea ice was to remain present in the WG in the future, it would maintain cold polar conditions, making the WG a potential refuge from climate warming for Antarctic organisms (e.g., Emperor penguins) and a prime location to study resilience of organisms to climate change.

While the global relevance of the WG is now appreciated, its functioning and properties remain overall poorly constrained relative to other Southern Ocean regions, as it stands as one of the most remote and inaccessible places on Earth. Data acquisition is particularly difficult in winter, when ships can hardly access it and satellites get almost no information in the visible wavelengths. This results in an alarming lack of observations from one of the key regions shaping our planet, calling for an efficient use of resources, data, and research to advance knowledge at a pace prompted by the observed atmospheric changes. This hurdle can only be overcome by crossing the boundaries of scientific disciplines in a coordinated international effort, as we propose in this overview.

This review emerged from a cross-disciplinary meeting on the WG at the Hanse Wissenschaftskolleg in Delmenhorst, Germany (Hoppema & Geibert, 2012). The event was driven by the realization that our understanding of a key region of the Earth System was limited by disciplinary boundaries and isolated working groups in different countries. By recognizing overlapping interests, sharing knowledge, and identifying overarching challenges, we believe that the research of this unique region can gain the momentum it deserves in light of our rapidly changing planet. The aim of this review is to present the existing challenges in our understanding of this complex system, in regard to both its properties and the mechanistic processes at play.

In recent years, several review papers of the present and future Antarctic climate have been published (e.g., Hellmer et al., 2016; Mayewski et al., 2009; Turner et al., 2013). Against this background of climate change and variability, we introduce challenges for research specific to the WG. To come to a reasonable assessment of the future changes in the different aspects of research in the WG system, we still need to constrain characteristics of the gyre's present-day configuration. Some challenges outlined here will particularly address current conditions, while others address future scenarios. The challenges identified in this review agree with, but expand upon, a recent Scientific Committee on Antarctic Research-Horizon Scan of future research directions in the Southern Ocean (Kennicutt et al., 2015). From the multiple research priorities presented there, many are worthy of further consideration; however, no further elaboration on how to address the questions was provided. Our approach in this review differs in that we include the scientific background available to date to make the connection with the identified challenge.

We propose the main scientific priorities that, if addressed, would significantly advance our knowledge of the climate, geology, physics, oceanography, geochemistry, and biology of the WG. Under increased greenhouse gas forcing, the Southern Hemisphere westerlies are intensifying, likely resulting in enhanced cyclonic wind forcing, increasing westward flow close to the Antarctic continent, and a southerly displacement of the ACC affecting the WG axis and its overall strength (Meijers et al., 2012; Z. Wang, 2013). Constraining processes in the WG using an interdisciplinary approach is therefore of the utmost importance as the region is likely to experience dramatic changes within the 21st century.

2. Geological Evolution and Present-Day Deposition Patterns

The space now occupied by the Weddell Sea opened by the breakup of the Gondwana supercontinent during temperate climate conditions in Jurassic times. This breakup was associated with massive but short-lived volcanic activity. The Weddell Sea subsequently grew larger during the Cretaceous and the Eocene. After the opening of Drake Passage at about 35 million years ago, the Weddell Sea became tectonically passive with approximately the same topography as it has today. A proto-WG probably existed already during this time, when a large ice sheet developed in East Antarctica. Numerous advances of ice sheets in both East and West Antarctica since 35 million years eroded large amounts of rock material on the continent and deposited it on the margins of the Weddell Sea.

The responses and feedbacks of the WG system to global climate changes during the last 35 million years and its glacial-interglacial cyclicity of the last 3 million years are poorly understood. The potential presence of sea-ice free regions even under glacial conditions, which enabled heat and gas exchange between ocean and atmosphere, and glacial-interglacial changes in the intensity of bottom water formation (which controlled ventilation of the world ocean) are likely to have played an important role in the global climate-ocean system.

Today's deposition patterns show mostly downslope transport at the margins of the Weddell Sea. In the central WG, very slow accumulation of mostly lithogenic material is observed, while biogenic remains are largely absent, although there is clear evidence for biological productivity in the surface waters.

2.1. The State of Knowledge

The Weddell Sea has undergone substantial changes through geologic time. The current picture of a sea-ice infested ocean basin, where only a few times in the recent history ships (USCGC *Glacier*, R.V. *Polarstern*, and H.M.S. *Endurance*) were able to enter its southwesternmost rim, that is, the waters offshore from the westernmost Ronne Ice Shelf, is just a snapshot in its geological evolution. During the Early Jurassic, some 180 million years ago, the Weddell Sea did not exist. Instead, the area was located within Gondwana and occupied by landmasses that today form Patagonia and the Falkland Islands (Figure 5). In contrast to the present-day cold climate, a temperate climate with landscapes that we know from present Europe prevailed in the Weddell Sea region during that time. However, during the Jurassic, the Gondwanan supercontinent

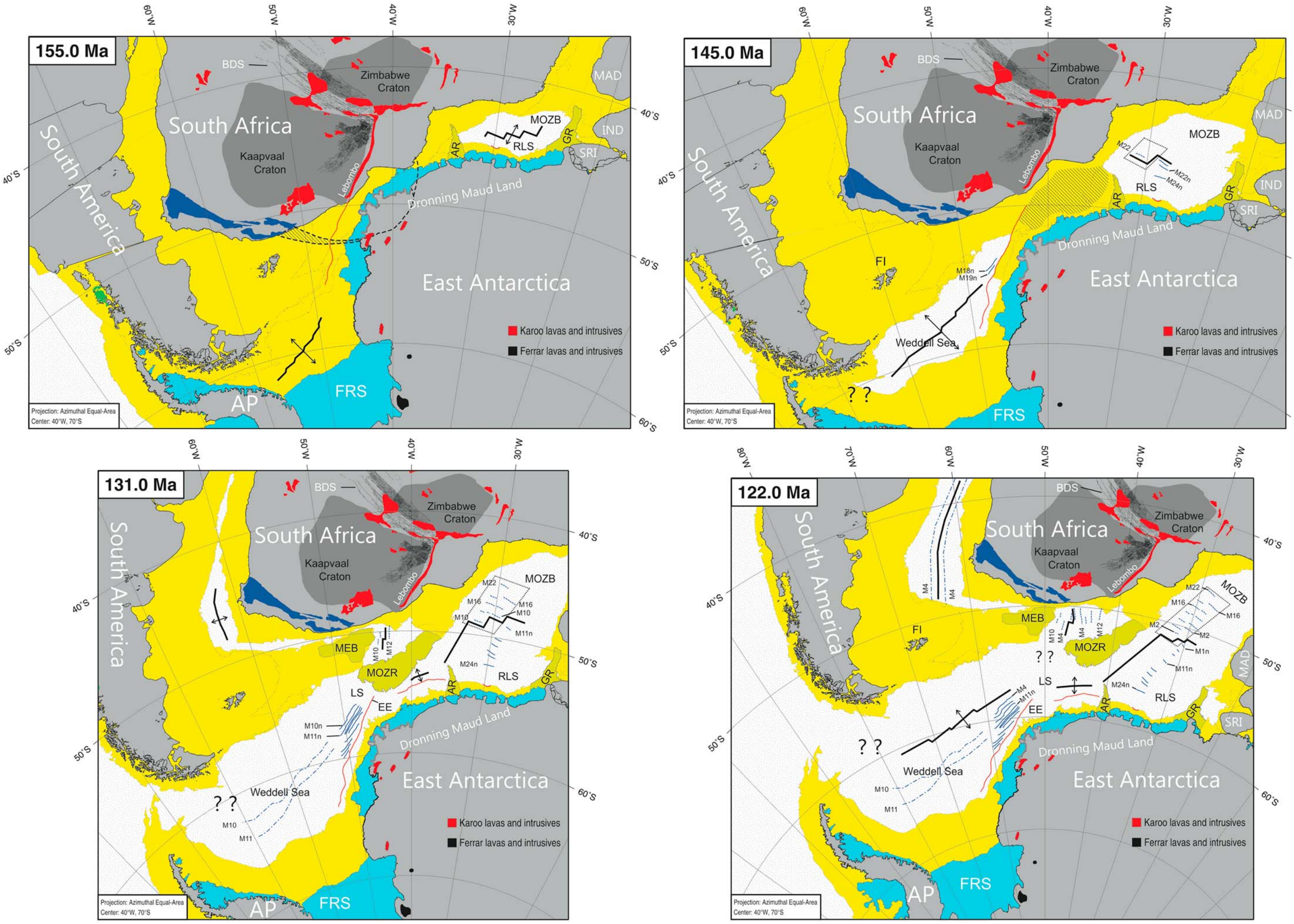

Figure 5. The four panels show the evolution of the Weddell Sea since Jurassic times (modified after Jokat et al., 2003). Dotted white areas: oceanic basins; bold black lines in oceanic basins: mid-ocean ridges; black arrows: spreading directions; dark blue solid and dash-dotted lines: magnetic chrons symmetric to the mid-ocean ridges, which formed by subsequent seafloor spreading; dark blue area: Cape Fold Belt; black dashed lines: magnetic anomalies; thin red lines: outlines of escarpments; yellow areas: extended continental crust; blue areas: present-day ice shelves around Antarctica, dark gray areas: outlines of South African cratons, which form the oldest crust in the region; red and black areas: locations of remnants of volcanic and intrusive rocks formed during the Gondwana breakup. In Antarctica, most of these magmatic rocks are covered by a thick ice sheet.

Abbreviations for all panels: AR = Astrid Ridge; AP: Antarctic Peninsula; BDS = Botswana Dyke Swarm; EE = Explora Escarpment; FI = Falkland Islands; FRS = Filchner-Ronne Ice Shelf; GR = Gunnerus Ridge; IND = India; LS = Lazarev Sea; MAD = Madagascar; MEB = Maurice Ewing Bank; MOZB = Mozambique Basin; MOZR = Mozambique Ridge; RLS = Riiser Larsen Sea; SRI = Sri Lanka.

(top left) Panel 155 Ma: This panel shows the configuration of tectonic plates ca. 10 Ma after breakup had started between Africa and Antarctica around 165 Ma. While oceanic crust was already forming in the MOZB, the Weddell Sea region was underlain by a largely extended, continental basin. The bold black line in the Weddell Sea (representing a possible spreading center) is hypothetical because little is known about the very early evolution of the Weddell Sea. The hatched area indicates a possible connection between magnetic anomalies in South Africa and Dronning Maud Land.

(top right) Panel 145 Ma: Magnetic spreading anomalies in the eastern part of the Weddell Sea put clear age constraints on the evolution of the entire basin. The Weddell Sea opened stepwise from east to west at a rate of 63 km per Million year (Jokat et al., 2003; König & Jokat, 2006). The South Atlantic was an area of largely extended continental basins at this time. The hatched area indicates the location of a shallow sea or a still subaerial region, highlighting the uncertainties in this reconstruction. The question marks in the northern part of the AP indicate that we have no information about the tectonic evolution of Patagonia relative to the AP. (lower left) Panel 131Ma: While the Weddell Sea, the RLS and its conjugate MOZB are already well-developed ocean basins, the South Atlantic starts to open from south to north. The spreading history between the MEB and the MOZR is rather complex. Here the panel shows the current state of knowledge.

(lower right) Panel 122 Ma: The configuration of the Southern Ocean is similar to today, with well-separated continents around it and in its center. Seafloor spreading is now widely active. Again, the kinematic evolution of the northwestern Weddell Sea ("??") is poorly constrained, since most of the crust was subducted at the present-day South Sandwich Trench during the Paleocene to Oligocene. All Mesozoic spreading anomalies on the South American side were destroyed by both this subduction and subsequent formation of the Drake Passage. It is highly unlikely that a connection between the Weddell Sea and the Pacific Ocean already existed at this time (as the panel suggests) because numerous microcontinents populated this gateway around 122 Ma (Eagles & Jokat, 2014).

began to rift apart and break up (Jokat et al., 2003). During its latest rift phase, the young Weddell Sea was a region with massive but short-lived volcanism caused by plate movement (Storey, 1995). During the period from the Cretaceous through the Eocene, while South America and South Africa moved equatorward and Antarctica drifted poleward, the Weddell Sea grew. The oldest sedimentary strata recovered by Ocean Drilling Program (ODP) Leg 113 are of Early Cretaceous age and were drilled in the eastern Weddell Sea (Kennett & Barker, 1990). In this area, rock dredging along the scarps of Wegener Canyon sampled a thick section of Late Jurassic to Early Cretaceous, organic rich sedimentary rocks (Fütterer et al., 1990). In addition, ODP Leg 113 drilled a Miocene-Pleistocene turbidite sequence in the abyssal plain of the central Weddell Sea and a sequence of Eocene to Pleistocene hemipelagites and contourites on the West Antarctic continental margin flanking the southwestern part of the Weddell Sea (Kennett & Barker, 1990). After the Drake Passage had opened between South America and Antarctica about 35 million years ago (Eagles & Jokat, 2014), the Weddell Sea became tectonically passive, with the bathymetry we know today.

Little is known about the sedimentation and oceanography before the opening of the Drake Passage. However, there is some evidence from seismic (Lindeque et al., 2013) and geological data (Carter et al., 2017; Mackensen & Ehrmann, 1992) as well as from modeling results (Douglas et al., 2014; Ladant et al., 2014) that a proto-WG existed already during the Late Eocene-Early Oligocene and potentially even during Cretaceous times. Due to the absence of a large ice sheet during the Cretaceous, the oceanic circulation within the Weddell Sea was controlled by wind, while the sediments deposited within the WG predominantly originated from the discharge of large river systems, which supplied material eroded from the temperate interior of East Antarctica to the ocean. During that time, the WG might not have had the important role for global thermohaline circulation that it has today because the mechanisms that are responsible for the production of precursor water masses for dense AABW today (i.e., supercooling of ocean water under ice shelves and brine rejection in polynyas) were not active under global atmospheric “greenhouse” conditions and in the absence of large ice sheets.

This situation changed with the stepwise opening of the Drake Passage between Patagonia and the northern tip of the Antarctic Peninsula. While Antarctica slowly drifted southward and the Passage widened (Eagles & Jokat, 2014; Lawver & Gahagan, 2003; Livermore et al., 2007; Scher & Martin, 2006) and global atmospheric CO_2 dropped significantly (Beerling & Royer, 2011; Pagani et al., 2011; Y.G. Zhang et al., 2013), the Antarctic continent became oceanographically isolated, and its environment changed to a polar climate (Cristini et al., 2012; DeConto & Pollard, 2003; Escutia et al., 2019; Galeotti et al., 2016; Mackensen & Ehrmann, 1992; Pollard et al., 2013). Little is known about the exact size and dynamics of the Oligocene Antarctic Ice Sheet. There is ample far-field (e.g., deep-sea records of oxygen isotopes in benthic foraminifera shells, global deep-water temperature records, and global sea-level records) and proximal evidence (e.g., deposition of subglacial tills on the Antarctic shelf, Southern Ocean surface water temperature records, and geochemical and mineralogical evidence for a shift from chemical to physical weathering) that a large ice sheet formed in East Antarctica at the Eocene-Oligocene boundary (e.g., Basak & Martin, 2013; Coxall et al., 2005; Ehrmann & Mackensen, 1992; Galeotti et al., 2016; Gulick et al., 2017; Hambrey et al., 1991; Lear et al., 2000; Miller et al., 2005; Zachos et al., 2008). Ephemeral ice sheets had previously existed there during the Late Eocene (Carter et al., 2017; Gulick et al., 2017; Scher et al., 2014, and references therein). The first large, continent-wide, stable ice sheet developed during the Mid-Miocene cooling step (e.g., Lear et al., 2000; Miller et al., 2005; Shevenell et al., 2004). Consequently, the sediments were no longer eroded by rivers but by glaciers, and the oceanography of the entire Weddell Sea changed from that of a temperate to a polar ocean basin since at least the Mid-Miocene (Anderson et al., 2011). The extent of Antarctica's massive erosion is well documented by the thick sedimentary sequences deposited on the continental margin in the Weddell Sea. For example, at its southernmost margin, just in front of today's Filchner-Ronne ice shelf, sediments deposited since the Jurassic are up to 12 km thick (Huang et al., 2014; Jokat & Herter, 2016). Miocene continental shelf progradation, its impact on AABW formation, and the evolution to present sediment transport processes and deposits in the southeastern Weddell Sea were reconstructed by seismic-stratigraphy and modeling (Huang et al., 2017). These data show that the Antarctic ice sheets were quite dynamic since at least the Mid-Miocene.

Several advances/retreats of the ice sheets in the southern Weddell Sea throughout the Quaternary are documented in seismic and acoustic data and in marine sediment cores from the continental margin (Anderson et al., 1991; Diekmann et al., 2003; Elverhøi, 1981; Gales et al., 2014; Gilbert et al., 1998; Grobe & Mackensen, 1992; Grobe et al., 1990; Hillenbrand et al., 2014; Kristoffersen et al., 2000; Kuhn & Weber, 1993; Larter et al., 2012; Mackensen et al., 1989; Melles & Kuhn, 1993; Michels et al., 2002; Ó Cofaigh et al., 2001; Pudsey et al., 1988; Shimmield et al., 1994; J.A. Smith et al., 2010; Weber et al., 2014), with sedimentary sequences from the southeastern Weddell Sea providing archives for decadal and perhaps annual resolution of ice dynamics because of the very high sedimentation rates (Sprenk et al., 2014; Weber et al., 2011). Sediment core and geophysical data indicate that the size of the Antarctic ice sheets in the Weddell Sea sector varied significantly during the last 2 million years, possibly in tandem with the global climate cycles of the Quaternary (e.g., EPICA Community Members, 2004; Lisiecki & Raymo, 2005).

For the Last Glacial Maximum (LGM; ca. 19–23 Kyr BP), the extent of grounded ice-sheet advance across the southern Weddell Sea shelf, and the magnitude of ice-sheet growth in the Weddell Sea drainage sector is not sufficiently understood (Bentley et al., 2010, 2014; Hillenbrand et al., 2014; Larter et al., 2012; Whitehouse et al., 2017). This has significant implications for (a) the estimate of Antarctica's contribution to the eustatic sea-level lowstand of ~130 m below present at the LGM and phases of rapid, major global sea-level rise ("meltwater pulses") during the last deglaciation (Weber et al., 2014) and (b) the production of precursor water masses (High Salinity Shelf Water and Ice Shelf Water [ISW]) for Weddell Sea Deep Water and Weddell Sea Bottom Water, and thus AABW, during the LGM. Grain size data from deep-sea cores indicate that outflow of AABW from the WG into the South Atlantic basin was apparently reduced during glacial times (Diekmann et al., 2003; Gilbert et al., 1998). The same may apply to the current vigor of the WG due to the expanded glacial-time sea-ice cover that shielded the surface ocean from the westerly winds (McCave et al., 2014). However, high salinities reconstructed for glacial bottom waters from the South Atlantic basin suggest that AABW production persisted through glacial periods (e.g., Adkins, 2013; Adkins et al., 2002). While most of the deep Weddell Sea may have been covered by perennial sea ice at the LGM (Collins et al., 2013; Gersonde et al., 2005), geochemical and micropaleontological data from sediment cores revealed that even then (seasonal), polynyas must have persisted within the sea ice covering the Antarctic continental margin in the WG (Grobe et al., 1990; Smith et al., 2010; Weber et al., 2011). Thus, precursor water masses for AABW might have formed within these glacial-time polynyas by brine rejection (Mackensen et al., 1996).

During the present interglacial, deposition in the Weddell Sea is dominated by an interplay between ice rafting and current transport (Diekmann & Kuhn, 1999; Howe et al., 2007; Pudsey & King, 1997). Geographically, the present-day sedimentation pattern in the WG can be split into three geomorphological regions (Jerosch et al., 2016): (i) a southern and eastern region, influenced by downslope transport of sediments near the glaciated continental margin with channel-levee deposits and contourites (Michels et al., 2002); (ii) the central and northern WG, dominated by slowly accumulating sediments of mostly terrigenous origin (Geibert et al., 2005; Howe et al., 2007); and (iii) a northwestern part with contourites and hemipelagites (Gilbert et al., 1998; Pudsey et al., 1988). In addition, Maud Rise in the eastern part of the gyre (Huang & Jokat, 2016) and the Polarstern Bank in the southern Weddell Sea (Bart et al., 1999) stand out from the surrounding abyssal plain with a cover of biogenic sediments (Abelmann et al., 1990; Figure 2).

From the coast, dense water masses and turbidity currents flow down the continental slope through channels and feed into large canyon systems extending across the continental rise (e.g., Amblas & Dowdeswell, 2018; Howe et al., 2007; Michels et al., 2002; Pudsey et al., 1988). Thus, these currents deposit sediments on the continental slope and rise but transport detritus even further down to the Weddell abyssal plain. The downslope transport processes are assumed to be much less active during interglacial times than during glacial periods (e.g., Gales et al., 2014; Lindeque et al., 2013) because of the "bulldozing" effect and subglacial till transport to the shelf break by advancing grounded ice with subsequent downslope redeposition during the latter times (e.g., Grobe & Mackensen, 1992). Sediment distribution is not spatially homogenous, with large fans extending hundreds of kilometers from the shelf break and sediment drifts forming depocenters on the rise (Gilbert et al., 1998; Huang et al., 2014; Kuhn & Weber, 1993; Kuvaas & Kristoffersen, 1991; Michels et al., 2002). Toward the inner WG, deposition dominated by gravitational downslope transport changes gradually to pelagic sedimentation, with sediments being deposited at rates of only a few millimeters per 1,000 years (e.g., Pudsey et al., 1988) and consisting predominantly of lithogenic components (e.g., Howe et al., 2007; Schlüter et al., 1998). This contrasts with sediments on the Antarctic continental margin to the south and southwest of the WG and those underlying the ACC to the north, which both contain higher amounts of biogenic particles and are deposited at higher rates (e.g., Howe et al., 2007; Pudsey &

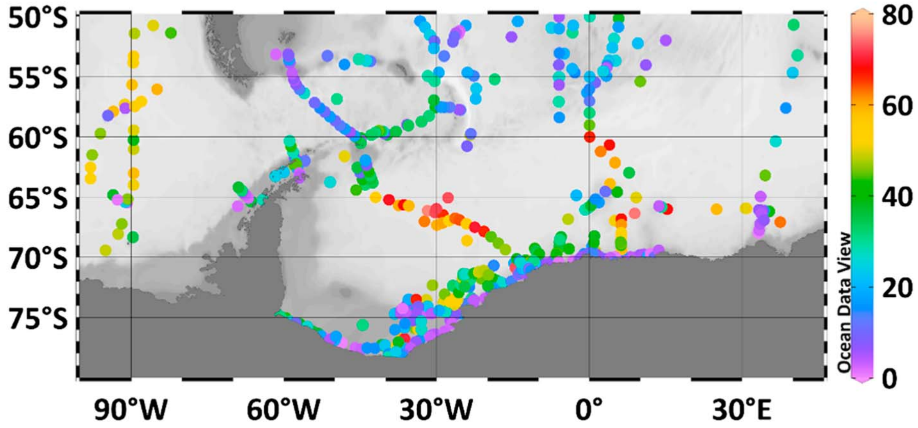

Figure 6. Weight percentage of the clay-sized fraction (<2 μm) in seabed surface sediments (data from Diekmann et al., 2003; Schlitzer, R., Ocean Data View, odv.awi.de, 2018).

King, 1997; Schlüter et al., 1998). The fact that the surface waters overlying the opal-poor sediments of the Weddell abyssal plain are known to be at least occasionally highly productive, with diatom blooms occurring during spring and summer, implies that biosiliceous particles must be affected by strong dissolution in the water column and at the seafloor (Howe et al., 2007; Pudsey & King, 1997; Schlüter et al., 1998; Usbeck et al., 2002). This is confirmed by both the observation of diatom dissolution in seafloor surface sediments (Zielinski et al., 1998) and detailed geochemical studies on these sediments (Schlüter et al., 1998), and it is also reflected in the spatial grain size distribution (Figure 6) which shows a clear dominance of clay-sized particles (<2 μm) in the sediments underlying the central WG (Diekmann et al., 2003; Petschick et al., 1996). In summary, the fluxes of biogenic opal, that is, mainly the remains of diatoms and radiolaria, are high toward the WG boundaries but very low in the center (Geibert et al., 2005).

Toward the north and east of 20°E, a transition to biogenic sediments is seen, but the band of siliceous sediments surrounding Antarctica is clearly a feature of the ACC, not the WG. The outflow of deep-water masses from the WG into the South Atlantic basin is restricted by bathymetric highs in the Scotia Sea (Naveira Garabato et al., 2002). In general, surface currents follow the same flow paths, thereby steering the drift paths of icebergs. Therefore, sedimentary records deposited at locations along these pathways offer unique opportunities for reconstructing past changes in bottom water flow and iceberg discharge (Diekmann et al., 2003; Gilbert et al., 1998; Ó Cofaigh et al., 2001; Weber et al., 2014), hence the International Ocean Discovery Program (IODP) Expedition 382 to "Iceberg Alley" in early 2019 (https://iodp.tamu.edu/scienceops/expeditions/iceberg_alley_paleoceanography.html).

2.2. Research Priorities

1. Despite the progress achieved over the last two decades in reconstructing the timing and processes of gateway opening around Antarctica during the last ca. 45 Ma, questions relating to the development of the WG are still unresolved, because of the lack of scientific drill holes (Escutia et al., 2019). For example, it is unclear (a) how the WG's current system changed in response to the opening of the nearby Drake Passage gateway and thus numerical models simulating these changes (e.g., England et al., 2017; Yang et al., 2014) remain untested, and (b) whether the subsidence of the seafloor in Drake Passage and the neighboring Scotia Sea below a distinct paleo-bathymetric threshold depth was required for the full development of the ACC and the WG.
2. The role of the growing Antarctic ice sheets since the late Paleogene for steepening the atmospheric meridional temperature gradient and thus strengthening the Southern Hemisphere westerlies and initiating the clockwise flowing ACC and the WG is still unclear. Some model experiments suggest that the existence of a large ice sheet on Antarctica could be essential for driving both the ACC and the WG (e.g., Ladant et al., 2014). Other modeling results imply that alternative mechanisms, such as variations in heat supply to the gyre caused by changes in regional thermohaline circulation, could be one of the main driving forces (e.g., Douglas et al., 2014).

3. Other issues that need to be addressed are how the fluctuations in the size and volume of the Antarctic ice sheets during the last 2 million years influenced the WG and how the resulting changes in the WG influenced the global oceanic circulation system and thus global climate. Even for the LGM, it is still unclear whether precursor water masses for AABW could have been produced under ice shelves fringing the southern margin of the WG or whether ice grounded on the entire continental shelf prevented the production of those water masses (Hillenbrand et al., 2014), and precursor-water masses for AABW formed in polynyas over the continental slope instead (Mackensen et al., 1996).
4. Another important open question addresses our understanding of biogeochemical processes: Why are sediments rich in biogenic opal absent from the seafloor underlying the central WG despite significant diatom productivity taking place in the surface waters?

3. Physical Oceanography

Alongside the atmosphere, the ocean plays a key role in the climate system in redistributing excess heat from the Equator toward the polar regions. The key ways in which this latitudinal transport of heat is achieved are through the system of vertically separated ocean currents known as the global overturning circulation and through a series of wind-driven, large-scale horizontal circulations known as ocean gyres. The Southern Ocean is a particularly important region for the transport of ocean heat, nutrients, and carbon as in this region, these climatically important quantities are exchanged between the main Atlantic, Pacific, and Indian Ocean basins within the ACC. The Southern Ocean is also the location where deep ocean waters, which have been isolated from the atmosphere for hundreds to thousands of years, are returned to the surface and can exchange their properties with the atmosphere. Moreover, certain regions around the Antarctic continent itself (the WG being a prime example) are key locations for the formation and modification of the densest ocean waters. These waters, made dense through rapid surface cooling and rejection of salt as sea ice is formed, subsequently fill the deepest parts of the Southern Hemisphere's ocean abyss. The complexity of the physical processes occurring within the WG region, our lack of knowledge, particularly of the wintertime physics, and the subsequent impact of the newly formed waters on the global ocean circulation mean that addressing the gaps in understanding in this region is a global climate priority.

3.1. The State of Knowledge

While the WG might appear externally as a relatively hermetic bowl with few isolated, well-defined inflows and outflows (e.g., Orsi et al., 1993), it is dynamically complex and highly variable, influenced by physical processes spanning several orders of magnitude in spatial and temporal scales.

At low frequencies, climate modes have been shown to influence both the strength of the gyre and Weddell Sea sea-ice extent. Martinson and Iannuzzi (2003) observed a spin-up in WG during El Niño conditions, with a weakening occurring during La Niña events. Furthermore, they found a strong dipole in sea-ice extent, with positive anomalies in the WG occurring contemporaneously with reduced Amundsen/Bellinghausen sea ice (and vice versa). More recently, Armitage et al. (2018) investigated the seasonal variability in the gyre strength from sea surface height estimates and concluded that the month-to-month circulation of the WG is strongly controlled by local wind stress curl ($r = -0.67$ between gyre strength and wind stress curl). Furthermore, both modeling (Mathiot et al., 2011) and observations (Núñez-Riboni & Fahrbach, 2009) support the observation that the seasonal mass transport of the Antarctic Slope Front (ASF), which runs along the southern side of the WG, is controlled largely by the strength of the easterly winds. The properties of the AABW, the dense water mass which forms on the continental shelves of the Weddell Sea before being exported to the South Atlantic, are also sensitive to change in wind forcing over the gyre. Jullion et al. (2010), using 15 hydrographic sections, argued that wind stress variability over the WG forced changes in AABW properties in the SR1b hydrographic section to the east of Drake Passage, with a lag time of around 5 months, which they ascribed to a baroclinic adjustment of the gyre. This forcing has effects outside the Weddell Sea, with Meijers et al. (2018) arguing that the intermittent presence of a cold, fresh slope current north of Elephant Island is attributable to variable export from the Weddell Sea, driven in turn by wind-driven acceleration of the boundary current. Kerr et al. (2018) have also argued for a long-term decline in the densest constituent of this bottom water, the WSBW, though some recovery in volume has occurred since 2005.

In contrast to these large-scale controls, there is also clear evidence that highly variable small-scale turbulent mixing has a key role in setting the thermohaline properties of the AABW that leaves the Weddell Sea through deep outflows to the north. Polzin et al. (2014), for example, have argued that elevated boundary mixing just north of Orkney Passage is key to transforming WSDW on entering the Scotia Sea. Understanding such small-scale processes is vital to interpreting long-term changes in AABW properties that have been observed in the South Atlantic, including the both warming and reduction in volume (Meredith et al., 2008; Purkey & Johnson, 2010). Furthermore, it has been argued that the occasional but periodic opening of the Weddell Polynya (Carsey, 1980), an area of open water of up to 3 × 10^5 km^2, which appeared in winters of 1974–1976 and again in 2016 and 2017, can short-circuit the usual deep-water formation processes by rapid open ocean buoyancy loss and deep convection. However, recent literature (e.g., Dufour et al., 2017) has emphasized the importance both of an interior heat reservoir and of mesoscale eddy activity on triggering polynya development. It has also been suggested that a prolonged negative phase of the Southern Annular Mode (SAM) might be implicated in allowing the polynya to open (Gordon et al., 2007).

Away from the densest waters, mixing processes are much less well understood. Muench et al. (1990) have observed large (up to 100 m) thermohaline staircases in the northwestern Weddell Sea, separating the warm CDWs from the overlying cooler and fresher layers, suggesting that these staircases were a key mechanism for vertical heat transport within the thermocline. While it appears that overall mixing rates in the interior waters are likely to be weak, the lack of observations of turbulent mixing processes in the Weddell Sea remains a key limitation on our knowledge of the gyre interior.

In general terms, the Weddell Sea is characterized by a strong coupling between the WG (transporting about 50 Sv [1 Sv = 1 × 10^6 m^3/s] cyclonically; Klatt et al., 2005; Q. Wang et al., 2012) and an intense overturning circulation, both forced by westerly and easterly winds over its northern and southern edges, respectively, as well as by buoyancy fluxes. Relatively warm and salty CDW (the voluminous midlayer water of the ACC, with its origins in the North Atlantic; Callahan, 1972) enters the gyre at its eastern edge and while concurrently rising (Donnelly et al., 2017) flows toward the southwestern Weddell Sea. This is illustrated in Figure 7, which shows the temperature of the subsurface temperature maximum from objectively mapped Argo float data (modified from Reeve et al., 2016). Here it intrudes onto the continental shelves as modified WDW and mixes with cold shelf waters that are made more saline by brine rejection during sea-ice formation. The intense air-sea-ice interaction that precedes brine rejection occurs on the shelf in front of the extensive ice shelves—the Filchner-Ronne (Nicholls et al., 2009) and Larsen (van Caspel, Absy, et al., 2015; van Caspel, Schröder, et al., 2015; Huhn et al., 2008)—and the resultant mixture of WDW and dense shelf waters leads to dense water formation. As this water mass descends, it entrains midlayer waters, before being exported northward and westward (Kilworth, 1974; Wilchinsky & Feltham, 2009). Ultimately, only the shallower portion of this water—the WSDW—is sufficiently light to overflow into the Scotia Sea. The WG has traditionally been considered as the main source of AABW production (Orsi et al., 1999). However, recent work has challenged this view, arguing that up to 50% of the AABW exported across the northern rim of the gyre is imported into the gyre from the Indian sector of the Southern Ocean (Jullion et al., 2014; Naveira Garabato et al., 2014). It has been proposed that one of these pathways is dominantly advective, while the other is dominantly eddy driven (Ryan et al., 2016).

3.2. Research Priorities

Despite recent oceanographic advances in our understanding of the WG, several pressing issues remain unresolved. These are discussed in the following subsections.

3.2.1. Connection With the Rest of the Southern Ocean

As stated previously, the eastern boundary of the WG remains poorly resolved with no well-defined boundary current marking its eastern periphery. The traditional defined limit of the WG is located around 30°E but can extend as far as the Kerguelen Plateau at ~70°E (Y.H. Park et al., 2001). Two main routes for CDW into the WG have been identified (Cisewski et al., 2011; Fahrbach et al., 2011; Núñez-Riboni & Fahrbach, 2009; Ryan et al., 2016).

First, there is a direct, eddy-driven route in the northeastern corner of the WG carrying warm and salty CDW into the gyre (Jullion et al., 2014; Leach et al., 2011; Ryan et al., 2016; Schröder & Fahrbach, 1999). This is thought to be generated by a topographic discontinuity in the southwest Indian Ridge (55°S, 25°E) as

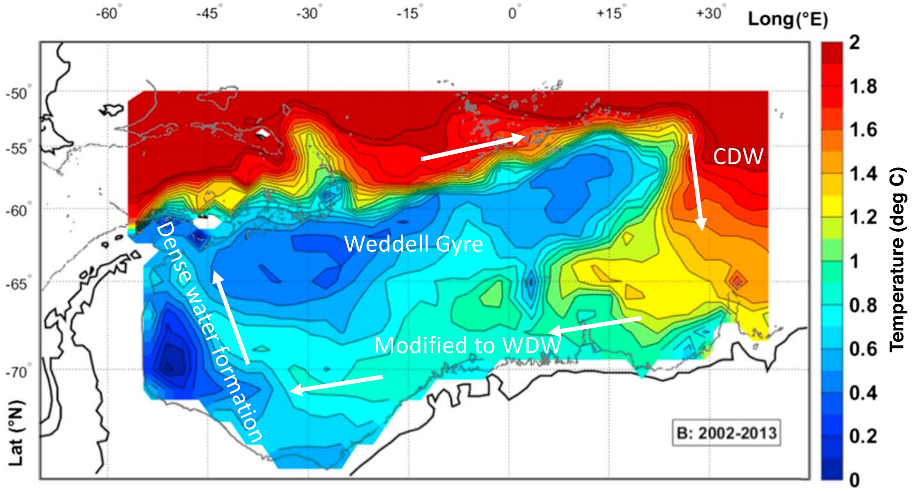

Figure 7. The Weddell Gyre: temperature (°C) at the subsurface temperature maximum, derived from optimally interpolated Argo float data (from Reeve et al., 2016; http://doi.org/10.17882/42182). The figure shows clearly the penetration of Circumpolar Deep Water (CDW) from the east on the southern side of the gyre, and the rapid cooling of the middepth waters on the western side of the Weddell Sea as Antarctic Bottom Waters are formed. The arrows show the dominant middepth flow direction of the gyre, and WDW is Warm Deep Water. The penetration of CDW from the east can clearly be seen, along with its cooling as it advects eastward along the southern edge of the Weddell Gyre.

suggested by elevated levels of mesoscale eddy kinetic energy just downstream of this trough and by idealized numerical simulations (Witter & Chelton, 1998).

A second advective pathway into the WG is via the ASF, which carries CDW (Cisewski et al., 2011) from the far east but also, more surprisingly, appreciable quantities of recently formed AABW (Archambeau et al., 1998; Hoppema et al., 2001; Meredith et al., 2000). This AABW, most likely originating from intense sea-ice formation within the Cape Darnley Polynya near 70°E (Ohshima et al., 2013), contributes significantly to the gyre's overall AABW inventory (Jullion et al., 2014; Naveira Garabato et al., 2014) and exhibits thermohaline variability that is not yet fully understood (Couldrey et al., 2013). While we now have some knowledge of the climatology of the eastern WG, there remain great uncertainties concerning its variability (both in terms of transport as well as water mass properties) and the relative contribution from eddies.

3.2.2. Changes in Air-Sea-Ice Interactions

The WG is subject to strong and complex thermodynamical forcing, including interactions with ice shelves (Nicholls et al., 2009) and a strong seasonal cycle of sea-ice extent (Zwally et al., 2002), as well as precipitation (Bromwich et al., 2011) and intense transfers of heat across the sea surface. Together, these lead to significant cooling and freshening of the inflowing CDW during the formation of new water masses (Jullion et al., 2014). The changes in buoyancy driving these water mass transformations are set by a delicate interplay between the different sources and sinks. In the center of the gyre, buoyancy gain occurs via sea-ice melt and the addition of meteoric water; these processes have a key role in driving the upper overturning cell. In addition, buoyancy loss also occurs, through sea-ice formation and large ocean-to-atmosphere heat fluxes near the southern and western margins of the gyre. The cancellation of these two terms leads to a small residual buoyancy gain in the gyre, but both contributions significantly contribute to both limbs of the overturning circulation of the Southern Ocean (Naveira Garabato et al., 2016). However, strong seasonality in both the gyre and the buoyancy forcing and the scarcity of wintertime observations greatly limit our ability to describe processes at play in detail and thus to predict future responses under climate change scenarios. The observed changes in sea-ice extent (Hobbs et al., 2016; P. R. Holland & Kwok, 2012) and glacial runoff (P. R. Holland et al., 2015) as well as the potential sensitivity to changes in the buoyancy forcing of both overturning cells highlight the urgent need to better understand the physical processes underpinning the buoyancy budget and their variability on monthly to decadal time scales.

3.2.3. Cross-Shelf Processes

In the WG, the ASF plays a central role in mediating cross-shelf exchanges of mass and heat and therefore directly influences processes such as ice-shelf stability, sea-ice production, and dense water formation. Despite its importance, little is known about the slope front variability and its dynamical drivers. Idealized numerical simulations show that export of bottom water from the shelf toward the deep ocean is sensitive to changes in easterly wind stress near the Antarctic coast (Stewart & Thompson, 2012). Novel glider observations (Azaneu et al., 2017) have also suggested that the passage of eddies and changes in bottom slope may also be responsible for high-frequency fluctuations in current strength. Hellmer et al. (2012) suggested that global-warming-related reduction of sea-ice cover in the southeastern Weddell Sea changes the oceanic surface stress, which may lead to an advection of warm water from the open ocean toward the Filchner-Ronne ice shelves causing a dramatic increase in basal melting by 2100.

Scarce observations in the Weddell Sea as well as idealized numerical modeling hint that mesoscale eddies contribute significantly to the cross-shelf transport of warm water onto the shelf. The current state of understanding is that both tides and eddies are significant in effecting cross-shelf exchange (Nøst et al., 2011; A.F. Thompson et al., 2014; Stewart et al., 2018; see also section 3.1), with onshore eddy and Ekman-driven transports being balanced by offshore near-bottom flows. However, most of this understanding comes from numerical simulations (both realistic and idealized). Further targeted process studies including high-resolution hydrography, velocity, and mixing measurements will be required to provide fundamentally improved understanding of the physical drivers underpinning the ASF dynamics and associated cross-shelf exchanges.

3.2.4. Water Mass Variability and Export

Hydrographic data reveal that the water mass properties associated with the WG overturning display interannual to decadal variability. Winter Water and WDW (the Weddell Sea variant of CDW) are subject to significant multiannual variability (Behrendt et al., 2011; Fahrbach et al., 2011), possibly forced by changes in the inflowing CDW (Cisewski et al., 2011). Moreover, the export of Central Intermediate Water (CIW) is thought to be strongly modulated by wind-driven acceleration of the boundary current of the gyre (Meijers et al., 2017). The observed variability of AABW is spatially strongly structured. Within the inner WG, AABW is dominated by multiannual variability in thermohaline properties (Fahrbach et al., 2011), though chlorofluorocarbon and sulfur hexafluoride (SF_6) data suggest a decrease of the AABW ventilation and related reduction of anthropogenic CO_2 uptake (Huhn et al., 2013). The cause (internal variability or an externally forced trend) and origin of this process (local or remote) remain elusive. In the eastern WG, AABW originating from Cape Darnley appears to have warmed during the 1990s and 2000s (Couldrey et al., 2013), and the AABW leaving the Weddell Sea has been consistently freshening since the 1990s (Hellmer et al., 2011; Jullion et al., 2013), but there also appears to be significant interannual variability forced by local adjustments in the circulation (Jullion et al., 2010; Meredith et al., 2008, 2011). The narrow gaps that dissect the South Scotia Ridge (the Orkney Passage in particular) through which bottom water is exported (Franco et al., 2007; Jullion et al., 2014; Naveira Garabato et al., 2002) actively contribute to the water mass modification via complex dynamical processes in bottom-intensified jets (Polzin et al., 2014). Over longer time scales, coupled general circulation models predict a rapid reduction in open-ocean convection in the WG over the coming century in response to greenhouse gas forcing (de Lavergne et al., 2014), though most coupled climate models do not accurately represent the scale and rate of convective processes occurring in the Weddell Sea (Heuzé et al., 2013).

Overall, our picture of the physical variability and time-varying flows in the WG has evolved greatly in recent years, and its role in large-scale circulation and climate has become much clearer. However, the gyre is a highly dynamic regime where small-scale processes play a significant role in the modification of exported water masses and where extant data sets are insufficient to generate the understanding required for accurate reanalysis, simulations, and predictions. Further targeted field campaigns, building on modern technology such as autonomous vehicles with ice capabilities, are needed to address each of the issues identified above. Following this, these new insights can be used to improve representations of key processes in numerical models. The recent application of radar altimetry from leads within the ice-covered ocean (Armitage et al., 2018; Bulczak et al., 2015; Peacock & Laxon, 2004) has, for the first time, allowed the circulation of the WG to be measured year-round, though the processes controlling, for example, the strong seasonal variability in the Slope Current in the WG sector, remain to be fully understood.

4. Cryosphere I: Interactions Between the Ocean and the Ice Shelves

The WG interacts with all components of the cryosphere, that is, sea ice, icebergs, and ice shelves. Sea-ice formation extracts freshwater from oceanic surface water causing the latter's densification due to brine release. On the broad southern continental shelf, dense water accumulates and fuels basal melting of the ice shelves, also contributing to the formation of deep and bottom waters. Wind and ocean currents in the southern Weddell Sea transport sea ice northward to melt and return freshwater to the central gyre during summer. A large hole—the Weddell Polynya—can sometimes appear near the seamount Maud Rise, where ocean heat melts sea ice also during winter.

Icebergs are driven by winds, ocean currents, and sea-ice drift in dense pack ice, following the general course of the WG until leaving to melt in warmer waters to the north. They provide freshwater to the ocean and carry iron-rich continental dust and debris, which might initiate phytoplankton blooms.

Melting of thick ice shelves is cooling and freshening the ocean along the coast. Even cold shelf water is efficiently melting the colder deep ice shelf bases. Fresher and hence more buoyant meltwater ascends along the ice base until the pressure release initiates the formation of ice crystals (combined with brine rejection), which attach to the ice base forming a body of marine ice. The water mass with temperatures below the surface freezing point produced in this way is called ISW and contributes to the formation of very dense Weddell Sea Bottom Water.

4.1. The State of Knowledge

Snow accumulation on the Antarctic ice sheet (up to 4,000 m thick) is balanced by the transfer of ice to the ocean, either by iceberg calving at ice shelf fronts or by melting of ice shelf bases (Jacobs et al., 1992). Recent studies, based on remote sensing, have revealed that both transfer mechanisms are of equal importance, and the ice shelf basal mass loss in the Weddell Sea was determined to be 118 ± 52 Gt/year (Depoorter et al., 2013), a figure that is largely controlled by the amount of oceanic heat available for melting. The decrease of the temperature maximum of the WDW from 1.2 °C at the Prime Meridian (Klatt et al., 2005) to 0.6 °C at the tip of the Antarctic Peninsula (Schröder et al., 2002) indicates significant exchange with the cold shelf waters. However, a well-defined ASF separates the warm waters from the coast.

Ocean-ice shelf interaction in the Weddell Sea can be separated into three characteristic regional regimes that are dominated by two different principal circulation modes on the continental shelf (Figure 8). In the east (regime I, approximately from 30° E to 30° W, Figure 2), a band of smaller ice shelves fringes the narrow continental shelf (Nicholls et al., 2009) and is in close proximity to the ASF, which is manifested by a strong halocline, deepening toward the shelf break (Figure 8a). Strong ocean currents increase basal melting where the ice overhangs the continental slope (Langley et al., 2014; Price et al., 2008), and basal melting driven by onshore transport of open ocean heat is directly linked to the frontal dynamics (Heywood et al., 1998). Recent observations and high-resolution modeling indicate the role of mesoscale eddies formed by ASF instabilities in transporting warm water onto the continental shelf and into the ice shelf cavities (Nøst et al., 2011; Stewart & Thompson, 2015). These eddies are also important in regulating the thermocline depth by balancing the wind-driven coastal downwelling (Sverdrup, 1954). The latter responds to surface wind stress and near-surface hydrographic conditions (Hattermann, 2018; Hattermann et al., 2014; Smedsrud et al., 2006) and interacts with bottom corrugations and troughs crosscutting the shelf break, to determine the inflow of warm water toward the ice shelf grounding lines (Hattermann et al., 2012; Nicholls et al., 2006). During summer, the prevailing easterly winds accumulate fresh surface water from sea-ice melt along the coast, where its admixture to the coastal Winter Water produces Low Salinity Shelf Water (Zhou et al., 2014), while downwelling of this solar-heated surface water seasonally increases melting beneath the ice shelves (Hattermann et al., 2012).

The Filchner-Ronne Ice Shelf, the largest floating volume of ice on the planet at $3{\cdot}10^5$ km^3, located in the southern embayment of the Weddell Sea (regime II, 30° W to 60° W, south of 70° S; Figure 2), is separated from WDW by a wide continental shelf. The circulation in the ice shelf cavity is driven by tides and density currents associated with High Salinity Shelf Water (Figure 7b) formed by intense sea-ice production in polynyas and leads near the ice shelf front (Haid et al., 2015; Tamura et al., 2008). The freshwater extraction due to sea-ice formation causes salinification, and thereby densification of the shelf water masses, ultimately contributing to the formation of deep and bottom waters at the continental slope and giving rise to a

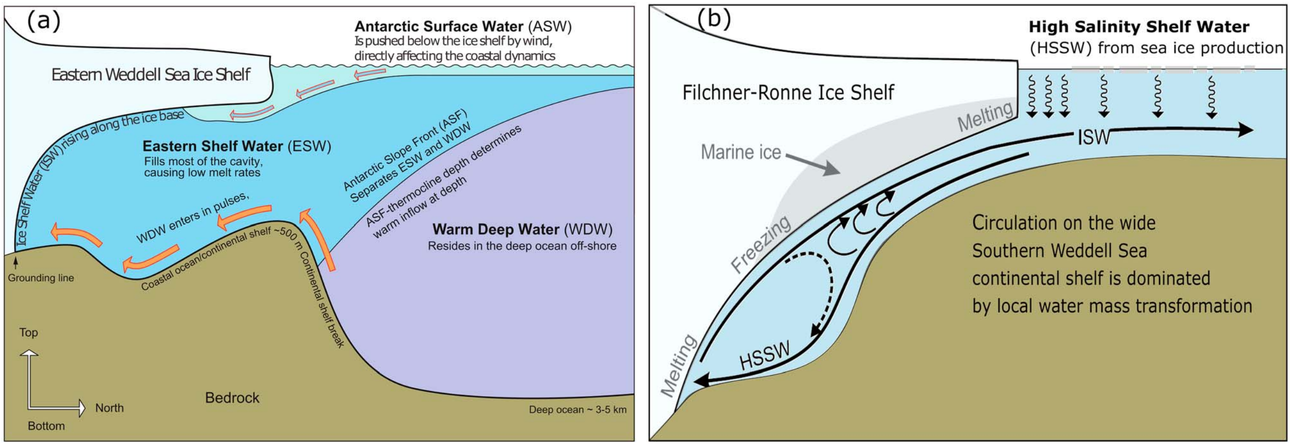

Figure 8. Schematic of the hydrographic conditions (a) along the continental slope of the eastern Weddell Sea (Hattermann et al., 2014, Open Access) and (b) on the wide continental shelf in the southern Weddell Sea (adapted from Nicholls et al., 2009).

V-shaped halocline that guides Low Salinity Shelf Water along the shelf break (Gill, 1973). High Salinity Shelf Water at the surface freezing point provides the oceanic heat for area-mean basal melt rates of a few decimeters per year (several meters per year locally along the deep grounding line), because the freezing point is reduced at higher pressure (Jacobs et al., 1992). The resulting water mass, ISW, partially refreezes at intermediate depths beneath the ice shelf as part of the "ice-pump" mechanism (Lewis & Perkin, 1986) and contributes to the formation of Weddell Sea Deep and Bottom Water (Nicholls et al., 2009). Traces of modified WDW have been observed to enter the continental shelf (Nicholls et al., 2008), mainly through topographic depressions such as the Filchner Trough, which connects the Filchner-Ronne Ice Shelf cavity with the open ocean (Darelius et al., 2016). The dynamics at the Filchner sill, which determine the inflow of WDW into the trough, involve a complex interplay between the topographically steered current and eddy-shedding of topographic shelf waves at the mouth of the trough (Daae et al., 2017). The topographically steered current cannot directly cross the sill: vortex stretching of an inflowing current due to the southward deepening of the trough imposes cyclonic relative vorticity, hence, deflecting the current out of the trough. However, model projections indicate that an irreversible regime shift might occur at the sill of the Filchner Trough as a consequence of a warmer climate during the 21st century. Warm waters, transported with the slope current, will reach the deep grounding line of the Filchner-Ronne Ice Shelf (Hellmer et al., 2012). In the model, the increased inflow of warm water was triggered by shoaling of the Slope Front thermocline at the shelf break (Figure 7a) and freshening of shelf waters due to reduced sea-ice formation. Increased basal melting in this warm state drives further shelf freshening, a positive feedback leading to irreversible WDW inflow independent of the applied climate forcing (Hellmer et al., 2017). The additional heat supply used for basal melting has an impact on the ice-shelf thickness distribution and, thus, ice flow velocities of the ice streams draining the Antarctic ice sheet (Timmermann & Goeller, 2017). The changing cavity geometry might also significantly influence the tidally induced sub-ice shelf heat transport (Mueller et al., 2018). However, large uncertainties remain around these simulations as the occurrence of such a regime shift in the southern Weddell Sea depends on the choice of the atmospheric scenario input (Timmermann & Hellmer, 2013). Other ice shelf-ocean model projections suggest that melt rates are likely to increase more in the eastern Weddell Sea (Kusahara & Hasumi, 2013).

At the Filchner sill, a well-documented overflow of ISW (Daae et al., 2019; Foldvik et al., 2004) exists. This flow, controlled by gravity and bottom topography, underrides the ASF and is a well-known source for Weddell Sea Bottom Water. Alternatively, the ISW may follow the continental slope toward the west, contributing to the formation of Weddell Sea Deep Water (Figure 3).

Along the western perimeter (regime III, west of 50° W, north of 70° S), the WG passes the ice shelves of the Antarctic Peninsula, where observed changes are mostly found to be associated with atmospheric processes (e.g., Scambos et al., 2000). Here the continental shelf circulation combines elements of both principal modes depicted in Figure 7. Due to the perennial sea-ice cover in the western Weddell Sea, observations across the ASF are rare, but those few all show the leakage of modified WDW onto the continental shelf beneath a weak V-shaped Slope Front (Absy et al., 2008; Gordon, 1998). These measurements also indicate a rapid decrease of the temperature toward the coast with no deep troughs allowing WDW derivatives to access the Larsen Ice Shelf front (Nicholls et al., 2004). Thus, the ocean's contribution to the decay of parts of the Larsen Ice Shelf is assumed to have been minor. However, observations of two drift experiments (Gordon et al., 1993; Hellmer et al., 2008) revealed the western continental slope to be covered with a thin layer of cold water, which contains a significant amount of glacial melt in its upper part, underlain by High Salinity Shelf Water (Huhn et al., 2008). The sources are located in or near the southwestern Weddell Sea (Figure 3). Observations near the tip of the Antarctic Peninsula (Schröder et al., 2002) indicate that the "nearby" contribution must happen in pulses, but the cause of the intermittent flow remains speculative.

4.2. Research Priorities

4.2.1. Dynamics of the ASF

The dynamics, temporal and spatial variability, and environmental sensitivity of the ASF and the associated slope current are still poorly understood. There is evidence that the water masses of the slope current originate at the southeastern margin of the Weddell-Enderby Basin (Figure 2; Hoppema et al., 2001; Ohshima et al., 2013). Here differential mixing with southward-flowing ACC waters as part of the eastern rim of the WG (Figure 3) might be responsible for the observed temperature variability in the deep (Ryan et al., 2016) and abyssal (Couldrey et al., 2013) waters of the Weddell Sea.

4.2.2. Monitoring Wintertime Conditions

Animal platforms such as instrumented seals (http://www.meop.net) have provided unique insights over the recent years into seasonal variability and processes (Årthun et al., 2012; Fedak, 2013). However, observations of wintertime conditions along the Weddell Sea coast are still rare, and our mechanistic understanding of the seasonal (as well as climate change-related) availability of warm water for glacial ice shelf basal melting (and its implications for the response to future changes) remains limited (Hattermann et al., 2012; Semper & Darelius, 2017). Even during the austral summer season, sea-ice conditions in the Weddell Sea do not allow for unimpeded ship operations. This calls for the design of new autonomous technologies, capable of reliably operating in heavy sea-ice conditions and underneath an ice shelf. A promising step forward is the development and use of under-ice Argo floats, which track the circulation of water masses in the whole WG, including its largely sea-ice-covered southern and western part (Reeve et al., 2016; Figure 7).

4.2.3. Numerical Modeling

Numerical simulations are challenged by (a) the need for high-resolution models that capture the Slope Front's dynamics (Nakayama et al., 2014; St-Laurent et al., 2013; Stewart & Thompson, 2015) and (b) requirements for accurate boundary conditions for the slope current that follows most of the east Antarctic coast. Accurate modeling also requires a realistic representation of synoptic atmospheric forcing, sea-ice processes, and the interaction with the ocean, for which very little data for validation exists. Finally, understanding and simulating ocean-ice shelf interaction in the Weddell Sea requires a better knowledge of the sub-ice shelf bathymetry, which is presently poorly constrained or entirely unknown beneath most of the ice shelves fringing this part of Antarctica's coast.

5. Cryosphere II: Sea Ice

Sea ice forms by the freezing of seawater that is exposed to cold air and covers the polar oceans as a 0.5- to 3-m-thick layer of ice. The areal extent and thickness of sea ice are important climate indicators. Sea-ice extent and thickness have strongly decreased in the Arctic over the past 40 years. In contrast, ice extent has slightly increased in the Southern Ocean during the same time period. There is no reliable information about changes of the thickness of Antarctic sea ice. Interpretation of sea-ice variability and change is complicated by the fact that sea ice moves under the forces of winds and currents. Therefore, the presence of sea ice and the location of the ice edge in a certain region can be more indicative of the drift of sea ice than of regional air temperatures or ocean heat fluxes. Similarly, the thickness of sea ice is strongly affected by rafting and ridging in regions of convergent ice drift, and ice thicknesses in a certain region may not be good indicators of local thermal conditions. Sea ice is closely interlinked with the air above and water below and plays important roles for the polar climate system due to its high albedo, stored latent heat, accumulation of snow, dense brine production upon formation, and freshwater release upon melting. Sea ice is also home to large standing stocks of marine microorganisms which supply food for higher trophic levels and are the base of the Antarctic marine food web.

5.1. The State of Knowledge

Sea ice in the WG is characterized by a mean, large-scale cyclonic circulation, which is a result of the mean atmospheric geostrophic forcing (e.g., Kottmeier et al., 1992). It is commonly believed that the wind stress associated with this circulation is a major driver of the surface ocean currents in the region (e.g., Youngs et al., 2015). The resulting surface currents cause a mean inflow of ice in the east and a mean outflow of ice in the northwest. Within the drift, sea ice survives the summer melt, eventually forming second-year ice, of which the Weddell Sea possesses the most of all Antarctic seas. The WG drift, together with the large seasonality of the ice cover with vast regions showing rapid ice growth in fall typically by way of pancake ice formation (Lange et al., 1989), leads to the predominance of seasonal ice in the eastern Weddell Sea with maximum, end-of-winter thicknesses of between 0.8 m in the north near the marginal ice zone and 1.5 m near the coast of Dronning Maud Land (e.g., Harms et al., 2001). The second-year ice farther to the southwest and west can be much thicker with mean thicknesses of more than 3 m (Haas et al., 2008; Harms et al., 2001) and is some of the thickest sea ice of the Southern Ocean. Accordingly, the western Weddell Sea is one of the regions with the largest northward sea-ice-derived freshwater fluxes in the Southern Ocean (e.g., Haumann et al., 2016). The mean drift also leads to the widespread occurrence of polynyas, in particular along the coast of Queen Maud Land and the Filchner Ronne Ice Shelf (e.g., Paul et al., 2015), which are key regions for High Salinity Shelf Water and AABW formation. New ice formed in the Ronne polynya often advects northward between bands of old ice, reaching thicknesses of up to 2 m after 7 months of ice drift owing to much lower oceanic surface heat fluxes in the western than in the eastern Weddell Sea (Haas et al., 2008).

Snow plays an important role for the sea-ice mass balance, and its thickness varies with region, season, and ice age. In winter, mean snow thicknesses on first-year ice range between 0.15 and 0.25 m (Arndt & Paul, 2018; Eicken et al., 1994 ; Massom et al., 2001). On second-year and older ice, mean snow thicknesses range between 0.5 to more than 1 m (Arndt & Paul, 2018; Haas et al., 2008; Kwok & Maksym, 2014). On thinner ice, the presence of a thick snow cover often leads to negative ice freeboard, flooding, and subsequent snow ice formation (Eicken et al., 1994; Massom et al., 2001). However, Eicken et al. (1994) estimated that snow ice contributes about 4% to total ice thickness in the Weddell Sea, which is less than in other Antarctic regions with thinner ice. In summer, thaw-refreeze metamorphism dominates snow processes and decay, with the formation of superimposed ice but an absence of melt ponding (Haas et al., 2001; Nicolaus et al., 2009). The latter is due to the fact that Antarctic sea ice with a generally colder, dryer atmosphere is dominated by different components of the surface energy balance than Arctic sea ice (Andreas & Ackley, 1982; Nicolaus et al., 2006).

In contrast to the rapid decline of Arctic sea-ice extent during all seasons since 1979, both the summer and winter Antarctic sea-ice extent has increased during the same time period by 1.6% and 0.6% per decade, respectively (Figure 9; e.g., Parkinson & Cavalieri, 2012; Simmonds, 2015). However, the influence of interannual variability is large, and there are extensive regional differences (Polvani & Smith, 2013; Simpkins et al., 2013; Swart & Fyfe, 2013; Zunz et al., 2013). This may explain why, in the Weddell Sea, the winter sea-ice extent has retreated at a moderate rate (−0.8% per decade), while the summer sea-ice extent has rapidly expanded (10.7% per decade; Figure 9; Hobbs et al., 2016; Turner et al., 2015). Changes to the sea-ice extent in the Ross, Amundsen, and Bellingshausen Seas have been linked to variations of stratospheric circulation related to ozone depletion and its imprints on surface winds (e.g., Bitz & Polvani, 2012; Ferreira et al., 2015; Sigmond & Fyfe, 2010, 2014; D.W.J. Thompson et al., 2011; Turner et al., 2009) as well as to increased surface freshening of the Southern Ocean due to basal ice shelf melt (e.g., Bintanja et al., 2013, but see Swart & Fyfe, 2013). However, there is comparatively little understanding of the underlying reasons for the observed changes in the Weddell Sea (Figure 9). Our present knowledge is further hampered by a lack of observational data of ice drift, thickness, and melt required for better model development, interpretation, and prediction. From these limitations in our present understanding, the

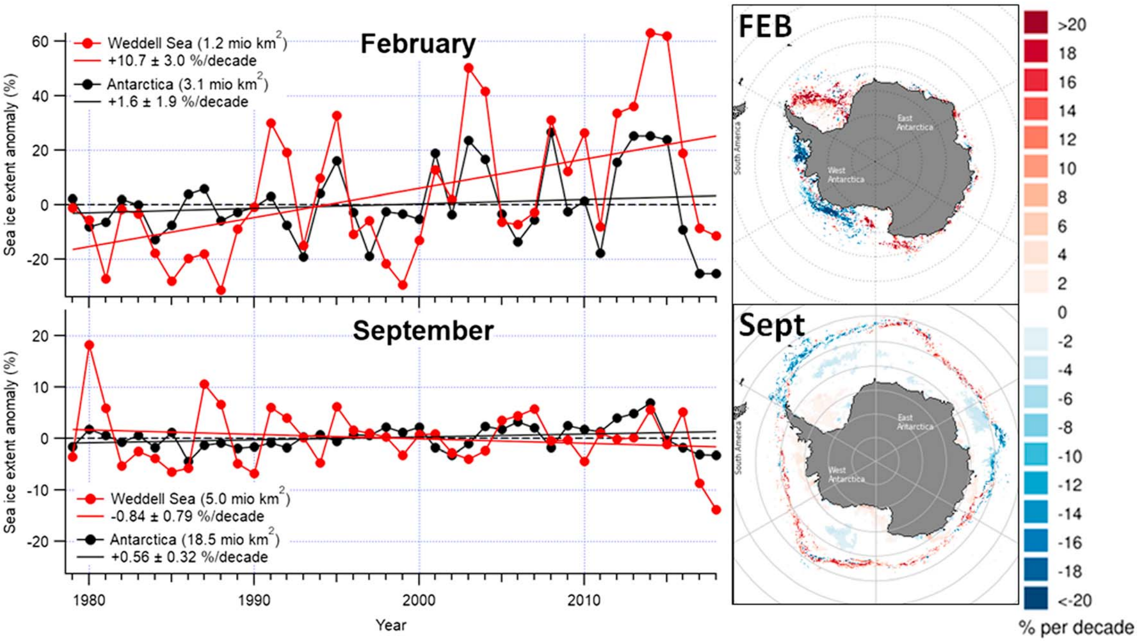

Figure 9. (left) Weddell Sea (between 0°W and 60°W) and Antarctic sea-ice extent trends in 1979–2018 for months with minimum (February [summer]; top) and maximum (September [winter]; bottom) ice extent. Parentheses show mean ice extent between 1981 and 2010 as reference. Data computed from sea-ice concentration provided by NSIDC (Cavalieri et al., 1996). Note different *y* axis scales. (right) Regional ice concentration trends in 1979–2018. Figure panels modified from NSIDC Sea Ice Index database, accessed 20 November 2018 (http://nsidc.org/data/seaice_index).

following research challenges can be identified, each requiring a close collaboration between sea-ice scientists and researchers from other disciplines.

5.2. Research Priorities

5.2.1. The Relationship Between Large-Scale Surface Winds, Ice Drift and Deformation, Ice and Snow Thickness, and Polynya Activity

Further understanding and prediction of large-scale WG sea-ice variability and change is a major scientific challenge. While the importance of wind forcing on ice drift and surface currents is well established (see above), freshwater may play an important role as well. For example, Naveira Garabato et al. (2016) and Pellichero et al. (2016) showed that the buoyancy, water mass transformations, and overturning circulation in the Weddell Sea are strongly influenced by sea-ice formation (negative freshwater flux) in the south and sea-ice melt in the north (positive freshwater flux). The sea-ice cover may also affect the momentum transport into the ocean, which could reduce Ekman pumping and upwelling into the center of the gyre. This has been shown only for the Beaufort Gyre in the Arctic, though the impact that this may have on the Antarctic, and the WG, has not yet been explored.

Recent work has shown that there is some correlation between observed changes in ice drift, winds, and changes in ice extent (P. R. Holland & Kwok, 2012). Those results suggest that the reduced ice drift in the western WG since 1992 might explain the observed retreat of winter ice extent (Figure 9). The reduced ice drift has also reduced sea-ice-associated freshwater fluxes in that region of the Weddell Sea, with potential consequences for reduced overturning circulation (Haumann et al., 2016). These observed trends in the western WG are in contrast with the changes observed in the Pacific sector of the Southern Ocean, where models show that regions with strongly increasing sea-ice concentration also have thicker ice (P.R. Holland et al., 2014). Better drift information from buoys and satellite data and better reanalysis products supported by more widespread marine meteorological observations are required to study the sea-ice observations with more detail in order to differentiate between distinct dynamic and thermodynamic processes in different regions of the WG (e.g., Heil et al., 2008; Schwegmann et al., 2011).

Of particular importance to estimates of sea-ice volume in the WG is a better determination of ice thickness. Satellites provide the most accessible data on the relevant spatial and temporal scales. Laser and radar altimetries are the most effective methods to date and have produced the first sea-ice thickness results over the Antarctic and WG (Giles et al., 2008; Kurtz & Markus, 2012; Yi et al., 2011; Zwally et al., 2008). However, altimetric sea-ice thickness retrievals are subject to large uncertainties, particularly due to the unknown thickness and microwave scattering properties of the overlying snow (e.g., D. Price et al., 2015; Xie et al., 2011). These uncertainties are further complicated by the snow-ice interface, which can fall below the local water level as a result of heavy snow fall on the relatively thin sea ice, causing a negative freeboard and widespread flooding (e.g., Massom et al., 2001). Similarly, frequent thaw-freeze events in summer cause strong snow metamorphism (e.g., Nicolaus et al., 2009), leading to large changes of snow microwave properties (e.g., Willmes et al., 2014).

Given the shortcomings listed above, new technologies need to be considered to measure sea-ice and snow thickness; for example, submarine or airborne measurements that have so far been used in small-scale studies should be applied more extensively in the future. Upward-looking sonars on moorings (e.g., Behrendt et al., 2015; Harms et al., 2001) and with autonomous vehicles (e.g., Dowdeswell et al., 2008) have been used to study long-term changes of sea-ice flux and thermodynamic processes in the WG. Airborne electromagnetic (EM) induction sounding provides coincident thickness and roughness information (Haas et al., 2008; Tan et al., 2012). Very promising efforts to obtain better snow thickness measurements have been presented by Kwok and Maksym (2014), using an airborne broadband Frequency-Modulated Continuous Wave radar. The studies by Haas et al. (2008) and Kwok and Maksym (2014) have demonstrated particularly well how regional ice and snow thickness variations are related to different ice regimes of first- and second-year ice of different age, origin, and roughness.

Variations in ice divergence and ice export influence the occurrence and extent of polynyas, which impact deep-water formation (see section 3.1). Therefore, polynyas may indicate sea-ice divergence and sea-ice thickness. Using either satellite passive microwave or thermal infrared data, Kern (2009), Drucker et al. (2011), and Paul et al. (2015) have shown that polynya area and ice production in the Weddell Sea are small compared to other Antarctic seas. Further, these quantities may have decreased slightly since 1992 and could therefore explain the observed decrease in ice velocity and ice divergence (P.R. Holland & Kwok, 2012). Hellmer et al. (2011) have also raised the possibility that changes in the Larsen Ice Shelves along the east coast of the Antarctic Peninsula could alter polynya activity and impact deep-water formation.

5.2.2. The Role of Ice Shelf Melt on Sea-Ice Mass Balance and the Platelet Ice Indicator

The melting of ice shelves produces freshwater (so-called ISW; see section 4.1) that contributes to stronger upper ocean stratification and therefore a reduced oceanic surface heat flux (e.g., Jacobs et al., 2011). There is some controversy about whether this negative climate feedback could in part be responsible for the observed expansion and potential thickening of the Antarctic sea-ice cover (e.g., Bintanja et al., 2013; P.R. Holland et al., 2014; Swart & Fyfe, 2013). ISW can also contribute to sea-ice expansion by causing platelet ice to form, which occurs under sea ice when ISW ascends to the surface and becomes supercooled (Langhorne et al., 2015). Remote sensing by airborne and ground-based EM sounding can measure the presence and thickness of platelet ice under the sea ice (Hunkeler et al., 2016; Rack et al., 2013). Extensive repeat EM surveys near prominent melting ice shelves could therefore be used to indirectly observe regional and interannual variations of ice shelf melt and its impact on changes of sea-ice thickness. Also, a network of fixed EM stations at the surface of fast ice could observe the seasonal evolution of platelet ice and ISW.

5.2.3. Understanding and Predicting the Weddell Sea Sea-Ice System's Present and Future Role for the Regional Ecosystem and Global Climate

Although we might have a reasonable understanding of the mean state of sea ice in the WG and depiction of some of its variability (Figure 9), a thorough understanding of its role in the WG system with its interactions and feedbacks with the atmosphere-ocean and ecosystems is still in development. From that end, it is imperative that systematic long-term observations be continued or initiated. These observations need to be linked to the development and validation of coupled atmosphere-ice-ocean models or climate models. Climate models still suffer from coarse resolution and insufficient parameterization of relevant processes but provide reasonable estimates of sea-ice extent (e.g., Landrum et al., 2012; Turner et al., 2013). However, regional models with higher resolution and more sophisticated sea-ice rheologies have been able to reproduce the observed sea-ice extent due to increases in circum-Antarctic winds, slight increases of Antarctic and Weddell Sea ice thickness, ridging, and volume (e.g., P. R. Holland et al., 2014; Massonnet et al., 2013; J. Zhang, 2014; see section 5.2.1).

Sea-ice characteristics and seasonal dynamics are key habitat parameters that also affect the cycling of carbon and nutrients, limit gas exchange between ocean and atmosphere, and trigger phytoplankton ice-edge blooms (see section 8.2.3). There is still an urgent need for large-scale assessments of sea-ice biomass and of the effects of changing ice conditions on ice algal production (e.g., Constable et al., 2014). This requires observations and models that capture the complexity of physical and biogeochemical processes or algal productivity in sea ice and their interaction with the ocean underneath or at the ice edge (e.g., Vancoppenolle et al., 2013; see section 10.2).

5.2.4. Will Antarctic Sea Ice Continue to Expand?

Despite the decadal expansion of Antarctic sea-ice cover, the ice extent reached a record low several days during the 2016–2017 and 2017–2018 austral summer season (Figure 9). The Weddell Sea polynya also appeared near Maud Rise for the first time since the 1970s during the same years. These events, which coincided with rapid summer sea-ice retreat, occurred during very low values of the Southern Annular Mode (SAM) index. The SAM index describes the sea level pressure differences between Antarctica and the lower latitudes, which can affect the latitude and intensity of the westerly wind belt that circles Antarctica. Years with a low SAM index and, therefore, a low-pressure difference are associated with a stronger than usual meridional atmospheric flow (Schlosser et al., 2017; Turner et al., 2017). Although the changes observed during the fall of 2016 and 2017 were within or only slightly exceeded the observed long-term average after accounting for interannual variability (Figure 9), they demonstrate that the Antarctic sea ice is sensitive to changes in atmospheric boundary conditions and related oceanic boundary conditions. These results illustrate the need to better understand the complicated interactions between atmosphere, ice, and ocean.

6. Chemistry I: Carbon Cycle of the WG

The global oceans slow the increase of carbon dioxide in the atmosphere by absorbing more than one quarter of all human-derived carbon dioxide (CO_2) emissions. The WG greatly contributes to this: the reduction in surface ocean carbon levels driven by heat loss during deep water formation and strong biological activity (that drives a CO_2 flux into the ocean) now more than compensate for the upwelling of old, carbon-rich waters (that tend to drive a CO_2 flux into the atmosphere). It is also an important region for how it connects the carbon cycle of the surface and deep oceans, both through the production of dense bottom waters that sink to the global abyss carrying carbon with them and the biological export of carbon to middepths where it is remineralized and transported to depth on climatically important time scales. The response of the delicate balance of these systems to global warming and increased atmospheric CO_2 levels are the key uncertainties for WG carbon cycle research. Specifically, what are the pathways and processes through which biology uses nutrients to produce new organic material? How is this transferred to depth and turned back into nutrients, and how will these intricate ecosystems react to warming surface waters; how will changing winds, stratification, and circulation affect the ability of surface waters to absorb CO_2 from the atmosphere? And will acidification ultimately impact the ability of the local biota to survive and the carbon uptake capacity of the gyre region as a whole?

6.1. The State of Knowledge

The Southern Ocean (>44°S) is an integral part of the global carbon cycle, both through its strong uptake of excess (anthropogenic) carbon dioxide (CO_2) from the atmosphere (~0.7 Pg C/year; Khatiwala et al., 2009; Mikaloff Fletcher et al., 2006) and as the location of strong outgassing of upwelled natural carbon (~0.4 Pg C/year, Mikaloff Fletcher et al., 2007). Within the net present-day Southern Ocean sink of 0.21 to 0.34 Pg C/year (Gruber et al., 2009; Landschützer et al., 2015; Takahashi et al., 2009), only weak CO_2 uptake occurs in the WG on an annual basis (Brown et al., 2015; Hoppema et al., 1999). However, a strong seasonal cycle in CO_2 disequilibrium and air-sea fluxes exists (Figures 10 and 11). Uptake of CO_2 mostly occurs following photosynthetic activity of phytoplankton in spring and summer (Hoppema et al., 1999), for example, as a rapid bloom development when sea ice retreats (Bakker et al., 2008) and when sudden cooling of surface waters occurs in autumn (Hoppema et al., 2000). In late autumn and winter, deepening of the mixed layer combined with upwelling of CDW high in CO_2 (Jullion et al., 2014) contributes to increasing surface

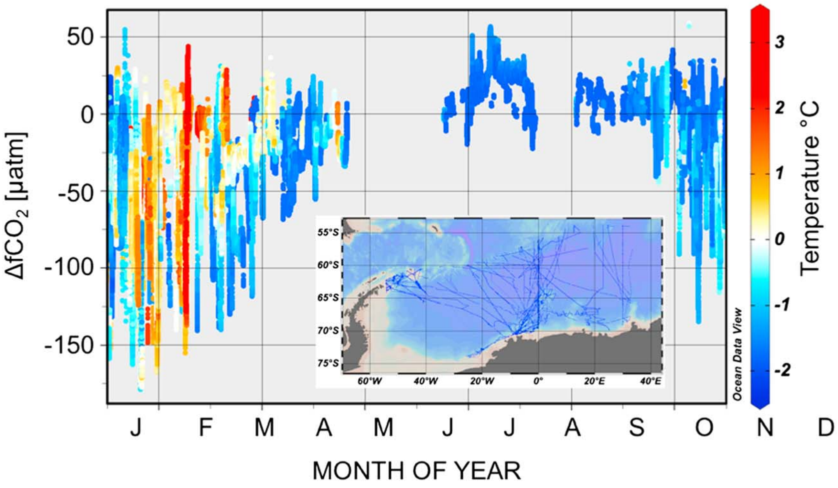

Figure 10. Seasonal cycle in air-sea CO_2 disequilibrium in the Weddell Gyre from historical sea-surface CO_2 measurements from 1984 to 2013 of SOCATv3 database—see inset for data locations (Bakker et al., 2016). Color scale is sea-surface temperature, showing that undersaturation cannot be fully driven by temperature. $\Delta fCO_2 = fCO_2$ in the ocean minus fCO_2 in the atmosphere.

ocean CO_2 concentrations and CO_2 release to the atmosphere (Brown et al., 2015). The advent of wintertime sea ice, however, caps the surface ocean and limits further CO_2 outgassing (Bakker et al., 2008; Brown et al., 2015).

In preindustrial times before 1850, the balance of these processes made the WG region a net source of carbon to the atmosphere, as winter outgassing outweighed spring-summer uptake (Hoppema, 2004a). However,

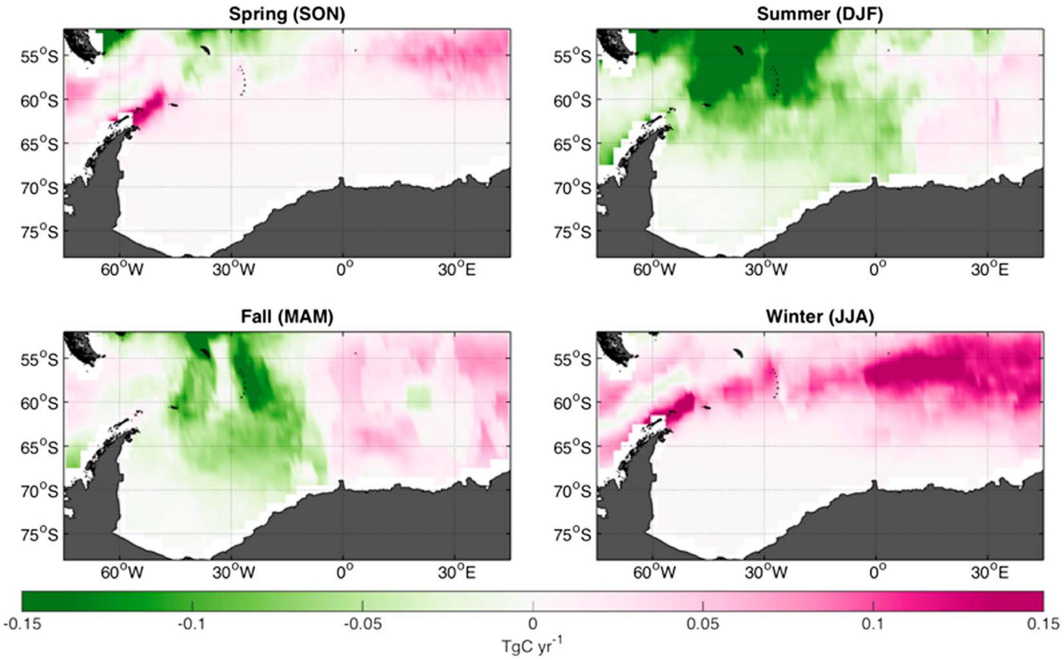

Figure 11. Seasonal average air-sea CO_2 fluxes for the 1998–2011 period for the Weddell Gyre and adjacent regions adapted from Brown et al. (2015), calculated using neural network-derived pCO_2 interpolation scheme sensu Landschützer et al. (2013) based on the SOCATv2 database (Bakker et al., 2014) where fluxes are proportional to the fraction of sea-ice-free water (from monthly satellite-derived sea-ice fields). Positive fluxes indicate outgassing, negative fluxes uptake. SON = September-October-November; DJF = December-January-February; MAM = March-April-May; JJA = June-July-August.

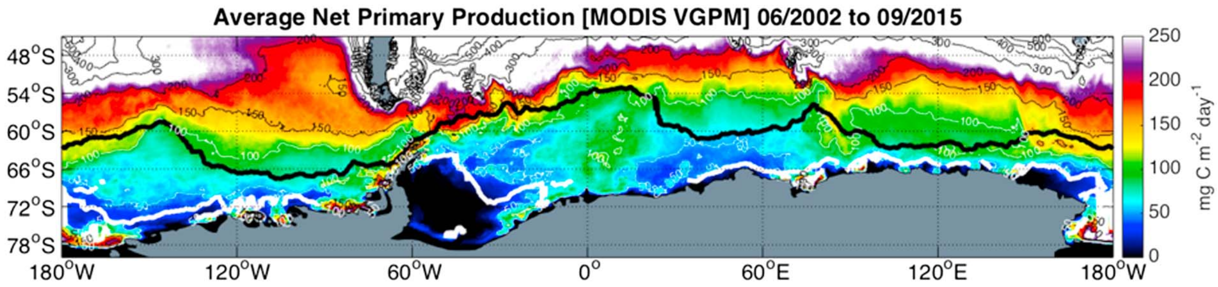

Figure 12. Average net primary production for the Southern Ocean, derived using monthly Aqua MODIS chlorophyll estimates and the Vertically Generalized Production Model (VGPM; http://science.oregonstate.edu/ocean.productivity/index.php; Westberry et al., 2008). Thick black line indicates maximum sea-ice extent for, for example, 2007. Thick white line indicates minimum sea-ice extent for, for example, 2008. Years chosen as middle of net primary production time series. MODIS = MODerate resolution Imaging Spectrophotometer.

elevated atmospheric CO_2 levels have shifted seasonal CO_2 disequilibria, reducing outgassing and enhancing biological drawdown of CO_2 (Hauck et al., 2015). The small net annual sink that results thus corresponds to a substantial anthropogenic CO_2 sink partly masked by a “natural” CO_2 increase or even outgassing. The region's environmental importance, however, relates to its location of intimate connection between the surface and deep oceans, the rate-limiting step in determining long-term CO_2 sequestration (Ito et al., 2010). Dense and bottom water formation on the region's continental shelves (Nicholls et al., 2009) and export into the abyss (Jullion et al., 2014; Orsi et al., 2002) transfers anthropogenic carbon to the deep ocean (0.01–0.04 Pg C/year locally in the WG and 0.06 Pg C/year globally south of 55°S; Ito et al., 2010). Combined with the biological export production of carbon to middepths and transport northward within the CIW (a shallow variety of CDW in the interior) of the WG (0.02 Pg C/year; Hoppema, 2004b), the region plays a critical role in removing CO_2 away from atmospheric interaction on long time scales that are climatically relevant (>100 years).

There are three major strands that determine the behavior of the carbon cycle in the WG: (i) biological factors that determine spring and summer CO_2 undersaturation levels, carbon remineralization at shallow depths, and partitioning between organic and inorganic as well as dissolved and particulate components; (ii) physical factors that act on this disequilibrium and determine the atmospheric CO_2 uptake and transport to depth; and (iii) chemical factors (i.e., buffering capacity/Revelle factor) that determine the ongoing capacity of surface waters to accommodate increasing CO_2 concentrations (Lenton et al., 2013). Climate change will impact the CO_2 balance of the surface ocean via direct effects on circulation, sea-ice cover, and glacial melt (Abernathey et al., 2016) and through indirect effects on the biological carbon pump as well as the rate at which these changes are propagated to depth. While progress has been made toward understanding their impacts, substantial aspects still suffer from limited understanding and raise a number of challenges.

6.2. Research Priorities

6.2.1. Biological Factors

Primary production in sea-ice and surface waters of the WG is accompanied by carbon uptake by biota, causing CO_2 undersaturation in the surface water with respect to the atmosphere, which drives the uptake of CO_2 from the atmosphere (Figures 10 and 11). Highest production levels in the Weddell region are observed in the seasonal ice zone (Figure 12), sustained by the upwelling of dissolved inorganic carbon (DIC) and nutrient-enriched WDW (Figure 4) and the supply of micronutrients (particularly iron, Fe) from sea ice and iceberg melt. Elevated levels are also observed at coastal polynya locations that are periodically ice free and within sea ice (Arrigo et al., 2008; Geibert et al., 2010; Thomas et al., 2001), possibly enabled by an elevated supply of scarce Fe from the continent. Estimates of annual net primary production (NPP) for the gyre vary between 0.11 and 0.25 Pg C/year (Brown et al., 2015, and references therein). The large range is representative of the disparate data sets, methods used, and extrapolations applied to extend temporally and spatially poor data to the entire WG. Overall, thorium isotope-derived export production estimates, inverse modeling, and a relatively shallow nutrient maximum (250–400 m) in the central WG together suggest that much of the NPP undergoes very shallow remineralization within the CIW of the WG (particle export is thought to be negligible below 250 m) prior to resupply to the surface layer through upwelling

(Usbeck et al., 2002). Note the latter vertical transport mechanism is not thought to apply for Fe, where sea-ice and iceberg melt are thought to be the major supply routes (Geibert et al., 2010). A small proportion of NPP may reach deeper layers during major bloom events (Wedborg et al., 1998) before being transported from the gyre at depth, as part of the outflow of CIW that sinks below the ACC on exiting the Weddell region (Hoppema, 2004b). A research question that arises here is the following: How large is the reduction in the partial pressure of CO_2 in the surface layer of the WG due to biological production and how large is its subsequent effect on air-sea CO_2 fluxes?

During biological activity, net organic material production occurs in both particulate and dissolved phases. In contrast to North Atlantic blooms (and due to ecosystem dynamics unique to the Weddell region), ~70–99% of total organic carbon produced during Southern Ocean blooms is preferentially partitioned to particulate organic carbon (POC) over dissolved organic carbon (DOC; Carlson et al., 2000). Substantial fractions of this particulate organic material are quickly converted into dissolved organic molecules by extracellular release, cell lysis, heterotrophic turnover, photochemical degradation, and solubilization from particles (Carlson & Hansell, 2015). Together with the unique circulation in the WG, this has several consequences for the organic carbon cycle: (1) In the WG, new DOC in the surface is rapidly remineralized or exported by mixing, advection, and deep-water formation; this is known from the Ross Sea (Carlson et al., 2000) where a recent study found a positive net community production that was rapidly exported to below 200-m water depth (DeJong et al., 2017). Surface water DOC concentrations in the WG are among the lowest of any major open ocean region (~46 μmol/L; Lechtenfeld et al., 2014; Zemmelink et al., 2008). (2) POC production in turn introduces fresh carbon into the shallow subsurface layers, which can be readily utilized by bacteria. However, in addition to direct remineralization, bacteria in the Weddell Sea can produce 5–9% of nonlabile DOC from simple organic substrates (Koch et al., 2014) on time scales comparable to the residence time of surface waters (~2.5–3 years; Hoppema et al., 1999). This process contributes to the residence time of carbon in the water column. (3) Radiocarbon data have helped identify a significantly higher reservoir age of DIC available for assimilation (i.e., utilization for organic matter production) compared to other upwelling regions (Leboucher et al., 1999; Schlosser et al., 1994) where phytoplankton-derived POC export may carry radiocarbon-enriched material back to depth. However, a recent study states that the DIC-radiocarbon content and DOC concentrations in the deep Southern Ocean largely reflect the conservative mixing of several deep-water masses (Bercovici & Hansell, 2016).

Several research questions arise from our lack of knowledge of organic carbon in the WG:

1. How large is the flux of fresh DOC into the mesopelagic and bathypelagic by solubilization of sinking particles or deep-water formation in the WG?
2. What are the turnover times of different organic matter fractions at various depths and what are the control variables?
3. What causes the partitioning of new production into POC, DOC, and DIC pools, enabling the middepth sequestration of carbon and how sensitive is it to external forcing?

Our understanding of the magnitude of carbon sequestration and its variability and the specific processes and pathways that contribute to and drive the local biological carbon pump are still limited. Thus, for a satisfactory and comprehensive understanding of the organic carbon cycle, a quantitative assessment of organic matter reactivity is fundamental. In the WG, the bioavailability of organic matter is particularly sensitive to and controlled by low temperatures and low substrate concentrations (Pomeroy & Wiebe, 2001), in addition to other controls such as substrate composition (Ksionzek et al., 2016) and micronutrient availability.

6.2.2. Physical Factors

Uptake of anthropogenic CO_2 by the ocean is driven by the continuous increase of CO_2 in the atmosphere and the ocean's continued ability to remove it from the atmosphere through the formation of mode and intermediate waters in the ACC and deep and bottom waters in the WG as part of the large-scale global overturning circulation. Natural CO_2 fluxes, meanwhile, will be influenced by upwelling rates, stratification, and changes in wind regimes. The carbon balance in the WG has been perturbed by a strengthening of the subpolar westerlies (Marshall, 2003; D.W.J. Thompson et al., 2011) that cause stronger upwelling of carbon and nutrient-rich subsurface water (Hauck et al., 2013; Hoppema et al., 2015). On the one hand, additional CO_2 input from depth can lead to a reduction of CO_2 uptake (le Quéré et al., 2007). On the other hand, the concomitant vertical supply of macronutrients and micronutrients (Hoppema et al., 2015) could stimulate

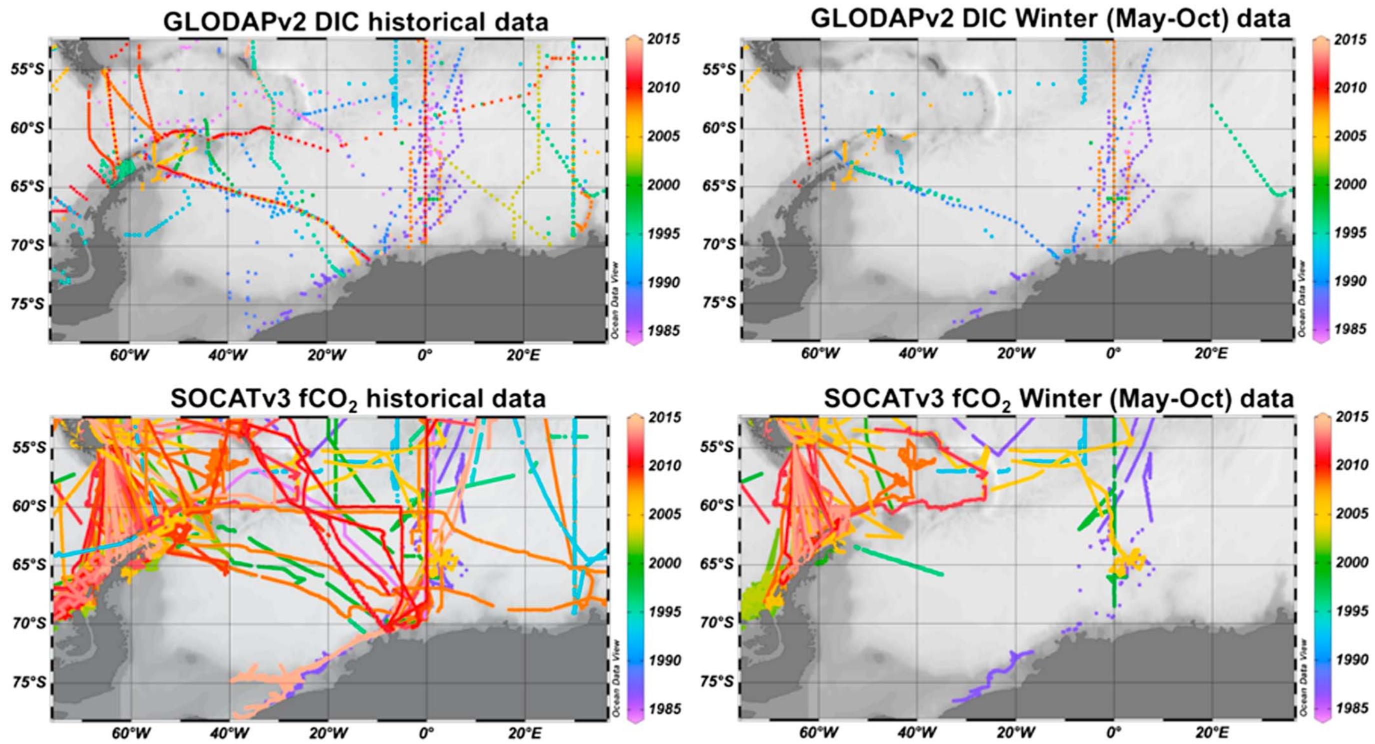

Figure 13. Carbon data availability in the Weddell Gyre region. (top) All Dissolved Inorganic Carbon (DIC) data (left) and for winter only (May-October, right) as available from the GLODAPv2 data set (Olsen et al., 2016; Schlitzer, R., Ocean Data View, odv.awi.de, 2018). (bottom) All surface fCO$_2$ data (left) and for winter only (May-October, right) as available from the SOCATv3 data set (Bakker et al., 2016). The colors indicate the year of collection (same scale for each plot). fCO$_2$ is the fugacity of CO$_2$, nominally quite similar to the CO$_2$ partial pressure, pCO$_2$.

biological production, leading to more atmospheric CO$_2$ uptake, thereby moderating the reduction in CO$_2$ uptake (Hauck et al., 2013). While the dynamics of the seasonality of the carbon cycle in the WG are roughly understood (Brown et al., 2015), winter data are scarce (Bakker et al., 2014; Figure 13), and a concerted effort is needed to reduce the uncertainties on annual CO$_2$ uptake and variability on subseasonal to interannual time scales (Monteiro et al., 2015) as a benchmark for future change. The WG's air-sea CO$_2$ flux is a result of a delicate balance of large-scale horizontal and vertical water mass transports that are highly variable and sensitive to heat and freshwater fluxes and wind forcing (Brown et al., 2015; Meredith et al., 2008; Naveira Garabato, MacGilchrist, et al., 2017). However, the relationship between changes in ocean circulation and CO$_2$ fluxes is still not well understood. Against this background, the following research questions arise:

1. What changes in the climate system are critical for the carbon cycle? For example, are the westerly winds increasing in a zonally symmetric manner or are other atmospheric processes becoming dominant in the WG (Landschützer et al., 2015)?
2. The upper ocean overturning and mixing could increase due to a strengthening of winds or decrease due to global warming and glacial/sea-ice melt (Hauck et al., 2015)—which will it be and how will this impact carbon fluxes?
3. Adjustments in circulation and overturning rate (Waugh et al., 2013) may already be impacting the ventilation of water masses within the WG (Huhn et al., 2013). Models, however, suggest that decreased rates of net CO$_2$ uptake may be due to increased natural outgassing rather than reduced anthropogenic carbon uptake (e.g., Lovenduski et al., 2013). What is the main driver of carbon uptake – deep-water formation that sequesters anthropogenic carbon or wind-driven upwelling, forcing natural carbon outgassing?

Considering the small but significant difference between surface and deep DOC concentrations, deep-water formation potentially creates an annual export flux of refractory DOC in addition to anthropogenic carbon export. Microbial breakdown of DOC in the deep ocean or export with bottom waters may be the ultimate sink for this DOC, but these pathways, magnitude, and their relation to DIC transports are still unresolved and their response to circulation change unknown.

6.2.3. Chemical Factors

The sustained buildup of DIC concentrations in the surface layers across the gyre over the past 40 years (Hauck et al., 2010; van Heuven et al., 2011) suggests continued anthropogenic CO_2 uptake, with local sea-surface pCO_2 time series indicating a sink of increasing magnitude for waters south of the Antarctic Polar Front in the Drake Passage (Munro et al., 2015) and shelf waters at the Antarctic Peninsula typically being close to saturation with anthropogenic CO_2 (van Heuven et al., 2014). However, accumulation of DIC in the water column is not significant at all depths (Hauck et al., 2010; van Heuven et al., 2011). Storage rates of DIC in both Weddell Sea Deep Water and Weddell Sea Bottom Water are location-dependent and determined by source region and formation process. For example, waters deriving from interaction with the Filchner-Ronne Ice Shelf show only small CO_2 increases in contrast to larger trends of increasing DIC concentrations observed in waters formed at the western boundary (van Heuven et al., 2014). A substantial divergence in anthropogenic CO_2 inventory estimates calculated using different estimation techniques (Khatiwala et al., 2013) means that much work remains in optimizing these methods in the Weddell region. Projections propose sustained anthropogenic carbon uptake in the WG (Hauck et al., 2015; Ito et al., 2015); however, continued uptake of additional CO_2 from the atmosphere will be subject to multiple opposing stressors, such as changes in the wind regime, sea-ice trends, and warming and circulation-related stratification. Important research questions arising from these issues are the following:

1. Deep-water formation is mainly driven by brine rejection in the Weddell Sea shelf region, a process that is sensitive to the temperature and salinity balance that affects the equilibrium concentration of dissolved carbon dioxide. How will the system respond to accelerated warming and an intensified freshwater cycle?
2. A significant unknown is the response of biota in the face of warming and acidifying surface waters and decreasing carbonate ion availability—how will this affect carbon drawdown driven by biological production that leads to surface CO_2 undersaturation and thus uptake of anthropogenic CO_2 from the atmosphere sensu Hauck and Völker (2015)?

There is thus great scope to improve our comprehension of the carbon cycle in the WG, with important questions remaining and large gaps existing in available data sets; poor spatial and temporal resolution in observations (Figure 13) is currently the limiting factor for advancing understanding and projecting changes in CO_2 throughout the water column and its physical and biological drivers. To address the above challenges, current observational arrays (such as repeat hydrographic cruises and biogeochemical floats) must be sustained and extended where possible. A dense network of high-quality marine CO_2 observations and auxiliary biogeochemical and ecosystem data is essential to be able to detect and to attribute carbon flux changes to their multivariate drivers, taking advantage of new advances in autonomous technologies to capture new, currently elusive, wintertime measurements and higher resolution spring-summer-fall observations. Targeted observational process studies across the spectrum of WG ecosystems, combined with the probing of higher-resolution eddy-permitting general circulation models and ocean biogeochemical models and satellite-derived reanalysis products, should enable substantial progress to be made in our understanding of these problems.

7. Chemistry II: Elemental Cycling in the WG

This section focuses on chemical elements playing a foremost role in oceanography, marine biology, and marine chemistry. These include the major nutrients like nitrate, phosphate, and silicate, as well as trace metals like iron, manganese, and zinc. In addition, we highlight natural radioactive tracers for oceanic processes like radium, actinium, and thorium. The distributions and cycling of those elements are determined by different processes in the water column and for some by input from the atmosphere. Underlying the elemental cycling in the WG is its unique physical processes: deep upwelling (rising of water from depth), combined with clockwise circulation and bottom-water formation. The rapid upwelling of deep water masses causes the concentrations of major nutrients to be among the highest found anywhere in the surface ocean. Due to iron limitation (insufficient iron available for full growth), algal growth cannot remove all major nutrients from surface water. A partial depletion of nutrients occurs in the productive summer

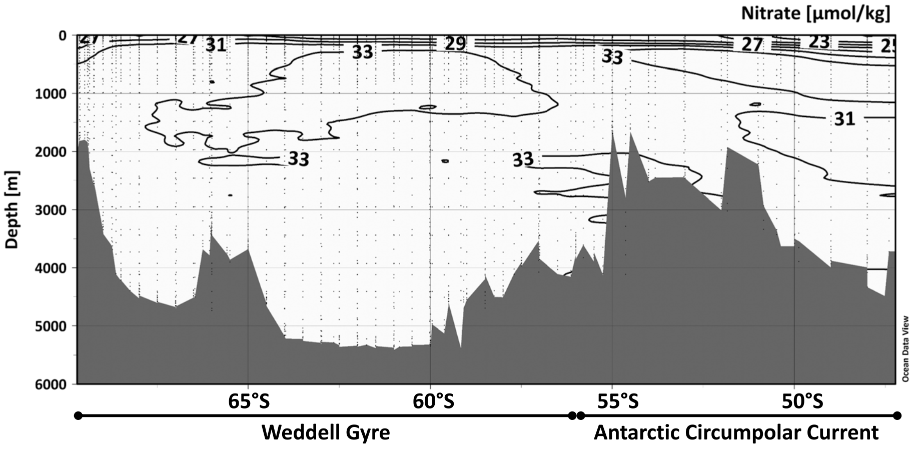

Figure 14. Nitrate along the Prime Meridian (data source: World Ocean Circulation Experiment). In the Weddell Gyre, high nitrate values reach the sea surface, and a maximum is observed at subsurface to intermediate depths.

season, resulting in biological patterns superimposed to the physical ones. Oxygen is taken up where the biomass is remineralized, resulting in local signatures in the oxygen content. Intermediate waters display low iron concentrations, limiting the potential iron supply by deep upwelling. The formation of bottom water transfers surface signatures of elements and isotopes to the deep WG, from where they can penetrate the world's oceans as AABW. Natural radionuclides are used to constrain the time scales of circulation, upwelling, and balance between particle flux, remineralization, and water exchange.

7.1. The State of Knowledge

We discuss three characteristics of elemental cycling of the WG, which makes it unique within the global ocean context. (1) The WG is one of the few regions in the global ocean where deep upwelling contributes strongly to the distribution of tracers (Rutgers van der Loeff & Berger, 1993) and nutrients (Dieckmann et al., 1991; Hoppema et al., 2001, 2015; Papadimitriou et al., 2007). In consequence, many chemical species enriched in intermediate and deep waters can be found at elevated concentrations in surface layers. (2) The combination of physical and biogeochemical processes in the WG influences the world's oceans through the formation of water masses that spread north, well beyond the equator. (3) The distributions of macronutrients, micronutrients, and tracers with no active role in biogeochemistry are closely linked in this region (Hellmer et al., 2016).

7.1.1. Distribution of Macronutrients and Micronutrients

The major nutrients, nitrate (N) and phosphate (P), have very similar distributions in the WG since they are involved in and influenced by the same processes. In the surface layer, there is a nutrient minimum, which is lower in the summer than in the winter due to uptake of nutrients for photosynthesis (Figure 14). It should be noted that, albeit locally a minimum, the Antarctic surface waters are known to have the highest nutrient concentrations in the world's upper-ocean layers.

A nutrient maximum is found in the WDW beneath the surface layer (Figures 4 and 14). The high concentrations there result from (1) advection of nutrient-rich CDW from the ACC into the WG and (2) local enhancement due to remineralization of organic matter at relatively shallow depths. This water mass is known as the CIW (Whitworth & Nowlin, 1987). Overall, nutrient concentrations decrease monotonically between a maximum located at the WDW lower boundary (~1,500 m) and a minimum at the bottom water. Together with the comparatively nutrient-poor surface water, the nutrient-rich WDW is the main source of the locally formed Weddell Sea Bottom Water. Thus, at or near the bottom of the basin, the nutrient minimum results from the low-nutrient content of the surface water mixing component of Weddell Sea Bottom Water. In the bottom layer, smaller local maxima or minima occur, which are due to small-scale circulation features as well as local nutrient inputs (especially silicate) from the sediments (Schlüter et al., 1998). The distribution of silicate (or silicic acid, Si) is subject to the same processes but with some significant differences. The silicate maximum occurs comparatively deeper due to the dissolution length scale (relative

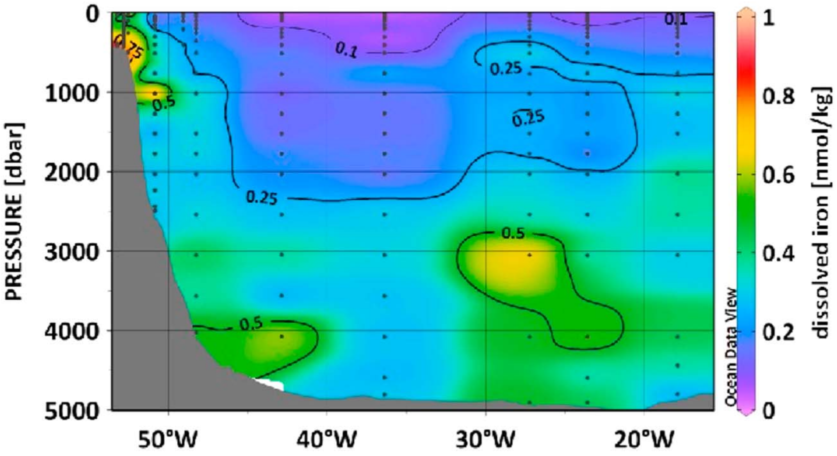

Figure 15. Dissolved iron distribution on a transect from the tip of the Antarctic Peninsula to near Kapp Norvegia (about 18°W; data from Klunder et al., 2014; Schlitzer, R., Ocean Data View, odv.awi.de, 2018).

to organic matter remineralization) of siliceous phytoplankton (mainly diatoms). In the WG, concentrations at the silicate maximum are much higher than those from the ACC waters entering the WG. In the bottom layer, the variability of silicate is greater than that of the other nutrients. Consistent with other nutrients, a silicate minimum is observed in recently formed bottom water. However, in some regions with intense interactions with the sediments, for example, off Kapp Norvegia, very high silicate concentrations are found (Hoppema et al., 2015), the highest in the entire WG.

The distribution of dissolved oxygen (DO) in the WG is overall reciprocal to that of the macronutrients. A DO maximum at the surface results from O_2 production by algae (photosynthesis) and equilibration with atmospheric oxygen. A DO minimum occurs just below the pycnocline, advected from the ACC. This DO minimum is enhanced in the gyre's interior (within the CIW) due to degradation of organic matter. In the bottom water, a DO maximum is found owing to the high-DO surface water mixing component of Weddell Sea Bottom Water. Given that deep and bottom waters of the WG contribute to the AABW, the latter is also rich in oxygen. Freshly ventilated AABW spreads far north, filling the bottom layers of a large part of the world's oceans, and thus contributes to the availability of oxygen during early diagenesis in surface sediments underlying AABW.

The distribution of iron is presented as an example of trace metals (Figure 15). Iron and manganese play a role in biological processes, with Fe being a limiting micronutrient for primary production in the Southern Ocean (Moore et al., 2013). There are indications that manganese may be colimiting in certain areas (Middag et al., 2011). Fe and Mn exhibit a minimum in the surface layer just like the macronutrients, driven by net uptake from primary producers. These trace metals are also supplied to the surface layer from the atmosphere (Croot et al., 2004) or advected laterally from shallow continental shelves (Klunder et al., 2014), particularly around the Antarctic Peninsula (de Jong et al., 2012). Mn exhibits two maxima, one at about 100 m and another one at the seafloor of the Weddell Sea. In the deep water separating those maxima, a Mn minimum around 2,000 m is caused by scavenging (Middag et al., 2013). Concentrations here are very low with only 0.1–0.2 nmol/L. Over the shelves and continental slope, the Mn and Fe concentrations are strongly enriched. At the bottom of the northern boundary ridge of the Weddell and Enderby basins (Figure 2), both Mn and Fe are very high due to input from hydrothermal vents (Klunder et al., 2011; Middag et al., 2011). Still, deep water Fe concentrations in the WG (0.3–0.4 nmol/L) are among the lowest in the global ocean, reflecting low local inputs to deep waters. In contrast, the micronutrient zinc (Zn) that is present at low concentrations in large parts of the ocean is relatively abundant in the WG (Croot et al., 2011; Zhao et al., 2014).

Generally, correlations between the macronutrients (N, P, or Si) and micronutrients (Fe or Mn) occur in the upper layers of the WG (Klunder et al., 2014). In the case of Fe, correlations also are found in deep waters (Klunder et al., 2011). Another example of a relationship between macronutrient and micronutrient vertical distributions is that of cadmium (Cd) and phosphorus (P; Boyle, 1988), leading to Cd distribution reflecting biological activity. This yields distinctive elemental ratios (Cd:P) and Cd isotope patterns in the high-nutrient setting of the WG. This ratio has a pronounced effect on the signature of water masses outside the WG, in particular by the export of Antarctic Intermediate and Mode waters (Abouchami et al., 2014; Baars et al., 2014).

7.1.1.1. Physical Controls on Nutrient Distribution

Following transport of water into the WG through its northern and especially eastern boundaries (see also section 3), nutrients associated with the CDW are redistributed throughout the WG by its cyclonic circulation. Vertical mixing and diffusion change the nutrient concentrations on the way through the gyre, leading to spatial differences and specific patterns. For example, the nutrient maximum of the WDW is eroded along the way within the main westward flow of the gyre in the south. Special features include the ASF with deep mixing and high productivity in its associated current (decreasing nutrient concentrations), the central gyre with subsurface enrichment due to shallow mineralization (Usbeck et al., 2002), and Maud Rise with its Taylor cap circulation and enhanced upwelling (increasing nutrient concentrations). Owing to the divergent gyre circulation, upwelling occurs at all depths in the WG interior. For the surface layer, this means that the nutrient inventory is replenished and high concentrations are maintained. Only due to the uptake of nutrients during photosynthesis are their concentrations lowered to some extent. At the onset of the austral winter, deep mixing creates a cold surface mixed layer with relatively homogeneous (high) nutrient concentrations, that is, the winter mixed layer.

The formation of deep and bottom waters, bringing down surface waters rich in nutrients and oxygen to the sea floor (see section 7.1.1), also serves as a pathway of molecular tracers and elements to the deep sea. This allows very young tracer signatures such as tritium, chlorofluorocarbons, and ^{14}C to penetrate the deep ocean. Iodide also has potential as a tracer of newly formed Weddell Sea Bottom Water (Bluhm et al., 2011) as it is only slowly oxidized to iodate in these cold waters. Bottom water formation brings down Mn and Fe into the abyssal basin (Middag et al., 2011) resulting in the elevated Fe in the bottom water (Westerlund & Öhman, 1991) but only close to the source. It appears that Fe is scavenged rapidly as lower values are found away from the coast in the deep waters of the central WG (de Jong et al., 2012; Klunder et al., 2011).

Summarizing, nutrients enter the WG along its northern boundary, and there is a large advective term associated with the eastern inflow. Within the residence time of water in the WG, macronutrients cannot be depleted. There is particulate export from the sea surface, which is remineralized mostly at shallow depths, leading to a nutrient maximum within the 300- to 1,500-m depth range. Physical export of nutrients takes place from the surface layer equatorward along the northern boundary, via Antarctic Intermediate Water and via bottom waters (Weddell Sea Bottom Water/AABW).

7.1.1.2. Biological Controls on Nutrient Distribution

Primary production adds to the redistribution of all nutrients in/from the surface layer via the uptake of nutrients for phytoplankton growth and eventually via the biological carbon pump (i.e., vertical export of biogenic particles and subsequent remineralization at depth). Phytoplankton uptake lowers nutrient concentrations during spring and summer, when the surface mixed layer is relatively shallow and light and Fe are available. During the growth season and toward autumn, a proportion of the dead organic material in the surface layer is degraded, bringing nutrients back into solution. On an annual basis, there is a net loss of nutrients from the surface layer; a fraction of nutrients is transferred into the dissolved organic matter pool (see section 7.1.1) and as particulate nutrients into higher trophic levels (see section 9) or down the water column as dead organic matter. Degradation and remineralization of this flux of organic matter occurs mainly at shallow depths of 300–800 m, de facto pumping down nutrients from the surface. Generally, only a small proportion of nutrients is transferred to the abyss in this way; although, in some areas of the gyre like the western Weddell Sea (Wedborg et al., 1998), this might be larger. Biological activity in the surface water near the regions of deep and bottom water formation will also codetermine the nutrient content of the AABW.

Sea ice plays an ambiguous role in nutrient cycling. While thick layers inhibit light penetration and therefore productivity and nutrient uptake, there is enhanced stability of the water column near the sea-ice edge. There, favorable microenvironments and micronutrient sources promote phytoplankton growth in many places (Lannuzel et al., 2008). Within the sea ice, the nutrient concentrations appear to be highly variable, depending on the ice texture and age (Dieckmann et al., 1991), still the nutrient concentrations mostly

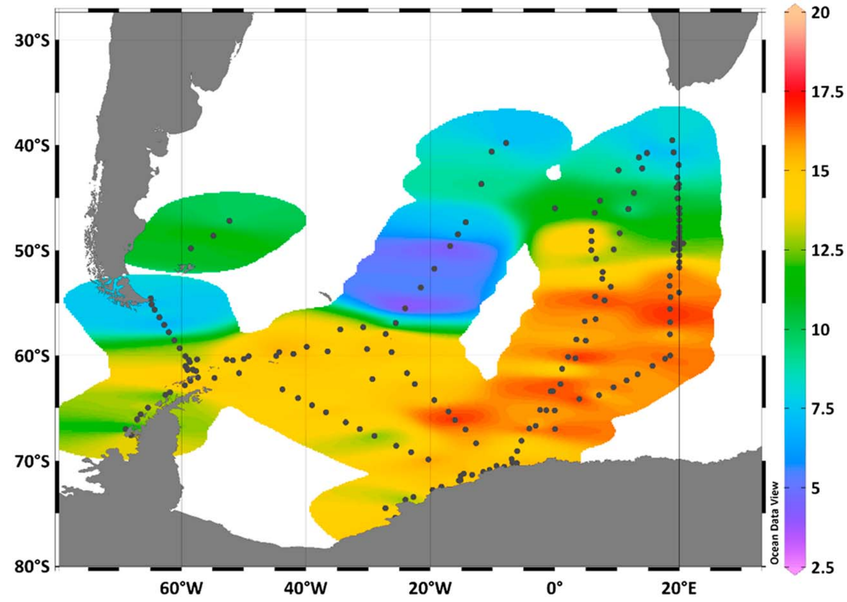

Figure 16. Distribution of ^{226}Ra (dpm/100-kg seawater) in the surface layer of the Weddell Gyre, whose main source is the deep-sea floor (half-life 1,600 years; data from Hanfland, 2002; Schlitzer, R., Ocean Data View, odv.awi.de, 2018). Its distribution reflects the effect of deep upwelling, which is more pronounced in the eastern part.

appear to be lower than in the water column owing to photosynthesis within the ice during summer (Papadimitriou et al., 2007).

7.1.1.3. Natural Radionuclides: Tracers for Productivity and Circulation

Natural radionuclides as chronometers and flux gauges have played a significant role in understanding physical processes and biogeochemical cycling in the WG. The distribution of soluble naturally occurring radionuclides like ^{226}Ra, ^{228}Ra, or ^{227}Ac is controlled by supply from sediments, loss by decay with a specific half-life, and redistribution due to diffusion and advection. Particle-reactive radionuclides like ^{230}Th, ^{231}Pa, or ^{234}Th are additionally affected by the delicate balance of production by their parent isotopes and removal by particles. The distribution of radionuclides therefore reflects all these processes integrated over a time scale of the respective half-lives or removal rates. This integrated picture of processes redistributing radionuclides is particularly useful for deriving process rates in remote parts of the ocean that are rarely accessible.

In surface waters of the Southern Ocean, ^{226}Ra is found to be correlated to silicate (Chung, 1980; Hanfland, 2002), but in the Weddell Sea, ^{226}Ra displays high values throughout the water column due to upwelling (Chung & Applequist, 1980), with the highest values toward the east where more localized upwelling occurs (Figure 16). ^{210}Pb shows a maximum at ~500-m depth, an unusual profile that has been ascribed to the combination of upwelling of WDW and near-absence of atmospheric input. ^{228}Ra is supplied by sediments and activities build up in oceanic shelf regions. In surface waters south of the Antarctic Polar Front, ^{228}Ra is found to be very low (Hanfland, 2002; Li et al., 1980) because the larger depths of the shelf seas in combination with short residence times prevent it from accumulating to high concentrations (Rutgers van der Loeff, 1994). ^{227}Ac, a daughter nuclide of ^{231}Pa, is released from deep-sea sediments. The distribution of the ^{227}Ac activity in excess of ^{231}Pa in the water column has been used to estimate an upwelling rate of 55 m/year in the eastern WG (Geibert et al., 2002). In summary, the distribution of soluble natural radionuclides reflects the effect of deep upwelling, and some may even serve to determine rates of deep-water replenishment.

Due to the rapid circulation of the WG, particle-reactive radionuclides with oceanic residence times of a few decades may still be significantly affected by advective processes. In fact, the Weddell Sea was the first place where the effect of advection and upwelling on the distribution of ^{230}Th was demonstrated (Rutgers van der Loeff & Berger, 1993). As a result of rapid circulation (residence of Lower CDW in the WG of about 35 years) and extremely low particle fluxes to the deep ocean (Fischer et al., 1988; Leynaert et al., 1993; Usbeck et al., 2002), a significant proportion of ^{230}Th can be exported from the WG. ^{230}Th burial rates in sediments are only 40% of the ^{230}Th production rate (Walter et al., 2000). Thorium is usually deposited near its production site (Henderson et al., 1999), making it a well-defined constant flux tracer elsewhere, but corrections to the constant-flux model are required in the WG (Geibert et al., 2005). ^{231}Pa is preferentially removed by biogenic opal as produced by diatoms, affecting the ^{231}Pa/^{230}Th ratio south of the Antarctic Polar Front, including the WG (Walter et al., 1997). Export of ^{230}Th and ^{231}Pa from the WG also affects their budgets in the ACC and further north of it (Rutgers van der Loeff et al., 2016). The ^{231}Pa/^{230}Th pair in marine sediments is a widely used proxy pair to reconstruct ocean circulation and particle fluxes in the past, supported by global circulation models incorporating particle fluxes (e.g., van Hulten et al., 2018). Export of radionuclides from the WG therefore is essential in paleo-reconstructions of adjacent regions.

The short-lived radiotracer ^{234}Th (24.1 days half-life) is well suited to study particle fluxes in the upper ocean on a time scale of about 2 months as it is constantly replenished by decay from uranium. In the WG, ^{234}Th has been used to quantify export production and mineralization depth (Usbeck et al., 2002). It has confirmed a shallow remineralization of biogenic particles that explains the low fluxes to the deep WG in spite of considerable production at the surface and has been applied to derive the vertical fluxes of organic carbon and trace metals (Rutgers van der Loeff et al., 2011). Anomalous distributions of ^{234}Th exceeding the equilibrium value with ^{238}U are found in the presence of ice algae (Rodriguez y Baena et al., 2008), pointing to enrichment of ^{234}Th in sea ice. In summary, natural radiotracers indicate that the WG is a region of intense vertical upwelling of Lower CDW with a short residence time on the order of 35 years, in which biogenic particles from the sea surface do not reach the sea floor but are remineralized just below the euphotic zone. Dissolved components are then being exported via Antarctic Intermediate Water and AABW and available for elemental cycles equatorward of the WG.

7.2. Research Priorities

7.2.1. Nutrient Temporal Trends

Being steady-state tracers, macronutrients are expected to show large variability in their distributions but no temporal trends. Along a section across the Weddell Sea, highly accurate nutrient data were collected on a repeat section between 1996 and 2011 (Hoppema et al., 2015). There was a highly significant increasing trend of all nutrients in the surface layer, some increases in the WDW, and major trends in the bottom layer for silicate only ($p < 0.05$). Near-surface nitrate increased by 0.1 μmol·kg^{-1}·year^{-1}, while bottom layer silicate increased at a rate of up to 1 μmol·kg^{-1}·year^{-1} (Hoppema et al., 2015). For example, in the surface layer, the silicate concentration increased by as much as 20 μmol/kg in these 15 years. The cause was suggested to be enhanced upwelling. Nutrients thus appear to be useful indicators of changes occurring in the WG system. Nutrient measurements along this Weddell Sea section and possibly other sections should be continued to monitor future changes in upwelling. Helpful to this cause is the current availability of certified reference materials for nutrient analysis in seawater (Aoyama et al., 2012), which will cater for consistent data among different laboratories. There is no understanding whether the upwelling rate will continue to intensify. Teleconnections like the Southern Annular Mode may play a role in nutrient changes in the WG, which requires more investigation. For example, a wind-driven increase in deep upwelling might enhance nutrient and micronutrient supply to the surface, but observations are scarce, and more investigations are needed to confirm the actual role of this mechanism. Local feedbacks may also be imminent, for example, changes in biological activity in the gyre following changes in sea-ice cover. In addition to the macronutrients, data for detecting changes or trends in micronutrients, especially Fe, are lacking. These will have a direct impact on the regulation of phytoplankton growth and thus also on the CO_2 sink of the WG.

7.2.2. Reason for Shallow Remineralization and Understanding of Subsequent Nutrient Export From the WG

In the interior WG, a subsurface water layer known as the CIW forms, with elevated nutrient concentrations and depleted oxygen (compared to its source water). Two factors contribute to the shallowness of this core: first, a pronounced shallow degradation of organic material exported from the surface layer and, second, the upwelling of water masses in the interior gyre, pushing the core with elevated nutrients to shallower depths (Usbeck et al., 2002). Owing to the shallow remineralization, the export of particles from the surface to the deep sea is exceptionally low, as observed with sediment traps (Fischer et al., 1988). Thus, the amount of organic matter reaching the sediments is low (except along the margins and around the promontory Maud Rise close to the Prime Meridian; Hulth et al., 1997). In fact, both opal and carbonate are absent in sediments of the deep central WG for reasons not yet understood. The shallow remineralization combined with isopycnal transport of CIW toward the north is a mechanism that very efficiently brings nutrients (just like CO_2) down to the abyssal depths of all ocean basins (Hoppema, 2004b). To date, the reason for the shallow remineralization in the interior WG is not clear. CIW with its elevated nutrients is, therefore, a crucial part of a mechanism with significance beyond the WG. Currently, there are still many unknowns regarding the generation of this nutrient-rich water mass and its temporal and spatial variability within the WG, as well as its far-field effects after export from the WG. Regional models are a promising approach to better understand what controls the distribution of CIW within and beyond the WG.

7.2.3. The Role of Sea Ice on Lateral Redistribution of Nutrients

Sea ice does appear to play a role in nutrient cycling since the concentrations in it are quite different from those in the underlying water (Dieckmann et al., 1991; Papadimitriou et al., 2007). Sea ice is transported within the gyre circulation where it roughly follows the wind-driven cyclonic water flow (see section 5). There is a net transport of sea ice from regions closer to the Antarctic continent, where it is formed, to the north where it subsequently melts (P.R. Holland & Kwok, 2012). During recent decades, this equatorward transport has decreased in the WG (Haumann et al., 2016), pumping less freshwater from south to north. Such processes will also have an influence on the nutrient redistribution in the water column of the WG. We envision that nutrients and micronutrients are taken up from within the sea by ice algae. After melting of the sea ice in more northerly areas and death of the plankton, these biogenic particles are remineralized in a different area than where they originated. The quantitative significance of this mechanism of nutrient transfer is not known.

8. Biology I: Phytoplankton of the WG

Phytoplankton, microscopic primary producers, constitute the “invisible forest” in the sea. The name phytoplankton is derived from the greek planktos, meaning “wanderer” or “drifter,” describing organisms that float in the water; (horizontal) currents and (vertical) mixing control their location in the water column. Phyto refers to organisms that photosynthesize, combining light, carbon dioxide, and minerals to form organic matter, the basis of life in the ocean. Most of the phytoplankton are composed of unicellular algae, which in polar environments, in contrast to temperate and tropical regions, are dominated by eukaryotes, cells with true cellular membranes. Phytoplankton species organize themselves in populations and communities, interacting with their environment and other components of the marine ecosystem. The goals of ecologists are to understand the role of organisms in the ocean as well as their interaction with each other and their environment and the what, when, and how of organisms' distribution in time and space. In the WG, organisms are exposed to an extreme polar climate and the physicochemical properties characteristic of the region. This includes the presence of sea ice and meltwater, high winds, variable availability of nutrients, and extremes in temperature and light. Plankton physiologists explore the ways phytoplankton function and adapt to these extreme conditions. Biologists collaborate with oceanographers to better understand the environmental conditions to which plankton are exposed. For example, there is an active collaboration to understand the response of plankton to an increase in oceanic carbon dioxide, a process termed ocean acidification. Although the ecology of plankton can be studied with autonomous vehicles and remote sensing techniques, findings regarding the physiology of plankton and the interactions between organisms are mainly based on field studies and experiments in the laboratory.

8.1. The State of Knowledge

Single-celled autotrophs and heterotrophs named *protists* populate the water column of the ocean and are collectively referred to as phytoplankton. Whereas autotrophic phytoplankton thrive in the sunlit surface layers where they convert carbon dioxide into biomass, heterotrophic single-celled organisms are found from the sea surface down to the seabed, feeding on existing organic carbon. Protists represent an important part of the food web in the ocean comprising a wide spectrum of species and size classes, from picoplankton (0.2–2 μm) to nanoplankton (2–20 μm) and microplankton (>20 μm; also known as net plankton). The most prominent groups of phytoplankton in the WG are diatoms, haptophytes, autotrophic picoflagellates, and nanoflagellates as well as dinoflagellates (Nöthig et al., 2009). However, some heterotrophic unicellular

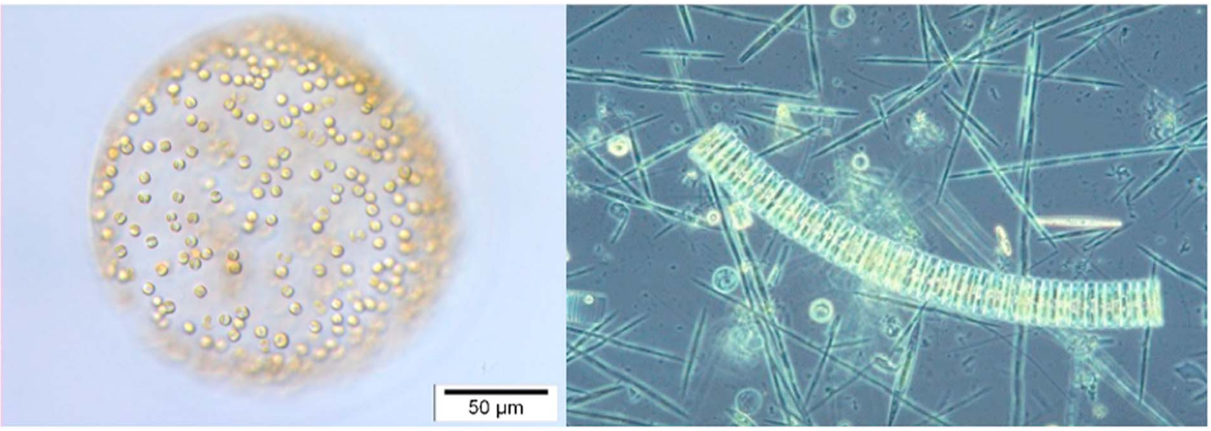

Figure 17. Dominant phytoplankton forms in the Weddell Gyre: (left) *Phaeocystis antarctica* colony composed of several tens of cells within a mucilaginous matrix. Individual cells can acquire flagella and swim as single cells: A colony (200–800 μm) is part of the microplankton, and single cells are nanoplanktonic (4–7 μm; photo courtesy of S. Gäbler-Schwarz, Alfred Wegener Institute). (right) Diatom assemblage of an iron (Fe)-induced bloom in the Southern Ocean (courtesy P. Assmy, Norwegian Polar Institute).

organisms such as ciliates, tintinnids, foraminifera, acantharians, and radiolarians can also photosynthesize with the help of autotrophic endosymbionts or by assimilating their prey's chloroplasts (Hewes et al., 1998; Nöthig et al., 2009).

Microplankton play a critical role in the WG marine ecosystem. Diatoms, the main drivers of phytoplankton productivity in the Southern Ocean, dominate the microplankton size fraction in net hauls, are easily analyzed by microscopy, and are ideal food for large zooplankton (Schmidt & Atkinson, 2016; Figure 17). They have historically been classified in four ecological groupings within ice-free waters of the WG: (1) spring forms, (2) spring forms but occurring throughout the year, (3) spring and autumn bloom forms, and (4) spring forms present in the winter and summer/fall species (Hart, 1934, 1942). More recent genomic surveys have helped unravel the greater diversity in diatoms and underlined their importance as a major photosynthetic group in the Southern Ocean as a whole (Malviya et al., 2016). Another dominant taxon of the net plankton is the colonial form of the haptophyte *Phaeocystis antarctica* (Figure 17; Hart, 1942; Karsten, 1905; Nöthig et al., 1991). *P. antarctica* appears to be of major importance in the entire WG, often dominating blooms, with increases in abundance typically noted after the spring sea-ice melt.

The role of picophytoplankton and nanophytoplankton in the WG is expected to be of equal importance to net plankton, albeit not during high cell concentrations or blooms which are most times dominated by large cells (see above). Contrary to net plankton and as a result of their small size, their occurrence is typically documented by diagnostic pigment markers and flow cytometry (Bidigare et al., 1986; Buma et al., 1992; Detmer & Bathmann, 1997; Peeken, 1997). The distributions of cryptophytes appear restricted to the northwestern WG (Buma et al., 1992; Mendes et al., 2012), while haptophytes, including single-cell *Phaeocystis* sp., are present in both northwestern and northeastern regions (Bracher et al., 1999; Buma et al., 1990, 1992; van Leeuwe et al., 2015; Mendes et al., 2012). In addition to provide identification of flagellates, pigment markers confirmed the high abundance of nanoplanktonic diatoms (Bracher et al., 1999; van Leeuwe et al., 2015; Peeken, 1997). In-depth characterization of the biodiversity of the phytoplankton is still scarce, but results from next-generation sequencing are promising (Wolf et al., 2014). Initial data resulting from these new methods have, for example, highlighted the occurrence of autotrophic dinoflagellates in the northwestern Weddell region (de Vargas et al., 2015).

The spatial and temporal variability of autotrophic phytoplankton communities in the WG system, characterized by their species composition, biomass, and phenology (timing of seasonal blooms), have primarily been studied by shipboard observations and ocean color remote sensing. These techniques provide data in the northern sectors of the WG during summertime when the ice extent and thickness are at their minimum. As a result, the seasonal coverage of sampling shows a bias toward the summer, with only sparse observations available for austral fall and winter. High summer phytoplankton biomass, measured as chlorophyll a (chl *a*), tends to be found in the open ocean (1.5–4 μg chl *a*/L), typically associated with the marginal ice zone (e.g., Bathmann et al., 1997). However, higher concentrations have been reported in the northwestern Weddell coastal polynyas (Cape et al., 2014), where values can reach in excess of 10 μg chl *a*/L. In the northeastern WG/Lazarev Sea region, the highest reported chl *a* concentration was observed during fall (3 μg chl *a*/L; Hayes et al., 1984). Elsewhere in the WG, the few winter studies available show low chl *a* never exceeding 0.08 μg/L (Nöthig, Bathmann et al., 1991; von Harbou et al., 2011) and biomass in the range of 0.3–2 μg C/L (Buma et al., 1992).

Primary production estimates derived from satellite remote sensing show that the WG accounts for ~25% to the Southern Ocean carbon uptake (Arrigo et al., 2008). Relative to the Southern Ocean as whole, the WG has the most productive marginal ice zone with one of the highest area-normalized primary production rates (70.2 g $C{\cdot}m^{-2}{\cdot}year^{-1}$ compared to an average of 57 g $C{\cdot}m^{-2}{\cdot}year^{-1}$ for the Southern Ocean; Arrigo et al., 2008). However, the oceanic waters in the northern WG contribute more to the regional production due to the larger areal extent of ice-free waters (Schlitzer, 2002; Taylor et al., 2013). While elevated, these productivity rates may in fact be underestimates, as remote sensing of ocean color can underestimate productivity in pack ice (Assmy et al., 2017; Geibert et al., 2010) or under thin ice (Arrigo et al., 2012; see section 5.1). Any future changes in sea-ice cover and dynamics could translate into a significant increase in carbon sequestration in the Southern Ocean, as observed in the Arctic (Kahru et al., 2016; see section 5.2.3). Within the WG, productivity is closely linked to species abundance and composition, showing high rates at frontal zones, sea-ice edge, and coastal environments (J. Park et al., 2010) correlating positively with biomass (Dierssen et al., 2000).

8.2. Research Priorities

As in other parts of the Southern Ocean, a primary target for research involves constraining the main drivers of phytoplankton diversity and production in the WG. Low light exposure to phytoplankton due to deep mixing and snow-covered sea ice as well as Fe limitation are prevalent drivers of phytoplankton ecology and physiology in the WG pelagic system. Timing of primary production is determined by relief from light limitation (i.e., greater exposure to light), through the establishment of a shallow mixed layer from sea-ice melt. However, it is the availability of nutrients (in this case Fe or silicate) that ultimately sets the annual extent of primary production (e.g., the yield; Llort et al., 2015). These factors control the distribution and growth rate of phytoplankton species, their abundance, and productivity, as well as differentiate between timing and yield in the seasonal variation of communities. Understanding the relationship of phytoplankton growth and composition to these environmental drivers will shed light on their role in shaping the variability of the biomass and productivity in the WG.

8.2.1. Light and Iron Availability

In a broad sense, primary production in Southern Ocean areas such as the WG is governed by the availability of iron and light (e.g., Boyd et al., 2010; see also sections 6 and 7). The availability of Fe, the principal limiting nutrient in the Southern Ocean, remains a primary factor in controlling productivity in open waters, where nutrient supply is dependent on large-scale fertilization processes, for example, glacier meltwater and suspended sediments act as Fe sources to the western WG (de Jong et al., 2012; see section 7). The relative importance of these two environmental drivers is difficult to disentangle, as low Fe availability and deep mixing regimes, resulting in low average irradiance, often coincide (Boyd et al., 2008). Some studies suggest that blooms are mainly controlled by the low irradiance levels that phytoplankton experience in the prevailing deep mixed layers of the open Southern Ocean (de Baar et al., 2005; van Oijen et al., 2004), although laboratory experiments indicate that irradiance conditions typical of shallow mixed layers can also inhibit phytoplankton growth efficiency (Hoffmann et al., 2008). Underlying these observations is the interaction between light and Fe at the physiological level (Sunda & Huntsman, 1997). Fe has a central role in cellular redox reactions; limiting concentrations of this trace metal alter the architecture of the photosynthetic apparatus, decreasing photoadaptive capabilities and lowering rates of chl *a*-specific primary production (Behrenfeld & Milligan, 2013; Greene et al., 1991). These impairments of photosynthetic processes strongly decrease the ability of phytoplankton to acclimate to low-light conditions experienced by phytoplankton in their environment (Alderkamp et al., 2012; Galbraith et al., 2010).

Given the interplay between nutrient and light availability in the upper ocean, an understanding of mixed layer processes, including factors affecting its depth, is critical to constrain the relative importance of these bottom-up processes. Shallow mixed layers in the marginal ice zone of the WG resulting from sea-ice melt, contributing to an increase in the average irradiance experienced by phytoplankton, are the proximate cause of large phytoplankton blooms occurring in this area (W. Smith & Nelson, 1990). Conversely, deepening of the mixed layer has been shown to result in suboptimal irradiance (van Oijen et al., 2004), leading to depressed photosynthetic rates. However, the relationship between primary productivity and mixed layer depths is far from linear. Cells mixed up to the surface may experience photoinhibition, for example, a physiological response contributing to decreased photosynthetic efficiency (Alderkamp et al., 2011; Cheah et al., 2017), as a result of excessive light availability. On the other hand, a deepening of the mixed layer can also entrain Fe-rich waters from depth and therefore increase productivity during Fe-limited conditions (de Jong et al., 2012; Tagliabue et al., 2014). Adding even more complexity, deeper mixed layers could also relieve grazing pressure on phytoplankton by diluting the microzooplankton grazers (Behrenfeld, 2010). The complexity between phytoplankton and the mixed layer can be further enhanced if, during erosion of the mixed layer, deep phytoplankton cells are injected back into well-illuminated waters, either by destabilizing a subsurface chl *a* maximum (Huisman et al., 2006) or bringing up large cells that accumulate at the top of the pycnocline (Cefarelli et al., 2011). Strass et al. (2002) have shown that mixed layer depth variability (and hence light limitation) and nutrient supply by upwelling, as well as the vertical and horizontal advection of plankton, can all be affected and linked to each other through mesoscale dynamics. Whether this is restricted to the fronts and eddies of the ACC is unresolved. Thus, nonlinear relationships between mixed-layer depth variability and primary productivity (Vernet et al., 2008) need further exploration in the WG.

In order to resolve this, the relative importance of light and Fe limitation in the Southern Ocean needs to be understood. While the role of iron in the WG marine ecosystem remains poorly characterized, this topic has been the subject of significant research interest over the last decades in adjacent regions. In an in situ Fe fertilization experiment in the Antarctic Polar Frontal Zone, Smetacek et al. (2012) showed that massive phytoplankton blooms can develop in a mixed layer as deep as 100 m. In naturally occurring blooms in this region, patterns of primary production did not correlate negatively with mixed-layer depth but instead appeared to be controlled by Fe supply earlier in the growing season (Hoppe et al., 2017). Provided Fe allows for bloom development, primary production was positively correlated with increases in depth-integrated biomass, the latter of which was not impaired by deep mixing but instead increased with increasing mixed-layer depth (Hoppe et al., 2017). There is growing evidence that Fe availability and its effects on the process of phytoplankton photoacclimation, rather than light limitation per se, may control primary productivity highlighting the need for a greater understanding of Fe pathways, sources, and sinks. In this vein, a more nuanced understanding of the role of icebergs in iron cycling is needed, given their potentially large-scale contribution to phytoplankton growth and biogeochemical fluxes in the Southern Ocean (Strass et al., 2017).

8.2.2. Biological Drivers

Our present understanding of phytoplankton diversity and productivity in the WG, and Southern Ocean in general, is based upon bottom-up rather than top-down processes, in spite of the importance of grazing in the polar carbon cycle (le Quéré et al., 2016; Smetacek et al., 2012). Field experiments indicate that viral lysis of phytoplankton (both picoplankton and nanoplankton) can be relatively minor in the WG, while grazing by microzooplankton exerts significant control of phytoplankton growth (Agustí & Duarte, 2013; Evans & Brussaard, 2012; Froneman & Perissinotto, 1996; Henjes et al., 2007; Jacobsen et al., 2007). In contrast, larger zooplankton are of secondary relevance in controlling phytoplankton biomass (Garcia et al., 2016). Although the grazing impact of both krill and salps on phytoplankton is moderate through most of the Southern Ocean, it should be noted that salps can in some instances control bloom formation (Perissinotto & Pakhomov, 1998) and that the impact of krill grazing may be underestimated due to the difficulty of measuring krill abundance and distribution (Smetacek et al., 2004).

Primary productivity is affected by community composition, an important determinant of growth and losses of phytoplankton standing stocks. The relationship between species diversity and productivity of phytoplankton in the WG is expected to be nonlinear with maximum diversity at intermediate productivity (Rosenzweig, 1995). Low diversity is associated with low productivity, increasing as productivity increases. Zooplankton grazing on the most abundant species is the main determinant of increased diversity under these conditions of lower productivity. Once diversity has reached a maximum, further increases in productivity are associated with decreasing diversity. High productivity and low diversity are found in the absence of grazers, as observed in high-latitude spring blooms where few bloom-forming phytoplankton species dominate (Vallina et al., 2014). If grazing increases diversity, widespread grazing by krill and salps as well as microzooplankton could be an important agent to maintain phytoplankton diversity in the WG (Cermeño et al., 2013; Ross et al., 2000). Existing estimates on diversity that rely only on microscopy and pigment analysis are inadequate to test this hypothesis at the ecosystem level. A more complete suite of taxonomic tools, including molecular techniques, will enhance diversity estimates and complete our understanding of picoplankton and flagellate diversity (Wolf et al., 2014) and its relationship to productivity.

The WG is an ideal location to test the role of higher trophic levels in exerting control on phytoplankton diversity and productivity (see section 9.2.2). It has been suggested that whaling in the early twentieth century decreased predation on krill (Willis, 2007); such a reduction of grazers could have facilitated increases in phytoplankton biomass. Additionally, it has been proposed that whales, when feeding near the surface, can inject Fe to the mixed layer through whale defecation, a small-scale process similar to the “storm effect” (i.e., biomixing) of krill diel vertical migration (DVM; Nicol et al., 2010). Measurements of the solubility of Fe from whale fecal material were recently reported and suggest a significant contribution to the available Fe pool (Ratnarajah et al., 2017). In addition, whale seasonal migration to lower latitudes in winter could be a critical vehicle to export organic matter from the WG, increasing connectivity to lower latitudes.

8.2.3. Phytoplankton Distribution in Relation to Sea Ice

The distribution pattern of phytoplankton biomass, expressed as chl *a*, and its interannual variability are tightly linked to sea ice, highlighting the importance of the receding ice edge for the ecosystem in the WG (Figure 18). During spring and summer, sea-ice melt contributes to the freshening of surface waters as a result of melt, increasing the stability of the water column and decreasing mixed layer depth. Sea-ice melt also releases Fe to the upper ocean, with previous studies showing two orders of magnitude higher Fe concentrations in sea ice than the underlying water column (Lannuzel et al., 2007). Fe originally accumulates in sea ice via scavenging of particles by ice crystals during sea-ice formation (Lancelot et al., 2009), with Fe distributions and concentrations expected to be modified by cycling within the sea-ice matrix. In the western Weddell Sea, sea ice has been estimated to provide an additional 7 nmol/L Fe to the mixed layer, in comparison to an average 0.34 nmol/L in surface waters of the WG (Klunder et al., 2011; Lannuzel et al., 2008). Dynamic redistribution of sea ice by winds suggests melting sea ice could have a role in fertilizing blooms anywhere in the WG (Geibert et al., 2010). Concentrations of exopolymeric substances originating for phytoplankton or sloppy feeding, which enhance the bioavailability of Fe, are in high concentrations within Antarctic sea ice. As a result, the exopolymeric substances are also likely to impact phytoplankton production when released to the water column during sea-ice retreat (Hassler et al., 2015; van der Merwe et al., 2009). Moreover, studies from the Arctic provide an alternative or supplementary mechanism on Fe enrichment by sea ice, where particulate Fe oxides within the sea-ice matrix are converted to dissolved bioavailable Fe by ultraviolet radiation (Kim et al., 2010). Due to the colimitation of light and Fe to primary production (see section 8.2.1), the study of new natural sources of Fe is of the utmost importance for both average as well as regional primary production in the WG.

8.2.4. How Will Productivity in the WG Respond to Ocean Acidification and Other Climate Change Drivers?

The high solubility of CO_2 in cold waters, in combination with upwelling of CO_2-rich water masses from depth, contributing to a decline of seawater pH and carbonate saturation state, making the WG waters prone to ocean acidification (Pörtner et al., 2014). However, laboratory- and field-based experiments with phytoplankton species from the Southern Ocean have also revealed potential beneficial effects of CO_2 for Southern Ocean primary production. This effect is thought to be mediated by substrate limitation (e.g., Rubisco) of photosynthetic carbon fixation (Riebesell et al., 1993; Tortell et al., 2008; Trimborn et al., 2013). The carbon concentrating mechanisms' efficiency and flexibility differ between phytoplankton species and groups, for example, between Southern Ocean diatoms and *Phaeocystis antarctica* (Trimborn et al., 2013); therefore, ocean acidification can be expected to affect species competition in their natural environment. In incubation experiments with natural phytoplankton assemblages from the Weddell Sea, floristic shifts within the diatom assemblage have been observed under increasing pCO_2 levels, which can have far reaching consequences for biogeochemistry; however, under the applied ocean acidification scenario, the diatom *Chaetoceros* sp. increased its relative abundance by 50% at 800 μatm compared to 380

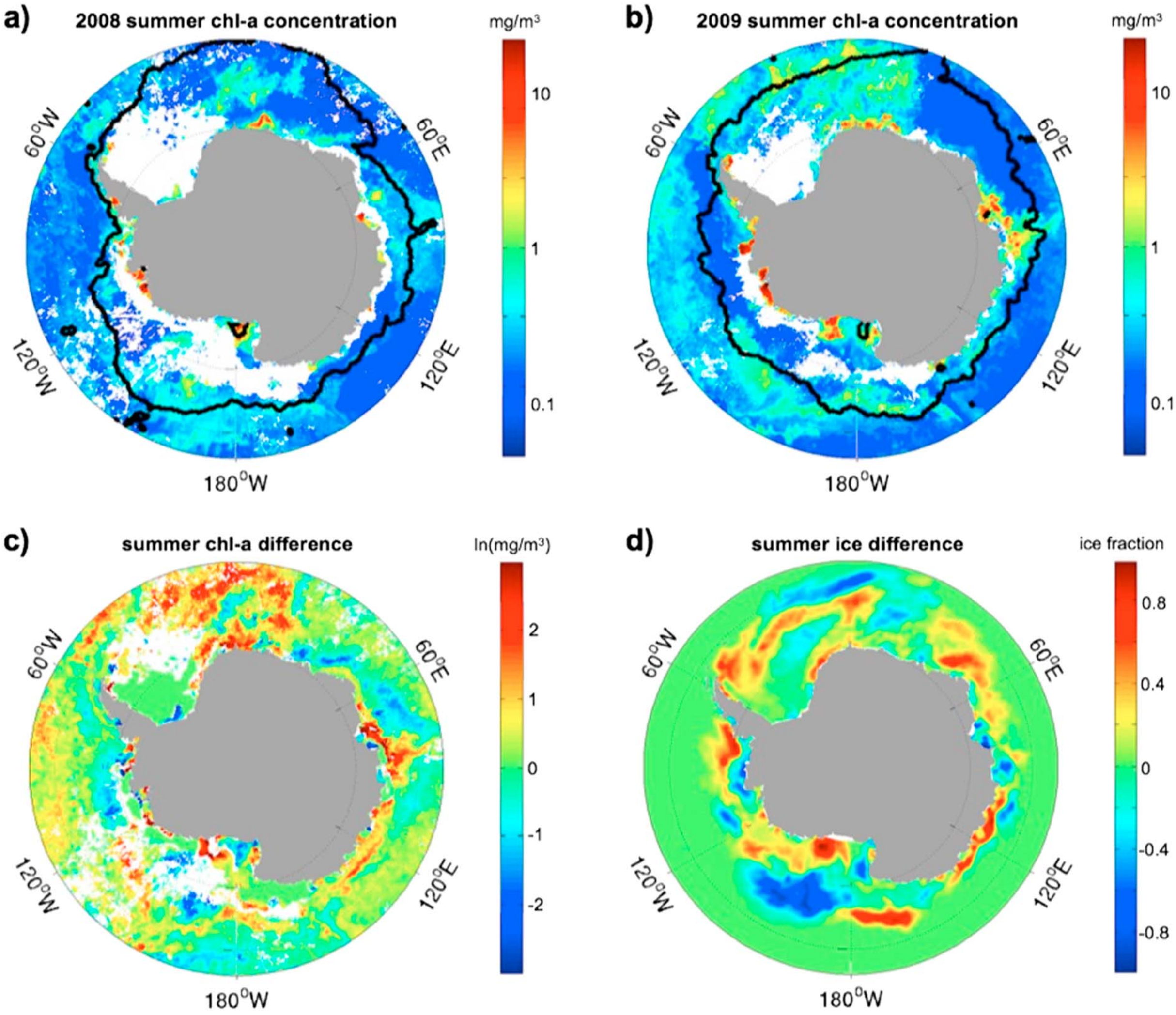

Figure 18. Relationship of phytoplankton to sea ice in the Southern Ocean, using 2008 and 2009 as an example: (a and b) chl *a* concentration in log scale (indicated by the color scale) with maximum sea-ice extension (black line); (c and d) anomalies of chl *a* and sea-ice concentration. Note how changes in chl *a* concentration at the edges of the Weddell Gyre from one year to the next relate to changes in sea-ice concentration. Note that in the eastern Weddell Gyre, summer sea-ice difference (d) shows an area of sea-ice decrease (blue) surrounding sea-ice increase (red) that correlates with higher chl *a* (c).

μatm CO_2 (Hoppe et al., 2013). A shift toward larger and more strongly silicified species may also significantly impact grazing success and food-web dynamics, as it is mainly larger diatoms and *Phaeocystis* colonies that escape the grazing pressure of Southern Ocean zooplankton. More strongly silicified frustules also allow survival of diatoms after copepod gut passages, affecting vertical flux of carbon to the seafloor (Assmy et al., 2013; see also section 6.2.1).

Reliable projections on the future effects of ocean acidification on primary production depend critically on an understanding of the integrated and interactive effects of multiple environmental drivers, including changes in CO_2, temperature, Fe input, melt-water supply, and light regimes (Boyd et al., 2010; Rost et al., 2008). For example, the increased productivity and shifts in species composition described above were not observed under Fe-limiting conditions (Hoppe et al., 2013). Ocean acidification may decrease Fe bioavailability, which in turn could lead to more severe Fe limitation of phytoplankton under high pCO_2 levels (Hoppe et al., 2013; Sugie et al., 2013). Laboratory competition studies (Xu et al., 2014) showed that future conditions would likely favor growth of the haptophyte *P. antarctica* over the diatom *F. cylindrus*, leading to potential ecosystem changes in the wider WG. For highly dynamic light regimes, as imposed by the deep mixed layers in the Weddell Sea, species-specific differences in the balance between these two processes may determine future species abundances. In an Antarctic strain of the diatom *Chaetoceros debilis*, for example, dynamic light fields mimicking a coastal spring mixed layer scenario led to a drastic decline in productivity under ocean acidification, while acidification or dynamic light in isolation had a much smaller impact (Hoppe et al., 2015). Thus, the combined effects of different environmental drivers on species competition and overall productivity need to be investigated to improve our understanding of climate change effects on WG productivity.

8.2.5. How to Address the Challenges in Biology?

Climate models and long-term data sets indicate that polar waters are particularly prone to global change. As in the entire Southern Ocean, expected changes in the WG for the coming century include warming, decline in sea-ice cover, and concurrent freshening of surface waters as well as a shoaling of the upper mixed layer, despite increasing westerly winds which may increase coastal upwelling (Pörtner et al., 2014; Sallée et al., 2013; Stammerjohn et al., 2012). Shallow mixed layers and increased stratification may reduce the input of nutrients such as Fe from deeper water, while increasing average irradiance levels experienced by phytoplankton. However, the recent increase in macronutrient concentrations in surface waters of the WG suggests increased upwelling within the WG (Hoppema et al., 2015; see section 6.1), a process which may ultimately offset the negative consequences of increased stratification on primary production.

In order to assess the impact of climate change on protist diversity and productivity of the WG, we require a profound and wide-ranging baseline knowledge of the protist community and distribution. Present understanding is restricted to ice-free locations and temporally restricted from late spring to early fall due to inaccessibility by ships. The scarcity of data precludes any examination of species change or climatic trends. Comparison of historic and recent studies of phytoplankton abundance, showing similar concentration, suggests no real changes over the larger WG domain (e.g., Nöthig et al., 1991; Olguín & Alder, 2011). However, because very little is known for the large region encompassing the permanently ice-covered regions of the south and central WG, this conclusion needs to be taken with caution.

Further research on the effect of multiple environmental drivers, both with natural phytoplankton assemblages and detailed physiological studies on single species, is necessary to predict WG phytoplankton responses with confidence. Most importantly, the interactive effects of ocean acidification, Fe limitation, and (dynamic) light should be investigated, even though such multifactorial experiments remain particularly challenging in remote areas such as the WG (cf. Feng et al., 2010).

For a more complete understanding of phytoplankton productivity and distribution, we suggest the following approaches:

1. Extend the sampling to include fall and winter and to under-ice environments with year-round mooring or ice-tethered buoy deployments, even in permanently ice-covered areas. In order to disentangle the relative importance of bottom-up and top-down controls and to reveal possible climatic changes, time series measurements are needed (Cisewski & Strass, 2016).
2. Sediment traps and other autonomous water samplers can collect ice algae and phytoplankton as sinking cells, marine snow and zooplankton fecal pellets (Turner, 2015). These platforms could potentially also be equipped with water sampling devices, camera systems, fluorescence, and nutrient sensors, as well as other probes which would provide a better understanding of mixed layer and upper water column processes across seasons. Combined with ^{234}Thorium measurements (see section 7), we can obtain sedimentation and remineralization of sinking phytoplankton carbon.
3. To further the understanding of the drivers of phytoplankton diversity and abundance, more comprehensive and multidisciplinary process studies are needed that take into account growth, as well as losses, and consider possibly all compartments within the pelagic system. Such deepened understanding of processes and their relative roles is a prerequisite for the identification of essential ecosystem variables (Constable et al., 2014).
4. Remote sensing techniques that can detect main phytoplankton groups, integrated with discrete shipboard studies, should be explored for targeted locations of the WG to extend our understanding of community composition and its evolution across the WG from October to March (Bracher et al., 2009; Soppa et al., 2014).
5. A more widespread application of molecular biological tools is needed to understand species selection in relationship to the environment and to their inhabitants.

6. As field measurements will always be restricted by logistics (cruises) or the method's limitations (e.g., remote sensing), modeling can provide additional understanding and help plan more targeted field sampling (e.g., Borrione & Schlitzer, 2013; Schlitzer, 2002).

9. Biology II: Food Web of the WG

Food webs describe the connectivity among organisms that share a common space and the transfer of matter and energy from one organism to another. A food web is represented by arrows pointing toward the organism that consumes or feeds on another one. When food is consumed or digested, several processes are present in a food web including consumption, remineralization, and recycling. Organisms are classified as grazers when they feed on plants, predators by consuming other animals, scavengers when feeding on dead prey. In this way, a herbivore only consumes plant material, carnivores only consume other animals, and an omnivores feed on both plants and animals, including dead matter. A detritivore remineralizes, or converts dead organic matter, back to its inorganic constituents. In Antarctica, and in particularly the WG, different food webs are expected to be associated with different ice and water conditions, nearshore and open ocean communities, and in particularly periods of 24-hr light in summer or complete darkness in winter. As species composition and abundance experience transitions during the yearly cycles of light, temperature, and sea ice, the volume and type of matter vary, facilitating or hindering different species associations.

9.1. The State of Knowledge

The role of biological factors in cycling organic carbon, and its eventual sedimentation below the productive surface layer, is primarily thought to depend on the size distribution of cells in phytoplankton communities, albeit modulated by selective zooplankton grazing (Barton et al., 2013; Smetacek et al., 2004). The relative importance as well as spatial and temporal dynamics of the microbial and classical food webs to WG productivity has not been quantified (Froneman et al., 2004). The "classical" food web is a short carbon transfer from larger diatoms to krill in turn supporting whales, seals, and penguins (El-Sayed, 1971; Hart, 1942; Figure 19). This type of planktonic community, observed in the spring bloom and in productive regions, is believed to be critical to overall regional productivity. In contrast, the "microbial" food web is based on small autotrophic flagellates such as solitary cells of *Phaeocystis* sp. Protists channel a large proportion of the phytoplankton carbon to higher trophic levels or recycle the carbon in the upper ocean (Smetacek & Nicol, 2005; see section 6.1). This type of food web may occur in areas of low productivity, for example, before and after the spring/summer blooms and during winter (Tang et al., 2009). Snapshots of these communities have been observed in summer regimes in the Weddell Sea (Nöthig et al., 1991).

Phytoplankton is at the base of the WG food web. Composition, abundance, and productivity are discussed in section 8. Here we present the heterotrophs, or consumer organisms, in the system who feed on autotrophic phytoplankton, DOC, and bacteria as well as other zooplankton and fishes.

9.1.1. Bacteria

The global distribution of major marine heterotrophic bacterioplankton is determined by oceanic water masses and controlled by their environmental and biogeochemical properties (Selje et al., 2004). Abundance and composition of bacterial communities in polar waters change in response to phytoplankton abundance, from pre-bloom to bloom conditions, from $0.3 \pm 0.06 \times 10^6$ to $0.8 \pm 0.3 \times 10^6$ cells per ml (Topping et al., 2006). On average, bacteria abundances are low when light is high, equal to 10–20% of phytoplankton biomass. Bacteria biomass to chl *a* and bacterial production to primary production relationships in the Southern Ocean are different from elsewhere in the world's oceans: Bacterial Production is only 0–10% of primary production due to tight control of bacteria by their predators, as protist grazers sustain faster growth rates than bacteria in cold waters (Duarte et al., 2005). However, heterotrophic bacteria become more abundant in the dark, at depth, during the winter and beneath sea ice and ice shelves (Delille, 2004). In austral winter, bacterial biomass can reach an integrated value of 1 g C/m^2, higher than phytoplankton biomass. The microbial loop can thus be a substantial subsidy to upper trophic levels, recovering carbon that would otherwise be lost from the system (Delille, 2004; Figure 19).

9.1.2. Zooplankton

Zooplankton are planktonic heterotrophs usually classified by their size: microzooplankton (size range 20 to 200 μm; Pasternak et al., 2008), composed of unicellular protists; mesozooplankton (0.2 mm to 2 cm) mainly composed of copepods; and the macrozooplankton (>2 cm) mostly represented by the Antarctic krill,

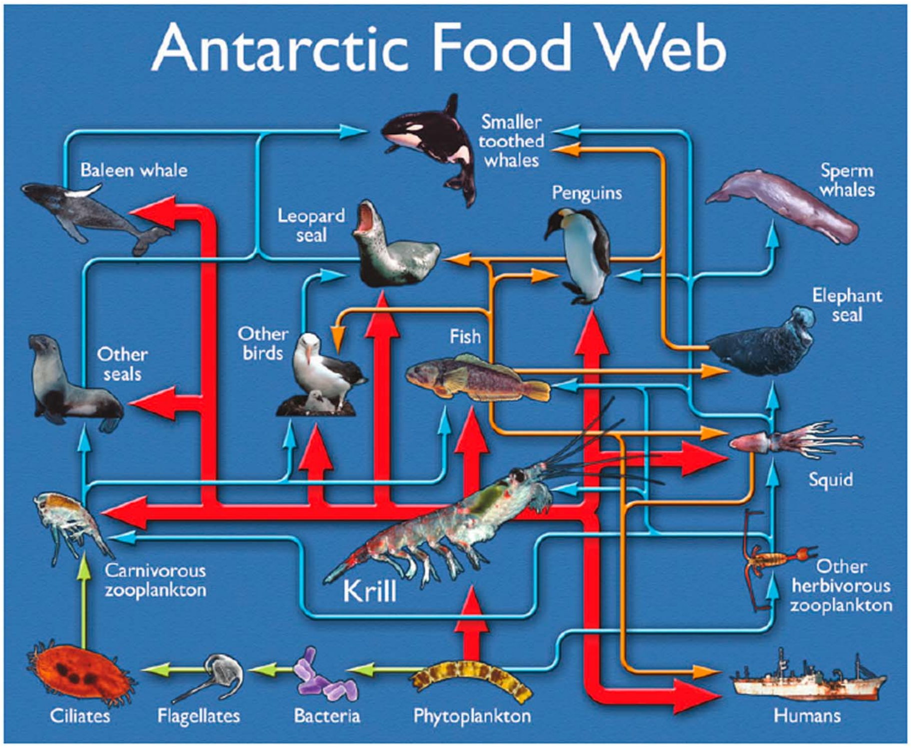

Figure 19. Antarctic Food Web describing short “classic” (red arrows), “microbial” (light-green arrows), other herbivore zooplankton (i.e., copepods, blue arrows), and catch by fisheries (orange arrows). Illustration courtesy of the British Antarctic Survey (http://www.photo.antarctica.ac.uk/external/guest/detail/search/10007839/1/8).

Euphausia superba, and salps, *Salpa thompsoni* and *Ilhea racovitzai* (Pakhomov et al., 2011). This classification originates in part from the different gear and instrumentation used to sample the different size classes, such as net mesh sizes, as well as their ecology, owing to the importance of size structure for ecosystem interactions and functioning. Microzooplankton can reach 25% of the planktonic biomass, occasionally exceeding phytoplankton biomass (Nöthig et al., 1991). Composed of heterotrophic protists, the microzooplankton are considered important in shaping the overall planktonic biodiversity and standing stocks in the WG (Buck et al., 1990; Gonzalez, 1993). They are found under conditions dominated by autotrophic flagellates, usually in winter or after the spring/summer bloom. Hence, heterotrophic protists can typically be seen ingesting small diatoms and nanoflagellates (Nöthig & von Bodungen, 1989). Observations to date suggest that bacteria are also a major food source for these unicellular heterotrophs in the Southern Ocean (Delille, 2004). For example, Vaqué et al. (2002) found that ~67% of bacterial production is consumed by heterotrophic protists each day during summer, although variability was very high (0% to >100%). Highest bacterivory was observed in locations with low chlorophyll (chl *a*), low primary production, and higher temperatures, coinciding with high heterotrophic nanoflagellate biomass and high bacteria production. Gast et al. (2014), utilizing culture-based grazing rates from algae, estimated that the summed impact of three mixotrophs on bacterial populations was typically <2% of the overall grazing. However, nanoprotozoa, feeding on bacteria and phototrophic flagellates <5 μm, were found to ingest up to 32–40% of the bacterial and primary production in the field (Becquevort, 1997). Thus, while heterotrophic protists, the smallest type of zooplankton, are found everywhere in the WG, constraining their impact on algae and bacteria has been elusive due to the variability in field observations (Vaqué et al., 2002).

The spatial distribution of polar mesozooplankton, e.g. copepods, is driven primarily by water depth and by sea ice dynamics (Hopkins & Torres, 1989); these environmental drivers affect overall copepod distribution in the water column as well as the ability of species to reproduce and develop (Makabe et al., 2017). Temperature is another factor expected to contribute to copepod niches, while dietary modes are related to chl *a* and wind stress (McGinty et al., 2018). In the Weddell Sea, copepod observations have primarily been collected in January and February, although data exist for all other months, including samples analyzed by gender and development stages (Cornils et al., 2018).

Antarctic krill is most abundant toward the north and northwest WG, where temperatures are low (Atkinson et al., 2017; Siegel, 2016; Figure 8), while salps proliferate in warmer waters north of the WG. However, the relative importance of these two groups of macrozooplankton could be changing. A study analyzing data for the last century showed that as krill densities decreased, salps appear to have increased in the southern part of their range (Atkinson et al., 2004). Since 1980, larger salp abundances have been observed south of 60°S, their natural boundary of distribution (Pakhomov et al., 2002).

E. superba consumes between 5% and 15% of their body carbon daily. Using a feeding basket with a large filtering surface (fine mesh size of 2–3 μm), they efficiently capture nanoplankton and microplankton, particles ranging in size 10–50 μm (Pakhomov et al., 2002). Even though phytoplankton—in particular diatoms—are the preferred food for krill, protozoans and small copepods are also ingested simultaneously and represent an important supplementary food source year-round (Schmidt & Atkinson, 2016; Pakhomov et al., 2002; Figure 19). Krill, like copepods, benefit from eating diatoms, including sea-ice and planktonic species (Cleary et al., 2018; Schmidt et al., 2018). The importance of sea-ice microbial community as food source for WG zooplankton is thought to increase during winter, although this pattern has been challenged by recent data. While some studies suggest that ice algae can provide up to 67% of the zooplankton carbon (Kohlbach et al., 2018), others conclude that pelagic production is more important as a food source and that sea ice mainly provides shelter (Meyer et al., 2017).

The salp *S. thompsoni* densities remain moderate throughout much of the WG south of 60°S, varying on average between 0.1 and 30 individuals per m^2 (Pakhomov et al., 2002). *S. thompsoni* summer ingestion rates and ambient phytoplankton concentrations have been documented within a chl *a* range of 0.2–1.2 mg/L, with results suggesting that salps are unable to graze at particle concentrations >1 mg/L due to clogging of the mucous filtering apparatus (Pakhomov et al., 2002). As nonselective filter feeders, salps retain particles within a wide size spectrum ranging from 1 to 1,000 μm (Fortier et al., 1994). Based on analysis of ingested pigments and DNA analysis, salps are thought to consume primarily flagellates, with an overall lower consumption of diatoms (von Harbou et al., 2011; Metfies et al., 2014). However, this understanding of salp habitat and feeding has also recently been challenged by studies showing that despite nonselective feeding mechanism and cooccurrence in the eastern WG, the two salp species seem to inhabit different niches (Metfies et al., 2014).

Emerging technologies are providing the first year-round insight into zooplankton abundance and ecology. For example, the backscatter signal recorded by moored acoustic Doppler current profilers enables the estimation of zooplankton abundance. Using this technique, Cisewski and Strass (2016) reported highest zooplankton abundance in the WG at the end of summer and lowest abundances during winter, following a seasonal pattern similar to that of phytoplankton. Furthermore, these authors found that DVM, characteristic of zooplankton, persisted throughout the austral night, with annual modulation centered at local noon in winter and summer solstices. The DVM ceased in late spring, potentially as a result of the presence of phytoplankton blooms or because of an overall increase in predator pressure associated with 24-hr daylight. These authors also observed large (sevenfold) interannual and spatial variability in zooplankton abundances correlated with variability in the magnitude and distribution of the spring bloom. Time series measurements such as these are needed to disentangle the relative importance of bottom-up and top-down controls of phytoplankton and ultimately reveal changes resulting from climate change.

9.1.3. Pelagic Fishes

The diversity of fish in Antarctica is low, ~66% less than that in the Arctic. This is mainly attributed to the presence of deep continental shelves and paucity of island groups, which decreases access to the shallow and coastal waters that fish prefer. Although there is low diversity, endemism is 88% of the total fish community, at least threefold higher than in any other isolated marine environment (Eastman, 2005). Fifty percent of the fish species found in Antarctica, and 90% of the individuals are of the suborder Notothenioidei, a group of bony fishes mostly confined to the cold waters of Antarctica (Eastman & DeVries, 1986; Eastman, 1991). This is the largest group represented in the Antarctic pelagic fish. Another important group of pelagic fishes is the lanternfish. There are ~250 species of these myctophids found worldwide, and 17 species reside in Antarctic waters (Christiansen et al., 2018; Griffiths et al., 2011). Data suggest that the nature of the ACC acts to homogenize distant fish populations living in the open ocean (van de Putte et al., 2012). For example, *Electrona antarctica* found around Antarctica reveals communities of high genetic diversity but low differentiation between the WG and other regions. The diversity observed could originate from local evolution, after an initial colonization by this family of fishes or by multiple colonizations in high latitudes bringing ancestral phenotypes. Present-day data are still insufficient to answer these evolutionary questions. Preliminary results indicate that Southern Ocean myctophids originated from three distant subfamilies, suggesting colonization occurred more than once (Christiansen et al., 2018). In contrast, notothenioids are a monophyletic group, principally benthic, that have diversified in Antarctic waters to form an adaptive radiation that includes semipelagic and pelagic species (Eastman, 1991; La Mesa et al., 2004).

Important evolutionary adaptations in notothenioids are noteworthy: The ability to produce compounds with antifreeze properties, or glycoproteins, depresses the freezing point of body fluids (DeVries & Cheng, 2005). Fishes can live with body temperatures ~1 °C lower than ambient temperature. However, the presence of ice crystals underneath sea ice threatens fish subsistence as they can penetrate fish gills, supercooling body fluids and freezing the entire organism. Due to their production of antifreeze proteins, notothenioids can withstand body temperatures of −2.2 °C (Eastman & DeVries, 1986). This antifreeze glucoproteins come with a price: Stabilized ice crystals within notothenioid bodies are hard to melt (Cziko et al., 2014). These fishes have the ability also to create neutral buoyancy without a swim bladder. Lipid sacs located under the skin and between muscle fibers are considered an adaptation to flotation, enabling the fish to float in the pelagic zone without excess energy expenditure (Friedrich & Hagen, 1994). Pelagic species of notothenioids in the Weddell Sea have 47–61% of their body weight as lipids, while mesopelagic species have ~20%, and benthic notothenoids have ~12% (Friedrich & Hagen, 1994).

Fishes are the second most important food source, after krill, for higher trophic predators, in particular the energy-rich pelagic myctophids and the Antarctic silver fish *Pleuragramma antarcticum* (Barrera-Oro, 2002; Schaafsma et al., 2018; Figure 19). Neutrally buoyant *P. antarcticum* constitutes >90% of abundance and biomass of midwater fishes in coastal waters, playing an important role in the food web (Donnelly & Torres, 2008; Hubold, 1985; Hubold & Tomo, 1989). This species feeds on euphausiids, mostly the smaller *Euphausia crystallorophias*, and also copepods (Hubold, 1985). Higher trophic level predators feed on *P. antarcticum*, and they replace *E. superba* as the main prey when the latter is absent (La Mesa et al., 2004; Ponganis & Stockard, 2007).

The pelagic life history of *P. antarcticum* takes advantage of shelf circulation to obtain a circumpolar distribution (Brooks, Caccavo, et al., 2018). Initial studies in the southern Weddell Sea showed that the postlarvae of *P. antarcticum* concentrate over the continental slope and inner shelf depressions reaching densities as high as 88 individuals per m^2 and representing 85–98% of all planktonic fish larvae (Hubold, 1984). While the postlarvae concentrate in a layer between 50- and 100-m depth, the year 1 fish is found deeper, below 200 m, at the depth of the ISW. Postlarvae accumulate in the slope front and eddies by Ekman transport of the surface layer toward the outer edges of the shelf. Ashford et al. (2017) have hypothesized that *P. antarcticum* larvae dispersing from the west coast of the Ross Sea encounter the shelf outflow along the western side of troughs. This jet advects them toward the shelf break and productive areas associated with nutrient-rich Modified CDW. As the larvae grow older, they either mix with the trough inflow, bringing them back toward the inner shelf, or reach the slope front, as observed by Hubold (1984), and become entrained in the westward flow along the Antarctic Slope Current and Front system.

Antarctic toothfish (*Dissostichus mawsoni*) is a native species of notothenioids in the Southern Ocean and, as of 2005, has been commercially exploited as a fishery in the northern edges of the WG at SST less than −0.6 °C (Roberts et al., 2011). This species inhabits the bathyal zone, living as deep as 2,500 m. *D. mawsoni* is a relatively fast-growing species, at least until 10 years of age, and can live up to 35 years, reaching >2 m in length and 100 kg in weight (Horn, 2002). The toothfish is a top predator feeding on other finfishes and cephalopods and is considered an opportunistic carnivore not showing prey selectivity (Roberts et al., 2011). *D. mawsoni* diet changes during fish development as size and habitat change: Juvenile phases on the continental shelf feed mainly on other small notothenioid fishes such as *P. antarcticum*. Krill (*E. superba*), squid, and other pelagic fishes can supplement their diet. As the fish grows and moves to the continental slope (500–1,000 m), their diet is dominated by larger fish. Older toothfish at depths >1,000 m feed on deep-water shrimp and squid and scavenge penguin remains and fishery discard that falls to depth (Roberts et al., 2011, and references therein). Toothfish is a key prey to Weddell seals (*Leptonychotes weddellii*) and orcas (*Orcinus orca*) and compete for smaller fish with Adélie penguins (Brooks, Ainley, et al., 2018). An estimated 52 tons of toothfish could be consumed yearly by the Weddell seals at McMurdo Sound alone, as high as the present-day catch by the fishery at the southern Sandwich Islands (Brooks, Ainley, et al., 2018). In comparison, the Convention for the Conservation of Antarctic Marine Living Resources (CCAMLR)-managed fisheries annual catch is ~20,000 tonnes of Antarctic toothfish (Ainley & Siniff, 2009; Ponganis & Stockard, 2007).

9.1.4. Higher Trophic Levels

9.1.4.1. Seabirds

Most of the birds in the Southern Ocean are seabirds, obtaining their food from the ocean. Of approximately 250 different species of seabirds in the world, 26 are found south of the Polar Front (Knox, 2006). Antarctic seabirds include flying seabirds such as albatrosses and petrels, only breeding on land, as well as skuas, shags, gulls, and terns (Croxall et al., 1984). Penguins are specialized seabirds that are flightless, adapted for life swimming through frigid waters, with four species breeding on the continent proper: Emperors (*Aptenodytes forsteri*), Adélies (*Pygoscelis adeliae*), Chinstraps (*Pygoscelis antarctica*), and Gentoos (*Pygoscelis papua*). Penguins constitute 90% of the avian Antarctic biomass. They depend on the peak of food abundance which coincides with the short period of bird reproduction in the summer. Seabirds typically nest on land. Some can be found at the coast where rocks are exposed, or on islands, congregating in large colonies. The seabirds congregating in coastal margins have important local effects. They can take large quantities of organic matter and nutrients from the ocean, where they feed and carry it to the land, where they nest. Although part of the organic matter, as guano, feathers, or egg shells will be transported back to the ocean with snow melt, the portion staying on land can sustain local terrestrial vegetation, mosses, and lichens. In comparison, few birds are found in the open waters of the WG, with the exception of those feeding at the sea-ice edge (e.g., Adélie penguins and Snow Petrels [*Pagodroma nivea*]).

Of the 46 Emperor penguin colonies found along the Antarctic coastline, about 12 of them are located along the Weddell Sea coastline, including two on the Ronne and Filchner ice shelves. Most of these colonies are found in the eastern coast of the Weddell Sea, three of which were recently discovered using satellite imagery (Fretwell et al., 2012). Two of the four largest Adélie colonies are in the NE tip of the Antarctic Peninsula, in the NW Weddell Sea, with no other Adélie colonies in the region (Borowicz et al., 2018). Antarctic petrel (*Thalassoica antarctica*) and Snow Petrel colonies also become more frequent toward the northeastern coastline (Croxall et al., 1995; van Franeker et al., 1999). An estimated $2.7 \pm 0.5 * 10^6$ individual flying seabirds were encountered in a band 300-km north and 150-km south of the ice edge in the Weddell Sea (van Franeker, 1996).

The main food eaten by seabirds includes crustacea, squid, fish (adults and larvae), or other seabirds' eggs and chicks (Figure 19). Squid is dominant in subantarctic waters and mostly consumed by albatrosses and larger petrels. Plankton in the form of Antarctic krill, *E. superba*, serves as an abundant food source to 55% of all seabirds. For penguins, krill constitutes between 61% and 100% of their diet by weight (Klages, 1989). The smaller krill *E. crystallorophias* is also known as seabird prey; other crustaceans of less importance are copepods and amphipods. Both adult and larval fishes are consumed by seabirds, mostly the silverfish *Pleuragramma antarctica*, abundant close to land and ice shelves. It is estimated that in the SO, penguins consume close to $18.1 * 10^6$ tons of krill (de L. Brooke, 2004; Joiris, 2000). On average, penguins consume more krill biomass than whales and two thirds of seal consumption (Knox, 2006). Stable isotope analysis supports the hypotheses of "classical" Antarctic food webs, or a short number of trophic steps, where a relatively small subset of available prey provides most of the nutrition to seabirds (Rau et al., 1992). The krill, *E. superba*, is thus a key trophic intermediary between phytoplankton production and higher trophic level consumers in the WG.

Antarctic seabirds have developed different feeding modes of targeted prey. Penguins are pursuit divers feeding as deep as ~250 m (e.g. Emperors), while other species only dive to 10–25 m (e.g. Gentoos). Other diving seabirds include the Blue-eyed shags and diving petrels which can dive and swim underwater due to their reduced wing span, allowing them to paddle through the water. However, their dives are usually restricted to the surface few meters of the water column. Antarctic shag (*Phalacrocorax antarctica*) and South Georgia shag (*P. georgianus*) have the deepest and longest dives of Antarctic flying seabirds feeding predominantly on demersal fish, such as Antarctic cod (*Notothenia coriiceps*; Casaux & Barrera-Oro, 2006). Most albatrosses and large petrels take prey from the surface. Finally, most seabirds can be scavengers, but some species are specialists, like Giant Petrels, mostly feeding from seal, penguin, and small seabird carcasses. Skuas, in particular the Brown Skua, are known to prey on penguin eggs and chicks. In this way, scavengers can have an important local impact close to colonies during breeding season, when the chicks need to grow before winter arrival.

In the open ocean, seabird distribution follows water masses (Joiris, 2000), upwelling, and frontal and sea-ice edge regions (Ainley & Jacobs, 1981; Bost et al., 2009; Fraser & Ainley, 1986). In the WG, seabird distribution is characterized by ice thickness and bathymetry (Flores et al., 2008), also by open water, sea-ice edge, and closed pack ice (Joiris, 2000). Although densities of seabirds at the sea-ice edge are similar in the Weddell Sea than in the European Arctic (Greenland, Norwegian, and Barents Seas), there is a higher species diversity in the Arctic. In contrast, the WG pack ice has an order of magnitude higher abundance of Adélie penguins, minke whales, and crabeater seals (Flores et al., 2008; Joiris et al., 2013). Seabirds integrate and reflect the availability of food with highest daily food intake located at the ice edge, followed by the polar open waters and lowest daily food intake at the closed pack ice (Joiris, 2000). At the ice edge, two bird communities overlap, one associated with the pack ice and another associated with the open waters of the sea-ice zone (Fraser & Ainley, 1986). The pack ice community is dominated by Emperor and Adélie penguins and Snow and Antarctic (*Thalassoica antarctica*) petrels. Penguins dominate the avian biomass although petrels are more numerically abundant. North of the ice edge the Southern Fulmar, the Cape Pidgeon, the Wilson's Storm Petrel, and the Mottled Petrel dominate. A hypothesis that birds are more abundant and have a more diverse community and higher biomass at the ice edge, due to increased productivity and access to prey, was proposed by Fraser and Ainley (1986) when they observed low bird abundance at leads and polynyas in the Weddell Sea and close to multiyear ice (Ainley et al., 1986). Sea ice has further effects on adult penguin population survival of ice-obligatory species: Adult survival of Emperor penguins and proportion of breeders of Snow Petrels depend on the presence of ice (Jenouvrier et al., 2005). Any long-term trend in decreasing sea-ice extent could therefore force penguin migration southward (Cimino et al., 2016; Fretwell et al., 2012; see section 5.2.3). Paleoecological data suggest that penguins are more likely to migrate than to adapt to new environmental conditions brought by climate change (Forcada & Trathan, 2009).

9.1.4.2. Marine Mammals

E. superba is a major food source for higher trophic levels and a link between phytoplankton and mesozooplankton and large consumers like marine mammals (Figure 19). Predator and prey distribution and behavior in the WG are tightly linked: Whales and leopard seals songs show a diel pattern coinciding with zooplankton DVM (Menze et al., 2017). Baleen whales, feeding on krill, show migratory patterns that are complex and flexible, presumably adapting to food availability (Thomisch et al., 2016). Leopard seals, inhabiting the pack ice, also consume krill in variable amounts (12–90% of their diet), but they complement their diet with fish, penguins, other seals and cephalopods (Figure 19; Botta et al., 2018). Crabeater seals also inhabit pack ice and consume mostly krill. Fish are another important prey for marine mammals: Fish of the family Nototheniidae comprise 80% of Weddell seals diets, with the Antarctic silverfish *Pleurogramma antarctica* being ~50% of the consumed fish (Daneri et al., 2018).

The top predator in Antarctic food webs is the orca (*Orcinus orca*; Cornejo-Donoso & Antezana, 2008). There are three recognized types of orcas in Antarctic waters, and all of them are found in the WG. Classified by their foraging habitats and movement patterns, the three ecotypes are designated types A–C, with a new type D recently identified for subantarctic waters (Pitman et al., 2011). Type A lives in open water and preys mainly on Antarctic minke whales (Pitman & Durban, 2012). Type B lives in loose pack ice where it preys mainly on seals, particularly Weddell seals. Fish is the only known prey of type C, foraging deep in pack ice and even among leads in fast ice (Andrews et al., 2008). However, little is known about consumption of these top predators and their role in the WG carbon cycling which could be changing, as observed in the Ross Sea (Ainley & Grant, 2012).

Estimating the impact of marine mammals in the WG food web and carbon cycling is dependent on quantitative assessment of their abundance and understanding their distribution. Blue whales (*Balaenoptera musculus intermedia*), fin whales (*Balaenoptera physalus*), and Antarctic minke whales (*Balaenoptera bonaerensis*) are all found in the WG, where minke whales are more abundant in the winter and fin whales in the summer (Menze et al., 2017). Blue whales are present in the Weddell Sea and eastern WG year around (Thomisch et al., 2016). New technology has helped understand whale patterns of distribution in space and time: Blue whale calls are detected by passive acoustic data in austral winter underneath 90% ice cover, suggesting coastal polynyas play a crucial role in providing food and access to air for breathing (Thomisch et al., 2016). Minke whales are often sighted within sea ice where ships cannot survey, leading to potentially significant underestimation of their abundances. Aerial surveys from helicopters reveal substantial numbers of these minke whales concentrated 150 km on either side of the ice edge (Williams et al., 2014). The difficulty in estimating abundances by whales often leads to an underestimation of their role in the polar food webs (e.g., Murphy et al., 2012).

9.2. Research Priorities

9.2.1. What is the Role of the Food Web in the Carbon Cycle of WG?

Focusing on the key species of the Antarctic ecosystem, it is well known that *E. superba* is channeling a large fraction of ingested material into a long-lived (e.g., food web transport-based) carbon pool (sensu Fortier et al., 1994; Figure 19), while salps are more important in sequestering biogenic carbon into the deep ocean, by producing large, fast sinking feces (le Fèvre et al., 1998). Indeed, as previously stated (section 9.1.3), Antarctic krill is a well-known food source for numerous Antarctic top predators in the Southern Ocean, including squid, demersal and mesopelagic fish, flying birds and penguins, seals, and whales (Knox, 2006). These top predators are thought to consume between 166×10^6 and 450×10^6 ton of Antarctic krill annually (Everson & Miller, 1994; Laws, 1985). The role of salps in mediating carbon to top predators is far less known than for krill (Pakhomov et al., 2002). However, a substantial number of Antarctic zooplankton species, including Antarctic krill, birds, and fishes, is known to consume salps in abundance (Pakhomov et al., 2002). It is becoming clear that salps are not a trophic cul-de-sac in the Antarctic ecosystem and may play a more important role in the ecosystem budget than previously believed (Pakhomov et al., 2011). Given their difference in feeding strategies (see section 9.1.2), it is critical to determine the relative importance of these species in channeling carbon to higher trophic levels of the WG food webs.

Krill fecal pellets are known to contribute to 60–85% of the POC in the upper 200 m, with great variability (Belcher et al., 2017). Vertical flux out of surface waters, attributed to krill feces, range from 0.01 to 15.6 $\text{mg C}\cdot\text{m}^{-2}\cdot\text{day}^{-1}$ (Pakhomov et al., 1997, 2002; Ross et al., 1998). While swarming, Antarctic krill contribute a much larger vertical flux, between 44.9 $\text{mg C}\cdot\text{m}^{-2}\cdot\text{day}^{-1}$ and 1.3 $\text{g C}\cdot\text{m}^{-2}\cdot\text{day}^{-1}$ (Clarke et al., 1988; Pakhomov et al., 1997). *S. thompsoni* feces may also account for 0.1–88 $\text{mg C}\cdot\text{m}^{-2}\ \text{day}^{-1}$ (Pakhomov et al., 2002). However, recent studies are challenging the role of macrozooplankton feces to efficiently transfer carbon to the deep sea in the WG. Iversen et al. (2017) showed that while 30% of the salp fecal pellets were collected by sediment traps at 100 m, only 13% of the pellets were caught at 300 m. The authors suggested retention of the carbon in the upper layers of the ocean, due to pellet break up, presumably by zooplankton grazing. Hence, salp fecal pellets could also have a role in recycling organic matter in the upper mesopelagic waters, a process not very well understood (Iversen et al., 2017). Similarly, krill fecal pellets break up in surface waters, which decreases their sinking rate, thereby facilitating recycling of organic matter and elements in the upper water column (Schmidt & Atkinson, 2016).

9.2.2. How Do We Best Model Food Webs in the WG?

Murphy et al. (2012) summarized in their review the type of data and models needed to understand the biological controls of carbon transfer in Antarctica and predict future change: first, a fundamental understanding of the factors influencing structure and function of food webs at multiple scales. Second, the development of a range of mechanistic models that consider links across physical scales, biogeochemical cycles, and feedbacks, and the central role of zooplankton would serve to constrain trophic interactions (e.g., Llort et al., 2015). Third, development of methodologies to test scenarios of the effect of past and future changes across trophic levels would guide prediction of future ecosystem states. These authors conclude that no single model is likely capable of adequately capturing all aspects of ecosystem dynamics required. Instead, a suite of conceptually separate yet complementary modeling approaches (ranging from 1-D to fully coupled global models) is required to provide alternative views of ecosystems and allow comparative analyses of structure and function. Such a multiple-model approach also provides the potential basis for acknowledging and dealing with the inherent complexity and uncertainty in analyses of the structure and function of oceanic ecosystems.

Recently, a new understanding of how plankton in the 2- to 200-μm size range (microplankton) may function as mixotrophs has emerged (D'Alelio et al., 2016). In an ocean where planktonic organisms can be considered mixotrophs instead of either autotrophs or heterotrophs, each cell can either ingest smaller particles as well as photosynthesize. Photosynthesis occurs in autotrophic algae that contain chloroplasts or by heterotrophic protists ingesting and maintaining functional chloroplasts in their bodies, as is the case for dinoflagellates ingesting cryptophytes (Hewes et al., 1998; Nöthig et al., 2009). Models with generalized mixotrophy are providing more realistic estimates of global distribution of primary production and vertical flux than classical ones while also providing better estimates of plankton particle size distribution (Ward & Follows, 2016). A mixotroph benefits in nutrient-limited regions by ingesting nutrients and extra carbon as prey. In the WG, mixotrophy can be proposed as a mechanism to survive during periods of darkness (McKie-Krisberg et al., 2015). This assumption implies that ingested organic carbon can be substituted for photosynthetically derived carbon and that bacterivory constitutes a major contribution to the carbon budget of photosynthetic algae, such as cryptophytes. Implementing mixotrophy in WG food web models could improve our understanding of the role predators have in food web robustness (Jacob et al., 2011), while also serving to constrain uncertainties in carbon export estimates (Schlitzer, 2002; Ward & Follows, 2016).

10. Modeling the WG

The WG is remote, making observations limited. Ocean surface observations from satellite are also limited due to the presence of sea ice. To increase understanding of the observations, we rely on models. An ocean model is a computer program that evolves the ocean state in accordance with the laws of physics. Model components representing ecosystem dynamics are often incorporated as well. Model inputs are the initial conditions and surface boundary conditions, with these boundary conditions describing exchanges between the ocean and the atmosphere, solid earth, or ice. Once the initial and boundary conditions are given, the model solves for the ocean state at future times. The primary challenge in modeling the WG is the need to simulate a large range of time and space scales on which important phenomena are occurring.

Models simulating the ocean are useful for estimating the ocean state and testing hypotheses seeking to explain observed ocean phenomena. Studying the statistics of the models can also help guide observational programs. There are some aspects of the ocean that are extremely difficult to observe, for example, vertical transport, and models informed from properties that are readily observed help illuminate these processes. Finally, ocean models with data assimilation can be used to recreate past ocean states or to forecast future ocean conditions. This section highlights research efforts using models to increase our knowledge of the WG.

Models provide a tool to understand and predict the coupled atmosphere-cryosphere-ecosystem-ocean system. Simultaneously synthesizing the laws governing these components and their interactions is a grand challenge, but it is the most promising way to test competing hypotheses regarding causal mechanisms that give rise to the observed WG. Though the coupled system is a more accurate representation, it is more challenging to interpret and verify than modeling stand-alone components. This problem is exacerbated by the fact that the most uncertain model components involve coupling parameterizations (e.g., representation of momentum or nutrient exchanges). Studies of the WG in Earth system models are just beginning to emerge, with most of these focused on open ocean polynya formation (e.g., Heuzé et al., 2013; Kurtakoti et al., 2018). However, the majority of WG modeling work has utilized models that represent a subset of components. The emphasis in this section is primarily on the physical modeling of the ocean and sea ice as these components have thus far received the most attention.

For climate modeling, a pressing challenge is to determine the substantial exchanges occurring between the atmosphere and the WG. Processes governing these exchanges occur on all scales, and due to a small Rossby radius of deformation, the influences of dynamics on scales less than 10 km are fundamental to regional properties (Chelton et al., 1998). The difference in representations of the gyre from two similar model setups run at different resolutions is apparent in the transport stream function (Figure 20). High resolution is needed to simulate the fluxes between the flows under the Filchner-Ronne Ice Shelf, the continental shelf waters, the interior WG, and the ACC. Polynyas in coastal and open-ocean regions have lasting profound effects on the atmosphere and ocean state, yet they are often inadequately represented in numerical models (Dufour et al., 2017). Challenges to modeling the WG go far deeper, however, than simply an issue of resolution.

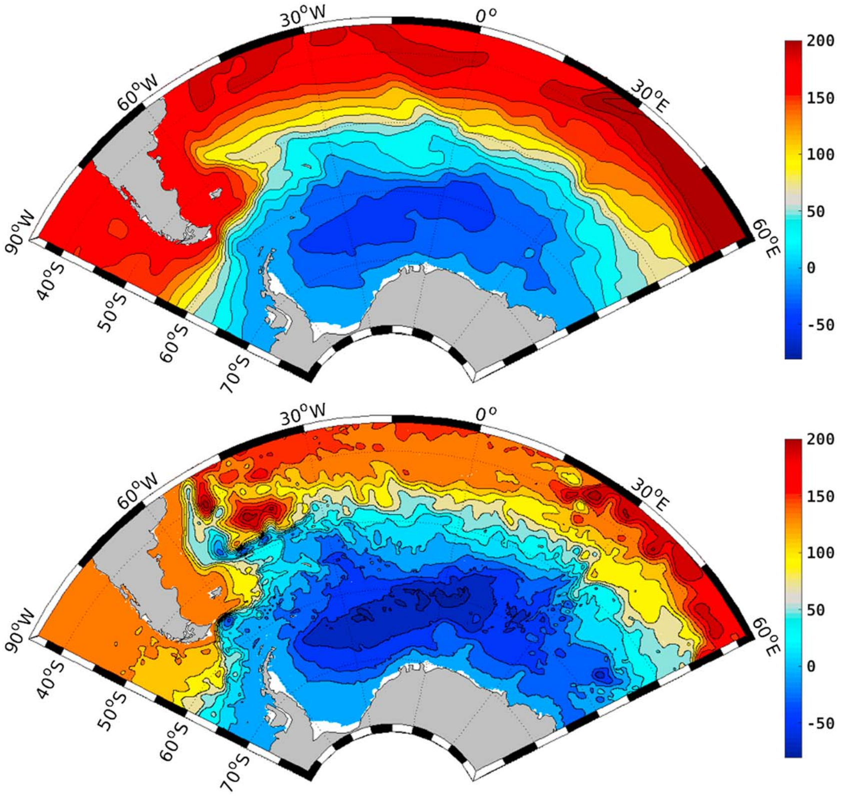

Figure 20. The 2008 mean vertically integrated transport stream function in Sv from a 1/3° run with 52 vertical levels and from a 1/12° run with 104 vertical levels. The contour interval is 20 Sv. The Weddell Gyre circulation is much stronger in the 1/12° simulation, with a 2008 mean transport of about 45 Sv in the 1/3° and about 75 Sv in the 1/12°. Moreover, even in the annual mean it is clear that eddy fluxes between the Weddell Gyre and the Antarctic Circumpolar Current are extremely different. Model output available online (sose.ucsd.edu).

At the surface, the WG is forced by a highly variable and energetic atmosphere that is poorly represented by atmospheric reanalyses, largely due to a paucity of observational constraints. Intense katabatic winds and polar cyclones are often underestimated. The regional hydrological cycle is also poorly quantified (e.g., Hellmer, 2004). Sea-ice transport plays a large role in freshwater distribution (Abernathey et al., 2016; Harder & Fischer, 1999). Kjellsson et al. (2015) also found the strength of the WG in their model to be sensitive to freshwater supply from the Antarctic Ice Sheet, as well as vertical mixing and initial sea-ice conditions. An often over looked component of the hydrological cycle is glacial meltwater transport by icebergs, and this contribution is difficult to model and poorly constrained (e.g., Silva et al., 2006; Tournadre et al., 2012). Moreover, the challenge of modeling iceberg transport projects onto biogeochemistry, as there is evidence that release of terrigenous Fe by melting icebergs fertilizes the surface ocean (Duprat et al., 2016).

Compounding issues of uncertainty in atmospheric reanalysis is the fact that little is known about air-sea exchanges in the presence of a lead-filled and highly dynamic sea-ice cover. Capturing intraocean exchanges is also nontrivial. The northern flank of the WG is bounded by the ACC fronts and subject to intermittently vigorous cross-frontal exchange. Modeling this interaction realistically requires simulating the ACC, leading to difficult questions regarding the model domain size necessary. Exacerbating these issues is the fact that the ocean bathymetry near the Antarctic continent is not accurately known. Since bathymetry has a great influence on all dynamical budgets, and thus on the circulation, 100-m errors are a significant obstacle to attaining model realism.

When modeling the WG physics and biogeochemistry, one must prioritize how to deal with the challenges described above. Global models capture nonlocal effects but can rarely afford to resolve much of the dynamical scales of the WG. Regional models incur errors at the open-ocean boundaries but make higher resolution runs practical. By simplifying the system, idealized models provide understanding of specific processes in the context of the WG environment. Data assimilation can be used to bring the model into consistency with the observations, and when constraints are adequate provides insight regarding errors in the model assumptions.

The challenges to modeling the WG are significant, but the modeling technologies are rapidly evolving. Below we review recent modeling work and discuss insights gained with respect to the circulation and governing dynamics. We organize by approach: idealized, regional, global, and data assimilating. We then look forward and discuss how modeling activities may help address unresolved critical scientific questions.

10.1. The State of Knowledge

Idealized models bring a deeper understanding of the mechanisms involved in WG processes. They employ a conceptual framework to explore the nature of a specific mechanism or to assess its relative importance for the regional dynamics. These studies allow informed decisions to be made in regard to how to parameterize physics that must be omitted due to computational constraints. A recent example is the work by Su et al. (2014), highlighting the importance of mesoscale eddies in setting the WG properties and in modulating the export of AABW through narrow boundary layers along the Antarctic continent. The formation of deep and bottom waters is also sensitive to ice-ocean interactions on the continental shelf, which have been investigated with thermodynamic models of ice growth in leads (Eisen & Kottmeier, 2000; Nicholls et al., 2009). Q. Wang et al. (2009) and Kida (2011) used idealized models to study the overflow of ISWs into the WG through deep channels, demonstrating the impact of local topography (Wang et al., 2009) and of local wind stress, which can modulate AABW transport by changing the position of the density front relative to the shelf break (Kida, 2011).

The sensitivity of the circulation under the Filchner-Ronne Ice Shelf to increased ocean temperatures and reduced sea-ice formation has been investigated using idealized models (Grosfeld et al., 1997; Grosfeld & Gerdes, 1998). More recently, exchanges between the open ocean and the ice-shelf cavities have been explored by Årthun et al. (2013). High-resolution (eddy resolving) idealized and regional models have also been used to study the slope front dynamics and its interactions with the adjacent ice shelves in the eastern Weddell Sea (Hattermann et al., 2014; Nøst et al., 2011; Zhou et al., 2014) and in the Filchner Trough (Daae et al., 2017). Significant questions remain over the role of eddy dynamics in transporting High Salinity Shelf Water under the ice shelves, also the interactions between water formed under melting ice shelves and the open ocean (P.R. Holland et al., 2007; Naveira Garabato, Forryan, et al., 2017). This lack of knowledge is in part owing to the extremely limited set of observations that exist beneath the floating ice shelves, though recent advances in hot water drilling technology has started to address this issue (e.g., through the Filchner Ice Shelf System project, by the British Antarctic Survey).

Idealized models have provided insight on the impact of sea ice on atmosphere-ocean exchange and biological productivity (D.M. Holland, 2001a, 2001b; Jeffery & Hunke, 2014; Lemke et al., 1990; Martinson et al., 1981; Timmermann et al., 1999). Local ecosystem dynamics have also been investigated using idealized mathematical models (e.g., Lancelot et al., 1991; Schalk et al., 1993). There is, however, a wide range of biological model complexity and structure; additional studies that address the relative importance of various ecological mechanisms in the region should be encouraged. Examples include natural Fe fertilization on the continental shelf, competition between phytoplankton types, and food-web interactions, all of which contribute to carbon production and export in the WG.

Ocean General Circulation Models that numerically solve the fundamental fluid dynamics equations can be run in a global or regional domain and can be coupled to atmospheric, sea-ice, shelf-ice, biogeochemical or ecological, or other model components. Though there are many examples of studies that employ General Circulation Models to investigate ice-ocean interactions in the WG, studies that focus on physical-biogeochemical interactions in the region are notably lacking.

Because they can be run at higher spatial resolution than global models, *regional models* are convenient for tackling the challenges of adequately representing processes occurring in the WG region, such as sea-ice dynamics, ice shelf-ocean interactions, and deep convection. Thus, similar to idealized models, regional models are useful to investigate water mass formation (Beckmann et al., 1999; Timmermann & Beckmann, 2004). Circumpolar domains are a common choice to avoid dealing with lateral boundary conditions; the WG properties are impacted by remote forcing particularly via coastally trapped waves and currents propagating along the Antarctic continent (e.g., Schodlok et al., 2001). Circumpolar studies also allow contrasting the slope coastal currents in different Antarctic regions (Mathiot et al., 2011). Studies of

WG processes in Southern Ocean models include those of Harder and Fischer (1999), Beckmann et al. (1999), Timmermann and Beckmann (2004), and Hellmer (2004). Hellmer et al. (2008) analyzed ice-ocean interactions in the shallow shelf areas. Thoma et al. (2006) found the eastern Weddell Ice Shelf region important for the preconditioning of bottom water formation. The study of Matano et al. (2002) addressed the interaction of the WG with the Scotia Sea. The role of tides has also been investigated, and it has been suggested that tidal mixing and tidally rectified flows play a significant role in the circulation and heat fluxes (Flexas et al., 2015; Pereira et al., 2002; Robertson et al., 1998). Variable spatial resolution was used to simulate convective plumes at the continental shelf (van Caspel, Absy, et al., 2015; van Caspel, Schröder, et al., 2015). Circumpolar models with enhanced resolution in the WG have been used by Schodlok et al. (2002) to investigate water mass export, which they find to be sensitive to atmospheric cyclone activity, WG strength, and sea-ice conditions, and by Meccia et al. (2013) to study ice shelf-ocean interactions.

Though *global models* typically cannot afford to resolve the small-scale dynamics that idealized and regional models are able to reproduce, they are valuable in determining the significance of the WG, especially with regard to air-sea fluxes and interbasin exchanges, including ventilation of the abyssal global ocean (e.g., Marinov et al., 2006; Renner et al., 2012; Sarmiento & Orr, 1991; Séférian et al., 2012). Bernardello et al. (2014) emphasized the role of Weddell Sea convection in governing ocean carbon uptake. WG circulation and water mass formation is the focus of the studies by Goosse and Fichefet (1999) and Renner et al. (2009). De Lavergne et al. (2014) used global models to suggest that deep convective overturning in the WG, which is a mechanism through which large polynyas may be maintained, has been weakened due to surface freshening associated with anthropogenic forcing. The model used by Hellmer et al. (2012, 2017) predicts that changes in the WG circulation during the 21st century will bring more CDW under ice shelves, irreversibly accelerating basal melt of the Flichner-Ronne ice shelf. How resolution of small-scale dynamics alters these hypotheses will soon be assessed, as global simulations run at resolutions of 1/12° (e.g., Kerr et al., 2012) and higher (Rocha et al., 2016) are becoming increasingly available.

Data assimilation models have been used to produce hindcasts of the WG. However, the analysis domain of all these products is large. As such, they are run with coarse resolution and often have errors representing processes such as bottom water formation. Dotto et al. (2014) highlight deficiencies in water mass properties in all five of the hindcast products they evaluated. Nevertheless, constraints are impactful, and Hellmer et al. (2005) found that adding them in a global assimilating model impacted the ACC circulation, water mass properties, and the spreading of abyssal waters.

Despite remaining problems in data-constrained WG models, analyses of these products has still proven insightful. Jullion et al. (2014) investigated a box inverse model to diagnose the multiple exchange pathways between the WG and its surroundings. A regional state estimate was produced for the Southern Ocean (Mazloff et al., 2010), and several studies of the WG utilized this resource, including those of Dong (2012), van Sebille et al. (2013), Jullion et al. (2014), Brown et al. (2015), and Meredith et al. (2015). Data synthesis has also been carried out, focused on improving the realism of Southern Ocean biogeochemical cycles (Melbourne-Thomas et al., 2015; Schlitzer, 2002). The Biogeochemical Southern Ocean State Estimate assimilates both physical and biogeochemical observations to produce a hindcast of the carbon, oxygen, nutrients, and physical state (Verdy & Mazloff, 2017). Rosso et al. (2017) analyzed the upper WG area integrated carbon budget in B-SOSE, showing that biological processes and dilution by addition of freshwater acted together to considerably draw down the carbon content in austral spring and summer. Some of this was compensated by an increase in atmospheric CO2 uptake. Carbon content increases during fall and winter partly due to concentration from ice formation. Air-sea CO2 exchanges are negligible in winter due to the insolating effect of sea ice in the model. Meanwhile, circulation transports carbon into the WG year round. Although it is a challenging task, continuing the efforts to develop coupled ocean-ice-biogeochemistry data-assimilating high-resolution models will lead to significant improvements in our ability to simulate the biogeochemistry and water mass evolution of the WG.

10.2. Research Priorities

WG models have provided insight into the regional circulation and dynamics. Questions of fundamental importance remain across a variety of scales. Process models are needed to understand property exchanges at the air-ice-ocean interface. How does better resolving this interface change the simulated biological productivity and the large-scale ocean properties? Interior property exchanges are also of fundamental importance. How do internal ocean processes (e.g., turbulent mixing) affect the properties and transports of dense waters being produced at the margins of the Weddell Sea and their subsequent advection into the interior?

Models allow probing the sensitivity of the system and the testing of hypotheses. For example, assessing the impact of tides may be achieved by running a simulation with and without tidal forcing. This assessment is rarely straightforward, with complications arising due to the nonlinearity of the system. Nevertheless, this highlights the importance of models for capturing the far-reaching and unexpected effects and interactions of components of this complex system. Analogous sensitivity experiments can be carried out to analyze the impact of differing atmospheric state, bathymetric resolution, or choices in model parameterizations. Experience has already shown that the carbon system is highly sensitive to Fe concentrations. Models can put the sparse in situ Fe concentration observations into context by mapping oceanic Fe transport pathways from atmospheric dust sources and from ocean floor sediments. Models can also be used to show where the carbon cycle is most sensitive to these inputs (e.g., Dutkiewicz et al., 2006).

These exercises inform the influence of various dynamics and prioritize observational and modeling development efforts. Critical questions to address using high-resolution models include the following:

1. What are the processes governing the interactions and exchanges between the Filchner and Ronne Ice Shelves, the continental shelf waters, the interior WG, and the ACC?
2. What governs the air-sea exchange of properties and gases in the presence of a seasonally varying ice cover?
3. What physical-biogeochemical-ecological mechanisms control the spatial and temporal variability of organic carbon production and export?
4. How do air-ice-ocean-sea floor exchanges and intergyre exchanges influence global ocean properties and biological productivity?

Modeling and observational analysis efforts are investigating these issues. However, the large-scale observing platforms that oceanographers typically rely on (i.e., Argo floats and satellite measurements) provide limited information in this region due to the complications from the presence of sea ice. Novel observational platforms will have to be exploited in order to constrain and validate models. A recent advancement is the growing “Marine Mammals Exploring the Oceans Pole to Pole” consortium (http://www.meop.net/), which is placing sensors on marine mammals. Another promising development is the deployment of Argo floats equipped with ice avoidance software. These floats are also beginning to carry sensors that measure nitrate, oxygen, pH, and fluorescence, and several have already been deployed in the WG (Briggs et al., 2018; Talley et al., 2019). Adding the ability to acoustically track the location of these floats when they are under the seasonal ice will improve their utility (Chamberlain et al., 2018).

The numerous challenges to modeling the WG are gradually being overcome. As computational resources continue to grow, simulations capable of resolving the processes that govern the WG state in global domains, or at least domains containing adjacent seas, will soon be possible. Meanwhile, WG observations are becoming more abundant. It is likely that high-resolution data-constrained simulations of the WG state, including biogeochemistry, will begin to be produced in the coming years.

11. Synthesis and Outlook

This review identifies challenges for future research in the WG to bring forward the understanding of this vital component of the climate system and be prepared for the expected changes. Unfortunately, for many of these changes, we do not know the exact signature; in many cases, we do not even know the sign, which direction the change will take (positive or negative feedback). The WG has already experienced significant perturbations, although it is difficult to discern whether the observed slow changes are long-term trends or multiyear to decadal variations. Before the expected and probable changes accelerate, we need to understand the system's present state. The WG has also experienced abrupt changes, such as the breakup of ice shelves off the eastern Antarctic Peninsula, creating oceanographic conditions not present for the last 10,000 years (Domack et al., 2005). Understanding processes associated with and developed after this abrupt change can facilitate the construction of future WG scenarios for the next 100 years. These recently changed environments are natural laboratories to study the consequences of abrupt change and provide information on the evolution of Antarctic coastal systems experiencing new modes of water circulation, increased sea-ice conditions, and underwater light availability (Cape et al., 2014, 2015; Gutt et al., 2013).

11.1. Winter and Under-Ice Conditions

A number of challenges are related to sea-ice cover and the season in which the latter is most prominent; this should not come as a surprise for a region that is so cold and remote. A major reason for lack of knowledge and data is the relative inaccessibility of the area in winter and the limited access to remotely sensed properties during the polar night. Winter is characterized by large heat losses to the atmosphere and entrainment of Fe in sea ice by scavenging of ice particles and is a period critical to the survival of microalgae species and multiyear planktonic organisms. Late-winter conditions are the starting point of many processes for the ensuing spring and summer. Not knowing the initial state seriously hampers the interpretation of seasonal process development, which have a large amplitude in the extreme environment of the WG. Only few cruises have been conducted in winter in the region, at a critical time for WG water mass formation near the ice shelves in the south and west (Figure 8). Automation, as in the form of hydro-acoustic moorings combined with under-ice hydro floats, could provide needed communication for winter sampling in this extreme environment.

11.2. Connectivity to Lower Latitudes

Understanding the WG has important implications for the world oceans through advection and transport of Antarctic waters and their characteristics. The newly formed, very dense water masses circulate in the WG for some years to decades; after leaving the gyre confinement, they contribute to the connection of subpolar and temperate oceans via the equatorward flowing AABW. As part of these deep waters, Antarctic characteristics and properties are transferred into the world oceans, first into the abyss and then into shallower ocean layers (during which the properties are modified). Ventilation of the abyssal waters of all oceans is thus affected, but also CO_2 (natural and anthropogenic), nutrients, and DOC and suspended particles, including microorganisms, are transported northward from the WG. Since limited sampling is available during water-mass formation, it is unknown what the physical, chemical, and biological properties of the contributing water masses are at their start, even when some knowledge about water mass fractions within the WG may be gained from data analysis (Kerr et al., 2018). The state of the water mass upon formation is the preformed state, which is needed in calculations of water properties further downstream, for example, the back-calculation of anthropogenic CO_2, or the transit time distribution.

Not only deep and bottom waters are exported from the WG but also surface and subsurface waters. By way of Ekman transport, the surface waters enter the ACC, where part of this water will participate in the formation of Antarctic Intermediate Water. The latter water mass exerts influence on the subsurface water layers of the world oceans in the Southern Hemisphere subtropical gyres (Saenko et al., 2003). Also, the subsurface water of the WG (the so-called CIW) is transported into the deep ocean north of the gyre and contributes significantly to the level of CO_2 in the deep oceans at temperate latitudes (Hoppema, 2004b). Probably, this also holds for many other properties, such as DOC, particles, and nutrients. At the surface, whale and bird migration enhances export of carbon and genetic material from the gyre into lower latitudes beyond passive transport of planktonic species via currents.

11.3. Intensification of the WG

Enhancement of the WG circulation by the increased westerlies will not only magnify existing processes, it will likely increase the connectivity of the system's pools. In this way the effect of changes in circulation, sea ice, and meltwater on the chemistry, biology, exchange with the atmosphere, water mass formation, and sedimentation will cascade through the system at increased rates. The properties of the gyre, such as the exchange with the ACC, the deep upwelling in the center of the gyre, the properties of the subsurface low oxygen, and high nutrients present in the central gyre at bathyal depths, could be enhanced. Two different scenarios could be envisioned, where changes are transferred linearly or exponentially through the system or where the establishment of a different equilibrium condition for the WG is attained. It is not yet possible to assess which scenario will prevail in the future, although a kind of linear change is likely because of many compensating effects (Hauck et al., 2013).

11.4. Future in Modeling

From ecosystem structure to air-ice-sea interaction, numerical models representing physical laws and biological behavior are used to answer scientific questions. The skill of these models is limited not by our understanding of these relationships but primarily by our ability to represent processes across scales from micrometers to thousands of kilometers. We are also limited by our knowledge of the initial conditions of models. The common strategy to overcome obstacles regarding scale is to use process studies to reveal how one can represent small-scale processes in large-scale models. Inverse methods can be used to estimate initial conditions. With growing computational resources and understanding of small-scale dynamics, model skill is growing. WG models now resolve processes in the order of 10 km and include many components of the system. As we move to simulating more and more processes, we reduce model omission error. These models then become more skillful for understanding system sensitivity and predictability. Moreover, these models can then be used to give realistic assessments of the observing system. By sampling these complex models just as the current or planned observing system samples the true WG, we can determine the uncertainty in our inferences. In other words, having a model that represents the true variability of the WG informs how much information our observing system is not capturing. By performing ensembles of this procedure with different proposed observing systems (i.e., observing system simulation experiments), we can derive a WG observing system that provides information to all stakeholders with associated accurate uncertainty estimates at minimal cost. Thus, a primary challenge for the coming decades is to produce models representing dynamics of the complete system at all scales and thus the full range of processes that govern the trajectory of the WG ecosystem and physical states.

11.5. The WG as Sink or Source of CO_2

Physical, chemical, and biological processes impinge on the cycling of inorganic carbon in the WG. Understanding these processes and their quantification is central for estimating the role of the WG as an atmospheric carbon sink or source, which is affected by air-sea exchange and the deep transfer of carbon to lower latitudes by transport along isopycnals (section 6.1). Estimates suggest that the WG transfers into the abyss between 16% and 66% of the Southern Ocean atmospheric carbon south of 55°S (Hoppema, 2004b; Fahrbach et al., 1992; Ito et al., 2010). The biological processes of respiration and bacterial remineralization return carbon to the inorganic pool. The organic carbon fixed by phytoplankton and not remineralized back to CO_2 remains in the gyre as food for fish, birds, and marine mammals (see section 9). However, estimates of carbon fixation in the WG vary by one order of magnitude. Rates of Southern Ocean primary production of 1.95 ± 0.07 and 2–3 Pg C/year by remote sensing (Arrigo et al., 2008, and Westberry et al., 2008, respectively; section 8.1) were estimated, while the summertime C uptake from the atmosphere is 0.7 Pg C/year (e.g., Mikaloff Fletcher et al., 2006). This difference also influences our assessment of the role of the WG in carbon uptake, from 0.477 Pg C/year by ocean color remote sensing to 0.033 ± 0.021 Pg C/year by a WG mass balance (Arrigo et al., 2008; Brown et al., 2015). One should note that not all carbon fixed is replenished by CO_2 from the atmosphere.

Errors in CO_2 estimates originate on the paucity of data for the WG (Figure 10, insert), and the seasonal, temporal, and spatial extrapolations based on short cruises needed to calculate estimates on an annual and WG wide basis (section 6.1). Modeling of primary production is also based on large uncertainties of atmospheric corrections, the climatological data on mixed layer and nutricline depths, the sensitivity of the conversion of chlorophyll estimates to carbon (e.g., the Chl:C ratio), and the importance of phytoplankton photoacclimation affecting the model parameters (Westberry et al., 2008). New space-borne lidars in satellites that can measure chlorophyll in the winter are now available (Behrenfeld et al., 2017), which supports obtaining year-round observations and estimates. The use of model outputs for mixed layer depth and nutricline depths (e.g., www.sose.ucsd.edu) and remote detection of phytoplankton composition (Bracher et al., 2009; Soppa et al., 2014) and cell size (Uitz et al., 2010) will be needed for the next generation of productivity estimates.

11.6. Looking Forward

If our understanding of the structure and function of the WG is to be enhanced, research efforts must be bolstered, and existing projects must find continuation. In addition, community building around the WG should get more attention as a way to pool resources available to the international scientific community. A useful initiative working toward that aim is the Southern Ocean Observing System (www.soos.aq). Generation or continuation of time series can be accomplished in different modalities. Certain lines, for example, the one along the Prime Meridian, that have been sampled since the 1980s should be sampled at intervals pertinent to the scales of critical atmospheric and oceanographic processes. Long-term changes, trends, or variability will thus become visible (Fahrbach et al., 2011; Hellmer et al., 2016; van Heuven et al., 2011). With a similar purpose, a Weddell observatory would be most useful in order to obtain time series in locations identified as characteristic of the gyre or where critical processes are more pronounced, for example, hot spots (e.g., www.soos.aq). Moreover, at these locations, samples could or should be taken on all cruises that visit the region.

Sampling needs to take advantage of new technology that can help bypass existing limitations, such as sampling during winter or under the ice. Moorings with hydrographic instruments and sediment traps are still useful to obtain year-round data even if they only collect data at single locations. Their usefulness will be enhanced if deployed in previously identified regional hot spots. However, they may be combined with moving devices like floats or sea-ice buoys for spatial extrapolation. The use of Autonomous Underwater Vehicles may also constitute a means to be present in locations not accessible by ship. The Autonomous Underwater Vehicles should be equipped with different geophysical, physical, chemical, and biological samplers or sensors (Camilli & Duryea, 2007). Modeling and data assimilation will get a boost when new (winter) data become available, and this will in turn enhance the understanding of the gyre in a synergistic way.

Glossary

Abyssal	Relating to deep waters in the ocean, usually between 3,000 and 6,000 m.
Abyssal plain	Vast, deepest part of seafloor in an ocean basin, which is mostly flat and spans water depths from 4,000 to 6,000 m.
Acoustic data	Information on the structure and shape of the subsurface (here seafloor) obtained by studying the propagation of sound in a medium.
Advection (**Advective pathway**)	Horizontal movement of sea water, or ice, transporting matter, tracers, and heat in the ocean.
Air-sea flux	The movement of gases/heat/freshwater through the interface of the ocean and the atmosphere and vice versa.
Airborne electromagnetic induction sounding	**EM sounding** carried out from an aircraft, for example, a helicopter or an airplane.
Algae	(singular: alga) In the ocean, free-floating unicellular plants, which increase their biomass through photosynthesis and thus provide food in the sea, e.g., phytoplankton.
Antarctic Circumpolar Current	(ACC) – An ocean current flowing clockwise from west to east around the continent of Antarctica. Also called West Wind Drift.
Antarctic Ice Sheet	Largest single mass of land ice on Earth with a size nearly 14×10^6 km^2 and a mean thickness of 2,160 m. It is divided in to the East and West Antarctic Ice Sheets and the Antarctic Peninsula Ice Sheet.
Antarctic Polar Front	This is a front of the Antarctic Circumpolar Current, where cold, fresh Antarctic surface waters from the south subduct beneath warmer, saltier subantarctic waters. At the surface, the front is characterized by strong gradients in temperature, nutrients, and distinct biological communities.
Antarctic Slope Front (ASF)	Well-defined hydrographic front over the Antarctic continental shelf break that separates shelf and open-ocean water masses.
Anthropogenic	Derived from human activities, typically in reference to carbon dioxide (such as the burning of fossil fuels, concrete production, and land use change); man-made.
Autotrophs	Organisms that convert simple inorganic compounds into organic matter. Photoautotrophs use sunlight as energy source, with the aid of the photosynthetic pigment chlorophyll *a*, sunlight, water, and nutrients.
Bacteria	Prokaryotes (cells without true membranes), typically only a few micrometers in length, which have a number of shapes, ranging from spheres to rods and spirals. They appear in large numbers, for example, a million bacterial cells in a milliliter of ocean water.
Basal melting	Ocean induced melting at the underside of the ice shelf.
Bathyal	Pertaining to the deep ocean, usually between 1,000 and 3,000 m deep, in between the continental shelf and the abyss.
Bioavailability	The quantity of substance that can be actively used by organisms for growth.
Biodiversity	Refers to the variety and variability of life forms inhabiting the planet.

Biogeochemical float	Autonomous profiling robots that move around in the oceans and periodically measure the top 2,000 m of the ocean water column by onboard biogeochemical sensors (including temperature, salinity, oxygen, fluorescence, pH, irradiance, and backscatter).
Biogeochemistry	Scientific discipline that studies biological, chemical, geological, and physical processes—and their interactions—that drive the composition and the changes of the natural environment.
Biological drawdown	The action of photosynthesis that reduces the concentration of carbon dioxide in seawater, generating a disequilibrium (undersaturation) with the atmosphere. Refers also to the reduction of nutrients in seawater by phytoplankton and bacteria consumption.
Biological export production	The quantity of organic matter produced in the ocean by primary production that sinks out of the sunlit waters. Phytoplankton cells can sink when they lose buoyancy, or organic matter sinks as fecal pellets from consumers. Organic matter can also be transported to depth by vertical migration of zooplankton and fishes.
Biomass	The mass of living organisms in a particular ecosystem during a specific time. Here we refer to the amount of algae during different seasons in the Weddell Gyre ecosystem.
Bloom/phytoplankton bloom	A large, temporary mass occurrence of phytoplankton that generally appears when there is a surfeit of available light, nutrients, and heat energy (mostly in spring).
Bottom Water	Dense water that is formed by sinking and mixing. It flows (circulates) close to the bottom of the ocean and thus is the densest, heaviest water.
Bottom-up processes	Biological processes subject to bottom-up control, that is, where environmental factors such as light, carbon dioxide, and nutrients control primary production.
Boundary current	Ocean currents with dynamics determined by the presence of a coastline or continental shelf.
Brine channels	When sea water freezes, the freshwater ice is formed leaving behind droplets of salty liquid. These droplets connect to small channels within the sea ice, which can be inhabited by small organisms, for example, ice algae.
Brine rejection	When seawater freezes, the newly formed ice consists almost of freshwater. The excess salt is “rejected” into the remaining seawater and increases its salinity, thereby creating a denser brine.
Buffering capacity	The ability of the ocean to absorb additional carbon dioxide from the atmosphere with only minor ensuing changes in acidity (i.e., pH). The presence of large quantities of bicarbonate and carbonate ions in seawater allow new CO_2 to be absorbed. This ability is only finite, however, and reduces in strength as additional CO_2 is absorbed.
Buoyancy fluxes	Heat or freshwater fluxes at the ocean surface that cause changes in the density of ocean water masses and subsequent water movement.
Buoys	Floating, autonomous instruments equipped with various meteorological, ice, biological, or oceanographic sensors that transmit data by satellite. Within the sea-ice zone, buoys are normally deployed in ice floes. The simplest buoys just record their position by means of Global Positioning System navigation, this providing records of ice drift.
Calving	Ice shelf edge process generating icebergs.
Canyon	Submarine canyons in the ocean are steep valleys cut into the sea floor of the continental slope, from the shelf into the deep ocean.

Carbon concentrating mechanisms	To overcome substrate limitation of carbon fixation in the marine environment, phytoplankton employ mechanisms that increase the concentration of CO_2 inside the cells.
Carbon cycle	Series of processes by which carbon compounds are interconverted and exchanged within different reservoirs of the environment, allowing life to be sustained on Earth. It involves the incorporation of carbon dioxide by photosynthesis into living tissue, its return to the atmosphere through respiration or decay, its sequestration to long-term carbon stores in soils or at the seafloor, and its release from fossil fuels.
Carbonate saturation states	Carbonates are widely used by marine organisms as structural components. The carbonate saturation states describe whether the chemical conditions in seawater such as pH and calcium concentrations favor the formation or dissolution of carbonates.
Cell lysis	When the cell bursts or is broken down by viral, enzymic, or osmotic mechanisms, releasing its contents.
Channel-levee deposits	Channels eroded into the seafloor serve as preferred transport routes for brines and flows of solid particles (e.g., turbidity currents). Due to the Coriolis effect, the particles will be deposited preferably on one particular flank of the channel, where their accumulation through time will build up a large bank (=levee).
Chlorofluorocarbons/ sulfur hexafluoride	Gases released into the atmosphere mainly in the twentieth century, either as refrigerants or as a by-product from the electrical industry. Their rapid increase in the atmosphere and their relative inertness in the ocean make them excellent tracers of ocean water mass age.
Chlorophyll *a*	The main light-harvesting pigment in photosynthesis. Since it is specific to algae and present in all of them, it is often used to determine the algal biomass.
Chloroplasts	Can be considered as the “chemistry factory” of the algae, in which the process of photosynthesis takes place.
Choanoflagellates	Choanoflagellates are ovoid or spherical in shape (~3–10 μm), free-living protozoans, and exist as one cell or in colonies; each of them has one flagellum with a conical transparent food-gathering collar at its base to gather food, for example, bacteria and detritus.
Ciliates and tintinnids	The ciliates are a group of protozoans characterized by the presence of hair-like organelles called cilia, which are identical in structure to eukaryotic flagella, but are in general shorter and present in much larger numbers, with a different undulating pattern than flagella. Tintinnids are ciliate protists that are virtually all restricted to the marine plankton with the common characteristic of having a lorica whose architecture forms the basis of classic taxonomic schemes.
Circulation	Large-scale movement of water in the oceans, also over the continental shelves.
Classical food web	A *food web* describes the pathways of food (as carbon or nitrogen) that cycles from plants to heterotrophic organisms. In the Weddell Sea, the *classical* food web comprises large diatoms being consumed by Antarctic krill (*Euphausia superba*) that feeds whales, penguins, and seals.
Colimiting	“Limiting” is mostly used to convey that a certain nutrient or property (e.g., light) restricts photosynthesis/phytoplankton growth beyond a certain level. Colimiting denotes that an additional factor is also restricting further growth. The term is linked to the specific stoichiometric needs of primary producers.

Community	A collection of species coexisting in space and time within an ecosystem, interacting with the environment and with each other (interspecific interactions).
Continental plate	Large piece of the rigid outer part of the Earth consisting of magmatic, metamorphic and sedimentary rocks that is mainly covered by land.
Continental shelf	describes a relatively shallow area of the seafloor, surrounding continents.
Contourite	Marine sediment type deposited from ocean currents (commonly formed on the lower continental slope and the continental rise).
Copepods	Metazoan zooplankton of the group Crustacea.
Coupled general circulation models	Large-scale computer models of the climate system, with full coupling of the physics between the atmosphere and ocean.
Cretaceous	Geological time period spanning 145 to 66 Ma.
Cryptophytes	A common name for a class of unicellular, microscopically small algae, with a reddish, bluish, or brownish color. The two flagella protruding from a *crypta* (scourges) allow them some small-scale movement in the water.
Cyclonic relative vorticity	Measure of local rotation (clockwise in the Southern Hemisphere) in a fluid flow, defined mathematically as the curl of the velocity vector.
Cyclonic/anticyclonic	Movement of air masses or ocean circulation in the same/opposite sense as the Earth's rotation. In the Southern Hemisphere, this equates to clockwise/anticlockwise motion.
Data assimilation	Data assimilation is thetechnique wherebyobservational data arecombined with output from anumerical model to produce anoptimal estimate of the evolvingstate of the system.
Decay	In the context of radioactivity, the mechanism by which one isotope converts into another isotope, usually of a different chemical element.
Deformation	Because the **drift** of sea ice can be divergent or convergent or can be directed toward the coast, the ice cover deforms by either fracturing and creating open water leads or by crumbling and rafting creating pressure ridges and ice rubble. Deformation causes strong variations in the sea-ice thickness distribution.
Degradation of organic matter	The microbially mediated process in which organic molecules are broken down to smaller molecules, to nutrients, or to their elements.
Density currents	Ocean currents driven by horizontal density differences in the water column.
Depocenter	An area of preferred sediment deposition.
Detritus	Unconsolidated material (such as rock fragments or organic particles) that results directly from disintegration of larger rock units or decaying organisms.
Diagenesis	Modification of sediments by biological, chemical, or physical processes (e.g., compaction), which turns them into different sediments or consolidated rocks.
Diatoms	The most important photosynthetic unicellular organisms of the cold seas. These microscopic organisms have siliceous (= opaline) shells, which consist of two halves and which are inserted into each other like a box (frustules). They can occur in enormous densities (blooms).
Diel vertical migration	Large zooplankton are able to swim up and down the water column, feeding close to the surface at night and moving to depth to avoid predators during the day.
Diffusion	The spreading of a substance or a property that is controlled by the random movement of its individual particles and the resulting statistical distribution.

Dinoflagellates	Mostly unicellular organisms with two long flagella, which run along a longitudinal and a transversal furrow. They are naked or bear a sort of shell (surrounded by cellulose plates). Several modes of nutrition have been observed among heterotrophic dinoflagellates: engulfment of prey, pallium feeding, and peduncle feeding.
Disequilibrium	A state when the concentration of a property in two reservoirs is not equal. For carbon dioxide, a disequilibrium that exists between the ocean and atmosphere will result in the flux of CO_2 between them until they return to equilibrium.
Dissolved inorganic carbon	The combined concentrations of dissolved carbon dioxide gas, hydrogen carbonate (bicarbonate), and carbonate ions in seawater, also known as total carbonate or total CO_2.
Divergence	Occurs at the surface of the ocean where two currents split, creating upwelling.
Downwelling	Sinking of surface waters to greater depths.
Drift	The lateral motion of sea ice as a result of winds, currents, and other forces.
Drivers	An environmental or biological factor that influences organisms or biological processes.
Dynamical scales	The distances over which oceanic motions are happening.
Earthquake	Natural consequences of the moving Earth's plates, which create seismic shock waves traveling through the entire Earth.
Ecosystem	A biological community of living and interacting organisms and their physical environment.
Eddy	Rotational flow of the ocean when the fluid is in a turbulent regime. Mesoscale eddies typically have space scales of 10–500 km and time scales of 10–100 days.
Eddy kinetic energy	Energy associated with the turbulent part of the motion within a fluid.
Electromagnetic (EM) sounding	A geophysical method to measure the thickness of sea ice. EM sounding is sensitive to the electrical conductivity of the underground and can accurately sense the distance to the interface between the resistive ice and conductive sea water, that is, to the bottom of the ice.
Endosymbionts	Autotrophic unicellular organism that live inside another heterotrophic organism. They provide food for the host organism by sharing photosynthetic produced carbon. In return, they are protected inside the host against grazers.
Eocene	Geological epoch spanning 56.0 to 33.9 Ma.
Eukaryotes	Organisms with a real nucleus; they also show a high degree of compartmentalization in the cell (e.g., organelles). Eukaryotes can be single celled (e.g., protists) but also multicellular organisms such as plants and animals.
Eustatic sea-level	Height of the global ocean, which changes in response to a change in the volume of seawater (e.g., caused by the formation/melting of an ice sheet on land) and/or in response to a change in the volume of ocean basins.
Extracellular release	Process by which organisms excrete dissolved organic matter into their surroundings.
Fast ice	Also called landfast ice; sea ice that is attached to the coast, an ice shelf or grounded icebergs and therefore does not move.
Feedback	This is the situation where the output or outcome of a process influences the input. Feedback can be positive or negative. In the first case, the process reinforces itself; in the second case, the outcome of the process leads to a diminution of the original input.

Flagellates	Organisms that can move with the help of one or more flagella (singular: flagellum). Nanoflagellates are less than 20 μm in size comprising a highly diverse group of unicellular autotrophs and heterotrophs.
Flooding	The flow of seawater onto the ice and into the lower snow layers as a result of **negative freeboard**. The resulting water saturated snow is called slush. During the winter, this slush refreezes to form snow ice.
Flow cytometry	A method to obtain abundance, type, and cellular characteristics of very small algae (less than 20 μm), which cannot be identified with an optical microscope.
Foraminifera, acantharians, and radiolarians	Amoeboid organisms that produce pseudopodia, thin extensions of the cell membrane and plasma forming a radial sticky web used for locomotion, floating, and feeding. All groups consist of shell bearing organisms: Foraminifera produce calcium carbonate shells; radiolarian skeletons are made of silica and acantharians of strontium sulfate. Many of these organisms do not ingest their food but digest it outside the body. Foraminifera, acantharians, and radiolarians also host symbionts with chloroplasts.
Freeboard	Freeboard is the height of the ice surface above the water level.
Front	The boundary between two water masses of different properties; across a front, strong gradients in temperature and salinity (and other properties) occur.
Frustules	Silica (glass) cases that are part of a diatom cell wall.
Geomorphological	Concerning the shape of the Earth's surface.
Geostrophic wind	Wind that results from a balance between sea level pressure differences (flowing from high to low pressure) and the Coriolis force (directed to the left of the direction of motion in the Southern Hemisphere). As a result, without friction or other disturbances, geostrophic winds are directed parallel to isobars, flowing clockwise (counterclockwise) around regions of low (high) sea level pressure.
Glacial	(1) Relating to or denoting the presence or agency of ice; (2) a time interval characterized globally by cold conditions and abundant glaciation (=ice age).
Glacial runoff	Total meltwater discharged from glaciers.
Global model	See Ocean Global Circulation Model.
Global Overturning Circulation	See *Thermohaline/overturning circulation*.
Grazing	The process of plant matter consumption by an animal, in this case by zooplankton feeding on phytoplankton.
Greenhouse gas	An atmospheric trace gas that absorbs infrared radiation and contributes to the warming of the atmosphere; the latter process is known as the greenhouse effect.
Grounding line	Zone where the grounded ice sheet starts to float on the ocean.
Gyre	System of rotating ocean currents, associated with large-scale wind motions.
Habitat	Natural environment, or home, of a living organism, plant or animal.
Half-life	The time after which statistically one-half of the atoms of a radionuclide has decayed and become an isotope of another element (rarely another isotope of the same element). The half-life may range from fractions of a second to billions of years, but it remains constant for a particular isotope.
Halocline	Characterizing a strong vertical salinity gradient in the water column.

Haptophyte	The common name for a clade of algae, which have two slightly unequal flagella involved in swimming and a third organelle called haptonema, which looks also like a flagellum but has a different structure and is involved in feeding, mostly bacteria. Some haptophytes have an outer skeleton of calcareous plates, and others are known to be mucilaginous, sometimes building up foam on beaches.
Heat Flow	The loss or acquisition of heat between two bodies of different temperature, between atmosphere and surface ocean, or between two water masses.
Hemipelagite	Marine sediment type consisting of a fine-grained mix of biogenic and lithogenic particles (commonly deposited in the vicinity of continental margins).
Heterotrophic	When respiration/remineralization (that produces carbon dioxide) is greater than primary production (that consumes carbon dioxide via photosynthesis).
Heterotrophs	An organism that cannot manufacture its own food and instead obtains its food and energy by taking in organic substances, usually plant or animal matter, also DOC.
Hydrographic cruise	Generally cross-ocean-basin research expedition that analyzes the water column every 50–100 km for its physical and chemical characteristics.
Hydrothermal vents	Waters that have undergone heating and chemical modification by heat from the Earth's interior are released from the seafloor, preferentially in the vicinity of volcanic structures. The sites of this release are called hydrothermal vents.
Ice edge	The boundary between the ice-covered and open ocean. Note that the ice edge can often be very diffuse with ice concentration increasing only slowly over tens or hundreds of kilometers from the open ocean to the closed pack-ice zone. The ice edge is commonly defined as the 15% ice concentration isoline.
Ice extent	The area of sea ice, defined as the ocean area enclosed by the **ice edge**. Because ice extent is defined by the location of the **ice edge** only, its area does not account for regions of open water or low ice concentration within the ice-covered area. Therefore, the true ice covered area is smaller than the area of ice extent.
Ice rafting	Transport by drifting icebergs and/or sea ice.
Ice shelf	A thick, floating slab of continental ice connected to ice resting on land, sometimes described as a "tongue of ice" when connected to a single glacier.
Ice shelf cavity	Portion of the ocean beneath a floating ice shelf.
Ice streams	Zones of high ice flow velocity draining an ice sheet.
Iceberg	A large mass of continental (not sea) ice, floating in the ocean, detached from a glacier front or an ice shelf.
Idealized models	A representation of specific oceanic processes using mathematical equations that are solved analytically or numerically. Either the equations or the domain over which they are solved are simplified in order to eliminate processes that are not relevant to the purpose of the model.
Interglacial	A time interval characterized globally by warm conditions and very limited or absent glaciation.
Isopycnal transport	Mass transport along lines of equal (potential) density.
Isotope	The atoms of a chemical element occur in varieties that differ slightly by mass, but their chemical properties are the same. Most chemical elements consist of a mixture of isotope, where the ratios of the isotopes can serve as a fingerprint of the element's origin

Jurassic	Geological time period spanning 201 to 145 Ma.
Katabatic winds	Winds flowing down the slope of a large-scale topographic feature, such as a glacier, due to gravity pulling the high-density air down from a high elevation.
Large-scale Fe fertilization experiments	Primary production in the Southern Ocean is primarily limited by low availability of the trace element iron. By artificially increasing iron availability at the ocean surface, scientists can study the effects of this important controlling factor on productivity and species succession.
Laser and radar altimetry	An airborne or satellite method to measure the elevation of the earth surface above a reference datum. Over sea ice, laser and radar altimetry can be used to retrieve the height of the ice or snow surface above the local water level. This height is called **freeboard** and is a proxy for ice thickness. The thicker the (floating) ice, the higher its surface will extend above the water level.
Leads	Up to several hundred meters wide areas of open water in consolidated sea ice.
Length scale	This is the typical distance over which a certain process takes place, for example, the vertical distance of remineralization within the water column. For nitrogen and carbon, the length scale is smaller than for silicate and iron.
Lithogenic components	Sedimentary particles originating from the erosion of hard rocks and soils on land (as opposed to biogenic).
Macronutrients	Chemical elements (N, P, and Si) essential for life in the ocean. In oceanography, the term typically refers to the most abundant and potentially growth limiting (for phytoplankton) nutrients: nitrate, phosphate, and silicate (silicic acid). Nutrients in seawater are commonly present in micromolar concentrations but can be present at nanomolar levels or below the limit of detection of current analytical techniques.
Marginal Ice Zone	This is the transition zone between the open ocean and the sea ice. Since this zone strongly depends on factors like wind direction and ocean currents, its shape varies from isolated floes to a compact edge of small ice floes pressed together. Usually, the Marginal Ice Zone is a hot spot of biological activity during periods of light.
Marine mammal	Aquatic mammal that lives in the ocean and relies on oceanic ecosystems to survive.
Marine sediment core	An elongate sample of unconsolidated seafloor deposits with their original sequence of deposition preserved.
Mechanistic processes	Fundamental processes involved in describing an oceanic system.
Meteoric water	Water derived directly or indirectly from snow and rain (including output from rivers and glacial ice melt).
Microbial	Pertaining to microbes or unicellular organisms; in its narrow definition, it only includes bacteria, Archaea, and viruses. Also describing phytoplankton and unicellular zooplankton in its broader sense.
Micronutrient	Just like macronutrients, micronutrients are essential for life in the ocean. These are elements that are typically required for enzymatic functions within organisms (e.g., iron [Fe], manganese [Mn], and zinc [Zn]). They are present in seawater at nanomolar concentrations (or lower). Absence of micronutrients can limit primary productivity, as is the case of Fe in Southern Ocean waters.
Micropaleontology	Scientific study of the preserved or lithified small hard parts of dead life forms (fossils) under a microscope.

Microwave scattering properties	Most satellite observation methods use microwaves to sense the presence and properties of sea ice. The emission and scattering of microwaves by snow and ice depend on the media's dielectric properties, particularly on their density and stratification.
Microzooplankton	Heterotrophic organisms (needing organic carbon as food) that are usually unicellular, ranging in size from 20 to 200 microns.
Miocene	Geological epoch spanning 23.0 to 5.3 Ma.
Mixed layer	The upper waters of the water column (typically 5 to 200 m) in contact with the atmosphere where the chemical properties and density are homogenous because of turbulent mixing caused by winds.
Mixotrophic	This term classifies organisms which are flexible in their diet, for example, they can switch from obtaining energy from photosynthesis and from ingesting other organic compounds (i.e., heterotrophy).
Mode waters	Waters masses that are nearly vertically homogeneous and are formed by deep winter convection at the surface, for example, Subantarctic Mode Water.
Model domain size	The geographical region over which a model solution is sought.
Moorings	Strings of autonomous oceanographic sensors and sediment traps moored to the sea floor. These sensors provide long time series of changes of water properties and the sedimentation of organic or inorganic particles. In ice-covered regions, moorings are often equipped with upward-looking sonar to obtain records of ice thickness variability.
Mortality	As part of their life cycle or by processes such as grazing or cell lysis, organisms die; the rate of individual disappearance from the population is their rate of mortality.
Negative freeboard	Refers to the surface of the ice depressed below the seawater level. If snow is deposited onto the ice, its mass pushes the ice surface down, reducing freeboard. If the thickness of the snow exceeds approximately one third of the thickness of the ice, the surface of the ice will be depressed below the water level causing negative freeboard. As a result, *flooding* may occur, that is, water will flow onto the ice saturating the lowest layers of the snow with seawater.
Neural network	A computing system inspired by the activity of animal brains that learns and progressively improves its performance in set tasks without a priori information.
Next-generation sequencing	This is high-throughput technology to measure DNA with massive parallel sequencing of millions of DNA fragments in a single analysis.
Nonlabile	Slowly reactive, or refractory, and in biogeochemistry refers to material that is not readily available for organisms to use as a food source.
Nutrients	Chemical components indispensable for supporting life, or fertilizers. (See also macronutrients and micronutrients.)
Ocean acidification	Oceanic absorption of human-made carbon dioxide causes perturbations of the marine carbonate chemistry, which collectively are termed ocean acidification, such as the decrease in seawater pH and carbonate saturation states as well as the increase in dissolved CO_2. Acidification means becoming acidic; however, seawater will not easily become an acidic fluid in absolute sense, that is, its pH will stay above 7.

Ocean General Circulation Models	A representation of the ocean constructed with a set of mathematical equations describing oceanic dynamics and thermodynamics over time and in three-dimensional space. Equations for temperature, salinity, pressure, and velocity are solved numerically on a grid that is a discretized representation of the modeled domain. General Circulation Models optionally include sets of equations for processes occurring on scales smaller than the grid size and for complementary models such as sea ice or biogeochemistry.
Ocean state	The conditions of the ocean: its temperature, salinity, density, velocity, phase (liquid or ice), and chemical composition. The ocean state varies in space and in time; the mean state refers to the time-mean conditions.
Oceanic plate	Large piece of the rigid outer part of the Earth that consists mainly of magmatic rocks and is mainly covered by ocean water.
Oligocene	Geological epoch spanning 33.9 to 23.0 Ma.
Opal-poor	Containing only small amounts of siliceous (=opaline) hard parts of single-celled marine plants (diatoms) or marine animals (radiolarians, silicoflagellates, and sponges).
Outgassing	The release of a gas, typically carbon dioxide, to the atmosphere from the ocean
Overflow (of Ice Shelf Water)	Water masses crossing a sill at the continental shelf break to descend along the continental slope.
Overturning cells	Two vertically differentiated current systems in the Southern Ocean, the lower of which forms the Antarctic Bottom Waters and the upper of which forms the Intermediate Waters.
Ozone depletion	Reductions of ozone concentrations in the stratosphere due to the degradation of ozone (O_3) by chlorofluorocarbons, leading to cooling of the stratosphere and intensification of *stratospheric circulation*.
Pack ice	Consists of drift sea ice which is characterized by its constant movement influenced by winds and ocean currents. Once the coverage of the ocean surface by sea ice is >70%, the ice accumulation is called pack ice.
Particle-reactive radionuclides	Radionuclides that tend to adsorb to particles, which makes them prone to removal by particles (dubbed: scavenging). They are ofen used as tracers to understand biogeochemical cycles.
pCO_2	Partial pressure of carbon dioxide describes the concentration of this gas in a certain volume. It is the unit commonly used to describe the increase of anthropogenic greenhouse gasses in the atmosphere and the oceans.
Pelagic sediment	A fine-grained marine sediment type that accumulates as the result of the settling of particles to the floor of the open ocean, far from land.
Photoacclimation	Photosynthetic organisms employ several mechanisms (termed photoacclimative) on different time scales, such as changes in pigment concentrations and their positions inside the cells, to respond to changes in light availability.
Photoadaptation	Depending on the ecological niche, photosynthetic organisms have evolved different photoadaptive mechanisms, such as changes in their cellular pigmentation, that allow them to effectively harvest light without facing high light stress.
Photochemical degradation	This is the process by which some organic matter is broken down by the action of sunlight.
Photosynthetic activity/ photosynthesis	A process by which plants or phytoplankton convert light energy (sunlight), carbon dioxide, and water to biomass and chemical energy, generating oxygen as a by-product. When environmental conditions are conducive to phytoplankton growth, biomass accumulates.

Photosynthetic apparatus	The cellular structures in which photosynthesis takes place. It consists of light harvesting pigments, the photosynthetic electron transport chain, and ATPases (enzymes) to store part of the newly generated chemical energy.
Photosynthetic yield	The term Fv/Fm (variable fluorescence/background fluorescence), or yield in short, describes the maximum quantum efficiency of charge separation at photosystem II and is a measure of stress experienced by photosynthetic organisms. Low Fv/Fm indicates cellular stress, usually due to nutrient limitation.
Physiology	Physical, chemical, and biological processes that drive the functions and activities of living systems such as cells and organisms.
Phytoplankton	Microscopic marine algae that contain chlorophyll and require sunlight to live and grow. They generally reside in the surface ocean, floating in shallow waters where light can reach them. They are the base of the marine food web.
Pigments	Organic compounds which are able to absorb light from the sun and thus are able to capture the light energy necessary for photosynthesis. A second type of pigments are photoprotective, helping the cell loose excess radiative energy under high light conditions.
Platelet ice	Large (>1 cm), disc-shaped ice crystals forming by rapid ice-growth in *supercooled* water. Ice platelets raise through the water column and accumulate at the surface or under sea ice to strongly contribute to sea-ice growth.
Pleistocene	Geological epoch spanning 2.6 Ma to 10 ka.
Polar cyclones	Large-scale winds flowing clockwise in the Southern Hemisphere around a low-pressure area.
Polynya	An area of the ocean that is not covered by sea ice in an otherwise sea-ice covered region.
Potential temperature	The temperature a parcel of water would have if raised adiabatically to 1,000 mb (surface pressure).
Primary Production	The rate of energy conversion, to form organic substances from inorganic sources, usually fueled by sunlight. In the absence of solar radiation, by organic or inorganic chemical sources can act as electron donors (e.g., nitrate and methane).
Progradation	Lateral, seaward directed outbuilding of sedimentary strata on a continental margin.
Protists	All unicellular organisms in the ocean.
Pycnocline	A strong vertical density gradient.
Quaternary	Geological time period spanning 2.6 Ma to 0.0 ka.
Radiocarbon	The name for the ^{14}C isotope of carbon. It is produced in the upper atmosphere by cosmic radiation from where it travels to the Earth's surface to be absorbed by the oceans/living biota. Although only present in trace amounts, as ^{14}C is radioactive (with a half-life of ~5,700 years), it is radiometrically detectable. This lends itself to be a very useful tracer for examining ocean circulation, given its half-life is of the same order of magnitude as that of ocean ventilation. In the 1950s and 1960s, the testing of atomic bombs led to a large pulse of ^{14}C being injected into the atmosphere, that on transferal to the ocean, enabled it to be used as a tracer of ocean ventilation.
Radionuclide	Isotope (see above) of an element that converts spontaneously to isotopes of a different element (rarely the same element) by emitting radiation from the atom's nucleus. Each radionuclide has a specific, unique, and constant probability to undergo this transition, leading to the decay of the radionuclide as a well-defined function of time (also see Half-life)

Reactivity	This is the tendency of a substance to undergo chemical reaction, either by itself or with other materials, and to release energy.
Reanalysis products	Global weather and wind information produced by weather forecasting centers based on meteorological observations and numerical weather models. They are often used as input data for global climate models.
Redox reactions	Chemical reduction-oxidation (redox) reactions changing the oxidation state of atoms. In biology, redox reactions are the basis of all electron transfer processes within the cell, such as photosynthesis and respiration.
Refractory	Used mostly for carbon material that is very difficult to remineralize and is thus poorly or nonlabile.
Regional model	A general circulation model that is solved over one region of the global ocean; it has one or more boundaries open to other oceanic regions through which properties are exchanged.
Remineralization	The breakdown of organic matter to its simplest, constituent inorganic forms, like nutrients or CO_2. This process consumes oxygen.
Residence time	Refers to the average length of time that a property persists within a system (the ocean or an ocean region), given its sources and sinks. If supply and removal of the property/element are in balance, the residence time is defined as the ratio of reservoir size and input/output terms.
Resilience	The capacity of an organism to maintain normal life processes after being exposed to damaging conditions.
Rheology	The laws that govern viscous or plastic *deformation* of continua, in this case of sea ice.
Ridging	The formation of pressure ridges as a result of ice *deformation* under compression or shear.
Rift phase	A time interval when a continent breaks up and when its individual fragments start to drift apart.
Rossby radius of deformation	The length scale at which flows are significantly affected by the rotation of the Earth; it determines the size of eddies and varies strongly with latitude, values being smaller at high latitudes. The Southern Ocean has an average radius of deformation of 10 m.
Roughness	Due to the deformation of sea ice forming pressure ridges and due to windblown snow forming sastrugis, the surface and bottom of sea ice are rarely smooth and level but often very rough. The roughness of the ice surface and bottom modify the impact of winds and currents on ice *drift*.
Salps	A primitive tunicate animal, of the group Chordata (with spinal chord), that feeds on small phytoplankton through a mucilaginous net that once full is ingested. They are sometimes found in large quantity or swarms.
Saturation	Regarding dissolved gases, they are said to have reached saturation when the seawater and atmosphere are at equilibrium.
Scavenging	The uptake of an element or a compound by particles and its subsequent removal from the ocean by gravitational settling.
Sea-ice flux	The volume of sea ice that *drifts* through a certain region during a certain time comprises a flux of sea ice. Because sea ice is less saline than seawater, sea-ice flux contributes to the oceanic freshwater balance.
Sea-air exchange	The process of exchange of material, usually gases, and heat, between the surface ocean and the atmosphere across the air-sea interface.
Seasonal ice zone	The region of the ocean that is covered by sea ice only during part of the year, mostly winter. In spring and summer, the sea-ice melts, enabling the seawater-atmosphere interactions.

Sedimentary	A rock originating from the continuous deposition and subsequent solidification of small particles raining out from a medium (e.g., seawater).
Sedimentation	The process of settling particles from the surface of the ocean to depth; mostly refers to matter being deposited on sediments.
Seismic	Relating to or denoting geophysical surveying methods for imaging the subsurface with artificially generated sound waves.
Sequestration	The long-term storage of carbon on climatically important time scales.
Sill	Topographic rise at the rim of troughs or ocean basins.
Sink	A location of the ocean where there is net absorption of a gas from the atmosphere.
Size classes	The phytoplanktonic community contains a wide spectrum of different size particles/cells. The most prominent are picoplankton (0.2 and 2 μm), nanoplankton (2–20 μm in size), and microplankton (larger than 20 μm).
Snow metamorphism	Changes of snow properties after deposition. These changes can be a result of, for example, compaction, temperature and radiation variations, moisture fluxes, or thaw.
Solubilization	The process of particles dissolving into the seawater.
Standing stocks	The amount, or biomass, of a species, per unit volume, at a certain location and time, used mostly for animals but also plants or phytoplankton.
State estimate	A best estimate of the physical and/or biogeochemical properties of the ocean at a particular snapshot in time which is consistent with the available, imperfect observations and the dynamical constraints provided by a time-evolving ocean model.
Stratosphere	The stratosphere is the Earth's atmospheric layer above the troposphere, the lowest layer of the atmosphere, situated from the Earth's surface to about 6 km in polar regions, extending from the tropopause to around 50 km above the surface.
Stratospheric circulation	Winds in the stratosphere. Over Antarctica and the Southern Ocean, stratospheric circulation is characterized by a strong clockwise (cyclonic) vortex centered near the South Pole.
Substrate	The reactant which is consumed or acted upon during a catalytic or enzymatic reaction.
Supercooled water	Water that is colder than its freezing temperature. Supercooled seawater can form when cold, deep water ascends to shallower depths, thus experiencing pressure-dependent increases of its freezing temperature. This process is common with Ice Shelf Water formed by bottom melt under ice shelves. Supercooling is reduced by the release of latent heat from the formation of *platelet ice*.
Tectonically	(1) Pertaining to the structure of the Earth's crust; (2) referring to the forces or conditions within the Earth that cause movements of the crust.
Teleconnection	This term describes a situation where processes at two distant sites are linked.
Terrigenous	Originating from erosion on land.
Thermal infrared	In the absence of clouds, thermal infrared sensors (radiometers) can observe temperature differences at the surface of the oceans. This is an effective means to distinguish between cold ice and warm water.
Thermohaline/overturning circulation	Global system of vertically and horizontally separated ocean currents, driven by the differences in ocean density created by surface heat and freshwater fluxes.

Thorium isotope	Thorium-234 (^{234}Th; half-life 24.1 days) is the natural radioactive daughter of uranium-238 (^{238}U; half-life 4.468×10^9 years). Once produced by the decay of ^{238}U in seawater, ^{234}Th is rapidly (days to months) scavenged by surrounding particles. If the rate at which these particles sink is greater than the rate at which ^{234}Th is produced, then the upper ocean will become depleted in ^{234}Th relative to ^{238}U. This deficit can be quantified to derive an estimate of ^{234}Th and particle export fluxes, and by combination with a site-specific ^{234}Th-carbon ratio, an estimate of upper ocean carbon export.
Tintinnids	see *Ciliates* above
Top-down control	When a population or biological community is limited in its abundance by other organisms that feed upon it, it is described as a top-down control on the population.
Topographic shelf waves	Waves in the ocean guided by topography, for example, the continental shelf break.
Trace element	A chemical element found only in very low concentration, such as micronutrients.
Tracer	Property or chemical compound that is transported through the oceans as part of the water movement and that if sampled and analyzed can thus be used to track these water masses. Temperature and salinity are the classical (physical) tracers in the oceans, as well as dissolved oxygen. Well-known anthropogenic tracers are chlorofluorocarbons which occur in extremely low concentrations.
Transport stream function	A quantity from which the oceanic flow can be inferred. The volume of water passing through a point over a period of time, or transport, is calculated by taking the derivative of the stream function, which is a measure of how abruptly the stream function varies over space. The flow is the largest where the lines of constant stream function are closest together.
Trench	Describes a depression in the sea floor, often narrow and long.
Trophic levels	Trophic levels start with the life forms that produce biomass from light or chemical energy as their primary energy source. Organisms of higher trophic levels eat organisms of lower trophic levels. "Trophic" derives from the old Greek word for food.
Trough	A long depression generally wide and flat bottomed with symmetrical and parallel sides.
Turbidite	Sediment type deposited from a gravitationally driven downslope flow of particles suspended in water; occurs often on continental margins.
Turbulent mixing	Centimeter-scale, chaotic fluctuations in water velocity which ultimately convert eddy and internal wave energy in the ocean into heat.
Undersaturation	When the concentration of a gas in seawater is lower than the solubility of that gas. It causes a flux of that gas from the atmosphere to the ocean until equilibrium is reached.
Upward-looking sonar	An underwater instrument to measure the thickness of sea ice. The upward-looking sonar measures the two-way travel time of an acoustic pulse transmitted and received by the instrument and reflected by the bottom of the ice. The depth of the bottom of the (floating) ice (called draft) is related to its thickness, similar to how ice *freeboard* is related to ice thickness.
Upwelling	Denotes rising, in this case rising seawater from depth toward the surface of the ocean.

V-shaped slope front	A topographically controlled front in southern and western Weddell Sea, marked by strong subsurface gradients in ocean temperature and chemistry and by stronger alongshore currents than most of the adjacent continental shelf.
Ventilation	The process that transports ocean water with trace gasses, including oxygen, from the surface, where gases have equilibrated with the atmosphere, to the deep ocean.
Viral infection	Like other organisms, marine species are subject to infections by viruses. With respect to phytoplankton, viral infections are thought to be an important reason for collapse of blooms.
Vortex stretching	Rotational enhancement of a water column as it enters deeper waters.
Water mass	A body of water with a common formation history and different physical properties to the surrounding water.
Westerlies	Semipermanent belt of winds blowing from the west, surrounding the Antarctic continent.
Whaling	The fisheries that catch whales for their useable products, meat, oil, and blubber, very prominent in the Weddell Sea in the 1800s and early 1900s; the hunting of whales was banned worldwide in 1986, the Southern Ocean was declared a whale sanctuary in 1994.
Winter Water	Remnant water of previous winter's mixed layer, residing beneath the shallow summer thermocline.
Zooplankton	From the greek *zoo*, animal and *plankton*, drift or wanderer. Zooplankton comprises animals that are heterotrophic, consuming algae and/or other animals as food.

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