Arctic Cloud Properties and Radiative Forcing From Observations and Their Role in Sea Ice Decline Predicted by the NCAR CCSM3 Model During the 21st Century

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Arctic sea ice is sensitive to changes in surface radiative fluxes. Clouds influence shortwave radiation primarily through their high albedo and longwave radiation by changing atmospheric emissivity and determining the height (temperature) of the layer of the highest emission. We review Arctic cloud properties affecting radiative fluxes, estimate sea ice effect on the top-of-atmosphere albedo, and discuss cloud response and contribution to the Arctic sea ice decline during the 21st century predicted by the National Center for Atmospheric Research Community Climate System Model, version 3 (CCSM3). Over perennial sea ice, clouds decrease incoming shortwave flux at the surface compared to clear skies from zero in winter to ~100 W m$^{-2}$ during the summer. On average over the Arctic Ocean, sea ice retreat decreases the shortwave radiation reflected at the top of the atmosphere within the same range for all-sky conditions. In addition, Arctic clouds warm the surface increasing the annual mean downwelling longwave flux by ~40 W m$^{-2}$. During the 21st century, CCSM3 predicts a drastic sea ice decline accompanied by larger cloud cover and liquid water content, which increase both cloud cooling and warming effects at the surface. The surface albedo decrease caused by sea ice retreat is partly compensated but not canceled by stronger shortwave cloud cooling. Warming of the near-surface atmosphere is an additional factor increasing the downwelling longwave flux at the surface. The ultimate effect of cloud changes in this model is facilitating the sea ice decline.

1. INTRODUCTION

During the last 30 years, satellite observations have shown a large reduction in the Arctic sea ice cover [Overpeck et al., 2005; Stroeve et al., 2005; Serreze et al., 2007]. Significant future reduction in the extent and thickness of the Arctic sea ice is predicted by coupled models participating in the Fourth Assessment Report of Intergovernmental Panel
on Climate Change (IPCC AR4) [Zhang and Walsh, 2006]. When forced with Special Report on Emission Scenarios (SRES) A1B scenario, in which atmospheric CO₂ concentration doubles by 2100, half of the model population simulates an ice-free Arctic Ocean in late summer by the end of the 21st century [Arzel et al., 2006]. In reality, the sea ice decrease has accelerated during the past several years, and seasonally ice-free Arctic can happen even faster than the most pessimistic forecast [Serrze et al., 2007; Stroeve et al., 2005]. According to satellite data, September 2007 average Arctic sea ice extent was already as low as $4.28 \times 10^{6}$ km$^2$, which is 23% lower than the previous record minimum for this month in 2005 and 39% below the long-term average from 1979 to 2000 (National Snow and Ice Data Center press release of 1 October 2007, available at http://nsidc.org/news/index.html).

Studies reviewed by Serrze et al. [2007] showed that a combination of the ocean, ice, and atmosphere changes have produced feedbacks, which worked together toward the rapid sea ice reduction. The ice-albedo feedback is considered to be a major process accelerating the sea ice reduction. Analyzing the 21st century SRES A1B simulations of the National Center for Atmospheric Research Community Climate System Model, version 3 (NCAR CCSM3), Holland et al. [2006] found that the Arctic sea ice retreat accelerates because of the larger open water area and increased absorption of solar radiation, triggered by increased oceanic heat transport into the Arctic. Similarly, Winton [2006] named both the ice albedo feedback and ocean heat flux as two primary drivers for sea ice removal in one of IPCC standard experiments, where CO₂ increases by 1% year to quadrupling. The increase in the oceanic heat flux has been supported by observations, which showed warming of the Atlantic layer residing at intermediate depths within the Arctic Ocean [Polyakov et al., 2004] and the increase in both transport and temperature of the waters entering the Arctic from the Atlantic [Schauer et al., 2004; Swift et al., 1997]. The exchange between relatively warm and saline water entering from the Atlantic and the sea ice ultimately depends on the strength of the Arctic halocline, which has significant spatial, seasonal, and interannual variability [Björk et al., 2002; Martinson and Steele, 2001; Rudels et al., 2004]. In the Bering Strait, observations (from 1990 to 2004) also suggest an increase in ocean heat flux in the Chukchi Sea due to both increased volume flux and temperature [Woodgate et al., 2006]. While the time series is relatively short, the 2004 ocean heat flux is the highest on record, and the 2001–2004 heat input could melt an equivalent of 640 km$^3$ of sea ice.

Another possible contribution to the Arctic surface warming is the increase in the downwelling longwave (LW) radiation associated with increased cloud cover and water vapor. Satellite data analysis by Francis et al. [2005] indicates that the sea ice edge annual anomalies are most highly correlated with the fluctuations in the downwelling LW flux, more than with the changes in wind advection, sensible heat transport, or shortwave (SW) radiation. Negative correlation was found between the sea ice edge anomalies and downwelling SW flux, indicating that stronger ice retreat occurred when the incoming SW flux at the surface was smaller. Observations in the Canadian Arctic showed that the earlier snowmelt onset and decrease in the annual mean albedo have been possibly driven by the increase in winter incoming LW radiation [Weston et al., 2007]. According to the modeling study by Zhang et al. [1996], it is the downwelling LW flux that controls interannual variability in the timing of melt onset. Sea ice thinning in response to increased cloudiness was demonstrated by several studies using thermodynamic sea ice models of different complexity [Ebert and Curry, 1993; Shine and Craine, 1984]. Satellite observation have shown a strong positive trend in spring cloudiness over the Arctic Ocean since 1980 [Schweiger, 2004; Wang and Key, 2003]. Cloudier atmosphere during spring could also contribute to the recent sea ice retreat in addition to the oceanic heat increase discussed above.

Surface cloud radiative forcing is defined as the difference in the downwelling and/or net radiative fluxes at the surface during cloudy conditions compared to clear skies. Clouds have competing effects on the surface radiative budget: they cool the surface by reflecting the SW radiative flux and warm the surface by increasing the downwelling LW radiation. The uniqueness of the Arctic cloud effects on radiation is in the strong seasonality of their SW and LW forcing. Cloud ability to cool the surface in the polar regions strongly depends on the surface albedo. Clouds reflecting solar radiation over the sea ice covered with fresh snow have zero or very little effect on the surface net SW flux because of the high surface albedo. When sea ice is covered with melt ponds during the summer or over the ice-free ocean areas, clouds greatly reduce the amount of the SW flux absorbed at the surface. Cloud ability to warm the surface depends only on the marcoophysical and microphysical cloud properties, which influence the downwelling LW flux. The upwelling LW flux, in turn, responds to changes in surface temperature. During the winter, cloud presence is most of the times associated with warmer surface temperatures [Walsh and Chapman, 1998]. During the summer melt period, when the surface temperature is constrained by 0°C, increased downwelling LW flux is used to melt the snow and ice.

There is a high uncertainty in the cloud response to already occurring and future sea ice changes. Cloud response to the reduction in sea ice thickness and area can have two opposite effects on the surface radiative budget. First is the
“umbrella” effect: melting of ice promotes greater cloud formation, which reduces the amount of SW radiation absorbed by the ocean. This diminishes the ice-albedo feedback and mitigates the increase in surface temperature. Second is the “greenhouse” effect: increased cloudiness and atmospheric temperatures accompanying the sea ice reduction in the Arctic enhance LW cloud warming of the surface. This extra energy causes earlier and/or greater snowmelt, which, in turn, triggers the ice-albedo feedback by reducing the surface albedo.

In the present chapter we use satellite and ground-based observations to estimate and compare the effects of clouds on the shortwave and longwave radiation. Further, we use the NCAR CCSM3 coupled global climate model to examine the change of sea ice and clouds during the 20th and 21st centuries. The chapter is structured as follows: Section 2 describes observations and the model. Section 3 reviews the studies based on observations subdivided into two parts: section 3.1 discusses the effects of sea ice and clouds on the SW fluxes at the top of atmosphere using satellite data, and section 3.2 reviews Arctic cloud radiative properties using ground-based observations. Section 4 presents the CCSM3 model simulations of sea ice, clouds, and radiation during the 20th and 21st centuries. The summary and the discussion of the study are given in section 5.

2. DATA AND METHODOLOGY

2.1. Observational Data

A suite of accurate up-to-date data of Arctic cloud properties, radiative fluxes, and sea ice are used in this study. Shortwave radiative fluxes and albedo spectrally integrated over the 0.2–5.0 μm band at the top of the atmosphere (TOA) derived from the Earth Radiation Budget Experiment (ERBE) data are used to analyze sea ice effects on the TOA shortwave radiation for all-sky conditions. The program combines the ERBS, NOAA 9, and NOAA 10 satellite measurements for the period from November 1984 to February 1990 [Barkstrom et al., 1989; Barkstrom and Smith, 1986]. We use the narrow field of view product with a spatial resolution of 2.5° × 2.5°. Global error for ERBE monthly SW fluxes is 5.5 W m⁻² [Wielicki et al., 1995]. Larger errors in the polar regions (up to 20 W m⁻² for all-sky fluxes and 50 W m⁻² for clear-sky fluxes) are caused by inaccuracies in surface scene and cloud identification over sea ice (a review of the ERBE data errors is given by Gorodetskaya et al. [2006]). Large uncertainties in clear-sky identification over ice surfaces make separate analyses of the data for clear-sky conditions and cloudy-sky conditions unreliable [Li and Leighton, 1991]. For all-sky data, the errors contribute to the scatter, but they are substantially smaller than the changes in the SW fluxes associated with seasonal variations in sea ice concentrations. Significant improvements in the radiative flux retrievals in the polar regions are achieved in the new product from the Clouds and the Earth’s Radiant Energy System (CERES) program, which is currently becoming available after a series of validations [Kato et al., 2006]. However, the TOA albedo from the CERES data differs significantly from the ERBE [Bender et al., 2006] and requires more comparison with observations prior to be used in sensitivity studies. The choice in the present study is given to the widely used ERBE data set, taking into account its shortcomings.

Sea ice concentration (SIC) data used for the analysis of sea ice effects on the TOA shortwave fluxes are from the UK Met Office Hadley Centre’s sea ice and sea surface temperature data set (HadISST1) available from 1870 to the present on a 1° latitude-longitude grid [Rayner et al., 2003]. Beginning in 1978, the data are derived from Special Sensor Microwave Imager and the Scanning Multichannel Microwave Radiometer [Gloersen et al., 1992]. The microwave radiance data have a monthly averaged SIC error of about 7%, increasing up to 11% during the melt season [Gloersen et al., 1992]. The biases are greatly reduced in the HadISST1 homogenization process using other satellite and in situ sea ice concentration and sea ice extent data [Rayner et al., 2003].

Cloud and surface radiative flux ground-based data are obtained during the Surface Heat Budget of the Arctic Ocean (SHEBA) program conducted on an ice floe that drifted more than 1400 km in the Beaufort and Chukchi seas between 74°N and 81°N latitudes and between 165°W and 140°W longitudes from late October 1997 to mid-October 1998 [Uttal et al., 2002]. The major meteorological and surface energy budget measurements were conducted on multiyear pack ice with summertime melt ponds and occasional nearby leads [Perovich et al., 2002; Tschudi et al., 2001]. Upwelling and downwelling fluxes were measured using broadband radiometers [Persson et al., 2002]. The uncertainty in the downwelling and upwelling LW flux was estimated at ±2.5 W m⁻² and at ±4 W m⁻² for the net LW radiation. The estimated uncertainty in the downward and upward SW flux is ±3% with a bias from −5 to +1 W m⁻² for the downward and 0 to −6 W m⁻² for the upward SW flux [Persson et al., 2002]. Cloud presence and base height were derived from a combination of lidar and radar measurements [Intrieri et al., 2002]. Column-integrated liquid water path (LWP) was derived from brightness temperatures measured by microwave radiometer with 25 g m⁻² accuracy [Intrieri et al., 2002; Westwater et al., 2001].

2.2. NCAR CCSM3 Model

The output from the 20th and 21st century simulations of the NCAR CCSM3 global coupled model is examined to
improve our understanding of the role of clouds in the Arctic sea ice decline. The model consists of components simulating the Earth’s atmosphere, ocean, land surface, and sea ice connected by a flux coupler [Collins et al., 2006]. This paper shows results from the configuration used for climate change simulations with a T85 grid for the atmosphere and land and a 1° grid for the ocean and sea ice. Cloud amount is diagnosed by the relative humidity, atmospheric stability, and convective mass fluxes [Boville et al., 2006]. Cloud ice and liquid phase condensates are predicted separately [Rasch and Kristjánsson, 1998; Zhang et al., 2003]. This links the radiative properties of the clouds with their formation and dissipation. However, after advection, convective detrainment, and sedimentation, the cloud phase is recalculated as a function of cloud temperature. Cloud liquid and ice are assumed to coexist within a temperature range of −10°C and −40°C with the ice fraction linearly increasing with decreasing temperature [Boville et al., 2006]. Clouds are all liquid above −10°C, and all ice below −40°C. Compared with observations, the model produces too much atmospheric moisture in the polar regions and too little in the tropics and subtropics, suggesting that the poleward moisture flux is excessive [Collins et al., 2006]. As a consequence of the excessive moisture advection and allowing cloud liquid water to exist at low temperatures, the model Arctic clouds contain large amounts of cloud liquid water [Gorodetskaya et al., 2008].

The sea ice in the CCSM3 is represented by a dynamic-thermodynamic model that includes a subgrid-scale ice thickness distribution, energy conserving thermodynamics, and elastic-viscous-plastic dynamics [Briegleb et al., 2004]. The surface albedo for the visible and near infrared bands is a function of ice and snow thickness and surface temperature.

In this paper, the model output during 1950–1999 is obtained from the “Climate of the 20th century experiment” 20C3M scenario simulations conducted from 1870 to present. The model output for 2000–2100 time period is from the SRES A1B experiment, where simulations are initialized with conditions from the end of the 20C3M simulations and run to 2100 under imposed SRES A1B conditions. The SRES A1B scenario assumes moderate population growth and rapid economic growth according to the estimates in the end of 1990s with atmospheric CO₂ concentration increase to 720 ppm by 2100 (doubling the 1990 amount) [Houghton et al., 2001].

2.3 Methods

Satellite data analysis of sea ice effects on the TOA shortwave fluxes is performed during the ERBE time period from November 1984 to February 1990 on 2.5° grid. Cloud properties influencing radiative forcing are obtained from point ground-based observations during 1 year from 1997 to 1998. The CCSM3 model output is analyzed on monthly mean and annual mean timescales on the 1° grid for the time period from 1950 to 2100. Although the data are available on different time periods and spatial resolutions, they are used for sensitivity estimates rather than for providing the absolute magnitudes. Observations give a picture of the sea ice/cloud/radiation relationships for modern climate. These relationships are then applied to understand the role of cloud and radiation changes on sea ice cover during the 21st century as predicted by the CCSM3 model.

3. SEA ICE AND CLOUD EFFECTS ON RADIATION: OBSERVATIONS

3.1. Effect of Sea Ice on the Top-of-Atmosphere Albedo

Top-of-atmosphere albedo represents a fraction of the incoming solar energy reflected from the surface and atmosphere system. Clouds shield the surface from solar radiation mitigating the effect of melting sea ice on the TOA albedo. While the surface albedo decreases drastically when sea ice cover declines (exposing dark ocean surface) the TOA albedo changes are much smaller if the skies are cloudy. The magnitude of the ice-albedo feedback is defined by the changes in the TOA albedo, rather than in the surface albedo, in response to the sea ice changes. Thus, cloud changes can modify the magnitude of the ice-albedo feedback. To show the role of clouds in mitigating the ice-albedo feedback, we estimate sea ice effect on the SW radiation reflected at the TOA for average cloud conditions.

In the Arctic, the ice-free ocean areas are almost always associated with overcast skies, which reduces the difference between the TOA albedo over the open ocean and sea ice. Figure 1 shows spatial distribution of the all-sky TOA albedo over the Arctic and adjacent ocean area during March and July, together with the climatological sea ice extent for these months. During July the sea ice surface has the lowest albedo because of ubiquitous melt ponds, while the sea ice extent reaches the minimum area only in September. In March, Arctic sea ice extent is at maximum, and insolation starts to increase after the polar night (incoming SW flux at the TOA averaged over the ocean north of 70°N reaches 100 W m⁻² in March). The monthly mean TOA albedo over sea ice in March lies between 60 and 70% (Figure 1a), reducing to 50–60% during the summer (Figure 1b). During the summer, the ice-free ocean, including the area of the ocean occupied by ice during the winter, has TOA albedo ranging between 40 and 50% (Figure 1b). Thus, while the difference between the TOA albedo over the ice-free ocean and over the sea ice is reduced by high cloud amount, seasonal
changes in sea ice cause about 10–20% change in the all-sky TOA albedo.

Sea ice effect on the TOA albedo is summarized in Figure 2, which shows the TOA albedo and the surface albedo as a function of sea ice concentration. The radiative effectiveness of sea ice is defined as $\text{RE} = \text{TOA albedo} (\text{SIC} = 100\%) - \text{TOA albedo} (\text{SIC} = 0\%)$, following Yamano et al. [1997]. The average RE of the Arctic sea ice with respect to the TOA albedo is 0.22 for all-sky conditions [Gorodetskaya et al., 2006]. Arctic sea ice experiences extensive melt during the summer, which decreases the sea ice surface albedo down to about 38%, while during winter, freshly fallen snow increases surface albedo up to 84% according to ground-based observations [Curry et al., 2001]. Surface albedo over the open ocean are calculated as a function of the solar zenith angle increasing from 3% during summer to almost 30% during winter. These surface albedo values define an envelope of the surface albedo ranges shown by thin lines in Figure 2: the upper line representing winter/fall/spring surface albedo and the lower line representing the melt season.

Monthly mean TOA albedo averaged for all-sky conditions is above the range of the surface albedo at low SICs and within the surface albedo range at high SICs. While the solar zenith angle causes a significant change in the open ocean surface albedo, the effects of clouds overwhelm these changes and increase the mean TOA albedo over the open ocean well above the maximum observed ice-free surface albedo. In contrast, over the 100% SICs the combination of sea ice and cloud effects results in the mean TOA albedo lying within the extremes of observed surface albedo values (Figure 2).

Figure 3 shows seasonal changes in the downwelling and reflected SW flux at the TOA, and the sea ice RE, calculated as the product of the downwelling SW flux averaged over the Arctic Ocean and sea ice RE with respect to the TOA albedo for each corresponding month [Gorodetskaya et al., 2006]. The largest impact of sea ice on the SW flux occurs during the summer: for an incident solar flux of about 450 W m$^{-2}$ reaching the TOA in the polar latitudes in June (averaged over the ocean north of 70$^\circ$N), local reduction of SIC from 100% to 0% results in about 100 W m$^{-2}$ decrease in reflected SW radiation at the top of the atmosphere. Sea ice effect on SW flux is relatively small during the fall and early spring reducing to zero during the winter months proportionate to the incoming SW flux. Thus, variations in monthly
mean sea ice induce noticeable changes in the TOA albedo, yet much smaller than the associated changes in the surface albedo because of the compensating effect of clouds.

3.2. Cloud Properties and Surface Radiative Forcing

In the previous section we showed that the sea ice effect on the top-of-atmosphere albedo is strongly compensated by clouds. Thus, changes in cloud shortwave radiative forcing should be taken into account when calculating the magnitude of the ice-albedo feedback. In turn, the effect of clouds on the SW flux absorbed at the surface (net SW radiation defined as the difference between the incoming and reflected SW fluxes at the surface) depends on the surface albedo: for the same cloud albedo, the cloud radiative forcing is greater over the surface with a smaller albedo, while over the highly reflective surface, clouds have little effect on the surface net SW flux. In addition, clouds have a warming effect on the surface by increasing the downwelling LW radiation. Thus, the surface radiative budget is strongly influenced by complex interactions between the sea ice and clouds. In this section we discuss separately cloud effects on the surface SW and LW radiative fluxes using ground-based observations obtained during the year of the SHEBA program [Intrieri et al., 2002a, 2002b; Persson et al., 2002].

The annual mean SW cloud forcing with respect to the net flux is $-10 \pm 0.5$ W m$^{-2}$; that is, clouds decrease the net SW flux by about 10 W m$^{-2}$. The annual mean LW cloud forcing with respect to the net flux is $38 \pm 3$ W m$^{-2}$ [Intrieri et al., 2002a]. However, cloud radiative forcing has a strong seasonal variability. Variations in the cloud effects on the net SW flux at the surface depend on cloud transmittance, the solar zenith angle (insolation), and also on the surface albedo. The cloud effect on the net LW flux at the surface is dominated by the downwelling component, which is dependent on cloud properties, such as cloud base temperature, cloud liquid and ice content, and cloud particle radius. Classical definition of cloud radiative forcing as a function of cloud presence (cloudy versus clear sky) or cloud fraction is insufficient in the Arctic, where a wide range of cloud radiative forcing exists for the same cloud fraction.

At SHEBA location (74°–81°N) polar night lasts from November to February. The effect of cloud on the net SW
radiation outside of the melt season is negligible because of the small amount of solar radiation reaching the surface and high surface albedo (Figure 4). Figure 5 from Intrieri et al. [2002a] shows annual cycles of the downwelling, upwelling, and net radiative fluxes and cloud radiative forcing with respect to each flux. During the melt season, from early July until mid-August, clouds reduce the downward solar flux reaching the surface by about 100 W m$^{-2}$ (Figure 5a). At the same time, cloud effect on the surface net SW flux strongly depends on the surface albedo. The cloud SW forcing with respect to the surface net SW flux is small until the end of May when sea ice is covered by snow with albedo of about 0.85 (Figure 4). The importance of cloud cooling increases dramatically with the progression of melt, when the surface albedo is lowered by snow melting and formation of melt ponds. Figure 4 shows a large drop in the surface albedo during June–July over the sea ice surface without even including the melt ponds. While downwelling SW flux reaches 200 W m$^{-2}$ in May, the net SW flux is below 50 W m$^{-2}$ until the onset of melt in June when the net SW flux doubles (Figure 4). The largest cloud effect on the surface net SW flux occurs in the beginning of July (50 W m$^{-2}$, Figure 5a), when surface albedo is the lowest.

The cloud forcing with respect to both the downwelling and net LW flux is large throughout the year with the annual range of 45 W m$^{-2}$ (from the minimum of 15 W m$^{-2}$ in December to the maximum of 60 W m$^{-2}$ in September) (Figure 5b). During the winter, the downwelling LW flux is the largest source of energy for the surface, and changes in the winter surface air temperature are closely tied to changes in the LW radiation budget [Walsh and Chapman, 1998]. The LW cloud forcing is dominated by the downwelling component, while the effect of clouds on the upwelling LW flux is small [Intrieri et al., 2002a].

Cloud surface longwave forcing depends mostly on the cloud emissivity (or opacity) and cloud emitting temperature [Chen et al., 2006]. Because of the frequent presence of liquid water in the Arctic clouds, their opacity is mostly determined by the cloud liquid water path (cloud liquid water content integrated over the entire atmospheric column). Cloud longwave forcing is highly sensitive to the cloud LW for low LWP values, while the impact of changes in LWP on cloud longwave forcing is practically zero for LWP values greater than 30 g m$^{-2}$. This property of clouds is known as the longwave saturation effect; that is, a cloud with large LWP emits as a blackbody, and thus any further increase in its LWP has no effect on the cloud longwave forcing.

The cloud emitting temperature, in turn, is most often determined by the cloud base temperature. Large values of the downwelling LW flux correspond mostly to clouds with base temperatures warmer than −15°C (Figure 6a) [Shupe and Intrieri, 2004]. Some clouds with colder cloud base temperature are almost indistinguishable from clean-sky emission. The strongest longwave cloud forcing was also found for clouds residing at heights lower than 0.5 km (Figure 6b) [Shupe and Intrieri, 2004]. Cloud base height, however, has mostly an indirect effect on cloud longwave forcing by influencing the cloud base temperature. Chen et al. [2006] showed that the clouds with bases between 1 and 3 km have decreasing longwave forcing with increasing cloud base height, while clouds with a base height lower than about 0.6 km have increasing longwave forcing with increasing height. This relationship shows if the cloud resides in or above the temperature inversion layer (typically about 1 km). The cloud base temperature effect on the LW cloud forcing is enhanced when clouds reside at or near the peak of the near-surface temperature inversion layer. It has to be taken into account that when a cloud resides below the temperature inversion layer, the maximum emitting
temperatures can be located within the cloud rather than at the cloud base [Stramler, 2006]. Presence of liquid in Arctic clouds substantially increases their impact on both the SW and LW fluxes [Shupe and Intrieri, 2004; Zuidema et al., 2005]. The annual mean LW (SW) cloud forcing with respect to the net surface fluxes was estimated at 52 (−21) W m⁻² for liquid-containing clouds compared to 16 (−3) W m⁻² for ice-only clouds [Shupe and Intrieri, 2004]. The time mean cloud optical thickness of liquid clouds is estimated at 10.1 ± 7.8 corresponding to LWP of 37 g m⁻². This by far exceeds the mean optical thickness for ice-only clouds of 0.2 [Zuidema et al., 2005]. Similarly, in mixed phase clouds, the cloud liquid content dictates cloud optical thickness [Zuidema et al., 2005]. For both LW and SW radiative fluxes, cloud optical thickness has a threshold value above which further increase in the liquid water path has no influence on SW or LW cloud forcing. Clouds become saturated in the LW at lower values of LWP than in the SW; that is, as LWP increases, the cloud SW cooling effect continues to increase after the LW warming effect reaches saturation (clouds emit as blackbodies regardless of further increase in LWP) [Shupe and Intrieri, 2004]. To compute cloud SW forcing as a function of cloud LWP over a reflective surface is difficult as cloud SW forcing also depends on the downwelling solar radiation and surface albedo. Monthly mean LWP at SHEBA is about 10–20 g m⁻² during the winter and increases to 90 g m⁻² by the end of summer [Gorodetskaya et al., 2008]. Low sensitivity

Figure 5. Annual cycle of (a) shortwave surface cloud forcing (SW flux SCF) and (b) longwave surface cloud forcing (LW flux SCF) for the downwelling (solid line), upwelling (dashed), and the net flux (dash-dot). Year day (1997) zero is defined as 1 January 1997, 0000 LT. Year days for 1 December, 1 March, 1 June, and 1 September are 335, 425, 517, and 609, correspondingly (adapted from Intrieri et al. [2002a]).

Figure 6. Distribution of the longwave cloud forcing (CFₜₜ) defined as difference between net surface LW fluxes for all-sky and clear-sky conditions distinguished by (a) cloud base temperature and (b) cloud base height. The clear-sky distribution (an uncertainty estimate for these calculations) is also shown in both Figures 6a and 6b. From Shupe and Intrieri [2004], with permission of the American Meteorological Society.
of cloud LW forcing to LWP has been found during the late spring and summer when LWP is large, while high sensitivity of cloud LW forcing to changes in cloud LWP is found in winter and spring when cloud LWP values are small [Chen et al., 2006]. As will be discussed in section 4, the magnitude of the cloud LWP during the present climate has a substantial effect on the cloud SW and LW forcing changes during the 21st century.

Figure 7 reproduced from Zuidema et al. [2005] illustrates the change in the cloud forcing with respect to the net radiative fluxes as a function of cloud optical depth based on low-level mixed phase clouds observed at constant height during 4 days in May 1998 over the SHEBA site. It shows that cloud LW forcing dominates the total cloud forcing for small cloud optical thickness (less than 3). After passing a threshold of 6, further increase in the cloud optical thickness has no influence on the LW radiation, and net cloud forcing is mostly determined by the shortwave component. Cloud effect on the net SW radiation depends on surface reflectivity causing large spread in cloud SW and total forcing, in particular for large cloud optical thickness.

The relative magnitude of the cloud SW and LW forcing discussed above has a large effect on the surface temperature and sea ice thickness. Over perennial sea ice, cloud warming effect overpasses the cooling effect during most of the year. During the winter polar night and transition seasons, cloudy conditions are associated with increased downwelling LW flux and warmer surface air temperatures over sea ice [Walsh and Chapman, 1998]. During the summer melt period, surface temperatures hover around 0°C, while the intensity of net surface melt is strongly influenced by the surface albedo and timing of melt onset: earlier melt onset increases the amount of solar radiation absorbed during the entire melt season [Perovich et al., 2007]. Perovich et al. [2002] explained the melt onset during SHEBA by the decrease in surface albedo due to a rain event which occurred at the end of May. On the other hand, in late spring warm air masses enter the Arctic from lower latitudes. This increases both the cloud liquid water content and effective emitting temperature [Stramler, 2006]. Increased downwelling LW flux during spring can provide energy for initiating the surface melt [Zhang et al., 1996; Weston et al., 2007]. Several studies using sea ice thermodynamic models found a strong relationship between changes in cloud radiative forcing and sea ice thickness. Curry et al. [1993] showed that a large increase in sea ice thickness occurs in response to reduction in annual mean cloudiness. According to Shine and Crane [1984], surface albedo can reverse the effects of cloudiness increase during the summer. They found that cloud increase outside of summer months leads to sea ice thinning, while if clouds increase during July and August, their radiative cooling will be greater than their longwave forcing, slowing down sea ice melt and thus leading to overall thicker ice if the forcing persists for years. In the next section, we will discuss changes in the Arctic cloud forcing predicted for the 21st century and their role in the sea ice cover changes as simulated by the NCAR CCSM3 coupled global climate model.

4. ROLE OF CLOUDS IN SEA ICE CHANGES DURING THE 21ST CENTURY

Coupled models participating in the IPCC AR4 assessment show large differences in future Arctic sea ice thickness and extent when forced with CO₂ emissions as specified by
SRES A1B scenario [Arzel et al., 2006]. The CCSM3 model simulates close to the observed sea ice changes during the 20th century and predicts the most drastic sea ice loss during the 21st century when compared to other models. Already by 2040, CCSM3 predicts that majority of the Arctic basin will be ice free in September. Holland et al. [2006] linked the drastic sea ice decline in the CCSM3 model to increased oceanic heat flux, which triggers the ice-albedo feedback by melting large areas of sea ice. Below we demonstrate that cloud changes also play an important role in sea ice decline. The results are based on one realization (run 1) of a group of seven ensemble members from the SRES A1B simulations.

Figure 8 shows the annual mean sea ice concentration averaged north of 70°N simulated by CCSM3 from 1950 to 2100. The model simulates SIC decrease from about 50% in 1950 to 45% in 2000 and 25% by 2100. The sea ice decline is accompanied by significant warming of the atmospheric boundary layer and a strong increase in the cloud liquid water content: the model predicts a 5 K increase in the boundary layer air temperature and about 50 g m⁻² increase in the cloud liquid water path during the 21st century (Figure 9a). As discussed in the previous sections, the increase in cloud liquid water path increases cloud optical thickness and thus increases both the cloud SW cooling and LW warming. At the same time, cloud effect on the downwelling LW flux has an additional strong contributing factor: increase in the near-surface air temperature.

Figure 9b disentangles the surface albedo and cloud effects on the shortwave radiation and compares cloud effects on the shortwave and longwave radiative fluxes. The thin black line represents the difference between the annual mean values of reflected radiation from 1950 to 2100 and the 1990–1999 average denoted in the equations below by \( \langle \ldots \rangle_9 \) (in Figure 9b values of the reflected flux are multiplied by minus showing anomalies in absorbed radiation due to surface albedo):

\[
\Delta SW^\uparrow = SW^\uparrow - \langle SW^\uparrow \rangle_9 = SW^\downarrow \alpha_s - \langle SW^\downarrow \alpha_s \rangle_9, \tag{1}
\]

where \( SW^\uparrow \) is reflected SW flux, \( SW^\downarrow \) is downwelling SW flux, and \( \alpha_s \) is surface albedo.

The thick black line in Figure 9b represents the anomalies in the net SW flux with respect to the 1990–1999 average:

\[
\Delta SW^{\text{net}} = SW^{\text{net}} - \langle SW^{\text{net}} \rangle_9 = SW^\downarrow (1 - \alpha_s) - \langle SW^\downarrow (1 - \alpha_s) \rangle_9
= SW^\downarrow - SW^\downarrow \alpha_s - \langle SW^\downarrow \alpha_s \rangle_9 + \langle SW^\downarrow \alpha_s \rangle_9. \tag{2}
\]

The difference between the anomalies in the net absorbed SW flux (\( \Delta SW^{\text{net}} \)) and the anomalies in the absorbed radiation due to surface albedo (\( -\Delta SW^\uparrow \)), that is the difference between thin and thick black lines in Figure 9b, shows by how much cloud offsets the increase in the absorbed SW radiation at the surface:

\[
\Delta SW^{\text{net}} - (-\Delta SW^\uparrow) = SW^\downarrow - \langle SW^\downarrow \rangle_9. \tag{3}
\]

The sea ice decline lowers the surface albedo, which decreases the surface reflected SW flux. If no changes in clouds occurred, the surface would gain about 42 W m⁻² more during the summer months in 2100 compared to the 1990s because of the surface albedo drop (thin black line in Figure 9b). At the same time, less shortwave flux reaches the surface because of the larger cloud liquid water path. This cloud effect offsets but does not cancel the increase in the net SW flux due to the strong drop in the surface albedo. Increase in cloud reflectivity reduces the actual gain in the net SW by 2100 compared to the 1990s to about 18 W m⁻² (thick black line in Figure 9b). At the same time, larger cloud liquid water path increases the cloud longwave emissivity, while warmer atmospheric temperatures increase the effective emitting temperature. This leads to a significant increase in the downwelling longwave flux (grey line in Figure 9b). The yearly average increase in the downwelling LW flux by 2100 compared to the 1990s is as high as 34 W m⁻², which is almost twice the magnitude of the net SW flux increase during the summer.

Figure 10 compares seasonal cycles in cloud LWP and boundary layer atmospheric temperature during the first and last decade of the 21st century, and Figure 11 shows
the effect of these cloud properties changes on the surface radiative fluxes. CCSM3 simulates high amount of liquid water in Arctic clouds during the entire year compared to ground-based observations performed during SHEBA from 1997 to 1998 [Gorodetskaya et al., 2008]. Its monthly average LWP exceeds the threshold value for cloud saturation in both SW and LW radiation (about 30 g m$^{-2}$), especially during summer months (Figure 10a). This is limiting the impact of increasing cloud LWP on SW cloud forcing. The largest increase by year 2100 in LWP is predicted during the winter months, when clouds contain smaller amounts of liquid compared to summer (Figure 10a). As discussed in the previous section, downwelling LW flux is strongly influenced by increasing LWP for thin clouds. This allows the increased LWP during winter months to have a noticeable effect on the downwelling LW flux (Figure 11a).

In addition to increase in cloud LWP, the atmospheric layer from 1000 to 850 mbar, where low stratus clouds usually reside, warms up significantly during the entire year but especially strongly during the winter (Figure 10b). Warming

---

Figure 9. Anomalies relative to 1990–1999 in (a) air temperature (T) averaged for 850 and 925 mbar pressure levels (thick line) and cloud liquid water path (LWP) (thin line) and (b) May–August mean decrease in the surface reflected shortwave flux (thin black line), May–August mean surface absorbed shortwave flux (thick black line), and annual mean downwelling longwave flux (grey line) simulated by the NCAR CCSM3 model averaged over the ocean north of 70°N. Data during 1950–1999 are from the 20C3M simulations run 1, and data during 2000–2100 are from the SRES A1B scenario simulations run 1, performed as part of IPCC AR4.

Figure 10. Seasonal cycles during 2000–2010 and 2090–2100 of (a) cloud liquid water path (LWP) and (b) air temperature averaged over 1000, 925, and 850 mbar pressure levels simulated by the NCAR CCSM3 model forced with SRES A1B scenario and averaged over the ocean north of 70°N. The error bars for the model output are standard deviations based on monthly means. SHEBA data are monthly averaged LWP based on daily values available from October 1997 to September 1998. SHEBA error bars are standard deviations based on daily values.
of the low troposphere contributes to the large increase in downwelling LW flux in winter and is the main factor responsible for increased downwelling LW flux during the summer when clouds are optically thick (Figure 11a).

Figure 11b shows seasonal cycles of the Arctic Ocean surface net SW fluxes during the first and last decades of the 21st century together with the contribution of the cloud and surface albedo changes to the difference between the two

Table 1. Differences in Cloud Fraction, Cloud Liquid Water Path, Surface Downwelling Longwave and Shortwave Fluxes, and Net Shortwave Flux Between the Last and First Decades of the 21st Century Averaged Over the Ocean North of 70°N Predicted by the NCAR CCSM3 Model SRES A1B Scenario for Each Montha

<table>
<thead>
<tr>
<th>Month</th>
<th>Cloud (%)</th>
<th>LWP (g m²)</th>
<th>DLF (W m²)</th>
<th>DSF (W m²)</th>
<th>Net SW (W m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>December</td>
<td>9</td>
<td>62</td>
<td>52</td>
<td>0</td>
<td>0</td>
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<tr>
<td>January</td>
<td>7</td>
<td>36</td>
<td>38</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>February</td>
<td>6*</td>
<td>34</td>
<td>33</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>DJF</td>
<td>7</td>
<td>44</td>
<td>41</td>
<td>0</td>
<td>0</td>
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<td>30</td>
<td>-3</td>
<td>0.8</td>
</tr>
<tr>
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<td>5</td>
<td>37</td>
<td>25</td>
<td>-15</td>
<td>2.2</td>
</tr>
<tr>
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<td>2.5</td>
<td>14</td>
<td>14</td>
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<td>17</td>
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<td>27</td>
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<td>-12</td>
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<tr>
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<tr>
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<td>3</td>
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<td>17</td>
</tr>
<tr>
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<tr>
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<td>6</td>
<td>45</td>
<td>35</td>
<td>-2.5</td>
<td>0.8</td>
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</tbody>
</table>

a Abbreviations are LWP, liquid water path; DLF and DSF, surface downwelling longwave and shortwave fluxes, respectively; and SW, shortwave flux. Seasonal means are for December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON). All differences are significant at 95% level, except for those marked with asterisks.
decades. In CCSM3, increased cloud cooling of the surface in the end of the 21st century only partly compensates for the large surface albedo decrease. The largest increase in the net SW flux by 24 W m$^{-2}$ is found in June. At the same time, a large increase in the LW cloud forcing is simulated during the entire year with a maximum of 52 W m$^{-2}$ in December and minimum of 10 W m$^{-2}$ in July (Figure 11a).

Cloud changes are often described in terms of cloud fraction. The CCSM3 model indeed predicts an increase in both cloud fraction and cloud LWP during the 21st century. However, magnitudes of the increase differ depending on the month, and in some months comparable increase in cloud fraction can be accompanied by different increase in cloud LWP and as a consequence different changes in downwelling LW and SW fluxes (Table 1). Changes in the net SW flux, shown in Table 1, also strongly depend on the insolation and surface albedo.

5. SUMMARY AND DISCUSSION

Drastic sea ice retreat has been observed in the Arctic during the last decade of the 20th century and beginning of the 21st century. Most global climate models forced with today’s trends in atmospheric greenhouse gas concentrations predict drastic sea ice decline in the Arctic by the end of the 21st century. The response of the Arctic climate system to initial warming is not linear and involves multiple feedbacks able to accelerate or diminish the surface warming. The sea ice-albedo feedback is considered one of the main factors accelerating sea ice disappearance by increasing the amount of absorbed solar radiation at the surface [see Winton, this volume]. Increased cloud formation is thought to mitigate the Arctic warming by replacing the highly reflective sea ice surface. However, clouds also contribute to the surface warming by increasing the downwelling longwave flux, thus enhancing the greenhouse effect. In the present chapter, we examined satellite and ground-based observations in attempt to disentangle cloud effects on the shortwave and longwave radiative fluxes. We also discussed changes in sea ice, clouds and surface radiative fluxes predicted by the coupled global climate model NCAR CCSM3 during the 21st century.

Over the Arctic perennial sea ice, clouds have a net warming effect on the surface during most of the year by increasing the downwelling longwave flux. The magnitude of the downwelling longwave flux strongly depends on cloud properties, such as cloud liquid water content and cloud base temperature. Clouds in the Arctic are usually mixed phase with frequent presence of liquid during winter and continuous large amounts of liquid during the summer. During winter, when cloud events are occasional and clouds are thin, even a small increase in the cloud liquid water content significantly increases the downwelling longwave flux. Cloudy skies in winter are always associated with warmer surface temperatures because of the efficient energy transfer from the relatively warm cloudy atmosphere to the surface by increasing the downwelling LW flux. CCSM3 simulations for the 21st century predict an increase in the cloud liquid water content together with warming of the atmospheric boundary layer. The largest increase in the cloud liquid water content and near-surface atmospheric temperature is predicted during the winter with large impact on the downwelling longwave flux. Thermodynamic model studies showed that this is the time when cloud changes have the largest impact on the sea ice thickness [Curry et al., 1993; Shine and Crane, 1984].

In summer, clouds are present practically continuously and contain large amounts of liquid. Above certain threshold, clouds emit as blackbodies, and further increase in the cloud liquid content has no effect on the cloud longwave forcing. For this reason, increase in the cloud base temperature plays a more important role for downwelling longwave flux during the summer. Cloud cooling effect also becomes significant during the summer because of both high cloud optical thickness and thus high cloud albedo and the lowered surface albedo over melting sea ice. Clouds reduce monthly mean shortwave radiation reaching the surface by up to 100 W m$^{-2}$ on average over the Arctic Ocean during the summer. At the same time, an increase in the shortwave radiation absorbed by the surface-atmosphere column up to the same magnitude can occur when sea ice gives way to the open ocean for all-sky conditions. Simulations with the CCSM3 model showed that during the 21st century, clouds significantly diminish but do not cancel the effect of reduced surface albedo on the surface-absorbed shortwave flux.

The relative role of the shortwave and longwave cloud forcing during the 21st century depends strongly on how the model simulates cloud properties. Mixed phase cloud parameterization in the CCSM3 model allows liquid water to be present at temperatures between $-10$ and $-40^\circ$C. Together with excessive poleward moisture flux, this leads to very high liquid water content in the Arctic clouds, which is overestimated compared to the SHEBA ground-based observations. There are not enough Arctic-wide observations to say if this overestimation is significant. Part of the model’s overestimation of the cloud LWP comes from the fact that CCSM3 values are averaged over the ocean and sea ice areas north of 70°N, while SHEBA measurements are from perennial ice area. In order to improve model simulation of cloud properties, there is a need for more year-round ground-based observations of clouds and radiative fluxes for various sea ice conditions. Correctly predicting seasonality
of cloud changes plays an important role. Overestimation of cloud LWP during summer decreases changes in both the shortwave and the longwave cloud forcing. During winter, CCSM3 cloud LWP is comparable to the observed values when cloud LWP changes are largest and the sensitivity of the cloud longwave forcing to LWP is high.

The NCA CCSR3 model predicts an increase in cloud fraction, cloud liquid water content, and near-surface atmosphere warming in the Arctic during the 21st century. These atmospheric changes provide additional energy to the surface; thus increase in cloudiness facilitates rather than reduces the surface warming accompanying sea ice decline. The ultimate sea ice changes depend on a combination of factors. The present study showed that clouds directly respond to initial Arctic warming and are among the key factors accelerating the Arctic sea ice decline. Given the larger magnitude of the longwave flux increase induced by clouds compared to the absorbed shortwave flux, as predicted by CCSR3, the cloud surface warming may be a more dominant driver of sea ice loss than the sea ice albedo feedback.

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