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Extreme Anomalous Atmospheric Circulation in the West Antarctic Peninsula Region in the Austral Spring to Summer of 2001/2, and its Profound Impact on Sea Ice and Biota.

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ABSTRACT

This paper describes extraordinary sea-ice conditions that occurred in the western Antarctic Peninsula region from September 2001 to February 2002, resulting from a strongly positive atmospheric-pressure anomaly (blocking high) in the S. Atlantic coupled with strong negative (low-pressure) anomalies in the Bellingshausen-Amundsen and SW Weddell Seas. This unprecedented atmospheric circulation pattern created a strong and persistent north/northwesterly flow of warm moist air across the Bellingshausen Sea ice zone. Using *in situ* observations and oceanographic, satellite and NCEP/NCAR Reanalysis data, we examine the profound and complex impact on regional sea ice, snowcover, oceanography and biota. On the one hand, persistent and unseasonable high air temperatures (to ~0ºC) caused significant sea-ice melt, resulting in a freshening of the ocean mixed layer and widespread ice-surface flooding (exacerbated by enhanced snowfall), and leading to major snow-ice formation and phytoplankton growth. On the other hand, the strong winds from the N/NW caused simultaneous extreme ice deformation and dynamic thickening of first-year ice (to >15m). Following an early winter maximum sea-ice extent, the resultant extreme sea-ice compaction against the Peninsula in the austral spring led to an unusually rapid seasonal ice-edge retreat, a highly-compact marginal ice zone, and an anomalously low spring and summer sea-ice extent. At the same time, it led to the persistence of a larger-than-average extent of highly-compact ice through summer. Ecological effects were both positive and negative, the latter including an impact on the growth rate of larval Antarctic krill. The unusual conditions contributed to the formation of a major phytoplankton bloom not only seaward but also poleward of the ice edge. As the initial bloom occurred within a 100% cover of sea ice, it could not be detected in SeaWiFS ocean-colour data. The anomalous persistence of heavy ice conditions combined with extraordinary snowfall to contribute to a disastrous

Adélie penguin-breeding season in 2001/2. This analysis suggests that sea-ice extent alone is an inadequate descriptor of the regional sea-ice state/conditions, from both a climatic and ecological perspective.

1. Introduction

The West Antarctic Peninsula (WAP) is under close international scrutiny as a region of high climate sensitivity. No other part of the Southern Hemisphere has experienced such a rapid warming over the past half century, of fully 0.5º C per decade (King, 1994; King and Comiso, 2003; King and Harangozo, 1998; Smith et al., 1996; Vaughan et al., 2001). Moreover, the WAP and adjoining southern Bellingshausen Sea regions are the only Antarctic sectors to have undergone a statistically-significant decrease in the extent of winter sea ice over the past few decades i.e., over the satellite passive-microwave data era (Weatherley et al., 1991; Jacobs and Comiso, 1993; Stammerjohn and Smith, 1997; Zwally et al., 2002). This trend has resulted from a decrease in the duration (but not magnitude) of winter sea ice (Smith and Stammerjohn, 2001). It has been proposed that the long-term warming trend is associated with changes in sea-ice area extent in the Bellingshausen Sea, as fluctuations in the latter appear to correlate strongly with variations in austral winter temperatures along the WAP (Jacobs and Comiso, 1997; Smith et al., 2003a). Stammerjohn and Smith (1996) provide a detailed discussion of sea ice characteristics and variability in the region. Evidence has emerged that with the recent shortening of the regional sea-ice season, there has been an apparent gradual replacement of the continental, polar system characteristic of the southern WAP with a more maritime and warmer system that is characteristic of the northern WAP (Smith et al., 2003a). Ecosystem responses sensitive to sea-ice changes are becoming increasingly apparent (Chapman et al., 2004; Fraser et al., 1992; Fraser and Trivelpiece, 1996; Ross et al., 1996; Smith et al., 1999; Smith et al., 2003b).

Evidence is also emerging of significant oscillatory behaviour in atmospheric-circulation anomaly patterns over the Southern Ocean - over a range of spatio-temporal scales and with teleconnections to lower-latitude phenomena (Carleton, 2003; Liu et al., 2002a, b; White et al., 2002; Yuan and Martinson, 2000). Major patterns include the Semi-Annual Oscillation (Enomoto and Ohmura, 1990; Meehl et al., 1998; Simmonds, 2003; Van den Broecke, 1998), El Niño/La Niña (Carleton et al., 2003; Harangozo, 2000; Kwok and Comiso, 2002a,b; Turner, 2004; Yuan, 2004), and the Antarctic Circumpolar Wave (Connolley, 2003; White and Peterson, 1996). On the decadal scale, recent change has been observed in large-scale tropospheric circulation in the form of a strengthening and contraction of the circumpolar vortex, and concomitant strengthening of the circumpolar westerlies (Gillett and Thompson, 2003; Hurrell and van Loon, 1994; Thompson and Solomon, 2002). This is associated with the so-called Southern Annular Mode (SAM), or Antarctic Oscillation (Fyfe et al., 1999; Hall and Visbeck, 2002; Karoly, 1990; Kidson, 1988; Raphael, 2003; Thompson and Wallace, 2000). The SAM is the dominant pattern of atmospheric variability in the Southern Hemisphere. According to Turner and Colwell (2004), localised factors have also played a role in the recent weather and climate of the Antarctic Peninsula region. Thompson and Solomon (2002) suggested that the behaviour of the SAM, and therefore its impact on atmospheric circulation and temperatures within and around Antarctica, is linked to seasonal and interannual variability in the stratospheric ozone hole. Other work e.g., Cai et al. (2003) and Fyfe et al. (1999), suggested a link between changes in the SAM and an enhanced greenhouse effect. Other dominant patterns of variability include the Pacific-South American mode and the Antarctic Dipole (Yuan and Martinson, 2001), which entails atmospheric anomalies of opposite sign in the southwest Atlantic and southeast Pacific Oceans. For an in-depth evaluation of modes of variability in the Southern Hemisphere atmosphere, see Carleton (2003), Simmonds (2003) and Simmonds and King (2004).

In this paper, we report upon extraordinary atmospheric and resultant unprecedented sea-ice conditions encountered during and after research cruise NBP01-5 on the R/V *Nathaniel B. Palmer* to the Bellingshausen Sea in the late winter to early spring of 2001. The study region and cruise track are shown in Fig. 1; the ice phase of the experiment lasted from September 11 to October 22 2001. Very heavy ice conditions resulted from extreme ice compaction and major dynamic thickening over an extended period, and coincided with a major decrease in regional sea-ice extent through the austral spring and summer of 2001/2. This in turn coincided with enhanced snow accumulation and, paradoxically, extensive melt. Cruise NBP01-05 was carried out as part of the US National Science Foundation's Palmer Long-Term Ecological Research (LTER) program. The overall aim of this program is to understand the structure and function of the Antarctic marine ecosystem in response to variability in oceanic, atmospheric and sea ice forcing (Smith et al., 1995, 2003a; Ross et al., 1996;

http://www.icess.ucsb.edu/lter/lter.html). Here, we also examine the major ecological implications of the extreme conditions in 2001/2.

We show that the protracted anomalous sea-ice conditions first encountered in the Bellingshausen Sea in late 2001 persisted through the austral spring and summer periods of 2001/2 and resulted from large-scale anomalies of opposite sign in the pattern of atmospheric circulation. Dominant were a persistent positive atmospheric pressure anomaly (blocking high) of unprecedented magnitude in the South Atlantic Ocean, coupled with large negative (low-pressure) anomalies in the Bellingshausen-Amundsen and SW Weddell Seas. We then assess the profound impact of the anomalous atmospheric circulation on regional i) sea ice and snowcover melt and ice microstructure, ii) upper-ocean structure, iii) sea-ice dynamics, iv) sea-ice concentration and extent, v) ice- and snow-thickness distributions. An assessment is then made of the resultant impacts on sea-ice biota and the subsequent ecological response to these unusual conditions. Finally, the regional conditions are examined in light of the circumpolar impacts of a wider wavenumber-3 pattern in atmospheric circulation, of which the WAP region is a component part. This paper complements that of Turner et al. (2002), who examined the impact of this anomalous atmospheric circulation pattern in creating unprecedented heavy ice conditions in the eastern Weddell Sea. Sea-ice conditions in the region and in the period prior to the anomaly (i.e., the austral winter of 2001) are discussed by Perovich et al. (2004).

2. Data and Techniques

Data from various sources were used in this analysis. The first comprises *in situ* data collected during the cruise. Detailed analysis was carried out on the physical properties of sea-ice core and snowcover samples. Many of these were collected within Marguerite Bay (Fig. 1), where the ship became beset for nearly two weeks. The relative contribution of the meteoric component of the ice cores (i.e., snow ice) was determined by oxygen isotope (_^{18}O) analysis of ice core samples and by crystallographic analysis of vertical thin sections under polarized light. Due to the extreme ice thicknesses encountered, the core samples are primarily from the upper few metres of the ice cover only. This cover typically comprised a number of rafted blocks with an individual thickness of ~0.5-1.0 m, and separated by water-filled interstices which were centimetres to tens of centimetres wide. Oceanographic data from an onboard CTD (conductivity-temperature-density) sampling programme were used to determine changes in the water-column characteristics related to surface melt. The experiment also included a detailed programme to measure biomass, primary production and nutrient concentration both in the ice and water column; this was carried out in tandem with the sea ice and oceanographic programmes. Meteorological conditions were analysed using i) data collected on the ship and ii) 12-hourly maps of mean sea-level pressure (MSLP) and 500 hPa geopotential height data from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis project (henceforth referred to as NNR) (Kalnay et al., 1996). Wind velocity data are computed from 10m zonal and meridional monthly mean winds from the NNR dataset. These data are in equal-area format, with a grid size of ~1.89º (meridional) by 1.875º (zonal). Monthly wind-field anomalies were computed using monthly means from the period 1980 to 2001.

In situ growth experiments were carried out with larval Antarctic krill (*Euphausia superba*), in a manner described in detail in Quetin et al. (2003) and Ross et al. (2004) and using SCUBA diving. Chlorophyll samples were also taken to estimate biomass i.e., ice algae and phytoplankton standing stocks. Chlorophyll was measured using a Turner Designs digital 10-AU-05 fluorometer, calibrated using a chlorophyll a (chl a) standard from Sigma Chemicals. Water aliquots were collected using Niskin bottles attached to a conductivity-temperature-density (CTD) rosette, filtered onto HA Millipore filters (0.45 microns pore size) under vacuum and extracted for 24 h in 10 ml of 90% acetone (Smith et al., 1981). A similar procedure was used for the ice samples. Ice cores were cut into 5, 10 or 20 cm sections, then melted in containers in the dark at room temperature and filtered immediately after the sample was completely melted and at a temperature of about 2 or 3º C. Once again, ice cores for biological sampling were only collected from the upper few metres of the ice, given the extreme over-rafted ice thickness. Water aliquots for the measurement of nutrient concentrations were sampled in a similar fashion to the chlorophylls and analyzed within 12 hours of sampling. Silicic acid (Si(OH)2-), orthophosphate (PO4-), nitrate (NO3-) and ammonium (NH4+) concentrations were measured according to the methods described in Johnson et al. (1985), using a Perstorp/Alpken segmented-flow nutrient-analysis system and Labtronics data-collection software.

Satellite data were used to extend the analysis both spatially and temporally. Daily maps of sea-ice concentration and extent (pixel size 25 x 25 km) were derived from brightness temperature data from the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave/Imager (SSM/I, 1987-present) and Nimbus 7 Scanning Multi-channel Microwave Radiometer (SMMR, 1978-1987), using the NASA Bootstrap algorithm (Comiso, 1995). Cloud patterns and sea-ice distributions were analysed using visible and thermal infrared data from the NOAA Advanced Very High Resolution Radiometer (AVHRR, ~1km resolution) and the DMSP Operational Linescan System (OLS, 0.55-2.7 km resolution). These data were collected onboard the R/V *Nathaniel B. Palmer* using a SeaSpace TeraScan system. Also, cloud patterns were analysed using Antarctic 3-hourly composite images comprising NOAA AVHRR and geostationary satellite data from the University of Wisconsin Space Science and Engineering Center (http://www.ssec.wisc.edu/data/composites.html).

Regional maps of the upper-ocean chlorophyll concentration were estimated from ocean colour satellite data from SeaWiFS and available from NASA at: http://daac.gsfc.nasa.gov/data/dataset/SEAWIFS/. The data used were Global Area Coverage (GAC) standard mapped data, derived by applying the OC4V4 algorithm (O'Reilly et al., 1998) and averaged over the months October 2001 to January 2002 inclusively using the maximum likelihood estimator mean. The pixel resolution is 9.77 km in the north-south and 4.2 km in the east-west direction in the area of interest. For further details, see Dierssen and Smith (2000) and Smith et al. (in prep.).

3. Anomalous Atmospheric Circulation in Late 2001-Early 2002

The initial establishment of a blocking-high pattern centred on ~50ºS, 35ºW in the South Atlantic in late-September 2001 is depicted in Fig. 2. A blocking high is an anticyclone that remains quasi-stationary for several days in the mid-latitudes where zonal flow usually prevails and the frequent passage of cyclones is the norm. Various objective definitions of the phenomenon have been developed for the Southern Hemisphere (see, for example, van Loon, 1956; Lejenäs, 1984; Trenberth and Mo, 1985; Sinclair, 1996; Wright, 1974). While the South Atlantic is one of the preferred locations for blocking-high activity (Sinclair, 1996), the late-2001 pattern was unprecedented in both its magnitude and persistence. Indeed, it lasted from September 2001 to February 2002 to dominate the regional meteorology, as shown in the monthly-mean composite anomaly map of Southern Hemisphere 500-hPa geopotential height over the period 1980-2001 (Fig. 3). Standout features in the hemispheric map are: i) an amplified wavenumber-3 pattern in the Southern Ocean coupled with ii) deep low-pressure (negative) anomalies of –150m in the Bellingshausen-Amundsen Seas and –135 m in the SW Weddell Sea. The largest high-pressure (positive) anomaly (of >150m) was that located in the South Atlantic. The unprecedented nature of this anomalous pattern is illustrated by the fact that the high-pressure recorded at South Georgia (~54.5ºS, 37ºW) was fully four standard deviations above the 1971-2000 mean, and represented a one-in-a-century event according to Turner et al. (2002). As stated by the latter, these negative anomalies were substantially different from the typical "Antarctic Dipole" regional pattern of atmospheric circulation (Yuan and Martinson, 2001).

Part of the anomalous pattern shown in Fig. 3 and described above bears a resemblance to the Southern Annular Mode (SAM), as represented by the first mode of variability in principal component analysis of the hemispheric atmospheric-pressure field (Simmonds and King, 2004; Thompson and Wallace, 2000; Thompson and Solomon, 2002; Hall and Visbeck, 2002). This dominant mode of variability is characterised by a wavenumber 3 pattern i.e., zonal asymmetry (please see Kidson [1999] and Raphael [2003] for further information on this type of pattern). In recent decades, the SAM has assumed an increasingly positive polarity (Simmonds and Keay, 2000; Thompson et al., 2000), leading to a strengthening of the westerly wind field (Liu et al., 2004). What is different about conditions in 2001/2, and described here, is the addition of deep negative pressure anomalies to the south of the South-Atlantic blocking high i.e., centred on the Amundsen/Bellingshausen Seas and the western Weddell Sea. The net effect of the anomalous configuration in Fig. 3 was a significant increase in the surface pressure gradient between South Georgia and the Antarctic Peninsula, setting up a strong predominantly northwesterly airflow across the WAP. This pattern of airflow was unprecedented in its persistence and strength, and over both an extended area and period of time.

Under more "normal" circumstances, the WAP region experiences prevailing westerly winds, with high synoptic-scale variability (King and Turner, 1997; Smith et al., 1996). An analysis of the mean-monthly climatological surface wind field derived from NNR data for August through February over the period 1980 to 2001 (not shown) indicates that the near-shore region of the WAP, and including Marguerite

Bay, exhibits something of a northwesterly component with speeds of <10 m s^{-1}. This changes to more of a westerly component offshore. An extraordinary departure from the climatological mean occurred from September 2001 through February 2002, however, as shown in Fig. 4 (an anomaly map of monthly mean wind speed and direction based on data from 1980 to 2001 inclusive). The near-surface winds over this period are remarkable in their consistency in both direction i.e., from the north-northwest, and in their strength (>10 m s^{-1}), and also over a broad band extending across the Bellingshausen Sea and Drake Passage.

The over-riding impact of the anomalous atmospheric pattern on the regional meteorology is further illustrated in the time series of measurements collected at the ship, in the Marguerite Bay region, for the period September 12-October 24 i.e., the onset and early part of the anomalous phase (Fig. 5). Strong winds (in excess of 20 m s^{-1}) with a predominantly meridional component persisted over a prolonged period, resulting in the transport of relatively warm moist air of mid-low latitude origin across the Bellingshausen Sea ice zone. Analysis of cloud patterns in satellite image time series (not shown) indicate that this air originated from as far north as ~30ºS in the South Pacific Ocean. This warm moist airflow from the N/NW resulted in i) persistent high surface air temperatures (of –1 to +1ºC), ii) ice convergence, over rafting and compaction against the Peninsula, and iii) frequent snowfalls under blizzard conditions, leading to significant snow accumulation and redistribution on the rough ice surface.

The anomalous nature of the blocking-high pattern in the South Atlantic in 2001 is confirmed by the application of an objective atmospheric blocking index. In this case, we have employed a blocking index (BI) developed by the Australian Bureau of Meteorology (Pook and Gibson, 1999) as defined as:

$$BI = 0.5(u_{25} + u_{30} + u_{55} + u_{60} - u_{40} - u_{50} - 2u_{45}) \qquad (1)$$

where u_x is the zonal (westerly) component of the wind speed (in m s^{-1}) at the 500 hPa atmospheric level at latitude x. Positive (negative) values of the index result when the westerly or zonal flow is weak (strong) at mid-latitudes and well (poorly) developed at low and high latitudes. In other words, positive anomalies represent a tendency towards blocking. We have computed the BI from NNR data to obtain an anomaly along the 30˚W meridian in each October relative to the mean October value for the period 1970 to 2001. As NNR analyses for the Southern Hemisphere have been found to contain systematic biases prior to 1970, particularly at high latitudes (Hines et al., 2000), we have confined our analysis to this 32 year period.

The results presented in Fig. 6a show that the blocking index at 30˚W in October 2001 was, at more than 1.5 standard deviations from the mean, the highest positive anomaly since 1976. Moreover, it followed on from 11 years of no substantial anomalous activity in October. Similarly, the blocking-index anomaly (BIA) time series for the austral spring (September-November) from 1970 to 2002, and shown in Fig. 6b, demonstrates that the positive 2001 BI anomaly was also the highest since 1976 (note that high indices also occurred in 1987 and 1989). Also shown, in Fig. 6c, is the BIA time series for the austral summer (December-February)-) from 1970 to 2002. In the case of the summer plot, the "year" of the particular summer refers to the year in which the austral summer period ended. Once again, a strong blocking-index anomaly is apparent for the austral summer of 2001/2 (marked A), and is more than two standard deviations above the mean. Note that the summer ending in February 2000 also exhibits a significant blocking anomaly. Only 1985/6 exhibits an anomaly that is in any way comparable to the 2001/2 and 1999/2000 events.

The anomalous nature of the 2001/2 event is further confirmed through analysis of NNR-derived monthly-mean meteorological variables at a height of 850 hPa from 65ºS, 70ºW (in the vicinity of Marguerite Bay) from 1990 to 2002 (Fig. 7). Other years, such as 1993/4 and 1999/2000, exhibit unusual constancy in wind direction from the north/northwest. The 2001/2 event alone is, however, notable not only for the persistent airflow over the WAP with a predominantly northerly component but also for the relative constancy of strong winds and the advection of low-latitude air masses with high moisture content meridionally across the sea-ice zone. Specifically, Fig. 7e depicts monthly mean precipitation rates. It shows that the peak in late 2001 was unusual in the 12-year record, with precipitation rates exceeding 5.5 mm day^{-1} (water equivalent) over a prolonged period.

4. Profound Impact on the Regional Sea-Ice Cover

The profound impact on sea-ice conditions in the WAP region was both complex and paradoxical, and was felt over a wide range of scales – from micro- to regional. In this section, we examine the net effect on sea-ice and snowcover properties, extent and thickness, and on the underlying ocean structure.

4.1. Widespread Sea Ice and Snowcover Melt

The incursion of warm air over a prolonged period caused widespread unseasonable melt of both the ice and its snowcover. The ice cover was typically warm and isothermal, with vertical temperature profiles on the order of -1.6ºC to -1.8ºC during prolonged warm spells. Analysis of core samples revealed the considerable impact on sea-ice microstructure and properties. For example, cores collected on September 26, about 3 days after the onset of the atmosphere anomaly, were extracted from upper rafted blocks of first year sea ice (see Section 4.3 for more details on rafting), were 0.38 to 0.46 m thick with a thick (~0.50 m) snowcover, and had an isothermal vertical temperature profile of –1.7 to –1.6 ºC . (See online Supplementary Material for a photo of such a representative core.) The air temperature at this time was -1.4ºC, with strong (20 m s^{-1}) northerly winds. The upper 30% of these ice cores were comprised of opaque and semi-consolidated slushy horizons, indicating incorporation of snow into the sea-ice matrix (as confirmed by subsequent thin section and $\delta^{18}O$ analysis). Separated by a horizon of slushy melt voids, the ice below this was harder, but characterised by significant melt in the form of enlarged brine channels, in this case ~1-2 cm in diameter, rendering the ice highly permeable. Golden et al. (1998) show that sea ice with a salinity of 5 psu becomes permeable once its temperature exceeds ~-5ºC. Both of these thresholds were strongly exceeded for long periods. This has important implications in that the enlarged brine channels coalesce to allow the downward transport of meltwater and/or the upward incursion of seawater (and macro-nutrients and algae) onto the snow-ice interface.

Similarly, snowcover properties were dominated by melt characteristics, e.g., crystal enlargement and rounding and the creation of icy melt layers, in upper horizons (Colbeck, 1982), and by saturation of the lower layers by flooding to form slush. Intermediate layers were themselves dampened and affected by the upward capillary “wicking” of brine from the slush horizon, in the manner described by Massom et al. (2001). This led to dampness, salinities of >10 psu, and significant snow metamorphism. Even in the absence of flooding, the lower snow layers were typically damp and saline. Snow salination in this manner affects the melt characteristics of the surface in response to high air temperatures i.e., a saline snow melts at a lower temperature than a fresh snow layer. It also impacts the optical and microwave properties of the ice cover.

In spite of the prevailing high temperatures, the base of the individual rafted blocks comprising the floes was typically characterized by apparent sea-ice growth, in the form of a layer of dendritic crystals 1.5-2.0 cm in length and protruding vertically downwards. Why basal ice growth should occur contemporaneously with significant melt above is unknown. It may relate, however, to the existence of a sheltered “micro-climate” of lower salinity water lenses at the base of floes and in the ~10 cm-wide seawater-filled interstitial gaps between the rafted-floe layers. These lenses, termed “under-ice melt ponds” in the Arctic by Hanson (1965), result from the release of meltwater percolating downwards through the overlying ice. This lower-salinity meltwater would refreeze on contact with the colder ocean (at a temperature of approximately –1.8º C). Due to the lower salinity of the “under-ice melt pools”, the temperature of the underlying water is below the melting point of the dendrites. This basal growth under melt conditions may be the equivalent of so-called “false bottoms” observed in the Arctic sea ice in summer (Eicken et al., 2002; Hanson, 1965; Notz et al., 2003). That basal growth occurred at a time of melt is borne out by oxygen-isotope analysis of ice cores cut into 10-cm samples. This gives information on the vertical distribution of ice with a component of meteoric origin. Compared to seawater, Antarctic snow is relatively depleted in the heavy stable isotope, ^{18}O, and therefore has a significant negative $\delta^{18}O$ value (Eicken, 1998; Jeffries et al., 1998). The $\delta^{18}O$ values measured from the basal strata ranged from ~-0.2 to –18 ppm.

4.2 Significant Freshening of the Ocean Mixed Layer

The persistent melt described above led to noticeable freshening of the upper ocean in the early austral spring of 2001. Fig. 8 presents CTD data from 2 downcasts at roughly the same location (~3 km apart due to ship drift) in Marguerite Bay (at ~68ºS, 70ºW) but four days apart (October 9 and 13, 2001, when the ship was “beset” by heavy ice conditions). The temperature and salinity (T/S) profiles in Fig. 8a clearly show a freshening of the ocean surface layer due to the widespread melt driven by the anomalous atmospheric circulation pattern. This occurred at a time when no significant change was observed in the upper ocean temperature profile, i.e., the mixed layer remained at freezing point (–1.8ºC). Fig. 8b is a close-up of the salinity profiles across the approximate depth of the mixed layer (dashed line depicting our "mixed layer" depth), with the salinity difference shaded. Fig. 8c is a plot of potential temperature versus salinity illustrating the temporal change in salinity in the ocean mixed layer while the deeper waters remained virtually the same.

These data were used to calculate the magnitude of ice melt responsible for the observed ocean freshening, using the relationship:

$$h_i = \_[(S1_j - S2_j)/(S2_j - S_i)/WMLD] \qquad (2)$$

where h_i is the ice thickness which has to be melted, S_i is the salinity of the ice to be melted (5 psu, based on analysis of the ice cores), $S1_j$ is the salinity of the first ocean profile at depth interval j, $S2_j$ is the salinity of the second profile at depth interval j, and WMLD is the winter mixed-layer depth (-125 dB, corresponding to a depth of 123.7 m for the October 9 cast). As such, _ $(S2_j - S_i)$/WMLD is the average salinity difference over the ocean mixed layer. In this case, h_i = 0.08 m of ice melt over the 4-day period, requiring, on average, ~70 Wm^{-2} heating.

An assumption is made that the ocean freshening was due to sea-ice melt, rather than the horizontal under-ice advection of freshwater derived from glacier melt. However, the later may be in evidence from longer profile comparisons, e.g. comparing profiles from before October 9, which showed us drifting over a front between October 7 and 9. The profiles crossing the front show dramatic intrusions at the base of the mixed layer showing considerable freshening (almost 0.40 m of ice melt if that is the source, and at this depth, we assume that it is from the melting of glaciers in the region).

4.3. Sea Ice Dynamics - Major Compaction and Anomalous Ice Extent.

The anomalous atmospheric circulation had an immense impact not only on sea-ice thermodynamic but also dynamic processes, and once again over a large area. The persistently strong north-northwesterly winds caused extreme over rafting and compaction of ice into bays and against the western Antarctic Peninsula and adjacent islands. The highly-convergent nature of the ice cover at this time is shown in the drift behaviour of three satellite-tracked ARGOS buoys deployed on ice floes in and around Marguerite Bay in August 2001 i.e. prior to the anomaly onset (Fig. 9). Please see Beardsley et al. (2004) for detailed information on buoy drift in the region. After initially heading westwards, the buoys drifted virtually due south, and covered only a relatively small distance. Dominant meridional drift of buoy 7949, for example, occurred along the 71° W meridian towards the coast of Alexander Island from late September 2001 onwards. Heavy ice conditions, extreme convergence and a lack of free drift account for relatively short distance traveled over this period – an impact also experienced by the R/V Nathaniel B. Palmer, which became beset only ~130 km from the open ocean. Under normal circumstances, the ice drift at this time of year would exhibit a much higher degree of variability, with a predominantly northwestward component. The net dynamic effect of the strong persistent north-northwesterly winds was to create a highly-compact (high concentration) and deformed ice cover with a rough surface in the Bellingshausen Sea. The unprecedented atmospheric circulation also produced unusual ice-edge conditions. The circumpolar Antarctic marginal ice zone is typically highly diffuse and composed of series of ice-edge bands rather than a distinct ice-ocean boundary, with periods of compaction being limited to periods of on-ice winds on synoptic scales only. It can be seen from the NOAA AVHRR image in Fig. 10 that the situation was quite different in late 2001, however, with an extraordinarily compact marginal ice zone and linear ice edge that persisted over a broad section of the Bellingshausen Sea. The polynya adjacent to Adelaide Island is also noteworthy, insofar as it persisted in spite of the extreme ice compaction elsewhere in the region. These dominant features of the physical landscape will be further discussed in Section 5 within the context of the ecological impacts.

The dynamic processes outlined above also resulted in a highly anomalous seasonal sea-ice areal extent in the austral spring through summer periods of 2001/2 in the NW Antarctic Peninsula sector. Compared to the long-term mean, sea-ice extent was greatly reduced in October to December 2001, as shown in the sequence of SSM/I monthly mean ice concentration and extent images in Fig. 11, with two isoline contours based on the 15% (white) and 75% (black) ice-concentration thresholds. The white dashed line in each represents the long-term mean ice-edge location (taken to be the 15% ice-concentration threshold) for that particular month over the period 1980-2001, derived from both Nimbus-7 Scanning Multi-channel Microwave Radiometer (SMMR) and SSM/I data combined. The black dashed line is the

1980-2001 mean 75% ice-concentration threshold, and is included as a proxy indicator of the extent of the consolidated pack-ice "core" region. Further analysis of the SSM/I data shows that sea-ice areal extent in the sector 60-80ºW decreased by fully ~43% (from ~0.62 to 0.36 x 10^6 km^2) over the period October 1-14, 2001, i.e., immediately after the onset of the anomalous atmospheric circulation pattern. Monthly-mean wind velocities from the NNR dataset are superimposed on the ice-concentration/extent data in Fig. 11. The two datasets combined again clearly show the profound impact of the extraordinary persistence of strong winds with a dominant north/north-westerly component over the west Antarctic Peninsula region over the period from October 2001 to February 2002 inclusive. It also is apparent that the low ice extent persisted throughout October and beyond, coinciding with an increase in ice compactness (concentration). Taken together, these results indicate that the anomalous decrease in regional sea-ice extent was predominantly driven by dynamic rather than thermodynamic processes. This is in line with similar findings from a synoptic-scale study of the region by Stammerjohn et al. (2003) and Harangozo (2004).

The anomalously low seasonal extent in late 2001 coincided with an unusually short period of sea-ice retreat for the region. Generally, the period of ice retreat is relatively long (~7 months), compared to that of ice advance (~5 months) (Stammerjohn and Smith, 1996). Moreover, maximum ice extent in the WAP region occurred in June which is atypically early compared to the average month of maximum sea ice extent for the WAP region (August), or compared to the late maximum in 2000 (September-November) (Meredith et al., 2004). The average month of maximum winter sea ice extent for Antarctica overall is September (Gloersen et al., 1992).

4.4. Immense Dynamic Sea-Ice Thickening.

The dominant ice convergence against the Peninsula and islands in response to persistent north-northwesterly winds led to immense dynamic thickening of the sea-ice cover. To stress once again, this thickening occurred at the same time as the ice was also undergoing significant and extensive melt. This created an ice cover dominated by heavily ridged and rafted fragments of first-year ice, individually 0.5-1.5m thick and frozen together to form a very thick stratum. Extraordinary ice thicknesses resulted, presumably over a large area given the magnitude of the anomalous atmospheric circulation. Indeed, SCUBA divers on the cruise, working on the krill programme, estimated ice thicknesses in mid-Marguerite Bay of 10-15 m on October 3, and even ~20 m on October 13 (Langdon Quetin, pers. comm., October 2001) (Fig. 12). For comparison, mean ice (and snow) thicknesses measured on other cruises to the approximate region in different years are significantly lower, at <1 m (Table 1). For Antarctica as a whole, the usual largely divergent nature of the sea ice results in a seasonal cover that is more typically of the order of ~0.5 to 2 m in mean thickness (Worby et al., 1998). Contemporary ship observations during NBP01-05 of ice thickness using the standard protocol of Worby and Allison (1999) seriously underestimated the actual thickness at this time. They yielded estimates of only 1-3m (e.g., on October 3), being based on observations of upper rafted blocks only, which were dislodged and overturned by the ship's hull. This was also the case with standard drill hole and ice core measurements, as these typically failed to penetrate the entire ice profile and therefore represent a sample of the upper horizons only i.e., a few rafted blocks, separated by interstitial water gaps of some tens of centimetres. As a result of these difficulties, it was impossible to measure the mean regional sea-ice thickness in late 2001.

The rafted floes making up the very deep ice were largely frozen together (albeit with interstitial gaps), in spite of the warm temperatures in the surface air and through the ice column. It is thought that "cementing together" of stacked floes in this fashion under melt conditions occurred by the rapid refreezing of meltwater percolating downwards through the porous sea-ice matrix, as described in Sections 4.1 and 4.2. This low-salinity water would refreeze at a higher temperature than seawater at a salinity of ~35 psu. Unfortunately, deep ice-core data were not collected to test this hypothesis, which is based on upper core analysis and visual observations by the dive team. This scenario implies that the ice mass would not relax and disperse (i.e. thin), with a subsequent release of lateral ice pressure.

Substantial dynamic thickening was observed to occur during individual "cataclysmic" events, when the ice periodically "gave" mechanically in response to the immense pressure exerted by almost constant convergence against the Peninsula and islands over a number of days. One such event took place on October 8. Over a matter of an hour or so, the relatively flat floe comprising an ice station was transformed to a jumbled mass of ice rubble by the severe mechanical pressure buildup within the ice cover. As a result of the extreme forces, ice was even rafted onto the back deck of the ship. This "icequake" coincided with winds of 15-20 m s^{-1} from the north (330-30º T) and air temperatures of ~0.0 to –1.0ºC i.e., typical conditions at this time.

The extreme compaction and thickness of the ice would help to explain another anomalous feature of the regional sea-ice cover in 2001/2, namely the persistence of unusually heavy sea ice conditions throughout the austral summer. As shown in Fig. 11, the overall areal extent is significantly less than the 1980-2001 mean in January and February 2002, while the compact core for the same period, and demarcated by the 75% ice-concentration contour, extends beyond the mean. The lack of open water within the pack would greatly diminish the normal seasonal ice decay process of the absorption of incoming shortwave radiation in leads and lateral ice-floe melt (Perovich et al., 2003). Moreover, the extreme ice thickness would reduce the ability of vertical melt processes to completely remove the ice cover over the annual melt season in 2001/2. Other factors likely contributing to the persistence of a heavy ice cover, and related to the atmospheric anomaly, include the enhanced snowcover thickness. The enhanced snow accumulation, both observed and modeled (see below), would also reduce the penetration/transmission of incident shortwave radiation into the underlying ice mass for heating (Eicken et al., 1995), given the high optical extinction coefficient of snow (Warren, 1982).

It should be noted that the lower concentrations apparent in the outer pack in Fig. 11 may be underestimates of the actual concentration – a common problem experienced in the retrieval of ice concentrations from satellite passive microwave data. Due to the persistence of northerly winds, the entire sea-ice zone was highly compact in late 2001, with the exception of a few coastal polynyas and most notably the one adjacent to Adelaide Island (see Fig. 10). However, the marginal ice zone was affected by strong wave-ice interaction, resulting in a predominance of brash ice grading into larger floes with distance from the ice edge together with snow removal and ice-surface wetting by wave over-washing and floe-floe buffeting. Although forming a 100% cover, such ice would have a significantly modified microwave emissivity compared to consolidated first-year ice with a dry snowcover (Comiso et al., 1989; Massom et al., 1999).

4.5. Enhanced Snowcover Thickness, Surface Flooding and Snow-Ice Formation

Another major impact resulted from the fact that the unprecedented persistent strong north-westerly airflow advected more moisture over the Bellingshausen Sea ice zone than climatologically expected. This led to anomalously high precipitation rates over an extended period, as shown in the plot of monthly mean precipitation rates derived from NNR data at 65ºS, 70ºW (Fig. 7e). High snowfalls occurred at other times (e.g. in 1994 and 2000), but at no time were they as protracted as in late 2001. Prolonged storminess and frequent snowfalls under blizzard conditions led to significant aeolian redistribution and an average snowcover thickness (Z_s) of 0.3-0.4 m over the highly deformed first-year sea ice (determined from in situ thickness transects at 0.5-1.0 m intervals). Enhanced localised snow accumulation occurred due to the rough nature of the ice surface, with significant dune and sastrugi formation adjacent to ridge sails and other surface-roughness features (Massom et al., 2003). The mean snow thickness observed in September-October 2001 was ~0.37 m (standard deviation = ~0.13 m, n = 286). This is significantly higher than previous observations in the approximate region and during the austral winter-spring e.g. Jeffries et al. (1994) reported a Z_s of 0.23 m for the sector 75-110ºW for 1993. Mean snow thickness values reported from other previous cruises to the region are given in Table 1 for comparison.

Widespread flooding of the snow/ice interface occurred as a result of enhanced snowcover loading and ice deformation. These processes acted to depress the ice surface below sea level and induce a negative ice freeboard. The consistently high temperatures played an important role by increasing the porosity/permeability of the ice, as discussed in Section 4.1. This in turn facilitated the upward incursion of seawater onto the ice surface through enlarged brine channels, while flooding also occurred laterally from floe edges. The resultant change (increase) in the incidence of ice-surface flooding as the anomalous atmospheric circulation pattern developed is shown in Fig. 13. This presents snow and ice thickness (drill hole) transects across two large and relatively flat first-year floes from roughly the same region taken about a week prior to the onset of the anomalous atmospheric circulation pattern (figure 13a), and a few days after it (Fig. 13b). This clearly shows an increase in the incidence of flooding after the anomalous pattern took hold. Table 2 presents mean values from these transects (plus an intervening transect), showing the major increase in negative freeboard and thicknesses of the slush and wicked layers in time as the atmospheric anomaly developed and took hold. Based upon 235 measurements along transects across floes, the mean freeboard overall was –6.4 cm (standard deviation = 14.7 cm). In viewing the above tables and Fig. 13b, it is important to recall that the ice thickness Z_{ice} measured in late

September-October represents only that portion of the total over-rafted ice thickness that could be sampled from the surface i.e., typically only the top few metres.

Surface flooding of Antarctic sea ice within inner pack regions is in general largely associated with the snow-loading effect, whereby the ratio of snow to ice thickness tends to be large enough in Antarctica to affect floe isostasy and depress the ice surface below sea level (Eicken et al., 1994; Jeffries et al., 1998; Lange et al., 1990; Massom et al., 2001). That flooding took place over such an extremely thick ice cover in October 2001 is due to a number of factors. Although the total ice thickness was substantial, the ice mass comprised a series of rafted first-year floes, stacked on top of each other and individually separated by interstitial seawater gaps. Key factors at this time were the extreme ice deformation and warm conditions. The former would act to depress a significant proportion of the ice surface below sea level, with seawater then flooding the surface by lateral intrusion from floe edges or by vertical intrusion through the warm and porous ice (see Section 4.1). Typical surface conditions encountered under the high-temperature regime included a thick layer of slush overlying the sea ice proper, with a large negative (up to –0.35 m) freeboard (see Supplementary Material for photographs). The high-salinity freezing slush layer can comprise more than half of the total snow column from the ice surface upwards, followed by a layer of wicked damp and saturated snow with a mean salinity of ~5 psu. This was typically capped by a layer of relatively dry and low-salinity (i.e., <1 psu) snow. The enhanced salinity and wetness of the snow likely have a significant effect on both the optical and microwave properties of the ice, with implications for the sea-ice biota, the surface energy balance and the interpretation of satellite data.

Although temporary, periodic decreases in air temperature shown in Fig. 5 were of sufficient magnitude and duration to freeze the slush, leading to substantial snow-ice formation. The resultant increase in the occurrence of snow-ice (the meteoric component) after the onset of the anomalous atmospheric circulation pattern is borne out by $\delta^{18}O$ analysis of ice core samples collected during the cruise and shown in Fig. 14. In this figure, profiles of ice $\delta^{18}O$ as a function of depth below the ice surface are split into three 10-day segments to highlight the impact of the anomaly. The first plot combines all ice-core data from September 16-26, corresponding to the period immediately prior to the anomaly onset, while the second plot spans the onset and a few days beyond and the third plot (October 8-18) illustrates data from after the establishment of the anomalous pattern. Again following Jeffries et al. (1998) and Eicken (1998), snow-ice i.e., ice with a meteoric component rather than exclusively frozen seawater, is defined as granular ice with a $\delta^{18}O$ value of <0. The peak in the incidence of snow-ice below ~1 m in the ice column corresponds to the downward movement of a floe surface by rafting, and/or the refreezing of downward-percolating snow meltwater in the form of dendrites on the base of rafted blocks (see Section 4.1). Taken overall, these findings are in line with recent observational and modelling studies e.g., Fichefet and Morales Maqueda (1999) and Wu et al. (1999), which show that snow-ice plays a major role in the seasonal growth and decay cycles of Antarctic sea ice. In this case, the impact is profound and complex.

5. Ecological Impacts

As stated above, the structure and function of all levels of the Antarctic marine ecosystem are intimately coupled to the annual advance, retreat and intervening behaviour and characteristics of the sea-ice cover (Smith et al., 1995; Ross et al., 1996). Due to its magnitude and persistence, the anomalous atmospheric circulation impacting the West Antarctic Peninsula sea-ice habitat was complex and profound, and arguably both positive and negative. The impacts of the anomalous atmospheric and sea-ice conditions on the sea-ice ecosystem continued throughout the 2001/02 season and possibly beyond.

5.1 Positive Impacts

At the lower trophic level, the widespread flooding of the snow-ice interface had a major impact on regional primary production by bringing phytoplankton and nutrients to the ice surface, where they concentrated to form a thriving "infiltration" community. These observations support the results of Arrigo et al. (1997) and Fritsen et al. (1998). The latter are that increased flooding in regions with thick snow cover (and in this case extreme ice deformation and persistently warm [porous] ice) enhances primary production in the infiltration (surface) layer, and that productivity in the freeboard (sea level) layer is also determined by sea ice porosity, which varies with temperature (Golden et al., 1998). Enhanced primary production of sea-ice phytoplankton resulted, in response to higher levels of photosynthetically-active radiation (PAR) at the ice surface compared to depth (Horner et al., 1992).

This resulted in a significant contrast between the ocean and ice, with phytoplankton biomass and primary production in the water column being consistently low compared to levels in the ice.

Algal biomass within the ice, measured as chlorophyll *a* (chl-*a*), showed a vertical distribution characterized by several layers of high concentration. These layers exhibited a high degree of discolouration to the naked eye (see Supplementary Material). Biomass concentration was 3-9 times greater in the layers compared to the background concentration. As shown in Fig.15, the layers averaged 28.3, 21.1 and 47.7 mg chl-*a* m^{-3} within a background of 4.7, 6.4 and 4.7 mg m^{-3} for the three time periods analysed. These correspond to the periods in Fig. 14 i.e., prior to, during and after the establishment of the anomalous atmospheric circulation pattern. The distinct layers in Fig. 15 were probably formed by the burial of individual infiltration communities by the extensive over-rafting of floes. This is supported by the distance between individual layers, which approximately corresponds to the thickness of individual first-year floes making up the rafted thick ice, as determined from the *in situ* drilling programme.

A marked temporal progression was observed in biomass distribution within the ice. This coincided with the observed increase in ice compaction and thickness and the change in ice conditions described above. Chlorophyll *a* (chl-a) was found throughout the upper ice column (note again that only the upper ice column was sampled), independent of ice thickness and all through the sampling period (Fig. 15). Note again that only the upper ice was sampled, due to the extreme thickness overall. The average number of layers of high chl-*a* concentration, however, decreased from September to October 2001 i.e., from 3 to 1. Secondly, the total integrated chl-*a* showed a maximum at the beginning of October. Integrated chl-*a* values for the three 10-day period depicted in Fig. 15 are 195.5 mg chl-*a* m^{-3}, 309.5 mg chl-*a* m^{-3} and 222.1 mg chl-*a* m^{-3} respectively. By comparison, average values of chl-*a* in the top few metres of the water column were: i) 16 to 26 September: 0.13 ±0.08 mg chl-*a* m^{-3} (n = 8); ii) 27 September to 7 October: 0.09 ±0.04 mg chl-*a* m^{-3} (n = 4); and iii) 8 to 18 October: 0.15 ±0.08 mg chl-*a* m^{-3} (n = 4).

Unlike chl-*a*, macro-nutrient concentration (nitrate, nitrite, silicic acid and orthophosphate) showed a less widely-varying distribution with depth (Fig. 15). The overall concentration was lower than in the water column e.g., the concentration of nitrate was 2.5 μM in the ice versus 28 μM in the water column. Some layering was also observed, but this was not as pronounced as for the chl-*a* distribution. In the case of nitrate, there is a suggestion that the nutrient layers could originate from biological activity, as local minima coincide with chl-*a* maxima.

Results presented in Table 3 show that all of the analytes covaried with time. Average nutrient concentration did not change with time, however. Vertical distributions in the ice column changed dramatically from a fairly uniform distribution in mid-September to low concentrations at the surface a month later. By mid-October, nitrate concentration became apparently limiting in the upper 0.5 m of the ice column (0.2 μM). High ice porosity and homogeneous temperature profiles in the ice column suggest that flooding of seawater at the snow-ice interface was initially the source of nutrients to the ice communities. This scenario is consistent with the homogeneous nutrient distribution found in mid-September. By mid-October, the increase in ice convergence and the extensive over-rafting of ice floes could explain the marked increase in nutrient concentration with depth, although the exact mechanism required to maintain such a distribution is not apparent from the data. Another factor could again be the enhanced surface melt, with gravitational downward percolation of meltwater through enlarged brine-drainage channels flushing nutrients down through the ice column. The increase in ice porosity with temperature could have a double-edged effect, both by allowing i) the upward percolation of seawater through the ice column to flood the surface, and ii) the downward movement of meltwater. Further work is necessary to examine these processes and their biological role on the micro-scale.

The high ice biomass observed during this cruise is believed to be related to the particular sea ice dynamics encountered in 2001. While there are few measurements of chl-a in sea ice in the WAP region for comparison, available data exhibit great variability in austral autumn/winter concentrations. Chlorophyll a concentrations averaged 0.55 mg m^{-3} during June-July 1999, with coincident water-column chlorophyll-a concentrations that were again relatively low (0.03-0.05 mg m^{-3}) throughout the sampling area during this early winter period (Ukita et al., in prep.). In May-June 2001, chl-a concentrations in newly-formed pancake ice were high, with an average of 13.8 mg m^{-3} where n = 2 (M. Vernet, unpublished data). These results suggest that the high chl-a concentrations found in September and October were due to a combination of factors, from high chl-a entrainment during formation in late austral autumn to a persistent late spring sea ice, described in this study.

5.1.1 A Major Phytoplankton Bloom Both Within and Outside the Ice Edge

The unprecedented conditions in late 2001 also contributed to major phytoplankton bloom conditions in the WAP. Enhanced ice deformation, snowfall and flooding in September-October 2001 resulted in the incorporation of a major "reservoir" of biological material into the sea ice (as described above). On traversing the marginal ice zone on October 22 (at ~64.75ºS, 64.90ºW), the ship encountered an extraordinary phytoplankton bloom poleward of the sea-ice edge, and within a 100% ice cover. Wave-ice interaction was a key physical and biological process, with a 2-3m swell from the NW leading to snowcover removal and surface wetting by wave over-washing and floe-floe buffeting, the latter process introducing additional algae onto the ice surface. It also compacted the marginal ice zone (see Fig. 10) and induced floe-floe pulverization. These processes mechanically broke down the sea ice to release the phytoplankton contained within it. This created an unconsolidated 100% cover of brash-ice fragments (up to 1-3m in diameter) separated by an interstitial dark-green slurry of frazil ice/slush and biological material combined, which was constantly re-worked by the swell. In this manner, sea-ice algae were exposed to significantly higher levels of photosynthetically active/available radiation (PAR) and macro-nutrients, compared to a consolidated ice cover with a snowcover. The net effect was the creation of an "intra-pack" bloom relatively far to the south and unusually early in the season.

The intensity of this bloom, which extended across a zone some tens of kilometres wide, is indicated by the darkness of the green discolouration. The unprecedented darkening of the ocean surface by the combined frazil-phytoplankton slurry represents a significant decrease in surface albedo, again compared to that of consolidated pack with a snowcover. This would lead to a positive feedback effect, whereby enhanced absorption of shortwave radiation as spring progressed would combine with the warm air being advected across the region to enhance melt of the slurry and phytoplankton release into the underlying water column. Stratification of the ocean mixed layer due to meltwater input over an extended period, both by ice melt and snow removal by wave overwashing, likely also contributed to the bloom conditions.

This "biological soup" resulted in a high degree of foraging activity at higher trophic levels. High concentrations of Adélie penguins (*Pygoscelis adeliae*) were observed inside the ice edge, typically in groups of 10-20, as well as Crabeater seals (*Lobodon carcinophagus*), hunting Leopard seals (*Hydrurga leptonyx*) and Killer whale (*Orcinus orca*). Concentrations of Chinstrap penguins (*Pygoscelis antarctica*), Snow Petrels (*Pagrodama nivea*, one group consisted of over 40 individuals), Antarctic Petrel (*Thalassoica antarctica*) and Antarctic Tern (*Sterna vittata*) occurred at and just seaward of the ice edge proper. Previous studies (e.g. Smith and Nelson, 1986) have reported on the occurrence of open-ocean blooms associated with the ice edge retreat, where there is an increase in water-column stability due to meltwater input coupled with the release of algae "trapped" in the ice. What is unusual about the bloom encountered in October 2001 is that it initially occurred relatively early in the season, within the sea-ice zone itself and over a zone some tens of kilometres wide. Unfortunately, no direct measurements were made of this biological "soup".

The intensity of the bloom, and its extensive coverage, is illustrated by analysis of ocean colour data derived from the SeaWiFS satellite. The data used are from the climatology of Smith et al. (in prep.), comprising seven seasons of SeaWiFS data (1997/98 through 2003/04) for an extended region west of the Antarctic Peninsula. Monthly-averaged Global Area Coverage (GAC) SeaWiFS data for this extended grid were themselves averaged to compute a chl-*a* climatology for each month for which the solar elevation and sea ice coverage permit adequate ocean colour data to be obtained i.e., September through April. The impact of cloudcover was minimised by the averaging process. Anomaly maps for each of the seven seasons, and each month, were then computed as the difference between the monthly average data and the mean climatology. The current paper uses a small subset of this dataset i.e., from October 2001 through January 2002 inclusive, to illustrate the extraordinary nature of conditions in the austral spring-summer of 2001/2.

Results are given in Fig. 16, and show that the months following the establishment of the atmospheric anomaly exhibited significant ocean-colour anomalies in the WAP region. Of immediate note is an apparent stark contrast between phytoplankton distributions in October compared to November through

January. In October 2001, an anomalously low pigment biomass occurred seaward of the compacted ice edge (see Fig. 16e). Indeed, this was the lowest October value for the seven seasons of SeaWiFS data under study. These low values are attributed to the persistence of anomalously very strong NNW winds, which continued to pack the sea ice against the Peninsula and created a deep ocean mixed layer unsuitable for bloom conditions. Note, however, that the open-ocean negative biomass anomaly occurred at the same time i.e., October 2001, as the extensive bloom within the sea-ice zone itself. An important point here is that the latter could not be detected by SeaWiFS. Average wind strength diminished in November and, with the increasing solar radiation, a strong bloom appeared over the shelf and along the shelf break throughout the region. The phytoplankton bloom continued in the WAP area through the following months and produced the highest November and December chlorophyll anomalies for the seven seasons from 1997/98 through 2003/04. It is hypothesized that the compacted and over rafted sea ice contributed melt water stability, nutrients and a copious phytoplankton seed population that gave rise to the anomalously high late-spring and early-summer bloom. Note again that this was detected in SeaWiFS data only when the ice edge receded.

5.1.2 Complex Impact on Krill

Antarctic krill (*Euphausia superba*) are a major element of the Antarctic marine ecosystem and are closely coupled to sea ice at various key stages of their life cycle (Quetin and Ross, 1991, 2001, 2003). Anomalous extreme ice conditions, such as those observed in late 2001 to early 2002, are likely to affect the under ice habitat, which larval krill use as a source of food in winter and spring when food in the water column is scarce. Although SCUBA diving in late 2001 was limited by the heavy ice conditions, an abundance of larval krill was observed during the dives within the gaps and crevices created by rafted ice blocks, and extending to greater depths than observed during previous winter cruises to the west of the Antarctic Peninsula (Quetin et al., 1996; Frazer et al., 1997).

Although the larval krill were observed feeding on the floors, sides and "caves" formed by the over-rafted pack ice, one question concerned whether larval krill were able to find sufficient food to grow since primary production would normally be low at high latitudes within a 20-m thick ice column. *In situ* growth experiments with larval krill, as described in detail in Quetin et al. (2003) and Ross et al. (2004) yielded anomalous results compared to previous cruises. During two previous cruises in the September/October time period in waters west of the Antarctic Peninsula, growth increments in larval krill were positive and relatively high, ranging from 7 to 13 % per intermolt period (Quetin et al., 2003). In September/October of 2001, however, the growth increments were negative, showing that larval krill were shrinking rather than growing (Fig. 17). The median intermolt period was 31 d. The following January, the young-of-the-year caught during the Palmer LTER annual cruise were still abundant, but were found in two modes with mean total lengths of 16.75 and 25.35 mm respectively (Quetin and Ross, 2003). The mode of smaller young-of-the-year might be derived from a late second spawning or from that part of the larval population living in the deeply over-rafted pack ice where growth rates remained negative late into the austral spring.

Fig. 17. *In situ* growth increments (% growth per intermolt period) for larval Antarctic krill collected from under the ice in austral spring of 1991 (open squares), 1993 (open triangles) and 2001 (closed circles). Larvae collected in 2001 averaged from 9.6 to 11.1 mm in total length. Points are the mean growth increment for an experiment with from 7 to 15 individual measurements. Error bars are standard error.

Previous work has shown interstitial cavities between ice-rafted blocks to play an important role in the larval-stage ecology of krill (Frazer et al., 1997; Ross et al., 2004). The likely benefits of rafting gaps to krill result from the fact that they form: i) a relatively quiescent and stable environment, enabling krill to drift with the ice with minimal energy expenditure; ii) protection from large predators; and iii) a concentrated food source relative to the underlying water column. Having said this, and in spite of the apparent downward injection of phytoplankton biomass by the rafting process (Section 5.1 and Fig. 15), it appears that insufficient food was available under the extreme ice conditions in 2001/2 to engender the growth of larval krill inhabiting the ice cavities. Clearly, more work is clearly necessary to better understand krill ecology relative to such habitat complexity.

5.2 Negative Impacts

The anomalously heavy ice conditions in the WAP region from late-winter through the austral spring to summer of 2001/2 had some adverse effects not only on krill but also on apex predators, such as Adélie penguins (*Pygoscelis adeliae*). This species lays its eggs in early- to mid-November, with the young birds fledging after 55 days and departing for the sea (Ainley and DeMaster, 1990). Of key importance is the regional behaviour of sea-ice relative to this breeding cycle (Smith et al., 1999). Work by Fraser and Trivelpiece (1996) shows that maximum population growth occurs during moderate and divergent sea-ice coverage years. Locating sufficient prey (mainly krill) within the foraging range (~100 km for the colonies around Anvers Island) e.g. during the chicks' crêche stage in mid-late January, is critical to the birds' reproductive success. Conditions in 2001/2 were such that the resident Adélie penguin population recorded a disastrous breeding season. It appears that two sets of circumstances associated with the unprecedented atmospheric circulation pattern may have contributed. On the one hand, the extreme, convergent ice conditions created a heavy ice cover that persisted through the spring-summer to severely delay the arrival of penguins at their breeding sites e.g., Torgersen Island (~64.79º S, 64.1º W) close to Palmer Station (Cohenour, 2001). Indeed, Torgersen Island remained locked in ice at a time when open water is typical. These same heavy ice conditions necessitated longer foraging trips (http://www.cybamuse.com/antarctica/seaice.htm), compounded by the lack of leads and the necessity to traverse a rough ice surface with a thick snow cover. This likely resulted in less food, and/or food less often, for chicks plus greater expenditure of adult energy (Ainley et al., 1998). In addition, the unusually heavy seasonal snowfall covered the eggs, and effectively drowned them when it melted under the accompanying warm conditions (W. Fraser, personal communication, October 2004).

The Adélie penguin population at Anvers Island (64.8ºS, 64.1ºW) has decreased over the past decade (Fraser et al., 1992; Fraser and Trivelpiece, 1996). This is apparently in response to the observed steady decrease in winter sea ice duration and the pronounced autumn and winter warming in the WAP region (Jacobs and Comiso, 1997; Smith and Stammerjohn, 2001; Vaughan et al., 2001). Wilson et al. (2001) suggest that sea-ice extent in the region has been reduced beyond the optimum for Adélie penguins, and that good conditions for them only occur now in years of significant winter sea ice extent duration. Fraser et al. (1992) further suggest that this trend relates to a reduction in suitable nesting habitats resulting from the observed warming trend. For example, enhanced snowfall such as that associated with the 2001/2 anomaly may diminish the availability of suitable snow-free nesting sites (Wilson et al., 2001). Another important unresolved question relates to possible changes in penguin prey distribution, and krill in particular (Smith et al., 1999), the abundance and availability of which is intimately linked to sea-ice conditions (as discussed earlier).

The impact of the anomalously heavy and compact ice conditions on other pack-loving seabirds, including the small, diminishing and endangered Emperor penguin (*Aptenodytes forsteri*) colony on the Dion Islands (67.87ºS, 68.70ºW) in NW Marguerite Bay, and pinnepeds is unknown, but is also likely to be significant. Birds and marine mammals rely heavily on the presence of leads/polynyas in the inner pack and on a diffuse marginal ice zone (Hunt, 1991). Virtually no open water was present within the regional pack in the spring of 2001/2, apart from the polynya adjacent to Adelaide Island (Fig. 10). Similarly, few whales were sighted within the pack, again in response to the extreme lack of open water (leads) observed.

5.3. The Bigger Picture

As shown earlier in Fig. 3, the anomalous atmospheric circulation around the Antarctic Peninsula from September 2001 through February 2002 was part of a circumpolar amplification of the mean wavenumber 3 pattern. This also had a major impact on other regions around Antarctica. For example, it caused the heaviest austral-summer sea-ice conditions in the NE Weddell Sea in 2002 for at least 50 years (Turner et al., 2002). It also resulted in unusual intrusions of mid-latitude air into the continental ice-sheet interior. The high-pressure anomaly at ~110ºE (in the South Tasman Sea), which is another preferred location for blocking-high activity (Pook and Gibson, 1999; Trenberth and Mo, 1985), was responsible for periods of slight to moderate snowfall at Dome C in East Antarctica (74.5ºS, 123.0ºE, elevation 3280 m) (Massom et al., 2004) – a rare event for this extremely dry region of the high interior Ice Sheet plateau. This episode also caused anomalously high surface wind speeds (snow redistribution) and air temperatures over the continental interior. On January 11 2002, Vostok (78.5ºS, 106.9ºE, elevation 3488 m) experienced a "balmy" temperature of –16.5ºC, close to the record high of –13.3ºC (on January 6, 1974), as reported by Turner et al. (2002). Similarly, South Pole station had a maximum temperature of –18.1ºC and a peak wind speed of ~16.5 m s^{-1}, also on January 11. In both cases, the annual mean temperature is <-50ºC. Typically, calm conditions usually prevail in all of these interior regions during this time of year. Further to the west, the high-pressure anomaly centred on ~90ºE had a significant influence on the weather in southern Australia, bringing anomalously cold and wet conditions in the austral summer of 2001/2.

6. Discussion and Conclusions

An anomalous pattern of atmospheric circulation, dominated by strong, warm and moisture-laiden winds with a predominant meridional component, affected a wide area of the West (and East) Antarctic Peninsula region over a 5-6 month period in late-2001 to early-2002. This had a profound impact on regional sea-ice and ocean conditions and thus marine ecology. This impact was highly complex, and indeed paradoxical, with counteractive processes occurring simultaneously. On the one hand, persistently high temperatures caused widespread ice and snow melt which had a major impact on ice/snow microstructure and properties and on upper-ocean structure i.e., major freshening and stratification. On the other hand, the same strong northerly winds that transported heat and moisture across the sea ice led to simultaneous and immense dynamic ice thickening by convergence against the Peninsula/islands and into bays. This compaction created extremely heavy ice conditions and an ice cover up to 20 m thick, composed of rafted first-year floes individually <1 m thick when undeformed. Due to the extreme over-rafted ice thickness, the standard ship observation and ice-drilling protocols proved to be inadequate sampling tools and yielded serious underestimates of the actual ice thickness. Regarding melt, the available evidence suggests that the observed significant freshening of the mixed layer contained contributions not only from melting sea ice and its snowcover but also from glaciers in the region i.e., both vertical and lateral processes were at play.

Another major outcome of the anomalous atmospheric circulation pattern was high snowfall under blizzard conditions, resulting in a snow cover that was significantly thicker than the climatological mean. Resultant snow loading combined with enhanced ice deformation and the warm conditions to cause widespread flooding of the snow/ice interface, resulting in enhanced snow-ice formation. This to some extent compensated for the loss of snow volume by melt. At the same time, apparent sea-ice growth occurred at the base of rafted ice blocks during melt periods, by the refreezing of downward-percolating meltwater. Work in the Arctic has shown that this "redistribution" of meltwater has important implications for the mass balance of the ice cover (Eicken et al., 2002), as well as the salt and heat balance of the ocean mixed layer (Kadko, 2000). Limited suitable data were collected on this experiment, and more work is required to examine this process in Antarctica.

By persisting throughout the austral spring through summer, the unprecedented pattern of atmospheric forcing had a major impact on sea-ice extent and its seasonal behaviour in the WAP region in the austral spring through summer of 2001/2. Under "normal" circumstances, sea ice coverage in the LTER and wider Bellingshausen Sea regions is usually distinct from other Southern Ocean regions in that the period of ice advance is relatively short (~5 months) in comparison to the period of ice retreat of ~7 months (Stammerjohn and Smith, 1996). Dynamic compaction against the Peninsula resulted in 2001, however, in a period of seasonal sea-ice retreat that was both early and rapid (short) compared to the climatological mean, resulting in an unusually low regional extent in the austral spring. As a result, maximum ice extent in the WAP region occurred atypically early in 2001 i.e., in June, compared to the more typical August. Minimum sea ice extent occurs in March in the WAP region compared to February in the Weddell and Ross Seas. Since the 1990s, however, there has been increased variability associated with the timing of sea-ice advance and retreat with underlying trends of later advance and earlier retreat that translate into a decreasing trend in the length of the winter ice cover season (Smith and Stammerjohn, 2001; Stammerjohn et al., 2003). In other words, the decreasing trend in winter sea-ice extent that has been observed in the WAP and greater Bellingshausen Sea region is not due to a decrease in the magnitude of winter sea-ice extent, but rather to a decrease in the duration of winter sea-ice extent. This has important ecological implications, and is a subtle factor that is often overlooked. Moreover, the early and rapid sea ice retreat that occurred in 2001/2 has important implications for the radiation balance at the surface, given that an open ocean absorbs significantly more incoming shortwave radiation than an ice-covered ocean. In addition, a decreased spring sea ice extent would likely lead to changes in cyclone trajectories (Menendez et al., 1999; Murray and Simmonds, 1995). Such complex feedback mechanisms require further study.

The extreme ice compaction by strong winds with a dominant meridional component also created an extraordinarily compact marginal ice zone and linear ice edge, which formed a distinct boundary between pack ice and open ocean. Moreover, heavy ice conditions persisted throughout the austral summer in Marguerite Bay – in spite of the warm conditions and in a region where seasonal meltback normally occurs. This coincided with a below-average sea-ice extent in January through March 2002, following on from the below-average extent in October through December 2001. The persistence of heavy ice through the austral summer in 2002 likely relates to three factors. These are: i) the wind-driven compaction (few leads for the absorption of incoming shortwave radiation and lateral floe melt); ii) the sheer thickness of the sea ice (to counteract total melt); and iii) the insulative effect of a thick snow cover. The latter also has a high optical extinction coefficient to diminish the penetration of shortwave radiation into the underlying ice (Eicken et al., 1995; Perovich, 1996).

Taken together, these results strongly suggest that sea-ice extent alone is an incomplete and at times over-simplistic indicator or descriptor of climate and/or ecological variability and change, in that it fails to take account of the degree of ice compaction and accompanying dynamically-driven changes in ice thickness. Had surface observations and additional information not been available, then the anomalous ice-edge retreat observed in the satellite passive microwave time series for late-2001 may have been erroneously interpreted as simply a meltback of the ice cover due to the high air temperatures, rather than a dynamically-driven compaction of the entire regional sea-ice zone (with snow-ice formation also playing a role). As we have seen, the observed decrease in sea-ice extent in fact coincided with a major increase in ice thickness – a factor that was only resolved by contemporary SCUBA-diver observations. Clearly, additional accurate information is required on the degree of compaction and the ice thickness distribution, and on a regional scale, in order to resolve temporal changes in ice volume. This is a major issue in sea ice-related climate research, given the current difficulty in accurately measuring ice thickness using standard ship observation techniques and *in situ* techniques, and in remotely sensing ice thickness from space. It remains to be seen how well emerging satellite technologies such as CryoSat (Drinkwater et al., in press) will perform in measuring sea-ice freeboard/thickness in the Southern Ocean (and particularly in a highly compact ice cover such as that encountered in 2001, where few leads are available as reference surfaces). Promising results are, however, emerging from new spaceborne laser measurements of sea ice freeboard from IceSat (Kwok et al., 2004; Jay Zwally [NASA], pers. comm., October 2004).

Anomalous atmospheric circulation in late 2001-early 2002 had a profound and complex impact on the regional sea-ice ecosystem, both positive and negative, by creating a significant decrease in sea-ice extent but a major increase in ice thickness, deformation and compaction and snowcover thickness. This resulted in a persistently very heavy ice cover characterised by virtually no leads, but a flooded ice surface. An infiltration phytoplankton community formed, creating optimal relatively high illumination conditions for phytoplankton growth. Persistent melt played a role by freshening and stratifying the ocean mixed layer. Subsequently, floe mechanical breakdown by pulverization driven by wave-ice interaction released the phytoplankton to produce a rich frazil ice-algal "soup" within the marginal ice zone itself. While previous work has concentrated on algal blooms seaward of the ice edge, this one initially occurred within a compact (albeit unconsolidated) ice cover. As it occurred in the presence of a 100% sea-ice cover, the bloom was impossible detect in satellite ocean-colour data alone. Indeed, analysis of SeaWiFS data shows that a strong negative anomaly occurred in chlorophyll concentration seaward of the ice edge in October 2001 i.e., at the same time as the extraordinary bloom within the pack that remained undetected from space. It appears that this bloom was conditioned by the earlier incorporation of phytoplankton into the ice mass deeper in the pack, mainly by flooding of the ice surface. Flooding within the marginal ice zone by wave/swell action would also play a role in introducing water, macro-nutrients and additional algae onto the ice surface. Floe over-washing associated with wave-ice interaction also removes snowcover to create more optimal optical conditions for phytoplankton growth. Moreover, snow removal into the upper ocean would provide another input of freshwater into the mixed layer. As such, the intra-ice bloom in the early austral spring through summer of 2001/2 appears to relate both directly and indirectly to the anomalous pattern in atmospheric circulation. This may help to explain why the bloom occurred relatively far to the south at such an early stage in the season. Only later in the season, when algae were released into the water column seaward of the receding ice edge, could the intensity of this bloom be detected by SeaWiFS. The implications are that the initial timing and distribution of sea ice-related algal blooms may go undetected from space when they occur within a compact sea-ice zone.

There has been much recent discussion on the impact of ice-extent variability on ecological differences, both regional and temporal. For example, Loeb et al. (1997) related the abundance of salps (*Salpa thompsoni*) and Antarctic krill (*Euphausia superba*) in the west Antarctic Peninsula region to interannual variations in sea-ice extent. In East Antarctica, *Nicol et al*. (2000) have also shown that phytoplankton, primary productivity, Antarctic krill (*Euphausia superba*), seabirds and whales tend to concentrate where winter sea-ice extent is maximal, while salps (*Salpa thompsoni*) occur where the extent is minimal. The new findings raise a major issue about the validity of using sea-ice extent alone as a proxy indicator of ecosystem structure and dynamics. Observations during 2001/2 further suggest that the intimate relationship between krill (and other organisms) and sea ice in the West Antarctic Pensinsula region is determined by more complex sea-ice characteristics than ice extent alone. These include the degree of deformation as it affects over-rafting, and the temperature, snowcover and flooding of the ice as they affect phytoplankton (prey) growth. Indeed, Frazer et al. (1997) show that sea-ice structural characteristics appear to be the prime determinant of larval krill abundance and distribution in winter, with larval krill having an affinity for interstitial gaps within over-rafted ice. As such, the conditions in late 2001-early 2002 may have been beneficial for krill, in spite of the anomalously low ice extent in late-winter to spring. On the other hand, negative growth appears to have occurred in larval krill over this period – a negative impact of the anomalously heavy ice conditions.

Similarly, the degree of compaction is a key factor affecting the ability of apex predators to locate and access prey, air to breath and to haul out. Ainley et al. (1994) and Smith et al. (1999) have shown how ice-obligate Adélie penguins can be adversely affected by too little ice, particularly in the winter immediately preceding the breeding season. Additional complications in 2001/2 were that minimal ice extent was accompanied by an almost total lack of leads, and that sea ice persisted through the austral summer in the coastal regions of the WAP, including Marguerite Bay and the area around Palmer Station. Taken together, the new results strongly suggest that more complex sea-ice parameterizations than ice extent alone are required in ecological research, including an index of ice "heaviness" to account for ice thickness/degree of deformation and also the degree of compactness. In other words, there is an important distinction between ice extent and heaviness that is of great ecological importance. As shown by the anomalous patterns in 2001/2, extreme heavy ice conditions can coincide with low ice extent (areal coverage), whereas there is a tendency in the literature to equate ice "heaviness" with extensive sea-ice coverage. In fact, the latter may result from divergent forcing and therefore be less thick and less compact. In all cases, it is the timing and magnitude of ice "events" or regimes relative to key phases in the breeding cycle and ecology of ice organisms that appears to be the key factor (Smith et al., 1998). What made the situation ecologically extraordinary in 2001/2 was the sheer persistence of the combined ice convergence, melt and heavy snowfall over a period of 5-6 months. Moreover, the major ecological impact persisted through summer, with very heavy ice cover remaining in a region that is generally ice-free. This combined with an unusually heavy seasonal snowfall to adversely affect Adélie penguin breeding success in the region.

The important role of oceanic processes in determining sea-ice distribution is beyond the scope of this paper. In their study of recent low ice extents in the Bellingshausen-Amundsen Seas, Jacobs and Comiso (1993, 1997) stipulate that increased upwelling of warm Circumpolar Deep Water - the most prominent water mass west of the Peninsula (Hofmann et al., 1996; Martinson, in prep.) - could reduce both sea ice coverage and thickness (Dinniman and Klinck, 2004). While this may indeed be the case, the current study emphasises that ice dynamics cannot be neglected given their profound impact on the sea-ice thickness distribution, even under melt conditions and particularly during blocking-high episodes. Once again, a less extensive ice cover does not necessarily equate with a thinner one.

The complex conditions in 2001/2 also likely had a major impact on snow and ice optical and microwave properties, with implications for the surface energy budget, ice-ocean-atmosphere feedback effects, phytoplankton growth, UVB penetration and the unambiguous interpretation of satellite data. The limitation of PAR by thick ice with a thick snowcover may be counterbalanced by deformation cracks providing pathways for enhanced downward light penetration. Moreover, it appears that the rafting process itself plays a role in the vertical distribution of phytoplankton in the thick ice by exporting existing infiltration communities downwards. This may be of key importance as a food source for larval krill in interstitial rafted gaps, and may to some extent counterbalance the expected drop-off in primary production with depth in the anomalously thick ice column. Clearly, more research is required, given the contemporaneous negative growth rate observed for larval krill. Also, Perovich et al. (2004) have contrasted sea-ice properties during the austral winters of 2001 and 2002 and documented morphological properties, growth processes and internal permeability of sea ice in this region. Comparison with ice conditions reported in Perovich et al. (2004) again underlines the anomalous nature of ice conditions in the west Antarctic Peninsula region in the austral spring through summer of 2001/2.

In their analysis of the effect of this anomalous atmospheric circulation pattern on sea-ice conditions in the NE Weddell Sea, Turner et al. (2002) suggest that large atmospheric anomalies of this kind are extremely rare. They estimate that the resultant sea-ice conditions in the NE Weddell Sea in the austral summer of 2001/2 were the heaviest for at least 50 years. In addition, Turner et al. (2002) stated that the South Atlantic high pressure was the most anomalous within the approximately 100-year meteorological record at South Georgia. They go on to suggest that while there is a known link between the El-Niño-Southern Oscillation (ENSO) and atmospheric conditions (Harangozo, 2000; Karoly, 1989) and sea-ice variability (Kwok and Comiso, 2002) around the Antarctic Peninsula, the very anomalous conditions in 2001/2 result from the natural variability in high-latitude circulation.

The anomalous atmospheric circulation pattern in 2001/2, which Turner et al. (2002) attribute to natural variability, transported warm air over a region that has been experiencing a significant warming trend over the past 55 years (King, 1994; Smith et al., 1999). Smith et al. (1999) further note that the warming trend at Faraday Station (65.25ºS, 64.26ºW) has been strongest in mid-winter (June), and suggest that this will affect the extent, concentration and thickness of the sea ice, and therefore the associated marine ecology. More work is required to understand the long-term role of blocking-high episodes in the South Atlantic, and how they tie in with, and possibly contribute to, the observed warming trend and associated ecological changes. The latter include the recent increase in the number of Chinstrap penguins and the appearance of Gentoo penguins (*Pygoscelis papua*) - both ice-intolerant species. At the same time, numbers of ice-dependent Adélie penguins have diminished greatly (Fraser and Trivelpiece, 1997; Smith et al., 1999, 2003). Also poorly understood is how the unprecedented coupling of the positive pressure anomaly with major negative pressure anomalies in the Bellingshausen-Amundsen Seas and the Weddell Sea in 2001/2 fits in with recently-identified modes of variability in large-scale atmospheric circulation outlined in Section 1. The latter include El Niño-Southern Oscillation activity (Gloersen, 1995; Kwok and Comiso, 2002; Simmonds and Jacka, 1995), the Southern Annular Mode (Raphael, 2003; Thompson and Solomon, 2002), and the circumpolar propagation of coupled atmospheric-oceanic anomalies within the Antarctic Circumpolar Wave (White and Peterson, 1996). Further questions remain as to whether blocking-high episodes, which tend to be abrupt and difficult to predict, will become more or less prevalent in a global-change scenario.

In an historical perspective, the contribution of anomalous blocking high-related atmospheric circulation related to the palaeo-climate record (as discussed by Smith et al., 1999) is also unknown. The anomalous southward flow of low-latitude maritime air over Antarctica from early spring through summer also has important implications for the interpretation of long-term climate data from terrestrial ice core records. As Smith et al. (1999) point out, the $\delta^{18}O$ isotopic composition of an ice core e.g. from the Dyer Plateau on the Antarctic Peninsula (Thompson et al., 1994), reflects a complex combination of 2 factors in addition to the surface elevation and latitude of deposition and local conditions during the metamorphism of the snow to ice. These are: i) atmospheric processes along the air-mass trajectory from the evaporation source to the site of condensation; and ii) the air temperature at which condensation occurred. The need to understand the full fractionation history in order to interpret ice-sheet isotope signatures is also stressed by Noone and Simmonds (2002, 2004).

Given the key role that sea ice plays in the global climate system by influencing the regional heat budget, surface albedo, and consequently oceanic and atmospheric circulation, variability related to blocking-high events has significant implications. This, and the other factors outlined in this paper, are testament to the highly complex impacts discussed above, and underline the critical need for further research.

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REFERENCES

Ainley, D. G., and D. P. DeMaster, 1990: The upper trophic levels in polar marine ecosystems. *Polar Oceanography*, W. O. Smith, Ed., Academic Press, 599-630.

Ainley D., P. Wilson, K. Barton, G. Ballard, N. Nur, and B. Karl, 1998: Diet and foraging effort of Adélie penguins in relation to pack-ice conditions in the southern Ross Sea. *Polar Biol.,* **20**, 311-319.

Ainley, D. G., C. A. Ribic, and W. R. Fraser, 1994: Ecological structure among migrant and resident seabirds of the Scotia-Weddell Confluence region. *J. Anim. Ecol.,* **63**, 347-364.

Arrigo, K. R., D. L. Worthen, M. P. Lizotte, P. Dixon, and G. Dieckmann, 1997: Primary production in Antarctic sea ice. *Science,* **276**(5311), 394-397.

Beardsley, R. C., R. Limeburner, and W. Brechner Owens, 2004: Drifter measurements of surface currents near Marguerite Bay on the West Antarctic Peninsula shelf during austral summer and fall, 2001 and 2002. *Deep-Sea Res. II*, **51**(17-19), 1947-1964.

Boyd, P. W., C. Robinson, G. J. Savidge, and P. J. Williams, 1995: Water column and sea ice primary production during austral spring in the Bellingshausen Sea. *Deep-Sea Res., II,* **42**, 1177-1200.

Cai, W. J., P. H. Whetton, and D. J. Karoly, 2003: The response of the Antarctic Oscillation to increasing and stabilized atmospheric CO_2. *J. Clim.*, **16**(10), 1525-1538.

Carleton, A. M., 2003: Atmospheric teleconnections involving the Southern Ocean. *J. Geophys. Res.*, **108**(C1), doi 10.1029/2000JC000379.

Chapman, E. W., C. A. Ribic, and W. R. Fraser, 2004: The distribution of seabirds and pinnepeds in Marguerite Bay and their relationship to physical features during austral winter 2001. *Deep-Sea Res. II*, **51**(17-19), 2261-2278.

Cohenour, T., 2001: Palmer: Piles of snow and ice. *The Antarctic Sun* (http:www.polar.org/antsun/2001_1104/continent.html).

Colbeck, S. C., 1982: An overview of seasonal snow metamorphism. *Rev. Geophys.*, **29**(1), 81-96.

Comiso, J. C., 1995: SSM/I ice concentrations using the Bootstrap algorithm. *NASA Rep. 1380*, NASA Goddard Space Flight Center, Greenbelt, Maryland, USA, 40 pp.

Comiso, J. C., T. C. Grenfell, D. L. Bell, M. A. Lange, and S. F. Ackley, 1989: Passive microwave *in situ* observations of winter Weddell Sea ice. *J. Geophys. Res*., **94**(C8), 10,891-10,905.

Connolley, W. M., 2002: Long-term variation of the Antarctic Circumpolar Wave. *J. Geophys. Res.*, **107**(C12), doi:10.1029/2000JC000380.

Dierssen, H. M., and R. C. Smith, 2000: Bio-Optical properties and remote sensing ocean color algorithms for Antarctic Peninsula Waters. *J. Geophys. Res*., **105**(C11), 26,301-26,312.

Dinniman, M. S., and J. M. Klinck, 2004: A model study of circulation and cross-shelf exchange on the west Antarctic Peninsula continental shelf. *Deep-Sea Res. II*, **51**(17-19), 2003-2022.

Drinkwater, M. R., R. Francis, G. Ratier, and D. Wingham, In press: The European Space Agency's Earth Explorer Mission CryoSat: Measuring variability in the cryosphere. *Ann. Glaciol., 39*, in press.

Eicken, H., 1998: Deriving modes and rates of ice growth in the Weddell Sea from microstructural, salinity and stable-isotope data. *Antarctic Sea Ice Physical Properties and Processes*, M. O. Jeffries, Ed., American Geophysical Union, 89-122.

Eicken, H., H. Fischer, and P. Lemke, 1995: Effects of the snow cover on Antarctic sea ice and potential modulation of its response to climate change. *Ann. Glaciol*., **21**, 369-376.

Eicken, H., H. R. Krouse, D. Kadko, and D. K. Perovich, 2002: Tracer studies of pathways and rates of meltwater transport through Arctic summer sea ice. *J. Geophys. Res.,* **107**(C10), 10.1029/2000JC000583.

Emery, W. J., C. W. Fowler, and J. A. Maslanik, 1997: Satellite-derived maps of Arctic and Antarctic sea ice motions: 1988-1994. *Geophys. Res. Lett.,* **24**, 897-900.

Enomoto, H., and A. Ohmura, 1990: The influences of atmospheric half-yearly cycle on sea ice extent in the Antarctic. *J. Geophys. Res.*, **95**, 9497-9511.

Fichefet, T., and M. A. Morales Maqueda, 1999: Modelling the influence of snow accumulation and snow-ice formation on the seasonal cycle of the Antarctic sea-ice cover. *Clim. Dyn.,* **15**, 251-268.

Fraser, W. R., and W. Z. Trivelpiece, 1996: Factors controlling the distribution of seabirds: Winter-summer heterogeneity in the distribution of Adélie penguin populations. *Foundations for Ecological Research West of the Antarctic Peninsula*, R. M. Ross, E. Hofmann, and L. B. Quetin, Eds., American Geophysical Union, 257-272.

Fraser, W. R., W. Z. Trivelpiece, D. G. Ainley, and S. G. Trivelpiece, 1992: Increase in Antarctic penguin populations: Reduced competition with whales or a loss of sea ice due to environmental warming? *Polar Biol.,* **11**, 525-531.

Frazer, T. K., L. B. Quetin, and R. M. Ross, 1997: Abundance and distribution of larval krill, *Euphausia superba*, associated with annual sea ice in winter. *Antarctic Communities: Species, Structure and Survival*, B. Battaglia, J. Valencia and D. W. H. Walton, Eds., Cambridge University Press, 107-111.

Fritsen, C. H., J. N. Kremer, S. F. Ackley, and C. W. Sullivan, 1998: Flood-freeze cycles and algal dynamics in Antarctic pack ice. *Antarctic Sea Ice: Biological Processes*, M. L. Lizotte, and K. R. Arrigo, Eds., American Geophysical Union, 1-21.

Fyfe, J. C., G. J. Boer, and G. M. Flato, 1999: The Arctic and Antarctic Oscillations and their projected changes under global warming. *Geophys. Res. Lett.*, **26**(11), 1601-1604.

Gillett, N. P., and D. W. J. Thompson, 2003: Simulation of recent Southern Hemisphere climate change. *Science*, **302**(5643), 273-275.

Gloersen, P., 1995: Modulation of hemispheric sea ice cover by ENSO events. *Nature,* **373**, 503-506.

Gloersen, P., W. J. Campbell, D. J. Cavalieri, J. C. Comiso, C. L. Parkinson, and H. J. Zwally, 1992: Arctic and Antarctic Sea Ice, 1978-1987: Satellite Passive Microwave Observations and Analysis, *NASA Special Pub. SP-511*, NASA.

Golden, K. M., S. F. Ackley, and V. I. Lytle, 1998: The percolation phase transition in sea ice. *Science,* **282**, 2238-2241.

Hall, A., and M. Visbeck, 2002: Synchronous variability in the Southern Hemisphere atmosphere, sea ice, and ocean resulting from the Annular Mode. *J. Clim.*, **15**, 3043-3057.

Hanson, A. M., 1965: Studies of the mass budget of Arctic pack-ice floes. *J. Glaciol.,* **5**, 701-709.

Harangozo, S. A., 2000: A search for ENSO teleconnections in the west Antarctic Peninsula climate in austral winter. *Int. J. Climatol.,* **20**, 663-679.

Harangozo, S. A., 2004: The impact of winter ice retreat on Antarctic winter sea ice extent and links to atmospheric meridional circulation. *Int. J. Climatol.*, **24**, 1023-1044.

Hines, K. M., D. H. Bromwich, and G. J. Marshall, 2000: Artificial surface pressure trends in the NCEP/NCAR reanalysis over the Southern Ocean and Antarctica. *J. Clim.,* **12**, 3940-3952.

Hofmann, E. E., J.M. Klinck, C.M. Lascara, and D. Smith, 1996: Water mass distributions and circulation west of the Antarctic Peninsula and including Bransfield Strait. *Foundations for Ecological Research West of the Antarctic Peninsula,* R. M. Ross, E. Hofmann, and L. B. Quetin, Eds., American Geophysical Union, 61-80.

Horner, R., S. F. Ackley, G. S. Dieckmann, B. Gulliksen, T. Hoshiai, L. Legendre, I. A. Melnikov, W. S. Reeburgh, M. Spindler and C. W. Sullivan, 1992: Ecology of sea ice biota. 1. Habitat, terminology, and methodology. *Polar Biol.,* **12**, 417-427.

Hunt, G. L., 1991: Marine birds and ice-influenced environments of polar oceans. *J. Marine Systems*, **2**, 233-240.

Hurrell, J. W., and H. van Loon, 1994: A modulation of the atmospheric annual cycle in the Southern Hemisphere. *Tellus*, **46A**, 325-338.

Jacobs, S. S., and J. C. Comiso, 1993: A recent sea-ice retreat west of the Antarctic Peninsula. *Geophys. Res. Lett.,* **20**(12), 1171-1174.

Jacobs, S. S., and J. C. Comiso, 1997: Climate variability in the Amundsen and Bellingshausen Seas. *J. Clim.,* **10**, 697-709.

Jeffries, M. O., and U. Adolphs, 1997: Early winter snow and ice thickness distribution, ice structure and development of the western Ross Sea pack ice between the ice edge and the Ross Ice Shelf. *Antarct. Sci.,* **9**(2), 188-200.

Jeffries, M. O., R. A. Shaw, K. Morris, A. L. Veazey, and H. R. Krouse, 1994: Crystal structure, stable isotopes ($\delta^{18}O$) and development of sea ice in the Ross, Amundsen and Bellingshausen Seas. *J. Geophys. Res*., **99**(C1), 985-995.

Jeffries, M. O., S. Li, R. A. Jaña, H. R. Krouse, and B. Hurst-Cushing, 1998: Late winter first-year ice floe thickness variability, seawater flooding and snow ice formation in the Amundsen and Ross Seas. *Antarctic Sea Ice Physical Processes, Interactions and Variability*, M. O. Jeffries, Ed., American Geophysical Union, 69-88.

Johnson, K. S., R. L. Petty, and J. Tomsen, 1985: Flow injection analysis in seawater macronutrients. *Mapping Strategies in Chemical Oceanography*, A. Zirino, Ed., *Adv. Chem. Ser*., **209**, 7_30.

Kadko, D., 2000: Modeling the evolution of the Arctic mixed layer during the fall, 1997 Surface Heat Budget of the Arctic Ocean (SHEBA) Project using measurements of ^{7}Be. *J. Geophys. Res.,* **105**, 3369-3378.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K.C. Mo, C. Ropelewski, J. Wang, A. Leetmaa, R. Reynolds, R. Jenne, and D. Joseph, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Met. Soc.*, **77**, 437-471.

Karoly, D. J., 1989: Southern Hemisphere circulation features associated with El Niño-Southern Oscillation events. *J. Clim.,* **2**, 1239-52.

Karoly, D. J., 1990: The role of transient eddies in low frequency zonal variations of the Southern Hemisphere circulation. *Tellus*, **42A**, 41-50.

Kidson, J. W., 1988: Interannual variations in Southern Hemisphere circulation. *J. Clim.*, **1**(12), 1177-1198.

Kidson, J. W., 1999: Principal modes of Southern Hemisphere low-frequency variability obtained from NCEP-NCAR reanalyses. *J. Clim.*, **12**, 2808-2830.

King, J. C., 1994: Recent climate variability in the vicinity of the Antarctic Peninsula. *Int. J. Climatol.,* **14**, 357-369.

King, J. C., and J. Turner, 1997: *Antarctic Meteorology and Climatology*, Cambridge University Press, 409 pp.

King, J. C., and S. A. Harangozo, 1998: Climate change in the western Antarctic Peninsula since 1945: Observations and possible causes. *Ann. Glaciol.,* **27**, 571-575.

King, J. C., and J. C. Comiso, 2003: The spatial coherence of interannual temperature variations in the Antarctic Peninsula. *Geophys. Res. Lett*., **30**, 2, 1040, doi:10.1029/2002G015580, 2003.

Kwok, R., and J. C. Comiso, 2002a: Southern Ocean climate and sea ice anomalies associated with the Southern Oscillation*. J. Clim.*, **15**(5), 487-501.

Kwok, R., and J. C. Comiso, 2002b: Spatial patterns of variability in Antarctic surface temperature: Connections to the Southern Hemisphere Annular Mode and the Southern Oscillation. *Geophys. Res. Lett.*, **29**, 10.1029/2002GL015415.

Kwok, R., H. J. Zwally, and D. Yi, 2004b: ICESat observations of Arctic sea ice: a first look. *Geophys. Res. Lett.*. In press.

Lange, M. A., P. Schlosser, S. F. Ackley, P. Wadhams, and G. S. Dieckmann, 1990: 18O concentrations in sea ice of the Weddell Sea, Antarctica*. J. Glaciol.,* **36**(124), 315-323.

Lejenäs, H., 1984: Characteristics of southern hemisphere blocking as determined from a time series of observational data. *Quart. J. Royal Met. Soc.*, **110**, 967-979.

Liu, J., D. G. Martinson, X. Yuan, and D. Rind, 2002a: Evaluating simulated Antarctic sea ice variability and its global teleconnections in global climate models. *Int. J. Climatol.*, **22**(8), 885-900.

Liu, J., X. Yuan, D. Rind, and D. G. Martinson, 2002b: Mechanism Study of the ENSO and Southern High Latitude Climate Teleconnections. *Geophys. Res. Lett.*, **29**(14), 1029/2002GL015143.

Liu, J., J. A. Curry, and D. G. Martinson, 2004: Interpretation of recent Antarctic sea ice variability. *Geophys. Res. Lett.*, **31**, L02205, doi:10.1029/2003GL018732.

Loeb, V., V. Siegel, O. Holm-Hansen, R. Hewitt, W. Fraser, W. Trivelpiece, and S. Trivelpiece, 1997: Effects of sea-ice extent and krill or salp dominance on the Antarctic food web. *Nature,* **387**, 897-900.

Massom, R. A., J. C. Comiso, A. P. Worby, V. I. Lytle, and L. Stock, 1999: Regional classes of sea ice cover in the East Antarctic pack observed from satellite and *in situ* data during a winter time period. *Rem. Sens. Environ*., **68**(1), 61-76.

Massom, R. A., H. Eicken, C. Haas, M. O. Jeffries, M. R. Drinkwater, M. Sturm, A. P. Worby, X. Wu, V. I. Lytle, S. Ushio, K. Morris, P. A. Reid, S. Warren, and I. Allison, 2001: Snow on Antarctic sea ice. *Rev. Geophys*., **39**(3), 413-445.

Massom, R. A., M. J. Pook, J. C. Comiso, N. Adams, J. Turner, T. Lachlan-Cope, and T. Gibson, 2004: Precipitation over the interior East Antarctic Ice Sheet related to mid-latitude blocking high activity. *J. Clim.*, **17**, 1914-1928.

Meehl, G. A., J. W. Hurrell, and H. van Loon, 1998: A modulation of the mechanism of the semiannual oscillation in the Southern Hemisphere. *Tellus,* **50A**, 442-450.

Menendez, C. G., V. Serafini, and H. Le Treut, 1999a: The storm tracks and energy cycle of the Southern Hemisphere: Sensitivity to sea-ice boundary conditions. *Ann. Geophys.*, **17**(11), 1478-1492.

Meredith, M. P., I. A. Renfrew, A. Clarke, J. C. King and M. A. Brandon, 2004: Impact of the 1997/98 ENSO on the upper waters of Marguerite Bay, western Antarctic Peninsula. *J. Geophys. Res*., **109**(C9), C09013, 10.1029/2003JC001784.

Murray, R. J., and I. Simmonds, 1995: Responses of climate and cyclones to reductions in Arctic winter sea ice. *J. Geophys. Res.*, **100**, 4791–4806.
Nicol, S., T. Pauly, N. L. Bindoff, S. Wright, D. Thiele, G. W. Hosie, P. G. Strutton, and E. Woehler, 2000: Ocean circulation off East Antarctica affects ecosystem structure and sea-ice extent. *Nature,* **406**(3), 504-507.
Noone, D., and I. Simmonds, 2002: Annular variations in moisture transport mechanisms and the abundance of d18O in Antarctic snow. *J. Geophys. Res*., **107**(D24), 4742, doi:10.1029/2002JD002262.
Noone, D., and I. Simmonds, 2004: The sea ice control on water isotope transport to Antarctica and implications for ice core interpretation. *J. Geophys. Res*., **109**, D07105, doi:10.1029/2003JD004228.
Notz, D., M. G. McPhee, M. G. Worster, G. A. Maykut, K. H. Schlünzen, and H. Eicken, 2003: Impact of underwater-ice evolution on Arctic summer sea ice. *J. Geophys. Res*., **108**(C7), 3223, doi:10.1029/2001JC001173.
O'Reilly, J. E., S. Maritorena, B. G. Mitchell, D. A. Siegel, K. L. Carder, S. A. Garver, M. Kahru, and C. McClain, 1998: Ocean color chlorophyll algorithms for SeaWiFS. *J. Geophys. Res.*, **103**(C11), 24,937-24,953.
Orsi, A. H., T. Whitworth, III, and W. D. Nowlin, Jr., 1995: On the meridional extent and fronts of the Antarctic Circumpolar Current. *Deep-Sea Res.,* **42**, 641-673.
Perovich, D. K., 1996: *The Optical Properties of Sea Ice*. Cold Reg. Res. and Engin. Lab. (CRREL) Report 96-1, Cold Regions Research and Engineering Laboratory, Hanover, NH, USA.
Perovich, D. K., T. C. Grenfell, J. A. Richter-Menge, B. Light, W. B. Tucker III, and H. Eicken, 2003: Thin and thinner: Ice mass balance measurements during SHEBA. *J. Geophys. Res.,* **108**, doi:10.1029/2001JC001079.
Perovich, D. K., B. C. Elder, K. J. Claffey, S. E. Stammerjohn, R. C. Smith, S. F. Ackley, H. R. Krouse, and A. J. Gow, 2004: Winter sea-ice properties in Marguerite Bay, Antarctica. *Deep-Sea Res. II*, **51**, 2023-2039.
Pook, M., and T. Gibson, 1999: Atmospheric blocking and storm tracks during SOP-1 of the FROST Project. *Aust. Met. Mag., Special Edition*, 51-60.
Quetin, L. B., and R. M. Ross, 1991: Behavioral and physiological characteristics of the Antarctic krill, *Euphausia superba. Am. Zool.,* **31**, 49-63.
Quetin, L. B., and R. M. Ross, 2001: Environmental variability and its impact on the reproductive cycle of Antarctic krill. *Am. Zool.,* **41**, 74-89.
Raphael, M. N., 2003: Impact of observed sea-ice concentration on the Southern Hemisphere extratropical atmospheric circulation in summer. *J. Geophys. Res*., **108**(D22), 4687, doi:10.1029/2002JD003308.
Ross, R. M., E. E. Hofmann, and L. B. Quetin, 1996: *Foundations for Ecological Research West of The Antarctic Peninsula*, American Geophysical Union, 448 pp.
Ross, R. M., L. B. Quetin, T. Newberger, and S. A. Oakes, 2004: Growth and behavior of larval krill (*Euphausia superba*) under the ice in late winter 2001 west of the Antarctic Peninsula. *Deep-Sea Res. II*, **51**(17-19), 2169-2184.
Scambos, T. A., C. Hulbe, M. Fahnestock, and J. Bohlander, 2000: The link between climate warming and break-up of ice shelves in the Antarctic Peninsula. *J. Glac.,* **46**(154), 516-530.
Simmonds, I., 2003: Modes of atmospheric variability over the Southern Ocean. *J. Geophys. Res.*, **108**(C4), 10.1029/2000JC000542.
Simmonds, I. and T. H. Jacka, 1995: Relationships between the interannual variability of Antarctic sea-ice and the Southern Oscillation. *J. Clim*., **8**, 637-647.
Simmonds, I., and K. Keay, 2000: Variability of Southern Hemisphere extratropical cyclone behaviour 1958-97. *J. Clim*., **13**, 550-561.
Simmonds, I. and J. C. King, 2004: Global and hemispheric climate variations affecting the Southern Ocean. *Antarct. Sc.*, **16**, 401–413.
Sinclair, M. R., 1996: A climatology of anticyclones and blocking for the Southern Hemisphere, *Mon. Wea. Rev*., **124**, 245-263.
Smith, R. C., and S. E. Stammerjohn, 2001: Variations of surface air temperature and sea ice extent in the western Antarctic Peninsula (WAP) region. *Ann. Glaciol.,* **33**, 493-500.
Smith, R. C., and S. E. Stammerjohn, 2003: Palmer LTER: Seasonal process sea ice cruise June-July 1999 (NBP99-06). *Antarct. J. U.S*., in press ??????.
Smith R.C., K. S. Baker, and P. Dustan, 1981: Fluorometric techniques for the measurement of oceanic chlorophyll in the support of remote sensing. *Scripps Institution of Oceanography Ref. 81-17*, Visibility Laboratory, Scripps Institute of Oceanography, University of California.

Smith, R. C., K. S. Baker, W. R. Fraser, E. E. Hofmann, D. M. Karl, J. M. Klinck, L. B. Quetin, B. B. Prézelin, R. M. Ross, W. Z. Trivelpiece, and M. Vernet, 1995: The Palmer LTER: A long-term ecological research program at Palmer Station, Antarctica. *Oceanography,* **8**(3), 77-86.

Smith, R. C., S. E. Stammerjohn, and K. S. Baker, 1996: Surface air temperature variations in the Western Antarctic Peninsula region. *Foundations for Ecological Research West of the Antarctic Peninsula*, R. M. Ross, E. Hofmann, and L. B. Quetin, Eds., American Geophysical Union, 105-121.

Smith, R. C., K. S. Baker, and S. E. Stammerjohn, 1998: Exploring sea ice indexes for polar ecosystem studies. *BioScience,* **48**(2), 83-93.

Smith, R. C., K. S. Baker, M. L. Byers and S. E. Stammerjohn, 1998: Primary Productivity of the Palmer Long Term Ecological Research Area and the Southern Ocean. *J. Mar. Syst*., **17**(1-4): 245-259.

Smith, R. C., D. Ainley, K. Baker, E. Domack, S. Emslie, W. Fraser, J. Kennett, A. Leventer, E. Mosley-Thompson, S. Stammerjohn, and M. Vernet, 1999: Marine ecosystem sensitivity to climate change. *Bioscience,* **49**(5), 393-404.

Smith, R. C., K. S. Baker, H. M. Dierssen, S. E. Stammerjohn, and M. Vernet, 2001: Variability of primary production in an Antarctic marine ecosystem as estimated using a multi-scale sampling strategy. *Amer. Zool.,* **41**, 40-56.

Smith, R. C., W. R. Fraser, S. E. Stammerjohn and M. Vernet, 2003a: Palmer Long-Term Ecological Research on the Antarctic marine ecosystem. *Antarctic Peninsula Climate Variability: Historical and Paleoenvironmental Perspectives*, E. W. Domack, A. Leventer, A. Burnett, R. Bindschadler, P. Conley, and M. Kirby, Eds., American Geophysical Union, 131-144.

Smith, R. C., W. R. Fraser and S. E. Stammerjohn, 2003b: Climate variability and ecological response of the marine ecosystem in the western Antarctic Peninsula (WAP) region. Climate Variability and Ecosystem Response at Long-Term Ecological Research (LTER) Sites, D. Greenland, D. G. Goodin and R.C. Smith, Eds., Oxford University Press, 158-173.

Smith, R. C., D. G. Martinson, S. E. Stammerjohn. R. A. Iannuzzi, and K. Ireson (In prep.). Western Antarctic Peninsula region pigment biomass and sea ice spatial/temporal distributions and annual variability.

Smith, W. O., and D. M. Nelson, 1985: Phytoplankton bloom produced by a receding ice edge in the Ross Sea: Spatial coherence with the density field. *Science,* **227**, 163-166.

Stammerjohn, S. E., and R. C. Smith, 1996: Spatial and temporal variability in Western Antarctic Peninsula sea ice coverage. *Foundations for Ecological Research West of the Antarctic Peninsula*, R. M. Ross, E. E. Hofmann and L. B. Quetin, Eds., American Geophysical Union, 81-104.

Stammerjohn, S. E., and R. C. Smith, 1997: Opposing Southern Ocean climate patterns as revealed by trends in sea ice coverage. *Clim. Change,* **37**, 617-639.

Stammerjohn, S. E., M. R. Drinkwater, R. C. Smith and X. Liu, 2003: Ice-atmosphere interactions during sea-ice advance and retreat in the western Antarctic Peninsula region. *J. Geophys. Res*., **108**(C10): 10.1029/2002JC001543.

Thompson, D. W. J., and J. M. Wallace, 2000: Annular modes in extra-tropical circulation, Part I: Month-to-month variability. *J. Clim.*, **13**, 1018-1036.

Thompson, D. W. J., J. M. Wallace, and G. C. Hegeri, 2000: Annular modes in extra-tropical circulation, Part II: Trends*. J. Clim.*, **13**, 1000-1016.

Thompson, D. W. J., and S. Solomon, 2002: Interpretation of recent Southern Hemisphere climate change. *Science*, **296**(5569), 895-899.

Trenberth, K. E., and K. C. Mo, 1985: Blocking in the Southern Hemisphere. *Mon. Wea. Rev.,* **113**, 3-21.

Turner, J., 2004: The El Niño-Southern Oscillation and Antarctica. *Internat. J. Climatol*., **24**, 1-31.

Turner, J., and S. Colwell, 2004: Antarctic climate change over the last 50 years – Results from the SCAR READER project. *Terra Nostra, Schriften der Alfred-Wegener-Stiftung 2004/4*, Abstract Volume, Abstract S8/O06, 214-215.

Turner, J., S. A. Harangozo, G. J. Marshall, J. C. King, and S. R. Colwell, 2002: Anomalous atmospheric circulation over the Weddell Sea, Antarctica, during the austral summer of 2001/02 resulting in extreme sea-ice conditions. *Geophys. Res. Lett.,* **29**(4), 10.1029/2002GL015565.

Ukita, J., et al. (in prep.).

Van den Broecke, M. R., 1998: The semi-annual oscillation and Antarctic climate, part 2: Recent changes. *Antarct. Sc.,* **10**, 184-191.

van Loon, H., 1956: Blocking action in the Southern Hemisphere. Part 1. *Notos*, **5**, 171-175.

Vaughan, D. G. and C. S. M. Doake, 1996: Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula. *Nature,* **379**(6563), 328-331.

Vaughan, D. G., G. J. Marshall, W. M. Connolley, J. C. King and R. Mulvaney, 2001: Climate change - devil in the detail. *Science,* **293**, 1777-1779.

Warren, S. G., 1982: The optical properties of snow. *Rev. Geophys. Space Phys.,* **20**, 67-89.

Weatherly, J. W., J. E. Walsh, and H. J. Zwally, 1991: Antarctic sea ice variations and seasonal air temperature relationships. *J. Geophys. Res.,* **96**, 15,119-15,130.

White, W. B., and R. G. Peterson, 1996: An Antarctic circumpolar wave in surface pressure, wind, temperature and sea ice extent. *Nature,* **380**(6576), 699-702.

White, W. B., C. Shyh-Chin, R. J. Allan, and R. C. Stone, 2002: Positive feedbacks between the Antarctic Circumpolar Wave and the global El Niño-Southern Oscillation Wave. *J. Geophys. Res.*, **107**(C10), 10.1029/2000JC000581.

Wilson, P. R., D. G. Ainley, N. Nur, S. S. Jacobs, K. J. Barton, G. Ballard, and J. C. Comiso, 2001: Adélie penguin population change in the Pacific sector of Antarctica: Relation to sea ice extent and the Antarctic Circumpolar Current. *Mar. Ecol. Prog. Ser.,* **213**, 301-309.

Worby, A. P. and I. Allison, 1999: A ship-based technique for observing Antarctic sea ice. Part I: Observational techniques and results. *Antarctic CRC Rep. 15*, Antarctic Cooperative Research Centre, University of Tasmania.

Worby, A. P., M. O. Jeffries, W. F. Weeks, K. Morris, and R. Jaña, 1996: The thickness distribution of sea ice and snow cover during late winter in the Bellingshausen and Amundsen Seas. *J. Geophys. Res.,* **101**(C12), 28,441-28,455.

Wright, A. D. F., 1974: Blocking action in the Australian region. *Techn. Rep. 10*, Australian Bureau of Meteorology, 29 pp.

Wu, X., W. F. Budd, V. I. Lytle, and R. A. Massom, 1999: The effect of snow on Antarctic sea ice simulations in a coupled atmosphere-sea ice model. *Clim. Dyn.*, **15**(2), 127-143.

Yuan, X., 2004: ENSO-related impacts on Antarctic sea ice: a synthesis of phenomenon and mechanisms. *Antarct. Sc.*, **16**, 415–425.

Yuan, X. J., and D. G. Martinson, 2001: The Antarctic Dipole and its predictability. *Geophys. Res. Lett.,* **28**(18), 3609-3612.

Yuan, X., and D. G. Martinson, 2000: Antarctic sea ice variability and its global connectivity. *J. Clim.,* **13**, 1697–1717.

Zwally, H. J., J. C. Comiso, C. L. Parkinson, D. J. Cavalieri, and P. Gloersen, 2002: Variability of Antarctic sea ice 1979–1998. *J. Geophys. Res.*, **107**(C5), 3041, doi:10.1029/2000JC000733.

Fig. Captions

Fig. 1. Map of the study region, showing the cruise track voyage NBP01-05 of the R/V *Nathaniel B. Palmer*. The inset map is of the Marguerite Bay region, where the sea-ice work was concentrated (in September to October 2001).

Fig. 2. Synoptic charts of NNR mean sea level pressure in the South Pacific, South Atlantic and Antarctic Peninsula region at 23:00Z on a) September 24, b) September 27, c) September 29, 2001, and d) October 8, 2001, showing the initial establishment of the blocking high in the South Atlantic and the deep low-pressure centres in the Amundsen/Bellingshausen and western Weddell Seas. Note the resultant strong northwest-to-southeast alignment and close spacing of isobars in the West Antarctic Peninsula region. Pressure values are in hPa.

Fig. 3. Monthly mean composite anomaly map of NNR 500 hPa geopotential height for the Southern Ocean and Antarctica for September 2001 to February 2002 (based upon the mean September to February 1980-2001). X is the approximate location of Marguerite Bay and the 2001 field experiment, and BH and LP denote the high-pressure (positive or blocking-high) and low-pressure (negative) anomalies respectively.

Fig. 4. Anomaly maps of monthly-mean 10-m wind fields for July 2001 through February 2002, for the West Antarctic Peninsula region, derived from monthly-mean NCEP/NCAR Reanalysis Project data from 1980 to 2001 inclusive. A wind-vector arrow scale is provided in the bottom left of each panel.

Fig. 5. Time series plots of a) air temperature, b) wind speed and direction, and c) relative humidity measured at the R/V *Nathaniel B. Palmer* whilst in the Bellingshausen Sea ice zone, September 12 to October 24, 2001. These are hourly averages of data collected every minute.

Fig. 6. Time series plot of the blocking index anomaly (BIA) at 30ºW from 1970 to 2001 for a) October, b) the austral spring (September to November inclusive), and c) the austral summer (December to

February inclusive). These data were derived from monthly mean values from NCEP/NCAR Reanalysis dataset. In (c), A marks the high BIA in the summer of 2001/2.

Fig. 7. Time series of meteorological variables at a height of 850 hPa and at 65ºS, 70ºW (in the vicinity of Marguerite Bay), computed from NNR data for the period 1990-2002. Monthly mean values are plotted for a) wind direction, b) wind speed, c) air temperature, d) relative humidity, and e) precipitation rate. The arrows highlight the period of interest.

Fig. 8. CTD data from October 6 and 13, 2001, in Marguerite Bay (at ~68ºS, 70ºW). a) Temperature and salinity (T/S) profiles of the upper ocean; b) a close-up of the salinity profiles across the approximate depth of the ocean mixed layer, with the salinity difference shaded; and c) is a plot of potential temperature versus salinity, with isopycnals and the freezing point (dashed line) marked. The pressures of 125 and 300db correspond to depths of 123.7 m and 296.8 m (for event 178).

Fig. 9. Drift tracks of 3 satellite-tracked ARGOS sea ice buoys in the Marguerite Bay region, August to October 2001, with start and stop dates marked.

Fig. 10. NOAA AVHRR channel 1 (visible) image of the Marguerite Bay region, 21:22Z on October 13, 2001, showing the extraordinary linear ice edge. Note the polynya on the downwind (lee) side of Adelaide Island. The ship's location is marked as NBP.

Fig. 11. Images of monthly-mean DMSP SSM/I sea ice concentrations from July to February 2002, with 15% and 75% ice contours marked (the white and dark lines respectively), and the Palmer LTER study region marked (white rectangle). The 1980-2001 mean ice-edge location for each month is marked as a white dotted line for the 15% ice-concentration extent and a black dotted line for the 75% ice-concentration extent. Contemporary monthly-mean wind velocity data from the NNR dataset are superimposed.

Fig. 12. Underwater photograph of the vertical side profile of a typical rafted floe, taken by SCUBA divers in Marguerite Bay in October 2001 during LTER cruise NBP01-5.

Fig. 13. Snow and ice thickness transects across two floes: a) at ~67.5ºS, 70.7ºW on 16-17 September 2001 (prior to the onset of the atmospheric circulation anomaly), and b) at 68.25ºS, 69.8ºW on September 30 (after the anomaly onset), where 0 cm is sea level. Thickness profiles of the flooded layer, damp and saturated snow are also marked.

Fig. 14. Paired plots of the frequency distributions of $\delta^{18}O$ values of ice cores and $\delta^{18}O$ as a function of depth in the ice core below the ice-floe surface, over three 10-day periods: a) and b) September 16-26; c) and d) September 27-October 7; and e) and f) October 8-18, 2001.

Fig. 15. Profiles of ice algal biomass as chlorophyll *a* in mg m^{-3} (a to c) and nutrients as nitrate in μM (d to f) for three time intervals during the study period, 16 to 26 September (a and d), 27 September to 7 October (b and e) and 7 to 17 October (c and f). For the first period, averages for 3 cores are presented, sampled at 5 cm intervals (20 depths), where n = 44. During the second period, 6 cores were sampled at a 5-cm resolution and 23 depth intervals, with n = 82. During the third period, 5 cores were sampled at 5 cm intervals, at 28 different depths, n = 91.

Fig. 16. Monthly averaged Global Area Coverage SeaWiFS images of the West Antarctic Peninsula region for the period October 2001 through January 2002 showing a-d) pigment biomass, and e-h) monthly anomalies of pigment biomass, computed as the difference between the actual month and the seven-year SeaWiFS climatology (not shown here). In (a-d), the following are marked: 500m and 1000m ocean-depth contours (white and black lines respectively), sea ice extent at the beginning and end of each month (dotted grey and black lines respectively), and monthly averaged NCEP/NCAR Reanalysis Project wind speed vectors (black arrows). Sea ice is masked in white. AP is the Antarctic Peninsula and MB Marguerite Bay.

Fig. 17. *In situ* growth increments (% growth per intermolt period) for larval Antarctic krill collected from under the ice in austral spring of 1991 (open squares), 1993 (open triangles) and 2001 (closed circles). Larvae collected in 2001 averaged from 9.6 to 11.1 mm in total length. Points are the mean

growth increment for an experiment with from 7 to 15 individual measurements. Error bars are standard error.

POSSIBLE SUPPLEMENTARY ONLINE MATERIAL:

Fig. SM1. A composite mosaic of AVHRR imagery, October 8, 2001, with clouds tracing out the meridional advection of warm moist maritime air to the Antarctic Peninsula region. The approximate direction of airflow is indicated by the arrows. The white X indicates the study region (around Marguerite Bay). Courtesy AMRC, University of Wisconsin.

Fig. SM2. Photograph of typical snowpits on first-year ice during the September-October field experiment in the WAP region.

Fig. SM3. Photograph of an ice core from the upper " block" of a typical melting (rotting) rafted first-year ice floe, September 26, ~68.03ºS, 69.85ºW.

Fig. SM4. Photographs of a) a snowpit showing snow-ice formation, September 27, 2001 (??S, ??W), and b) an upturned floe at 68.18ºS, 69.78ºW on October 4, showing the broad upper horizon of snow ice. Note the discolouration within the snow-ice layer due to the presence of phytoplankton.

Fig. SM5. Sea-ice conditions in the marginal ice zone, showing the extraordinary brash-frazil-phytoplanton "soup", as observed from the ship on 22 October 2001 (at ~64.75ºS, 64.90ºW in the N. Bellingshausen Sea)

TABLES

Table 1. Mean ice and snow thicknesses measured on other cruises to West Antarctica. **JEFFRIES et al. (1994) NEEDED**.

Distance from ice edge (km)	*Zi (m)*	*Zs (m)*	*Date*	*Region*	*Reference*
50-100	0.34	0.07	Aug-Sep 1993	BAS	*Worby et al.*, 1996
0-200	0.55 ±0.24	0.07 ±0.06	May-Jun 1995	WRS	*Jeffries and Adolphs*, 1997
0-200	0.22 ±0.11	0.04 ±0.03	Jun-Jul 1999	WAP	*Smith and Stammerjohn*, 2003
350-550	0.55	0.23	Aug-Sep 1993	BAS	*Worby et al.*, 1996
200-600	0.80 ±0.38	0.17 ±0.11	May-Jun 1995	WRS	*Jeffries and Adolphs*, 1997
200-550	0.49 ±0.11	0.04 ±0.03	Jun-Jul 1999	WAP	*Smith and Stammerjohn*, 2003
Average	0.48	0.19	Aug-Sep 1993	BAS	*Worby et al*, 1996
Average	0.51 ±0.27	0.11 ±0.10	May-Jun 1995	WRS	Jeffries and Adolphs, 1997
Average	0.36 ±0.18	0.04 ±0.03	Jun-Jul 1999	WAP	*Smith and Stammerjohn*, 2003

Table 2. Mean values of the thicknesses of snow (Z_s), ice (Z_{ice}), freeboard, slush layer (Z_{slush}) and wicked layer (Z_{wicked}), freeboard, and ice-surface temperature (T_{is}) along 3 transects. Transect 2 was at ~68.0ºS, 69.9ºW. Units are centimetres. Zero freeboard is sea level.

	Z_s	Z_{ice}	Freeboard	Z_{slush}	Z_{wicked}	T_{is}
TR1, Sept. 16-17 n = 80	30.2±6.5	93.6±33.0	-0.2±2.4	0.8±1.4	4.1±3.2	-2.1±0.4
TR2, Sept. 26 n = 50	51.6±9.0	89.9±32.8	-16.3±7.6	18.3±5.2	25.0±6.2	-1.8±0.1
TR3, Sept. 30 n = 30	32.1±4.9	71.4±19.1	-14.0±4.1	14.0±4.0	19.0±4.0	-1.9±0.1

Table 3. Chlorophyll *a* concentrations (in mg m^{-3}) and nutrient concentrations (in µM) in the ice column sampled and averaged in the three 10-day periods from mid-September to mid-October, 2001. In these periods, n = 44, n=82, and n = 91 respectively.

Period	**Chl *a***	**Phosphate**	**Silica**	**Nitrate**	**Ammonium**
Sept. 16-26	9.59±9.25	0.33±0.99	9.53±1.84	3.38±0.69	1.67±1.79
Sept. 27-Oct. 7	12.9±8.6	0.37±0.14	9.1±2.9	2.48±1.04	0.53±0.22
Oct. 8-18	7.7±13.1	0.35±0.19	10.76±4.6	2.2±1.55	0.75±0.41