Sensitivity of the southern hemisphere circulation to leads in the Antarctic pack ice

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SUMMARY

To assess the sensitivity of the southern hemisphere circulation to changes in the fraction of open water in the sea ice we have conducted four experiments with a July 21-wave General Circulation Model (GCM) with this fraction set to 5, 50, 80 and 100%. The mean surface temperatures and the surface atmospheric temperatures over the sea ice increased as the water fraction increased and the largest changes were simulated adjacent to the coast.

Significant anomalies in the surface heat fluxes, particularly those of sensible heat, accompanied the decrease in the sea ice concentration. Substantial atmospheric warming was simulated over and in the vicinity of areas in which leads were considered. In all but one experiment there were anomalous easterlies between about 40 and 60°S with westerly anomalies further to the south. The surface pressure at high latitudes appears to change in a consistent fashion with the fraction of open water, with the largest changes occurring in the Weddell and near the Ross Seas.

Some of the feedbacks which may enhance the responses here, but which are not included in our model, are discussed.

1. INTRODUCTION

The study of climate in polar regions is complicated by the existence of sea ice and the many feedback processes with which it is associated. The potential impact on hemispheric or global climate of forcings in the high latitudes, and particularly those associated with sea ice perturbations, are attracting attention. The Antarctic region is of considerable importance in this because of the very dynamic nature of Antarctic sea ice and the fact that it is located at lower latitudes and less confined by land masses than its northern counterpart.

When sea ice is present in a given region it acts as an insulator between the relatively warm ocean and the cold overlying air. The ice not only prevents significant fluxes of heat and moisture reaching the lower atmospheric layers but its high albedo reflects most of any shortwave radiation. Hence, atmospheric structure can be expected to depend strongly on whether there is a sea ice cover or not. However, even if sea ice covers most of the ocean in a given region, small areas of open water are able to affect substantially the fluxes of heat across the surface–atmosphere boundary (e.g. Andreas and Murphy 1986), and hence the allowance for the effect of the open areas (leads) is of importance.

A number of researchers have studied the concentration of sea ice in the two hemispheres and recent analyses suggest there are considerable amounts of open water in the ice pack in both polar regions, with large variations through the year. Among these we mention Zwally et al. (1983), the Commander, Naval Oceanography Command (1985), Parkinson et al. (1987) and Gloersen and Campbell (1988). This last study also points out trends in the sea ice concentration over the period of their analyses. The studies also show that the concentration varies considerably with location and, in general, becomes higher at higher latitudes, as one would expect. It follows that a realistic treatment of the effects of sea ice must take into account that it is not in the form of a continuous cover but has leads within it.

In this paper we employ a General Circulation Model (GCM) to study the impact on the atmospheric circulation of areas of open water in the polar ice packs. In the model we use (detailed in Simmonds 1985, and Simmonds et al. 1988), the sea ice distribution and thickness is specified and not permitted to change but it does incorporate the dynamic
and thermodynamic effects of leads. While this is by no means a complete treatment of sea ice in the GCM, we believe an appropriate consideration of leads is an important step in the treatment of sea ice and its effects.

2. DESCRIPTION OF LEADS PARAMETRIZATION AND DESIGN OF EXPERIMENTS

The parametrization of the effect of leads in the model used here takes into account two important modifications that the exposure of open water introduces: the impacts on the surface fluxes and on the vertical profile of the radiative heating rates. The method by which the scheme does this is described in detail by Simmonds and Budd (1990). For convenience we sketch here its central features. In the sea ice zones each 'grid box' is conceived to be broken into a sea-ice part and an open-leads part. The fraction of open water in the ice pack, \( f_w \), is, for simplicity, not considered to vary with position, and we also specify the ice thickness everywhere as 2 m. Budd (1986) has shown that the winter Antarctic sea ice displays considerable spatial variability in thickness and the overall average is probably closer to 1 m. However, the work of Maykut (1978) and others suggests there is little change in the surface fluxes over ice when its thickness exceeds 1 m, and our interest here is centred on the much greater sensitivity to the open-water fraction.

A simplification that is made in the scheme is that the horizontal mixing in the atmospheric boundary layer is sufficiently rapid that for each grid box we need only a single (average) vertical atmospheric profile, which is that computed by the model, rather than separate profiles above the sea ice and the leads. In particular this means that the temperature, moisture and wind at the lowest model level (which in part determine the various fluxes) apply to the whole grid box (i.e. over the ice as well as the leads). To what extent this simplification is justified depends to a great extent on the typical size of the leads which we imagine in our parametrizations, as compared with the height of the lowest level in the model (about 75 m above the surface). Fairall and Markson (1987) have pointed out that many studies assume that the 10 m wind is the same over sea ice and leads and produced evidence to indicate that this assumption is unrealistic. The assumption is probably much more justified at the 75 m level. If our population of leads (or ice floes) is made up of numerous, small (scales \( \leq 75 \) m) members the assumption is probably reasonable and the atmosphere is well mixed horizontally at the height of the lowest level. Direct ship observations reported by Allison (1989) indicate most frequent floe sizes in the pack ice off East Antarctica of less than 100 m. Note, however, that for large leads or polynyas one would not expect the assumption to be valid.

An important impact of leads in sea ice is their effect on the surface fluxes of moisture, temperature and momentum. The parametrization of these in the model over continuous sea ice and open water is discussed in Simmonds (1985). For a given atmospheric structure, these fluxes are very different over the two domains. To allow for the strong nonlinear effects the scheme performs separate flux calculations over the sea ice and open water parts of the box. (The surface temperature over the open-water part is specified as the freezing point of sea water, 271.4 K.) These fluxes are then averaged with the appropriate area weighting and the transports communicated to the lowest level of the model.

As far as the radiative processes are concerned, the assumption of small leads means that the radiative heating rates over the two surface types can be thought of as 'well mixed'. It is then necessary to perform only one radiation calculation for each grid-box column. The surface albedo for such a column is taken to be the area-weighted average of the sea ice and open-water albedo. These fluxes are used over the ice portions of a
grid box to calculate the surface temperature over that domain. The temperature and fluxes which we discuss later are the appropriate grid-box average taken over the two surface types.

The present experiments were designed to determine the sensitivity of the 'perpetual July' version of the model to changes in the leads fraction in all polar sea ice. We report here on four simulations in which $f_w$ is set to 0.05, 0.50, 0.80 and 1.00. We refer to these experiments as W05, W50, W80 and W100, respectively. (Note that the last experiment is equivalent to the complete removal of sea ice.) The observational studies referred to above report a range from 5 to 50% open water in the winter Antarctic pack ice. The fraction specified in two of the experiments are in this range. The other cases are designed to test the sensitivity outside these limits.

The 'control' simulation of the 21-wave GCM, with which these experiments are compared, was run with no leads (i.e. $f_w = 0.0$) and its climate was estimated from a run of 600-day period with 'perpetual July' conditions. (The climatology of this control is shown compared with observations, in Simmonds et al. 1988.) We display in Fig. 1 the July sea-level pressure simulated by the model in the southern extratropics, from which an assessment of the quality of the model product can be made. The climates of the anomaly runs were derived from the analysis of 300-day simulations, after an adjustment period of 90 days.

Figure 1. Mean-sea-level pressure simulated in the control run. The contour interval is 5 hPa.

3. Results

The most immediate impact of the change of $f_w$ is the modification of the grid-box average temperature over the sea ice zone. The differences between the experiments W50 and W100 and the control in the southern hemisphere (SH) are displayed in
Fig. 2. (Because the sea surface temperature outside of the sea ice zone had not been changed, the zero contour on these difference plots corresponds to the location of the ice edge.) The decrease in sea ice concentration is accompanied by an increase in the surface temperature over the ice–water mix. In all simulations the smallest temperature differences are near the edge; they become greater at higher latitudes, reaching in excess of 30 K in the W100 experiment. Virtually all of the Antarctic continent undergoes surface warming, the magnitude of which decreases with distance from the coast.

These changes of surface temperature have an impact on the surface heat fluxes. In Fig. 3 we show the change in the grid-box average sensible-heat flux in these same two experiments. It can be seen that in the ice region this anomalous flux is everywhere positive (upwards), assuming typical values of 60–80 W m\(^{-2}\) and an extreme value in W100 in excess of 300 W m\(^{-2}\) in the western Ross Sea. (These values are the differences between fluxes of opposite sign). Just to the north of the ice edge the anomalies assume the opposite sign. This is consistent with the findings of Simmonds (1981) and Mitchell and Senior (1989). The juxtaposition of these positive and negative anomalies is presumably due to the fact that low-level air flowing off the continent picks up more sensible heat as it flows over ice with leads than it would over solid ice. Hence such air reaches the open ocean warmer, and the upward flux of sensible heat there is considerably reduced. The latent-heat flux anomalies in the experiments (not shown here) assume similar structure but their magnitudes are typically one third of those for the sensible heat. Even though these are modest by comparison, they still have an important impact in terms of the moisture budget. We shall return to this point later. Even though there are large local changes in these two surface heat fluxes, the globally-averaged values are quite small. The sensible-heat changes are 0.12, 0.78, 0.57 and 0.57 W m\(^{-2}\) in W05, W50, W80 and W100, respectively. The anomalous figures for the latent heat are 0.00, –0.28, –0.45 and –0.54, all less than 1 W m\(^{-2}\). Mitchell and Hills (1986) made a similar finding in their large-scale ice-removal experiment.

A consequence of these heat-flux anomalies is a disturbance to the atmospheric temperature structure. The latitude–height structure of the zonally-averaged temperature responses in experiments W05, W50, W80 and W100 are displayed in Fig. 4(a–d). (In this, and many of the plots shown in this paper, regions over which changes are significantly different from zero at the 95% confidence level in a univariate test—Simmonds 1981—are denoted by stippling.) In all cases there is a layer of significant warming above the sea ice in both hemispheres, the magnitude of which increases with \(f_w\). In the SH the maximum warming is centred at about 70°S. In W05 the tropospheric warming is relatively modest and its domain of significance does not extend above 700 hPa. In the other three experiments the region of significant warming extends through most of the troposphere between the pole and 50°S. In all cases, in the stratosphere, there is a significant cooling below 30 hPa in the southern polar area and, in all but W05, a region of warming further to the north. The significance and extent of the changes occurring in the Arctic increase with \(f_w\).

A similar presentation for the zonal component of the wind is given in Fig. 5. The only changes of consequence in the W05 experiment occur in the stratosphere. The anomalies in the other three simulations display a number of common features, including a weakening of the upper tropospheric westerlies between about 40 and 60°S and anomalous westerlies further south. The weakening extends all the way to the surface in the W50 and W100 experiments, whereas in W80 there is a small intensification at lower levels. The reasons for this different response in W80 are explored later.

The response of the sea-level pressure south of 30°S in the four experiments is given in Fig. 6. There is little change of significance in W05 (part (a)) and the only feature
Figure 2. Difference between the surface temperature of: (a) W50; (b) W100 and the control simulations. The contour interval is 5 K.
Figure 3. Difference between the (upward) surface sensible-heat flux of: (a) W50; (b) W100 and the control simulations. The contour interval is 40 W m$^{-2}$. The zero contour is accentuated and negative contours are dashed.
Figure 4. Difference between the zonally-averaged temperature of (a) W05; (b) W50; (c) W80; (d) W100 and the control simulations. The contour interval is 1 K. The zero contour is accentuated and negative contours are dashed. Regions of differences significant at the 95% confidence level are stippled.
Figure 5. Difference between the zonally-averaged eastward component of the wind: (a) W05; (b) W50; (c) W80; (d) W100 and the control simulations. The contour interval is 1 m s⁻¹. The zero contour is accentuated and negative contours are dashed. Regions of differences significant at the 95% confidence level are stippled.
Figure 6. Difference between the mean-sea-level pressure of: (a) W05; (b) W50; (c) W80; (d) W100 and the control simulations. The contour interval is 2 hPa. The zero contour is accentuated and negative contours are dashed. Regions of differences significant at the 95% confidence level are stippled.
Figure 6. Continued.
worth mentioning is the pressure increase to the north of the ice edge in the South Indian Ocean. Figure 6 (b) shows a greater response in the 50% open-water case, with significant pressure reduction in the Weddell Sea and over Oates Land. Just to the north of the ice edge there is a belt of increased pressure at about 60°S, while still further north there is a band of significant reduction in the South Indian Ocean. With increasing $f_w$, the area of significant change becomes larger. In Fig. 6 (c) and (d) (experiments W80 and W100) substantial reductions are simulated in the Weddell Sea and on most of the periphery of the continent. To the north of these areas there is a complementary increase in pressure, the significance of which is most marked in W50 and W100. Note that the pressure gradient differences near 65°S reduce fairly rapidly away from the surface, consistent with the thermal-wind differences implied by Fig. 4. The zonal wind differences in this belt are small at 900 hPa, as can be seen in Fig. 5.

We show two plots of the changes in the SH 500 hPa height field, as representative of the modification felt in the middle troposphere (Fig. 7). There is little change in W05 and the comparison of Fig. 7 (a) and Fig. 6 (a) reveals a rather barotropic structure. By contrast, as $f_w$ increases and the anomalous surface heat fluxes become greater, these high-latitude systems exhibit considerable tilt in the vertical. Figure 7 (b) shows that the 500 hPa height in W80 has increased significantly (by up to 100 m) over much of the continent and its surrounding sea ice, and anomalous highs tend to be located 30–60 degrees longitude to the east of surface lows (see Fig. 6 (c)).

To assist in the later discussion of the atmospheric moisture budgets in these simulations, we show the precipitation change induced in W50 (Fig. 8). The response is rather 'noisy' but, nonetheless, there are a number of regions which have undergone significant precipitation change. In general, precipitation appears to increase near the coast and on to the continent, while decreases are simulated over the northern parts of the sea ice and out over the nearby ocean.

4. DISCUSSION

(a) Changes over the sea ice and neighbouring regions in the southern hemisphere

As an aid to the interpretation of the above results we have calculated the mean over the SH sea ice area of some of the changes displayed above. We had remarked that there were changes of surface temperature over the SH sea ice and the continent and that the magnitudes appeared to increase as $f_w$ increased. To provide quantitative information on this we have calculated the surface temperature change averaged (in an area-weighted sense) over the Antarctic sea ice for the experiments. In Table 1 we present these differences, along with the average in the control simulation. For the purposes of later discussion we have also performed calculations, over the same domain, of the averages of surface air temperature (at the lowest model level ($\sigma = 0.991$)), surface sensible-heat and latent-heat fluxes, sea-level pressure and precipitation rate. The Table shows that over the sea ice domain the temperature change increases monotonically from 2.0 K in W05 to 17.5 K in W100. If the temperatures over the sea ice part of each grid box in the sea ice zone did not change, one would find a linear relation between this temperature increase and $f_w$. This is not so (e.g. in W50 we simulated an increase of 12.5 K, whereas a linear relationship would give 50% of 17.5 K, or 8.8 K) and the results emphasize the impact of nonlinearity.

We suggested that even small regions of open leads can have dramatic impacts on the atmosphere, but once certain features become 'saturated' there may be proportionally less impact from further increases in $f_w$. Ledley (1988) has conducted model experiments
Figure 7. Difference between 500 hPa height of: (a) W05; (b) W80 and the control simulations. The contour interval is 20 m. The zero contour is accentuated and negative contours are dashed. Regions of differences significant at the 95% confidence level are stippled.
EFFECT OF ANTARCTIC PACK ICE LEADS

TABLE 1. AVERAGE VALUES OF SURFACE TEMPERATURE, SURFACE AIR TEMPERATURE (AT LOWEST MODEL LEVEL ($\sigma = 0.991$)), SURFACE SENSIBLE-HEAT FLUX, SURFACE LATENT-HEAT FLUX, MEAN-SEA-LEVEL PRESSURE AND PRECIPITATION RATE. IN THE FIRST PART OF THE TABLE THE AVERAGES ARE PERFORMED OVER ALL ANTARCTIC SEA-ICE POINTS (AREA-WEIGHTED). THE AVERAGES IN THE SECOND PART ARE CALCULATED OVER A DOMAIN OF TWO GRID BOXES WIDE (ABOUT 7 DEGREES LATITUDE) IMMEDIATELY TO THE NORTH OF THE ANTARCTIC SEA-ICE EDGE. THE THIRD PART OF THE TABLE PRESENTS THE AVERAGES OVER A SIMILAR DOMAIN BUT LOCATED IMMEDIATELY TO THE SOUTH OF THE SEA ICE. THE UNITS ARE K, W m$^{-2}$, hPa AND mm d$^{-1}$.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Control (C)</th>
<th>W05-C</th>
<th>W50-C</th>
<th>W80-C</th>
<th>W100-C</th>
</tr>
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<td>2.0</td>
<td>12.5</td>
<td>16.0</td>
<td>17.5</td>
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<tr>
<td>S. air temp.</td>
<td>256.4</td>
<td>1.2</td>
<td>7.6</td>
<td>10.1</td>
<td>11.1</td>
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<tr>
<td>SH</td>
<td>-46.2</td>
<td>12.1</td>
<td>75.6</td>
<td>94.7</td>
<td>102.9</td>
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<tr>
<td>LH</td>
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<td>3.8</td>
<td>22.8</td>
<td>28.9</td>
<td>28.0</td>
</tr>
<tr>
<td>SLP</td>
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<td>-0.2</td>
<td>-1.2</td>
<td>-3.3</td>
<td>-2.3</td>
</tr>
<tr>
<td>Precip.</td>
<td>2.47</td>
<td>0.00</td>
<td>0.15</td>
<td>-0.05</td>
<td>-0.06</td>
</tr>
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</table>

Averages over Antarctic sea ice

<table>
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<tr>
<th>Parameter</th>
<th>Control (C)</th>
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<th>W50-C</th>
<th>W80-C</th>
<th>W100-C</th>
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<td>S. temp.</td>
<td>273.9</td>
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<td></td>
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<tr>
<td>S. air temp.</td>
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<td>0.2</td>
<td>1.8</td>
<td>2.5</td>
<td>2.7</td>
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<td>SLP</td>
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<td>0.6</td>
<td>1.8</td>
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<td>2.7</td>
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<tr>
<td>Precip.</td>
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<td>-0.03</td>
<td>-0.47</td>
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Averages to north of Antarctic sea ice

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<th>Parameter</th>
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<th>W80-C</th>
<th>W100-C</th>
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</thead>
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<td>5.4</td>
<td>7.7</td>
<td>9.3</td>
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<td>S. air temp.</td>
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<td>0.8</td>
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<td>Precip.</td>
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<td>0.12</td>
<td>0.46</td>
<td>0.57</td>
<td>0.80</td>
</tr>
</tbody>
</table>

Averages to south of Antarctic sea ice

To determine the impact of changing the lead fraction. She found a large impact on the annual mean surface air temperature structure in the polar regions, primarily due to changes in the sensible-heat flux between the surface and the atmosphere over the leads. Her Fig. 1(b) showed that increasing the minimum lead fraction from 0.0 to 4.3% produced, at 70°S, about 25% of the total change in simulated surface air temperature associated with the complete elimination of sea ice. (In the central Arctic the value was even greater, being about 40%.) Reference to Table 1 shows that in our GCM simulations the response over the winter Antarctic sea ice domain is not quite as marked, but is certainly nonlinear. For example, the surface air temperature increase is 11% of that of the ice-free run when the leads fraction is 5%. The analogous values are 68% in the 0.50 experiment and 91% in the 0.80 case.

The sensible-heat flux anomalies across the Antarctic sea ice zone also show a strong and monotonic dependence on $f_\omega$, ranging from 12.1 W m$^{-2}$ in W05 to 102.9 W m$^{-2}$ in W100. The nonlinear dependence on $f_\omega$ is even stronger than that displayed by the surface temperature response. Another feature of the sensible-heat response, to which we drew attention in Fig. 3, was the negative anomalies simulated to the north of the sea ice edge. (A number of the atmospheric variables considered here exhibited similar behaviour.) To quantify this compensation we have calculated the changes in sensible heat as above, but with the domain of averaging being the two (transform) gridpoints immediately to the north of the Antarctic sea ice edge. This belt around the hemisphere is about 7
degrees latitude wide and obviously has the same azonal structure as the ice edge. We have also performed calculations of the means of the other variables in the same way; these data are presented in the second part of Table 1. The third part of Table 1 contains similar information, but for it the averages were calculated over the belt covered by the first two gridpoints south of the ice (i.e. the coastal 7-degree latitude strip around the continent, including the permanent ice shelves). The Table reveals anomalous downward sensible-heat fluxes to the north of the ice edge in all runs. These also show a monotonic dependence on $f_w$ and assume values about one half of the upward fluxes over the sea ice region. By contrast, there is little change over coastal Antarctica.

The latent-heat flux changes over the pack-ice zone also show a monotonic sensitivity over the ice (except for W80, which is just 0.9 W m$^{-2}$ greater than the value in W100) and are numerically about one third of their sensible-heat counterparts. The latent heat also is subject to negative anomalies to the north of the edge. These changes in the spatial distribution of the evaporation affect the precipitation patterns at the high latitudes. It is clear from Table 1 that, despite the large increases in the mean evaporation over the southern ice zone, the precipitation changes over the same area in the various experiments are rather small and are in fact negative in W80 and W100. (A comparison of fluxes expressed in latent energy and moisture units can be made by noting that a latent-heat flux of 28.5 W m$^{-2}$ is equal to an evaporation of 1 mm d$^{-1}$. ) Hence the anomalous moisture derived from increased evaporation over the sea ice domain falls as precipitation elsewhere. The structure of Fig. 8 suggested significant precipitation increases were simulated near the coast and over the neighbouring parts of the continent, with reductions in the north, at least in W80. From Table 1 one can see that the evaporation has decreased over the open ocean points to the north by amounts which

![Figure 8](image.png)

Figure 8. Difference between the precipitation rate of W80 and control simulations. The contour interval is 0.5 mm d$^{-1}$. The zero contour is accentuated and negative contours are dashed. Regions of differences significant at the 95% confidence level are stippled.
are monotonically related to $f_w$. A similar statement is true for the precipitation. The magnitude of the latter reduction attains only about 70% of the former (in consistent units), implying that, in all experiments, the anomalous circulation is responsible for transporting moisture into the belt. Table 1 also reveals that the coastal 7 degrees latitude of Antarctica experience a modest reduction in evaporation and a large increase in precipitation. The magnitude of the differences increases with $f_w$, reaching 0.80 mm d$^{-1}$ in the ice-free case. Oglesby (1989) has conducted some GCM experiments of relevance to the question of precipitation changes over Antarctica in response to changes in SH sea ice. Consistent with our results he finds that replacing sea ice with open water increases rainfall over the continent. Oglesby suggests that, up to a point, increasing southern ocean sea surface temperature might enhance Antarctic glaciation and tie up more moisture in the ice cap, thus making a contribution to lowering global sea levels. However, too large a warming could decrease glaciation by increasing the ratio of rainfall to snowfall. These considerations have some relevance to the behaviour of sea level in climate change scenarios.

(b) Responses not monotonically related to $f_w$

We have seen that many of the changes to the variables in these experiments bear a clear monotonic relationship to the fraction of open water. This is true of those quantities which are directly affected by an increase in the fraction of open water in pack ice, such as the surface temperature over the Antarctic ice zone and the (closely associated) surface sensible-heat flux. However, for variables which are not unambiguously and directly related to the change in forcing, such monotonic behaviour need not be observed or expected. An example of such a variable is the zonal wind; its behaviour in these experiments is worth pursuing. In both W50 and W100 experiments the zonally-averaged surface westerlies were seen to weaken between 35 and 55 $^\circ$S, while they were simulated to intensify in most of this belt in the 'intermediate' simulation in W80 (Fig. 5). The explanation for this behaviour is associated with the regional response at the high southern latitudes. Reference to Fig. 6 shows that the largest responses in sea-level pressure simulated in the experiments were in the Weddell Sea region. This is not surprising as it is the region of the greatest latitudinal extent of sea ice and has been seen before to be very sensitive to ice removal (Simmonds and Dix 1986). Ackley (1981) stated that "... the Weddell Sea is looked on as a key area in understanding the southern hemisphere's climate and climate variations ...", and a number of studies (e.g. Schwerdtfeger 1979; Baines and Friedrich 1989) have pointed out some of the special characteristics of this region. The response of the sea-level pressure, even locally, is very complex because it is the result of many factors including topography, ocean temperatures, changes in heat fluxes and in baroclinicity. The strength of the anomalous lows in the Weddell Sea has a non-monotonic dependence on $f_w$, assuming magnitudes of 3, 7, 13 and 9 hPa in the four experiments, with the deepening in the W80 experiment much greater than might have been expected. We mentioned above some of the factors which go into determining the local Weddell Sea response of the sea-level pressure. Among these, the increase of the sensible-heat flux has an impact on lowering the pressure, and this flux has been seen to have a clear monotonic dependence (Table 1). However, in this region the baroclinicity also has a great impact on the behaviour of storms. Mechoso (1980) and others have pointed out that the region around Antarctica is baroclinically very active, with a complicated energy cascade. In our experiments the meridional surface temperature gradient over the Antarctic sea ice region is reduced with increasing $f_w$, and is completely eliminated in the W100 experiment.
To determine what effect this reduction has on baroclinic development in the lower troposphere we have compared the structures of the 700 hPa temperature fields in the experiments. Simplified baroclinic theory (e.g., Stone 1978) implies that in the 65–70 degree latitude belt (the location of the centre of the stationary anomalous lows in the experiments) a temperature gradient of about 1.5 degC per 5 degrees of latitude or equivalent is required for baroclinic instability. In the Weddell Sea the structure of the isotherms in the W80 and W100 experiments (Fig. 9 (a) and (b)) is rather different, as are the gradients. In the region of the anomalous low centres these gradients are 1.8 and 1.4, respectively, in the two runs. Hence, in the mean, the 700 hPa temperature field in the Weddell Sea is baroclinically supercritical in the W80 and subcritical in the W100 experiment. (The other two runs are even more supercritical.)

It is not immediately clear to what extent the arguments based on baroclinic stability are relevant in explaining the stationary response simulated in the experiments. Mole and James (1990) have discussed the differing behaviours of transient and stationary features and highlighted the role of downstream propagation. The mean wind in the Weddell Sea is very weak and hence the above suggested application of the theory may be reasonable. It should also be pointed out that the above arguments have to be modified somewhat owing to the considerable zonal asymmetry of the 700 hPa thermal field. Mole and James extended Stone's analysis to the case of zonally-varying forcing and pointed out that the local baroclinic processes are quite complex. Notwithstanding these points, however, the above discussion suggests that the strong anomalous low in the Weddell Sea in the W80 experiment is maintained by both large sensible-heat fluxes and baroclinic instability, while, on average, the latter is absent from the W100 experiment. Independent support for this chain of reasoning comes from the behaviour of the anomalous lows over and to the west of the Ross Sea area (another region of extensive ice coverage), which exhibit a monotonic strengthening with \( f_w \). In this vicinity the 700 hPa temperature gradients are supercritical on average in all experiments. This is evident for the W80 and W100 experiments in Fig. 9. The Figure also suggests the important role played by topography in these studies. Schwerdtfeger (1979) has also drawn attention to the very different mean atmospheric structures in the Weddell and Ross Sea areas and to some of the implications of this.

The above discussion suggests that simple monotonic dependence on \( f_w \) of variables which are determined by a number of influences need not be observed or expected. This then has implications for responses in mid latitudes forced by changes at the periphery of Antarctica. In particular, Fig. 6 makes clear the reasons for the rather different zonal-wind response in the W80 experiment between 40 and 60 °S compared to that in the W50 and W100 experiments.

(c) Comparison with earlier studies of the response in sea-level pressure in the ice-removal case

Comparison with the results of earlier studies allows us to assess the role of model resolution in sensitivity to sea ice changes. Simmonds and Dix (1987) (SD87) reported an experiment with a 15-wave version of this model in which all Antarctic sea ice was removed. The 21-wave version used here produces a more accurate climatology, and this improvement may be expected to change the sensitivities. Comparison of Fig. 5 in SD87 and Fig. 6(d) here shows that, after the removal of sea ice, the changes in sea-level pressure at high southern latitudes are statistically significant over a much greater area in the higher-resolution model. The magnitudes of the changes are greater, and tend to be more organized spatially. While the responses in the two experiments differ in a
Figure 9. Mean temperature field at 700 hPa in: (a) W80; and (b) W100 simulations. The contour interval is 1 K.
number of specific details, there are a number of features in common, such as the reduction in pressure over the Weddell Sea, and the anomalous ridges to the north of the sea ice edge in the South Atlantic and to the south of Australia. The zonally-averaged temperature responses (Fig. 3 in SD87, Fig. 4 (d) here) differ to some degree above about 700 hPa in mid to low latitudes but are very similar in the southern polar region. A similar comparison for the zonal component of the wind (Fig. 4 (SD87) and Fig. 5 (d)) reveals that the higher-resolution model replaces stratospheric westerly anomalies in the subtropics with easterlies, but the changes in the troposphere are very similar.

We can also make some comparison with the ice-removal experiment of Mitchell and Senior (1989) (MS). Although these authors removed Antarctic sea ice only equatorward of 67.5°S the pattern of change in sea-level pressure we simulate in the W100 experiment (Fig. 6 (d)) shows considerable resemblance to their Fig. 12. Both exhibit significant reductions in the Weddell Sea and further to the west, and also in the vicinity of Oates Land. In addition our model also simulates large, significant reductions in the Ross and southern Weddell Sea where ice had been removed but where the sea ice specification had not been changed in MS. In the present experiment there is a belt of pressure increase centred to the north of 60°S, attaining significance south of Africa, south of Tasmania and to the west of the Drake Passage. MS display pressure increases at a latitude a little further north, but none of these are significant, except over an area to the east of South America.

(d) Changes in the transfer coefficients and the potential impact of feedbacks not included here

There are a number of feedback processes which are not included in this study but which may have the potential to magnify effects or influence remote locations. In our study the sea ice is prescribed as fixed whereas in reality it is known to respond to changes in the wind stress. The large anomalies in sea-level pressure simulated over the Antarctic sea ice suggest significant changes taking place. Gordon and Taylor (1975) have suggested that the curl of the wind stress continuously generates open water within ice fields, and there is obviously the potential for feedback which the specification of unchanging sea ice precludes.

We have discussed, in Simmonds (1985), the dependence of the transfer coefficients on wind speed, stability and surface type. The zonal average of the drag coefficient, \( C_D \), in the four experiments and the control is displayed in Fig. 10(a). In the control it is about \( 1.0 \times 10^{-3} \) in mid latitudes and decreases slowly towards the sea ice edge. Because of the increased roughness of sea ice compared to that of open water the drag coefficient rises to over \( 2.0 \times 10^{-3} \) over the sea ice and drops back to about \( 1.0 \times 10^{-3} \) over the Antarctic continent. The drag coefficient displays this last behaviour because, even though the roughness over the continent is specified as the same as over the sea ice, the atmosphere over the continent is very stable, particularly at higher elevations (Phillpot and Zillman 1970). The structure of the curves is rather similar to those obtained by Mitchell and Senior (1989) when they removed Antarctic sea ice or artificially reduced the drag coefficient over part of the sea ice (their Fig. 13). They concluded that the change in surface roughness contributed substantially to the response to reduced (continuous) sea ice extents.

As areas of leads are opened up in the pack ice there are two competing influences in changing \( C_D \). Firstly, as warm water is exposed beneath low atmospheric temperatures the stability dependence in the drag coefficient will tend to increase \( C_D \). On the other hand, in the model the roughness length of the exposed water is considerably less than that of the sea ice, so this will tend to diminish the drag coefficient. It will be clear from
Fig. 10 (a) that it is this second influence which dominates. In all cases \( C_D \) increases with distance south over the sea ice zone (owing to the fact that more of the latitude circle is covered by sea ice), but not at the same rate as in the control, and is smaller at all latitudes over the sea ice. Also over most of the sea ice the drag coefficient is seen to have a monotonic dependence on \( f_w \). By contrast, the drag coefficients over the continent show the reverse of this last effect owing to the progressive warming over the continent as \( f_w \) increases. The roughness is the same in all cases, but the warming of the Antarctic surface makes the lower levels less stable in this region.

Andreas and Murphy (1986), among others, have studied the behaviour of bulk transfer coefficients over sea ice with leads. This change of drag coefficient with sea ice concentration is consistent with the observations and atmospheric boundary-layer model results of Burns (1990). She found that the drag coefficient decreased almost linearly with \( f_w \), and under 'rough' sea ice conditions (floe sizes between 5 and 50m) the drag coefficient decreased by about a factor of two as \( f_w \) was changed from 0.0 (solid ice) to 1.0 (open water). The data presented in Fig. 10 (a) show our modelled \( C_D \) to exhibit a very similar behaviour. The zonal averages presented here must be treated with a little caution because at certain latitudes the zonal average contains other than sea-ice points. To allow a better appreciation of changes over the southern sea ice region (as defined in the control simulation) we have calculated the mean drag coefficient over this region in the various experiments. The mean of \( C_D \) is \( 2.19 \times 10^{-3} \) in the control and is reduced to 2.14, 1.59, 1.17 and \( 0.83 \times 10^{-3} \) in the four anomaly runs. The reductions effected are rather similar to those found by Burns (1990).

In the real climate system the momentum flux over oceans forces the ocean circulation, which in turn influences the distribution of heat and ocean temperature (Budd 1986). A weakening of the mid-latitude westerlies in the SH, similar to that simulated in the W50 and W100 experiments, might be expected to change the strength of the large anticyclonic ocean gyres, which are fundamentally responsible, through advective and upwelling effects, for maintaining the east–west temperature gradient in the various ocean basins. It is known that regional-scale climate is quite sensitive to this gradient.
(e.g. Simmonds et al. (1989)) and that changes to it can induce large effects remotely. In our case we simply prescribe the sea surface temperature; hence the ocean is not able to respond in the manner outlined above. However, the high-latitude changes may be able to influence regions further removed than those considered here, through the medium of oceanic response. (Even without the transport mechanisms afforded by the ocean it is becoming clear, e.g. James (1988), that the remote response to high-latitude winter forcing is more likely in the southern than in the northern hemisphere.) There is much evidence to suggest that anomalies at high southern latitudes do have impacts of this nature. For example, van Loon and Shea (1985) have found that the (southern) winter before a ‘Warm Event’ in the tropical east Pacific Ocean is marked by weaker surface westerlies in the central and eastern Pacific between 30 and 60°S (and westerly anomalies to the north of this belt). It may be that changes in the atmospheric flow induce changes in the oceanic circulation, whose effects are felt some time later and at remote locations. It has been known for some time that the Pacific Ocean gyre shows significant oscillations (Wyrtki and Wenzel 1984).

The meridional distribution of the zonal-average surface wind stress in the control and the four experiments is displayed in Fig. 10(b). North of the sea ice edge (about 60°S) there is a small reduction in the stress in the W05 and greater reductions in the W50 and W100 experiments. The stress in W80 undergoes much less change than these two. Differences from control are seen as far north as 40°S. From Fig. 10(a) it is seen that there is very little change in \( C_D \) over this latitude domain, so any changes in the stress comes about as a result of changes in the wind speed. This is consistent with the surface wind changes displayed in Fig. 5. Over most of the sea ice zone (which is represented in the zonal averages between 60 and 75°S) there is a reduction of the westward stress in all cases. The fact that significant positive correlations have been found between Antarctic sea ice extent and the Southern Oscillation Index when ice area leads the Index (Chiu 1983) suggests that Antarctic sea ice is a forcing mechanism of some importance in this sequence: this is consistent with the picture presented above.

5. CONCLUDING REMARKS

We have conducted four experiments to assess the sensitivity of SH circulation to changes in the fraction of open water in the pack ice. This fraction was set to 5, 50, 80 and 100% in the experiments which were compared with the control with 0%. As expected the mean surface temperatures and the surface atmospheric temperatures over the sea ice increased with increasing water fraction, the largest changes being simulated adjacent to the coast.

Large changes in the surface heat fluxes, particularly sensible-heat fluxes, accompanied the presence of leads. Averaged over the sea ice area the sensible-heat flux changed from −46.2 W m\(^{-2}\) in the control to −34.0 W m\(^{-2}\) in the 5% open-water case. This change of 12.2 W m\(^{-2}\) is already 12% of the sensible-heat flux change between the ice-free case and the control. The changes in latent heat corresponded to about 1 mm d\(^{-1}\) over the southern ice in most of the experiments. This anomalous moisture falls as precipitation over the southern part of the ice and on the coastal parts of Antarctica and has some implications for changes in sea level. There are reductions in precipitation over the northern part of the sea ice and the ocean beyond the ice edge.

Significant warming is simulated over and in the vicinity of areas in which leads were considered and in all cases there is a cooling in the southern polar stratosphere. There are virtually no significant changes to the zonally-averaged winds in the troposphere in W05. In W50 and W100 there are anomalous easterlies between about 40 and 60°S with
westerly anomalies further to the south. The structure in W80 is similar except that the band of tropospheric easterly anomalies does not extend down to the surface.

We have found that for the atmospheric quantities directly affected by the opening up of the sea ice there is a clear and monotonic dependence of the anomalies on the open-water fraction. For secondary quantities, such as the sea-level pressure or precipitation change, this is neither observed nor expected. As a case in point we found in all cases the greatest pressure reductions were simulated over the Weddell Sea, but that the anomalous low was deeper in the $f_w = 0.8$ case than for $f_w = 1.0$. The analysis of this revealed the complexity of the response in sea-level pressure, depending as it does on a number of interacting factors including topography, ocean temperatures, changes in heat fluxes and in baroclinicity.

We have touched on the question of the potential impact of feedback processes which were not included in our study. A large reduction of the zonally-averaged eastward surface stress was simulated between 40 and 60°S, in two of the studies. These are similar to some of the precursors to ‘Warm Events’, which in turn have an influence on climate fluctuations on a global scale. The model results appeared to be consistent with published analyses of high-latitude forcing.

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