Recent changes in Arctic sea ice and ocean circulation

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Arctic sea ice and the upper ocean are important components of the climate system. Floating sea ice changes the surface albedo, insulates the ocean from heat loss, and reduces the exchange of momentum and gases such as water vapor and CO₂ between the ocean and atmosphere. Salt ejected by growing sea ice alters the density structure and modifies the circulation of the ocean. Regional climate changes affect the sea ice characteristics and these changes can feed back on the climate system, both regionally and globally.

Arctic sea ice has undergone rapid and dramatic changes over the satellite era (since 1978), characterized by a decline of roughly 14% per decade in pan-Arctic September sea ice extent (Comiso and Nishio 2008; Cavalieri and Parkinson 2012), substantial thinning (Rothrock et al. 1999; Kwok and Rothrock 2009) (Figure 1), a transition from multiyear to first-year ice (Kwok et al. 2009; Maslanik et al. 2011) (Figure 2), and longer melt seasons (Perovich and Polashenski 2012; Stroeve et al. 2014). The expectation that current trends will continue portends a future with

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In this edition, we focus on describing and understanding the interactions between ocean circulation and sea ice. In addition to atmospheric circulation and radiative changes, ocean circulation variability influences sea ice cover, while changes in sea ice can alter the ocean circulation beyond the immediate polar regions. In both the Arctic and Antarctic, the subsurface ocean contains warmer water, which can melt ice when mixed to the surface, while fresh surface waters reinforce stratification and insulate the ice from the subsurface heat. These two regions have very different ocean circulations. While the Southern Ocean is dominated by the upwelling and sinking associated with the meridional overturning circulation, the more stratified Arctic exchanges freshwater and heat with the global ocean through wind-driven gyres. The articles allow us to compare the different processes that influence variability in sea ice loss, using evidence from both observations and models. They also highlight new opportunities, such as the upcoming NASA ICESat-2 mission and the Sea-Ice Model Intercomparison Project of CMIP6, which may improve our understanding of the relative roles of ocean and atmospheric processes in driving polar sea ice changes.

Figure 1. Bar chart shows the mean sea ice thickness (m) in six Arctic regions for four periods (1958–1976, 1993–1997, 2003–2007, 2011-2015). Thicknesses have been seasonally adjusted to September 15.

larger expanses of open ocean, longer durations of ice-free conditions, and a more accessible Arctic. It is widely expected that Arctic sea ice coverage will continue to decline over the coming decades (Stroeve et al. 2012a), resulting in a seasonally ice-free Arctic before the end of the century.

The causes of ice loss are complex, involving changes in atmospheric and oceanic heat, freshwater and momentum fluxes, and internal feedbacks that result from the sensitivity of the coupled Arctic atmosphere–ice–ocean system to changes in the ice and snow cover (Serreze and Barry 2011; Stroeve et al. 2012b; Wassmann and Lenton 2012). Melting sea ice has consequences. If the sea ice disappears, heat from the sun – which mostly reflects off high-albedo ice surfaces and back into space – instead would be absorbed by the ocean, further accelerating the warming of the Arctic. And more warming means more ice melt. This is referred to as the ice-albedo feedback. The impacts of sea ice loss are potentially far reaching: ice cover anomalies affect atmospheric circulation patterns locally, and probably remotely (Grassi et al. 2013), and possibly even exert a significant influence on mid-latitude weather patterns; although the evidence for this is still equivocal (Francis and Vavrus 2012; Barnes 2013; Screen and Simmonds 2013).
Remarkably, the signal of global warming is amplified in the Arctic (known as Arctic amplification or AA), particularly during late summer and early fall when sea ice has declined most rapidly (Stroeve et al. 2012b). The period of AA began around 1990 (Cohen et al. 2014). AA results from both local and remote factors, amplified by a variety of positive feedbacks (e.g., Pithan and Mauritsen 2014; Burt et al. 2016). Decreased sea-ice extent (~40% summer loss in 30 years) reduced the surface albedo and air-ocean insulation, thereby increasing latent and sensible heat fluxes into the atmosphere, which increases surface temperatures and moistens the Arctic atmospheric boundary layer. Decreases in ice thickness (~75% in 30 years; Lang et al. 2017) also augment fluxes of heat and moisture into the atmosphere and allow the ice to be more responsive to mechanical forcing by winds and currents. Uncertainty about the causes of AA contributes to the challenge of identifying how much of the amplified warming signal is driven by local changes and feedbacks versus remote influences (Screen and Simmonds 2013). This distinction is relevant to the debate on possible Arctic and mid-latitude linkages; if a significant portion of AA is driven remotely, it may be partly viewed as a response to, rather than a forcing of, changing mid-latitude weather patterns.

Arctic Ocean circulation is important to climate by controlling the circulation of sea ice, mixing and exporting freshwater, and providing heat to melt ice. Under significant wind at short timescales, ice moves at ~2° and ~20° to the right of the wind direction (Thorndike and Colony 1982; Nansen 1902). However, averaged over many wind events, the ice velocity (Figures 3c and 3d) tends to match the surface geostrophic current (Figures

Figure 2. Decline in multiyear sea ice coverage (103 km^2) since 1999 from satellite scatterometer data (NASA’s QuikSCAT and ESA’s ASCAT).
Figure 3. Figure shows ocean controls on ice motion, including the feedback of ice ocean stress, $\sigma_{io}$, on sea surface gradient and ice velocity through Ekman transport (a) under high winds when ice velocity is greater than ocean velocity, and (b) under low winds when ice velocity is less than ocean velocity. Average Arctic sea ice velocity during a low AO period (c) and a high AO period (d) compared to sea level pressure (SLP) and geostrophic surface currents during a low AO period (e) and a high AO period (f) compared to dynamic ocean topography (DOT) (Morison et al. 2012). The red lines indicate the sea ice velocity zero vorticity lines in 2004-2006 (c & e) and 2007-2009 (d & f). The green line in (d) is the 2004-2006 zero vorticity line from (c) to illustrate the shift to a more cyclonic circulation in 2007-2009.

3e and 3f) because both ice and water move to the right of the gradient of sea surface height. And both are driven in roughly the direction of the average geostrophic wind because convergence of Ekman transport tilts the sea surface up to the right of the wind. For example, the average winds of the Beaufort High converge near surface water to dome the sea surface and form the anticyclonic Beaufort Gyre (Figure 3a). The gyre is stabilized by the combination of internal ice stresses and the ice-ocean stress related to the departure of ice velocity from the geostrophic water velocity (McPhee 2008). When winds drop (Figure 3b), the ice is slowed to below the geostrophic current velocity by internal ice stress ($\sigma_i$). Ice-ocean stress ($\sigma_{io}$) then becomes cyclonic, and Ekman transports act to decrease doming, which reduces water velocities until they match the ice velocity. The opposite equilibration occurs when winds become high.

Similarly, when the Icelandic Low extends into the Russian side of the Arctic Ocean, the resulting Ekman transport divergence creates a sea surface trough and cyclonic circulation (Sokolov 1962; Morison et al. 2012). Between this cyclonic circulation and the Beaufort Gyre lies the transpolar drift (TPD) of sea ice across the Arctic
Ocean from the Pacific side to the Atlantic side. The TPD aligns with the current due to the front between Pacific-derived waters in west longitudes and Atlantic-derived waters in east longitudes.

The relative strengths of these three features of ice and water circulation define modes of circulation related to global climate indices. In the anticyclonic mode, the anticyclonic Beaufort Gyre is dominant when the Beaufort High is strong (Sokolov 1962) or equivalently when the Arctic Oscillation Index (AO) is low (Morison et al. 2012). In the anticyclonic mode, the anticyclonic Beaufort Gyre is dominant when the Beaufort High is strong (Sokolov 1962) or equivalently when the Arctic Oscillation Index (AO) is low (Morison et al. 2012). In the cyclonic mode, the cyclonic circulation along the Russian side of the Arctic Ocean is dominant when the AO is high (Morison et al. 2012).

Through its effect on modes of Arctic Ocean circulation, global climate impacts the pathways of freshwater and availability of heat to melt ice. In the Arctic Ocean, the warm saline inflow of Atlantic water through the eastern Fram Strait is mixed with Eurasian runoff, freshwater contained in the inflowing Pacific water, and net precipitation-evaporation. The pathways of the resulting low-salinity, near-freezing mixture along with low-salinity sea ice are dependent on the modes of circulation.

The anticyclonic mode is thought to converge ice and freshwater in the upper ocean resulting in slowed flushing of both to the sub-Arctic seas. Freshwater collects near the surface of an expanded and strengthened Beaufort Gyre (Proshutinsky and Johnson 1997; Proshutinsky et al. 2001; Proshutinsky et al. 2009), and Eurasian runoff tends to move directly across the Eurasian Basin to exit Fram Strait. The TPD of near-surface freshwater to Fram Strait shifts the orientation clockwise and broadens into east longitudes (Figures 3c and 3e). This mode is favored by a low AO at interannual timescales; although at monthly scales, the Beaufort Gyre freshwater accumulation correlates with a positive AO with a two-month lag (Dewey et al. 2017).

In the cyclonic mode favored by a high AO, the ice motion is more divergent, which results in lower ice concentrations and arguably a stronger TPD with more rapid ice motion to the sub-Arctic seas (Kwok et al. 2013). Related to this, high winter AO correlates with reduced total ice extent the following summer (Rigor et al. 2002). The TPD of near-surface freshwater turns counterclockwise and shifts towards Canada (Figures 3d and 3f). The cyclonic circulation on the Russian side of the Arctic Ocean causes Eurasian runoff to flow eastward along the Russian coast as far as the Chukchi Sea to be mixed with Pacific water and injected into the Beaufort Sea halocline to increase freshwater content there at the expense of the Eurasian Basin (Morison et al. 2012). The diversion of Eurasian runoff to the east shrinks the cold-halocline of the Eurasian Basin that isolates the sea ice from the heat in the deeper Atlantic water (Steele and Boyd 1998).

Since reaching a maximum in 1989, in spite of a record minimum in 2010, the winter AO has been, on average, elevated one standard deviation above the average prior to 1989. This means that the possible increased Arctic Ocean outflows of freshened water and sea ice, which are important to climate because they increase the stratification of the sub-Arctic seas, are having a leveraged impact on the deep convection that is part of the global thermohaline circulation. In the extreme, this has led to the “Great Salinity Anomaly” and possible shut down of convection (Dickson et al. 1988). The strength of the outflow of cold low-salinity Arctic water along the East Greenland continental shelf may also impact ocean temperatures affecting the marine terminating glaciers of the Greenland ice sheet. There is also evidence that recent freshening of the whole water column of the Nordic sub-Arctic seas has freshened the entire North Atlantic (Curry et al. 2003).
References


An Arctic Ocean in transition

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Arctic sea ice plays an important role in the energy balance of our planet. Reductions in the amount of sea ice cover can lead to amplified warming in the Arctic (e.g., Serreze et al. 2009), with the potential to influence large-scale weather and ocean circulation patterns (e.g., Walsh 2014; Francis and Vavrus 2012; Jaiser et al. 2012), influence permafrost temperatures (Lawrence et al. 2008), and perhaps also Greenland melt (Stroeve et al. 2017). Sea ice is also crucial for polar ecosystems and marine biogeochemistry (e.g., Tynan 2015; Vancoppenolle et al. 2013), the livelihoods of coastal communities (Laidler et al. 2008), maritime activity (Emmerson and Lahn 2012), and resource extraction (Emmerson and Lahn 2012).

During the last four decades of satellite observations, the Arctic Ocean has gone through a period of profound transformation, with large reductions in the amount of sea ice remaining at the end of the summer melt seasons (e.g., Stroeve et al. 2012a; Serreze and Stroeve 2015). Since the late 1970s, the September sea ice extent has declined by more than 40%, dropping from about 7x106 km$^2$ in the 1980s to less than 4x106 km$^2$ in 2012, and the last decade has witnessed the ten lowest September sea ice extents recorded during the satellite data record. Forecasts submitted to this year’s June Sea Ice Outlook (SIO), which has been providing seasonal forecasts of September sea ice extent since 2008, predict that 2017 will likely, once again, see a September extent below 5x106 km$^2$. September mean extent has been below 5x106 km$^2$ for seven out of the last ten years.

While much attention has been paid to the summer ice loss, changes are not limited to summer. The last few winters have seen record low sea ice extents, with 2017 marking the third year in a row with a record low for the winter maximum extent. More surprising is the number of consecutive months between January 2016 through April 2017 with ice extent anomalies at least two standard deviations below the 1981-2010 mean (Figure 1). Additionally, October 2016 through April 2017 saw seven consecutive months with record low sea ice extents, something that had not happened before in the satellite record.

At the same time that the overall extent of the ice cover has declined, pronounced thinning of the multiyear ice cover has also been documented (Lindsay and Schweiger 2015), along with a transition of an Arctic Ocean from a predominantly multiyear ice regime to a seasonal one (Maslanik et al. 2007; 2011). The melt season is also starting earlier than it used to, and the ice is forming later, leading to longer periods of open water (Stroeve et al., 2014a). Together these changes have made the ice cover more vulnerable to melting out in summer, especially under atmospheric forcing conducive to extensive summer ice loss.
While ice loss is now occurring in all months of the year, larger changes in summer compared to winter amplifies the range of the seasonal cycle, increasing the seasonality (Figure 2). As larger parts of the Arctic Ocean become ice-free in summer, regional seas gradually change from having a perennial to a seasonal ice cover. The Barents Sea is already only seasonally ice covered, whereas the Kara Sea has recently lost most of its summer ice and is thereby starting to become a seasonally ice covered region (Onarheim et al. submitted). Most of the Arctic Ocean, however, still refreezes completely in winter, except for the Barents Sea that is rapidly losing its winter ice (Onarheim and Årthun, submitted). The Arctic sea ice is, thus, becoming more seasonal both in terms of increased range of the seasonal cycle (difference in maximum and minimum sea ice range) and in terms of new areas becoming seasonally ice covered.

Given the importance of sea ice in the climate system, numerous studies have looked to explain the drivers of this ice loss. Climate models are often used for these types of assessments in order to try to separate out natural climate variability from anthropogenic changes. The scientific consensus is that the current loss of the sea ice cover is attributed to two main factors: an overall long-term decline and thinning from warming as a result of increasing greenhouse gases (e.g., Notz and Stroeve 2016; Stroeve et al. 2012; Kay et al. 2011), together with shorter (interannual to decadal) changes from natural climate variability (e.g., Ding et al. 2017). These studies go further to implicate greenhouse gases as the dominant factor, estimating more than half (50-60%) of the current rate of ice loss is a direct result of increasing greenhouse gases, with the remainder a result of natural climate variability. However, despite the predominance...
of anthropogenic forcing, studies have also suggested the possibility of decade long periods of temporary sea ice recovery (Kay et al. 2011).

While predicting these more random events has proven challenging, forecasting large-scale atmospheric circulation anomalies, such as the Northern Annular Mode (NAM), may provide some seasonal- to decadal-scale predictability in ice conditions for specific regions of the Arctic (e.g., Yeager et al. 2015). Predictability on seasonal timescales has proved especially challenging given the chaotic nature of summer atmospheric circulation patterns. Sea ice forecasts of the September mean extent submitted each June, July, and August to the Sea Ice Prediction Network often fail to forecast the observed pan-Arctic September sea ice extent, especially when the observed extent falls far from the long-term trend line (e.g., Hamilton and Stroeve 2016; Stroeve et al. 2014b). Seasonal forecasts may become even more difficult to make in the future because of thinner sea

Figure 2. Seasonality index computed as the difference between the maximum and minimum extent normalized by the annual mean extent from the NSIDC Gridded Monthly Sea Ice Extent and Concentration dataset (Walsh et al. 2015) and the NSIDC Sea Ice Index (Fetterer et al. 2016).
ice cover and increased interannual variability. Climate models also indicate increased variability as the ice cover thins.

On the other hand, the ability to predict sea ice retreat and advance in the Chukchi Sea, using springtime atmospheric conditions and oceanic heat flux on seasonal timescales, shows promise (Serreze et al. 2016). In other regions of the Arctic, sea ice persistence may offer some predictability, whereas other regions show ice thickness and sea surface temperature persistence that can be exploited to improve seasonal ice forecasting, though this may change as the ice cover thins further (Cheng et al. 2016). Given the importance of the sea ice cover on ecosystems and human activities in the Arctic, there is an urgent need to advance sea ice predictions, not only in summer but also during the entire year.

On longer timescales, we often turn to climate models to provide future forecasts of sea ice conditions (Figure 3). However, climate models have, on average, failed to capture the pace of Arctic summer sea ice loss (Stroeve et al. 2012b, 2007; Massonnet et al. 2012). This has led to some speculation that a seasonally ice-free Arctic Ocean may be realized by the middle of this century. The failure of climate models to adequately simulate the observed rate of ice loss has led to studies that use subsets of models to make future forecasts (e.g., Snape and Forster 2014). However, internal climate variability limits the ability of any single model to reproduce the observed evolution of ice cover over the arbitrarily chosen time-period a model is evaluated for and does not necessarily provide a more accurate forecast (Stroeve and Notz 2015).

![Figure 3](image-url)

**Figure 3.** Observed (black line), modeled historical (gray dotted line), and future projections (colored solid lines) of September sea ice extent (10^6 km^2) under different climate forcing scenarios from CMIP5 climate models. Shading corresponds to one standard deviation (1σ) from the multi-model ensemble mean. The numbers in parenthesis correspond to the number of models used for the multi-model ensemble mean and 1σ.
New insights are likely to be gained with model output from the 6th phase of the Coupled Model Intercomparison Project (CMIP6). A central aim of the CMIP6 Sea-Ice Model Intercomparison Project (SIMIP) is to better understand biases and errors in sea ice simulations so that we can improve our understanding of the likely future evolution of the sea ice cover and its impacts on global climate. To reach this goal, a community-defined set of model outputs have been recommended that will better characterize the heat, momentum, and mass budget of Arctic sea ice (Notz et al. 2016). This will allow for better quantification of the role of internal variability, external forcing, and model deficiencies.

The loss of the Arctic sea ice cover has captured the world’s attention. What does the future hold for Arctic sea ice? While models may not get all features of Arctic sea ice changes correct, they are unanimous that Arctic sea ice will continue to decline as the concentration of greenhouse gases in the atmosphere increase. The linear relationship between cumulative CO₂ in the atmosphere and summer sea ice decline (Notz and Stroeve 2016) provides a clear limit on the additional amount of CO₂ society can add to the atmosphere before the Arctic Ocean becomes ice free in summer. At current emission rates, this will happen by 2045, though the uncertainty range of 20 years from natural climate variability gives a time span of 2035 until 2055.

Scientists are increasingly charged with society relevant science. Yet despite numerous efforts to document the impacts of sea ice loss on ecosystems, coastal communities, local and global ocean circulation, and weather patterns, foreseeing the consequences of an Arctic Ocean ice free in summers remains difficult. It is especially difficult to link current sea ice changes with impacts on mid-latitude weather despite growing awareness in the public that sea ice plays a role (Hamilton and Lemcke-Stampone 2014). While studies are not yet in agreement that current sea ice changes have already influenced mid-latitude weather, impact is expected to increase as the Arctic Ocean transitions towards a seasonal ice cover, affecting not just Arctic communities but also communities at lower latitudes.

References


The 2017 WCRP/IOC Sea Level Conference produced a statement, based on input from the broad community during the meeting. The statement is open to be adopted by conference participants as well as those who did not participate.

Signatories of the statement do so as individuals and not on behalf of their institutions.

Sign by August 31
Improving our understanding of Antarctic sea ice with NASA's Operation IceBridge and the upcoming ICESat-2 mission

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Antarctic sea ice is a crucial component of the global climate system. Rapid sea ice production regimes around Antarctica feed the lower branch of the Southern Ocean overturning circulation through intense brine rejection and the formation of Antarctic Bottom Water (e.g., Orsi et al. 1999; Jacobs 2004), while the northward transport and subsequent melt of Antarctic sea ice drives the upper branch of the overturning circulation through freshwater input (Abernathy et al. 2016). Wind-driven trends in Antarctic sea ice (Holland & Kwok 2012) have likely increased the transport of freshwater away from the Antarctic coastline, significantly altering the salinity distribution of the Southern Ocean (Haumann et al. 2016). Conversely, weaker sea ice production and the lack of shelf water formation over the Amundsen and Bellingshausen shelf seas promote intrusion of warm Circumpolar Deep Water onto the continental shelf and the ocean-driven melting of several ice shelves fringing the West Antarctic Ice Sheet (e.g., Jacobs et al. 2011; Pritchard et al. 2012; Dutrieux et al. 2014). Sea ice conditions around Antarctica are also increasingly considered an important factor impacting local atmospheric conditions and the surface melting of Antarctic ice shelves (e.g., Scambos et al. 2017). Sea ice formation around Antarctica is responsive to the strong regional variability in atmospheric forcing present around Antarctica, driving this bimodal variability in the behavior and properties of the underlying shelf seas (e.g., Petty et al. 2012; Petty et al. 2014).

Satellite passive microwave data have provided a long-term assessment of Antarctic sea ice coverage (i.e., concentration) since the 1970s. Unfortunately, our understanding of Antarctic sea ice is limited by inadequate observations of its thickness distribution and overlying snow cover. High-resolution ice-ocean models (e.g., Holland et al. 2014; Zhang et al. 2014) and model-based sea ice reconstructions (e.g., Massonnet et al. 2013) provide important insight into the potential ice thickness variability. However, these models are being calibrated/validated against the limited observations of Antarctic sea ice thickness currently available and have arguably received less attention compared to model calibration efforts in the Arctic.

The record low Antarctic sea ice cover observed in the austral spring/summer of 2016/2017, in stark contrast to the small, but positive, long-term trend (e.g., Turner et al. 2017), has presented the Antarctic sea ice community with new questions regarding the potential variability and future trends in Antarctic sea ice and their consequences for the Southern Ocean and global climate system.

Antarctic sea ice: A challenging medium to observe
Radar and laser altimetry have been used relatively successfully in the Arctic to derive basin-scale estimates of sea ice thickness (e.g., Kwok and Cunningham 2008;
The altimetry technique uses measurements of sea ice freeboard — the extension of sea ice above the local sea level — and estimates of snow depth to derive sea ice thickness (assuming the ice is in hydrostatic equilibrium). Satellite altimetry of Antarctic sea ice thickness has been hindered by uncertainty in the snow depth used to convert freeboard into thickness and, for radar altimetry (e.g., from ESA’s CryoSat-2 satellite), uncertainty in the penetration of the radar signal into the overlying snow layer (e.g., Giles et al. 2008; Willatt et al. 2010).

Passive microwave derived snow depths (e.g., Markus & Cavalieri 1998), provide the only pan-Antarctic snow depth estimate currently available. The reliability of this data over different ice surfaces, however, is still uncertain. Snow accumulation models forced by reanalyses have also been used to estimate Antarctic snow depth (e.g., Maksym and Markus 2008). But due to the strong winds around Antarctica and the large coverage of leads (fractures) within the Antarctic ice pack, the fraction of snow lost to leads is thought to be significant (Leonard and Maksym 2011), increasing the challenge of this approach compared to the Arctic. The increased prevalence of flooded and refrozen snow over Antarctic sea ice (e.g., Massom et al. 2001) is also thought to increase the remote sensing challenges compared to the Arctic.

The only pan-Antarctic sea ice thickness dataset publicly available is based on laser altimetry derived freeboard observations from NASA’s Ice, Clouds, and Land Elevation Satellite (ICESat), using the assumption that the ice-snow interface is at sea level, i.e., that the freeboard (draft) is all snow (ice) (Kurtz and Markus 2012). The passive microwave snow depth estimates have been used to estimate sea ice thickness regionally, such as in the Weddell Sea using ICESat data (Zwally et al. 2008; Yi et al. 2011). This approach, however, has not been extended to the entire Antarctic sea ice pack. Empirical relationships between freeboard and ice thickness have also been invoked as a more relevant solution to the thick/deformed Antarctic sea ice regimes (Xie et al. 2013).

Alternative sea ice thickness datasets include: ship-based observations from the Antarctic Sea ice Processes and Climate (ASPeCt) dataset, 1980-2005 (Worby et al. 2008); upward looking sonar (ULS) measurements of ice draft in the Weddell Sea, 1990-2008 (Behrendt et al. 2013); and more localized measurements (e.g., electromagnetic sensing and in situ drill holes). More recently, autonomous underwater vehicle (AUV) draft measurements in localized regions around Antarctica have been used to suggest that Antarctic sea ice might be thicker than previously assumed (e.g., Williams et al. 2015).

In the Arctic, the age of the ice is often used as a proxy for ice thickness (Maslanik et al. 2011; Tschudi et al. 2016). The Antarctic sea ice pack is generally younger, however, limiting the potential for age to act as a useful proxy for thickness. The limits of seasonal ice growth are also uncertain, given the strong variability and uncertainty surrounding atmospheric and oceanic forcing around Antarctica. A ‘Frontal Ice Zone’ — a region of more deformed, older ice adjacent to the sea ice edge — was recently suggested by Nghiem et al. (2016), based on the analysis of satellite radar backscatter data. This utilizes the principle that older/rougher ice tends to exhibit higher radar backscatter compared to younger, flatter ice. However, more direct measurements of sea ice thickness and roughness are needed to validate this idea.

**Operation IceBridge**

A relatively new data source for Antarctic sea ice research is the suite of measurements provided by NASA’s Operation IceBridge (OIB) mission. OIB began in 2009 with the objective to bridge the gap between ICESat and the upcoming launch of NASA’s ICESat-2 satellite (Markus et al. 2017). OIB has collected data over the western Antarctic sea ice pack in austral spring since 2009 (October/November, all years except 2015). The regions profiled by the OIB Antarctic sea ice campaigns include myriad ice types, such as the thick/deformed ice of the western Weddell Sea, and the younger, drifting ice over the Ross shelf sea. The OIB Antarctic sea ice flight lines
The primary instrument onboard OIB is the conically scanning Airborne Topographic Mapper (ATM) laser altimeter (Krabill et al. 2002), which profiles the ice surface across a width of ~250 m. Each laser shot has a horizontal footprint resolution of ~1 m and a vertical accuracy of ~10 cm. An ultra wide-band snow radar (Panzer et al. 2013) that flies with OIB is used to estimate snow depth, while coincident optical imagery helps identify leads/cracks in the ice cover; enabling the calculation of ice freeboard and thus thickness. Note that the increased prevalence and size of leads in the Antarctic ice pack make freeboard measurements easier than in the Arctic.

The Antarctic OIB data have arguably been underexploited due to the challenges of Antarctic sea ice altimetry described earlier (e.g., uncertain penetration of the OIB snow radar into the snow cover). An approach to derive snow depth from the OIB snow radar and the (snow plus ice) freeboard from the OIB ATM was presented by Kwok & Maksym (2014) for the 2010/2011 data in the Weddell and Bellingshausen seas, providing the first assessment of snow depth and freeboard variability from the Antarctic OIB data. Unfortunately, a lack of coincident in situ data preclude a more thorough understanding of the uncertainties of the snow radar derived snow depths, as is currently underway for the Arctic OIB data (Kwok et al., 2017).

Figure 1. (a) Antarctic Operation IceBridge (OIB) sea ice flights in austral spring (October/November) overlaid on the ICESat ice thickness estimates (2003-2008 October/November mean; Kurtz and Markus 2012). The red stars indicate the location of the Weddell Sea moorings (various times of operation between 1990 and 2008). (b) Mean height of surface topography features detected from the 2009-2014 OIB ATM sea ice data overlaid on the mean ASCAT radar backscatter (σ) over the OIB sea ice campaign time periods. Ice topography processing updated for the Antarctic from Petty et al. (2016).
Despite the challenges associated with deriving Antarctic snow depth and thickness, the OIB ATM laser data can still be used to directly assess the ice (plus snow) freeboard and topography. The OIB ATM data have recently been used to assess Arctic sea ice topography (Petty et al. 2016) and its contribution to the atmospheric form drag coefficient (Petty et al. 2017). Note that these topography estimates involve the explicit detection of surface features (e.g., pressure ridges), as opposed to the more bulk metric of surface roughness (normally calculated as the standard deviation of elevations within a given window). Petty et al. (2017) also demonstrated strong correlations between satellite radar backscatter data from the Advanced Scatterometer (ASCAT) and the OIB ATM derived form drag coefficients, enabling the OIB results to be extrapolated outside of the regions directly profiled — an exciting prospect for the Antarctic OIB data.

An extension of this analysis is currently underway for the Antarctic OIB ATM sea ice data. A preliminary analysis of the OIB Antarctic ice topography is shown in Figure 1b, overlaid on the ASCAT backscatter data. The results demonstrate a broadly consistent spatial pattern of ice topography, with higher topography in the western Weddell Sea and lower topography in the Ross Sea, as expected. The ice topography results, along with other metrics of ice surface roughness, can be used to assess where we might expect smooth/thin or rough/thick ice to be prevalent, helping guide the analysis of satellite altimetry retrievals. Petty et al. (2016) also demonstrated that the OIB ice topography data can provide a reasonable, albeit crude, assessment of the total sea ice thickness, while Kwok & Maksym (2014) found a strong correlation between their OIB estimates of ice surface roughness and snow depth, which they suggested may be due to the role of deformed ice acting as a catchment for snow.

The upcoming ICESat-2 mission
An exciting new avenue for Antarctic sea ice research will be provided by the launch of NASA's ICESat-2 mission (scheduled for launch in 2018). The Advanced Topographic Laser Altimeter System (ATLAS) on-board ICESat-2 is a multi-beam, photon counting laser altimeter, featuring six linear profiling beams that cover a total width of ~6 km (Markus et al. 2017; Figure 2) The three pairs of beams, together with its continuous operation, will result in about nine times better spatial coverage than the ICESat satellite. A surface elevation will be estimated from the return photon distribution — commonly referred to as a Figure 2. Schematic of Antarctic sea ice profiling with the NASA Operation IceBridge laser altimeter (ATM) and snow radar, and the expected profiling from the upcoming ICESat-2. Not to scale.
photon cloud. Measurements of sea ice freeboard (Arctic and Antarctic) are a direct mission requirement that will be provided after launch by the ICESat-2 project office. Similar to the OIB mission, one of the primary research priorities of ICESat-2 will be estimating sea ice thickness, with snow depth again expected to provide the primary source of uncertainty.

For the analysis of ice topography/roughness, the lower horizontal resolution laser footprint of ATLAS (~15 m) compared to the OIB ATM (~1 m) means it will not directly resolve the smaller surface roughness features. The Multiple Altimeter Beam Experimental Lidar (MABEL) has been deployed prior to the launch of ICESat-2 to provide relevant data for mission testing and development, especially in regards to the interpretation of the ATLAS photon cloud for sea ice elevation retrievals. Kwok et al. (2014) demonstrated that the MABEL and ATM retrievals co-vary along track and are similarly sensitive to surface elevation variability. MABEL (and thus hopefully ICESat-2) can detect surface elevation variability at this 15 m resolution, but the impact of ice topography on the photon distributions is still an open research topic.

Summary and suggestions for future research priorities

NASA’s Operation IceBridge and the upcoming ICESat-2 mission will provide crucial new data to help us understand the state of the Antarctic sea ice pack. One of the biggest remaining uncertainties is the snow depth on Antarctic sea ice. More sophisticated models of snow accumulation/redistribution could help improve the reanalysis derived snow depth estimates and provide critical context for the passive microwave and OIB derived snow depths being produced. Coincident in situ Antarctic sea ice campaigns to help validate these data would be invaluable.

Sea ice models provide a vital tool for understanding the variability and drivers of Antarctic sea ice mass balance, with their sophistication improving rapidly over recent years. For example, a sophisticated form drag parameterization scheme has recently been incorporated into the sea ice model CICE (Tsamados et al. 2014) to improve the atmosphere-ice-ocean coupling; however, the new parameterization scheme contains several free parameters that need to be better constrained with observations. Satellite emulators included in models (e.g., simulating sea ice freeboard) could provide an important bridge between the modeling and observational communities, along with more sophisticated calibration efforts currently underway. The depth of sea ice observations possible with OIB and the upcoming launch of ICESat-2 should provide the community with a wealth of data for advancing our knowledge of Antarctic sea ice.

References


Sea ice in the Southern Ocean exhibits substantial year-to-year variability in addition to a small long-term increase in recent decades (Comiso and Nishio 2008; Eisenman et al. 2014; blue line in Figure 1). The modest increase in Southern Ocean sea ice is a stark contrast to the significant declines in Arctic ice extent seen over the same time period over which global temperature has risen (Serreze et al. 2015). The delayed anthropogenic warming of the Southern Ocean relative to surface temperature increases in other regions is robustly predicted by climate models, and the underlying mechanism is well understood; the energy input to the surface of the Southern Ocean by anthropogenic forcing is transported equatorward by the basic-state ocean overturning circulation as surface water advects northward and is replaced by upwelled water that has not been exposed to the surface since long before the influence of anthropogenic forcing (Armour et al. 2016). Thus, one expects the delayed warming of the Southern Ocean to result in a muted signal of sea ice decline relative to the Arctic that may not be detectable above natural variability. While the observed increase in Southern Ocean sea ice extent is within the range of natural variability simulated by unforced coupled climate models (Polvani and Smith 2013), coupled climate models unanimously simulate Southern Ocean sea ice reductions under current levels of anthropogenic forcing (Turner et al. 2013).

The question remains as to what processes drive the observed year-to-year variability and long-term trends in Southern Ocean sea ice. A myriad of coupled atmosphere/ocean/cryosphere processes have been proposed:

- Stratospheric ozone depletion is leading to sea ice expansion as a result of intensified surface westerlies over the Southern Ocean (Thompson and Solomon 2002), and the associated equatorward Ekman ice transport (Turner et al. 2009).

- Ozone depletion is leading to sea ice retreat at longer timescales induced by the same enhanced surface westerlies (Bitz and Polvani 2012; Sigmond and Fyfe 2010), due to increased upwelling of warmer subsurface waters (in the salinity stratified Southern Ocean) by Ekman suction (Ferreira et al. 2015).

- Sea ice expansion from reduced interaction between the surface ocean and the subsurface warm waters due to increased ocean stratification as a result of either freshwater discharge from Antarctic glaciers (Bintanja et al. 2013), decreased surface salt input via brine rejection (Zhang 2007), or increased regional precipitation (Liu and Curry 2010).

- Wind changes of unknown origin are altering ice production in regions of anomalous wind convergence (Zhang 2014) and ice drift (Holland and Kwok 2012).
Clearly, there is no consensus on a mechanistic understanding of the processes that drive Southern Ocean sea ice variability. Furthermore, much of our understanding of Southern Ocean sea ice variability is borne from the use of coupled model simulations in which the relative roles of ocean circulation, atmospheric circulation, radiative processes, and their mutual interaction with sea ice growth and decay may not represent reality. The modeling of these processes is especially difficult in the Southern Ocean given significant model biases in simulating the mean state properties of the region. Specifically, the thermal stratification of the Southern Ocean (Kostov et al. 2016) and the amount of solar radiation reaching the Southern Ocean’s surface (Trenberth and Fasullo 2010) differ drastically between models and are biased relative to the observations. Given the biases in the simulated basic properties of the Southern Ocean in models, the physical processes that drive sea ice variability in coupled models may not occur in nature, while mechanisms that are absent in model simulations may be of central importance to the variability of Southern Ocean sea ice.

Figure 1. Time series of observational based estimates of anomalous Southern Ocean sea ice area (blue – from the National Sea Ice Data center (Fetter and Fowler 2006)) and energy fluxes into the Southern Ocean. The black line shows the $F_{WALL}$ expressed as the resultant average heating over the polar cap in Wm$^{-2}$ (monthly data has been low-pass filtered with a cut off period of 6 months). The anomaly in Southern Ocean average net shortwave radiation at the TOA (red) and outgoing longwave radiation (green) are calculated from CERES (Wielicki et al. 1996) energy balanced and filled data.
**A pathway forward: The energetic signature of Southern Ocean sea ice variability**

The growth and decay of Southern Ocean sea ice is influenced by ocean processes, atmospheric circulation, and radiative processes. We argue that the primary driver of sea ice variability can be identified from analysis of the (anomalous) energy budget of the Southern Ocean associated with an ice loss event, as summarized in Figure 2a-c. The energy budget of the Southern Ocean climate system (defined as the region between 55°S and 70°S) is composed of three primary terms: (i) the net radiative input into the top of atmosphere (RAD$_{TOA}$); (ii) the energy flux across the interface between the ocean-sea ice and the atmosphere (SHF) and; (iii) the poleward atmospheric energy flux across into the region (F$_{WALL}$).

We now consider the resultant changes in the above energy fluxes associated with three different sequences of events leading to a sea ice loss event (reverse signs for an ice growth event), with each driven by an initial change in a different component of the system:

**Mechanism One: Ocean Driven Sea Ice Loss**

Changes in ocean dynamics — such as thermal stratification, vertical energy fluxes by enhanced upwelling, or lateral energy fluxes — lead to enhanced energy input to the surface ocean resulting in ice loss (Figure 2a). As a result, the warmed and exposed ocean surface heats the atmosphere via enhanced upward

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**Figure 2.** Schematic of the different proposed mechanisms of Southern Ocean sea ice loss and the oceanic/atmospheric/radiative energy fluxes associated with each mechanism. Numbers indicate the chronological order of events in each mechanism: sea ice driven by (left panel) ocean circulation changes resulting in upward surface energy fluxes (blue arrow), (middle panel) atmospheric circulation (purple arrow), and (right panel) radiative processes (red arrow). The lower panels show coupled climate model simulations where sea ice loss events are driven by each of the above mechanisms. Each line shows the energy flux anomaly (Wm$^{-2}$) into the Southern Ocean region (spatially averaged poleward of 55°S) associated with a 2 standard deviation (σ) sea ice loss event.
turbulent energy fluxes (SHF). The atmosphere heats up and exports the anomalous energy via circulation anomalies (a negative $F_{\text{WALL}}$ anomaly) and radiatively via the Planck feedback. We note that the initial perturbation to the oceanic circulation may have been provided by a local wind stress anomaly and that we will distinguish this case from the changes in atmospheric energy transport (discussed below).

**Mechanism Two: Atmospheric Circulation Driven Sea Ice Loss**

Atmospheric circulation anomalies (originating as, for example, internal modes of tropically forced anomalies (Ding and Steig 2013) or mid-latitude circulation changes) drive an enhanced energy input to the atmosphere above the Southern Ocean (i.e., a positive $F_{\text{WALL}}$ anomaly; Figure 2b). The atmosphere heats up and energy is fluxed downward to the surface via enhanced downwelling longwave radiation and reduced upward sensible energy fluxes resulting in ice loss (a negative SHF). There is a loss of RAD$_{\text{TOA}}$ via the Planck feedback.

**Mechanism Three: Radiation Driven Sea Ice Loss**

Changes in local atmospheric optical properties (e.g., a reduction in Southern Ocean cloud cover) and surface albedo result in enhanced radiative input to the system (RAD$_{\text{TOA}}$), heating both the atmosphere and the surface — via an enhanced downward radiative surface energy flux — resulting in ice loss (Figure 1c). The enhanced radiative input is mainly stored in the ocean while some is fluxed away in the atmosphere (negative $F_{\text{WALL}}$) and damps the initial radiative input via the Planck feedback.

From an energy budget perspective, the driver of the sea ice loss can be identified from the process that adds energy to the atmosphere leading up to the sea ice loss. In this simplified framework, the driver of sea ice loss can be identified from the energy flux anomalies preceding, during, and following an ice loss event. What mechanism leads to Southern Ocean sea ice loss events in coupled climate model simulations? We demonstrate here that unforced Southern Ocean sea ice variability — defined as year-to-year changes associated with internal variability in the absence of external radiative forcing — simulated in long coupled climate model control runs (CMIP5; Taylor et al. 2012) exhibit all three theoretical mechanisms of sea ice loss introduced above.

The primary mechanism of sea ice variability differs between coupled climate models; ocean dynamics (Figure 2d), atmospheric energy transport (Figure 2e), and radiative processes (Figure 2f) can each serve as the dominant mode of sea ice variability in different climate models. In ocean-driven sea ice loss, oceanic processes impart energy to the surface that melts sea ice and provides upward surface energy fluxes to the atmosphere prior to the sea ice anomaly (note the positive values of the blue line in Figure 2d). In contrast, in sea ice loss driven by the atmospheric circulation, the atmosphere imports energy to the Southern Ocean prior to the sea ice anomaly (note the positive values of the purple line on the left of Figure 2e). In radiation driven ice loss, the net radiation delivers energy to the Southern Ocean prior to the sea ice loss (the positive values of the red line preceding the ice loss in Figure 2f).

Figure 2 demonstrates that the dominant mechanism of sea ice variability differs fundamentally between different coupled climate models. An analysis of CMIP5 models suggests that sea ice variability is dominated by ocean processes in about one third of the models, atmospheric circulation in one third, and radiation in one third. Given the diversity of mechanisms driving Southern Ocean sea ice loss in state-of-the-art coupled climate models, what is the dominant driver of Southern Ocean year-to-year sea ice loss in the observational record? Furthermore, can identifying the dominant mechanism of sea ice loss in nature help us to select the climate models that best represent the observed system to better inform our projections of future changes in Southern Ocean sea ice under global warming?

**Preliminary results: Observed relationship between Southern Ocean sea ice loss events and energy fluxes**

We present a preliminary analysis of the interannual variability of the Southern Ocean energy budget
associated with sea ice variability. Surface energy fluxes over the Southern Ocean are poorly constrained in observations. Thus, we focus on the relationship between the input of energy to the Southern Ocean via the atmospheric circulation ($F_{WALL}$) and the radiative anomalies associated with sea ice loss/gain events. The atmospheric energy flux across 55°S ($F_{WALL}$) is calculated from the vertical and zonal integral of the meridional moist static energy (MSE, the sum of latent, sensible, and potential energy) calculated from ERA interim six hourly data (Dee et al. 2016), using the methodology of Donohoe and Battisti (2013). We express $F_{WALL}$ variability as the resultant average heating (energy flux divergence) applied to the Southern Ocean by dividing $F_{WALL}$ by the surface area of the polar cap. The year-to-year variability (defined by low-pass filtering monthly anomalies with a six month cutoff period) of $F_{WALL}$ is on the order of 13 Wm$^{-2}$ (2σ). In the summer of 2001, $F_{WALL}$ was reduced by 30 Wm$^{-2}$ (half the climatology) in atmospheric energy input to the region peaking in the summer of 2001. Anomalies in the $F_{WALL}$ primarily result from changes in the strength of transient eddy sensible energy fluxes over the middle and lower troposphere, although stationary eddies in the stratosphere make a non-negligible contribution to the year-to-year variability (and likely do not have the same impact on surface climate as comparable magnitude fluxes in the lower troposphere).

Radiative anomalies — calculated from the CERES data from 2000 (Loeb et al. 2009) — make a substantially smaller (2σ = 3.3 Wm$^{-2}$) contribution to the interannual variability of the Southern Ocean energy budget as compared to $F_{WALL}$ (red and green lines in Figure 1). During sea ice loss events, there is a modest increase in absorbed solar radiation on the order of 1 Wm$^{-2}$, due to the reduced surface albedo (+1.9 Wm$^{-2}$) that is counteracted by enhanced cloudiness and cloud reflection (-0.9 Wm$^{-2}$). We emphasize that, given the small magnitude of radiative variability observed over the Southern Ocean compared to the impact of atmospheric processes, radiative processes cannot be the primary driver of Southern Ocean climate variability. The mechanism outlined in Figure 2c is not supported by our interpretation of the observational record.

The above results suggest that the dominant anomalous energy balance associated with observed Southern Ocean sea ice loss events is between the atmospheric energy input and surface energy fluxes. But are the atmospheric circulation anomalies driven by ice loss events associated with ocean and/or ice dynamics (as in Figures 2a,d) or are the ice loss events triggered by atmospheric energy input that is triggered remotely (as in Figures 2b,e)? There is no consistent lead/lag relationship between $F_{WALL}$ and Southern Ocean sea ice in the preliminary observational record shown here. This result suggests that there may be different flavors of Southern Ocean-wide ice loss events, some triggered by local dynamics and others responding to remote forcing. In future work, we hope to look more closely at the spatial structure (longitudinal and vertical) of the relationship between sea ice loss events, the resultant surface energy fluxes, and atmospheric circulation changes.

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**References**


Two Live Broadcasts at US CLIVAR SUMMIT

Polar Ocean & Sea Ice Interactions

Tuesday, August 8
5:45 - 7:45 PM ET

Featuring authors from this edition:
Ron Kwok (NASA JPL)
Julienne Stroeve (NSIDC and U. College London)
Alex Petty (U. Maryland and NASA GSFC)
Aaron Donohoe (U. Washington)

Understanding & Predicting Climate Teleconnections

Tuesday, August 8
10:00 AM - 12:00 PM ET

Featuring:
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Daniel Swain (U. California-Los Angeles)
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