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Antarctic Sea Ice— A Polar Opposite?

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RVIB *Nathaniel B. Palmer* alongside an ice floe in the Amundsen Sea, January 2009. *Photo credit: Maria Stenzel*

ABSTRACT. As the world's ice diminishes in the face of climate change—from the dramatic decline in Arctic sea ice, to thinning at the margins of both the Greenland and Antarctic ice sheets, to retreating mountain glaciers the world over—Antarctic sea ice presents something of a paradox. The trend in total sea ice extent in the Antarctic has remained steady, or even increased slightly, over the past three decades, confounding climate model predictions showing moderate to strong declines. This apparent intransigence masks dramatic regional trends; declines in sea ice in the Bellingshausen Sea region that rival the high-profile decline in the Arctic have been matched by opposing increases in the Ross Sea. Much of the explanation lies in the unique nature of the Antarctic sea ice zone. Its position surrounding the continent and exposure to the high-energy wind and wave fields of the open Southern Ocean shape both its properties and its connection to the atmosphere and ocean in ways very different from the Arctic. Sea ice extent and variability are strongly driven by large-scale climate variability patterns such as the El Niño-Southern Oscillation and the Southern Annular Mode. Because many of these patterns have opposing effects in different regions around the continent, decreases in one region are often accompanied by similar, opposing increases in another. Yet, the failure of climate models to capture either the overall or regional behavior also reflects, in part, a poor understanding of sea ice processes. Considerable insight has been gained into the nature of these processes over the past several decades through field expeditions aboard icebreakers. However, much remains to be discovered about the nature of Antarctic sea ice; its connections with the ocean, atmosphere, and ecosystem; and its complex response to present and future climate change.

INTRODUCTION

The decreasing Arctic sea ice extent has been perhaps the most conspicuous example of climate change anywhere on Earth (e.g., Perovich, 2011). A "canary in the coal mine," this thin, icy veneer over the ocean is among the most sensitive indicators of climate change, and one of the first warning signs of its impact. In contrast, Antarctic sea ice is a canary with a more enigmatic tune. Although annual average polar sea ice extent shows a statistically significant circumpolar decrease in the Arctic (–3.8 ± 0.2% per decade over 1979–2008), the Antarctic shows a small circumpolar increase (+1.2 ± 0.2% per decade) (e.g., Comiso, 2010). While the rate of sea ice decline in the Arctic has been even more rapid than model predictions (Stroeve et al., 2007), most climate models also predict decreases in the Antarctic (e.g., Arzel et al., 2006). This seemingly surprising increase has been seized upon by some to call into doubt the predictive power of climate models. Understanding this conundrum has societal relevance as this uncertainty, real or perceived, may influence public policy on climate change.

In fact, Antarctic sea ice is quite distinct from its northern counterpart. This stems from differences in geography, sea ice growth and decay processes, large-scale climate interactions, and ice-ocean interactions and feedbacks (Turner and Overland, 2009). Whereas sea ice in the Arctic Ocean is largely landlocked and relatively protected, and it can survive for many years, Antarctic sea ice is bounded to the south by the Antarctic Continent and exposed to the vast Southern Ocean to the north. Moreover, Antarctic sea ice extends into much lower latitudes (mostly between 60°–70°S, as compared to 70°–90°N in the Arctic). As a result, most sea ice in the Antarctic is seasonal—freezing in winter and melting again each summer. The extent of the ice pack varies from a winter maximum of about 19 million km^2 to a summer minimum of just 3–4 million km^2. Historically, the Arctic has held much more perennial ice (ice surviving for more than a year), with a winter maximum of 15–16 million km^2 reducing to about 7 million km^2 in summer, although in recent years sea ice extent has plunged to just 4–5 million km^2 (Comiso, 2010).

Around the unbounded Southern Ocean, the nature of the sea ice cover is shaped by the world's strongest prevailing (westerly) winds and highest waves as well as frequent storms. Consequently, the Antarctic sea ice environment is much more dynamic and ephemeral than its Arctic counterpart. Storms bring moisture southward, so that the Southern

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Ocean and its sea ice experience the highest snowfall rates of any region on Earth, while the Arctic Ocean is a desert (Serreze and Hurst, 2000). Despite its vast extent, Antarctic sea ice forms only a thin veneer on the ocean's surface—mostly less than one-meter thick (Worby et al., 2008a). Seasonal ice thickness is limited by the high ocean heat flux from the relatively warm Circumpolar Deep Water (CDW) that pervades the subsurface waters of the Southern Ocean (Martinson and Iannuzi, 1998).

These factors determine both the response of sea ice to climate change and variability and its impact on the atmosphere, ocean, and polar ecosystem. Much has been learned about the nature of Antarctic sea ice from science expeditions aboard icebreakers over the past three decades. Yet, much still remains to be learned. For example, while a record of sea ice extent variability over the past three decades has been possible with continuous satellite observations, a reliable estimate of even the average ice thickness and that of its snow cover still eludes us (Giles et al., 2008). Until such fundamental information is obtained, our ability to build and validate accurate predictive models, to attribute observed changes, and to assess the impacts of current and future change remains elusive.

THE CHANGING ANTARCTIC SEA ICE COVER

In contrast to the observed overall increase, most climate models simulate decreases in Antarctic sea ice extent over the past 30 years (Figure 1). These predicted decreases continue into the future, showing a 30% decline by 2100 (Arzel et al., 2006). However, most climate models also fail to accurately reproduce mean ice extent (particularly in summer) and overestimate its year-to-year variability (Figure 1). This failure is in part due to poor representation of the Southern Ocean in models (Russell et al., 2006), but may also reflect poor representation of sea ice processes and a lack of in situ data for validating models.

Warming over the Southern Ocean since the 1950s is observed in both the atmosphere (Chapman and Walsh, 2007) and surface ocean (Gille et al., 2002), and is predicted by climate models, albeit more modestly than over the Arctic (Bracegirdle et al., 2008). At first glance, this warming seems to be contradicted by the observed modest increase in sea ice. However, surface temperature alone may not be the primary driver of ice extent.

The Southern Ocean climate is dominated by strong prevailing westerly winds that circle the continent. There has been a pronounced poleward intensification of these winds since the 1970s (Hurrell and van Loon, 1994), primarily due to a change in the dominant mode

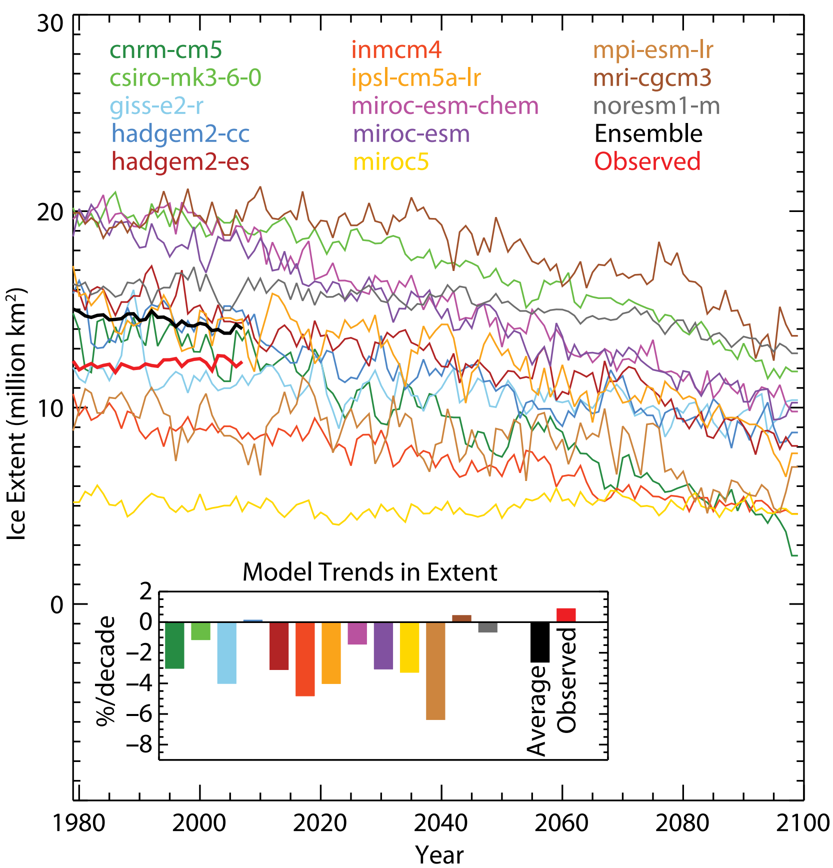

Figure 1. Annual mean modeled sea ice extent for the late twentieth century (1979–2007) and projected extent to 2100 for individual runs from 13 Coupled Model Intercomparison Project (CMIP5) climate models. The observed satellite extent for 1979–2007 is shown by the thick red line. The thick black line is the average of all models for the same period. (Inset) Trends are shown relative to the mean ice extent over 1979–2007 for each model. Ice extent is poorly represented in most models, and almost all model runs show a small but significant downward trend, while the observed ice extent has increased by a small but statistically significant fraction. Explanation of the model nomenclature can be found at http://cmip-pcmdi.llnl.gov/cmip5. *CMIP5 data courtesy of Tom Bracegirdle*

of atmospheric variability, the Southern Annular Mode (SAM). The SAM is characterized by sea level pressure differences between the continent and mid-latitudes (Marshall, 2003), and its phase has become more positive, meaning lower pressures at higher latitudes and stronger, more poleward westerly winds. These changes appear to be anthropogenically driven, with a more positive SAM caused by both increasing atmospheric carbon dioxide (Marshall et al., 2004) and tropospheric ozone depletion (Gillet et al., 2009; Thompson et al., 2011). In the Southern Hemisphere, increasing westerly winds push sea ice further north, due to the tendency of the ice to drift to the left of the wind (by a mechanism called Ekman transport). This process tends to increase sea ice extent, thereby opposing the expected melt due to warmer temperatures.

The More Things Change, the More They Stay the Same

The response of the sea ice is not so simple, however. The overall increase in ice extent obscures dramatic changes that have been observed regionally over the past three decades. Ice extent has increased in the Ross Sea (5% per decade) and decreased in the Amundsen and Bellingshausen Seas (7% per decade) (Turner et al., 2009). More strikingly, the length of the annual ice-free season has decreased in the western Ross Sea by over two months but increased in some areas of the Bellingshausen Sea by three months (Figure 2, and Stammerjohn et al., 2012). In fact, the rate of increase in ice-free conditions in the Bellingshausen Sea region is even *greater* than in regions of greatest ice decline in the Arctic.

The distinction between decreasing ice extent and length of ice-free summer is important, as it is open-water duration that controls solar heating of the upper ocean that then drives observed increases in sea surface temperature (Meredith and King, 2005) and ocean ecosystem impacts (Montes-Hugo et al., 2009). The longer ice-free summer and increased westerly winds also allow for greater wind mixing and upwelling of warm CDW onto the western Antarctic continental shelves (Martinson, 2011), increasing ocean heat content from below (Martinson et al., 2008). In some recent years (2008–2009), sea ice has completely disappeared from the Bellingshausen and eastern Amundsen Seas in summer, exposing continental ice shelves to open water for longer periods, possibly speeding their collapse (e.g., Massom et al., 2010).

There are two primary geographic factors that will modify any zonally symmetric forcing and can help explain the regionally contrasting sea ice trends. The continent itself is centered somewhat off the pole, with an irregular coastline

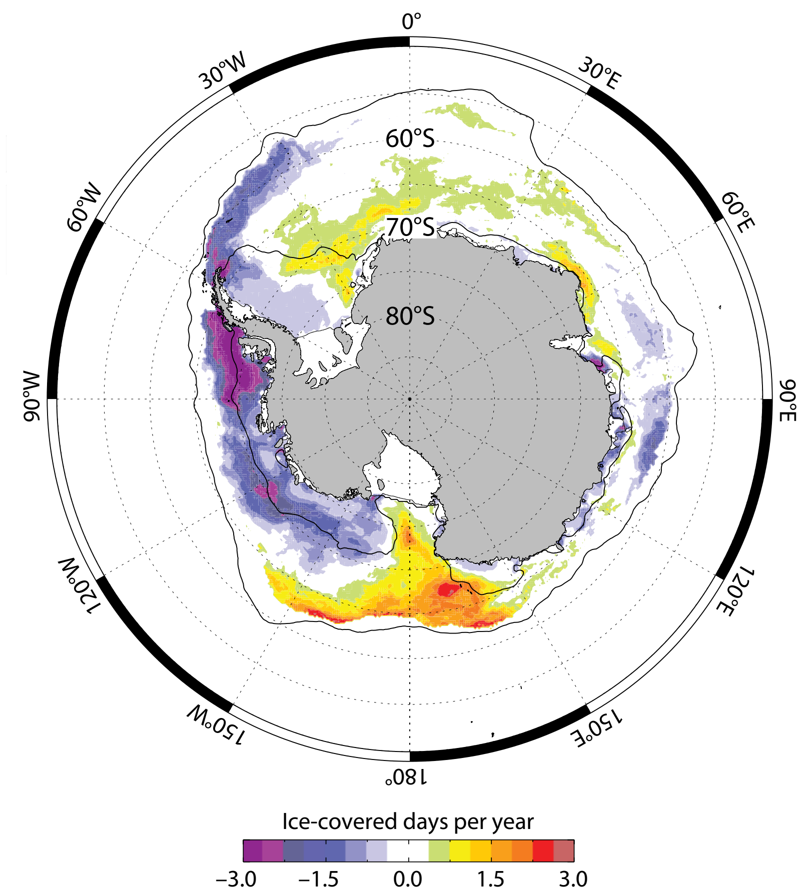

Figure 2. Trends in the length of the ice-covered season for 1979–2010 in the Antarctic, where large but opposing regional trends are evident. The greatest change has been an almost three month decrease in ice season duration in the Bellingshausen Sea over the past three decades. There is a comparable increase in ice season duration in the Ross Sea. These opposing trends resulted in a slight overall positive trend in ice extent for the Southern Ocean as a whole. Black lines show the mean minimum and maximum ice extent for the same time period.

that modifies westerly wind flow around the continent. For example, the westerly flow is perturbed as it passes into the Ross Sea region, pushing storms into the Amundsen Sea (Baines and Fraedrich, 1989) to create a quasistationary low-pressure system in the South Pacific, known as the Amundsen Sea Low (ASL). Increasing winds spin-up (deepen) the clockwise-rotating ASL, causing stronger northerly winds along its eastern limb (in the Bellingshausen Sea region), and southerly winds along its western limb (in the Ross Sea region).

Strong, warm winds from the north along the Antarctic Peninsula push the ice edge further south (Harangozo, 2004), delay ice advance in autumn, and speed its retreat in spring (Stammerjohn et al., 2008). At the same time, in the Ross Sea, increased cold air outbreaks from the south favor increased ice production and northward drift. The regional changes are strongest in autumn, which may be a consequence of a delayed surface signature of increased depletion of stratospheric ozone strengthening the SAM (Thompson et al., 2011), deepening the ASL, and increasing sea ice in the Ross Sea (Turner et al., 2009). When regional sea ice changes are averaged together, circumpolar sea ice extent shows only a small net change.

This scenario cannot account for all the regional sea ice changes or for the overall positive trend in ice extent (Lefebvre and Goosse, 2008). Tropical variability and the El Niño-Southern Oscillation (ENSO) strongly influence the Ross, Amundsen, and Bellingshausen Seas (Yuan, 2004). During El Niño events, the storm track shifts to the north in the South Pacific and to the south in the South Atlantic, bringing warmer air and less sea ice in the Ross Sea region and cooler air and more sea ice in the Bellingshausen and Weddell Seas (Yuan, 2004). The converse occurs during La Niña events. Furthermore, when a positive SAM accompanies La Niña, the ASL is deepest and further southeastward, and strong northerly winds greatly retard sea ice advance west of the Antarctic Peninsula. Modes of climate variation thus act in concert to affect sea ice extent in this region (Stammerjohn et al., 2008). Sea ice in the western Ross Sea region is less influenced by ENSO variability; here, intensification of westerly winds and increased Ekman transport have led to a later wind-driven retreat and earlier advance.

This explanation may elucidate sea ice variability in the South Pacific sector of the Southern Ocean—anthropogenic influences drive an increasing SAM, leading to small overall increases, while couplings between SAM and ENSO drive large regional changes. But again, this explanation is too simplistic, as it ignores the role of ocean forcing and feedbacks. Also, it ignores the role the sea ice itself may play in modulating the response to external (atmospheric or oceanic) forcing through processes governing ice growth and decay and ice-ocean feedbacks.

ICE PROPERTIES AND PROCESSES

The nature of Antarctic sea ice had only begun to be uncovered in the late 1970s and early 1980s as the first scientific forays into the summer pack ice began (e.g., Gow et al., 1987). Dedicated sea ice research cruises aboard modern icebreakers such as *Nathaniel B. Palmer*, *Polarstern*, and *Aurora Australis* have greatly improved our knowledge and understanding of the Antarctic sea ice environment since the mid-1980s.

In 1986, the first true winter expedition deep into the Antarctic pack ice discovered that the primary process of sea ice formation there was far different than in the Arctic (Wadhams et al., 1987; Lange et al., 1989). Rather than freezing vertically downward like a sheet of ice on a lake, loose "frazil ice" crystals form in the turbulent, wind- and wave-affected environment of the open ocean. As these crystals accumulate into a soupy mass, they begin to consolidate into circular pans known as "pancake ice" (Figure 3). As the ice dampens the waves, the pancakes consolidate into a continuous sheet. In the Southern Ocean, the swell can penetrate hundreds of kilometers into the pack ice before the ice becomes consolidated (e.g., Jeffries and Adolphs, 1997). Because the frazil crystals are constantly stirred deep into the water column, the ocean can continue to lose large amounts of heat during wind- and wave-driven frazil ice production. Once formed into pancakes at the surface, any continued wind/wave action will cause pancake rafting and rapid thickening to as much as 40–70 cm before consolidation occurs (Lange et al., 1989). Ice core data suggest that typically between about 20% and 60% of Antarctic sea ice is composed of frazil ice, depending on the region (Lange et al., 1990; Worby et al., 1998; Jeffries et al., 2001). Unfortunately, this key process is not yet included in sea ice models.

Another ~ 7–40% of Antarctic sea ice is composed of snow ice (described below). The remaining thickness is composed of ice formed by downward freezing, known as congelation ice—the principal mode of formation for Arctic sea ice (Weeks, 2011). There, the ice is insulated from the warm, very salty Atlantic water by the so-called cold halocline, a layer of cold, salty water that sits at the base of the (cold/fresh) Arctic surface mixed layer (Aagaard et al., 1981). The heat flux from ocean to ice is almost nil (typically less than 1 W m^{-2}; Steele and Boyd, 1998), allowing ice to grow several meters thick over several years. In the Antarctic, there is no such barrier, and warm CDW is easily mixed up from below. This mixing can happen due to mechanical stirring caused by sea ice motion on the ocean surface, particularly during storms, or by convection due to ice growth and brine release.

When the ocean freezes, salty brine is rejected from the growing ice, and the dense brine sinks and mixes with the water below. This process deepens the surface mixed layer. As it deepens, it entrains warm CDW from below, bringing heat to the surface that retards further ice growth. Averaged over the winter, this "ocean heat flux" is typically 15–35 W m^{-2} (Martinson and Iannuzi, 1998; Martinson et al., 2008)—enough to limit undeformed ice thickness to about 70 cm. Even in the depth of winter, melting of the ice underside is frequently observed (e.g., Jeffries et al., 1998). So how, then, does thick ice form at all?

Two key processes are at play—deformation and snow ice formation. Given the moist, maritime environment of the Southern Ocean, large amounts of snow fall on Antarctic sea ice. In fact, more snow falls on Antarctic sea ice than on any comparably sized region on Earth (Maksym and Markus, 2008). Because of high accumulation rates and thin ice, the weight of the snow often depresses the sea ice surface below sea level. Brine and seawater then infiltrate into the snow to create a slush layer. This layer subsequently freezes to form a salty, granular ice layer known as "snow ice" (Jeffries et al., 2001). The amount of snow ice varies temporally and spatially, comprising about 10% of the ice in the Weddell Sea (Lange et al., 1990) to as much as 40% in the Bellingshausen and Amundsen Seas (Jeffries et al., 2001). Because freezing now occurs from the ice surface, instead of at the bottom where ice growth is limited by the insulating effects of the snow and the high ocean heat flux, this can be a more effective means of ice thickening. In extreme cases, it creates an ice growth "conveyor belt," where ice grows on the top and melts from the bottom, such that the entire ice thickness can be composed of snow ice (Lytle and Ackley, 2001).

Two long-lived deployments of ice mass balance buoys (IMBs) in the Amundsen and Weddell Seas in 2009 illustrate the delicate balance between snow accumulation and ocean heat (Figure 4). Such devices have autonomously monitored the growth and decay of sea ice and its snow cover for nearly two decades in the Arctic, but to date only a handful have been deployed in the Antarctic (e.g., Perovich et al., 2004).

The IMB record from the Weddell Sea (bottom panel, Figure 4) represents a cold

Figure 3. Antarctic sea ice during initial formation (left) and during late summer decay (right). At the advancing ice edge, initial freeze-up occurs rapidly through the "frazil-pancake" cycle, whereby loose crystals formed in turbulent waters coalesce into pancakes that are both buffeted by and dampen the incoming waves. In summer, rich algal communities form in rotten "gap" layers found just beneath the freeboard layer of the ice. These extremely porous layers are readily seen from above, as a ship moving through the pack ice will shear the solid upper ice from the rotten, porous ice layer below. The formation of these layers is likely controlled by a complex interplay of physical (and possibly biological) processes, emblematic of the close coupling between sea ice physical and biological processes.

ice regime wherein a relatively thin snow cover provides poor insulation, allowing the ice to cool and thicken, much like what is observed in the Arctic winter. The following summer also brought minimal surface melt so that as the snow melted, the ice actually thickened as the melt-water refroze at the snow-ice interface to form "superimposed" ice (e.g., Kawamura et al., 2004). This process differs from that in the Arctic, where surface melt and melt pond formation have (until recently) been the dominant mode of summer ice decay. The IMB record from the Amundsen Sea represents a warm ice regime caused by a thick snow cover providing increased insulation. The ice floods, but snow accumulation is too fast for this layer to freeze completely into snow ice. The ice rots from within as the ocean melts it from below, overwhelming the ability of the "conveyor-belt" growth mechanism to maintain the ice thickness.

These data point to the importance of not only external processes but also internal, fluid-phase processes in the evolution of Antarctic sea ice. Sea ice does not reject all the salt in seawater as it freezes. It traps brine within a complex, porous microstructure (e.g., Weeks, 2011). This microstructure acts as a conduit for fluid transport, supplying seawater to the snow-ice interface, causing snow ice formation upon refreezing (Maksym and Jeffries, 2000). Convective overturning can be vigorous and represent a significant upward heat flux (Lytle and Ackley, 1996) that delays refreezing of the ice beneath, leaving it susceptible to internal and bottom melt.

Ice brine dynamics are a boon to the internal sea ice ecosystem. The porous brine network forms a rich habitat for microorganisms. Brine overturning during flooding and snow ice formation is critical for supplying nutrients to algal communities within the ice, fueling algal blooms well into autumn (Fritsen et al., 1994). In summer, rotten, highly porous "gap" layers can form (Figure 3). Such layers often harbor intense blooms of ice algae with algal concentrations as high as are found anywhere in the polar oceans (Ackley and Sullivan, 1994; Haas et al., 2001). Notably, neither internal community occurs in the Arctic. The role that brine dynamics plays in structuring the ice and its ecosystem is just beginning to be understood (Vancoppenolle et al., 2010; Saenz and Arrigo, 2012).

Figure 4 illustrates that there is an upper limit to how thick Antarctic sea ice can grow thermodynamically, determined by the balance of heat loss to the atmosphere, snow accumulation

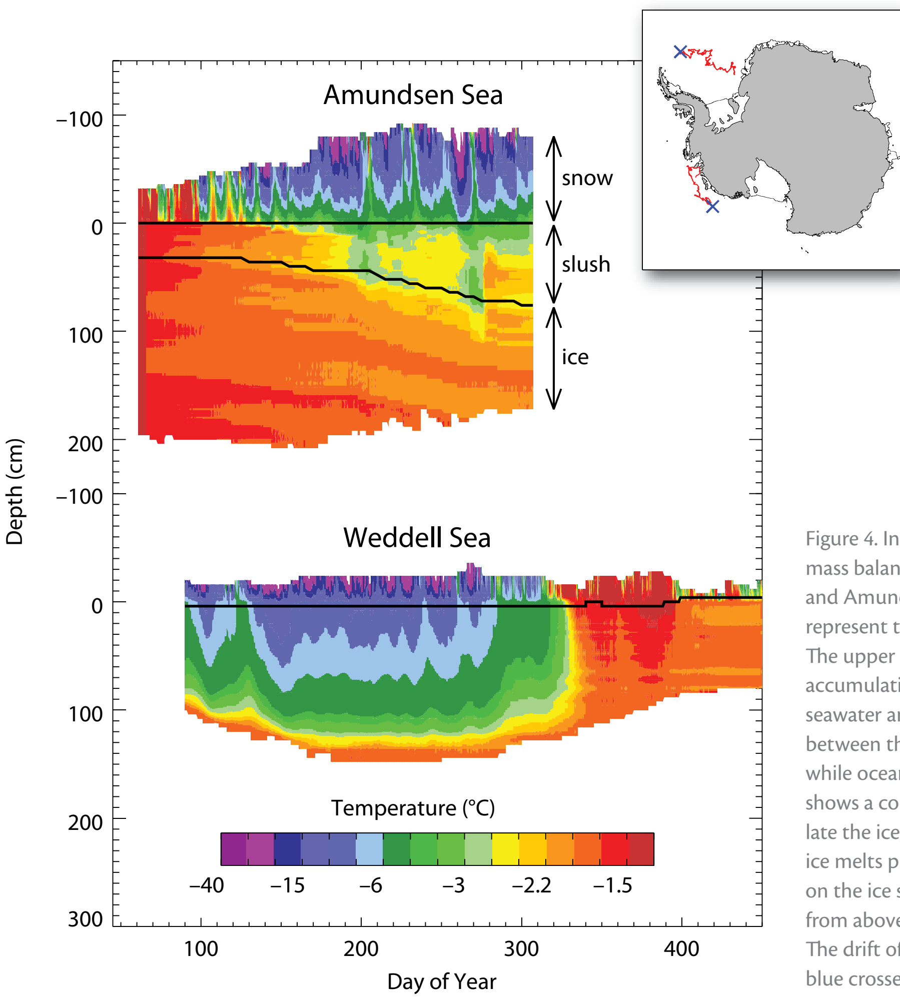

Figure 4. Internal sea ice temperature changes determined by ice mass balance buoys (IMBs) deployed in sea ice floes in the Weddell and Amundsen Seas in February 2009. The upper/lower bounds represent the snow surface and the bottom of the ice, respectively. The upper record shows a warm ice regime, where heavy snow accumulation both insulates the ice and floods the surface with seawater and brine, leading to a slushy layer (denoted by the region between the black lines) that partially freezes on the ice surface while ocean heat melts the ice from below. The Weddell record shows a cold ice regime. The thin snow cover is insufficient to insulate the ice, allowing ice to accrete on the bottom. In summer, the ice melts predominantly from below. Summer snowmelt refreezes on the ice surface to form superimposed ice, thickening the ice from above (denoted by the increase in the height of the black line). The drift of the IMBs is shown by the red lines in the inset map. The blue crosses denote their final positions when the IMB failed.

rates, brine dynamics, and ocean heat flux. Ice thickening beyond about two meters largely occurs by ice deformation (i.e., rafting and ridging). Antarctic sea ice is constantly in motion due to the strong winds and frequent storms in the Southern Ocean that cause alternating episodes of divergence and convergence. Areas of open water (leads) that are exposed when the pack ice diverges freeze rapidly, so that new ice is continually produced even when basal ice growth on thicker floes has ceased. This ice subsequently thickens mechanically during intervening periods of ice convergence, leading to deformed ice that is typically blocky and irregular (see photo on p. 140), in contrast to the "hedgerow"-like ridges found in the Arctic (Tin and Jeffries, 2003). Extreme ice deformation and mechanical thickening can occur where ice drift is blocked by coastal features (Massom et al., 2006) or by assemblages of icebergs grounded in shallow (less than about 350 m) waters (Massom et al., 2001). This mechanism produces much greater volumes of ice than would thermodynamics alone, and the thicker ice is more likely to survive the summer melt. Sea ice thickness in excess of six meters has been observed in some coastal locations in summer (e.g., Jeffries et al., 1994). In winter, such thick ice is inaccessible to even powerful icebreakers. To characterize the complete spatial distribution of ice thickness will require improved methods of observation.

Modulating the Response to Change

The processes coupling precipitation, ice growth and melt, and upper-ocean stratification control the response of sea ice to a changing climate through the balance of ice-ocean interactions and feedbacks. These interactions and feedbacks are strongly dependent on region and season. Their net effect is not well understood or known, but several scenarios have been proposed. For example, an earlier wind-driven sea ice retreat that leads to upper-ocean warming and continued decreases in sea ice creates a positive (amplifying) feedback (e.g., Meredith and King, 2005) and accelerates sea ice retreat in some regions. A negative (stabilizing) feedback can occur in other regions as enhanced ice divergence under strengthening westerly winds will tend to increase ice production, which in turn will cause more convective overturning and upward heat flux to inhibit ice growth (Sigmond and Fyfe, 2010). Several studies have also implicated freshening of the upper ocean in stabilizing the ice cover. Bitz et al. (2006) suggested that atmospheric warming would stabilize the water column through surface freshening, reduced upper-ocean convection, and thus reduced sea ice melt from below. Similarly, warmer air temperatures may actually increase ice extent by reducing new ice growth, which increases stratification and thus reduces convective overturning and upward heat flux (Zhang, 2007). In this scenario, the effect of reduced bottom melting is greater than the reduced ice growth. Thus, there is a net increase in ice thickness and extent. Liu and Curry (2010) have proposed that this increased stratification is achieved through an enhanced hydrological cycle that accompanies the warming. At the same time, the expected increase in precipitation accompanying atmospheric warming (Bracegirdle et al., 2008) could enhance snow ice production rates, leading to thicker ice and increased ice extent (Fichefet and Morales Maqueda, 1999; Zhang, 2007).

Whatever the precise mechanisms, the feedbacks among precipitation, ice growth, and upper-ocean stability and upward heat transport are critical for understanding present behavior of Antarctic sea ice and its future trajectory. And here, field measurements are sorely lacking. For example, while we have a relative wealth of data on snow depth, there are almost no measurements of precipitation and snow accumulation and redistribution on sea ice (Leonard and Maksym, 2011), and few continuous, season-long observations of ice-ocean interactions exist.

THICK OR THIN?

The processes described above most directly control ice thickness rather than extent. In fact, ice thickness may be even more sensitive to climate change than extent (Arzel et al., 2006). Ice thickness measurements made under Arctic sea ice since the late 1950s by US and British navy submarines show that perennial ice thickness there has declined by as much as 40% (Rothrock et al., 1999). No such large-scale, long-term record is available for the Antarctic.

Until recently, our entire knowledge of Antarctic ice thickness came from icebreaker expeditions into the pack ice. Techniques used include visual observations (ASPeCt protocol; Worby et al., 2008a), drilling transects (e.g., Jeffries et al., 1998), and electromagnetic sounding from the surface, ships, or helicopters (e.g., Haas, 1998). Although these methods have provided the most accurate and detailed data on ice thickness distribution and morphology available, they are severely limited in their spatial coverage.

From over 80 voyages in all regions and all seasons, thousands of visual assessments of ice conditions and thickness have been made since the 1980s, comprising the most extensive record of ice thickness in Antarctica. While the data are insufficient to determine trends, they represent the only circumpolar estimates of ice thickness (Figure 5), and the "best" available data for evaluating models and satellite data products. However, the underway observations are not without limitations; for example, ice thickness is visually (and subjectively) estimated from afar, ridged and thick ice may be significantly underestimated (Worby et al., 1998), and observations are local only to the ship's track (which is often biased toward thinner ice).

Spaceborne altimeters offer the tantalizing possibility of monitoring sea ice thickness and its temporal variability over large scales. Reasonable estimates of large-scale ice thickness, and its recent decline, have been obtained in the Arctic with both radar altimeters (Laxon et al., 2003) and the ICESat laser altimeter (Kwok et al., 2009). In the Antarctic, however, the thick snow cover shields the ice below from direct measurement (Giles et al., 2008; Zwally et al., 2008; Yi et al., 2011), requiring independent information on snow thickness (and density) to determine ice thickness. Snow depth on sea ice can be estimated using passive microwave data from satellites (Markus and Cavalieri, 1998), but this method can underestimate snow depth in deformed areas by a factor of two to three (Worby et al., 2008b). Where the snow is thick enough to depress the ice surface near sea level, the resultant uncertainty in satellite-derived ice thickness can be substantial.

In fact, sea ice drilling data from around the Antarctic sea ice zone show that the freeboard (the height of the snow/ice interface above sea level) is very often near zero (Maksym and Markus, 2008). This observation suggests that ICESat measurements of surface elevation provide a better proxy for snow depth than for ice thickness. The measurements are then a proxy for a minimum ice thickness—if the snow were any thicker, it would flood and freeze into snow ice; if the snow depth were less, then the actual ice thickness would be greater. Even so, this estimate exceeds the in situ observed ice thickness by a factor of two (Figure 5).

So where does the truth lie? That is not so easy to determine. However, point measurements of ice draft made by moored upward-looking sonars in the Weddell Sea suggest that the truth may lie somewhere in between (Strass and Fahrbach, 1998). While the in situ observations may be biased toward thinner ice, there are large potential errors in the satellite estimates due to uncertainty in snow depth and ice density. Also, satellites likely underestimate the fraction of thin ice, as it is difficult to distinguish from open water. Field data from the two International Polar Year Antarctic sea ice research cruises show that if corrections for the snow cover are made based

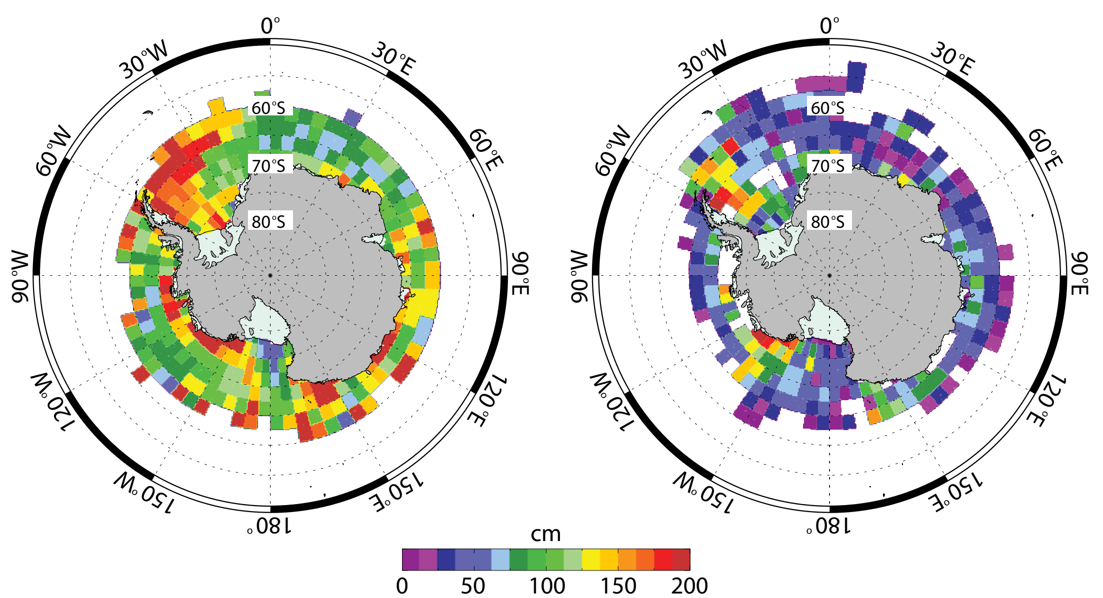

Figure 5. Estimates of the circumpolar Antarctic mean sea ice thickness distribution for October to November 2003 from ICESat satellite altimetry (left) and ship-based visual observations (ASPeCt data, at right). The ICESat ice thickness is a lower bound, determined by assuming that the surface elevation is almost entirely snow. While the patterns of ice distribution are similar, the ICESat estimate is almost twice that of the in situ observations. *ICESat data courtesy of Ron Kwok*

on local conditions, good agreement between in situ and satellite-derived ice thickness can be achieved (Worby et al., 2011; Xie et al., 2011). However, these corrections vary regionally, depending on local ice conditions.

Until we have sufficient in situ measurements of snow depth, ice thickness distributions, and snow and ice properties, this conundrum will remain. These data are of critical importance. As Antarctic sea ice is largely seasonal, ice thickness provides a measure of total sea ice production and, hence, a measure of the surface salinity flux in winter, the freshwater input to the ocean in summer, and total heat loss to the atmosphere. With uncertainties in these quantities of 50–100%, we lack the ability to properly evaluate models—models that cannot yet capture either the current state of Antarctic sea ice cover or its trends. This limitation in turn affects our confidence in the accuracy of future projections.

THE WAY FORWARD

Much has been learned about Antarctic sea ice and its interactions with the atmosphere and ocean over the past 20–30 years since ship-based research expeditions into the pack ice began in earnest. Until now, most studies have provided only a "snapshot" of ice properties and processes. However, we are now on the cusp of a new era in observational capability for monitoring the Antarctic sea ice zone. Autonomous drifting and ice-tethered platforms have matured such that continuous measurements of ice processes and air-ice-ocean interactions are now becoming possible.

But even as the Arctic has more than a decade of widespread deployments of autonomous observing platforms, the Antarctic remains relatively untrodden ground. While drifting buoys have been deployed sporadically for over 20 years, they have been too few and far between to provide a complete picture of ice drift behavior, let alone detection of trends. Only a handful of ice mass balance buoys, and almost no other platforms that can autonomously monitor air-ice-ocean interactions, have been deployed in the Antarctic. Autonomous vehicles are now capable, however, of reliable operation within the challenging environment of the Antarctic pack ice (e.g., Banks et al., 2006). These platforms have the potential to revolutionize our understanding of this remote and undersampled sea ice environment.

Such missions represent only the first step toward sustained observation of the Antarctic sea ice zone. With many secrets of the Antarctic sea ice cover yet to be revealed, the crucial need for direct observations from icebreakers like RVIB *Nathaniel B Palmer* remains. Indeed, our uncertain knowledge of the properties and behavior of the ice and its interactions with the climate system strongly argues for renewed impetus to measure and monitor the ice cover—with continuation and enhancement of field-based measurement programs—so that improved projections of the future state of the ice cover, and the potential impacts on its biota and the environment around it, can be made. Then perhaps, unlike the Arctic canary, we can know the future of the Antarctic ice pack before it has met its fate.

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