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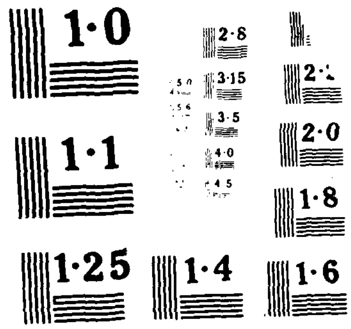
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EXAMPLES OF ICE PACK

RIGIDITY AND MOBILITY CHARACTERISTICS

DETERMINED FROM ICE MOTION

Science Applications International Corporation

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EXAMPLES OF ICE PACK RIGIDITY AND MOBILITY CHARACTERISTICS DETERMINED FROM ICE MOTION

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ABSTRACT

A method has been developed to determine ice pack rigidity and mobility using observed ice motion. Using this method, one may determine how solidly the ice pack is frozen in near real-time. In addition, spatial and temporal variations in the freezing and thawing of the ice pack can be studied. This method was developed to study how well numerical ice modeling techniques were reproducing actual sea ice processes. The ice pack rigidity parameter is now being used to compare observed periods of ice pack thawing and freezing with model results from the same regions and times. This report describes the development of the ice pack rigidity/mobility parameter. An application of this method to periods in 1975 and 1979 is discussed in detail.

Various degrees of ice rigidity and mobility were studied using remotely-sensed ice motion data off the north coast of Alaska during 1975 and 1979. Characteristics of the time histories of pack ice speed were used to infer changes in the rigidity of pack ice. Summer-time ice rigidities were detected first in late June 1975 and lasted through September. However, in 1979 considerably higher rigidities were found in August while summer-like rigidities were detected into late November. Analyses of atmospheric pressure distributions suggest that less mechanical breakup occurred in the summer of 1979, resulting in the greater rigidities during August of that year. In addition, minimum ice coverage was 20% less in the Beaufort Sea in 1979 than in 1975. The result was a relatively large percent of thinner ice for November of 1979 than for 1975, the likely cause of the less rigid conditions detected during the fall of 1979.

NOMENCLATURE

D - deviation in height (m) of a pressure level from the standard atmosphere (Huschke, 1959).

e_T - time lag (h) at which the autocorrelation of ice speed drops to e^{-1}.

SL - the large-scale disturbance or longwave component of a scalar field; in this study, the 500 mb circulation (Holl, 1963).

U - mean ice speed over a given time interval.

V - variance of ice speed over a given time interval.

I INTRODUCTION

To navigate and access natural resources in ice-infested waters, one must deal with the impact of ice with ships and production facilities. The consequence of such impacts depends upon the speed of the ice, the ice thickness, and the ice pack rheology. The importance of the latter variable, ice pack rheology, is in terms of how solidly the pack ice is frozen. In this case, we are not solely considering the material rigidity of the ice itself. Rather, we are concerned with the rigidity of a medium which is made up of individual ice floes and which responds in a particular way to external forcing (e.g., wind). A winter-like ice pack, with its stronger internal ice structure and greater concentration of floes, has an enhanced capability to transmit internal stresses (at a rate of up to 10 m/s). A summer time ice pack is less of a collision threat since there is a larger percent of open water and the internal ice structure has decayed with the warming of the ice.

Thus, it would be useful to have an ice pack rigidity/mobility parameter by which one may judge how solidly the ice pack is frozen. Such a parameter could be used to indicate when the ice pack undergoes its seasonal thaw and freeze so as to better schedule operational activities. Also, one could make better predictions of the potential harm of a forecasted ice/structure impact. Moreover, observed variations in rigidity could be used to verify predictions of numerical ice models. Finally, a rigidity parameter could aid climatologists in determining the times and spatial extents of arctic air/sea interaction modifications that result from the thawing and freezing of the ice pack.

Seasonal variations in the characteristics of ice parameters have been noted in numerous data sets (Hibler and Tucker, 1977; Hibler, 1980). Rather abrupt changes were seen in the spectral nature of ice kinematic parameters using data from 1975 and 1976 (Denner, Lewis, and Chase, 1984). These changes occurred in July, October, and December 1975 and in April 1976. During the summer, the kinematic parameters were characterized by high frequency, large amplitude fluctuations. Fall and spring fluctuations were smaller in amplitude and lower in frequency. Winter fluctuations were seen to be infrequent and the smallest in amplitude. These changes and the times at which they occurred are clear indications of how freezing and thawing of the ice pack play a role in determining the spectral nature of ice motion. Thus, the spatial and temporal variations of ice pack rigidity and mobility might be readily monitored in real-time using the spectral characteristics of ice motion.

In this paper, we use specific characteristics of ice speed time histories as an index of the rigidity and mobility of pack ice. Ice rigidity and mobility during 1975 and 1979 were studied using this index, and the results show some interesting differences between the two years. Additional environmental data are considered in determining those factors causing the differences. These factors appear to be 1) a greater mechanical breaking up of the ice pack by summer winds in 1975 than in 1979 and 2) a greater percent of thin ice in the fall of 1979 than in 1975. As a result, the larger amplitude, higher frequency ice motion similar to the summer of 1975 was seen to persist into November in 1979.

II CALCULATION OF ICE RIGIDITY/ MOBILITY

Examples of the seasonal differences in the characteristics of ice speed in the Arctic Ocean are shown in Fig. 1. The summer motion is similar to free drift conditions, with speeds varying at the inertial period (12 to 12.5 h). The oscillations of the speed are caused by the variations of the compaction of the field of ice (traveling faster as the ice field diverges and slower as the ice field converges). The free drift force balances are primarily between the inertia of the ice and the Coriolis effect. This balance is altered during non-summer months when the internal ice stresses replace the ice inertia in importance. Thus, the fall, winter, and spring ice speed characteristics shown in Fig. 1 do not show the higher frequency oscillations, and the magnitudes of the speeds are more damped out. This is especially true for winter.

The above example is for pack ice in the Beaufort Sea which had an average thickness of 2.3 m in August 1975. Under the same wind conditions, one would expect thinner ice (less mass) to have larger speed amplitudes and thicker ice (greater mass) to have smaller

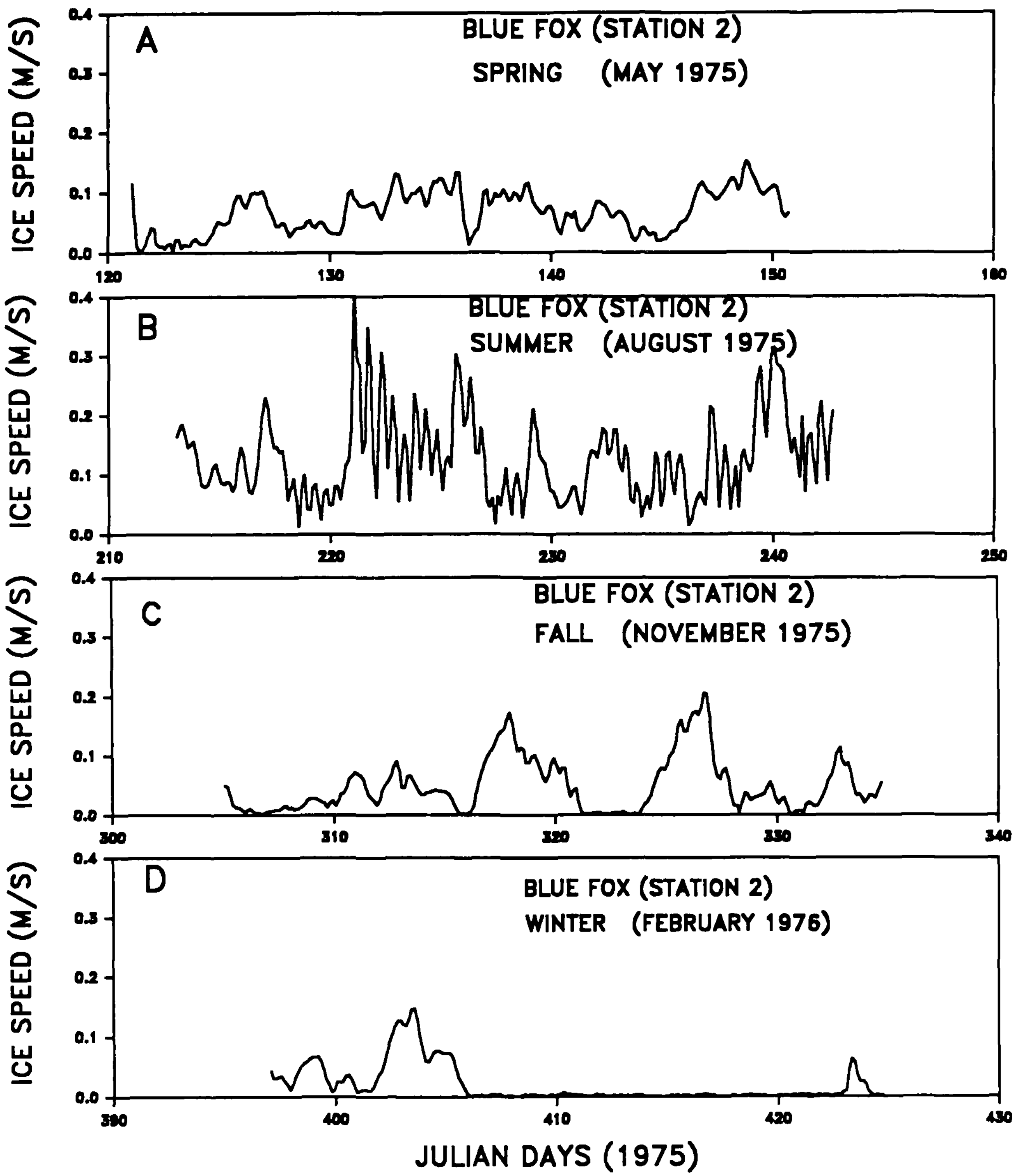

Fig. 1. Examples of arctic ice speed variations for spring, summer, fall, and winter. These data were collected in the Beaufort Sea during 1975 and 1976.

amplitudes. Thus, in the development of an ice rigidity and mobility parameter, one can not specifically deal with ice speed if the analysis scheme is to be applied over large areas. One must actually deal with momentum per unit area (i.e., speed x density x thickness of the ice). Determining ice thickness using observed ice motion and wind data has been discussed recently in the literature (Feldman, 1986; Lewis, Crissman, and Denner, 1986).

Parameterizing the Speed Time Histories. Parameterizations of the ice speed characteristics were developed using the test data from the Beaufort Sea (Fig. 1). The parameters used are mean ice speed U, variance V, and an e-folding time scale, e_T. The mean ice speed is used as an indicator of the energy at lower frequencies while the variance is a measure of the total energy. The e-folding parameter is the time lag at which the autocorrelation of the ice speed drops to e^{-1}. It is a measure of the persistence of the ice speed. Thus, e_T is longer when most of the energy is concentrated at lower frequencies, and shorter when there exists significant energy at higher frequencies (for example, 12 to 13 hours).

The data from Fig. 1 are used in 6 day time periods to calculate e_T, U and V. This was done at three hour intervals, and the results are shown in Figs. 2 and 3. The differences between the summer and the more viscous fall, winter, and spring ice regimes are readily seen in these figures. The mean ice speeds and variances decrease from summer to winter and then increase from spring to summer. However, the higher U's and V's are associated with shorter e-folding times due to the energy associated with the higher frequency inertial oscillations of the ice speed (the motion is more variable with time). As the speed and variance decrease, the motion becomes more persistent in time, and the e-folding time increases. Summer e-folding times are of the order of only 15 hours or less while non-summer e-folding times can be as high as 80 hours.

An interesting aspect of the data shown in Figs. 2 and 3 is the differences of the two transitional seasons, fall and spring. Although these seasons have similar e-folding times and mean ice speeds (Fig. 2), their total energy contents (speed variance) are quite different. The fall variance is considerably larger than the spring variance, a reflection of the less solidly frozen conditions and smaller ice concentrations

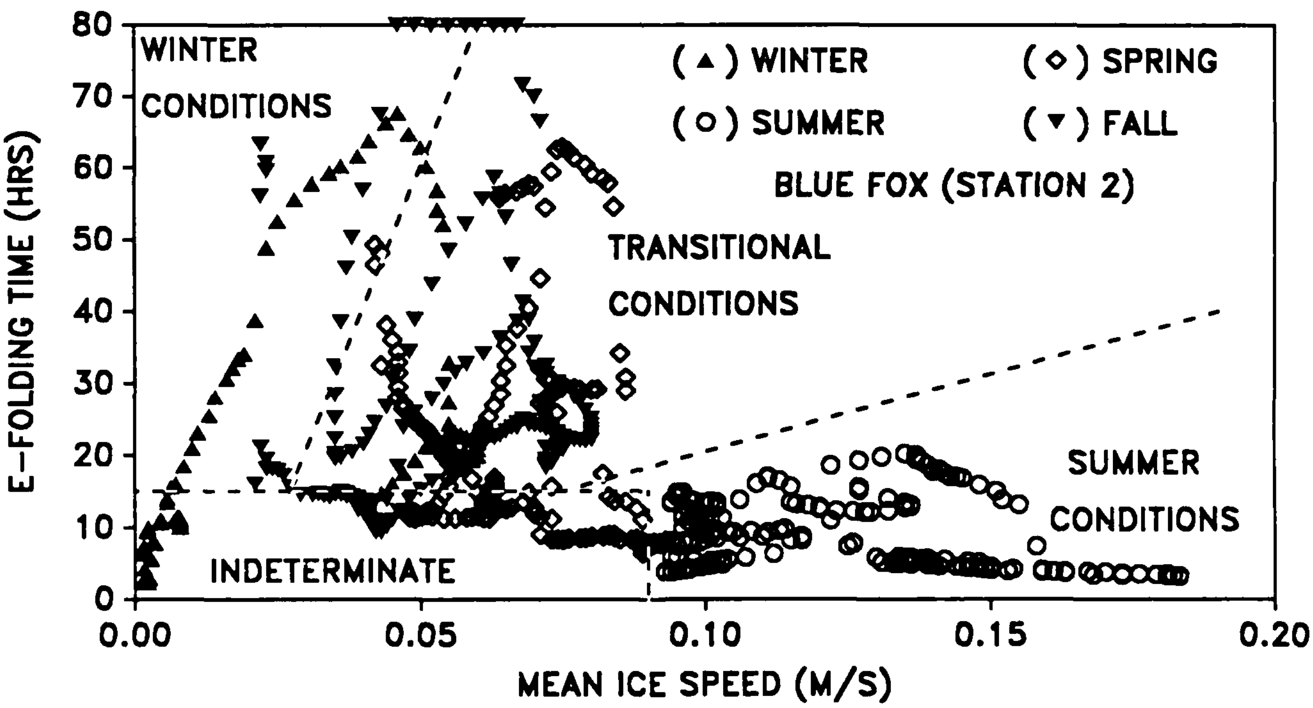

Fig. 2. Two-dimensional characteristics of six day mean ice speed and corresponding e-folding time of the speed time history. Dashed lines represent divisions into summer, transitional, and winter regions.

of fall. This difference is used to produce a rigidity parameter which can differentiate between fall and spring situations.

Calculation Scheme. The time history of our rigidity/mobility parameter is calculated in the following manner. Six days of momentum data at three hour intervals are used to calculate U, V, and e_T. In the six day span, one would expect at least one atmospheric front to pass through the area where the ice drift is being observed. This would allow the study area to be energized by the wind field and allows one to differentiate between situations such as summer free drift conditions under light winds and the more rigid fall conditions under stronger winds.

Using the e_T versus U space, we give the rigidity parameter R a value of 1 if the point value is in the summer regime region, 3 if it falls in the fall/spring region, or 5 if it falls in the winter region (see the regions in Fig. 2). We then use the e_T versus V space and add 1 to the R value if the point falls in the summer region, 3 if it falls in the fall region, or 5 if it falls in the winter/spring region (see the regions in Fig. 3). Dividing this value of R by 2 gives a rigidity/mobility parameter that varies from 1 (least rigid) to 5 (most rigid).

The computer code that performs these calculations steps ahead every three hours and calculates a new R value. It was found that there were some instances in which the point in our two-dimensional space would slightly cross the boundary from one region to another and then jump back. The result is a spike in the rigidity parameter which is not a true indicator of a change in conditions. To prevent such spiking, the R value used is actually a weighted average parameter. The average uses the present R value with a weight of 1.0 and previous R values, the n^{th} value having the weight of 2^{-n}. In addition, there were instances in which the data points fell into the indeterminable range of low U and V values along with a low e_T value (lower left of Figs. 2 and 3). These situations result from not having a significant wind event during the 6 day period. In this case, the previously calculated R value is used.

Thus, our final rigidity/mobility parameter reflects average conditions over a six day period, with some memory for previous conditions. We can define a range of conditions using this parameter in the following way. Summer-time conditions will have an R value

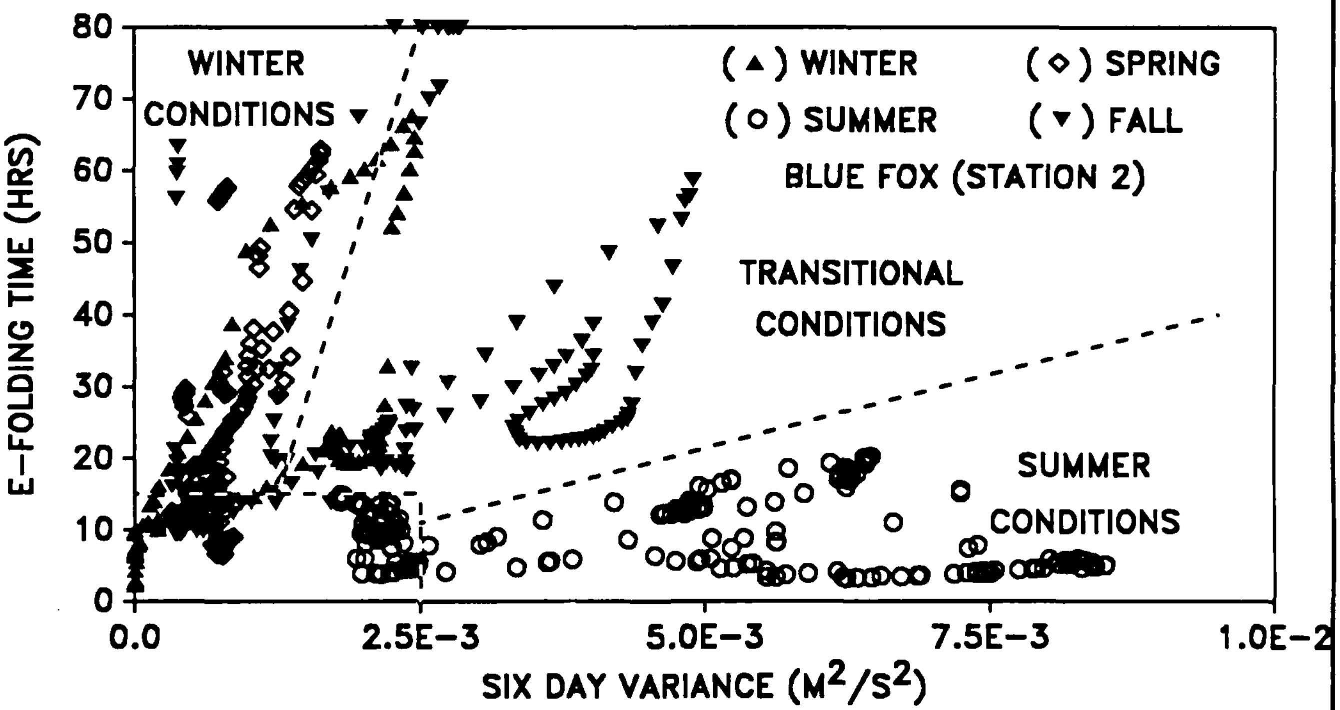

Fig. 3. Two-dimensional characteristics of six day ice speed variance and corresponding e-folding time of the speed time history. Dashed lines represent divisions into summer, transitional, and winter regions.

of 1 to 2. The fall season will have R values that range from 2 to 4. Spring values will tend to have slightly higher values, such as 3 to 5. Winter lock-up conditions will have the highest values of R at 4 to 5.

Ice Rigidity/Mobility, 1975 and 1979. Sample cases of ice pack rigidity variations were constructed using additional buoy drift data from the Beaufort Sea in 1975 (Thorndike and Cheung, 1977) and in 1979 (Thorndike and Colony, 1980). The rigidity results point out the inter-annual differences that can occur in arctic ice pack conditions. During 1975, summer-time ice rigidities were detected first in late June (Julian day 185, Fig. 4a). Summer conditions persisted from about Julian days 220 through 270, with several periods of apparent refreezing and thawing. As one would expect with the eventual melting of all of the thinner ice, the least rigid, most mobile conditions occurred during late August and September, days 240-270. After September, the pack ice went through its fall freeze-up, with rigidities reaching as high as 4.0. A short period of thawing appears to have occurred during late October (Julian days 295-300). Afterwards, the ice pack tended toward greater rigidities, with intermittent periods of thawing to some degree.

In 1979, the calculations for the Beaufort Sea indicate relatively rigid conditions during late spring which lasted until late August (Julian day 235, Fig. 4b). The exception to this trend was the summer-like rigidities in July 1979. The motion characteristics of 1979 indicate relatively solidly frozen ice pack conditions during August, with rigidities reaching as high as 4.0 (see day 247, Fig. 4b). We see that the fall freeze-up in 1979 occurred at almost the exact same time as in 1975, around Julian day 270. However, November 1979 tended to be less rigid than November 1975. Some summer-like conditions were detected as late as the end of November in 1979. These November 1979 conditions existed even though air temperatures were relatively low (-20° to -30° C).

To account for the differences in ice pack rigidity between the two years, we will consider several sets of environmental data. These include the 1975 and 1979 500 mb atmospheric pressure distributions, surface air temperatures, and extent of ice coverage by 40° longitudinal sectors.

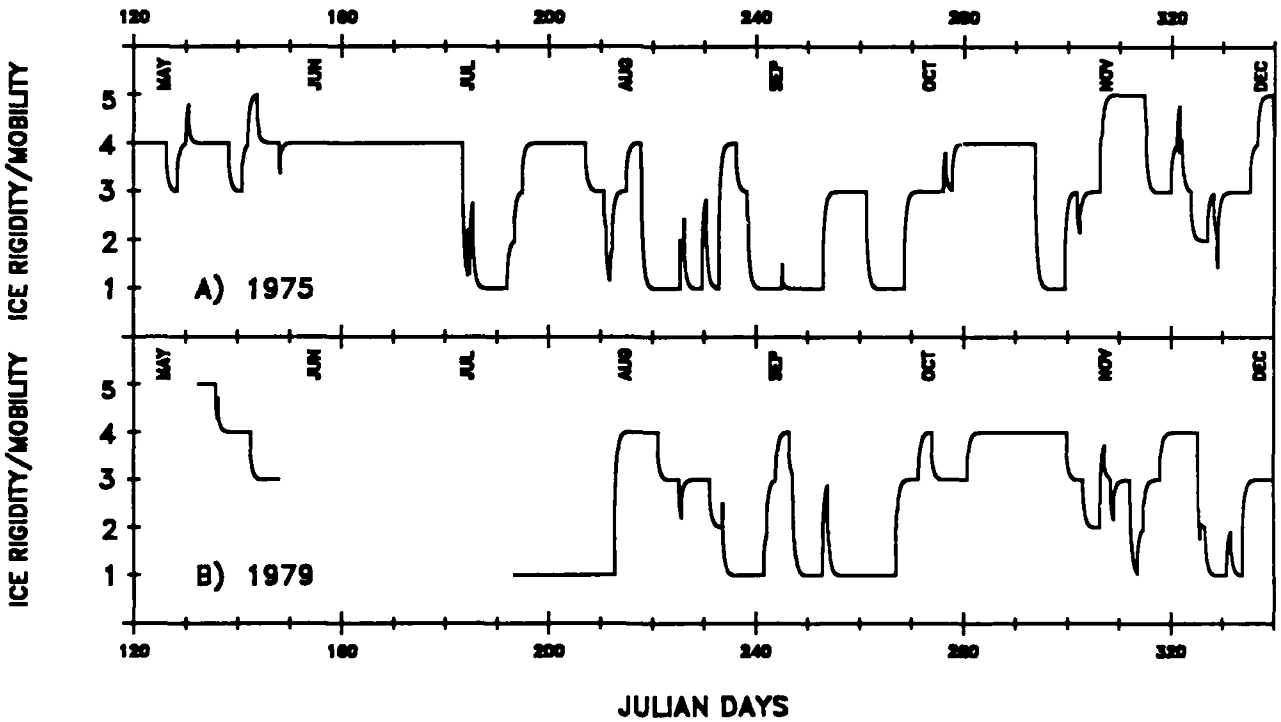

Fig. 4. Ice pack rigidity/mobility values in the Beaufort Sea for May-November 1975 (a) and 1979 (b).

III ICE COVERAGE CHARACTERISTICS, 1975 AND 1979

We first consider two data sets pertaining to ice extent and concentration. One set is the ice edge maps produced at weekly intervals by the Naval Polar Oceanography Center in Suitland, Maryland. These maps give an indication of the geographical distribution of pack ice as well as percent of ice coverage. The second set is actual ice coverage (km²) by 40° longitudinal sectors radiating from the North Pole.

Ice Extent. We define the edge of the pack ice as that line which encloses ice concentrations of greater than 75 percent. At the end of May 1975, this ice edge extended through the Bering Straits and well into the Bering Sea (Fig. 5a). The line of 0°C average air temperature reflects this ice edge pattern, with below freezing air temperatures found as far south as 61° N in the Bering Sea. By the end of August (Fig. 5b), both the ice limit and 0°C average air temperature lines had moved northward to about 70° to 71°N. Although the Bering Straits and the far eastern Beaufort Sea were ice free, most of the northern coast of Alaska was still ice-bound. This type of situation persisted into the end of September (Fig. 5c), with the ice edge moving somewhat southward. Toward the end of October 1975 (Fig. 5d), few regions of open water could be found in the western Arctic Ocean while the Bering Straits remained ice free. The 0°C average air temperature line had moved well south into the Bering Sea.

The ice extent conditions at the end of May 1979 (Fig. 6a) did not reach as far south as in 1975. Open water was found in the Bering Straits and in the far eastern Beaufort Sea. Moreover, the southern limits of the line of 0°C average air temperature reached only as far south as 69°N. By the end of August (Fig. 6b), the ice extent had receded further to 72-74°N, and the entire coasts of Alaska and eastern Siberia were practically ice free. The 0°C average air temperature was found well north of the ice edge at this time. Conditions had changed little by the end of September 1979 (Fig. 6c) except for the fact that the 0°C air temperature line had moved southward along the Alaskan coast and crossing eastern Siberia. However, by the end of October (Fig. 6d) the ice edge had moved southward and only the northern sector of the Bering Straits was ice free.

Areal Ice Coverage. We now consider monthly variations in the actual ice coverage for two 40° longitudinal sectors radiating from the North Pole (Walsh and Johnson, 1979b). The first sector is from 120°W to 160°W and will be referred to as the Beaufort Sea sector (see Figs. 5 and 6). The second sector, from 160°W to 160°E (Figs. 5 and 6), will be referred to as the Chukchi Sea sector. Table 1 shows the ice coverages for these two sectors for May through December, 1975 and 1979.

In the Chukchi Sea sector, the May 1975 coverage was 18491 km², and this was reduced to 12375 km² by August. By December 1975, the coverage had increased to 21648 km². During 1979, the May coverage was initially 16852 km². The minimum coverage was 11242 km², but this occurred in September 1979, not August. In December, the coverage had increased to 21043 km², similar to that of 1975.

The Beaufort Sea sector had a coverage of 14366 km² in May 1975, but this dropped to ~13200 km² during June, July, and August. By September 1975, the coverage was back to the levels of May, and continued to increase slightly through December to 14773 km². The variations in 1979 were considerably different. A maximum coverage of 13926 km² was found in June 1979, and this decreased to only 10879 km² by September. However, in the following month, October 1979, the coverage increased significantly to 14608 km². This coverage continued to increase slightly to 14740 km² in December 1979.

IV ATMOSPHERIC PARAMETERS

The influence of the patterns of synoptic scale highs and lows have been related to ice extent and advection near the outer ice edge limits (Walsh and Johnson, 1979a; Niebauer, 1980; Overland and Pease, 1982). This prompted us to study the atmospheric patterns for 1975 and 1979 to enhance our understanding of the differences in ice conditions for those two years. The 500 mb level atmospheric circulation can be separated into three ranges of scale. These are the synoptic scale, the longwave scale (SL), and the polar vortex scale. The atmospheric longwave pattern is of interest here since it determines

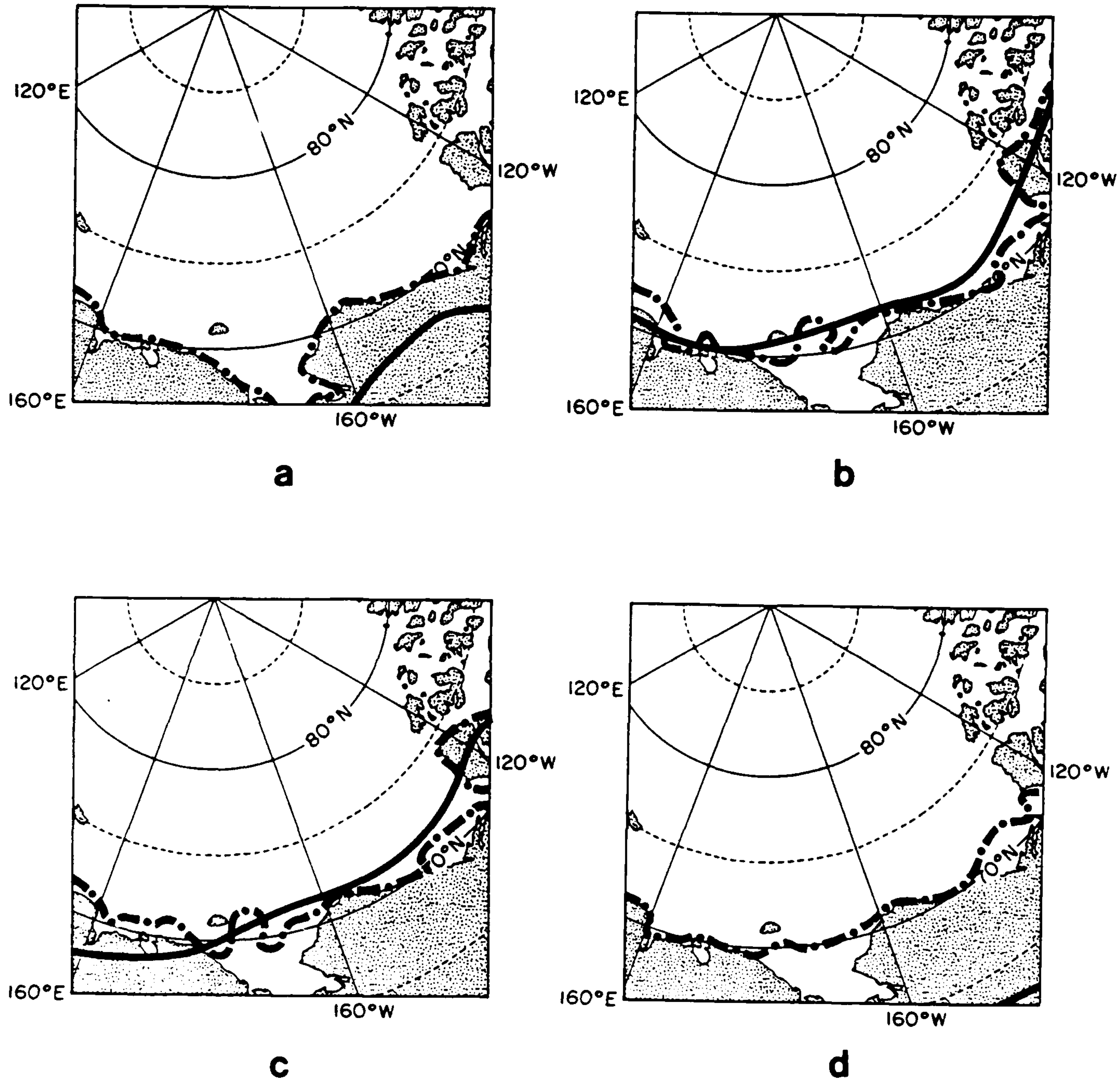

Fig. 5. Ice edge extent and location of the 0° C average air temperature line for the western Arctic Basin in 1975: end of May (a), end of August (b), end of September (c), and end of October (d).

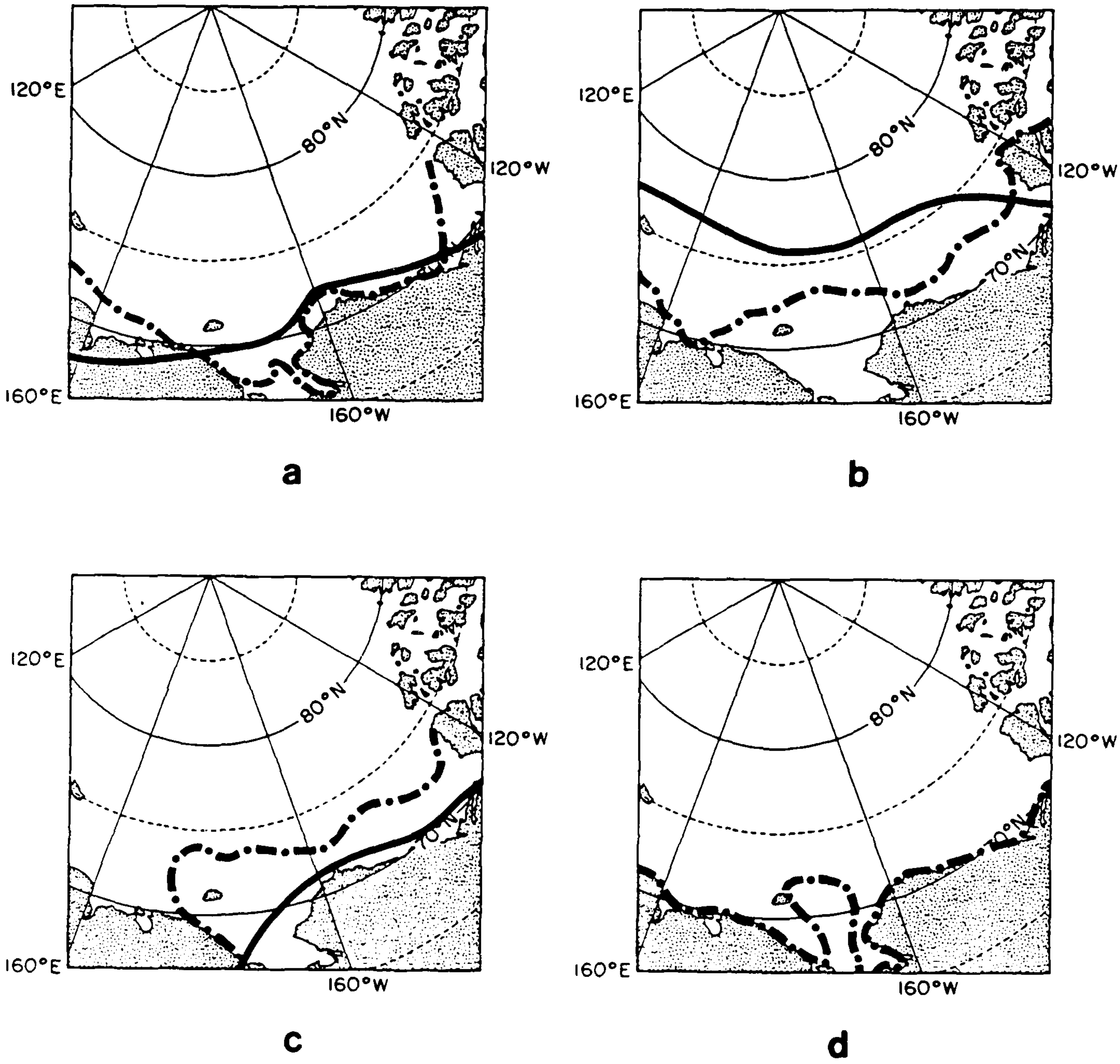

Fig. 6. Ice edge extent and location of the 0° C average air temperature line for the western Arctic Basin in 1979: end of May (a), end of August (b), end of September (c), and end of October (d).

the steering of synoptic scale storms (storm tracks).

A pattern separation program used by the Navy at the Fleet Numerical Oceanography Center (FNOC) produces a 500 mb SL pattern on a daily basis (see comments in nomenclature section). The monthly distribution patterns of SL high and low centers were developed by direct summing of the count of centers from the daily FNOC pattern-separated 500 mb SL analyses. These daily high and low centers were assigned to the nearest grid point in a 63x63 northern hemisphere grid (Englebretson, 1985). The 500 mb SL center intensities, D, are given as the deviation (meters) of the 500 mb height at the center of the high or low from the standard atmosphere 500 mb height. The cumulative value of the highs and lows can be used as a proxy for the surface pressure anomaly and, therefore, the wind and weather conditions. This proxy relationship results from the barotropic nature of the atmospheric long-waves; minimum tilt but large vertical extent (Palmer and Newton, 1969). Thus, the nature of the high or low cells at 500 mb will be reflected at the surface with, at most, only a slight horizontal displacement. This was verified for 1975 and 1979 using monthly mean sea-level pressure charts for July through November.

The relationship between the D values for 1975 and 1979 are shown in Fig. 7 for the western Arctic Ocean (north of 70°N and from 90°W through 180° to 90°E). The vertical scale in the figure is the sum of the 500 mb SL high and low center D values for a given month. A marked reduction of atmospheric low activity in the Beaufort Sea during 1979 is clearly indicated by the more positive values of D as compared to the negative values of 1975. The only exception to this reduction occurred in July and November.

Additional details on the number and intensity of 500 mb SL centers are provided in Table 2. The total monthly D values for each center type, number of low and high centers, and average intensities are listed for May through November, 1975 and 1979. The more frequent high activity in the Pacific sector during 1979 is again indicated, with the July and November exceptions shown. Note that, during 1979 in the Pacific sector, the monthly count of highs equals or exceeds that of lows in all but July and November. The total count

Total Ice Coverage (km^2)

	Chukchi Sea Sector		Beaufort Sea Sector	
	1975	1979	1975	1979
May	18491	16852	14366	13794
June	15543	14850	13156	13926
July	13233	13266	13167	13343
August	12375	11297	13255	11440
September	13530	11242	14355	10879
October	15257	14245	14641	14608
November	17853	17303	14553	14696
December	21648	21043	14773	14740

Table 1. Monthly coverages of sea ice for two 40° longitudinal sectors in the arctic.

Month, 1975	Low		High	
May	$\frac{-253}{6}$	(-42)	$\frac{364}{6}$	(61)
Jun	$\frac{-0}{0}$	(0)	$\frac{29}{1}$	(29)
Jul	$\frac{-529}{9}$	(-59)	$\frac{235}{8}$	(29)
Aug	$\frac{-549}{12}$	(-46)	$\frac{64}{3}$	(21)
Sep	$\frac{-374}{6}$	(-62)	$\frac{175}{7}$	(25)
Oct	$\frac{-885}{15}$	(-59)	$\frac{0}{0}$	(0)
Nov	$\frac{-208}{3}$	(-69)	$\frac{287}{7}$	(41)
1979				
May	$\frac{-287}{6}$	(-48)	$\frac{511}{13}$	(39)
Jun	$\frac{-47}{2}$	(-24)	$\frac{285}{15}$	(19)
Jul	$\frac{-620}{18}$	(-34)	$\frac{258}{10}$	(26)
Aug	$\frac{-66}{2}$	(-33)	$\frac{33}{2}$	(17)
Sep	$\frac{-245}{7}$	(-35)	$\frac{531}{19}$	(28)
Oct	$\frac{0}{0}$	(0)	$\frac{302}{14}$	(14)
Nov	$\frac{-195}{4}$	(-49)	$\frac{77}{3}$	(26)

Table 2. Total D value, number of low or high 500 mb SL centers, and average intensity (D value) by month for the western Arctic Basin (90° W through 180° to 90° E) north of 70°N for May through November of 1975 and 1979.

for 1979 is 76 highs and only 39 lows, while the 1975 totals are 32 highs and 51 lows.

V DISCUSSION

There are some interesting differences between the calculated Beaufort Sea ice pack rigidities of 1975 and 1979. Firstly, August of 1975 was a month of relatively less rigid conditions while August of 1979 showed considerably higher rigidities. This seems to be somewhat of a paradox, considering the greater extent and coverage of the ice pack in 1975 and the northern reach of the 0°C average air temperature in 1979 (Figs. 5b and 6b; Table 1). Such conditions would seem to imply a more solidly frozen ice pack in August 1975 than in 1979, similar to the rigidities found during July of the two years. A second interesting aspect of the rigidity data is the persistence of non-rigid conditions into November 1979. This occurred even though the ice extent and coverage was near their maximums and recorded air temperatures in the Beaufort were in the range of -20° to -30°C. These conditions were similar to those of November 1975, yet considerably more solidly-frozen conditions were detected during that year.

To show some of the effects of the differences in rigidities of the two years, we consider ice divergence in the central Beaufort Sea. These were calculated using ice station drift data, 3 or 4 stations encompassing a region of ~8.3×10^3 km^2 in 1975 (Thorndike and Cheung, 1977) and 4 or 5 stations encompassing a region of about 27×10^3 km^2 in 1979 (Thorndike and Colony, 1980). The less rigid conditions of August 1975 (Fig. 8a) allowed the ice pack to oscillate between convergence and divergence at an approximately 12.5 h period (the inertial period). The result is the opening and closing of leads on a regular basis, and this allows a significant interplay between the ocean and atmosphere (fluxes of momentum and latent and sensible heat). In August 1979 (Fig. 8b), such an interplay was not occurring in the Beaufort Sea due to the more rigid ice pack conditions. However, the November 1979 ice divergence variations (Fig. 8d) were quite similar to those seen in August 1975, indicating a considerable amount of air/sea interaction during the fall of 1979.

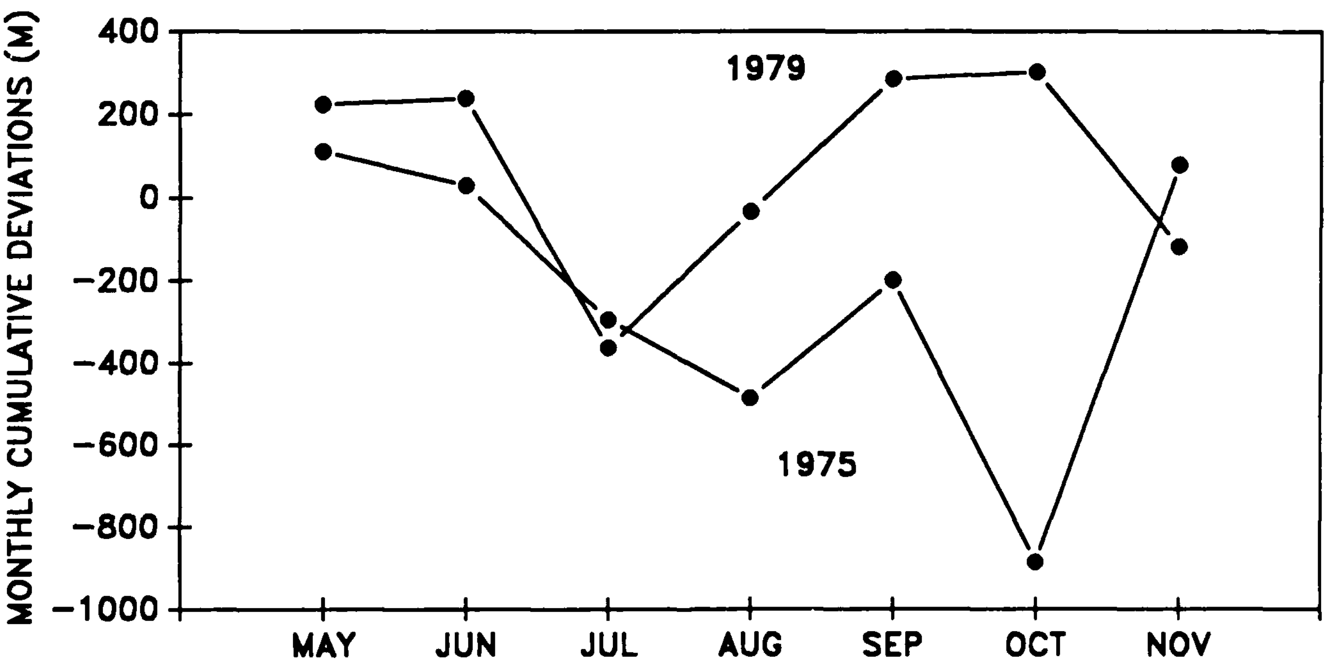

Fig. 7. Variations in the monthly cumulative 500 mb SL high and low center D values for 1975 and 1979 for the Pacific (Beaufort) sector of the arctic.

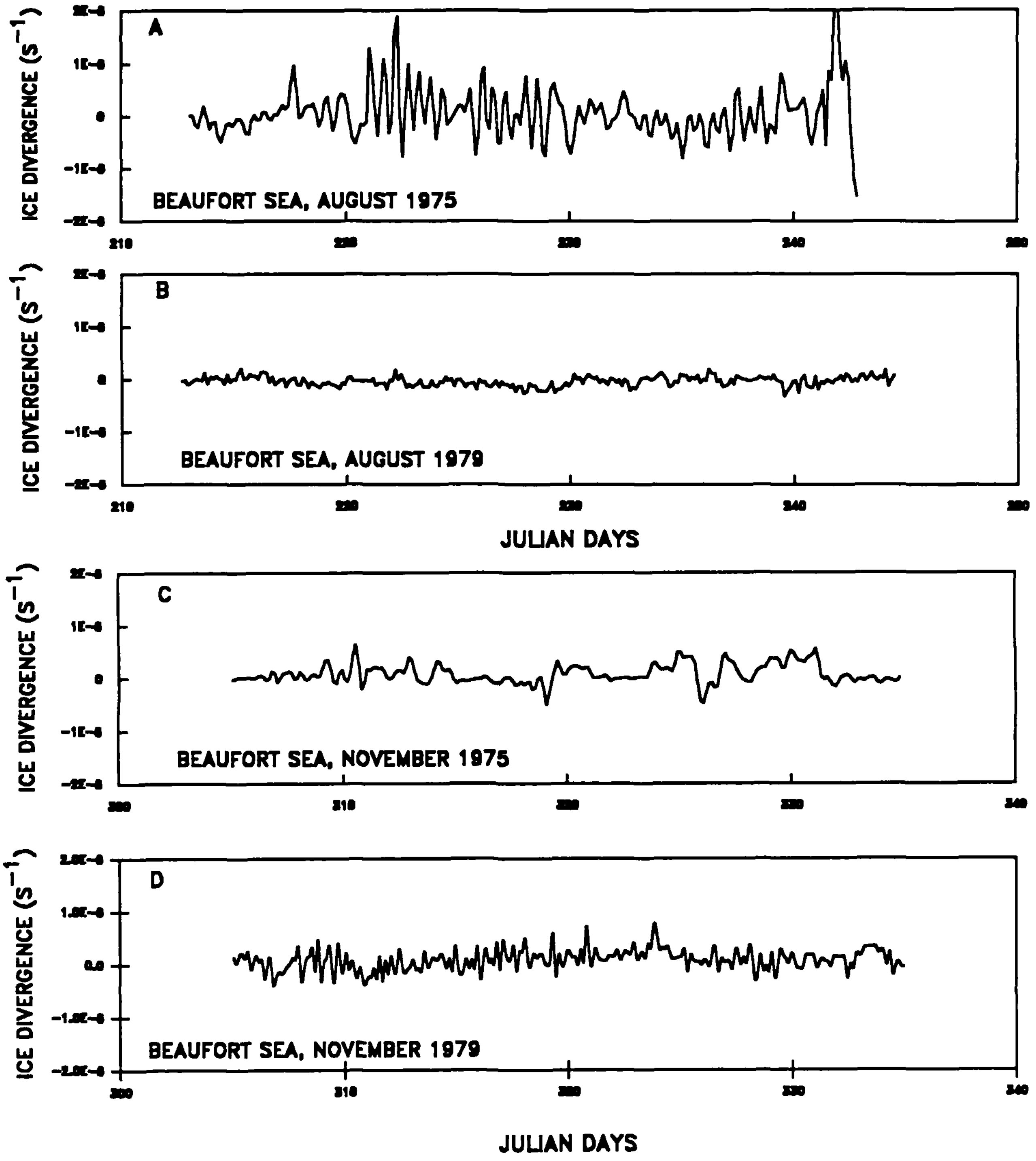

Fig. 8. Ice pack divergence time histories in the Beaufort Sea during August 1975 (a), August 1979 (b), November 1975 (c), and November 1979 (d).

The November 1975 ice divergence variations (Fig. 8c) did not show the high frequency oscillations seen in November 1979.

If we consider the thermodynamics of the ice/atmosphere, the data from the two years would imply less rigid conditions through July 1979. The 0°C average air temperature line was further north and there were more atmospheric lows in June and July 1979. If we consider the rigidities for July of the two years, we do see higher values in 1975 than in 1979. However, this trend reverses in August of the two years. We attribute this to the continued mechanical breakup of the ice pack by the summer storms of July and August 1975. Table 2 and Fig. 7 indicate that the storm activity during July through November 1975 was significantly more intense than for the same period in 1979 (monthly average 500 mb D values were -59 in 1975 versus -36 in 1979). Also, storm activity during August through October was greater in 1975 (33 lows) than in 1979 (9 lows). These more intense and frequent lows have a number of effects on arctic atmospheric processes. These include a well mixed or neutral atmospheric boundary layer without inversions, strong surface winds, and alternating periods of cloudy and nearly clear skies. The expected results of these conditions would be increased wind stress on the ice and active ice pack motion. This would enhance the mechanical breakup of the ice pack. Moreover, there would also be an increase in the air/ice heat exchange as a result of greater turbulence in the planetary boundary layer. Thus, the heat loss from the ice pack during the colder air but stronger wind conditions of July and August 1975 could well have been greater than that of the warmer air but lighter wind conditions of the same period in 1979. If this were true, the result would be a weaker ice structure in 1975, and this would further contribute to the breakup of the ice pack.

The atmospheric conditions for August of 1979 were those typically under high pressure systems, with 19 high centers and 7 low centers (Table 2). Arctic areas under high pressure systems generally exhibit light winds and overcast, low clouds (stratus). Moreover, there is typically a strong, surface-based temperature inversion. This alone can dramatically reduce fluxes of momentum and heat between the ice and the air. Thus, the evidence suggests that much less mechanical breakup of the ice pack took place into August of 1979, and the Beaufort was relatively frozen under the influence of the high pressure systems. Less rigid conditions eventually occurred in late August and September 1979.

We now turn our attention to the rigidity variations found in November of the two years. In 1975, the average rigidity value was approximately 4.0 while it was about 2.5 in 1979. As was noted before, the ice divergence during November 1979 was similar to that of August 1975. Considering the cold air temperatures of November 1979, such opening and closing of the Beaufort ice pack must have resulted in tremendous fluxes of sensible and latent heat from the ocean to the atmosphere.

The ice extent and coverage data indicate that near-maximums had been reached in November of both years. Yet the 1979 ice was free enough to have relatively large mean speeds and variances with significant energy at higher frequencies. This implies the horizontal motion of inertial circles, which is readily confirmed by studying the ice trajectory data. The primary difference between the two years is the fact that minimum ice coverage in the Beaufort sector in 1979 was 21% less than that of 1975 (Table 1). Moreover, this minimum ice coverage occurred in September while the minimum in 1975 occurred in June-August. By October, both years had approximately the same amount of ice coverage. Thus, the coverage data give an increase in the Beaufort sector of 3729 km² during October 1979 (124 km² per day). This implies a tremendous production of thin ice during a period dominated by 500 mb highs. The increase in cyclonic activity from October to November 1979 (Fig. 7 and Table 1) would then be driving an ice pack which, although extensive in coverage, was weak due to being 26% new ice. As a result, the older, thicker ice had the ability to move freely by crushing the thin, new ice.

VI SUMMARY AND CONCLUSIONS

An ice rigidity/mobility parameter has been developed to determine how solidly arctic pack ice is frozen. The use of an ice pack rigidity/mobility parameter using observed ice motion has many benefits. Not only can ice conditions be monitored remotely, but one is not forced to quantify how the ice is to respond to various dynamic and thermodynamic forcings.

Arctic ice drift data have been used to determine variations in ice pack rigidity and mobility in the central Beaufort Sea. Of the two years studied, a number of differences were noted. Specifically, summer-like rigidities were detected in August 1975 and November 1979 while more rigid conditions were detected in August 1979 and November 1975. To understand these differences, we considered ice extent and coverage data, air temperature data, and 500 mb large scale pattern (SL) data. As with the ice rigidity parameters, these data sets were considerably different for the two years. From the atmospheric data, we infer that the wind forcing during the summer of 1979 was considerably less than that of 1975. This, and the dominance of atmospheric high characteristics, appears to have resulted in the more solidly frozen conditions of August 1979. Also, minimum ice coverage in the Beaufort Sea region during 1975 occurred during June through August while the minimum for 1979 did not occur until September. Moreover, the minimum ice coverage during 1979 in the Beaufort Sea sector of the arctic was 21% less than the minimum in 1975. By November of both years, the total ice coverages were approximately the same. This indicates a larger percent of thinner ice for November of 1979 than for 1975, the likely cause of the less rigid conditions detected.

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