Low-cloud, boundary layer, and sea ice interactions over the Southern Ocean during winter

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Abstract

During austral winter, a sharp contrast in low-cloud fraction and boundary layer structure across the Antarctic sea ice edge is seen in ship-based measurements and in active satellite retrievals from Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO), which provide an unprecedented view of polar clouds during winter. Sea ice inhibits heat and moisture transport from the ocean to the atmosphere, and, as a result, the boundary layer is cold, stable and clear over sea ice, and warm, moist, well-mixed and cloudy over open water. The mean low-cloud fraction observed by CALIPSO is roughly 0.7 over open water and 0.4-0.5 over sea ice, and the low-cloud layer is deeper over open water. Low-level winds in excess of 10 ms\(^{-1}\) are common over sea ice. Cold advection off of the sea ice pack causes enhanced low-cloud fraction over open water, and thus an enhanced longwave cloud radiative effect at the surface. Quantitative estimates of the surface longwave cloud radiative effect contributed by low-clouds are presented. Finally, ten state-of-the-art global climate models with satellite simulators are compared to observations. Near the sea ice edge, seven out of ten models simulate cloudier conditions over open water than over sea ice. Most models also underestimate low-cloud fraction both over sea ice and over open water.

1. Introduction

Sea ice, low-clouds, and the atmospheric boundary layer modulate the climate of the Southern Ocean by influencing surface heat fluxes. During winter, sea
ice insulates the ocean from the cold atmosphere above, reducing the rate of ocean heat loss at the surface by a factor 10-100 [Gordon, 1991]. Low-clouds and moisture emit longwave (LW) radiation downward and heat the surface, and low-level winds control the surface turbulent heat and moisture fluxes. When sea ice forms, brine is rejected, adding salt to the near-surface waters. These processes modify the buoyancy of surface waters and are responsible for deep and intermediate water formation. Roughly two-thirds of the deep water in the global ocean is formed in the Southern Ocean [Johnson, 2008], making it a region of critical importance for the global overturning circulation of the ocean [Marshall and Speer, 2012; Talley, 2013]. Surface fluxes of heat and moisture in the polar regions are intimately linked to the atmospheric boundary layer and to sea ice, are poorly observed, and are a topic of high priority for improving our understanding of polar climate and climate change [Bourassa et al., 2013].

Interactions between sea ice and boundary layer clouds have previously been studied, but focus on this topic has generally been on the Arctic. Across the Arctic basin during summer and early fall, low-clouds are more abundant and optically thicker over open water than over sea ice when viewed from active satellite remote sensing products [Kay and Gettelman, 2009; Palm et al., 2010] and from surface observers [Eastman and Warren, 2010]. On the other hand, Schweiger et al. [2008] used passive satellite retrievals and found that, during fall, regions of low sea ice concentration coincide with enhanced mid-level cloudiness and reduced low-cloud cover. Barton et al. [2012] found that the sensitivity of Arctic low-cloud fraction to variations in sea ice concentration depends on synoptic regime. For
stable regimes, which support low-clouds, a significant but weak covariance between sea ice concentration and cloud properties occurs during most seasons [Taylor et al., 2015]. Near the sea ice edge, cold, off-ice advection is known to cause enhanced low-cloud cover, but, due to a lack of observations, previous work has focused on case studies of extreme events [e.g. Walter, 1980; Renfrew and Moore, 1999; Petersen and Renfrew, 2007]. It has also been argued that boundary layer moisture, or lack thereof, triggers the onset of sea ice melt and freeze-up when advected over Arctic sea ice [Kapsch et al., 2013] and that a warmer, moister atmospheric boundary layer has amplified Arctic sea ice decline in recent years [Serreze et al., 2009; Screen and Simmonds, 2010; Boisvert and Stroeve, 2015].

Unlike the Arctic, interactions between Antarctic sea ice and boundary layer clouds have been given relatively little attention. Bromwich et al. [2012] first pointed out that the total cloud fraction observed from active satellite retrievals is about 0.1-0.2 lower in sea ice covered regions of the Southern Ocean than over open water to the north. Fitzpatrick and Warren [2007] used ship-based measurements of downwelling solar radiation over the Southern Ocean to show that, during austral spring and summer, clouds tend to be optically thicker over open water than over sea ice. It is not clear if the relationship between low-clouds and sea ice in the Arctic is similar to that in the Southern Ocean.

In this study we describe the boundary layer properties and low-cloud fraction across the Southern Ocean during winter, but with an emphasis on the marginal sea ice zone. We use satellite-based active retrievals of clouds, which provide an unprecedented view of polar clouds during winter, as well as ship-based
measurements. We also use a radiative transfer model to compute estimates of the downward flux of LW radiation near the sea ice edge and its sensitivity to low-level warm and cold advection. Finally, we evaluate ten state-of-the-art climate models. This paper is organized as follows: datasets and the methodology of the radiative transfer calculations are described in Section 2, results are given in Section 3, and conclusions are given in Section 4.

2. Data and Methods

a. Datasets

Cloud observations are taken from the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) instrument onboard the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite. CALIOP is a lidar that measures high vertical resolution profiles of backscatter from which estimates of cloud properties are derived [Winker et al., 2007]. Because CALIOP is an active instrument, retrievals are not affected by lack of sunlight or near-surface inversions – two conditions that are common at high latitudes during winter and are problematic for passive satellite retrievals of low-clouds. As an example of the challenge of cloud detection with passive instruments over the polar regions, Liu et al. [2004] found that, during polar night, about 40% of all clouds went undetected by the cloud mask algorithm of the moderate-resolution imaging spectroradiometer (MODIS) used at that time. The algorithm has since been improved [Baum et al., 2012], but detection of low-clouds over the polar regions remains a major challenge.
for passive instruments [Ackerman et al., 2008]. Unlike passive instruments, the
signal-to-noise ratio of CALIOP is maximized in the absence of sunlight, making it
well suited for studying clouds during polar night.

We use the General Circulation Model-Oriented CALIPSO Cloud Product
(CALIPSO-GOCCP) version 2.9 [Chepfer et al., 2010; CALIPSO, 2015]. CALIPSO-
GOCCP provides cloud fraction on a 2° longitude, 2° latitude and 480 m height grid.
It also provides low (below 3.2 km), middle (3.2-6.5 km) and high (above 6.5 km)
cloud fraction, and estimates for how they are partitioned between liquid and ice.
For a given grid box and time interval, cloud fraction is defined as the number of
scenes in which a cloud was positively identified divided by the number of scenes in
which the lidar was not fully attenuated in the grid box. The lidar beam becomes
fully attenuated at an optical depth of ~3 [Winker et al., 2007], so CALIOP often does
not measure the bottom portion of low-clouds [Cesana et al., 2016]. The vertical
resolution of CALIOP is 30 m below 8 km and 60 m above 8 km, with a total of 583
vertical levels. Using the relatively coarse GOCCP vertical grid, which has only 40
levels, significantly increases the signal-to-noise ratio and provides a grid that is
better for comparison with global climate models [Chepfer et al., 2010].

We also use satellite-based observations of sea ice concentration from the
National Oceanic and Atmospheric Administration/National Snow and Ice Data
Center Climate Data Record of Passive Microwave Sea Ice Concentration dataset
[Peng et al., 2013; NOAA/NSIDC, 2015], and cloud liquid water path from the Multi-
Sensor Advanced Climatology of Liquid Water Path dataset [Elsaesser et al., 2015].
Cloud liquid water path is defined as the total mass of cloud liquid water above a
unit area of the earth’s surface. The latitude of the sea ice edge, which we define as
the northernmost point at which the sea ice concentration is 0.35, is computed from
the sea ice data. Our main conclusions are not sensitive to choosing a threshold of
0.50 or 0.25 for this definition. We also use temperature, specific humidity, and
wind fields from the European Center for Medium-Range Weather Forecasts
Reanalysis (ERA Interim; [Dee et al., 2011; ECMWF, 2015]). Boundary layer fields in
reanalysis products are poorly constrained by observations over the Southern
Ocean, and should be interpreted with caution. However, ERA-Interim is
consistently ranked among the most reliable reanalysis products in the high
southern latitudes [e.g. Bromwich et al., 2011; Bracegirdle and Marshall, 2012; Jones
et al., 2016].

All satellite and reanalysis data are analyzed on either monthly- or daily-
mean timescales during the months of June, July and August (JJA) from 2006
through 2014. ERA-Interim reanalysis data are available on monthly-mean and
instantaneous 6-hourly time resolutions, and daily-means are computed from the
instantaneous 6-hourly data. Monthly-mean fields are analyzed unless stated
otherwise.

Additionally, we use ship-based observations of air temperature and wind
speed from soundings, and cloud-base height measured by a ceilometer.
Measurements were made on two cruises that traversed the Weddell Sea during
June-August 2013 and May-August 1992 [König-Langlo et al., 2006; König-Langlo,
2013; König-Langlo, 2005]. The cruise tracks are shown in Figure 1. On the 2013
cruise, soundings were launched once per day and have a vertical resolution of
about 30 m in the lower troposphere, while, on the 1992 cruise, soundings were launched four times per day and have a vertical resolution of about 60 m. Sounding data are linearly interpolated to a vertical grid with a spacing of 30 m and 60 m on the 2013 cruise and 1992 cruise, respectively. A total of 57 and 161 soundings were taken poleward of 55°S on the 2013 cruise and 1992 cruise, respectively. Wind speed measurements used global positioning system (GPS) technology on the 2013 cruise and the OMEGA radio navigation system on the 1992 cruise. As a result, higher quality wind speed measurements were made on the 2013 cruise [König-Langlo et al., 2006]. Measurements of cloud-base height were made on the 1992 cruise but not the 2013 cruise. These ship-based observations complement the satellite observations because the soundings can resolve the vertical structure of the boundary layer and the ceilometer measurements can reliably detect the cloud-base height. The main weakness of the ship-based observations is that measurements are sparse.

Finally, we use output from ten global climate models that participated in the Coupled Model Intercomparison Project Phase 5 (CMIP5) [Taylor et al., 2012], including output from the CALIPSO simulator [Chepfer et al., 2008]. Models in fully-coupled and atmosphere-only configurations are evaluated, and the first ensemble member for each model is used.

**b. Radiative transfer modeling**

One goal of this study is to quantify the downward flux of LW radiation at the ocean surface, which we call “surface $LW_{\downarrow}$”, and its dependence on low-cloud cover. Because direct observations of surface $LW_{\downarrow}$ are not available, we compute it using a
radiative transfer model. The advantage of using a radiative transfer model
compared to using reanalysis data is that we are able to vary low-clouds while
holding middle- and high-cloud cover fixed to zero. This method isolates the
contribution to surface $LW_i$ made by low-clouds.

We use the Rapid and Accurate Radiative Transfer Model for Global Climate
Models [Mlawer et al., 1997; Clough et al., 2005; Iacono et al., 2008]. This model is
one-dimensional and takes vertical profiles of temperature, humidity, cloud liquid
and ice water content as inputs and computes surface $LW_i$. The temperature and
humidity profiles are taken from the ERA-Interim reanalysis, and for each profile
the surface $LW_i$ is computed with a clear-sky and with low-cloud completely
obscurring the sky. We refer to these values as $LW_{i,\text{clear}}$ and $LW_{i,\text{overcast}}$ respectively. A best estimate for the true value of the flux of downward LW
radiation at the surface, which we call $LW_{i,\text{alt-sky}}$, is also computed:

$$LW_{i,\text{alt-sky}} = LW_{i,\text{clear}}(1 - LCF) + LW_{i,\text{overcast}} LCF$$

where $LCF$ is the low-cloud fraction observed by CALIPSO. Daily-mean data are used
because it is the shortest time resolution for which CALIPSO-GOCCP cloud
observations are available.

The radiative transfer calculations are done only for regions of open water
near the sea ice edge. We focus on scenes between 1°-3° equatorward of the ice
edge, where the sea ice concentration is approximately zero. This restriction is made
because microwave, satellite-based retrievals of cloud liquid water path are not
available over sea ice. Although the domain is restricted to open water scenes only,
the surface heat budget over open water is of interest because extreme air-sea heat
fluxes can occur there [e.g. Papritz et al., 2014].

In the radiative transfer calculations, several assumptions about low-clouds are made that are based on observations presented in Table 1. First, low-clouds are assumed to consist entirely of supercooled liquid. CALIPSO-GOCCP observations of cloud fraction partitioned by phase are presented in Table 1 and show that low-clouds over the Southern Ocean are frequently composed of liquid, consistent with previous studies [Hu et al., 2010; Morrison et al., 2011]. Cesana et al. [2016] suggest that low-clouds with unclassified phase in the CALIPSO-GOCCP dataset are mostly mixed-phase clouds. Even so, at least three-quarters of the low-clouds detected by CALIPSO are liquid (low-cloud fraction is 0.50 for liquid clouds and 0.69 for all phases). Second, all cloud liquid water is assumed to reside in the lowest 3.2 km of the atmosphere. Under this assumption the column-integrated cloud liquid water path is equal to the total liquid water contained in low-clouds. This assumption is justified by the CALIPSO phase observations (Table 1), which show that liquid clouds are usually found at low levels. Third, the liquid water path is assumed to be 60 gm$^{-2}$, which is near the median of satellite observations (Table 1). Fourth, the height of low-cloud base and top are set to 500 m and 1000 m, respectively. The cloud-base height value comes from measurements from the 1992 cruise taken when surface observers reported open water or open pack ice near the ship (Table 1). During this cruise, observed cloud-base height was distributed nearly uniformly between the surface and 1000 m. Fifth, low-clouds are assumed to have a droplet
Our results are insensitive to modest changes in these five assumptions about low-cloud properties. In order to test the sensitivity of the radiative transfer calculations to the assumptions about low-cloud properties, runs were performed with cloud-base height set to 0 m and 1000 m, with liquid water path halved to 30 \( \text{gm}^{-2} \) and with cloud effective radius doubled to 32 \( \mu \text{m} \) and halved to 8 \( \mu \text{m} \). The results are not sensitive to modest changes in liquid water path because liquid clouds are nearly opaque to LW radiation for liquid water path values greater than \( \sim 20 \text{ g/m}^2 \) [Hartmann, 2016], which is much lower than observed values (Table 1). In fact, each modification of the assumptions resulted in a change in \( LW_{i,\text{overcast}} \) of about 7 Wm\(^{-2}\) or less, which is small compared to the contribution to surface \( LW_{i} \) made by low-clouds during overcast conditions (\( \sim 80 \text{ Wm}^{-2} \) – discussed in the Results section). In other words, to leading order, low-cloud fraction controls the surface LW radiative effect from low-clouds. The model error is less than 1 Wm\(^{-2}\) [Mlawer et al., 1997], which is much smaller than the uncertainty due to the five assumptions made about cloud properties.

### 3. Results

**a. Satellite observations of sea ice and low-cloud over the Southern Ocean**

We start with a brief description of Antarctic sea ice during winter. Figure 1 shows July-average sea ice concentration over the Southern Ocean. Two contours of
Sea ice concentration are shown: 0.35 and 0.95. These contours can be thought of as marking the boundaries between open water, fragmented sea ice, and a sea ice pack that covers the surface nearly completely. Throughout most of the Eastern Hemisphere, sea ice concentration rarely exceeds 0.95. This could be a result of the coastline extending equatorward and forcing the sea ice closer to the Antarctic Circumpolar Current. In regions where the coastline cuts poleward, like the Weddell and Ross Seas, sea ice concentrations greater than 0.95 are much more common. Average sea ice concentrations in June and August are similar in this regard (not shown). Wadhams et al. [1987] describe the winter sea ice pack in the Weddell Sea as observed from a cruise. They found the marginal sea ice zone to be a ~250 km band of fragmented pancake ice with pockets of exposed seawater. Farther south, they found sea ice organized into vast floes that covered the ocean surface nearly completely. We recommend viewing photographs of these features in Wadhams et al. [1987, their Figure 12].

Now, consider low-cloud fraction over the Southern Ocean. The 2006-2014 winter climatology of low-cloud fraction and the latitude of the sea ice edge are shown in Figure 2. The interannual standard deviation of the latitude of the sea ice edge ranges between about 0.5° to 1.5° latitude. One standard deviation on either side of the mean position of the sea ice edge is shaded in Figure 2 to show that the effects of interannual variability of the location of the ice edge are likely small. Throughout the Southern Ocean, cloudier conditions are seen over open water than over sea ice. Near the sea ice edge, low-cloud fraction is about 0.7 over open water and 0.5 over sea ice. The gradient of low-cloud fraction across the sea ice edge is
weakest in the Southern Indian and Western Pacific Oceans (20°E – 160°E). This weak gradient is likely because the sea ice pack is more fragmented in this region than elsewhere in the Southern Ocean (Figure 1). In this region, the low-cloud fraction is more variable over sea ice than over open water because gaps in the sea ice pack are found throughout the ice pack, but little sea ice is found equatorward of the sea ice edge. In the Weddell and Ross Seas, where the sea ice pack covers the surface nearly completely, the low-cloud fraction is about 0.4 or less and the gradient in low-cloud fraction across the sea ice edge is sharp.

The relationship between sea ice concentration and low-cloud properties near the sea ice edge is made clearer by stratifying the observations based on distance from the sea ice edge. For each grid point and time (monthly-means from June, July and August between 2006-2014 are considered), the meridional distance between the grid point and the ice edge is computed. Data are then composited by meridional distance from the ice edge, using a bin width of 0.5° latitude, and averaged. We analyze data from the Weddell and Ross Seas (defined as 50°W-0°E and 130°W-170°E, respectively; Figure 1), two regions where the sea ice pack covers the surface nearly completely and where the sea ice edge is located far offshore (Figure 1). This procedure was also done on the JJA-mean of each year, and the main conclusions are the same using either monthly- or seasonal-averages.

Figure 3a shows the vertical profile of mean cloud fraction in the lower troposphere over the Weddell Sea as a function of meridional distance from the sea ice edge, and Figure 3d is similar but for the Ross Sea. On average, low-clouds extend deeper and are more prevalent equatorward of the sea ice edge. The mean
low-cloud fraction and sea ice concentration are shown in Figure 3b and 3e, and the method for deriving the confidence interval for the mean is described in the Appendix. The domain can be split into three regions based on sea ice concentration: an “ice” zone where sea ice concentration is ~1 that is located poleward of 2° south of the ice edge, an “open water” zone where sea ice concentration is ~0 that is located equatorward of 1° north of the ice edge, and a “transition” zone between. Within the ice zone the mean low-cloud fraction is nearly uniform at around 0.5, and within the open water zone the mean low-cloud fraction is nearly uniform at around 0.7. The mean low-cloud fraction is significantly larger in the open water zone than the ice zone. From south to north across the transition zone, the low-cloud fraction increases smoothly as sea ice concentration decreases.

Figures 3c and 3f show vertical profiles of mean potential temperature and specific humidity from reanalysis data as a function of meridional distance from the ice edge. In current reanalysis data, the surface heat budget and the atmospheric boundary layer over the Southern Ocean are poorly constrained by observations, and therefore these data should be interpreted with caution. Nevertheless, the data suggest several differences between the boundary layer over sea ice and over open water. The lower troposphere is more stable over sea ice than over open water, as can be seen by the vertical spacing in the potential temperature contours. Over open water, near-surface temperatures are close to the freezing temperature of seawater, and across the sea ice edge, near-surface temperatures drop rapidly. Boundary layer specific humidity values are also nearly a factor of two larger over open water than over sea ice.
b. Boundary layer structure from ship-based observations

Soundings resolve the vertical structure of the boundary layer and provide further insight into the physical processes at work. In this section, sounding data are represented by probability distributions. For each height measured by the soundings, the probability distributions are computed by binning the data, computing the number of observations in each bin, and normalizing by the total number of soundings. Data are composited into measurements made between 55°-65°S and poleward of 65°S. Because the sea ice edge is typically located between 60°-65°S in the Weddell Sea during winter, it is likely that most of the soundings poleward of 65°S were taken over consolidated pack ice. Meanwhile, soundings between 55°-65°S are likely a mixture of some taken over consolidated pack ice and some taken where open water was exposed to the atmosphere.

Figure 4 shows the probability distribution of temperature at each height between 10-1500 m. The 2013 and 1992 cruises are shown separately in Figure 4a-b and 4c-d, respectively, because the cruises used different sounding technologies [König-Langlo et al., 2006] and had different times between successive launches.

Two boundary layer regimes are seen: a warm and a cold mode. The warm mode is characterized by having near-surface temperatures close to the freezing temperature of seawater and by a moist adiabatic lapse rate above. In this regime the boundary layer is well-mixed and moist. The cold mode is characterized by typical near-surface temperatures of about -15°C to -25°C and by a low-level inversion. Poleward of 65°S, the cold mode dominates (Figure 4b and 4d). Between 55°-65°S, both the warm and the cold modes are seen, albeit with different
likelihoods between the two cruises (Figure 4a and 4c). Differences in the relative occurrence of the warm and cold mode in the 55°-65°S composite between the two cruises could be a result of different weather events. The latitudinal distribution of the warm and cold modes suggests that the cold mode forms over consolidated pack ice, and the warm mode forms over open ocean or gaps in the sea ice.

The soundings also measured wind speed, and this is shown in Figure 5. The probability distribution of wind speed as a function of height is shown for all soundings taken poleward of 55°S. Sounding data are not composited by latitude here, but doing so results in composites that resemble Figure 5 but are noisier (not shown). The soundings reveal that wind speeds of 10 ms\(^{-1}\) or more are common at heights of 200-600 m. For both cruises, the average wind speed between 200-600 m is 10 ms\(^{-1}\) or more for 60-70% of the soundings. For the 2013 cruise, the strong low-level winds are often associated with a low-level jet. On this cruise, the modal value of wind speed is \(~12-15\) ms\(^{-1}\) at heights of 200-400 m and decreases with height to \(~8\) ms\(^{-1}\) at heights of 800-1000 m (Figure 5a). Data from both cruises show that strong low-level winds are common during winter.

Low-level jets are of interest because they indicate the presence of a stable boundary layer. Low-level jets exist at the top of stable boundary layers and, at least in temperate latitudes, are initiated when the boundary layer transitions from convective to stable. During this transition, the sudden shoaling of the boundary layer causes a reduction in drag from turbulent momentum flux, and therefore a sudden increase in wind speed, at heights above the stable boundary layer but below the top of the former convective boundary layer. The stable boundary layer
limits drag on the winds above and allows the jet to persist and follow an inertial
oscillation [Blackadar 1957; Thorpe and Guymer, 1977]. The mechanisms that
initiate low-level jets over Antarctic sea ice during winter are not fully understood.
One possible mechanism is warm advection from open water to sea ice covered
regions, which temporarily deepens the boundary layer and then allows a new jet to
form when the boundary layer collapses to a stable profile [Andreas et al., 2000].
Another possible mechanism is motions arising from baroclinic instability
associated with the thermal contrast between sea ice and open ocean.

We emphasize that a weakness of this study is the short time span of
sounding data. Soundings were taken over a total of 50 days between 55°-65°S and
59 days poleward of 65°S. Despite this drawback, the main conclusions are robust:
in both cruises, a warm and a cold boundary layer regime are seen, and low-level
wind speeds in excess of 10 ms⁻¹ are common.

c. Advection across the sea ice edge

How do clouds respond when cold air is advected equatorward, across the
sea ice edge, and vice versa? We start with an investigation of low-cloud fraction and
its sensitivity to advection across the sea ice edge. Cold air outbreaks, in which air is
advected from a cold land or ice surface to a warmer ocean, are known to cause the
development of low-clouds [e.g. Walter, 1980; Liu et al., 2006]. When the cold air
mass is heated from below by the warm ocean surface, convection occurs and low-
clouds form. Once formed, low-clouds are maintained by radiative cooling at cloud-
top, radiative heating at cloud-base, and the moisture source of the ocean.
Figure 6 shows the mean low-cloud fraction as a function of meridional distance from the sea ice edge, stratified by low-level advection across the sea ice edge. As a metric for low-level advection across the sea ice edge, the meridional wind at 1000 hPa is linearly interpolated to the latitude of the sea ice edge. We refer to this value as $v_{ice\,edge}$. Data are composited into scenes in which $v_{ice\,edge}$ is less than $-0.5\sigma \approx -3 \text{ ms}^{-1}$ and greater than $0.5\sigma \approx 3 \text{ ms}^{-1}$, where $\sigma$ is the standard deviation of $v_{ice\,edge}$. These composites correspond to on-ice flow and off-ice flow, respectively. These composites are made using daily-mean data over the Weddell Sea. The mean low-cloud fraction equatorward of the sea ice edge is significantly larger during periods of off-ice flow than periods of on-ice flow. The peak in low-cloud fraction during periods of off-ice flow is located at about $2^\circ$ equatorward of the ice edge, suggesting that low-clouds formed by cold advection can persist well away from the sea ice edge. The fact that the peak in low-cloud fraction is about $2^\circ$ latitude equatorward of the sea ice edge may be a result of the predominant low-cloud type transitioning from roll clouds near the sea ice edge to cellular convection downstream [Walter, 1980]. This hypothesis is also consistent with the composites in Figure 3a and Figure 3d, which show that, near the sea ice edge, the low-cloud layer deepens toward the equator. Finally, for latitudes $2^\circ$ south of the ice edge and poleward, where sea ice covers the surface nearly completely (Figure 3b), there is either no significant difference, or a very small difference, in low-cloud fraction between the on-ice flow and off-ice flow composites. This result suggests that low-clouds over open water are coupled to the surface and require the moisture source of the open ocean to exist, and therefore dissipate when separated from open water.
d. Impact of low-level advection on the surface heat budget

We have seen evidence of a warm and a cold boundary layer regime, and that cold, low-level advection off of the sea ice pack causes low-clouds to form over open water. How do low-level advection and the resulting boundary layer and low-cloud changes impact the surface heat budget? To address this question we use a radiative transfer model to compute surface $LW_1$ over open water near the sea ice edge and to estimate the contribution made by low-clouds. The Weddell Sea is again used as the region of study. Recall that estimates of surface $LW_1$ are computed for a clear sky, with low-cloud completely covering the sky, and using low-cloud fraction observed by CALIPSO. These values will be called $LW_{1,\text{clear}}$, $LW_{1,\text{overcast}}$, and $LW_{1,\text{alt-sky}}$ respectively. The $LW_{1,\text{alt-sky}}$ values are the best estimate for the real world, while the $LW_{1,\text{clear}}$ and $LW_{1,\text{overcast}}$ values help with interpretation. Also, recall that there are no middle- or high-clouds in these calculations, so the radiative effects of low-clouds are isolated here.

First, consider the average values of surface $LW_1$. The average values of $LW_{1,\text{clear}}$ and $LW_{1,\text{overcast}}$ are about 210 Wm$^{-2}$ and 290 Wm$^{-2}$, respectively. In other words, if a point at the ocean surface was located under a clear sky, and a low-cloud passed overhead, then the downward flux of LW radiation would suddenly increase by about 80 Wm$^{-2}$, a 40% increase from the clear-sky value. The average value of $LW_{1,\text{alt-sky}}$ is around 270 Wm$^{-2}$. The surface LW cloud radiative effect, defined as $LW_{1,\text{alt-sky}} - LW_{1,\text{clear}}$, is about 50-60 Wm$^{-2}$. During winter, low-clouds warm the ocean surface by about 50-60 Wm$^{-2}$ on average.
Furthermore, surface $L W_1$ depends on the strength of warm or cold advection at low-levels. In the calculations of $L W_{1,\text{clear}}$ and $L W_{1,\text{overcast}}$, temperature and specific humidity are varied but low-cloud fraction is held fixed, and therefore the surface LW cloud radiative effect is nearly constant. In the calculation of $L W_{1,\text{all-sky}}$, temperature, humidity, and low-cloud fraction are all varied. Thus, by comparing data from the $L W_{1,\text{all-sky}}$, $L W_{1,\text{clear}}$ and $L W_{1,\text{overcast}}$ calculations, the sensitivity of surface $L W_1$ to low-cloud variations can be separated from the effects of temperature and humidity variations.

Figure 7 shows surface $L W_1$ plotted as a function of $v_{\text{ice edge}}$. In the $L W_{1,\text{clear}}$ and $L W_{1,\text{overcast}}$ calculations, where low-cloud fraction is held fixed, the data are anticorrelated with $v_{\text{ice edge}}$ ($r = -0.61$ and $r = -0.64$, respectively). This happens because air masses that form over the sea ice pack are cold and have low specific humidity, and these features of the air cause it to radiate relatively weakly to the surface when advected over open water (when low-cloud fraction is held fixed).

However, in the $L W_{1,\text{all-sky}}$ calculation, where low-cloud fraction is varied according to CALIPSO observations, data are weakly anticorrelated with $v_{\text{ice edge}}$ (scatterplot in Figure 7). When the $L W_{1,\text{all-sky}}$ data are binned by $v_{\text{ice edge}}$ and averaged, the result agrees well with a linear regression (compare the blue and red lines in Figure 7). Because cold advection causes cloudy conditions, $L W_{1,\text{all-sky}}$ data approach the $L W_{1,\text{overcast}}$ regression line for large positive values of $v_{\text{ice edge}}$. The surface LW cloud radiative effect, seen in Figure 7 by the difference between $L W_{1,\text{all-sky}}$ and $L W_{1,\text{clear}}$, increases by $1.1 \pm 0.1$ Wm$^{-2}$ per 1 ms$^{-1}$ increase in $v_{\text{ice edge}}$. Put another
way, typical values of the average surface LW cloud radiative effect, estimated by the regression, range from 43 W m\(^{-2}\) to 65 W m\(^{-2}\) for \(v_{\text{ice edge}} = -10\) m s\(^{-1}\). As a result, the regression coefficient of \(LW_{\text{alt-sky}}\) on \(v_{\text{ice edge}}\) \((-0.7 \pm 0.1\) W m\(^{-2}\) per 1 m s\(^{-1}\) increase in \(v_{\text{ice edge}}\)) is significantly smaller in magnitude than the regression coefficients of \(LW_{\text{clear}}\) and \(LW_{\text{overcast}}\) (compare slopes in Figure 7). Therefore, when low-cloud fraction, temperature, and humidity are all allowed to vary, as they are in the real world, then surface \(LW_i\) is much less sensitive to warm or cold advection than when low-cloud fraction is held fixed. Low-clouds warm the surface most strongly during cold advection events, and therefore act to reduce the sensitivity of surface \(LW_i\) to cold advection.

We emphasize that these calculations are only able to capture one term of the surface heat budget: surface \(LW_i\). Surface turbulent heat fluxes are likely very important as well. Over open water in the Southern Ocean during winter, average values of surface turbulent fluxes of sensible and latent heat are around 30 W m\(^{-2}\) and 50 W m\(^{-2}\) respectively – on the order of the average surface LW cloud radiative effect – but surface turbulent heat fluxes can be several hundred watts per square meter during extreme cold air outbreaks [Papritz et al., 2014]. These turbulent heat flux values come from reanalysis data, and should therefore be interpreted with caution. Our work is progress towards constraining the Southern Ocean surface heat budget, but a complete understanding also requires knowledge of the surface turbulent heat flux.
e. Evaluation of Global Climate Models

We have seen that the surface LW cloud radiative effect from low-clouds is about 50-60 Wm\(^{-2}\) on average. Therefore, low-cloud biases in global climate models (GCMs) have the potential to significantly bias the modeled surface heat budget. Accurate representation of marine boundary layer clouds and their radiative effects are a major challenge for GCMs [Dufresne and Bony, 2008; Trenberth and Fasullo, 2010], and low-clouds are especially poorly represented in GCMs during polar night [Karlsson and Svensson, 2011]. Here we evaluate the climatology of low-cloud fraction near the sea ice edge in ten GCMs. Low-cloud fraction in the models is computed by a CALIPSO simulator that estimates what CALIPSO would observe if it were flying above the atmosphere in the model. Both fully-coupled and atmosphere-only (AMIP) configurations are analyzed between 1990-2004 [Taylor et al., 2012; Gates, 1992]. Fully-coupled models have prescribed atmospheric greenhouse gas concentrations from observations, while atmosphere-only models have prescribed sea surface temperature and sea ice concentration from observations. Most importantly, interactions between the ocean, sea ice, low-clouds and the atmospheric boundary layer are active in fully-coupled configuration and are suppressed in atmosphere-only configuration. Model climatologies are compared to the observed climatology between 2006-2014.

Figure 8 shows the July-mean low-cloud fraction observed by CALIPSO and simulated by six GCMs that have output from both atmosphere-only and fully-coupled configuration. Model behavior is quite diverse. Starting with the atmosphere-only runs, HadGEM2, MIROC5 and MRI-CGCM3 have cloudier
conditions over open water than over sea ice. However, the gradient in low-cloud fraction across the sea ice edge is weak in MIROC5. The IPSL models and MRI-CGCM3 significantly underestimate low-cloud cover over open water. Compared to the atmosphere-only runs, models in fully-coupled configuration generally have similar low-cloud fractions, and the MIROC5 and MPI-ESM-LR models have a sea ice edge that is located too close to the coastline. The models do not capture the weaker low-cloud fraction gradient across the sea ice edge in the Eastern Hemisphere that is seen in observations (Figure 2). Other than that, the modeled low-cloud fraction does not have any systematic geographic bias.

Model bias in low-cloud fraction is quantified and shown in Figure 9. As a metric for the low-cloud fraction near the sea ice edge, the mean low-cloud fraction is computed over open water and sea ice, again defined as 1°-3° equatorward and 2°-4° poleward of the ice edge, respectively. These will henceforth be referred to as LCF_{open water} and LCF_{sea ice}. One common feature between models is a bias of too little low-cloud fraction over open water, consistent with previous work [e.g. Zhang et al., 2005]. In fact, over open water, nine out of ten models underestimate low-cloud fraction, one model (GFDL-CM3) agrees with observations at 95% confidence, and no models overestimate low-cloud fraction. This can be seen by noting that LCF_{open water} for all of the models appears to the left of the observations in Figure 9a. The magnitude of the largest model bias is about 0.35 (IPSL models), meaning that low-cloud fraction in these models is roughly half of the observed value. Model bias in LCF_{sea ice} is more diverse and ranges from +.15 to -.26. However, out of the eight fully-coupled models, all but one model underestimate LCF_{sea ice}, one model agrees
with observations (MPI-ESM-LR), and no models overestimate LCF_{sea ice}. This can be seen by noting that all but one of the fully-coupled models are below the observations in Figure 9a. Again, models that underestimate low-cloud fraction the most have low-cloud fractions that are roughly half of the observed value. Most models underestimate low-cloud fraction over open water and over sea ice.

Having established the mean low-cloud fraction bias in the models, we now examine the difference of low-cloud fraction between the open water and sea ice regions, which is shown in Figure 9b. Seven out of ten models have significantly cloudier conditions over open water than over sea ice. Although most models capture the correct sign of LCF_{open water} − LCF_{sea ice}, the magnitude varies between 0.38 in the MIROC-ESM and MIROC-ESM-CHEM models and -0.11 in the IPSL models, while the observed value is LCF_{open water} − LCF_{sea ice} = 0.17 ± 0.01.

Finally, it is interesting to consider how low-clouds change with model resolution and configuration. Models that have both atmosphere-only and fully-coupled output provide an opportunity to compare low-clouds when the ocean and sea ice are prescribed and when they are interactive. Of these six models that have both atmosphere-only and fully-coupled output, three have little or no difference between low-cloud fraction in fully-coupled and atmosphere-only configuration (IPSL-CM5A-LR, IPSL-CM5A-MR and MRI-CGCM3). However, these models also underestimate low-cloud fraction the most (Figure 8 and Figure 9a). The MIROC5 and MPI-ESM-LR models have more realistic values of LCF_{open water} − LCF_{sea ice} in the atmosphere-only configuration than in the fully-coupled configuration (Figure 9b), but this may be due to the fact that the sea ice edge in these models is much
closer to shore, and therefore more exposed to cold continental air, in fully-coupled configuration (Figure 8). Finally, the IPSL models make for an interesting comparison because they differ only in resolution: IPSL-CM5A-LR and IPSL-CM5A-MR have horizontal resolutions of $1.9^\circ \times 3.75^\circ$ and $1.25^\circ \times 2.5^\circ$, respectively. In this model, finer resolution does not improve the low-cloud bias.

4. Summary and conclusions

During austral winter, active satellite retrievals from CALIPSO and ship-based measurements show a strong contrast in low-cloud fraction and boundary layer structure over Antarctic sea ice and the adjacent open ocean. Low-cloud fraction is roughly 0.7 over open water and 0.4-0.5 over sea ice, and the low-cloud layer is much deeper over open water. The boundary layer is cold, stable, dry and clear over consolidated sea ice and warm, moist, cloudy and well-mixed over open water. At heights of 200-600 m, wind speeds in excess of 10 m s$^{-1}$ are common over sea ice, and are often associated with a low-level jet. During periods of cold, off-ice advection, low-cloud fraction and the surface LW cloud radiative effect are enhanced over open water. This enhanced cloud radiative effect acts to substantially slow the rate of LW cooling of the ocean mixed layer compared to what would happen if low-cloud fraction were uncorrelated with warm or cold advection. Low-cloud fraction over sea ice is similar for on-ice and off-ice advection conditions, indicating that low-clouds that form over the open ocean are coupled to the surface and do not survive when separated from the moisture source provided by open water.
These results support the hypothesis of two-way interactions between the ocean surface and the atmospheric boundary layer during winter over polar oceans. Regions of open water have relatively warm surface temperatures and large surface fluxes of heat and moisture to the atmosphere. Moist and warm boundary layers with a strong greenhouse effect form over open water and thus reinforce the warm surface temperatures there. If such a region were to become covered by sea ice then the surface heat and moisture fluxes would reduce and the boundary layer would cool, dry, and become less cloudy, causing the greenhouse effect of the boundary layer to weaken. This would reinforce the cool surface temperatures and help the sea ice persist.

Additionally, ten state-of-the-art climate models with CALIPSO simulator output were examined. Seven out of ten models simulate a larger low-cloud fraction over open water than over sea ice. Nine out of ten models underestimate low-cloud fraction over open water, and seven out of eight fully-coupled models underestimate low-cloud fraction over sea ice. The observed low-cloud and boundary layer properties shown in this work can be used as a test in future model intercomparison projects when CALIPSO simulator output for more models is available.
Acknowledgements

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Appendix

Derivation of 95% confidence interval for the mean low-cloud fraction

Figure 3b and Figure 3e show the mean low-cloud fraction as a function of meridional distance from the sea ice edge in the Weddell Sea and Ross Sea, respectively. The 95% confidence interval for mean low-cloud fraction is computed assuming that low-cloud fraction measurements at each grid-cell and time are independent. The low-cloud fraction data is available as monthly averages on a 2° longitude by 2° latitude grid. To justify the assumption that measurements of low-cloud fraction at each grid cell and time are independent we must assess spatial and temporal autocorrelation.

Serial correlation in the meridional dimension

Our goal is to bin the data by meridional distance from the ice edge and compute the mean and 95% confidence interval for the mean of each bin. The CALIPSO grid is resolved in 2° latitude grid cells, and when compositing by meridional distance from the ice edge we use bins of width 0.5° latitude. Therefore, for any given time, no two grid cells of the same longitude and neighboring latitudes can be assigned to the same bin. We therefore do not need to consider serial correlation in the meridional dimension when estimating the effective degrees of freedom of each bin.

Serial correlation in the zonal dimension
We estimate a lower bound for the number of degrees of freedom in the zonal dimension by computing the correlation length scale $L$ for each latitude and time, and comparing it to the resolution of the CALIPSO grid. Following [Taylor, 1921; Keller, 1935], we define the correlation length scale as

$$ L = \int_0^\infty r(\tau) d\tau $$

where $r(\tau)$ is the spatial autocorrelation function of low-cloud fraction in the zonal dimension and $\tau$ is the separation distance. The distance between independent points in the zonal dimension can be thought of as $2L$. We computed $2L$ for each time and latitude within 10 degrees of the sea ice edge, and the maximum value was an order of magnitude smaller than the $2^\circ$ longitude resolution of the CALIPSO grid. We therefore treat each grid cell as an independent measurement of low-cloud fraction in the zonal dimension.

Serial correlation in the time dimension

Each longitude and latitude grid-cell contains a timeseries of low-cloud fraction observations. The number of effective degrees of freedom of the low-cloud fraction timeseries ($N_{eff}$) is related to the number observations of low-cloud fraction ($N$) and the lag-1 autocorrelation of low-cloud fraction ($r_1$) by the following equation [Bretherton et al., 1999]:
Our goal is to bin the observations of low-cloud fraction by their meridional distance from the ice edge and then to use this equation to estimate the effective degrees of freedom for each bin. For each longitude and latitude grid cell we compute $r_1$ over the entire timeseries. Then, for a given bin, say “bin A,” and a given grid cell, say “grid cell B,” we keep track of the number of times that grid cell B is assigned to bin A and then compute an estimate for the number of effective degrees of freedom contributed to bin A by grid cell B by scaling by the right hand side of the above equation. This procedure is done for every grid cell and every bin.

The total number of effective degrees of freedom for each bin estimated by this procedure is slightly greater than if one were to assume each observation of low-cloud fraction is independent in the time dimension. This happens because the lag-1 autocorrelation of low-cloud fraction over the domain tends to be slightly negative, probably because of random sampling variability. We therefore assume that each estimate of low-cloud fraction is independent in the time dimension.
References


doi:10.1038/nclimate1884


NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, Version 2, 2015.


Table Captions

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Figure Captions

**Figure 1.** Map of Antarctica and the Southern Ocean showing average sea ice concentration during July from 2006-2014 from satellite observations. Two contours of sea ice concentration are shown: 0.35, which we use as a metric for the sea ice edge, and 0.95. Boundaries for the Weddell (50°W-0°) and Ross (130°W-170°E) Seas are shown in the black dashed lines, and tracks for the cruises to the Weddell Sea are shown as colored lines. The cruises started near 0° longitude and finished near the Antarctic Peninsula.
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**Figure 5.** Vertical profile of wind speed from soundings in the Weddell Sea region poleward of 55°S. For each height, color shows the probability density function of wind speed. Bins of width 3 ms⁻¹ are used in the calculation. Data from the 2013 and 1992 cruises are shown in the top and bottom panels, respectively. Data were collected over 53 days on both cruises. Note that wind speeds of 10 ms⁻¹ or more are common at heights between 200-600 m, and that the signature of a low-level jet can be seen in the measurements from the 2013 cruise.

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