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Role of Arctic sea ice in global atmospheric circulation: A review

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ABSTRACT

Formed by the freezing of sea water, sea ice defines the character of the marine Arctic. The principal purpose of this review is to synthesize the published efforts that document the potential impact of Arctic sea ice on remote climates. The emphasis is on atmospheric processes and the resulting modifications in surface conditions such as air temperature, precipitation patterns, and storm track behavior at interannual timescales across the middle and low latitudes of the Northern hemisphere during cool months. Addressed also are the theoretical, methodological, and logistical challenges facing the current observational and modeling studies that aim to improve our awareness of the role that Arctic sea ice plays in the definition of global climate. Moving towards an improved understanding of the role that polar sea ice plays in shaping the global climate is a subject of timely importance as the Arctic environment is currently undergoing rapid change with little slowing down forecasted for the future.

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1. Introduction

The Arctic climate is rapidly shifting as documented by the unprecedented warming of its atmosphere, ocean, and land. Such changes have been in turn accompanied by various environmental modifications including those observed in the terrestrial and marine ecosystems, hydrosphere, and cryosphere (ACIA, 2005). Sea ice is a

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critical component of the Arctic's marine system that has changed substantially over the past several decades. Since the late 1970s, its extent has consistently declined in all months of the year and since the early 1990s the retreat has been accelerating. Each year since 2001, sea ice extent and thickness have declined throughout the entire Arctic with prominent minima (Serreze et al., 2007). In 2007, the September extents reached record lows since satellite measurements began in 1979 with average extent reaching 4.28 million km^2, a value 23% below the previous monthly record set in 2005, and 39% below the long-term 1979–2000 mean (Stroeve et al., 2007). Giles et al. (2008) found similar trends in the average Arctic winter sea ice thickness anomalies in 2007 that measured 0.26 m or 10% below the 2002–2007 average with greatest thinning observed in the western Arctic where the anomalies reached 0.49 m below the six-year mean. Projections of simulated Arctic sea ice cover trends are expected by some to result in a seasonally ice-covered Arctic by the end of the 21st century (Zhang and Walsh, 2006). The magnitude of the decline in Arctic sea ice over the past several decades has been, however, under predicted by most model simulations. For instance, Stroeve et al. (2007) estimate the recent sea ice extent minima to be approximately 30 years ahead of the ensemble mean model forecasts and argue that between 33 and 38% of the 1953–2006 summer trend may be anthropogenically forced. This percentage further rises to between 47 and 57% for the 1979–2006 period. Consequently, if the models indeed underestimate the magnitude of the anthropogenic influence in the Arctic, the ice-free September Arctic Ocean that is being predicted for some point after the year 2100 may be observed much sooner. In addition to anthropogenic influences, the recent loss of Arctic sea ice also has been attributed to factors such as natural climate variability in the Arctic and ice–ocean feedbacks (e.g., Overland et al., 2008). Overland and Wang (2005) argue that the accompanying atmospheric warming referred to by Overland et al. (2008) as the "Arctic warm period" may be related to a changing surface albedo that promotes the absorption of more solar energy in areas of reduced sea ice. When coupled with anthropogenic influences the conditions may be providing persistence to a new regime of reduced sea ice.

Arctic sea ice has the potential to be an integral player in the global climate system (McBean et al., 2005; Liu and Alexander, 2007). It has the capacity to significantly mediate the exchange of radiation, sensible heat, and momentum between the atmosphere and the ocean, impacting the climate through the aforementioned ice-albedo feedback mechanism (Curry et al., 1995). In the tropics, the sea surface is known to exert significant influence on the atmospheric flow as excess energy is transported through latent and sensible heating from the ocean to the atmosphere. The El Niño-Southern Oscillation (ENSO) is a classic phenomenon that illustrates this interaction when unusually warm sea waters in the central and eastern basin periodically alter atmospheric circulation patterns and weather regimes around the globe through atmospheric teleconnections (Philander, 1990). In the high latitudes, the interaction between the ocean surface, sea ice, and the atmosphere also plays an important role in climate variability (Walsh and Johnson, 1979; Kushnir et al., 2002), but here the fluctuations in atmospheric conditions are the dominant driving force of the sea surface conditions. Still, the air–sea interaction is not exclusively a one-way process as sea surface conditions are known to drive the ocean's impact on the atmosphere as well. Walsh and Johnson (1979) argue that the impact of winter sea ice on the atmosphere, for instance, outweighs that of the effect of the atmosphere on sea ice. These results have since been confirmed by Slonosky et al. (1997) and Alexander et al. (2004) at interannual to decadal timescales. As well, Deser et al. (2000) demonstrate that atmospheric circulation in the Atlantic sector is very sensitive to sea ice variations, particularly in the active storm track region east of Greenland and Dethloff et al. (2006) argue that changes in the surface processes have the ability to feed back on the global climate system in part through an atmospheric wave bridge between the Arctic region and the lower latitudes.

Simulations of future Arctic climate forecast a continued warming accompanied by an accelerated decline in sea ice cover. These changes have been in part attributed to the positive feedback mechanisms involving snow and sea ice (Kattsov et al., 2005) and anthropogenic greenhouse gas increases (e.g. Vinnikov et al., 1999; Holland et al., 2006; Zhang and Walsh, 2006). There are, however, large discrepancies in the exact predictions of Arctic's future climate and its sea ice, a problem that in turn introduces significant uncertainties about how the Arctic may potentially shape climate variability and change around the globe. The purpose of this review is to synthesize the published observational and modeling efforts that document the potential impact of Arctic sea ice on remote climates. The thrust of the review focuses on the synthesis of the influence of Arctic sea ice on large-scale atmospheric circulation, and the inherent modifications to atmospheric temperatures, precipitation patterns, and storm track behavior across the middle and low latitudes of the Northern Hemisphere during the cool seasons. Also addressed here are the challenges facing current studies that aim to improve our awareness of the role that Arctic sea ice plays in the definition of global climate. Fig. 1 depicts the various locations and areas in the Arctic mentioned in this review and Table 1 summarizes the primary characteristics of the modeling studies synthesized in this review, for reference.

2. Recent status of Arctic sea ice

The impact of Arctic sea ice on global climate has been studied through a number of its characteristics including concentration, extent, and more recently thickness. Ice concentration is defined as the fraction of the ocean covered by ice while sea ice extent refers to the area enclosed by the ice edge where the ice concentration is at least 15% (Lemke et al., 2007). The extent of Arctic sea ice varies considerably during any given year with a maximum in February–March and minimum in August–September (Comiso, 2006). Between 1979 and 2006, Arctic sea ice extent typically reached a maximum on 7-March, covering between 14 and 16 million km^2. This coverage gradually decreased during the spring and summer months to an annual minimum ranging between 5 and 7.5 million km^2 around 17-September (Deser and Teng, 2008b) (Fig. 1). Over this time period winter long-term mean sea ice marginal zones reached the Labrador Sea, the Greenland and Barents Seas, the Bering Sea, and the Sea of Okhotsk. The zones retreated northwards to coastal regions of the Arctic Ocean and the Canadian Archipelago in the summer. Arctic sea ice displays considerable spatio-temporal variation with two

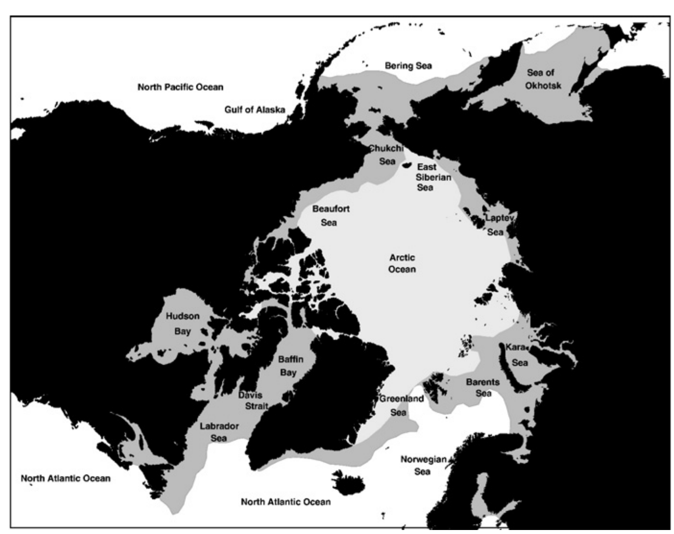

Fig. 1. Geography of the Arctic region. Shaded in dark (light) gray is the mean sea ice extent (1978–2007) during March (September) when sea ice is at its maximum (minimum) extent (after Deser and Teng, 2008a).

Table 1
Principal modeling studies that examined the large-scale atmospheric response to changes in Arctic sea ice conditions in winter.

Study	Model resolution (vertical/horizontal [lat × long])	Sea ice forcing	Model	Time of analysis
Fletcher et al. (1973)	2 layers/4° × 5°	Completely removed sea ice in all months	UCLA GCM; fixed annual SSTs and sea ice	Feb
Newson (1973)	5 layers/3° × 5°	Completely removed sea ice in all months	UKMO GCM	
Warshaw and Rapp (1973)		Completely removed sea ice in all months	UCLA GCM	Jan
Herman and Johnson (1978)	4° × 5°	Varied sea ice margins (maxima and minima) reflective of observed conditions 1973–1977 (Pacific) and 1960–1978 (Atlantic)	Godard (GISS) AGCM	Jan–Feb
Royer et al. (1990)	20 layers/2.8° × 2.8°	Completely removed sea ice in winter	French spectral GCM; T42	Dec–Feb
Raymo et al. (1990)	9 layers/8° × 10°	Reduced sea ice during all months; all NH sea ice removed in Sep; in Mar sea ice limits set equivalent to modern Sep limits	GISS II GCM (Hansen et al., 1983)	Dec–Feb; Jun–Aug
Rind et al. (1995)	9 layers/8° × 10°	Varied sea ice thickness and extent	Hansen et al. (1983) GISS GCM	
Murray and Simmonds (1995)	9 layers/5.75° × 3.33°	Gradually reduced Arctic sea ice through increase of leads fraction −5%, −20%, −50%, −80%, and −100% (complete ice removal)	GCM (Simmonds, 1985); rhomboidally truncated at 21 waves	Jan
(Honda et al., 1996; Honda et al., 1999)	30 layers/5.6° × 5.6°	Forced model with Sea of Okhotsk sea ice cover during heaviest ice years (1978, 1979, 1980, 1983) and lightest ice years (1975, 1984, 1991, 1994)	Global primitive equation model with a spectral transform method (Kodera et al., 1990) AGCM; triangular truncation at wave-number 21	Jan–Feb
Parkinson et al. (2001)	9 layers/4° × 5°	Varied sea ice concentration between −50% and +50% in 17 simulations from present-day values (1979–1986) in both hemispheres.	GISS AGCM version B224 (Hansen et al., 1997)	All months
(Magnusdottir et al., 2004; Deser et al., 2004; Deser et al., 2007	18 layers/2.8° × 2.8°	Used realistic spatial structure of sea ice extent; conducted various experiments with distinct ocean sea ice boundary conditions; (e.g. ICE1 simulation corresponds to fixed seasonal SST climatology + the observed 40-yr areal sea ice extent climatology climatology + trend in sea ice extent over eastern and western north Atlantic (1958–1997); ICE2 corresponds to the fixed SST climatology + 40-year sea ice climatology climatology + 2 × trend)	NCAR CCM3 AGCM (Kiehl et al., 1998); hybrid-sigma coordinate system; T42; 61 year runs; 240 member ensemble and linear baroclinic model (Peng et al., 2003) were also used in Deser et al. (2007)	Dec–Mar
Alexander et al. (2004)	18 layers/2.8° × 2.8°	Used realistic spatial structure of sea ice conditions; simulations forced with observed Arctic sea ice conditions with most (1982/83) an least (1995/96) ice coverage during 1979–99 period	AGCM — CCM3.6 (Kiehl et al., 1998; Hack et al., 1998); T42; used 50-member ensembles	Dec–Feb
Sewall and Sloan (2004)	Spectral 3.75° × 3.75°	Generated Arctic sea ice cover and SSTs to 2050; forced model with projected conditions	NCAR CCSM (Boville et al., 2001)	All seasons but only discussed DJF
Chiang and Bitz (2005)	5.8° × 6°	Model forced with Last Glacial Maximum (LGM) land ice cover distribution for both hemispheres (Peltier 1994), one at a time	CCM3 GCM (Kiehl et al., 1998); coupled to 50 m fixed depth ocean mixed layer that allows thermodynamic atmosphere–ocean interaction but no ocean dynamics; T31 × 15	All months
Singarayer et al. (2005)	19 layers/2.5° × 3.75°	Forced model with sea ice climatologies derived from NIC, NASA, and Bootstrap records using Met Office sea ice as control for total of 4 experiments	Atmospheric GCM HadAM3 (Pope et al., 2000)	Dec–Feb
Singarayer et al. (2006)	19 layers/2.5° × 3.75°	Forced model with observed sea ice from 1980 to 2000 and predicted sea ice reductions until 2100 under two scenarios (one where observed linear trends are extrapolated where Mar and Sep ice declines by 8% an 39%, respectively; another that assumes a faster rate of decline by 26% and 82% in Mar and Sep, respectively)	Atmospheric GCM HadAM3 (Pope et al., 2000)	Dec–Feb
Dethloff et al. (2006)	19 layers/0.5° × 0.5°	Forced model with improved sea ice and snow albedo feedback mechanisms; sea ice-albedo scheme takes into account snow, pure sea ice, melt ponds (Køltzow, 2007)	Coupled atmosphere–ocean GCM ECHO-G (Zorita et al., 2004)	Dec–Feb
Gerdes (2006)	24 layers/2.5° × 2°	Used four simulations with distinct seasonal cycles of sea ice concentration and thickness; THI95 experiment used monthly sea ice thickness averaged for 1994–1996 period; THI65 used thickness conditions derived from 1964–1966; CON95 and CON65 used 1994–96 and 1964–66 ice concentration fields, respectively; analyzed differences between ensemble means for 1990s forcing and 1960s forcing	Atmospheric model GFDL AM2 (Anderson et al., 2004)	Jan–Mar

distinct leading geographic patterns dominate both the winter and summer sea ice variability as determined by an Empirical Orthogonal Function (EOF) analysis outlined in Wilks (1995) (e.g. Walsh and Johnson, 1979; Slonosky et al., 1997; Deser et al., 2000; Deser and Teng, 2008a,b). In winter, the leading pattern exhibits an out-of-phase fluctuation in sea ice between the eastern and western sectors of the Atlantic and Pacific. More specifically, there is an out-of-phase variation between sea ice conditions in the Labrador and Greenland–Barents Seas and an out-of-phase variation between sea ice conditions in the Bering Sea and the Sea of Okhotsk. The second winter pattern displays a uniform behavior of sea ice variability throughout the Arctic, either showing above or below-normal conditions throughout the marginal ice zones. During the summer, the leading geographic pattern exhibits uniform behavior throughout most of the Arctic. The second pattern shows an out-of-phase variation between the Barents and Kara Seas and between the East Siberian and Beaufort Seas (Deser and Teng, 2008b). Observational records of Arctic sea ice thickness are sparse as satellite remote sensing techniques capable of mapping it have not been available until recently. Consequently, our current understanding of sea ice thickness in the Arctic is based largely on numerical models of (e.g., Rothrock et al., 2003). Thickest ice is present along the northern Canadian Archipelago and the thinnest ice is observed throughout the peripheral seas including Beaufort, Laptev, Chukchi, Barents, and Kara Seas. Highest winter variability is present in the southern Beaufort Sea and along the northern Barents, Kara, and Laptev Seas (Bourke and Garrett, 1987). Laxon et al. (2003) estimate the 1993–2001 mean winter ice thickness in the area between 65°N and 81.5°N at 2.73 m that increased from about 2 m near Siberia to 4.3 m off the coast of Canada and Greenland. These magnitudes showed an interannual variation of 24.5 cm or 9% during the examined time period.

3. Observed and projected trends in Arctic sea ice

Lemke et al. (2007) argue that sea ice decline in the Arctic may date back to the early 1970s. Between 1979 and 2007, Arctic sea ice extent decreased significantly by -0.52×10^6 km^2, or 5% per decade totaling an area of 1.76×10^6 km^2 (Deser and Teng, 2008a). The magnitude of this negative trend increased from -0.35×10^6 km^2 per decade during the 1979–1993 period, to -0.9×10^6 km^2 per decade observed between 1993 and 2007 (Comiso, 2006; Deser and Teng, 2008a). Between 1979 and 2007, the trend was greatest during summer at -0.7×10^6 km^2 per decade, or -3.4% per decade, whereas during the winter it was approximated at -0.5×10^6 km^2 per decade, or -9.0% per decade. Both, the summer and winter magnitudes are statistically significant at the 99% confidence level. Changes in sea ice extent and concentrations have not been geographically uniform across the polar region. Deser and Teng (2008a) found the trends to exhibit distinct spatial patterns depending upon time period and season examined. Between 1979 and 2007, for instance, winter sea ice declined in all the marginal seas, with the exception of the Bering Sea. During an earlier period, between 1979 and 1993, on the other hand, concentrations increased in the Labrador and Bering Seas and decreased in the Greenland and Barents Seas and in the Sea of Okhotsk. Since 1993, sea ice has been declining throughout the marginal seas in winter, with greatest amplitudes observed in the Atlantic sector. During summer between 1979 and 2006, sea ice concentrations have been declining throughout the Arctic with largest changes occurring in an area stretching from the Laptev Sea eastward to the Beaufort Sea. Up until 1993, however, sea ice declined in the East Siberian Sea and expanded in the Barents, Kara, and eastern Beaufort Seas. Since 1993, summer sea ice concentrations have been on the decline throughout the Arctic.

Arctic perennial or multi-year ice cover also has been rapidly declining at a rate of about -10% per decade since the early 1980s (Comiso, 2006; Mahoney et al., 2008). Belchansky et al. (2005) find the decline to be most dramatic in the Alaskan and Siberian Arctic in the Beaufort, Chukchi, and East Siberian Seas, between 1979 and 2004. There has been an overall decline since 1933 in multiyear sea ice in the Russian Arctic with periods of both advance and retreat (Mahoney et al., 2008). The authors find summer sea ice retreated during two periods, one between the 1930s and 1950s and the other that started in the mid-1980s and was ongoing in 2006. The ice partially recovered during the 30 year period between the mid-1950s and mid-1980s. The most recent retreat since the 1980s has been Arctic-wide and occurs during all seasons, whereas the retreat observed at the turn of the 20th century was only confined to the overall Russian Arctic during the summer months. Maslanik et al. (2007b) confirm these results as they find the overall mean age of the Arctic multiyear ice to have decreased since 1982 with the oldest ice types almost gone today as 58% of the multiyear ice now consists of relatively young 2- to 3-year old ice. This fraction is considerably higher than the 35% observed in the mid-1980s. The remaining older and thicker ice is also now confined to a much smaller portion of the Arctic Ocean. According to several studies, sea ice thickness of the central Arctic pack appears to be thinning. In the region around the North Pole, for instance, Haas et al. (2008) document a reduction of 44% in late-summer mean sea ice thickness between 2001 and 2007. During this time period, late-summer second year mean ice thickness declined by 25% or 1.81 m. Rothrock et al. (1999) were the first to provide an Arctic-wide analysis of sea ice thickness trends, finding a 40% decline throughout the central Arctic Basin between two time periods, 1958–1976 and 1993–1997. Changes were smallest in the southern Canada Basin and greatest in the Eurasian Basin. In their 2003 study the authors also note a decline in Arctic sea ice thickness of 0.6–0.9 m between 1980 s and 1996 with maximum thickness observed in the mid-1960s. Miller et al. (2007) confirm these trends by noting a steady decline by 3 to 4% in Arctic sea ice thickness between 1980 and 2000. This decline was found most robust in the Eastern Arctic basin where it has been occurring since the early 1960s and accelerating after 1980. Most recently, Belchansky et al. (2008) examine changes in Arctic sea ice thickness between 1982 and 2003 using a neural network algorithm for the Arctic basin. Their analysis, however, detected no significant linear trend in ice thickness over the entire 22-year period, although it suggested an increase of $+7.6\pm0.9$ cm per year in the mean January thickness values between 1982 and 1988 in most regions of the Arctic. Ice thickness then decreased through 1996 at a rate of -6.1 ± 1.2 cm per year, modestly increasing thereafter in the central Arctic through 2003 at a rate of $+2.1\pm0.6$ cm per year. Serreze et al. (2007) also report no clear long-term trend in sea ice thickness over the Arctic basin since 1979.

The annual sea ice areal extent in the Northern hemisphere is projected to further decline in coverage at -3.28×10^5 km^2 per decade, or -28.7% in the last 20 yr of the twenty-first century in comparison to the 1979–1999. By the end of the 21st century, the multiyear ice coverage also is predicted to decline, on average by -3.82×10^5 km^2 per decade, or by -56.8% (Zhang and Walsh, 2006). Bitz and Roe (2004) expect that sea ice will thin fastest in regions where it is originally the thickest and both, Gordon and O'Farrell (1997) and Zhang and Walsh (2006) predict summer sea ice extent to decline faster than winter ice areas in the 21st century, with the Arctic facing seasonal ice cover by the middle of this century.

The recent loss of Arctic sea ice has been attributed to a combination of factors including natural variability in the large-scale atmospheric circulation, global response to anthropogenic factors such as increased levels of greenhouse gases, and ice–ocean feedbacks (Overland et al., 2008). Between 1980 and 1999, the trends in surface warming and retreat of sea ice from the Arctic were largely attributed to the state of the large-scale atmospheric circulation marked by the presence of positive phases of the Arctic Oscillation (AO) (e.g., Serreze et al., 2007; Maslanik et al., 2007a; Deser and Teng, 2008a,b) and the Pacific North American-like (PNA*) (Quadrelli and Wallace, 2004) patterns (Overland et al., 2008) that were associated with primarily zonal wind flow (Overland and Wang, 2005). Since 2000, the large-scale atmospheric circulation has significantly changed as both the PNA* and the AO have become more variable and neutral and the wind circulation pattern shifted to meridional blowing towards the central Arctic. Overland and Wang (2005) point to a similar pattern of circulation in the late 1930s that coincided with anomalous winter surface warming of $+4$ °C at Spitzbergen. Regardless of this change, however, the warming and ice loss trends have since continued or even accelerated throughout the Arctic testifying to the presence of alternative mechanisms that must help drive the changes in sea ice. Overland and Wang (2005) and more recently Maslanik et al. (2007a), for instance, point to a “dipole pattern” synonymous with high pressure over the Canadian Arctic and low pressure over the Siberian

Arctic. Maslanik et al. (2007a) also identify a “central Arctic” pattern marked by a low pressure cell over the Arctic Basin. Both patterns continue to be present since the late 1980s and have contributed to the observed sea ice reduction in the western and central Arctic. As well, the recent warming of the Arctic may be related to changes in surface albedo that transpire from the absorption of more solar energy in areas of reduced sea ice (Overland and Wang, 2005). When combined with anthropogenic influences (Zhang and Walsh, 2006) these conditions may perpetuate a new state of climate and reduced sea ice throughout the Arctic.

4. Influence of Arctic sea ice on oceanic circulation

The strong potential of Arctic sea ice to influence global climate through oceanic circulation is widely recognized. Specifically, numerous studies have addressed the relationship between Arctic sea ice and the activity of the north Atlantic Thermohaline Circulation (THC). McBean et al. (2005) defines the THC as a global-scale overturning in the ocean that transports significant amount of heat towards the poles through warm and highly saline water at surface and cold, less saline waters towards the equator at depth. In the Northern hemisphere the overturning occurs in the western north Atlantic in the Greenland, Irminger, and Labrador Seas (Broecker, 1990) and is known to be sensitive to surface water density produced by temperature and salinity fluctuations that are in part affected by an outflow of fresh water from the Arctic. Melting of Arctic sea ice increases this outflow and is suspected to reduce the overturning and thus the oceanic flux of heat to northern high latitudes weakening the THC (Broecker, 1997). Dixon et al. (1999), however, argue that freshwater flux changes that result from increased precipitation dominate the THC activity, whereas Mikolajewicz and Voss (2000) and Gregory et al. (2005) find changes in surface heat fluxes to be of principal interest in driving the THC. Although most climate models forecast a weakened THC in a warmer climate with large melting of sea ice (e.g. Lemke et al., 2007), there are some models (e.g. Latif et al., 2000; Gent, 2001) that do not show a significant reduction of the THC under such conditions. Many attribute this discrepancy to our lack of understanding of the forces that drive the THC but Saenko et al. (2004) argue that some of it may be present due to the models' inability to accurately simulate the current climate of the high-latitude north Atlantic, including surface temperatures and sea-ice extent conditions. The authors demonstrate this by running a series of experiments where they vary sea ice extent and sea surface temperature conditions under similar radiative forcing scenarios. They find that the THC response to the forcing changes depending upon the initial climate conditions in the modeling scenario. Colder north Atlantic initial conditions with extensive sea ice cover, for instance, can produce more stable THC. The circulation can also strengthen during the same climate warming scenario when the initial cold bias becomes even more extreme with maximum sea ice extending well south of present-day latitude. Saenko's results suggest that sea-ice may have a stabilizing effect on the weakening of the THC under a global warming scenario. The hypothesis has been recently confirmed by Levermann et al. (2007) who propose two feedback mechanisms, one that reduces the strength of the THC and the other that enhances it. The first, called “the temperature effect” argues that initial atmospheric warming reduces the difference between surface and air temperature thereby reducing oceanic heat loss and thus the strength of the THC. The second mechanism stresses the importance of Arctic sea ice. It is referred to as a “sea ice effect” mechanism where a rise in near-surface temperatures enhances melting of sea ice exposing more of the ocean surface to the atmosphere. This process in turns causes the ocean to lose more heat strengthening the THC. Similarly to Saenko's results, the stabilizing effect of sea ice becomes stronger when the initial THC is weak and sea ice reaches farther south. Thus, an initially weak THC can weaken less under global warming. This information underscores the importance of Arctic sea ice in the state of oceanic circulation of the north Atlantic with global climatic implications.

5. Remote response to Arctic sea ice variations

The remote response to changes in sea ice is driven by the interaction between anomalous surface fluxes and planetary-scale circulation (Alexander et al., 2004) (Table 1). Herman and Johnson (1978) (Table 1) provide a brief overview of observational studies that focused on the statistical connections between polar sea ice variations and remote climate impacts dating back to the early 20th century. Hildebrandsson (1914) was among the first to suspect that there may be a correlation between winter conditions in the Northern Hemisphere, especially over Europe and ice conditions in the East Greenland Sea during the previous summer. A few years later, Wiese (1924) presented evidence suggestive of relationships between ice conditions in the East Greenland and Norwegian seas and subsequent storm frequency and precipitation over Northern Europe and Scandinavia. In 1936, Scherhag proposed a connection between temperature conditions in the north Atlantic and sea ice conditions in the East Greenland Sea. Walker (1947) noted a relationship between ice conditions around Newfoundland and Icelandic/north European pressures and Schell (1956, 1970) observed a correlation between ice margin in the Arctic and temperatures and pressures over Europe. Finally, Defant (1961) pointed to a positive feedback between ice extent and atmospheric pressure in the Arctic. Many studies have since added to this body of knowledge and contributed towards our understanding of the connections between Arctic sea ice and global climates. To date, the numerous observational and modeling efforts have explored and documented the influence of Arctic sea ice on various aspects of our climate including the mean state of surface temperature, precipitation, storm track activity, and general circulation of the atmosphere across many regions of the Northern hemisphere from the Arctic to the tropics.

5.1. Atmospheric circulation

Herman and Johnson (1978) were among the first to show that interannual variations in wintertime ice cover that resemble observed conditions in the seas of the north Atlantic and north Pacific can influence large-scale circulation features as far as the subtropics. Owing in part to their impact on latitudinal temperature gradients variations in Arctic sea ice can modify the dynamical atmospheric field that connects the Arctic to the lower latitudes. Altered are the strength of the westerlies and intensities of storms throughout the midlatitudes, the subtropics, and tropical regions of the Northern Hemisphere. Less ice in the Arctic results in a significant decrease in the speeds of the westerlies and intensities of storms poleward of 45°N, and a general increase in the westerlies in the lower latitudes of the tropical and subtropical regions. The opposite impacts are observed during times when the ice cover is more expansive (e.g. Royer et al., 1990; Murray and Simmonds, 1995; Honda et al., 1999). The reduction in the westerly flow in the midlatitudes has been explained by a concurrent increase in the mid-tropospheric geopotential heights over the high latitudes and a lessening of the meridional temperature gradient between the high and low latitudes. The pattern of reduction in the strength of the westerlies at midlatitudes and an accompanying increase in the subtropics, around 30°N, has been attributed to a small southward displacement and intensification of the westerly jet maximum.

Surface air temperatures generally decline throughout the middle latitudes upon the retreat of Arctic sea ice. These changes coincide with reduced geopotential heights over similar areas, conditions indicative of an equivalent-barotropic atmosphere where atmospheric pressure is only a function of density and isobaric surfaces are isopycnic surfaces and vice versa (Royer et al., 1990; Deser et al., 2000; Alexander et al., 2004). Royer et al. (1990) (Table 1) note that this mid-latitudinal cooling results in an atmospheric wave-number four pattern with four areas of reduced geopotential heights at the mid-tropospheric level centered above the east Pacific (40°N, 150°W), Sea of Japan (40°N, 130°E), the Black Sea and Caspian Sea (45°N, 45°E), and the west Atlantic (35°N, 70°W). The authors note a general decline in pressure over the north Pacific/American and an increase over the north Atlantic/Eurasian sectors when Arctic sea ice cover is removed. The rise in surface pressure occurs in two principal areas, one over Iceland and the Norwegian Sea and the other over Siberia and the western Bering Sea. The area of maximum decline in pressure over the north Pacific is located south of the Gulf of Alaska. Consequently, the Icelandic low pressure system over the north Atlantic undergoes a small southward shift and a weakening of its pressure minimum while the Aleutian low over the north Pacific deepens and shifts eastward reducing the East Pacific subtropical high off the California coast (Royer et al., 1990). In their study, Raymo et al. (1990) (Table 1), also notice an increase in the strength of the Azores high from the north Atlantic into western Europe and over Scandinavia together with a stronger high pressure system over the eastern sections of North America when Arctic sea ice declines.

The nature of the empirical relationship between Arctic sea ice variability and remote climate response has also been examined in several studies. Murray and Simmonds (1995) (Table 1), show reductions in Arctic sea ice concentrations to be non-linearly related to changes in 850 mb temperatures and midlatitudinal westerlies, suggesting that areas located farther away from the location of forcing may experience more non-linear responses. Parkinson et al. (2001) (Table 1) find both hemispheres to be more affected by sea ice concentration decreases than by increases but note a linear relationship between simulated average surface air temperatures and the magnitude of the ice concentration change. Most recently, Alexander et al. (2004) confirm these findings when they point to a linear relationship between ice extent and atmospheric response in the mid-troposphere of the Atlantic sector. Upon a closer examination, Deser et al. (2004) (Table 1) show the response of the atmosphere winter circulation to north Atlantic SSTs and sea ice anomalies to be linear in nature with respect to the size of the sea surface forcing, but nonlinear with respect to the direction of the forcing, supporting similar results presented earlier by Parkinson et al. (2001). The authors find positive sea surface temperature and negative sea ice anomalies over the north Atlantic to display larger impact on the general circulation of the atmosphere than negative sea surface temperature and positive sea ice anomalies.

Until recently, studies focusing on the assessment of the impact of Arctic sea ice on the general atmospheric flow have used sea ice concentration and extent as the principal variables to represent sea ice characteristics. With recent advancements in the estimation of sea ice thickness, Gerdes (2006) (Table 1) attempts to examine its effect on atmospheric circulation in various General Circulation Model (GCM) experiments. The study examines the differences between atmospheric conditions during the 1964–65 and 1994–96 time periods that represent extreme cases of large and small ice volume, respectively. The experiments replicate some of the large-scale atmospheric modifications associated with Arctic sea ice thinning that have also been shown during reduced sea ice extents or concentrations. These changes include, for instance, reduced sea level pressures over the central Arctic and their projection onto the north Atlantic Oscillation (NAO) (i.e. Section 5.1.1). Unlike with sea ice extents and concentrations, however, Gerdes finds the changes in sea ice thickness not as easily associated with any major atmospheric circulation patterns and the feedbacks not as straightforward as with sea ice extents. The author argues that it is possible that sea ice thickness may produce significant changes in atmospheric flow only once the thickness falls below a certain value over a large enough area.

5.1.1. Atlantic sector

Changes in Arctic sea ice have the potential to influence the formation of extratropical teleconnection patterns (Liu and Alexander, 2007). Over the Atlantic basin and the European continent, the large-scale atmospheric response to polar sea ice cover is most prevalent at middle and upper levels of the atmosphere and often resembles the internal mode of atmospheric variability, the north Atlantic/Arctic Oscillation (NAO/AO) that reaches equilibrium in 2–2.5 months after the initial change (Deser et al., 2007) (Table 1). Negative phases of the NAO/AO frequently coincide with reduced ice conditions in the Atlantic often (Murray and Simmonds, 1995; Alexander et al., 2004; Deser et al., 2004; Dethloff et al., 2006). The response is equivalent barotropic from the surface to the tropopause (e.g., Deser et al., 2004). Deser et al. (2000) performed an EOF analysis (Wilks, 1995) on winter Arctic sea ice concentration and found the leading pattern that exhibits an out-of-phase fluctuation in sea ice between the eastern and western sectors of the Atlantic and Pacific basins to be most related to the index of the NAO. Upon a careful examination, however, the authors also recognize that although the time series of the NAO and sea ice index as defined by the first PC are broadly similar with linear correlation coefficient of about 0.63, the individual winters also can be radically different at times. For example, the NAO index was high in both 1989 and 1990, yet the ice index was near normal in 1989 and low in 1990. In other years when the NAO has been extreme the ice index was near normal (e.g. 1961, 1973, and 1985). This information illustrates that although important, the NAO alone is inadequate in explaining the details of atmospheric forcing related to the sea ice margin.

Alexander et al. (2004) compare simulated and observed atmospheric anomalies to examine how the variations in sea ice may feed back onto the atmospheric circulation. In their assessment the authors take into account the fact that unlike simulated anomalies, the observed circulation anomalies result from many processes other than variations in sea ice. Such factors include internal atmospheric variability and atmospheric responses to conditions in the tropics such as the ENSO. The comparisons performed in the study reveal the observed 500 hPa geopotential height anomaly to be nearly opposite to the response of the simulated model output. This finding suggests that the interactions between Arctic sea ice and the overlying atmosphere in the northern Atlantic reduce the original atmospheric circulation anomaly, a hypothesis confirmed by others (e.g. Magnusdottir et al., 2004; Deser et al., 2004) (Table 1) and consistent with the presence of a negative feedback mechanism between sea ice concentrations and large-scale atmospheric response in the Atlantic sector.

5.1.2. Pacific sector

Dethloff et al. (2006) (Table 1) argue that the feedback of the sea-ice and snow cover on atmospheric circulation is stronger in the Pacific than in the Atlantic sector, as the changes in Arctic energy sinks generate a large-scale wave train across the Pacific basin. Modeled responses to sea ice variability across the Pacific basin resemble observed data as demonstrated by Honda et al. (1999) (Table 1), suggesting that, unlike in the Atlantic, the ice–atmosphere interactions in the north Pacific sector enhance the original atmospheric circulation anomaly and thus exert a positive feedback mechanism on the large-scale atmospheric circulation. The response is enabled by the alteration of surface turbulent heat fluxes in the vicinity of the sea ice change that subsequently impact meridional temperature gradients and the relative intensity and position of the Aleutian low and Siberian high (Liu et al., 2007).

Although the NAO may be a common response pattern to Arctic sea ice fluctuations in the Atlantic sector, this pattern does not control the response to the same extent in the north Pacific. Liu et al. (2007) show that, in the north Pacific, surface air temperature signatures synonymous to that of the AO predominate only when sea ice extents are out of phase in the Sea of Okhotsk and the Bering Sea. Instead, sea ice anomalies in the north Pacific sector excite a large-scale Rossby wave train with three centers of action. One center is located over Siberia and the Sea of Okhotsk, another over Alaska and the western Arctic Ocean, and the third is situated over the eastern Pacific Ocean and western North America (Honda et al., 1996; 1999; Alexander et al., 2004; Dethloff et al., 2006) (Table 1). The pattern consists of a trough located over eastern Siberia, a ridge over Alaska, and another trough that extends from the eastern north Pacific to central North America (Alexander et al., 2004). Honda et al. (1999) suggest that this Rossby wave train that is barotropic-equivalent in nature is thermally forced by anomalous surface heat fluxes in and around the Sea of Okhotsk as the anomalous sea ice cover in this area produces anomalous diabatic heating near the surface that yields a stationary low-level response. This process induces anomalous vertical motions in the atmosphere that connect the thermal anomalies present close to the surface with the upper level wavelike response. Peng and Robinson (2001), Kushnir et al. (2002), and others also have noted that the development of the equivalent-barotropic ridge over the central north Pacific is strongly dependent upon the relationship between the location of the oceanic heating or cooling and the climatological flow. For instance, Peng and Robinson (2001) find that a warm sea surface temperature anomaly over the western Pacific induces an equivalent-barotropic high in the center of the basin when the model shows a well-defined center of action over that region. In another study, Kushnir et al. (2002) note that the extent of the eddy feedback also depends on the position of the sea surface heating relative to the storm track. The authors find that synoptic eddies provide a strong positive feedback favorable for the development of an equivalent-barotropic ridge over the central basin when sea surface heating occurs over the western north Pacific. The feedback is drastically different when the heating is shifted elsewhere such as to the eastern Pacific.

5.1.3. Tropics and subtropics

Paleoproxy data and paleoclimate simulations provide evidence that suggest a linkage between high-latitude ice conditions and the marine tropics at longer timescales (Manabe and Broccoli, 1985). Chiang et al. (2003) illustrate, for example, that most of the tropical circulation anomalies in the Last Glacial Maximum (LGM) simulations are more directly attributable to land ice influences than to direct effects of orbital or greenhouse gas changes. More recently, Chiang and Bitz (2005) (CB) (Table 1) modeled the response of the climate to an artificially imposed increase in sea ice extent in one hemisphere. The authors show that the presence of increased ice cover in either hemisphere strongly influences the Intertropical Convergence Zone (ITCZ) over the oceans by modifying the radiative balance at the top of the atmosphere and the poleward transport of atmospheric energy. The authors suggest a variant of a coupled feedback between Wind, Evaporation, and Sea surface temperature (WES feedback) of Xie (1999) as the mechanism of energy transfer between the high and low latitudes as follows. The presence of ice reduces local latent and sensible heat fluxes that lead to a cooling and drying of air over the region. The cool air is advected into the mid-latitudes through atmospheric transport and mixing cooling land and ocean surfaces. Once in the midlatitudes the cooling produces anomalous surface pressure gradients that enhance the speed of the easterlies, which in turn promote evaporative cooling of the ocean surface that push the cool sea surface further towards the equator. In the tropics, the cool sea surface waters induce anomalous meridional pressure gradients across the ITCZ latitudes resulting in a cross-equatorial flow that further displaces the convergence zone southward. This process initiates a positive feedback mechanism that maintains its southern displacement through the presence of the anomalous meridional sea surface temperature gradient. The authors attribute this feedback in part to changes in latent heat flux and reduction in the clear sky downwelling longwave flux over the tropics. The displacement of the ITCZ towards the opposite hemisphere also alters the Hadley circulation increasing subsidence of dry air from the upper atmosphere in the hemisphere with the imposed ice and uplift in the other hemisphere.

The CB study suggests that the marine ITCZ displacements are the preferred way for the tropics to respond to ice changes in the high latitudes at shorter time scales as well. The experiments find fast initial cooling of surface temperature of over 8 K throughout the northern high latitudes within the first three months from the introduction of ice over the northern latitudes. This cooling gradually migrates into the lower latitudes for two to three years thereafter, penetrating into the northern tropical Pacific only after the first year. The ITCZ is displaced into the other hemisphere around 12–15 months after the introduction of additional ice. Hughen et al. (2000), Lea et al. (2003), and most recently Liu and Alexander (2007) confirm this rapid high to low latitude connection.

The tropical atmospheric circulation and sea surface conditions in the western tropical Pacific also have been shown to be affected by changes in sea ice extent over the north Pacific through modifications in atmospheric flow in the midlatitudes that change the character of the East Asian winter monsoon and the Hadley circulation. Liu et al. (2007) note that a reduction in sea ice cover in the north Pacific enhances surface convergence in the tropics between 0° and 15°N and subsidence in the subtropics between 20°and 40°N drying the troposphere and causing a precipitation deficit over subtropical East Asia. Such conditions are indicative of an intensified local Hadley cell.

5.2. Surface air temperature

Several studies have recently examined the sensitivity of simulated global atmospheric temperature responses to a variety of prescribed sea ice conditions. Parkinson et al. (2001), for instance, report a slight decrease by −0.10 °C (increase by 0.17 °C) in global and hemispheric average temperatures upon an increase (decrease) in sea ice concentrations in the Arctic by +7% (−7%). Both hemispheres were reported to be more affected by sea ice concentration decreases than increases. When viewed on a monthly basis and regionally, the impacts of sea ice on temperature can be as large as 6 °C. An investigation of the effects of sea ice changes on the climate sensitivity to doubled atmospheric CO_2 conducted by Rind et al. (1995) (Table 1) reveals the direct sea ice snow cover impact on global annual average surface temperature response to be approximately 0.4 °C when estimated with 1D and 2D models that do not allow any feedbacks, rising fourfold to 1.56 °C when the total sea ice effects are estimated using 3D model experiments that consider various feedbacks. In another study, Hansen et al. (1997) note that the specified sea ice and SST boundary conditions from the Atmosphere Model Intercomparison Project (AMIP) (Gates, 1992) produce a global surface atmospheric warming in the GISS GCM of 0.24 °C per decade for the time period between 1979 and 1993, much greater than the observed warming of 0.1 °C per decade, tracing the excess warming to discontinuities in the sea ice boundary conditions. Finally, Singarayer et al. (2006) (Table 1) predict that global annual mean surface air temperatures may increase by up to 0.3 °C as a result of a decreasing Arctic sea ice during the 21st century.

In the high latitudes, sea ice variations are known to affect local surface air temperature regimes, but other influences such as oceanic feedback mechanisms are important in explaining the observed spatial features of air temperature anomalies away from the Arctic region. Reduction (expansion) in sea ice coverage coincides with atmospheric warming (cooling) over that surface that expands to lower latitudes through horizontal advection and eddy transport (Royer et al., 1990). This process, however, has a limited influence on temperature very far from the initial source of the sea ice anomalies. The general consensus is that temperature anomalies drop off rapidly with distance upon the removal or expansion of Polar sea ice (e.g. North, 1984; Manabe and Broccoli, 1985; Raymo et al., 1990; Parkinson et al., 2001; Chiang and Bitz, 2005), the rate of this decline being especially pronounced over the oceans (Royer et al., 1990; Parkinson et al., 2001).

The initial Arctic surface warming or cooling can be more effectively advected to lower latitudes in areas of northerly flow, as has been observed on the western sides of the Aleutian and Icelandic lows. This flow is, however, more restricted to higher latitudes on the eastern sides of the pressure systems that in turn prevent the thermal anomalies from being transported south into the midlatitudes (Royer et al., 1990). Opposite responses in surface temperature to Arctic sea ice variations have been documented over the eastern and western hemisphere continents (Newson, 1973; Murray and Simmonds, 1995; Parkinson et al., 2001).

5.2.1. Western hemisphere

Low-level circulation patterns transport the Arctic air south advecting the surface temperature anomalies into lower latitudes of the North American continent. Jäger and Kellogg (1983) compute average maps of temperature anomalies for ten warmest Arctic winters between 1931 and 1979 and find an overall warming of about 1 K during the warm years extending centrally from Alaska to Florida along with a cooling in the eastern Canadian Arctic. Royer et al. (1990) document a minor warming in the southern and central parts of North America when Arctic sea ice is removed and Murray and Simmonds (1995) find positive 850 hPa height temperature anomalies extending from the Arctic to northern Canada and the western U.S. when Arctic sea ice is reduced. Raymo et al. (1990) report a warming of up to 18 °C over northern Canada and Alaska as well as a minor rise in temperatures in the southern and central parts of North America upon the reduction of sea ice limits in the Arctic during winter. The authors attribute this change to the warming of polar air masses that migrate from the Arctic latitudes southeastward into east central Canada. Deser et al. (2000) find that when the sea ice conditions in the Labrador and Bering Seas are out of phase with sea ice conditions in Greenland/Barents Seas and the Sea of Okhotsk the hemispheric response of surface temperature resembles that which typically accompanies the NAO. Positive sea ice anomalies in the Labrador Sea and Bearing Sea along with negative anomalies in the Greenland/Barents Seas and the Sea of Okhotsk coincide with positive temperature anomalies over northern Eurasia and northwestern Canada, and negative anomalies over northeastern Canada and Greenland, with anomalies in excess of 2 °C over the continents.

5.2.2. Eastern hemisphere

The general atmospheric circulation processes that dominate over the Eastern hemisphere limit the advection of Arctic air southwards over the Eurasian continent and inhibit the propagation of anomalous temperatures into the southern latitudes. Arctic sea ice reductions generally coincide with moderate cooling in the middle latitudes of Europe and Asia, a change that has been attributed in part to a southward displacement and weakening of the midlatitude westerlies over the area due to a decrease in the equator-to-pole thermal gradient (Royer et al., 1990). Although Jäger and Kellogg (1983) note a warming across northern sections of Eurasia and the Siberian Arctic during warm Arctic winters, at the same time the authors also find a general cooling throughout Asia south of 50°N stretching from the Mediterranean through Asia into Japan. Royer et al. (1990) report moderate cooling in a wide area that stretches in a southeast direction from Scandinavia to the Ural Mountains, and towards the south over Poland and Russia down to the Carpathians, the Black Sea, and the Caspian Sea during winters when Arctic sea ice is removed. At this time both, Raymo et al. (1990) and Royer et al. (1990) note significant winter warming over the northern coast of Asia with progressively smaller temperature effects towards the continental interior. Murray and Simmonds (1995) also find a warming at the 850 hPa level that extends to the southern latitudes of northern Eurasia, but a cooling throughout central Eurasia, in northern Africa, and the Middle East when ice extent is reduced in the Arctic during January.

Liu et al. (2007) investigate the associations between Arctic sea ice variability and the climate of the Asian–north Pacific in more detail by extracting the leading modes of spatio-temporal sea ice variation across the north Pacific using an EOF analysis (Wilks, 1995). The results reveal two dominant patterns of variability in sea ice during winter that correlate with anomalies in surface temperature across the area. During the positive phase of the leading EOF that signifies reduced ice cover in the Sea of Okhotsk and expanded ice cover over the Bering Sea, the local warming observed over the Sea of Okhotsk and Japan extends westward through northeast China, and into Siberia. The large-scale air temperature modifications observed across the Asian/north Pacific sector at this time generally reflect that of the AO, results confirmed by Deser et al. (2000). The local warming over the Sea of Okhotsk and cooling over the Bering Sea are typically accompanied by an anomalous anticyclone that extends from the northern Pacific into Siberia, and a weakened East Asian jet stream and trough. The associated anomalous southeasterlies/easterlies reduce the influence of the climatological northwesterlies/westerlies, leading to the warming observed throughout northeast China and central Siberia. The positive phase of the second leading EOF coincides with reduced ice cover throughout the north Pacific. These conditions lead to strong local warming over two areas, one over the Pacific north of 40°N and the other in the subtropical central north Pacific. Widespread cooling also is observed over much of continental East Asia that extends eastward into the central north Pacific. At this time, the warming in the northern Pacific coincides with an anomalous cyclone situated over the entire north Pacific, and unseasonably strong East Asia jet stream and trough. The associated anomalous northerlies intensify the East Asian winter monsoon, leading to cold and dry conditions along the east coast of Asia. Anomalously large winter sea ice extents in Russian Seas in the area between 60.5°N and 89.5°N and between 60°E and the dateline, also can strengthen the Siberian high bringing cool temperatures into China (Wang et al., 1997). Winter sea ice conditions in the Kara and Barents Seas show some of the largest variation that produce significant fluctuations in surface heat fluxes in the Arctic. They are known to influence strongly the accumulation of cold air over the seas and the frequency and strength of cold air outbreaks in lower latitudes of Europe and Eurasia (Wu et al., 1999).

Northern China experiences frequent dust storms and the state of the atmospheric circulation in the Northern hemisphere plays an important role in their formation (Peng et al., 2004; Zhang et al., 2006). As Arctic sea ice concentration increases, the temperature difference between the equator and the polar regions increases, enhancing the development of the longitudinal circulations that make dust storms more likely to occur. Dust storms become less frequent during years when Arctic sea ice concentrations decline. Zhang et al. (2006) also find that when the Arctic sea ice concentrations increase, the Kuroshio and north temperate zone sea surface temperatures increase while Californian sea surface temperatures decrease, weakening the Kuroshio current and increasing the temperature difference between the Kuroshio region and the Arctic. The resulting intensified circulation promotes frequent cold front outbreaks causing cool winters and springs along with more frequent occurrence of dust storms in northern China. The opposite effects are realized upon a decline in sea ice concentrations.

5.3. Precipitation and storm track activity

Fluctuations in Arctic sea ice exhibit a marked influence on precipitation regimes around the globe, primarily by modifying the behavior of local storm tracks and large-scale planetary wave trains. In the high latitudes where the ice edge is located in close proximity to the local storm track, the atmospheric modifications that arise from the presence of sea ice anomalies in the area also can influence low-level atmospheric baroclinicity and impact the path and intensity of storms (Deser et al., 2000; Alexander et al., 2004). These modified storm systems are subsequently carried by the atmospheric waves away from the initial location of change, thereby having the potential to alter precipitation regimes in various areas throughout the Northern hemisphere. Arctic sea ice decline during winter is generally accompanied by increases in precipitation throughout the middle latitudes of the Northern hemisphere, and as with temperature the large-scale geographic signatures closely resembling those of the negative phase of the NAO. Such conditions typically bring unseasonably wet weather into northern Europe and Alaska, and dry conditions into western U.S. and the Mediterranean region (Thompson and Wallace, 1998).

5.3.1. Western hemisphere

A general rise in precipitation throughout the middle latitudes during winters when Arctic sea ice declines or is removed has been noted across the Northern hemisphere by many. In one of the earlier studies Jäger and Kellogg (1983) examine the distribution of precipitation anomalies for the average of the 10 warmest Arctic winters and find Alaska and the southern U.S. (except Florida) to be wetter than expected. At the same time the authors also point to dry winters in an area that extends across middle North America from the Pacific coast through the Great Lakes and into New York. Royer et al. (1990) ascertain a reduction in winter precipitation over the north Atlantic due to a weakened and southwardly displaced Icelandic low but an increase in the east Pacific off the coast of California due to a weakened California high when Arctic sea ice is removed. Murray and Simmonds (1995) note a rise in precipitation in parts of the mid-latitudinal western and eastern Pacific, the western Atlantic, and in Europe upon the reduction of Arctic sea ice concentrations during January. Magnusdottir et al. (2004) note a broad area of weakly increased precipitation over the mid-latitudinal Atlantic between 35°N and 50°N in an experiment where winter sea ice in the Arctic in the Atlantic basin is forced to decline at twice the observed rate. Winter precipitation increases also have been reported by Singarayer et al. (2006) (Table 1) over regions of western Europe and off the east coast of North America along the north Atlantic storm track upon the decline of Arctic sea ice cover. The GCM experiments of the same study, however, note precipitation reductions over the western coast of the U.S. should Arctic sea ice continue to decline at its current rate to 2100. Similar reductions in precipitation over western North America also are predicted by Sewall and Sloan (2004), changes that the authors attribute to modifications in storm track position.

Changes in the behavior of Northern hemisphere storm tracks associated with Arctic sea ice variations have been examined by many as they are known to be responsible for large proportion of precipitation in the midlatitudes. The response of storm track and cyclonic activity is, however, less clear than that of surface air temperature, precipitation, and general large-scale circulation. Murray and Simmonds (1995) examine the changes in cyclone behavior associated with a reduction in Arctic sea ice. With the exception of the eastern north Pacific where a general increase in cyclone numbers occurs at this time, the study finds little impact on the number of cyclonic systems outside of areas where sea ice is removed. Storm speeds and intensities tend to decline north of 45° at most longitudes, a change attributed to modifications in the westerly flow. More recently, Singarayer et al. (2005, 2006) find the winter storm track over the Pacific basin to weaken and shift slightly south from the end of the Pacific Ocean across the North American continent when Arctic sea ice is reduced. This modification has been attributed to the waning of the meridional temperature gradient between the warmer Arctic and lower latitudes, and to the presence of a stronger trough over eastern North America and the north Pacific which promotes the formation of a stronger jet over the eastern seaboard of the continent. The studies find the storm track activity to increase and shift slightly to the south over the north Atlantic, southern parts of Greenland and Fram Strait and over Europe upon sea ice decline, a modification attributed in part to the presence of the stronger jet over the North American continent. In their study Magnusdottir et al. (2004), however, note a general decrease in storm activity over both of the central ocean basins north of 50°N. A modest increase in storm activity is only noted over the eastern half of the Atlantic basin south of about 55°N, and over the eastern Pacific off the coast of California when sea ice cover in the Arctic declines. Alexander et al. (2004) confirm these results as they find the main branch of the north Atlantic storm track to weaken as a result of a large-scale response to reduced ice cover to the east and enhanced cover to the west of Greenland.

5.3.2. Eastern hemisphere

Jäger and Kellogg (1983) find dry winters throughout most of central and southern Europe, western Russia, Japan, and East Asia and wet conditions in Scandinavia, central Russia, and India during the 10 warmest Arctic winters. Above-normal precipitation over much of China and Siberia is noted more recently by Liu et al. (2007) during winters when low sea ice concentrations are present in the Sea of Okhotsk and high sea ice concentrations dominate throughout the Bering Sea. The authors associate these anomalies with the unseasonably strong low-level southeasterlies and easterlies, a flow that tends to block off the expected northwesterly/westerly flow diverting cold and dry air from northeast China and central Siberia from the area. Instead, an anomalous anticyclone develops over an area stretching from the northern Pacific to Siberia. These conditions also typically are accompanied by a weakened East Asian jet stream near 30°–50°N and prevailing northwesterlies or westerlies over central Asia. The weakened jet coincides with a shallower trough anchored along the coast of East Asia, extending from northeast China through Japan to the Sea of Okhotsk. Warm and wet conditions dominate throughout northeast China and central Siberia. During winters when sea ice concentrations are low throughout the north Pacific Liu et al. (2007) find below-normal precipitation to be present over continental East Asia in an area that extends northeastward to the north Pacific. The East Asian trough deepens at this time; the expected northerlies strengthen, as do the East Asian jet between 25° and 40°N and the westerly flow in the north Pacific between 40° and 60°N. The lower levels of the atmosphere become dominated by anomalous cyclonic circulation over the entire north Pacific linking East Asia with western North America. Together, such conditions lead to a stronger winter Asian monsoon, cold and dry conditions along the eastern coast of Asia, and dry conditions over subtropical East Asia. The intensified East Asian winter monsoon strengthens the local Hadley cell which in turn strengthens the East Asian jet (Liu et al., 2007).

Wu et al. (1999) demonstrate variations in winter sea ice in the Barents and Kara seas to be related to the intensity of the East Asian winter monsoon through the Eurasian teleconnection pattern (Barnston and Livezey, 1987). The authors find heavy sea ice in these seas to easily excite the positive phase of the pattern with anomalously low 500 hPa heights over Siberia and positive height anomalies over East Asia. Such conditions weaken the intensity of the winter monsoon and decrease the frequency of cold air outbursts in China. Opposite effects are observed during light ice conditions in this area of the Arctic basin.

6. Challenges facing studies of Arctic–global climate connections

6.1. Observational data and studies

Understanding the interaction between polar sea ice conditions and climate requires accurate records of sea ice field conditions throughout the Arctic with its variation documented over multiple decades and seasons. Reliable records are essential in driving and evaluating the results from atmospheric and coupled General Circulation Models and conducting observational studies. Characteristics of sea ice conditions described by concentration, area, thickness, or extent are routinely used in studies that aim to assess the local and remote climate impacts of sea ice variability because of their significant influence on ocean–atmosphere exchange of heat and moisture, and surface reflectance properties.

Satellite technology has enabled Arctic sea ice conditions to be monitored continuously for three decades, since 1978. Sea ice conditions were derived from data collected from the *Nimbus-7* Scanning Multichannel Microwave Radiometer (SMMR) until 1987, and the Special Sensor Microwave Imager (SSM/I), thereafter. Various algorithms have been used to derive sea ice concentration values from the raw Passive Microwave Radiometer (PMR) data. The most widely used methods include the Bootstrap (Comiso, 1995) algorithm and that developed by the National Aeronautics and Space Administration (NASA) team (Cavalieri et al., 1991). Sea ice thickness changes are estimated using a combination of coupled ice–ocean models, satellite altimetry, and various in situ measurement technologies such as upward looking sonar and electromagnetic profiling (Belchansky et al., 2008). Between 1972 and 1994, the U.S. National Ice Center (NIC) made available operational ice charts of weekly sea ice concentrations that provided complete coverage of the Arctic (Tanis and Smolyanitsky, 2000). These charts have been recognized as one of the highest quality records from the satellite era, compiled by experts using a variety of sources including satellite data, ship, and aerial reconnaissance data (Singarayer et al., 2005). Before 1970, the records of sea ice conditions consist primarily of aircraft, ship, and coastal observations taken at scattered locations in the Arctic and irregular time intervals. Recently, Mahoney et al. (2008) introduced a Russian sea ice chart dataset of sea ice extent compiled by the Arctic and Antarctic Research Institute for the 1933–2006 period. Ship observations dating back to the 19th century are limited to regions (or parts) of the north Atlantic only (Lemke et al., 2007). Among the most widely utilized sea ice concentration and extent datasets are the Walsh and Chapman dataset (Walsh, 1978) and the recently released sea ice concentration dataset from the Hadley Center (HadISST1) that also incorporates the Walsh dataset (Rayner et al., 2003). The former includes records from ship reports, charts, and, satellites and incorporates these observations into a gridded at 1° latitude–longitude sea ice extent and concentration dataset for the Northern hemisphere dating back to 1870. The authors, however, caution users about the use of the pre-1950s data because large portions of this data were derived from climatology or were interpolated. This problem increases the difficulty of fully assessing the impact of sea ice variations on atmospheric circulation through observation analysis alone, especially when considering interdecadal and decadal scale variations in climate. The Hadley dataset replaces the global sea ice and SST (GISST1) dataset and contains monthly global complete fields of sea ice concentration on a 1° latitude–longitude grid since 1871 making it the most complete and consistent record of sea ice concentration for the globe, including the Great Lakes and the Caspian Sea (Rayner et al., 2003).

Significant differences among the various sea ice datasets have been documented posing multitude of challenges for effective comparisons of results from studies that utilize them. These distinctions arise because the continuous records are often derived from diverse sources and from different algorithms (Comiso et al., 1997; Singarayer and Bamber, 2003). In their study Singarayer et al. (2005) compare records of sea ice cover and extent from the Arctic simulated from three monthly sea ice climatologies including the NASA and Bootstrap derived PMR datasets, and the NIC records. The authors find the uncertainty in the estimation of sea ice cover to be a combination of random and systematic errors, both temporarily and spatially dependent. In general, the NIC records produce significantly larger values of ice cover than the PMR datasets. For example, the NIC ice-covered area is greater than the NASA team area by 5–10% for most of the year and expands to 23% larger during July and August. The difference between the bootstrap and NIC areas also is greatest in July and in late summer in the PMR datasets. The derived sea ice extents and areas are most similar during autumn and winter, especially in the central Arctic. The largest differences occur between the NIC and NASA datasets in the summer months and in areas of seasonal ice cover near the ice edge in the Bering and Labrador Seas and the Sea of Okhotsk where the NASA team concentrations are considerably lower than the NIC and the bootstrap data. In general, the lower concentrations stretch over a more extensive area in summer than winter. PMR data also suffer from an inability to distinguish areas of summer surface melt water from open water resulting in artificially low concentrations. Singarayer et al. (2005) argue that the fact that the largest differences between the sea ice concentration datasets occur in summer suggests that the effect of surface melt ponds on PMR ice concentration retrievals is one of the main causes of the discrepancies, an issue that is also particularly evident in the NASA derived datasets. Agreement between the PMR datasets is greatest in winter in the central Arctic, whereas differences are again greatest in the summer in regions of seasonal ice cover (Comiso et al., 1997). With respect to sea ice thickness, there is considerable disagreement among models regarding its precise estimation and its variability. This problem makes the detection of long-term trends a difficult problem compounded by the high interannual variation and sparse temporal sampling (Miller et al., 2006; Belchansky et al., 2008).

The quantification of the errors associated with each of the datasets is challenging. Gloersen et al. (1992) provide a best estimate of the overall accuracy for SMMR and SSM/I data at ±7%, and Cavalieri et al. (2002) estimate the overall NASA team dataset accuracy at approximately ±5%, increasing to around ±15% for the Arctic in summer with best accuracies being recorded in areas of thickest ice. Comiso (2002) estimates the overall accuracy of the Bootstrap dataset at ±5 to 10% outside of areas with large fractions of thin ice or melt ponds. Although the NIC dataset was compiled by experts and the charts are of high quality and accuracy, changes in utilized data sources and relative experience of diverse analysts have introduced errors or biases that are hard to quantify (Partington et al., 2003). Validation of the various sea ice datasets also has been proven difficult due to the lack of comparable spatial coverage in ground truth data but Singarayer et al. (2005) acknowledge that each dataset has its demonstrated strengths and weaknesses.

Discrepancies present in the sea ice records exert a considerable influence on the model simulations of climate variability and change. Although inconsistencies are smallest during winter, their effect on climate is the greatest at this time, especially in high latitudes. Singarayer et al. (2005) point to discrepancies of over 20% in ice-covered areas in summer to show little impact on the mean climate, while 5%–10% differences in winter have Arctic-wide consequences for surface climate conditions. This finding has been attributed to the presence of large ocean–atmosphere temperature gradients, and very low sea ice temperatures that induce large surface heat flux anomalies and result in a highly sensitive climate to sea ice during winter. As the magnitude of these gradients diminishes into the summer the climate system becomes less sensitive. Singarayer et al. (2005) find the response of the modeled climate to existing sea ice inaccuracies of comparable magnitude to the recently observed changes in Arctic climate. Parkinson et al. (2001) confirm these findings by demonstrating a significant sensitivity of simulated climate to sea ice concentration specifications even within the expected accuracy levels of ±7%. Global mean surface air temperatures decline by −0.10 °C when sea ice concentrations are increased by 7%, and rise by +0.17 °C when concentrations decline by a similar amount. Simulated values of surface air temperatures are affected by a 7% ice concentration increase throughout the globe in all months, with anomaly values in excess of 6 °C found in the high latitudes during winter. Together, these studies underscore the importance of accurate sea ice prescription in the assessment of the impact that Arctic sea ice variation can pose on local and remote climates around the world.

The specific characteristic chosen to describe sea ice conditions in a study also can significantly influence the magnitude of the resulting impact on local and remote climates. Through their experiments, Alexander et al. (2004) show that changes in wintertime ice concentrations have a more substantial impact on hemispheric circulation than do changes in ice extent alone. They reach this conclusion by comparing sea level pressure and 500 hPa winter height anomalies generated by ice extent against ice concentration anomalies. Although the pattern of response in the two experiments is similar at the surface, it is not the case in the free troposphere. The 500 hPa response is approximately 40–80% larger over the Atlantic–Asian portion of the Arctic, and the Rossby wave train that emanates from the Sea of Okhotsk is diminished over North America in the ice concentration simulations. The NAO/AO-like response to sea ice variability also is more prominent in experiments that used concentration instead of sea ice extent data. The authors speculate that the enhanced surface heat flux anomalies in the concentration simulations may preferentially excite the internal modes of atmospheric variability.

Observational studies that utilize these datasets and study the impact of polar sea ice on local and global climates face additional difficulties. The largest challenge is posed by the fact that the results of such studies are not simple responses to fixed ice conditions as is the case in model experiments, but instead highlight the atmospheric conditions which prevail during extremes in sea ice conditions. By design such studies contain climatic and oceanic influences not directly accounted for by the approach. Such factors may significantly confound the results making their interpretation difficult. For instance, wintertime atmospheric circulation over the north Pacific and the marginal seas tends to be influenced by the ENSO events and other large-scale modes of climatic variability (e.g. Pozo-Vázquez et al., 2001; Jeverjeva et al., 2003). In the real atmosphere as captured by observational data, a signal forced by sea ice anomalies may be masked not only by the dominant remote influence of the ENSO and such internally generated interannual variation as the AO/NAO but also by the atmospheric anomalies that have given rise to the sea ice anomalies themselves. In an attempt to address this problem Honda et al. (1999), for instance, seek to remove the influence of the ENSO using regression analysis. This process can become rather problematic, however, when negative feedback processes that act to dampen the original signal that may be at play, as is the case in the Atlantic basin where the interactions between Arctic sea ice and the overlying atmosphere reduce the original atmospheric circulation anomaly (e.g. Deser et al., 2000).

6.2. Modeling studies

Simulation studies and experiments that employ General Circulation Models (GCMs) of the atmosphere and the oceans allow us to determine the impact on climate that results from variations in polar sea ice in the most controlled manner. Regardless, the progress towards a better understanding of the atmospheric response to sea ice variability in the middle and high latitudes has been slow. Although there is a general agreement among modeling studies regarding the type of processes important in producing the atmospheric response to sea surface temperature or sea ice fluctuations in the polar areas (Kushnir et al., 2002), discrepancies arise when considering the magnitude of the response as that is dependent upon the specific physical parameterizations of the models. Other difficulties also can arise when attempting to extract the atmospheric signal in the middle and high latitudes. Realistic simulations of climatic conditions under various polar sea ice conditions also require the use of fully coupled ocean–atmosphere models that capture the complex two-way interactions between the atmosphere, oceans, and sea ice conditions. The method entails the simulation of simultaneous evolutions of atmospheric and oceanic conditions, a process that in turn introduces a challenge of isolating the precise contribution of the ocean and atmosphere to the sea/atmosphere interaction. In an attempt to overcome this problem, the influence of the extratropical sea surface temperature/sea ice anomalies on the atmosphere in coupled models has consequently been assessed by comparing the atmospheric variability of the coupled midlatitudes system with the uncoupled variability that develops in the presence of prescribed climatological conditions that describe the surface (Kushnir et al., 2002).

Almost 40 years ago, Namias (1969) recognized that extratropical oceanic conditions can impose considerable influence on the atmosphere. However, determining the nature and strength of the ocean's back interaction on the atmosphere has remained a challenge until today, and remains the primary reason for using GCMs in controlled experiments with a prescribed sea surface temperature forcing. For example, while the response to a sea surface temperature or sea ice anomaly can provide a significant signal at the 500 hPa level, this signal almost certainly is smaller than natural variability, making its detection difficult in realistic GCM integrations. Earlier studies tried to overcome this problem by using large amount of forcing (i.e. remove all sea ice) to enhance the signal-to-noise ratio of the model response to obtain a clearer picture of the interactions between the different physical and dynamic processes (Table 1). As well, the modest size of the modeled signal-to-noise ratio necessitates that models be run using an appropriate number of realizations or ensembles. Alexander et al. (2004), for instance, propose a minimum of 20–45 ensemble members be used to simulate the mid-tropospheric response to sea ice variations effectively during the winter when the mean magnitude of the sea ice extent response is modest at 20 m at 500 mb and 2–2.5 mb at the surface.

Model specifications also have been substantially diverse with respect factors such as model resolution, parameterization of physical processes and surface characteristics such as sea ice and sea surface temperatures, and simulation of natural variability of the ocean and atmosphere. Such distinctions make the interpretation of sea ice experiment results challenging. Table 1 highlights the major differences in the principal modeling studies discussed in this review that have examined the atmospheric response to changes in Arctic sea ice conditions outside of the polar region. It is evident that earlier modeling studies employed only uncoupled atmospheric models that used idealized ice configurations, coarse horizontal resolutions typically in the neighborhood of 5° latitude by 5° longitude, and limited number of vertical layers. Both, the horizontal and vertical resolutions were too large to specify the boundary condition anomalies and simulate the regional and global-scale climatic features and the inherent atmospheric responses, adequately. Rind et al. (1995) find the actual sea ice climate response to be severely oversimplified in models of poor vertical resolution, a conclusion confirmed more recently also by Kattsov et al. (2005) who claim that the current vertical resolution of atmospheric GCMs is still unable to simulate effectively the large-scale temperature gradients and inversions that often occur in the Arctic leading to an overestimation of sensible heat fluxes. Latest coupled ocean–atmosphere models rely on much finer resolution ranging in the horizontal from about $1.9° \times 1.9°$ to $4° \times 5°$ for the atmosphere, and from about $0.2° \times 0.3°$ to $3° \times 4°$ for the ocean. Oceanic models have undergone significant improvements over the past decade, with better representations of dynamics and thermodynamics of the ocean basins that now permit the simulation of currents, temperature, and salinity structures in three dimensions, and take into account the freezing, melting, and dynamics of sea ice and ice–ocean interactions (Kattsov et al., 2005). The vertical resolution of current models ranges from 12 layers to 56 layers in the atmosphere and from 13 layers to 47 layers in the ocean (Randall et al., 2007). Several of the recent studies assessing the atmospheric impact of Arctic sea ice using models with resolution of $2.8° \times 2.8°$ (i.e. Alexander et al., 2004; Deser et al., 2004) for hemispheric, down to $0.5° \times 0.5°$ for regional polar studies (Rinke et al., 2006; Dethloff et al., 2006) (Table 1).

Earlier modeling studies of Fletcher et al. (1973), Newson (1973) and Warshaw and Rapp (1973) (Table 1) completely removed the sea ice during their experiments in order to enhance the response signal. A more realistic prescription of sea ice and its dynamics has become more common in recent years as its proper representation significantly influences the resulting climate sensitivity, impact on atmospheric circulation, and the simulation of the global climate (e.g., Rind et al., 1995; Dethloff et al., 2006). The dynamics are important in the simulation of energy transfers between the ocean and the atmosphere through ice and snow, sea ice characteristics such as thickness, fractional cover, and the accumulation and melting of snow on the ice (Rind et al., 1995). For instance, overestimation in models of regions with thick ice probably will not affect climate sensitivity, but the same cannot be assumed in regions of thin or seasonal ice. Rind et al. (1995) demonstrate that this sensitivity increases with the introduction of more realistic sea ice thicknesses in place of approximating the ice surface by a uniform slab with inclusions of open water. A more precise ice thickness representation is known to improve simulation of thermodynamic variations in growth and melting rates that can significantly influence the ice–ocean albedo feedback processes (e.g. Randall et al., 2007). Much has been achieved with respect to stand-alone sea-ice dynamics modeling through the Sea Ice Model Intercomparison Project (SIMIP) that was initiated in the late 1990s. Its mission was to evaluate the various sea-ice rheology representations in various models, finding the viscous-plastic rheology to provide the best simulation results (Lemke et al., 1997). Regardless, some coupled GCMs still employ a simple free-drift scheme where ice is only allowed to be advected with ocean currents (Kattsov et al., 2005).

The actual sea ice climate response also is severely oversimplified in models that involve only simple dynamics and thermodynamics that allow for several important positive feedbacks but minimize the negative feedbacks (Rind et al., 1995). The understanding of sea ice feedbacks is complex due to their strong coupling to atmospheric and oceanic processes in high latitudes (Randall et al., 2007). In particular, the general circulation and ocean–atmosphere–sea ice interactions of the Arctic region continue to be only weakly represented in the coupled GCMs (Kattsov et al., 2005). Dethloff et al. (2001), for instance, point to the need to improve the vertical stratification and atmosphere–surface energy exchange processes. In their recent study, the authors examine the global atmospheric feedbacks connected to variations in sea ice and snow cover in the Arctic by forcing the atmosphere–ocean GCM with improved sea ice and snow albedo feedbacks instead of using a fixed value of about 80% for surface albedo. The improved snow albedo parameterization uses a surface temperature-dependent scheme and distinguishes between forested and non-forested areas, and the new sea ice-albedo scheme takes into account snow, pure sea ice, and melt ponds, and relies on snow cover and surface temperature. The study finds the new albedo scheme to provide a higher and more realistic albedo in winter and early spring, leading to improved simulation of Arctic surface air temperatures when compared to the climatological datasets. The improved parameterization in the global climate model in turn exerts strong influences on the global circulation pattern of the middle troposphere, affecting precipitation and storm tracks regimes across the mid-latitudes of the Northern hemisphere. Randall et al. (2007) point to other relevant feedbacks that need attention including the ice insulating feedback, the Meridional Overturning Circulation/SST-sea ice feedback, and ice thickness/ice growth feedback. Although our understanding of these mechanisms is growing their influence on climate sensitivity still needs to be quantified.

Studies that aim to explore the impact of polar sea ice on global atmospheric circulation are also typically forced with idealized and time independent sea surface temperature (SST) anomalies although the real SST field varies continuously in both space and time, as does the climatological state of the overlying atmosphere (Kushnir et al., 2002). This complexity must be taken into account if model results are to be relevant to the natural system. Failing to permit SSTs to change from one simulation to another, for instance, prevents feedback and propagation effects through the SSTs that restrict the water vapor and cloud responses and the albedo-temperature feedbacks, dulling the responses outside the polar regions (Parkinson et al., 2001). Many studies including the Atmospheric Intercomparison Project (AMIP) have attempted to resolve this issue by forcing atmospheric models with historically observed global SST variations. Doing so, however, blends the influence of extratropical sea surface conditions with the influence of the tropical SSTs that typically dominate the global SST variation, making the separation of the extratropical SST signal difficult. For example, during winter the SST forced variance can be as high as 80% of the total variance in the tropics, whereas in the mid and high latitudes it is generally only around 20%. The exception is the north Pacific where the SST forced variance can reach 60% (Parkinson et al., 2001).

A model that does not correspond to the realistic patterns of atmospheric variability cannot adequately represent the response to extratropical sea surface anomalies. Kushnir et al. (2002) argue that of particular importance is the quality of a model's simulation of atmospheric and oceanic variability as expressed by the climatological flow, transient eddy fluxes, and internal low frequency variability. GCMs generally deviate from the observed atmosphere in all three respects. It is not surprising, therefore, that a large variation exists in the simulated impacts on the atmosphere associated with sea surface characteristics in the extratropics. Peng and Whitaker (1999), for instance, find that the eddy response to an idealized heating in the north Pacific is significantly greater when the observed winter climatology is used as the basic state for models, instead of that from their GCM. On the other hand, Peng and Robinson (2001) discover the statistical association between low-level warmth and the equivalent-barotropic ridge aloft to be much stronger in observations than in the GCM. Kushnir et al. (2002) argue that as a result of such discrepancies, models may be underestimating the dynamical response to SST anomalies by as much as a factor of two. Finally, the response of sea ice anomalies is known to be highly sensitive to the location of the ice anomaly relative to certain atmospheric features such as the storm track, the mean jet, and the jet strength (e.g. Deser et al., 2000; Alexander et al., 2004; Magnusdottir et al., 2004; Dethloff et al., 2006). This finding is problematic since GCM models have significant errors in the location and strength of these features (Magnusdottir et al., 2004). The differences among the models and the experiments that study the climatic impact of Arctic sea ice variations inherently make the assessment of the details and nature of the relationships between the sea ice forcing and atmospheric response challenging.

7. Summary and conclusions

This review synthesizes published literature on the role that Arctic sea ice plays in shaping the winter global climate through atmospheric processes at interannual timescales. Sea ice has been recognized as the primary means by which the Arctic can impact the global climate because of its ability to influence the exchange of radiation, sensible heat, and momentum between the atmosphere and the ocean. As well, the evidence regarding our climate's sensitivity to changes in the Arctic sea ice conditions is mounting.

The study of Arctic's influence on the global climate dates back to the turn of the 20th century when scientists began to establish statistical correlations between sea ice conditions in the northern oceans and surface climate around Europe and Eurasia. Focus has since been placed on the study of the impact on various characteristics of the mean state of our climate system including that of the general circulation at various levels of the atmosphere, surface air temperature, precipitation rate, storm track activity, Asian monsoon systems, and the migration of the Intertropical Convergence Zone in the tropics. Greatest attention has been given in the literature and thus in this review to understanding such relationships on inter-annual time scales during the northern hemispheric cool season. Variations in Arctic sea ice conditions exert their greatest impact on the strength and path of the westerlies, the polar and subtropical jet streams, and associated cyclones. Arctic sea ice impacts the general circulation of the atmosphere in very distinct manner in the north

Pacific American and north Atlantic and Eurasian sectors. In the Atlantic sector the large-scale atmospheric response generally resembles the internal mode of atmospheric variability, the NAO/AO, the interaction between Arctic sea ice and the overlying atmosphere being governed by a negative feedback mechanism. In the Pacific sector sea ice anomalies excite large-scale Rossby wave trains with the sea ice–atmosphere interaction being governed by a positive feedback mechanism. The response of the climate in remote areas has been demonstrated to be controlled by linear dynamics with respect to the size of the forcing but nonlinear mechanisms with respect to the forcing's direction. These distinct responses help define the specific geographic distribution of resulting anomalies in surface weather conditions such as air temperature and precipitation regimes, and storm track activity.

Until the advent of GCMs the methods of inquiry were limited to observational studies that examined the statistical relationships between Arctic conditions and remote climates at various time lags. The introduction of GCMs of the atmosphere in the 1960s and 1970s and coupled ocean–atmosphere models in the 1990s, along with the use of more accurate and realistic sea ice condition prescriptions in recent years, enabled the inquiry to be performed in a much more controlled and realistic manner enhancing the potential to isolate the contribution of sea ice to global climate variability. The progress towards a comprehensive understanding of the connection between Arctic sea ice and global climate has been slow, however. Although there is a general agreement among studies regarding the type of processes important in producing the response to sea ice fluctuations, discrepancies arise when considering the magnitude of this response. Studies on the subject continue to face numerous challenges. Among these are issues regarding the quality and quantity of sea ice records. Other difficulties center on the model specifications such as horizontal and vertical resolution, and the prescriptions of sea ice dynamics, sea surface temperatures and other physical processes. Also of concern are the adequate simulation of natural variability of the ocean and the atmosphere, the isolation of the precise contribution of the ocean and atmosphere to the sea–atmosphere interaction, and the modest size of the atmospheric signal that needs to be extracted in such experiments.

Arctic sea ice is recognized as an integral player in the global climate system. Large discrepancies exist in the exact predictions of Arctic's future climate and its sea ice. As well, observational and modeling studies that aim to explore the impact of Arctic sea ice on remote climates continue to face a series of theoretical and practical difficulties. These challenges together introduce significant uncertainties about how the Arctic may potentially shape climate variability and change around the globe. Regardless, the importance of a sustained pursuit towards a better understanding of the role that polar sea ice plays in shaping global climate is underscored by the rapidly changing Arctic environment with little signs of slowing down into the future.

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