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A review of sea-ice weather relationships in the Southern Hemisphere

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ABSTRACT Within the last decade data on sea ice from satellite coverage have become available for the Southern Hemisphere. These are sufficiently detailed spatially and temporally to permit some examination of the "mean" and "anomaly" conditions of sea ice extent and the relative influence of thermodynamic and dynamic driving forces in the advance and retreat of the sea ice cover. The data record is reviewed with some consideration given to the different mechanisms of ice advection by wind forcing, thermodynamic growth, and ocean mixing. These mechanisms control the ice edge around Antarctica and lead to the characteristic advance-retreat relationships for the Weddell Sea, East Antarctica, and the Ross Sea. Recent statistical and function (EOF) analyses have shown two primary areas of higher annual variation of sea ice conditions which are presumed to be more sensitive to variations in forcing fields, probably of dynamic (winds and currents) rather than thermodynamic (temperature) origin. While the ice data indicating response to changes in the forcing fields are becoming more available, the actual mechanisms that caused the sea ice changes of the past few years are somewhat speculative because of the lack of both atmospheric and oceanic data on similarly resolved scales to the ice data. This lack of high confidence data also presents some difficulties in examining the other half of a possible sea-ice atmosphere positive feedback mechanism. Here it is postulated that atmospheric forcing of the sea ice system causes changes in air-sea energy transfers that then drive the atmosphere to its own anomaly condition. Some previous studies have indicated at least qualitative support for an atmospheric response to the ice changes. Examples of these are zonal wind response (circulation vigour), air temperature changes, storm track, and generation and lifetimes of cyclones. Further correlations that may define the mechanism of sea ice response to the forcing fields and supply stronger evidence of weather and climate responses to ice variations, may be available by analysis of the Global Weather Experiment drifting buoy data obtained during 1979. Drifting buoys fill the data gaps in the spatial and temporal forcing fields that exist in the oceanic regions of the Southern Hemisphere and buoys implanted on sea ice help resolve the detailed sea ice response while coincidently measuring

the forcing fields. Some sea ice buoy records will be briefly examined to indicate these potential correlations. These data sets will also provide sufficient data to test numerical models of ice cover changes that include both dynamic and thermodynamic forcing of the ice cover.

INTRODUCTION

Since the beginning of the satellite era (1960s) mapping of the extent and variability of the Antarctic ice cover has been improved. The frequency of coverage and the type of sensor have markedly improved our estimates of Antarctic sea ice extent which were formerly based primarily on occasional ship observations (e.g. Heap, 1964), aircraft observations from coastal stations (Treshnikov, 1967), island station reports (Fletcher, 1969), and inferred extension of these observations by "climatic estimates" of the ice edge in other areas which were assumed to correspond with the observed areas. As a result, comparisons between mean ice extents of various publications indicated consistently higher or lower ice extents in various regions depending on the data used to compile the mean values (e.g. Pease's, 1975, comparison of Soviet Atlas and US Hydrographic Office Data, also Parkinson & Washington, 1979). Although the new data record is not long in climatic terms (partial coverage 1967-1972, complete coverage 1973-1979), the regions of major variability in ice coverage have been delineated and some assessment can be made of the correlations between sea ice changes and weather around Antarctica. We first review the sea ice record and indicate the differences in sea ice advance-retreat mechanisms in various regions around Antarctica. Statistical analyses while still on a relatively small sample offer a quantification of the observed variability for objective correlation with weather phenomena. We then review the possible magnitude that the presence or absence of sea ice offers in influencing atmospheric heating and the consequent response of the atmosphere to these positive or negative heat sources. Correlations of sea ice changes with observed weather features have been attempted but are presently viewed with some caution because of the lack of sufficient weather observing stations where the sea ice changes are occurring to provide a solid quantitative basis for comparison. With this caution, these semi-quantitative correlations indicate a substantial interaction between ice and weather processes. The acquisition of longer records of sea ice variability along with the 1979 drifting buoy records of the Global Weather Experiment, should offer a firm basis for these correlations.

SOUTHERN HEMISPHERE SEA ICE DATA IN THE SATELLITE ERA (1967-PRESENT

Ice mapping
The mapping of the ice cover around Antarctica has been compiled primarily by the US Navy's Fleet Weather Facility's Ice

Forecasting Group, now the Navy-NOAA Joint Ice Center. Satellite coverage of three types is used, visible and infrared images obtained by the NOAA and DMSP satellites with the scanning radiometer of the NOAA satellites resolving features of about 4 km and the VHRR sensors and DMSP satellites resolving features of the order of 1.5-1 km, both under relatively clear conditions.

Passive microwave emission is mapped by the NASA NIMBUS V and VI satellites at about 32 km resolution (Zwally & Gloersen, 1977) but without the clear sky constraint. These satellite mappings are supplemented by available ship and aircraft observations of ice edge position, ice concentration and ice types although these observations are limited to the summer period when ships and aircraft are operating in the Antarctic region. The passive microwave images from two sensors, the Electrically Scanning Microwave Radiometer (ESMR) on NIMBUS V and the Scanning Multi-channel Microwave Radiometer (SMMR) on NIMBUS VI have been especially useful since they sense emissivity contrasts of the surface, and are not generally affected by clouds (although rain will affect the received signal). In the high cloud regions of

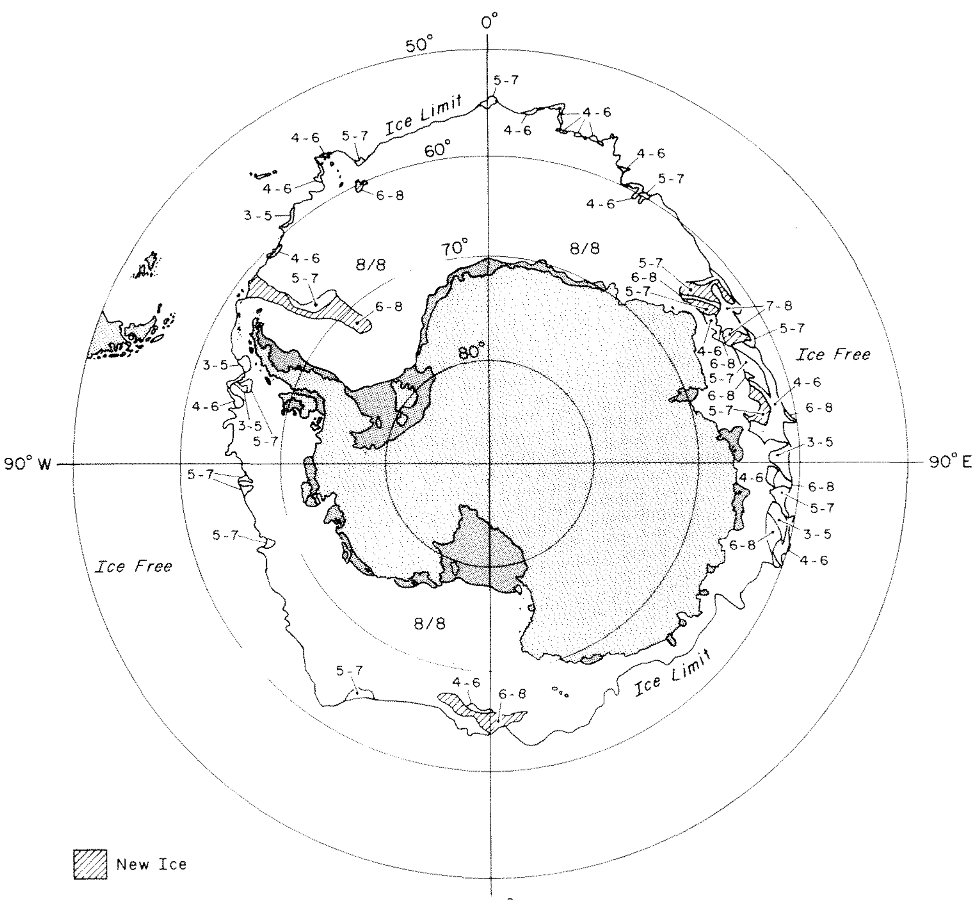

Fig. 1 Ice limit and concentration during Antarctic winter prepared from the 9 August 1979 ice chart (Navy-NOAA Joint Ice Center, Fle Wea Fac, Suitland, Maryland, USA). Ice concentration is expressed in fractional eighths of coverage.

the circumpolar trough that coincides generally with the position of the ice edge, the 80-90% average cloudiness seen there (van Loon, 1972a) gives some uncertainty of the ice edge position obtained by visible and infrared sensors but can be resolved by the microwave. Lack of daylight also affects the visible light radiometers during the winter period. Zwally & Gloersen (1977) give some examples of the microwave data while Gloersen *et al.* (1978) describe the determination of received signal in terms of the surface emissivity (dependent on surface temperature, ice type, and ice type concentration) and the atmospheric transmissivity (dependent on water content and temperature) that by suitable comparisons of frequency and polarization dependencies can be used to resolve ice types and concentrations within error limits of about 5-10%. The relationship for sensed brightness temperature T_B, for a single ice type only, is given by

$$T_B = (\varepsilon_i T_o)C + (\varepsilon_w T_w + A)(1-C) \qquad (1)$$

where ε_i, T_o, C are the ice emissivity, temperature and concentration; ε_w and T_w are the water emissivity and temperature; and A is the atmospheric water vapour contribution. Figure 1 shows a winter ice map derived primarily from the microwave satellite coverage. These maps are produced once a week throughout the year with the period of observation (usually 3-4 days of data) also indicated.

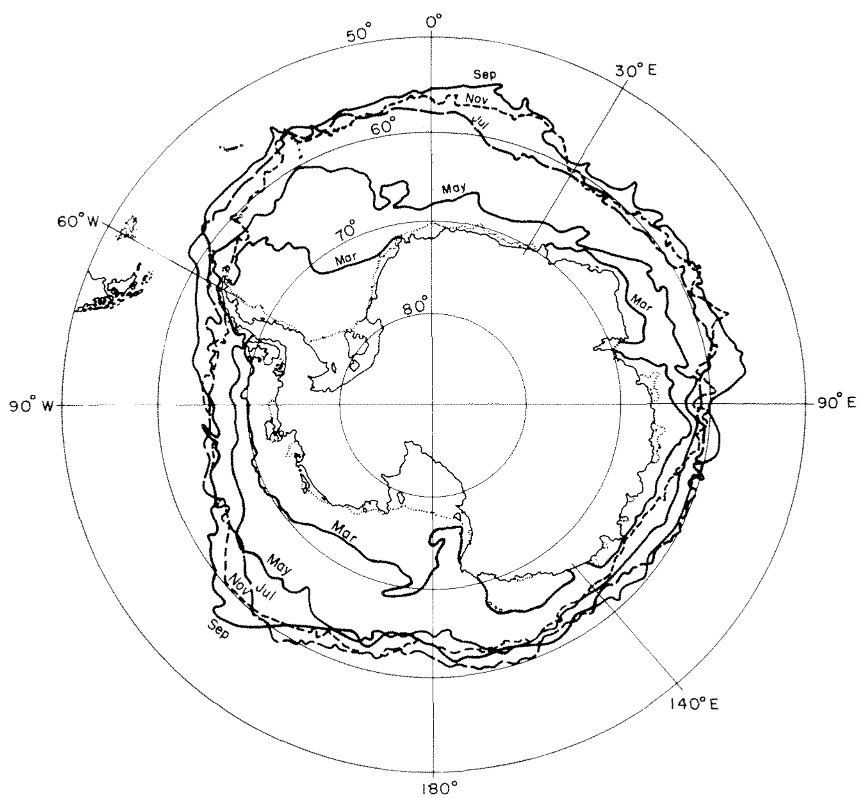

Fig. 2 Area bounded by the pack ice edge in 1974.

Seasonal and annual sea ice variability

In comparison with the more familiar Arctic pack ice, the greatest contrast arises because of the unbound nature of the Antarctic pack ice with the world ocean while the Arctic pack ice contact with warmer oceans is restricted to a few narrow passages. As a result, the Antarctic pack ice exhibits a greater annual variation in extent (Ackley & Keliher, 1976) with a greater than 5 - fold increase in area bounded by the ice edge at maximum extent compared to minimum. Figure 2 shows the area bounded by the pack ice edge through the annual cycle in 1974. Figure 3 shows the limits of pack ice for the years 1971-1976 and it can be seen that any given meridian will show variability in maximum pack ice extent. We note that the maximum and minimums vary regionally from year to year so the outer and inner boundaries for maximum and minimum extent did not occur simultaneously around the continent. We examine this more closely in the next section. Kukla *et al.* (1977) have compiled total snow and ice coverage data for various years. For the period 1972-1976, Kukla *et al.* (1977) conclude that the annual average for the period roughly parallels the mean extent of pack ice in November for the same period. The November curves for the pre-microwave years 1967-1972 are, therefore, considered indicative of the average annual variations of ice cover (Kukla *et al.*, 1977).

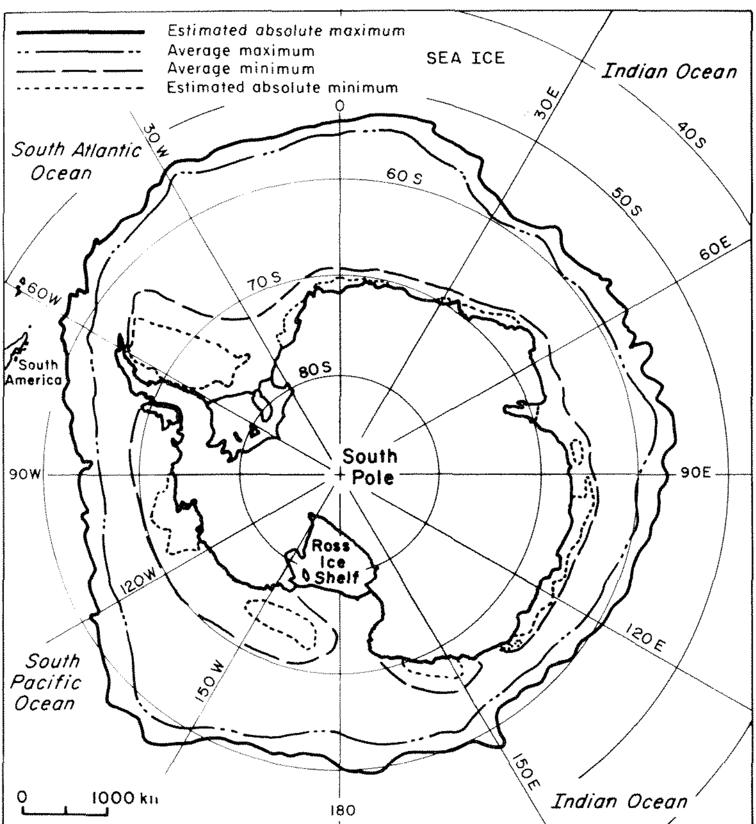

Fig. 3 Pack ice limits 1971-1976 (after *CIA Atlas of the Polar Regions*).

Looking at the November extents only, indicates an increase in ice coverage from 1966-1972 while ice coverage decreased from 1974-1976 (Fig. 4). The total range of the annual average variability is, however, of the order of 2 x 10^6 km^2 for 1972-1976 while the mean November range is nearly double this value (∿4 x 10^6 km^2) for the same period. The November values also show a decrease in area in 1974 compared to 1973, while the annual average for 1973-1974 shows an increase in 1974. The higher variability and the discrepancy may indicate that the November ice data, while the closest to the maximum ice extent available for the earlier years may not be a particularly good indicator of the Southern Hemisphere ice extent on an annual basis and probably should be used only cautiously to extend the ice record back from the years of higher quality mapping (1972 on). One reason for the enhanced range of the November variability may be because ice anomalies are apparently regionally confined (Ackley & Keliher, 1976; Budd, 1975) but the maximum extent varies in phase around the continent. Figure 5 (Ackley, 1979) indicates this phase variability by comparing the time series of the ice edge latitude for particular meridians. Here the maximum northward extension of the ice edge occurs during

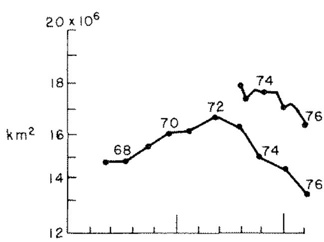

Fig. 4 Annual average and mean November Southern Hemisphere ice extents 1966-1972 (after Kukla *et al.*, 1977).

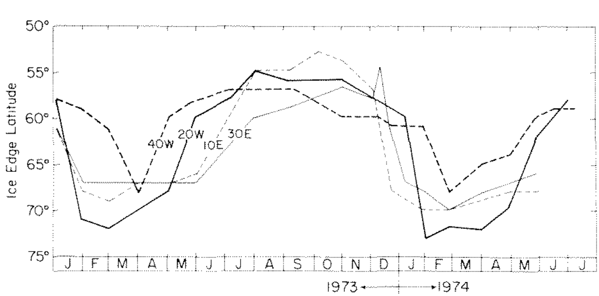

Fig. 5 1973-1974 time series of ice edge latitude for the 40 °W, 20 °W, 10 °E, and 30 °E meridians (after Ackley, 1979).

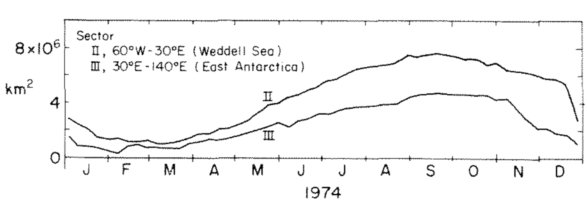

Fig. 6 Ice area in the Weddell (60 °W-30 °E) and East Antarctica (30 °E-140 °E) for 1974.

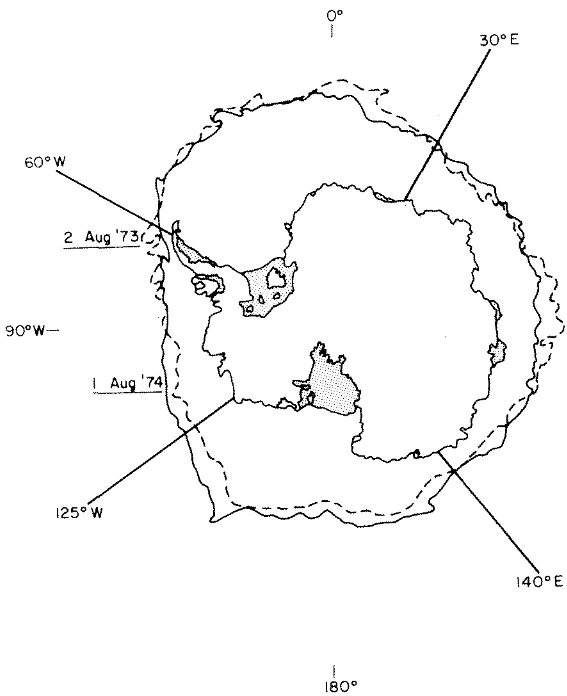

Fig. 7 Ice extents for 1 August in both 1973 and 1974. Less ice coverage during 1974 in the Weddell is roughly compensated by an increase in the Ross Sea sector.

July-August at the 40°W and 20°W meridians, September-October at the 10°E meridian and November-early December at the 30°E meridian. Figure 6 compares the ice area in the Weddell (60°W-30°E) and East Antarctica (30°E-140°E) sectors for 1974 and we see that the highest retreat rate of the East Antarctic ice occurs during November. A change in the timing of this retreat by one or two weeks could, therefore, sharply change the estimates of annual variability based on this month's data only. Picking a particular month may, therefore, enhance those regions that are either maximum in ice extent or of higher variability at that time over regions that may reach their maximum earlier or be of lesser variability at that time. We are more confident for those years (1972-1976) where the data for the entire year are available and this (Fig. 4) generally indicates lesser ice coverage in 1975-1976 than in 1973-1974 by about 1-2 x 10^6 km^2. Zwally *et al.* (1979) point out most of this decrease is due to a large polynya (open water area within the pack ice zone) which formed in the eastern Weddell Sea region in 1974-1976 but not in 1973. Aside from the polynya formation there is a tendency for increased ice extent in one region to be balanced by decrease in another, leading to relatively small changes in area bounded by the outer ice edge from year to year over this short data record. Figure 7 (Ackley & Keliher, 1976) shows the first week of August ice maps for 1973 and 1974 where lesser 1974 ice coverage in the Weddell (60°W-30°E) is about balanced by increased ice coverage in the Ross Sea region at this same time.

Figure 8 shows the ice variability in various sectors as well as totals for all areas around Antarctica (after Zwally *et al.*,

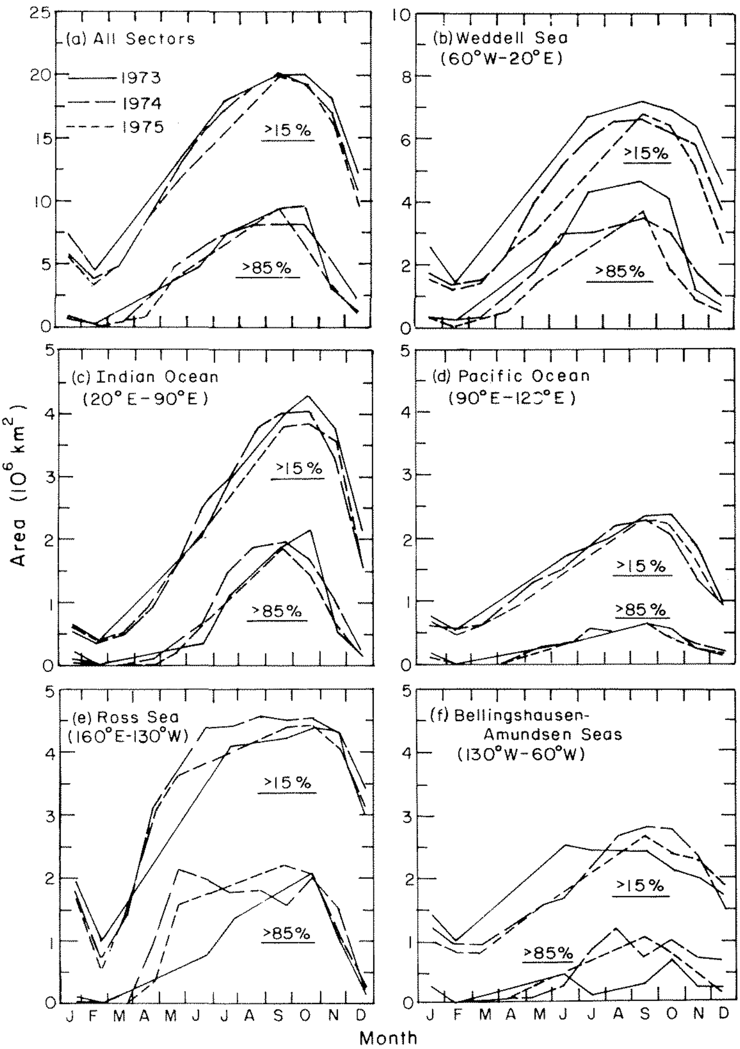

Fig. 8 Total ice and ice area by sectors around Antarctica (1973, 1974, 1975). The areas of > 15% ice concentration and > 85% ice concentration are indicated (after Zwally *et al.*, 1979).

1979). Two sets of curves are shown for each case, the first for the total area of ice greater than 15% concentration and the second for the area of ice of greater than 85% ice concentration. At the maximum and at most other times the lower ice concentration (<85%) accounts for over half the total "icy" area. Average concentrations are not available from these data alone, but other estimates using microwave data (Ackley & Keliher, 1976, low value, and Zwally & Gloersen, 1977, high value) indicate the probable range of average concentration is 65-80% ice within the >15% ice boundary defining the ice edge. These values, even the highest ones, are in sharp contrast to winter Arctic Ocean concentration estimates of well over 90%, at times approaching 100% ice cover (Gloersen *et al.*, 1978). Unfortunately, these values estimated from digitized microwave data also sharply disagree with Fle Wea Fac map coverage which currently list at least three-quarters of the Antarctic winter ice at 8/8 fractional coverage which is usually interpreted as at most no more than 6% open water. Clearly more experiments and comparative data are needed to validate these microwave estimates of ice concentration, a possibility with the assistance of drifting buoys on sea ice. Some consequences of varying the ice concentration over the range indicated are examined later in the computations of atmosphere-ocean heat exchange in the presence of pack ice.

Seasonal and inter-annual variation by region

The regions of least inter-annual variability are those bordering East Antarctica (20°E-160°E) ((c) and (d) of Fig. 8). The most notable differences from other regions are the relatively late peaking of the maximum ice extent tending toward October rather than the earlier mostly September dates of the other sectors. Figure 8(f) for the Bellingshausen-Amundsen sector displays about the least annual range with a maximum to minimum areal ratio (summer *vs.* winter) of about 2.5 compared to 4 for the Pacific Ocean sector, roughly 6-8 for all other sectors and about 5-6 for the total around Antarctica. One characteristic of the Bellingshausen-Amundsen Seas sector is the relatively early maximum in June 1973 relative to later years. Zwally *et al.* (1979) speculate that this may be an unusually fast advance related to eastward transport. Ackley & Keliher (1976) on the other hand, suggest later lesser ice coverage during July-September is related to an anomalous retreat during the June-July period apparently by an atmospheric blocking event. The retreat was well correlated with later cyclone activity, and is examined further in the Section on Effects of Sea Ice on Weather. The higher advance rate seen during early winter 1973 could be either the transport mechanism, since it correlates well with a decreased advance rate west of the region in the Ross Sea (Zwally *et al.*, 1979), or a combination of this and enhanced early winter growth caused by southerly winds and lower temperatures than usual.

Of most interest are the Weddell and Ross Sea sectors (Fig. 8(b) and (e)) since, as mentioned earlier, the inter-annual variability around all of Antarctica in total coverage is significantly influenced (about half) by the presence or absence of the polynya in the Weddell. Ice extent variations between the Ross Sea and Weddell (Fig. 7) also cancel somewhat to give a small overall change in total ice but account for considerable spatial variation between 1973 and 1974.

In ice characteristics, the Weddell Sea contains some unique features relative to the rest of Antarctic sea ice. It contains the major area of pack ice persisting through the summer period (Figs 3 and 6), and as pointed out above, it accounts for the highest amount of the total inter-annual variability around Antarctica.

Zwally *et al.* (1979) note that (Fig. 8(b)) the Weddell has a larger proportion, about half, of highly concentrated ice, than is present in other sectors of Antarctica. As Ackley (1979), Schwerdtfeger (1979), and Zwally *et al.* (1979) indicate, advection dominates the Weddell pack ice, influenced by the topographic boundary of the Antarctic Peninsula. While the maximum pack ice period is in September for the total Weddell region, some indications of the differences from other regions were shown in Fig. 5, where the timing of the maximum northward ice extension is shown to vary from west to east as the season advances. The rate of advance is also dominated by the early part of the season. In Fig. 6, we compare more closely the areas of 60°W-30°E (note that Zwally *et al.* use 60°W-20°E) and East Antarctica. Here it is seen the dominant ice advance period is March-June at a high rate while East Antarctica is at a lower rate that is nearly constant from March-September. Ackley (1979) and Sobey (1975) indicate this early season advance is northeasterly (through May) followed by a slower ice edge northern advance from July-September. In comparison, East Antarctica pack ice advances generally northward only throughout the fall-winter period (Ackley, 1979).

The Ross Sea, as the Weddell, is also characterized by higher variability and predominantly strong early season advance. Figure 8(e) compares the 3 years and we note the high rate for 1974 and 1975 in the March-June period and a levelling off of the ice area after this period. The advance during 1973 differed considerably from this with a more gradual rate of advance extending over a longer period. Zwally *et al.* (1979) mention the possibility of ice transport from the Ross Sea to the Bellingshausen-Amundsen Sea sector since the lower, early season, Ross Sea advance in 1973, coincides with a stronger than "normal" advance in the other region. The total area changes are provocative in that they nearly correspond, i.e. the "shortage" in the Ross Sea sector of about 1×10^6 km^2 of ice is nearly balanced by about the same ice area of "excess" in the Bellingshausen-Amundsen. One of the shortcomings of ice mapping, however, is that differences in ice edge position by physical movement or transport of ice from one area to another cannot be differentiated from freezing in one area and melting in another or even, as may be the case here, enhanced growth compared to reduced growth, i.e. dominantly thermodynamic rather than dynamic controls. Given the latitudes and usual wind directions at the ice edge during early season, one might expect a little movement of ice between the Ross Sea and Bellingshausen-Amundsen Seas but in the opposite direction to that indicated since the dominant winds at the higher latitudes in the early season are the polar easterlies which would move ice from east to west rather than the trend implies here. DeRycke (1973) in a study of ice motion using VHRR satellite imagery for December-March 1973 in the eastern Ross Sea indicates the movement is highly variable but with a general direction suggestive of westward movement attributed to the current (East Wind Drift) which would be opposite to the trend observed in the ice mapping as shown in Fig. 8(e) and (f). The drifts of trapped ships especially the *Belgica* in 1897-1898 (Arctowski, 1908) is also in agreement with this variable but generally westward rather than eastward movement for the Bellingshausen-Amundsen Seas. We would tentatively conclude that the anomalies in 1973 giving enhanced ice advance for the Bellingshausen-Amundsen and reduced ice advance for the Ross Sea are not coupled through ice removal from the Ross Sea and transport to the Bellinghausen-Amundsen. Some enhancement of ice growth through lower temperatures or greater dominance of southerly winds (or both) may be more applicable to the advance. As well, the anomalous retreat of the Bellingshausen-Amundsen pack in July 1973 appears atmospherically triggered and related to ice growth and melt and, therefore, a similar advance mechanism is more consistent with this retreat than transport to or from other regions, which appears generally weak for this region.

Possible mechanisms of ice edge control

As indicated in the previous section, several mechanisms may be operative in advancing and retreating the ice edge since it has different character in timing, amount and variability around Antarctica. A general total picture, however, may be achieved by examining the model of Gordon & Taylor (1975) which offers an overview as to the dominant mechanism and with some additional regional detail provides an acceptable picture of the advance mechanism for the Antarctic pack ice. They note that the seasonal variation of sea ice has important characteristics that influence the thermohaline structure of the ocean and that heat and moisture exchange affects the atmosphere. One assumption is that new ice is added to the outer fringe in winter and removed in summer (i.e. primarily a stationary thermodynamic effect). Their hypothesis on the other hand is that much of the ice growth and retreat is related to the wind stress. In Antarctica, the curl of the wind stress is

$$\text{curl}\ \tau = \frac{\partial \tau_x}{\partial y} - \frac{\partial \tau_y}{\partial x} \tag{2}$$

where τ_x, τ_y are the components of the wind stress acting in the east and north directions.

$$\tau_x = \rho_{air}\, C_D\, W_x\, (W_x^2 + W_y^2)^{\frac{1}{2}} \tag{3a}$$

$$\tau_y = \rho_{air}\, C_D\, W_y\, (W_x^2 + W_y^2)^{\frac{1}{2}} \tag{3b}$$

where ρ_{air} = density of air, C_D = drag coefficient, and W_x, W_y are wind velocity components east and north respectively.

The curl of the wind stress induces a general Ekman divergence of surface water and sea ice (assuming that the ice diverges with a divergent Ekman field), continuously generating open water regions within the ice fields. In winter, the open water freezes over and the ice field expands, extending northward as a function of the total divergence. The northern extent of the ice may be determined by an inability of the local heat balance to continually freeze over the open water generated by the Ekman divergence. In summer, the divergence would act to rapidly decay the ice cover since the generation of open water would provide a heat source due to its reduced albedo and laterally melt the existing ice. Figure 9 from Gordon & Taylor (1975) shows the annual fluctuation of sea ice area during advance and the calculated area due to Ekman divergence with two values of drag coefficient. The recent satellite evidence (Zwally *et al.*, 1979; Ackley &

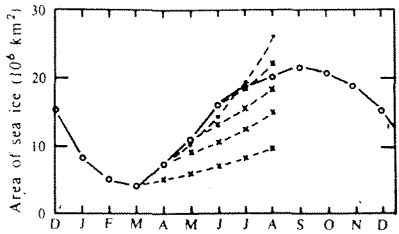

Fig. 9 (○) annual fluctuation of Antarctic sea ice (solid curve). The other curves show the sea ice area during advance based on Ekman divergence with a drag coefficient of 1.3×10^{-3} (×) or 2.6×10^{-3} (●). Each curve begins with the month for which the observed ice cover is used to initiate the calculation (after Gordon & Taylor, 1975). (Copyright 1975, by the American Association for the Advancement of Science.)

Keliher, 1976) is supportive of this mechanism, for example Fig. 8 (from Zwally *et al.*, 1979) which indicates that over half the total ice area at any given time (Fig. 8(a)) is less than 85% ice covered, i.e. extensive open water fields exist within the sea ice cover throughout the year.

A repeat of an earlier caution is that the interpretive ice mapping based on the satellite imagery indicates a more compact ice cover than the computed open water percentage based directly on the same digital satellite data. One difficulty is that estimates of both emissivity and surface temperature are necessary to compute the ice concentration directly. The differences may, however, be semantic since the rapid freezing-over of open water areas during winter may bias the ice mappers to indicate higher concentrations since on the time scale of navigational uses, any particular open water area may freeze so rapidly that it is basically not navigable. On earlier maps (prior to 1975), personal communication with the ice mappers indicated that only ice boundary mapping and not concentration variations within the winter Antarctic ice edge were mapped because of the navigational nature of the map usage and the absence of ship traffic at that time of year. More recent mapping (e.g. Fig. 1) has indicated more open water than previously mapped albeit still lower than that calculated from digital data, perhaps indicating more attention to this variability than previously shown by the mappers, even in the navigational "off season".

Two more quantitative models, but considering primarily thermodynamic effects, have also been attempted. Pease (1975) in formulating a thermodynamic model for the Antarctic calculated one dimensional thermodynamics along the 155°E meridian. This meridian was chosen because various sources of observational data showed less disagreement here and the advective characteristics of the Weddell and Ross Sea areas (which we discuss later) were thought not to be a factor at this location, that is the advance-retreat at this meridian may be more characteristically thermodynamic than dynamic. This reasoning is inaccurate since the drift track of the trapped ship *Aurora* in 1915-1916 from the Ross Sea embayment crossed the 155°E meridian at about 67°S and showed significant drift between 160° and 155°E (Wordie, 1921a).

This ship track is essentially a particle path for ice floes indicating significant advective ice contribution at this meridian from the deepest part of the Ross Sea embayment. The "residual" ice mass present between the Oates Land Coast and the Balleny Islands is also renowned for its summer persistence and heavy ice condition (Streten, 1973; Law, 1964) indicative of its advective contribution. Pease's adjustment of the parameters in the thermodynamic calculation in order to compare favourably with mapped ice extent values is clearly not valid due to the advective contribution to the ice at this meridian in the actual case.

Parkinson & Washington (1979) have developed a numerical model simulating annual ice variability in both the Arctic and Antarctic, which includes both dynamics and thermodynamics. This study has been essentially thermodynamic in nature and includes only a crude parameterization of ice motion. Hibler (1980) summarizes the problem as being that the ice velocity field obtained by assuming free drift is iteratively corrected to insure the maintenance of a fixed fraction of leads. This iteration is, however, performed without regard for conservation of momentum, and in practice tends to stop all motion rather than modify the relative motion as occurs in reality.

In the steady-state, the force balance on the ice in Parkinson & Washington's model is

$$0 = \tau_a + \tau_w + D + G \tag{4}$$

where τ_a, τ_w are the forces induced by wind and water; $D = \rho_i h_I f(\underline{V}_I x \underline{k})$ the Coriolis force; and G the force induced by the dynamic topography gradient.

The thermodynamics are computed from a surface energy balance involving either water, ice or snow-covered ice. Included are the radiation balance at the surface, that penetrating the snow and ice surface, sensible and latent heats, and oceanic heat flux including a parameterization of the oceanic mixed layer (for details see Parkinson & Washington, 1979).

Figure 10 shows the monthly simulated ice extent values from Parkinson & Washington, and indicates that the agreement of ice boundary is generally within 2° of latitude of atlas values. While that statement may be true, the entire seasonal range is only about 6° of latitude over major portions of East Antarctica and the simulated extents in that region can be off by over 100%. The timing of the seasonal cycle is in reasonable agreement with observation with minimum extent occurring in March and the modelled maximum in August, somewhat off the observed maximum in September. The more major conflicts with the limited data available result from the simple dynamics in the Parkinson & Washington model. They indicate the large majority (estimated about 80-90%) of the pack ice in winter is over 90% concentration in conflict with the more recent estimates by Zwally *et al.* (1979) indicating lower concentrations of only at most 50% of the ice at over 85% concentration. One possibility is the ambiguity in satellite imagery between open water and thin ice but for mechanics and heat balance purposes, labelling of the thin ice in an open water category is probably more realistic than in

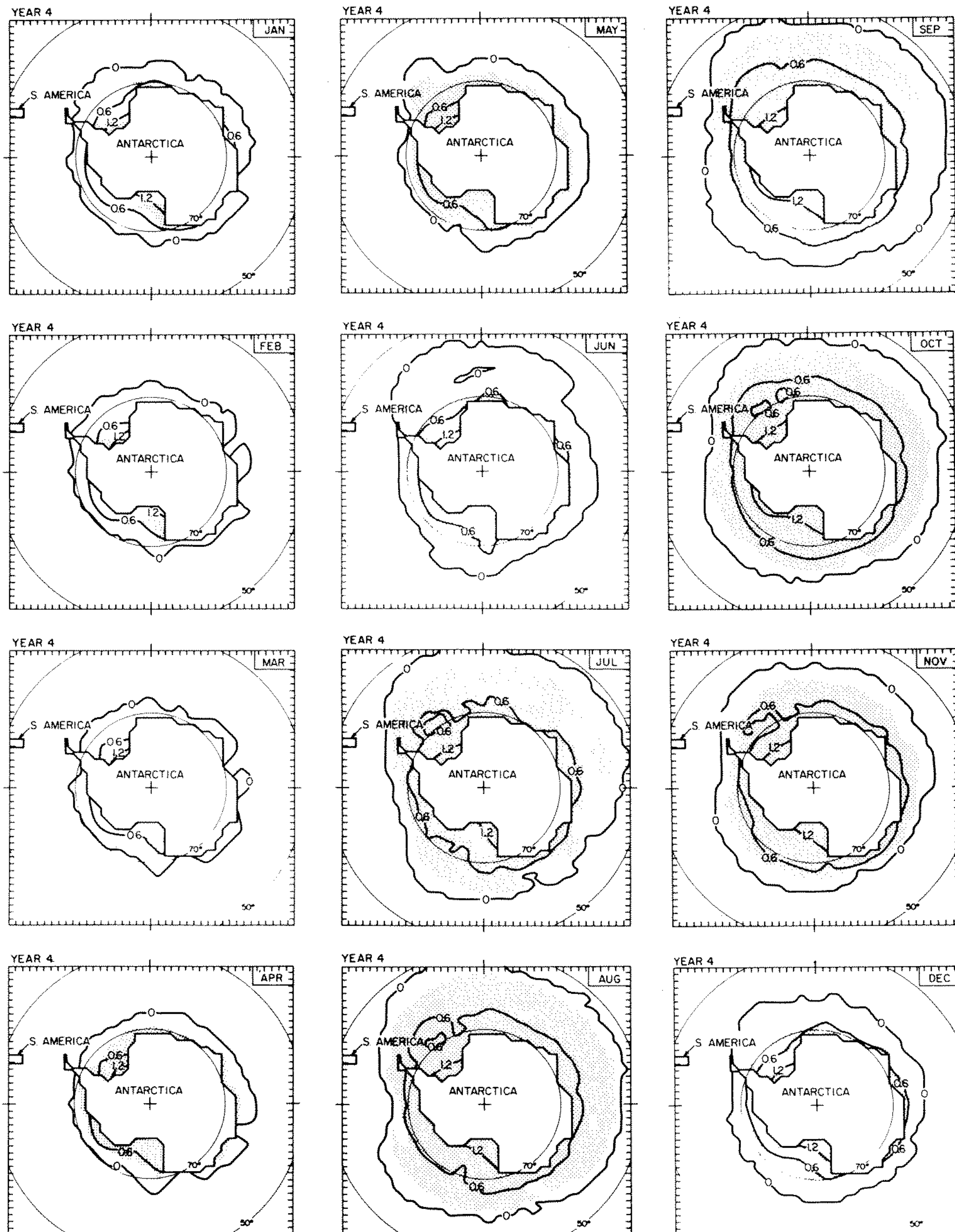

Fig. 10 Monthly simulated sea ice extents (after Parkinson & Washington, 1979). Contours show ice thickness while shading indicates compactness above 90%.

grouping it with thick ice (Hibler, 1979; Maykut, 1978). Similarly, the ice velocity patterns simulated indicate a shutting-down of motion in the interior pack especially during winter. Limited ship and drifting buoy data (Ackley, 1979, 1981) indicate, however, that even in the Weddell where ice conditions are considered the most severe, velocities are commensurate with the higher range observed in the Arctic (>4 km day^{-1}) with the highest values (>7 km day^{-1}) observed during winter in contradiction to the model results. Ice thickness contours are also in the opposite direction in the Weddell Sea to those observed (Ackley, 1979) again because of the lack of ice motion. The lack of dynamics is shown in Parkinson & Washington's (1979) Fig. 25 where relatively little change is seen between the "dynamics" and "no dynamics" (thermodynamics only) case. For the Arctic case Hibler (1979) using a momentum-conserving dynamic-thermodynamic sea ice model indicates dynamics induced transport significantly influences the results and indicates better agreement with observed ice thickness distribution patterns. It also points out the importance of advection in inducing enhanced ice growth in particular regions by the stripping away of the newly formed ice. This mechanism profoundly affects the distribution of major air-sea interaction regions. The advective contribution is expected to be similarly underestimated for the Antarctic case where for example in the Weddell region between 3 and 5 x 10^6 km^2 of ice are estimated advected into the southern Weddell/South Atlantic annually (Gordon & Taylor, 1975). Schwerdtfeger (1979) points out the lack of detail in the wind field and coarse grid-point spacing of the Parkinson and Washington model is at least partially responsible for the discrepancy since the effect of the Antarctic Peninsula, a strong topographic and climatic factor, is not properly resolved. A more fundamental problem is the dynamics formulation. Hibler (1980) points out that the thermodynamic model of Parkinson & Washington, by driving the simulation with observed air temperatures, which are in turn dependent on ice conditions, partially forces proper results. Simply, the question is, which comes first, the ice condition (driven by winds) or the lower air temperature fields?

In contrast to the dominantly thermodynamic approach of Parkinson & Washington, Baranov *et al.* (1976) have calculated a purely dynamic formulation. Here the intent is not to simulate ice extent changes, but to compute velocity fields within the ice and the corresponding convergence-divergence fields for given wind and current field forcing. Basically, additional terms are added to the force balance shown in equation (4) to account for the lateral transfer of momentum and ice resistance forces which are modelled similarly to those of a viscous fluid.

Only April average pressure values were used for driving and the system was spun up from free drift (no resistance) as an initial value. It is claimed that the resultant drift vectors will cause a pattern in ice compression and relaxation zones comparable to mean monthly values for these terms. While of interest in comparing data on ice concentration, direct comparisons are not possible since ice divergence is not computed but only the stress fields of comparative compression. One can surmise that only zero stress would give a region of ice divergence and the total of those areas is quite low in their simulation. The amount of ice concentration within these areas is also unknown. Other difficulties are the use of only a single month's mean pressure field which makes comparisons with long term data sets somewhat questionable. Within the limitations of our knowledge, the drift velocities are reasonably portrayed relative to the known values from ship tracks, primarily in the Weddell, and the patterns elsewhere are similar to those inferred from current measurements in the Antarctic regions. Without simulations of the more measureable observables of ice extent and concentration variations, this simulation cannot presently be confidently validated although the dynamics are preferable to those used elsewhere.

In summary, no coherent sea ice model is available to explain even our limited observations on Antarctic sea ice characteristics. Each attempt to dominate one process over another, thermodynamics, advective transport, or divergence (Ekman transport) in some way violates our sensibilities as to what is in fact the major cause. Gordon & Taylor (1975) offer an intuitively satisfying mechanism in the Ekman transport but the evidence for its operation, the continuous divergence of the interior parts of the pack, while mounting, conflicts with mapping efforts in the recent and more distant past. It also is not sufficiently quantitative nor does it contain a numerical thermodynamics to get any ice growth and, therefore, provides only an intuitive understanding of the phenomenon with no real measure as to its level of skill in explaining it. The Parkinson & Washington model severely limits the role of advective and deformational contributions and while simulating with some verisimilitude the advance-retreat characteristics, it may be a case whereby the cancelling of conflicting unknown terms (for example, strong oceanic heat flux and weak ice transport) may fortuitously give a close approximation to the actual characteristics. Comparisons with ice concentration data, drift, and the ice thickness distribution is considerably less satisfying with this model than the ice extents. The wind drift model of Baranov *et al.* (1976) gives us some measure of ice velocities that moderately satisfy the limited observable drift tracks available, but detail on how ice concentrations might vary, while possibly a function of the stress field terms that are given, is still not available. As well, no thermodynamics indicating the replenishment of divergent ice or modelling at the extension of the free ice edge is included here, and no temporal changes in the wind field are allowed.

We next examine the observable data by region to give some rank to the dominant processes extractable from the observations with reference to the ideas available from the existing modelling as described above. In the Weddell Sea, advection seems to be a major factor as illustrated in Fig. 5. Here we previously indicated the phase lag between the western and eastern longitudes of the Weddell region. It is also seen that the ice edge movement is characterized by an east-moving component (Ackley, 1979) since there is a sharp jump in ice edge latitude to the north, first observed at 40°W, and occurring at equally spaced time intervals to 20°W and finally to 10°E. In the later winter-early spring a second east-moving component successively passes 20°W to 10°E and finally to 30°E in December. The behaviour of the 30°E meridian strongly illustrates the west to east movement since the ice "wave" passes 30°E after the southern retreat of ice at 30°E has begun in November. The result is two maxima in the 30°E annual cycle for this year, one in November and one in December. The advection of the mid-winter growth that took place at more western longitudes is the cause of this feature since spring ice growth cannot be extensive at these lower latitudes (<60°S). Some rough estimates of the amount of ice area involved in the advective amount of the Weddell pack may be made. Figure 11 (Ackley, 1979) indicates that the major drift pattern in the Weddell carries ice from the southernmost regions to the northern-regions in about one year. The western Weddell pack persists throughout the summer (Fig. 3) and as indicated by microwave imagery, the pack maintains a "first-year" ice emissivity in accordance with the drift rates which would indicate it is "turned over" annually. The total area of the minimum summer pack (2-3 x 10^6 km^2) is therefore advected east and north annually. In volume of ice the amount of ice lost in summer melting should be subtracted from this total. Ackley (1979) and observations from the Shackleton expedition (Wordie, 1921b) indicate these ablation losses are minimal, so essentially little to no ice is lost by melting from the year-round pack.

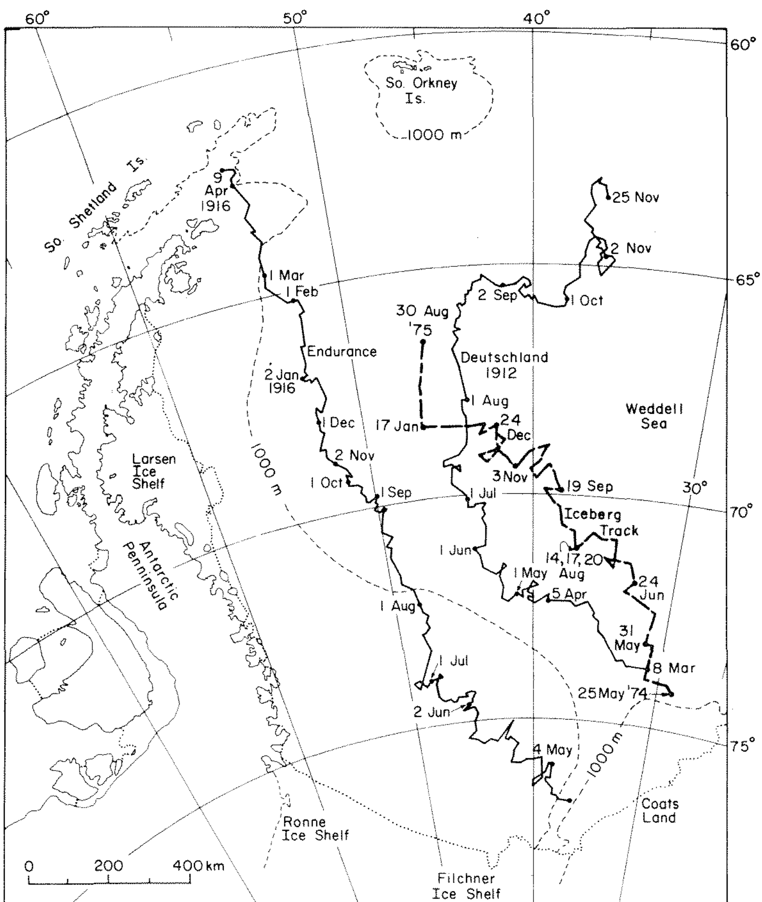

Fig. 11 Drift tracks for trapped ships and an iceberg in the Weddell region (Ackley, 1979).

In addition, there is a certain amount of ice production within the advected area. Data buoy records (Ackley, 1981) estimate this area, which is essentially dynamic production by divergence effects during the advance, as about 50% greater than its initial value during the ice growth period. A combination of divergence with advection of the existing year-round pack, therefore, accounts for 3-5 x 10^6 km^2 of ice area or about 40-50% of the ice in the region from 60°W to about 30°E. This figure is probably a minimum value since it refers only to the "old" ice production and Ekman divergence effects in newly formed ice can probably account for additional ice production in as yet an unknown amount thereby increasing the advective and Ekman divergence contribution. Gordon & Taylor (1975) and Gordon (personal communication) indicate the Ekman transport contribution is probably enhanced on the northern side of the Antarctic divergence where the primary ice production in the Weddell takes place since transport to the left of the predominantly westward winds at the northern lower latitudes acts to expand the ice cover more northerly than in the higher latitudes. Schwerdtfeger (1976, 1979) has discussed the meteorological implications of ice production at high latitudes being carried more northward in the Weddell than elsewhere.

In contrast, the sea ice area around East Antarctica does not show in the advance phase the east or northeast moving character of the Weddell pack. Instead, the advance is consistently northward and Fig. 8 shows the gradient of the area change is constant from March through October. The ice area left after summer melt is also relatively small so advection of the second-year pack in the beginning of the winter period is probably only a small contribution to the next year's advance. A sea ice advance that follows relatively closely the seasonal energy balance cycle implies a more dominantly thermodynamic forcing in this region. This characteristic is probably accentuated by the zonal symmetry of the continent and polar front in this region with the annual sea ice cycle waxing and waning between these two boundaries. The interannual variability also appears less for this region in contrast to the Weddell and Ross Sea regions (see next section). Again if primarily thermodynamic rather than dynamic controls are operative here, more constancy in sea ice area might be expected since extreme temperature fluctuations might be necessary to significantly affect the sea ice response. On a mean annual basis, temperature changes have been small in the Northern Hemisphere (Kukla *et al.*, 1977) so relatively smaller sea ice variability is expected than that of dynamically controlled areas such as the Weddell where annual winds can show great variability and commensurately more sea ice variability in response to these effects. Weller (1968) and Allison & Akerman (1980) provide data on the energy balance both over sea ice and open water at Mawson, an East Antarctic location. The period of energy balance change over open water is in phase with the change in northern extension of the ice boundary, in contrast to the western longitudes of the Weddell region where there is a strong early season peak in the sea ice advance followed by a levelling off of area change in the mid-winter. Again the sea ice boundary changes in phase with the seasonal temperature cycle is indicative of primary energy balance forcing of the sea ice cycle. In the Parkinson & Washington model, however, where thermodynamics account for the primary forcing, ice extents in this region disagree with the Soviet Atlas of the Antarctic and also seem to give greater extents in this region than recent satellite data would indicate. The reason for this disagreement is not known.

The Ross Sea (160°E-130°E) (Fig. 8(e)) is another area where advection effects seem strong, at least in the early part of the advance season (March-June). One characteristic is a sharp gradient of area change comparable to the Weddell Sea during this time. This change is attributable to strong southerly winds off the Ross ice shelf region that rapidly advect and also freeze rapidly new ice in the Ross Sea embayment. After June, however, the areal change is relatively small throughout the rest of the winter period (at least in 1974-1975). Schwerdtfeger (1979) points out the contrast between the Weddell and Ross Seas maybe the turning effect of the Antarctic Peninsula which because of its northward deflection of the ice until about 60°S, allows the ice to be injected into the strong westerlies of these latitudes. In contrast, the Ross Sea embayment only extends to ∿70°S so ice that flows from the high latitudes is deflected west (by the polar easterlies) rather than east as in the Weddell and Ekman effects might be more likely to pull it toward the south rather than the north once it reaches 69° or 70°S. As a result, the drift track of the *Aurora* from the Ross Sea (Wordie, 1921a) shows such a deflection and the area between the Balleny Islands (67°S, 163°W) and the Oates Coast (70°S, 163°W) is known for difficult sea ice conditions for ships, implying some piling up of ice in this region. Zwally & Gloersen (1977) have pointed out that the maximum ice edge seems to map the circumpolar current in this region implying the southward deflection of the current by the Macquarie Ridge (Gordon *et al.*, 1978) may be influential in either compacting the ice edge in this region or possibly changing the heat balance such that new ice is not formed at the ice edge. Continuous loss of ice in the edge region may be in the first approximation a tenable hypothesis since we note that the amount of compact ice maintains a relatively constant percentage of the total ice area throughout the winter period and does not increase. It seems reasonable that the advective effects would not turn off in mid winter so we might assume that ice is constantly being supplied to the ice edge during winter. Heat transport from the ocean, however, that is possibly greater at the northern latitudes here than elsewhere may be influential in melting the advected ice and keeping the total ice area and highly concentrated ice area at a relatively constant level during the latter part of the winter season.

The Bellingshausen-Amundsen sector may mix all of these effects depending on seasonal variations in forcing. 1973 showed a high rate of advance up to July 1973 followed by a sharp decline and then a constant ice area for the rest of that winter. Ackley & Keliher (1976) noted that the ice retreat was coincident with a blocking high over South America which apparently injected anomalously mild air into the region for over ten days. Temperatures rose above the freezing point at several Antarctic Peninsula stations at this time. In contrast, 1974 and 1975 showed relatively slower ice advances but a levelling off in advance rate as the maximum extent is approached, similar to the late season behaviour of the Ross Sea sector, implying some ocean mixing controls as discussed above. The rate of early season advance though is more similar to the East Antarctic advance implying more thermodynamic control (energy balance) of the ice advance during 1974 and 1975. The drift rates from the *Belgica* indicate the pack is more sluggish than observed elsewhere again reflecting a low dynamics and Zwally & Gloersen (1977) indicate that the emissivity variations would be indicative of a multi-year signal from the coastal band of ice, again implying low motion. It appears, however, from the 1973 behaviour, this sector while usually showing less dynamics than elsewhere, can be markedly affected by abnormal atmospheric events which change the advance rates here if their timing is appropriate. Without a longer data record, however, the frequencies of these abnormal events are difficult to determine.

STATISTICAL ANALYSES OF SEA ICE EXTENT AND VARIATION

Mean values and harmonic analysis

While mapping provides details of individual processes, some statistical description of the sea ice data is necessary to understand adequately the relationship between parameters and assign physical interpretations to these parameters. Rayner & Howarth (1979) have recently analysed the microwave sea ice data and their results will be briefly summarized here. Two series were generated at each meridian for these analyses: one of maximum ice extent (minimum latitude = MINL) and one of minimum continuous ice extent (maximum latitude = MAXL). Fourier analysis of the 180 components generated indicates the first seven harmonics are sufficient to describe more than 70% of the maximum ice extension boundary throughout the period and more than 90% of the winter seasons.

The first amplitude of the data, the zonal average, displays annual fluctuations. Extremes for the 1972-1975 period were, 59.93°S on 4 September 1974, and 68.84°S on 19 February 1975 for the minimum latitude series and 60.60°S on 30 October 1973 and 69.77°S on 16 February 1975 for the maximum latitude series. Based on the means, an asymmetry of the growth-retreat characteristics is shown, with winter lasting 205 days while summer is about 160. The spring-summer removal (120 days) is also more rapid than the autumn-winter growth (150 days). Least squares fitting of the maximum ice extent gives an overall annual mean latitude of 63.75°S and an annual range of 8.66°. Phase computation with asymmetry included for the difference in ice removal and ice growth times gives a summer minimum time of about 15-20 February and a winter maximum of 10-15 September.

The second amplitude of the harmonic analysis gives the centroid of the ice, including the continent. Due to the asymmetry of the continent, all of these lie in the Eastern Hemisphere between longitudes 15°E and 120°E. Surprisingly, the centroid is furthest from the pole (∿500 km) in winter similar to the centroid of the continental ice alone. In summer, the pack ice centroid is as little as 70 km from the pole, an indication that the pack ice, especially during periods of minima, is located so as to offset the asymmetry of the continental ice and land. Spatial variance of the boundary gives some measure of the north-south deviations from the mean latitude. The range is 5.3 to 15.3 degrees squared with an average of 9.30 degrees squared. In summary, the Fourier analysis indicates the 360 boundary points of the maximum extension can be adequately explained by 15 Fourier coefficients (8 amplitudes and 7 phases). These statistical analyses indicate general features of the pack ice distribution that may be comparable to modelling studies. These are that the area of ice (including the continent) may be expected to double between the summer minimum in mid-February and the winter maximum in mid-September. Asymmetry in the growth and decay times and consistent location of the centroid along approximately 30°E in winter should remain. Polynyas will exist throughout some winter seasons (as shown by meridian time series) but will dominate the spring.

EOF analysis and implications for stochastic climate models

Lemke *et al.* (1981) have also computed statistics on sea ice variability from digitized ice chart data. The annual cycle of variability from the analysis of 1973-1979 ice chart data is shown in Fig. 12 (Fig. 4 of Lemke *et al.*). As noted previously, and confirmed here, the strongest annual changes are found in the Weddell and Ross Seas. As well the anomaly variance, σ_a obtained by computing the deviation from the 6-year mean (these are annual average values), shows similar patterns. These are consistent with the primarily dynamic controls in these regions discussed earlier compared to the lower variance East Antarctica and Bellingshausen-Amundsen regions. They note the anomaly variance of the sea ice area exhibits only about one third of that of the annual cycle. This finding essentially puts a number on the possible inter-annual change albeit for a short data record which was earlier speculated (Fletcher, 1969) to perhaps be as large as the annual cycle. The recent data do not indicate such large changes have recently occurred. A further interesting finding is given by Empirical Orthogonal Function (EOF) analysis of the sea ice changes. It was found, as in Rayner & Howarth (1979) that about 80% of the total variance is contained in the first seven EOFs and that the lower-order EOFs are seen to be associated with larger spatial scales (Fig. 13). Lemke *et al.* note that the main contribution to the first EOF comes from the Weddell Sea while the second EOF is mainly determined by the Ross Sea area. This decomposition implies that these two main areas of variability are nearly decoupled, although the behaviour in 1973-1974 implied they were 180° out of phase in that year, at least for the seasonal extension rather than the annual averages given by the EOF analysis. Further analysis of these functions on a seasonal

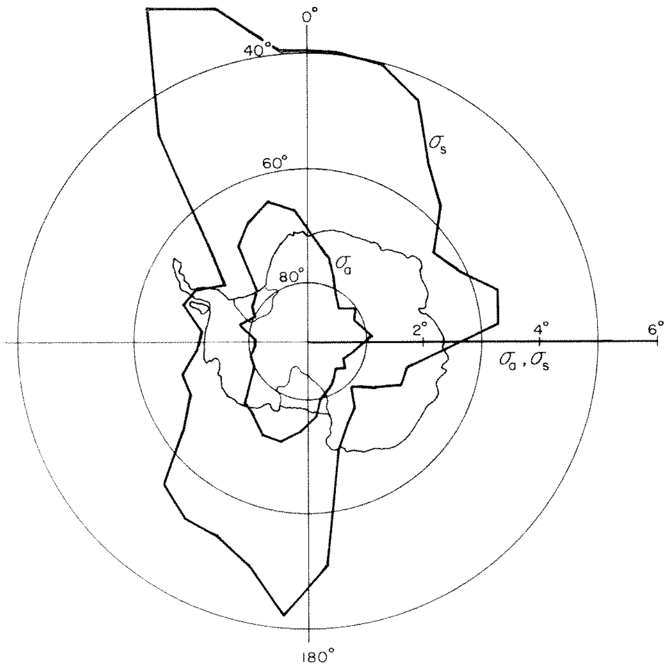

Fig. 12 Annual cycle (σ_s) and sea ice anomaly (σ_a) standard deviation as a function of region around Antarctica (after Lemke *et al.*, unpublished).

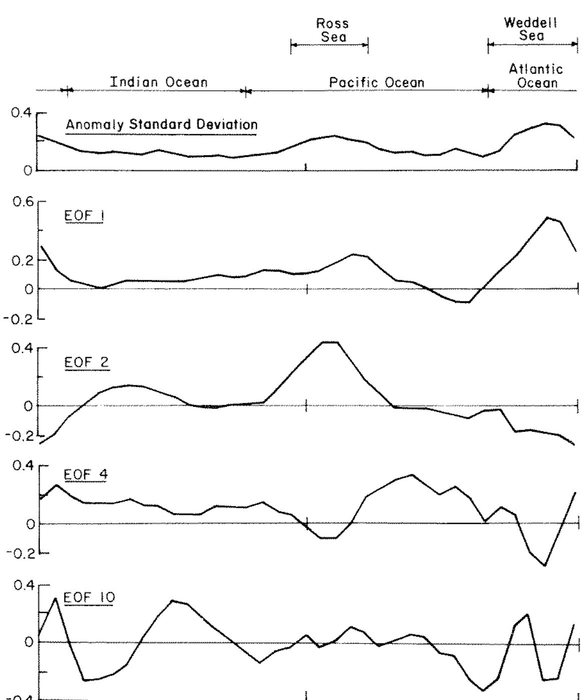

Fig. 13 Total sea ice anomaly standard deviation and decomposition of the variance into empirical orthogonal functions 1, 2, 4, 10 for the Antarctic region (after Lemke *et al.*, unpublished).

basis should also be of interest and Lemke *et al.* plan to conduct these in the near future.

In comparing the Arctic and Antarctic using linear stochastic climate models, Lemke *et al.* find that a first order Markov model employing only a local feedback correlation term is relatively complete in describing the Arctic ice variability as a function of white noise atmospheric forcing. The Antarctic on the other hand is more accurately described by three feedback parameters for each longitudinal sector, incorporating the local feedback, a diffusive or symmetric parameter to adjacent sectors and an advective or anti-symmetric coefficient to adjacent sectors. In a physical sense this higher order parameterization indicates more than local effects including both atmospheric feedback and ice advection are needed to properly simulate the ice response in Antarctica, in confirmation of the inferred ice edge controls postulated earlier.

EFFECTS OF SEA ICE ON WEATHER

Fletcher (1969) shows that rates of atmospheric heat loss vary greatly over the three major components of surface material in the Southern Hemisphere: open ocean, sea ice, and continental ice. Since sea ice shows the greatest annual variability, it is hypothesized to account for a significant amount of the seasonal and inter-annual variability of the weather. Figure 14 shows the annual pattern of heat loss by the atmosphere over these three surfaces. These values are based on seasonal budgets of the

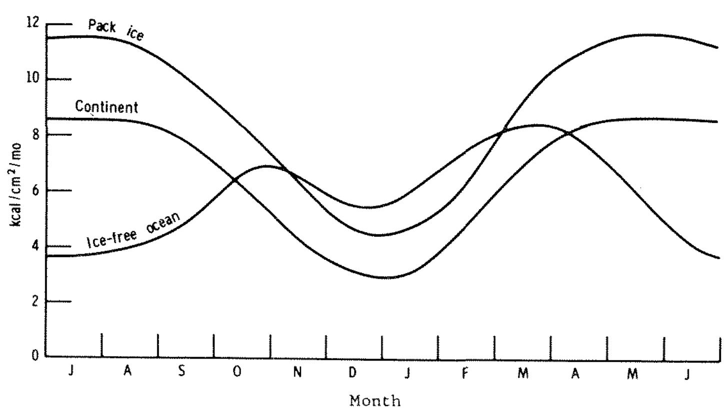

Fig. 14 Annual pattern of atmospheric heat loss over sea ice, open ocean and continental ice (after Fletcher, 1969).

radiation and surface heat fluxes over these surfaces. Ackley & Keliher (1976) use satellite data on sea ice coverage to compute the regional variation of atmospheric heat loss as well as the inter-annual change. They show that while 1973 and 1974 did not vary in total heat loss (comparing sea ice extents only and not concentrations within the ice pack for the 2 years), significant heat loss occurred regionally which could affect the character of circulation by changing the major areas of air-sea interaction. As well, ice concentration variations could also provide

Table 1 Total sea-air heat flux 60°-70°S, Q_T (values in cal cm^{-2} day^{-1}, positive indicates atmosphere to ocean heat transfer)

Month	A No ice	B 100% ice cover	C* Ice chart Conc. data	D† Same as C but with 30% leads
Jan.	+177	+64	+132	+145
Feb.	+66	+18	+56	+59
Mar.	−128	−26	−108	−114
April	−201	−46	−139	−158
May	−329	−56	−193	−233
June	−449	−54	−212	−283
July	−452	−54	−173	−257
Aug.	−377	−34	−120	−195
Sept.	−297	−21	−76	−142
Oct.	−130	−20	−53	−76
Nov.	−84	−12	−41	−54
Dec.	+55	+15	+35	+41
Average	−179	−19	−74	−106 cal cm^{-2} day^{-1}
	−65	−6.9	−27	−39 kcal cm^{-2} a^{-1}

Notes: $Q_T = (Q_s + Q_b) + Q_E + Q_h$ = radiation balance + evaporation + sensible heat loss

Q_s from Sasamori *et al.* (1972). An albedo of 0.4 is used for the direct beam and 0.066% for the indirect short wave radiation.

Wind speed: 5m s^{-1}
Relative humidity: 85%
50% middle cloud cover

T°C sea: −2°C May-November
0°C May, October-November
+2°C January-March

T°C air: −13°C June-September
− 8°C May, October-November
− 5°C April, December
− 3°C March
+ 1°C January-February

* Estimate of % ice cover taken from ice charts.
† Percentage of leads varies greatly, 30% may be upper limit. The Weddell Polynya not explicity included in these calculations.

significant changes in the heat loss pattern. The amount of heat exchange using microwave data on ice concentrations relative to totally ice covered area was found equivalent to about the total energy of one or two extra cyclones per week. Gordon (1979) has indicated the importance of the area south of the Polar Front Zone (PFZ) as possibly the "heat exchanger" for the entire world ocean, providing the major region for heat loss by the ocean that balances heat input at almost all other latitudes. The crucial question here is the amount of open water within the ice-covered area as it accounts for the highest values of the oceanic cooling. Table 1 indicates this comparison with no ice A, 100% ice cover B, ice chart average concentration data C, and an estimate with 30% open water D, a figure commensurate with the digital microwave estimates reported earlier here. As can be seen, annual average values for the two most reasonable estimates (columns C and D), differ by close to 50% in the total heat loss. Table 2, also from Gordon (1979), indicates the relationship between net air-sea heat flux by zonal region, including the PFZ.

Table 2 Net sea-air heat flux (positive values represent heat from atmosphere to ocean)

Zone	Latitude	Area ($\times 10^6$ km^2)	Q_T (kcal cm^{-2})
Sea ice zone	60°-70°S	16	−27 (Table 1 calculations)
Northern ice fringe	60°S	—	− 6.4 (Fletcher, 1969)
South of PFZ	53°-60°S	20	+10 (Zillman, 1972, Bunker, p.c.)
North of PFZ	45°-53°	22	−15 (Zillman, 1972, Bunker, p.c.)

Summary of total heat flux:

1. South of polar front zone: $-\ 7.4 \times 10^{13}$ cal s^{-1}
2. South of 60°S: -13.0×10^{13} cal s^{-1}
   South of 60°S with 30% less ice: -20×10^{13} *cal* s^{-1} *(a nearly 50% increase in ocean heat loss).*

Again the relative importance of the sea ice zone (with either estimate of ice concentration) in venting the ocean is shown. Weller (1981) in reviewing the South Polar energy balance has compared the fluxes of sensible and latent heat into the atmosphere only with the assumption that instead of open water, thin ice of only 5 cm thickness is used. These indicate fluxes of 552 cal cm^{-2} day^{-1} and 284 cal cm^{-2} day^{-1} respectively for this region indicating a further crucial step, i.e. with reduced ice concentration, how much of the open water is actually thin ice, a quantity that is presently indistinguishable from open water in the microwave data. In any case, the more conservative of these numbers, 284 cal cm^{-2} day^{-1} is still 3 or 4 times the value previously obtained by Fletcher (1969) indicating the substantially higher air-sea fluxes that may be present south of the ice edge than previously assumed. Weller also speculates that the primary inter-annual variability in the fluxes of sensible and latent heat into the atmosphere are a function of the ice concentration variability and that the energy balance in the pack ice zone remains the major unanswered question in our understanding of the thermal forcing of the atmosphere in the Southern Hemisphere.

To obtain some measure of the effects of sea ice on weather, Herman & Johnson (1978) have used a general circulation model where sea ice conditions are an input boundary condition to examine the possible sensitivity of the Northern Hemisphere general circulation fields to the observed variability in ice conditions. They use 17 years of North Atlantic ice conditions and 5 years of data in the Pacific sector. While this study is not directly relevant to our pursuit here of Southern Hemisphere relationships, it provides some confidence that effects can be expected, based on observed, rather than inferred, variability in ice conditions, and lends some credibility to the necessarily more qualitative and correlative search for connections in the more data-poor regions of the Southern Hemisphere. A control was defined as the mean of six January-February simulations with ice boundaries corresponding to the minimum ice cover occurring simultaneously in the North Atlantic and North Pacific. An anomaly was defined as the mean of two simulations with maximum ice conditions occurring simultaneously, both sets of conditions obtained from the recent data on ice charts since 1961. The comparison between maximum and minimum ice conditions on the January-February climatology showed higher sea level pressures over the ice areas (near the edge) by 4-8 mb with lowered pressures in the North Atlantic by about 8 mb in the vicinity of the Icelandic low and similarly in the Gulf of Alaska (Aleutian low). Geopotential heights at 700 mb were lowered by 80 gpm in the Gulf of Alaska and 100 gpm in the Icelandic region. Temperatures were also affected in the lower troposphere (below 800 mb), being lowered by about 2°C between 50 and 70°N. These differences from the control were statistically significant at roughly the 95% confidence level. While not including any feedback effects since the sea ice limits are imposed boundary conditions in the GCM used, these results are reassuring in that observable effects should be expected, and that the expected correlations are numerically of the order outside the noise figures of the recent weather record.

SOUTHERN HEMISPHERE CORRELATIONS

Budd (1975) has described some effects relating sea ice extents to surface temperature conditions, specifically comparing the sea ice conditions obtainable by satellite photos from 1967 to 1970 and Fleet Weather maps from 1971 on. The longest Southern Hemisphere record, that of Laurie Island was shown to be highly correlated between temperature and sea ice conditions. Fletcher (1969) used this record to infer total ice conditions around Antarctica and subsequently hemispheric effects on temperature and circulation. However, as the later record has shown, and Budd points out, anomalies may be regionally confined and not representative of total ice conditions around the continent. Linear regression of the Laurie Island data indicates a ±1°C mean annual temperature change *locally* corresponds to a change in ice duration of ±70 days *locally*. From the extent of sea ice in different months, the period of ice duration as a function of latitude was derived. In combination with the Laurie Island temperature-duration curve, this gives Fig. 15 (after Budd's Fig. 3) showing the relation between fluctuations in the mean annual temperature and the latitude of the maximum ice extent. At the latitude of ∿60°S corresponding to a mean position of the maximum ice (Rayner & Howarth, 1979), the slope of the variation gives a gradient of approximately 1° lat./°C. The propagation of anomalies was also observed with corresponding sea ice and mean annual temperatures anomalies of the Antarctic coastal stations apparently propagating from west to east at approximately 20 to 100° longitude/year. The tendency for opposite behaviour of anomalies, i.e. as some grow others weaken, was also shown, again consistent with more recent satellite sea ice data indicating small total change in sea ice extent while regions can have significant anomalies of opposite sign (Ackley & Keliher, 1976; Zwally *et al.*, 1979).

Ackley & Keliher (1976) compared cyclone data in the

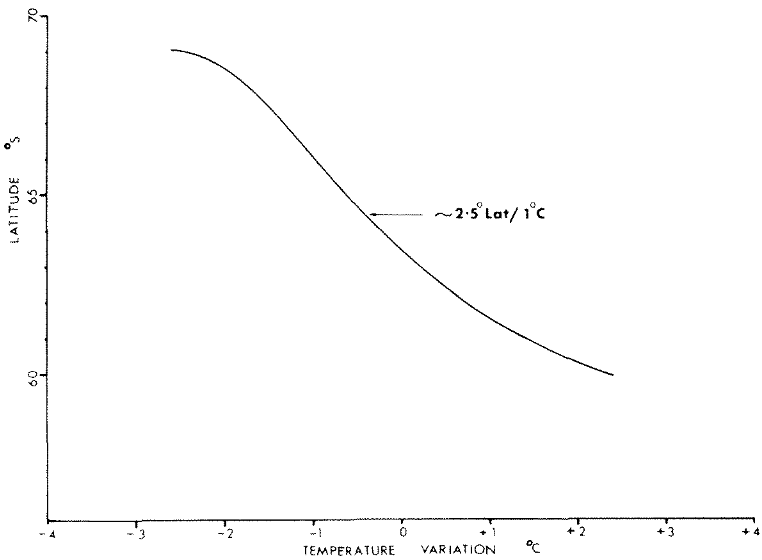

Fig. 15 Relationship between mean annual temperature at Antarctic coastal stations and latitude of maximum ice extent. The slope at 60 °S is about 1 ° lat./degree (after Budd, 1975).

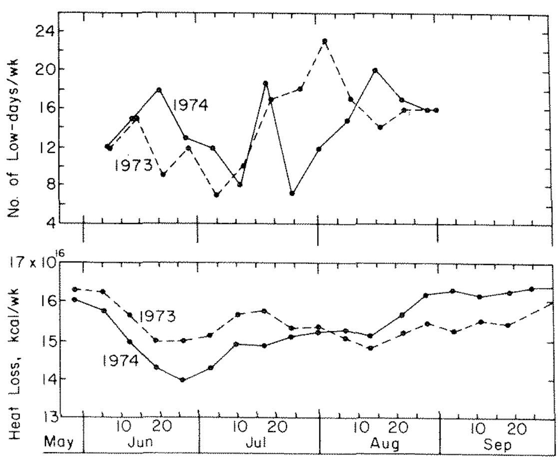

Fig. 16 Atmospheric heat loss comparisons and cyclone activity in the Bellingshausen-Amundsen sector for 1973 and 1974. Heat loss is based on mapped ice extent differences between the 2 years. (After Ackley & Keliher, 1976.)

Bellingshausen-Amundsen Sea sector between 1973 and 1974 where an apparent "step" anomaly in ice extent occurred reversing the seasonal trend at this location. The behaviour is shown in Fig. 8(f) where the increasing ice extent for 1973 suddenly drops during July for this sector. The ice loss period corresponded to a noticeable increase in numbers of cyclones before and after this event in this sector. Comparisons between 1973 and 1974 values (Fig. 16), both for atmospheric heat losses and cyclone counts show that the anomalous sea ice decrease in 1973 could be a factor in increasing the oceanic heat input to the atmosphere thereby increasing either the longevity of storms or the amount of cyclogenesis, or both. Even though this study dealt with numbers of cyclones which could be counted from maps or satellite photos, it is not a quantitative study since the total atmospheric energy contained in the systems may not have changed significantly between the two years. This energy cannot be computed accurately since typically Southern Hemisphere charts were contoured at a coarse 10 mb interval for this period, so quite large features can only be coarsely resolved. Barry (1980) in comparing National Weather Service chart data with simultaneously obtained buoy data that was not used in the chart analysis for the Arctic Ocean has indicated that individual grid point pressures over a 2 month period agreed for only 14 days, were too low on the charts for 38 days and were too high on 14 days. It was also found that major synoptic features were either missed or poorly represented for about half the period. Apparently, recent use of buoy data in Southern Hemisphere analysis has also indicated that major synoptic features were previously underestimated in Australian Bureau of Meteorology analyses.

In more general terms, the relationship between the sea ice boundary and the pack ice margin does not indicate strong evidence for a climatic front (Taljaard, 1972). As stated by van Loon (1972b) "In the transition seasons (spring and fall) when the mean position of the circumpolar trough is nearest Antarctica, the border of the ice is near its extreme poleward and equatorward positions. In the months when the trough on the average is farthest north, the ice border has reached no such extreme position". Schwerdtfeger & Kachelhoffer (1973) do, however, indicate a close seasonal relationship between the ice margin and the latitudinal band of maximum cyclone frequency in the transition seasons. More recently, Howarth (1980) has examined cyclone tracks and frequencies in 1973-1974 at the maximum and minimum sea ice periods. No close correspondence between the tracks and the ice boundary was found. In the Northern Hemisphere there appears to be a displacement south of the ice edge of the major depression track although it parallels the ice margin at least in the North Atlantic (Barry, 1980). The hypothesis has been that the ice margin may be related to cyclogenesis since air temperature contrasts on either side are expected to be particularly severe. The area could then be one of strong baroclinicity and classical polar front disturbances could form there and, therefore, lead to the enhanced development of vortices. The cyclone data record does not apparently bear this out and may suggest that with the newer ice data especially on concentration, the ice edge may not be in fact the optimum region for a zone of enhanced cyclogenesis. The reduction in assumed ice concentration from previous estimates, relatively increases the heat flux by about 4 or 5 times in the pack ice zone as a whole, thereby leading to a general reduction in the assumed baroclinicity at the ice edge since the temperature contrast is probably spread over a wider band. In fact, the strongest contrast may be between the continental ice and the sea ice zone where open water heat losses are a factor. Both the more recent cyclone studies (Howarth, 1980) and the quote from van Loon (1972b) allude to this, and with more data differentiating cyclones originating rather than moving into the ice zone we may find that there is a zone of enhanced heat flux at the continental rather than pack ice edge and a broader zone of cyclogenesis associated with the entire pack ice/open water region. At the Polar Front where again strong horizontal gradients occur in the surface temperature field, enhanced relative development of cyclones may also occur rather than further south at the ice edge.

Streten (1973, 1977) has examined the zonal indices of the Southern Hemisphere given by the pressure difference between pairs of stations on either side of the Polar Front, including the entire pack ice zone. In the earlier study, monthly mean values of the flow for October corresponded well with existing data on sea ice showing a generally greater ice extent in 1969 and relatively higher flow for this same period. The areas of opposite sign (lower flow in 1969) were generally in the Eastern Hemisphere where little change in ice extent or slightly greater extents in 1970 were observed. In the later study (Streten, 1977) seasonal mean indices of the westerly air flow are compared. It was found the zonal flow is strongest and most variable in spring and displays the greatest regional variability southeast of Australia and north of the Weddell Sea (Fig. 17 after Fig. 10b of Streten, 1977). Given the limited number of

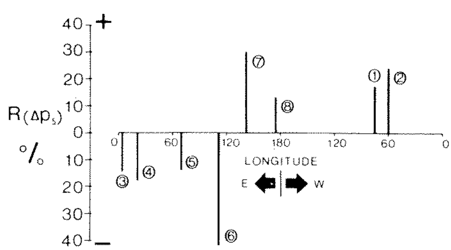

Fig. 17 Percentage departure from the annual zonal mean by longitude position of station pairs for all seasons (after Streten, 1977).

station pairs and their lack of total regional sampling, it is still interesting to compare Fig. 17 with Fig. 12 after Lemke *et al.* which indicates the regions of greatest ice variability correspond, within broad limits, to the regions of high zonal wind variability. Individually through numbers of cyclones (Ackley & Keliher, 1976), and collectively through mean zonal wind variations, sea ice variations are apparently related to the circulational features that are important on the weather time scales.

Schwerdtfeger (1979, 1976, 1975) has indicated the unique interaction between ice and weather conditions in the Weddell Sea region. Here the topography of the Antarctic Peninsula is a major control, presenting a meridional barrier in essentially a zonal flow. The effect is to deflect air and ice into a northern course parallel to the Peninsula, and with the additional "stacking" of the colder air against the 1500-2000 m mountain barrier, a wind effect leading to enhanced relative occurrence of cold southerly winds occurs on the eastern side of the Peninsula relative to conditions elsewhere. The barrier winds are expected to be larger than 100 km in lateral extent and, therefore, large enough to strongly influence the ice drift here. Since the Peninsula extends to far more northern latitudes than other regions, the ice stream is forced into the westerlies. This injection leads to changes in the isotherm pattern skewed further north in the Weddell Sea than elsewhere along with cyclone and other circulation indices as mentioned earlier. The Weddell Sea is looked on as a key area in understanding the Southern Hemisphere's climate and climate variations because of these ice-atmosphere interactions which apparently are felt from the Peninsula past the 0° meridian to approximately 30°E, i.e. about 90° of longitude.

CONCLUSIONS

While our data record on the sea ice variations over the past 10 years has increased measurably in the Southern Hemisphere, some good and some not-so-good prospects are on the horizon. The good news is that a better quantification of ice-weather relationships may be possible, especially for calendar year 1979, since close to 300 drifting buoys were emplaced in the oceanic regions of the Southern Hemisphere with some in sea ice regions (Ackley, 1981; Vinje, unpublished) that should assist greatly in resolving the weather in this region. Ice mapping has also been improved although the newer microwave sensor (SMMR) on NIMBUS VI has apparently suffered calibration problems that may make detailed knowledge of ice concentration especially, a little less sure than previous years. Ironically, it appears we are on the verge of a great step backward, for the expected lifetime of current microwave mapping satellites is not expected beyond roughly 1980 and no new launches are anticipated until 1983. Nevertheless, a concerted correlative effort for 1979 together with some knowledge of the detailed sea ice variations of the 1970s as shown earlier here, may be enough to establish the principal deterministic connections between sea ice and weather and allow the appropriate models to be constructed that conform to our knowledge on ice characteristics and provide some skill in establishing the appropriate two-way interactions between ice variations and weather. As always, more information will continue to be needed, and some critical areas of concern are the correct specification of cyclone characteristics in and near the sea ice belt, the differentiation and correct portrayal of sea ice concentrations, and energy balance measurements over drifting ice and open water in Southern Hemisphere winter conditions. The current results on sea-ice weather relationships are qualitative but intriguing and suggestive of the intermediate few weeks to few months time scale between the response of the atmosphere (a few days) and the ocean's upper layers (months to a year) that is both of high interest for long range forecasting and probably one of the least known components of the climate system.

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