Climate Variability in the Amundsen and Bellingshausen Seas*

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ABSTRACT

Satellite data reveal a 20% decline in sea ice extent in the Amundsen and Bellingshausen Seas in the two decades following 1973. This change is negatively correlated with surface air temperatures on the west side of the Antarctic Peninsula, which have increased $-0.5^\circ C$ decade$^{-1}$ since the mid-1940s. The recession was strongest during summer, when monthly average minima in 1991–92 removed much of the incipient multiyear ice over the continental shelf. This would have lowered the regional-mean ice thickness, impacting snow ice formation, brine production, and vertical heat flux. The northern ice edge contracted by $-1^\circ$ of latitude in all seasons from 1973–79 to 1987–93, returning toward mean conditions in 1993–95. The decline included multiyear cycles of several years in length, superimposed on high interannual variability. A review of atmospheric forcing shows winds consistent with mean and extreme ice extents, and suggests links to larger-scale circulation changes in the South Pacific. Historical ocean measurements are sparse in this sector, but mixed-layer depths and upper pycnoclines beneath the sea ice resemble those in the Weddell Sea. Weaker surface currents or changes in the upwelling of Circumpolar Deep Water on the continental shelf could have contributed to the anomaly persistence.

1. Introduction

On 30 January 1774, in what we now call the Amundsen Sea (Fig. 1), heavy sea ice near 71$^\circ$S, 107$^\circ$W forced Captain James Cook to abandon his attempt to reach the fabled southern continent. In 1820, Admiral Thaddeus von Bellingshausen was able to reach far enough south to discover Peter I Island and sight what is now Alexander Island. The first expedition to the south and west of those islands, in 1898–99 on the Belgica, was beset for more than a year in sea ice (Cook 1900). Sealers had earlier wintered over on the Antarctic Peninsula (Campbell 1992), but the Belgica crew, sustained in its ordeal by the fledgling polar explorers Roald Amundsen and Fred Cook, were probably the first to winter south of the Antarctic Circle. Since that time, the sea ice of the southeastern Pacific has enjoyed a reputation of being of “much greater age and thickness than usual, [making] the continental coasts between Alexander Island and the Ross Sea the most inaccessible” (Heap 1964).

Routine satellite monitoring of sea ice in the polar regions began in 1973. In 1993 we reported that a major decrease in sea ice extent had occurred in the Bellingshausen Sea from mid-1988 through early 1991. That retreat was strongly correlated with surface air temperatures on the west coast of the Antarctic Peninsula (AP), which reached a historic high in 1989 (Morrison 1990). The sea ice decline coincided with more northerly surface winds and greater cyclonic activity, and was particularly evident during the austral summer, extending during 1992 into the adjacent Amundsen Sea. The timing of the record Bellingshausen Sea ice recession corresponded with the low-ice phase of a double wave that propagates eastward around Antarctica every 7–10 yr, with regional ice edge expansion and contraction every 3–5 yr (Murphy et al. 1995; White and Peterson 1996).

In this paper we describe a longer-term sea ice recession over the larger Amundsen and Bellingshausen Seas (Fig. 1). We note that errors in quantification of the ice cover due to gridding, moisture effects and precipitation are small compared with the interannual variability. A discussion of the spatial and temporal extent of the regional temperature anomaly is followed by an evaluation of probable atmospheric and
oceanic forcing of and response to sea ice anomalies in this region. In reviewing the regional literature, we consider implications of the observations for model projections of future sea ice cover on the Southern Ocean.

2. The Amundsen and Bellingshausen (A and B) sea ice record

a. Data

Sea ice extent and concentration in the A and B near the annual extremes are shown in Fig. 2. The February 1985 and September 1982 panels represent typical summer and winter values during the 1973–94 period. The satellite passive microwave data we used were obtained from the Nimbus-5 Electrically Scanning Microwave Radiometer (ESMR; 1973–76), the Nimbus-7 Scanning Multichannel Microwave Radiometer (SMMR; 1978–87), and the Special Sensor Microwave Imagers (SSMI; 1987–94) of the Defense Meteorological Satellite Program. To fill gaps in these records during the 1970s, we used information derived from National Ice Center (NIC) charts, compiled from visible, thermal infrared, and passive microwave observations (NOCDA 1985). Weatherly et al. (1991) found no systematic difference between NIC and ESMR ice extents, but noted a slight negative bias relative to the SMMR records of Gloersen et al. (1992). During 6 full yr of overlap (1974, 1976, and 1979–82), our calculated ESMR and SMMR ice extents averaged 7% higher than the NOCDA data.

b. The sea ice extent anomaly

The A and B sea ice displays a low seasonal range relative to other Southern Ocean sectors, with a late summer ice field exceeded only by that in the Weddell Sea (Zwally et al. 1983b; Enomoto and Ohmura 1990). Gloersen et al. (1992) noted that these ice features might be caused by below-freezing temperatures for most of the year and by diversion of the Circumpolar Current (ACC) by the AP. However, the former condition applies to all of the circumpolar coastal region, and the ACC widens southward upstream of the A and B (Gordon and Molinelli 1982; Orsi et al. 1995). The west side of the AP is about 6°C warmer than the east side (Reynolds 1981), as northward barrier winds prevail in the Weddell Sea and the westerlies have a southward component in the Bellingshausen Sea (Schwerdtfeger 1984; Jacobs and Comiso 1993). These factors limit northward ice advance west of the AP, while southward retreat in the summer has, until recently, not penetrated widely onto the continental shelf.

At the summer minimum, monthly average ice extents from passive microwave observations in the A and B have ranged from a high of 1.03 Mkm$^2$ in 1983 to a low of 0.40 Mkm$^2$ in 1992, with no minima since 1987 above the 22-yr means of 0.76 and 0.82 Mkm$^2$. In the separate Bellingshausen and Amundsen Seas, minima of 0.06 and 0.08 Mkm$^2$ in February of 1991 and 1992 were 80% below average, near the threshold achievable with sensor resolution and continental mask accuracy. Using the National Aeronautics and Space Administra-
Fig. 2. Sea ice cover in the Amundsen and Bellingshausen Seas, 130°W to the Antarctic Peninsula, with the colors depicting ice concentration. Each image represents a 1–3-day average from satellite passive microwave brightness temperatures, using the algorithm of Comiso (1995). The February panels depict conditions near the summer minimum ice extent; the September panel shows ice cover near a winter maximum. Data were obtained from the Nimbus-5 ESMR, 1973–76; the Nimbus-7 SMMR, 1978–87, and the SSMI, 1987–94 of the Defense Meteorological Satellite Program.

Fig. 3. Latitude of the northern ice edge in the Amundsen and Bellingshausen Seas (70°–130°W), averaged over septennial periods. From a midmonth compilation of National Ice Center charts at 10° longitudes provided by H. Jacka, extended for 1993 from weekly northern ice limit charts (NPOC 1993).

c. Spatial and temporal variability

Heap (1964) indicated that mean sea ice conditions from 7° to 92°W were “less significant than the frequently enormous departures from the average,” and Gloersen et al. (1992) noted a large interannual variability in A and B monthly mean ice extents. On the other hand, Fig. 3 in Parkinson (1992) and Fig. 1 in Simmonds and Jacka (1995) suggest that variability in this region is below the Southern Ocean average, and generally lower than in the adjacent Weddell and Ross Seas. More recently, Parkinson (1995) observed that “the time series for the Bellingshausen Sea shows marked multiyear fluctuations, with increased wintertime ice coverage in the 1988–91 period [when] the summertime coverage was unusually low.” The longer record over the larger A and B region (Fig. 4) shows multiyear cycles of large and small annual range (1.14 to 2.15 Mkm²). The winter peaks suggest an interannual periodicity of 3–5 yr, which is also reflected in the summer minima, as noted earlier for shorter portions of the sea ice record (Zwally et al. 1983b). To evaluate longer-

Mkm² occurred in 1992, versus a two-decade average of 1.65 Mkm² and a 1986 maximum of 1.91 Mkm².
term variability, 5-yr moving means of seasonally averaged ice extent are shown in Fig. 5. The smoothed data reveal generally negative trends during all seasons, superimposed on shorter cycles that roughly coincide with the circumpolar wave (White and Peterson 1996). Linear regressions over the full record have slopes ranging from $-0.14\, \text{Mkm}^2\, \text{decade}^{-1}$ in winter (July–September) to $-0.20\, \text{Mkm}^2\, \text{decade}^{-1}$ in summer (January–March (JFM)), significant at the 90% to 99% levels, respectively. Over the shorter SMMR/SSMI period (1979–94), the trends were significantly negative only during autumn (April–June, >90%) and summer (JFM, >99%).

d. Potential errors due to gridding, instrumentation, algorithm, and surface wetness

Constructing a long time series of sea ice cover requires a combination of data from different sensors and satellites that can provide dissimilar information. To improve consistency, we have regridded the SMMR data to the format now used for SSMI. Land masking has been updated (Martino et al. 1995), and some additional land–ocean boundaries have been masked, based upon Advanced Very High Resolution Radiometer (AVHRR) and 85-GHz passive microwave observations. Residual errors in ice edge and concentration may remain due to the different sensor frequencies, antenna patterns, side lobes, and resolution. For example, ice concentration derived from ESMR data relied upon a 19.35-GHz channel at zero incidence angle, whereas SMMR (SSMI) used 18- (19.35) GHz and 37-GHz channels at vertical polarization and 50° (53°) incidence angles. A comparison of SMMR and SSMI ice extents during the July–August 1987 overlap period shows a discrepancy of less than 1%. The algorithm we used (Comiso 1995) gave ice concentrations that were typically 5% to 15% higher than the NASA team algorithm for most of 1992 in the A and B (Figs. 7 and 8 in Comiso et al. 1996).

ESMR winter brightness temperatures of apparent multiyear ice are $-20^\circ\text{C}$ colder in the A and B than in much of the rest of the Southern Ocean (Zwally et al. 1983b). While an apparent shift of $-20\%$ in ice concentration could result from this colder signature on the one-channel ESMR sensor, the use of data from at least two channels of SMMR or SSMI substantially reduces any error due to the multiyear ice signature (Comiso et al. 1992). During early spring, when the surface is slightly wet and highly emissive, the multiyear ice signature would be masked, making even an ESMR retrieval fairly accurate. During midsummer and summer, other factors such as flooding, melted snow, or a slushy surface may cause underestimates in ice concentration. However, snow wet by flooding and wicking of the brine (Jeffries et al. 1994a; Aldworth 1995) would tend to be masked by the formation of crusts and ice lenses reported by these authors (and by Cook 1900). Warm, moist, northerly winds were frequently recorded on the
beset Belgica: “A few days ago temperature rose to +0.5°C and has remained near −1°C for several days . . . everything is wet” (10 May log entry in Cook 1900). Air and snow surface temperatures above freezing were also encountered in the central A and B during August and September 1993 (Jeffries et al. 1994a). Time series and in situ ice observations are unavailable, but little surface wetness and no melt ponds were evident during a late summer cruise through the A and B (Jacobs et al. 1994). This suggests that wetness effects may be intermittent and not a significant source of error in monthly averaged data.

Inaccuracies in ice extent determination (the sum of areas with ice concentration >15%) should be minimal, due to the strong concentration gradient near the ice edge. Although high water vapor content in the atmosphere is typical near the ice edge and can affect estimates of ice extent, the use of a combination of 19-, 22-, and 37-GHz data minimizes this effect (Comiso 1995). From a comparison of microwave and AVHRR imagery at a time of particularly low ice cover (Fig. 6), the ice edge is better and more consistently defined with the SSMI than with the AVHRR data. The latter do not appear to give a larger ice extent, but more quantitative analyses of the AVHRR data can be compromised by the persistent cloud cover (Stammerjohn and Smith 1996). The fact that the observed A and B ice retreat was similar during all seasons (Fig. 2) is further evidence that surface effects do not cause large, fictitious reductions in ice extent.

e. Precipitation and basal melting

Sea ice fields on both sides of the AP are known to experience freeboard submergence and flooding at the ice–snow interface (Heap 1964; Lange et al. 1990; Eicken et al. 1994; Jeffries et al. 1994b). Moisture transport and accumulation on the continent is high in this region (Giovinetto and Bentley 1985; Bromwich et al. 1995) and may lead to thicker snow on the sea ice. There

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Fig. 6. Composite images of the western Antarctic region encompassing the Amundsen and Bellingshausen Seas on/or about 1 February 1994. From left to right, the panels show: AVHRR channel 1 (0.72 μm) data, SSMI passive microwave brightness temperatures (K) at 19.35 GHz with vertical polarization, and percent ice concentration derived from both 19.35- and 37-GHz data. The meridional lines are spaced at 45° of longitude around 90°W, and the zonal lines are at 5° of latitude, with most A and B ice between 70 and 75°S.
f. Open water and annual cycles

The SE Pacific sea ice field has a higher percentage of open water than the adjacent sectors, with annual means ranging from 26% to 33% of ice extent since 1978, versus 19% to 22% in the Weddell Sea. Greater ice divergence in the A and B could result from higher synoptic variability of sea level pressure (Bromwich et al. 1995; White and Peterson 1996), or from stronger upwelling. Divergence is limited in the Weddell Sea by the AP, which blocks westward ice drift, and by the colder temperatures, which will cause more rapid freezing in newly opened leads. The area of open water (including thin ice) within the A and B pack increases gradually from a minimum in late summer to a maximum in late summer–early winter, ranging from 19% to 28% during the winter months (Fig. 7). Open water seasonality was low and did not change markedly during the recent ice extent anomaly. Open water averaged 0.49 Mkm² from 1979 to 1993, much larger than the 0.29 Mkm² exposed by the recent ice edge retreat (Fig. 2). Persistent open water and thin ice in this region will tend to mask changes associated with the retreat of the northern ice edge.

An unusual characteristic of the Bellingshausen Sea is its symmetric annual cycle of ice growth and retreat. Roughly equal February–March minima are followed by 5 months of advance to an August–September maxima, then 5 months of retreat (Fig. 8). The more typical Southern Ocean cycle is one of slow growth and rapid retreat (Rayner and Howarth 1979). Stammerjohn and Smith (1996) identified an even shorter ice advance than...
retreat period in a smaller Bellingshausen sector, plausibly inferring a regional surplus of ice import and melting. However, the northern Amundsen Sea imports ice from the Ross Gyre and displays an asymmetric cycle. For most of the Southern Ocean, rapid spring retreat exceeds the available air–sea heat flux, a deficit that must be made up from the underlying deep water (Gordon 1981). A more gradual retreat in the Bellingshausen Sea could therefore imply less deep water influence, consistent with a relatively low-salinity surface layer (section 5).

3. Air temperatures and ice extents

a. The observed temperature anomaly

Since the report of a warm period on the northwestern side of the AP in the early 1970s (Schwerdtfeger 1976), several authors have cited the lengthening period of positive temperature anomalies in this region (e.g., Wetherly et al. 1991; Chapman and Walsh 1993). Sansom (1989) found no statistically significant trend in the 1958–88 air temperatures at Faraday Station, due to the large interannual variability, particularly during winter. However, King (1994) indicated that the more recent record-high temperatures have increased the statistical significance of the trends and that warmer winters have caused a decrease in the annual temperature range [see also Stark (1994)]. A composite record from the southwestern AP region confirms a secular climate change since the 1960s (Harangozo et al. 1995, unpublished manuscript). A longer melt season on the surface of eastern AP ice shelves (Ridley 1993) shows that warming was not limited to the Bellingshausen side of the peninsula. Raper et al. (1984) and Jones (1990) cited the strength and persistence of the AP warming relative to other Antarctic regions. Figure C5 of Folland et al. (1992) suggests that it also extended northwest from the AP to the region south of New Zealand, one of the largest regional annual surface temperature anomalies (+0.5°C) in the global ocean from 1981–90 (vs 1951–80).

The Faraday–Rothera (western AP) air temperature record begins in the mid-1940s and shows an increase of ~0.05°C yr⁻¹ from 1944 to 1991, averaging 1.1°C colder than at Orcadas Station northwest of the Weddell Sea (Fig. 9). The coherence of these time series is not high ($r^2 = 0.56$), but is consistent with the observation by Heap (1964) that similar climatic conditions prevail in the Bellingshausen and northwestern Weddell Seas. Linear regressions of the records show a slower rate of increase at Orcadas, +0.026°C yr⁻¹ during the overlap period (1945–91) and +0.018°C yr⁻¹ over the full term (1904–91), but both are significant at the 99% level. Although regional warming was apparent prior to the 1940s (Jones et al. 1993) and western AP temperatures have occasionally exceeded those at Orcadas, the lower Orcadas maxima from 1904 to 1944 suggest that the AP did not experience an episode earlier this century warmer than that in 1988–92. An ice core from the crest of the AP indicates that the last two decades have been the warmest in the last five centuries (Thompson et al. 1994).

b. Air temperature versus sea ice extent

A 1956 peak in the Faraday–Rothera temperature record (Fig. 7) coincides with a time of early ice breakout in Marguerite Bay (Heap 1964). Schwerdtfeger (1976) correlated the AP warmth in the early 1970s with more open water there, consistent with low summer ice extents in the southeastern Pacific during the ESMR period (Parkinson 1992). The Bellingshausen ice extent has been negatively correlated with western AP surface air temperatures (Weatherly et al. 1991; Chapman and Walsh 1993; Jacobs and Comiso 1993; King 1994), as is apparent from Fig. 7. From 1973 to 1993, western AP annual-mean air temperatures ranged over 6°C, while the upstream sea ice extent varied by 0.48 Mkm² (Fig. 10). Low annual-mean temperatures of ~6.0°C in 1977 and 1978 fit the relatively high ice extents for those years in the NOCDA (1985) data. From the Fig. 10 regression ($r^2 = 0.77$), a 1°C rise in surface air temperature corresponds to ~0.11 Mkm² less sea ice. The Amundsen Sea coastline lacks long instrumental temperature records, but Fig. 3 shows that the mean A and B ice edge shifted southward by ~1° latitude in the time between the first and last periods, comparable to a 1.5°latitude/°C projection for the circumpolar region (Jacka and Budd 1991).

The mean annual cycle of ice extent in the Bellings-
Fig. 10. Annual-mean ice extent in the Bellingshausen Sea (60°–100°W) versus the Faraday–Rothera air temperature mean. Sea ice data gaps in 1973 and 1975 (+) were filled as noted in the Fig. 9 caption. Years that diverge most from the linear regression, 1980 and 1986, also appear as extremes in Fig. 12.

Fig. 11. Annual cycles of monthly average ice extent in the Bellingshausen Sea and surface air temperature (note inverted scale) at Faraday–Rothera (Fig. 9 caption). The solid curves show means for the 1973–94 period, and the dotted curves show the historically warm 1989.

The Bellingshausen Sea lags air temperature by 1–2 months in all seasons (Fig. 11). Ice extent during the record-warm 1989 also lagged temperatures by about a month in summer and fall, but not during late winter and spring. While ice extent typically lags air temperature by several weeks in the Southern Ocean (Cavalieri and Parkinson 1981; Jacobs and Comiso 1989; Chapman and Walsh 1993), Weatherly et al. (1991) also found that ice leads the winter temperatures. Fletcher (1972) noted a correlation between winter iciness of the Southern Ocean and Southern Hemispheric zonal circulation intensity. Ackley and Keliher (1976) reported that ice extent influences synoptic-scale weather patterns, air temperatures, and the regional atmospheric circulation.

4. Atmospheric forcing

From changes in sea level pressure gradients between Antarctic coastal stations, Schwerdtfeger (1976) inferred that warmer air temperatures in the early 1970s were caused by increased northwesterly winds toward the AP. Westerlies in the Bellingshausen Sea typically have a southward component in the 1977–89 Australian Bureau of Meteorology records (Jacobs and Comiso 1993). Here, we estimate wind stress directions over the larger A and B region and divide the record at 68°S, roughly separating the prevailing westerlies and easterlies. Incorporating winds from the European Center for Medium-Range Weather Forecasting extends the wind record, but reveals an offset during the overlap period (Fig. 12). Wind speeds appear to increase through 1989 and then to decrease, but the scarcity of supporting instrumental data means that little significance can be attributed to these fluctuations. Seasonal variability in speed (direction) is strongest in the northern (southern) zone. For most of the year the sea ice edge lies north of 68°S (Fig. 3), where the typical wind stress will aid ice advance due to Ekman drift. The summer edge is south of 68°S, where the more southward wind stress will tend to retain ice along the coastline.

Sea ice advance in the A and B is typically monotonic through August or September, with a midwinter pause or retreat in some years, particularly near the AP (Ackley 1981; Harangozo 1994; Figs. 4 and 9). Summer and winter maxima in cyclone densities east of the Ross Sea, an area of frequent system stagnation and decay, also lie in the confluence of lows that have spiraled southeast from the Tasman Sea region (Jones and Simmons 1993). These extremes differ in season from those reported by van Loon (1971), but both support the existence of a semiannual wave that could induce a midyear retreat. While storms will alter the sea ice cover by breaking up and rafting floes and by increasing leads and surface water divergence, evidence for decadal changes in storminess is lacking. The marked “southward” (“northward”) peak in wind stress direction during 1980 (1986) in the 60°–68°S zone of Fig. 12 is consistent with convergence (divergence) and lower (higher) ice extent than might be expected from the air temperature that year in Fig. 10. However, the longest period of sustained southerly winds in the zone from 68° to 75°S (Fig. 12) extended into the period of record-low early summer ice extent in 1992. Since southerly winds will usually transport colder air and promote freezing, this suggests an oceanic role as the sea ice was advected into warmer waters.

Changes in the Southern Hemispheric atmospheric circulation over the past two decades were largest in the South Pacific (van Loon et al. 1993). These have involved the semiannual oscillation, circumpolar trough,
FIG. 12. Seasonal average wind speed and direction in the south-western Pacific from 60° to 68°S (upper panels) and from 68° to 75°S (lower panels). Dashed lines denote Australian Bureau of Meteorology (ABM) and solid lines European Center for Medium-Range Weather Forecasting (ECMWF) records. Wind direction is the surface stress direction, with the coordinate system rotated 30° clockwise to account for ice drift to the left of the wind in the Southern Hemisphere. The shading indicates where ice will be advected northward. ABM and ECMWF wind speed and direction changes are generally coherent during the 5-yr overlap period, with typically stronger ABM winds. The lower-resolution ABM grid contains only five points in the 68°–75°S region.

and polar vortex, and may be related to the concurrent rise of sea surface temperature at low latitudes (Hurrell and van Loon 1994). El Niño–Southern Oscillation (ENSO) cycles occur every several years, and have been characterized by more positive tropical Pacific air temperature anomalies since about 1976 (Folland et al. 1990). ENSO influence extends to high latitudes (Chiu 1983; Smith and Stearns 1993; Gloersen 1995) and could account for some of the low-frequency variability in Figs. 4 and 5. White and Peterson (1996) indicated that the 4–5-yr circumpolar wave in the Southern Ocean surface pressure, wind, temperature, and sea ice extent (see also Zwally et al. 1983a; Murphy et al. 1995) is probably associated with El Niño activity in the equatorial Pacific. It is not yet clear how large-scale changes in the atmospheric circulation over the South Pacific might focus a climate anomaly in the A and B – Antartic Peninsula region.

5. Oceanic response

Air–sea interactions in the Bellingshausen Sea may have led to the warmer temperatures in the western AP region (King 1994). Lying downwind from the prevailing westerlies, the peninsula would have experienced a more maritime climate as the ice edge moved southward (Fig. 3). The A and B sea ice retreat was comparable in area and latitude, but more persistent than the Weddell Polynya, which was well developed for only 3 yr in the mid-1970s (Carsey 1980). Air–sea interactions will be less intense in the warmer A and B, but winter sea–air heat flux is much larger over open water than over sea ice (Gordon 1981). In summer, the lower albedo accompanying a smaller sea ice cover will allow greater radiational heating of the surface layer. It is thus quite likely that the recent increase in the open water area in the A and B was a factor in the AP warming.

An initial sea ice anomaly would tend to be maintained by the greater thermal inertia of the ocean versus that of the atmosphere. Heat storage in the mixed layer has been correlated with longer summer seasons (Jacobs and Comiso 1989). Abnormal summer air temperatures can predispose near-surface waters to more or less ice the next fall or winter (Weatherly et al. 1991) and are a possible source of multyear anomalies (Gloersen et al. 1992). However, persistence within a limited area implies relatively weak surface currents or continual reinforcement. A mixed-layer temperature anomaly in the central A and B would drift east-northeast toward the Drake Passage (Sievers and Nowlin 1984; Webb et al. 1991). If it moved at more than 10 cm s⁻¹, a perturbation would be advected out of the A and B in less than 6 months; at half that speed, an anomaly could propagate beyond a single annual cycle within the region. Velocities near fronts in the deep Bellingshausen Sea exceed 10 cm s⁻¹ against a background of weaker currents (Pollard et al. 1995). From the wind data in Fig. 10, stronger currents could be expected northward in the ACC and a weaker flow with a more southward component toward the Antarctic continent.

Northeast drift of surface water and ice over the deep ocean west of the AP is one limb of a clockwise circulation, with southward flow on the continental shelf (Hofmann et al. 1996). Scattered islands, bays, and grounded icebergs along the irregular A and B coastline would retard and recirculate shelf currents, contributing
to persistence of the anomaly. The Belgica track revealed slow and erratic ice motion in the southern Bellingshausen Sea, with a stronger westward drift near the shelf break. Extensive open water areas on the Amundsen shelf in the summers of 1976 and 1992 lagged similar features on the Bellingshausen shelf by about 1 yr. If related to the transport of a heat storage anomaly, this would be equivalent to a westward drift of \(-4\) cm s\(^{-1}\). That direction conforms with the prevailing winds and with observed and modeled ice and ship drifts (Ackley 1981; Pollard and Thompson 1994; Read et al. 1995). Modeled eastward surface currents along this coastline (Webb et al. 1991) may have resulted from the use of a wind stress field that did not resolve the coastal easterlies.

It is not apparent that the Circumpolar Deep Water (CDW) played a strong role in the recent A and B sea ice retreat. Hofmann et al. (1996) note that heat flux from CDW west of the AP may moderate the amount of ice cover and regional climate. For sea ice extent to be reduced on a decadal timescale by that flux, more heat must reach the surface water, presumably through a more active vertical circulation. The relatively fresh surface water, lack of bottom water formation, and historically low seasonal range in ice extent have made the A and B region seem more like the Arctic than the rest of the dynamic Southern Ocean. However, its mixed-layer depths and salinities are similar to those in the Weddell Sea (Fig. 13), and its upper-pycnocline dT/dS ratios are intermediate between those of the Weddell “warm and cold” regimes. The A and B sea ice growth will thus be constrained by the high ratio of heat to salt across this gradient (Martinson 1990). Its entire pycnocline is stronger than in the Weddell Sea (greater salinity range from 50 to 300 m), limiting deep convection forced by sea ice formation. CDW entrainment during brine-driven mixed-layer deepening also implies ice growth prior to the observed decline. Greater upwelling of CDW could result from increased wind-driven divergence of the ice and surface water, as inferred from deeper lows during 1989 in the Bellingshausen Sea (Jacobs and Comiso 1993) and suggested by slight changes in modeled wind strength (Fig. 10).

Did deep water temperatures in the A and B change during the last two decades? At reoccupied stations along 67\(^\circ\)S in the Bellingshausen Sea, the CDW temperature maximum was often 0.05–0.15°C warmer in 1992 than a few decades earlier (Swift 1993). However, greater warm anomalies in the deep waters can be found in the archives, suggesting a variability related to mesoscale features or to movements of the southern ACC boundary and deep frontal zone near 67\(^\circ\)S (Orsi et al. 1995). Read et al. (1995) reported that transport and frontal structure from 85\(^\circ\) to 88\(^\circ\)W in the Bellingshausen Sea did not change substantially between 1964 and 1992. On the other hand, a recent southward shift of the ACC position in the southeastern Pacific is supported by temperature differences between the World Ocean Circulation Experiment meridional sections and archived data (Swift 1995), and consistent with a retreat of the northern ice edge. The evidence is thus mixed, and the limited historical data make it difficult to separate any possible trend from natural variability in the A and B deep water.

There are indications that a recent change may have occurred in the rate of upwelling of CDW onto the continental shelf. Deep water floods the deeper regions of the southeastern Pacific continental shelves and is only slightly cooled and freshened in the process (Jacobs et al. 1996). At depths below 200 m near Alexander Island (Fig. 1), CDW was up to several tenths of a degree warmer in 1994 than in the early 1980s data of Potter et al. (1988). Warmer CDW would enhance melt-driven upwelling near the coastline, but better time series data and models are needed to evaluate the potential impacts of such changes on mixed-layer temperatures and sea ice extent.

6. Discussion

a. Modeling implications

Coupled air–sea general circulation models (GCMs) incorporate feedback mechanisms, such as variable surface albedo, and project decreasing sea ice extents as atmospheric temperature rises. These results agree with the generally negative correlations found between ice extent and surface air temperature (Jacka and Budd 1991; Weatherly et al. 1991). Early model simulations with enhanced greenhouse gases did not include ice dynamics, leads, or salinity effects, compromising quan-
tative projections of sea ice extent and thickness (Mitchell et al. 1990). Some GCMs now incorporate sea ice dynamics (e.g., Pollard and Thompson 1994), and future models may benefit from the inclusion of variable snow cover and surface mixed layers. Zonally variable warming over the Antarctic sea ice in a model that includes the response to aerosols (Taylor and Penner 1994) is consistent with recent warming near the AP.

A few models emphasize the tendency for warming and freshening to stabilize the ice–ocean system with a deeper pycnocline and weaker ice divergence (Martinson 1990; Manabe et al. 1991). The latter project thicker sea ice on the Southern Ocean in a warmer climate as a cap of lower-salinity surface water caused by higher P – E and continental runoff strengthens the pycnocline and damps the vertical ocean circulation and heat flux. However, with a surface water residence time of ~2 yr (Gordon and Huber 1990), an improbable 5 m yr⁻¹ of freshwater would be needed to produce mixed-layer salinities as low as those in the Arctic (Fig. 13)—that is, sufficient to greatly decrease heat flux from the CDW. Manabe et al. (1991) modeled an air temperature increase of ~0.05°C yr⁻¹, similar to that observed in the A and B, and obtained a high runoff by transferring to the ocean any P – E > 20 cm yr⁻¹ on Antarctica. They noted that precipitation may be overestimated at high southern latitudes, a typical problem in current GCMs (Chen et al. 1995). A lower ice extent and a reduction of multiyear ice seems incompatible with greater ice thickness, and would imply that the observed warming and any related freshening in the A and B did not markedly increase the regional stability of the upper ocean.

b. Circumpolar ice extent

Satellite observations of sea ice extent now extend over more than two decades, revealing interannual changes and a declining sea ice cover in the A and B. While decadal and longer trends are common in regional records (e.g., Fletcher 1972), this retreat probably exceeded that at any other time this century and diminished the regional-mean sea ice thickness. In the full Southern Ocean, decreases in sea ice extent in one area are typically balanced by increases elsewhere (Ackley 1981; Zwally et al. 1983b; Parkinson 1994). Over the 1979–87 period, Gloersen and Campbell (1991) did not find a significant change in Antarctic sea ice extent, but Johannessen et al. (1995) reported a net decline of 0.13 Mkm² from 1978 to 1994. That would correspond to a warming of ~0.14°C over the full sea ice region if the air temperature–ice extent relationship in Fig. 8 were extrapolated to the circumpolar ice edge. In addition, the A and B ice edge recession of ~1° of latitude (Fig. 2), or 0.29 Mkm², implies sufficient net growth in other sectors to compensate about half of this regional decline. The retreat depicted in Fig. 4 may be reversed over the next several years, but continued warming in high southern latitudes (Jones 1990; Jacka and Budd 1991) will eventually register as a significant loss of circumpolar sea ice cover.

c. Future work

The significance of the recent southeastern Pacific sea ice retreat has been its duration and association with a lengthy regional warming trend. The forcing is not yet clear, but probably lies external to the region, associated with large-scale changes in the state and circulation of the atmosphere and ocean. The high variability of sea level pressure and increased open water in the southeastern Pacific sector probably enhanced air–sea interactions, heat and moisture fluxes, and divergence of the mixed layer. In situ ocean data are as yet inadequate to define reliable time series, but evidence of A and B surface water alterations might be found downstream in the ACC or in the Antarctic Intermediate Water. Sea ice time series should be lengthened by continued satellite monitoring, and by the use of ice core and tree ring proxies (e.g., Cook et al. 1992; Thompson et al. 1994). Understanding the cause and course of this kind of climate variability would benefit from the identification and monitoring of key indices of the subsurface ocean circulation, application of regional coupled circulation models, and more regular field measurements.

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