Wintertime Arctic Sea Ice Extremes and the Simultaneous Atmospheric Circulation

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ABSTRACT

Twenty-five years of Arctic sea ice data have been used in conjunction with data from the lower atmosphere (the surface and 700 mb) to establish some concurrent general circulation relationships. Five January climatologies for both maximum and minimum sea ice areas over the entire north polar cap were first determined. The differences between the two sets of mean atmospheric patterns were then found. The student’s t-test was used to establish the statistical significance.

The results indicate that the wintertime (January–February) atmospheric circulation in the Pacific tends to be weaker during heavy ice conditions while the differences in the Atlantic are not as significant. These results are compared with GCM (general circulation model) simulations with similar maximum ice conditions.

Extremes in Arctic sea ice during January were also examined locally. Although extremes in sea ice between Greenland and northern Europe were not found to be strongly associated with concurrent atmospheric changes, above-normal sea ice in the Bering Sea and the Davis Strait were found to be strongly associated with simultaneous surface northerly flow locally. The regional analysis also showed that above-normal ice is associated with intensification of the closest major low pressure centers and 700 mb troughs.

1. Introduction

At the present time, the circulation of the Arctic Ocean and of the Arctic atmosphere are not well understood. Yet some have postulated that the climates of the polar regions have a significant relation to global climate (e.g., Ewing and Donn, 1956).

Sea ice, the interface between ocean and atmosphere in the Arctic through much of the year, plays a potentially important role in this context. It acts to reduce the heat flux to a cold atmosphere from a relatively warm ocean below and also has a significant effect on the radiation budget in polar regions. Thus, to establish and understand the importance of this feature of Arctic climate, a careful examination of the interactions between sea ice, ocean and atmosphere is crucial. More specifically, we need to better understand the effect of Arctic sea ice variations on the circulation of the atmosphere and ocean, and determine how these variations in sea ice are caused.

One way of better understanding the effect of sea ice on the atmosphere is to develop a model of the atmospheric general circulation and then see how this model reacts to different ice boundary conditions. A number of researchers have tried this approach, including Williams et al. (1974), Warshaw and Rapp (1973) and most recently Herman and Johnson (1978). The inherent problem in this approach is that a non-interactive ice-atmosphere system is assumed, i.e., the atmosphere is not allowed to change the ice cover. Depending on the time scales used, this can have the disadvantage of modelling an unrealistic situation as changes in the atmospheric circulation do have a significant impact on current and subsequent sea ice anomalies on the order of a few months or less as found by Walsh and Johnson (1979b) and by Lemke et al. (1980).

Recent attempts to model sea ice have provided some insight into what factors might be important in determining sea ice extent. By treating sea ice as an elastic-plastic medium, and by using geostrophic wind and ocean currents as input, Hibler (1979) developed a model to simulate Arctic sea ice over a seasonal cycle. This numerical simulation was fairly successful in reproducing many of the observed velocity and thickness characteristics of the Arctic ice cover. Parkinson and Washington (1979) developed a similar sea ice model (in that winds and ocean currents were used as input and a seasonal equilibrium was obtained) although their ice-ice interactions contained more parameterizations and were not dependent on ice thickness. A linear stochastic dynamic model of Arctic and Antarctic sea ice based on white noise atmospheric forcing, local stabilizing relaxation, lateral diffusion and advection was constructed by Lemke et al. (1980).
Their model was capable of satisfactorily explaining the statistical space-time structure of sea ice anomalies for both regions; and within their model hierarchy, attempts to fit the data to still simpler models were unsuccessful, indicating lateral diffusion and advection are also important in determining sea ice variations. None of these modeling efforts, however, included feedback effects of ice on atmosphere or ocean, but the stage has begun to be set for more interactive ice-atmosphere-ocean models.

On the observational side, attempts have been made to establish some ice-atmosphere relationships, but until recently (i.e., before aircraft and satellite data became available) these attempts were hindered due to a lack of continuous sea ice observations, particularly away from the ice edge and in more remote polar areas (e.g., the Siberian Arctic). Because of these deficiencies, sea ice data analyses were quite restricted spatially and researchers tended to concentrate on local variations of sea ice to establish ice-atmosphere relationships. For example, Schell (1955) found summer sea ice in the Greenland-Barents Sea area a good predictor of subsequent sea surface and air temperatures to the south. Later, Schell (1970) conducted a similar study, but also included precipitation over northern Europe and western Asia as a predictand. Miles (1974) found a high positive correlation between the strength of the winter atmospheric circulation in the North Atlantic and the subsequent severity of sea ice off Newfoundland. Surface wind direction and especially air temperatures at Barrow, Alaska were found by Rogers (1978) to be primary parameters to consider in forecasts of the varying sea ice conditions in the Beaufort Sea. Rogers and van Loon (1979) found sea ice anomalies in the North Atlantic to be an integral part of the Greenland-Northern Europe “seesaw”; whereby the seesaw mode of unusually cold winters in Greenland and warm winters in northern Europe tended to be followed by more than normal sea ice in the Davis Strait and near Newfoundland, and less than normal sea ice in the Baltic Sea.

With the advent of satellite observations in the mid '60s came the possibility of extending sea ice statistical analyses to the entire polar region. While ten years of data are not really enough to attach much significance to year-to-year variations, the quality of the satellite-derived data set was much improved over what was previously available. The data used in this study consist of satellite-derived and aircraft data, which thereby increased the length of the study period from 10 to 25 years. Walsh and Johnson (1979b, hereafter WJ-II) used the same data set to evaluate correlations between the dominant modes of sea ice and atmospheric variability on monthly and seasonal time scales. They found that the surface temperature field showed the strongest correlation to sea ice fluctuations and the asymmetry about the zero-lag cross correlation, for all the atmospheric fields they correlated with sea ice, showed a stronger tendency for atmosphere to force sea ice (i.e., atmosphere to cause sea ice changes) on the order of a few months or less.

This research differs from WJ-II in that this analysis examines the simultaneous atmospheric fields during wintertime extreme ice conditions. First, these extreme ice conditions will be examined over the entire polar cap, then these results will be compared with recent model simulations. Finally, sea ice extremes will be examined locally and it will be shown that the corresponding atmospheric state exhibits a coherent and statistically significant pattern.

2. The data sets and method of analysis

a. The data sets

Monthly Arctic sea ice data from 1953 to 1977, inclusive, were digitized onto a set of rectangular grids centered over the North Pole. The data were compiled from a variety of sources; a complete list and the method of compilation are fully described by Walsh and Johnson (1979a, hereafter referred to as WJ-I).

A Cartesian coordinate system was used to describe this rectangular grid where the grid spacing was 60 nl mi in both the x and y directions. The value (on tape) at a given grid point for a particular month and year describes the fraction of ocean area covered by ice. There were several factors that led to uncertain values at the beginning of the study period, but overall the quality of the data improved with time. The data were from aircraft measurements principally, but also include the more recent satellite data from the late '60s.

The daily atmospheric data were obtained from NCAR and a number of other sources (cf. Jenne, 1975). Monthly means were used in this study, and the analysis of the atmospheric data is restricted to the period 1953–77 so that it is consistent with the sea ice data.

The atmospheric data include sea level pressure from 20–90°N, surface temperatures north of 60°N, and 700 mb heights and temperatures north of 20°N. The grid spacing is 5° in both longitude and latitude.

b. The method of analysis

The ice observations were taken at the end of each month, and only extremes in the January departures of sea ice will be considered in the following sections. However, the choice of January as a winter month was not crucial; ice anomalies do have high monthly persistence (WJ-I). To make a comparison
of simultaneous fields, the January and February atmospheric data were averaged together.

To examine sea ice extremes over the entire polar cap, the five Januaries of maximum sea ice area and the five Januaries of minimum sea ice area were first determined. Table 1 gives the January departure of sea ice area over the entire Arctic in square kilometers for each year, with H (heavy) or L (light) denoting the extremes. Here, January departure refers to the total sea ice area for a given January minus the mean sea ice area in January during the 25-year period.

For each type of atmospheric data, the January–February patterns were averaged together for the heavy ice Januaries then for the light ice Januaries. The differences between these two mean patterns were then found (heavy minus light).

There is an obvious drawback to this approach of relating sea ice and atmosphere. The two fields of ice and atmosphere are probably interrelated on many different time scales and through a variety of physical processes. This approach does not necessarily isolate one single process, nor determine that one field solely drives the other at zero time lag. Therefore, the possibility arises that while the results may be statistically significant, they may not be immediately interpretable in terms of the physical processes involved.

The heavy and light analyses were also applied to regional extremes in sea ice. Here, regional extremes refer to the twelve 30° longitudinal sections around the Arctic. Results will only be presented for the Bering Sea and the Davis Strait regions, since the other areas showed relatively insignificant results. The analyses of the Bering Sea and Davis Strait regions showed not only more significant atmospheric differences but also showed patterns that were consistent with each other. The precise areas defined by these two regions are shown in Fig. 1 and the January departures are shown in Table 2.

A statistical test of significance that lends itself well to this type of analysis is the student's t-test for the difference between two means (see Mitchell et al., 1966, p. 63). The significance levels were, in each case, computed using this test and are shown (by shading) superimposed on the difference patterns.

3. Results

a. Background information

First some gross features of the January–February circulation will be presented. This will be useful in understanding the "difference" patterns discussed later.

The mean January–February sea level pressure pattern for the years 1953–77 was computed and is shown in Fig. 2. The Icelandic low is centered at about 60°N, 40°W, and the Aleutian low is at 50°N, 170°E. There is also a predominant high pressure center over Siberia at about the same latitude as the Aleutian low.

The standard deviations shown in Fig. 3 were

![Fig. 1. Base map; longitude interval 20°, latitude 20° from 80°N southward. Arrows denote longitudinal boundaries used in regional heavy-light analysis.](image-url)
Table 2. January departures of sea ice area by region ($\times 10^4$ km$^2$).

<table>
<thead>
<tr>
<th>Year</th>
<th>Davis Strait (40°-70°W)</th>
<th>Bering Sea (160°W-170°E)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1953</td>
<td>0.58</td>
<td>-0.72</td>
</tr>
<tr>
<td>1954</td>
<td>-2.25</td>
<td>3.69</td>
</tr>
<tr>
<td>1955</td>
<td>-0.25</td>
<td>0.86</td>
</tr>
<tr>
<td>1956</td>
<td>-3.95</td>
<td>8.68</td>
</tr>
<tr>
<td>1957</td>
<td>9.82 (H)</td>
<td>-6.88</td>
</tr>
<tr>
<td>1958</td>
<td>2.62</td>
<td>9.12 (H)</td>
</tr>
<tr>
<td>1959</td>
<td>4.32</td>
<td>1.17</td>
</tr>
<tr>
<td>1960</td>
<td>-10.36</td>
<td>0.38</td>
</tr>
<tr>
<td>1961</td>
<td>-7.31</td>
<td>-6.46</td>
</tr>
<tr>
<td>1962</td>
<td>-13.91 (L)</td>
<td>-16.00 (L)</td>
</tr>
<tr>
<td>1963</td>
<td>-17.91 (L)</td>
<td>-12.62 (L)</td>
</tr>
<tr>
<td>1964</td>
<td>1.67</td>
<td>-2.26</td>
</tr>
<tr>
<td>1965</td>
<td>-3.77</td>
<td>3.68</td>
</tr>
<tr>
<td>1966</td>
<td>-18.27 (L)</td>
<td>-11.98 (L)</td>
</tr>
<tr>
<td>1967</td>
<td>-4.12</td>
<td>2.01</td>
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<tr>
<td>1968</td>
<td>4.45</td>
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<td>1969</td>
<td>-16.61 (L)</td>
<td>-12.47 (L)</td>
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<tr>
<td>1970</td>
<td>-3.33</td>
<td>8.80</td>
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<tr>
<td>1971</td>
<td>6.77</td>
<td>10.59 (H)</td>
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<td>1975</td>
<td>20.36 (H)</td>
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</tr>
<tr>
<td>1977</td>
<td>-10.89 (L)</td>
<td>11.61 (H)</td>
</tr>
</tbody>
</table>

Computed from the January–February averaged sea level pressure patterns. The largest standard deviations are just east of or coincident with the major surface low pressure centers.

The mean January–February pattern of the 700 mb height field (Fig. 4) is much smoother than that for the sea level pressures. There are two major troughs, each corresponding to the two major surface low pressure centers. The first trough which tilts slightly westward with height above the Icelandic low, extends from the pole through Hudson Bay, then southeastward into the Atlantic. In further discussions, we will refer to this as the eastern North American trough. The other trough also ex-
due to chance. Therefore, none of these differences can be viewed as significant.

An inherent assumption in examining extremes in sea ice over the entire cap to ascertain atmospheric relationships is that Arctic sea ice does, in fact, tend to vary in a wavenumber zero mode. In other words, this approach would bring out meaningful relationships between ice and atmosphere only if (among other factors) the sea ice variations had a strong wavenumber zero component. WJ-I found, however, that Arctic sea ice during this period had a much stronger wavenumber 1 component instead. Significant ice-atmosphere relationships may then be more likely to occur when ice extremes are specified locally, and this will later be shown to be the case.

The 700 mb height differences were also calculated and are shown in Fig. 6. Note that the 700 mb height differences are generally much more significant than the sea level pressure differences. Nevertheless, the differences at the two levels strongly resemble each other. At 700 mb, an anomalous low is again directly to the north of an anomalous high in the Atlantic. Thus (for heavy ice), while the surface Icelandic low is stronger and is displaced slightly northeast and the surface subtropical high is north of its normal position, the eastern North American trough lies further away from the continent and does not extend as far south.

The similarities between the 700 mb height and sea level pressure differences are also quite marked in the Pacific. Corresponding to the weaker surface

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**Fig. 5.** January–February sea level pressure differences for heavy minus light ice Januarys. Contour interval is 1 mb. Outer shading indicates 95% significance or greater, inner shading 99% significance or greater.

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**Fig. 6.** January–February 700 mb height differences for heavy minus light ice Januarys. Contour interval is 1 dam. Shading the same as in Fig. 5.

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**b. Heavy and light analyses over the entire polar cap**

The January–February sea level pressure differences between heavy and light ice Januarys over the entire cap (Table 1) are shown in Fig. 5. Note the large negative differences centered just off the east coast of Greenland with large positive differences centered to the south. The other main features to note are the weakened Aleutian low and Pacific subtropical high. So when Arctic sea ice is of greater extent than normal, the Icelandic low is stronger and is displaced slightly to the northeast, the subtropical high in the Atlantic is displaced northward, and the Aleutian low is weaker as is the Pacific subtropical high.

The only difference in Fig. 5 significant at the 99% level occurs over the southwestern United States where the standard deviations are quite small (on the order of 1 mb or less during both the heavy and light ice years). In addition, one would expect 1% of the area to be significant at this level simply...
subtropical high in the Pacific, there is a weaker 700 mb ridge off the west coast of the United States. The intensity of the Aleutian low (trough) is weaker at both levels as well. Finally, it should be noted that the differences in the Pacific are more statistically significant, overall, than those in the Atlantic.

For heavy ice conditions in winter then, it appears that the atmospheric circulation in the North Pacific is weaker while the strong circulation in the Atlantic is even stronger, but further north.

The temperature differences at the surface and at 700 mb along with their statistical significances are shown in Figs. 7 and 8, respectively. The results are surprising in that the surface temperatures are not lower over the entire cap, but only lower near the Bering Sea, the Greenland area and north central Siberia. However, these two former regions are the main areas where sea ice varies in wintertime (WJ-I, Fig. 1). The only significant differences occur in the Barents Sea, where the surface temperatures are higher during heavy ice winters.

The results of van Loon and Williams (1977) may lend some insight into these Barents Sea temperature differences. These authors investigated the mean winter surface and 700 mb temperature trends from 1949–72, as well as the number of tracks of lows which crossed 70°W between 30 and 40°N versus those between 45 and 55°N. The surface temperature trends over the Arctic, which they computed, are consistently of the opposite sign to those in Fig. 7. They relate the falling temperatures in the Barents Sea area to the decrease in storm tracks into that area, allowing the local stability to increase. They went on to suggest that if the trends were of the opposite sign, then the storm tracks in the North Atlantic would move further north, meaning stronger winds, more clouds and more frequent oceanic air masses that would destroy or diminish the surface stability and make for a drastic rise of surface temperatures over a large part of the eastern Arctic, particularly in the Barents Sea where the largest positive differences were found here.

The 700 mb temperature differences are quite similar (Fig. 8). For heavy ice, temperatures are warmer than normal from the pole to Scandinavia while the only colder polar regions are near Greenland and the Bering Sea.

c. Model comparison

Model results that may be of interest here, because of the similar temporal and spatial scales involved, are those reported by Herman and Johnson (1978). These authors used the GLAS (Goddard Laboratory for Atmospheric Sciences) general circulation model to simulate wintertime atmospheric differences between minimum and maximum Arctic sea ice coverage. The January extremes in Arctic sea ice were estimated from 17 years of observed data in the Atlantic sector, and from five years of Pacific data. The atmospheric differences they calculated include those from sea level pressures, 700 mb heights and temperatures, the zonal wind as a function of latitude and pressure, and the precipitation over northern Europe. They ran the

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**Fig. 7.** January–February surface temperature differences for heavy minus light ice Januarys. Contour interval is 1°C.

**Fig. 8.** January–February 700 mb temperature differences for heavy minus light ice Januarys. Contour interval is 1°C.
model several times with maximum and minimum sea ice conditions for the January and February period.

The Herman and Johnson (1978) model results for the sea level pressure differences show a deepened Icelandic low further to the southeast than the observational results but, in general, the Atlantic results are similar. They also found higher than normal pressure directly to the south of this low. In the Pacific, their results do not correspond as well with those of Fig. 5. They find the Aleutian low had deepened and extended further downstream for maximum ice conditions.

Examining the 700 mb height differences calculated by Herman and Johnson (1978), we find similar consistency between studies; the differences in the Atlantic are more comparable to the observational results than those in the Pacific. Their results show lower than normal heights near Greenland with higher heights directly to the south.

Although the differences in the Atlantic computed by Herman and Johnson (1978) did not compare unreasonably with the observational results, the statistical significances were low enough (in the case of the observed data) that few definite conclusions can be drawn. Since their study only included an ice-forming atmosphere, however, one might be tempted to think that where model and observations agree, the ice is forcing the atmosphere. In any case, it is interesting to note here that both model and observation show increased westerlies in the mid-Atlantic.

d. A regional approach to ice-atmosphere relationships

As mentioned previously, it may be more fruitful to see how ice anomalies in a more localized region affect the atmospheric circulation (or vice versa).

In this section, the polar ice cap was divided into twelve 30° sections, each extending from the pole southward. Fig. 1 shows the precise areas reported on here (marked by arrows). Five heavy and five light ice Januaries were then identified for each section. Because only a few areas vary much in Arctic sea ice during winter, many ice areas were not associated with significant atmospheric differences [e.g., the Barents, Laptev, East Siberian and Beaufort Seas, (see WJ-I)]. The areas that were associated with significant atmospheric changes were the Davis Strait region (40°–70°W) and the Bering Sea (170°E to 160°W), and Table 2 identifies the heavy and light ice Januaries for these two regions.

It must be mentioned that the region east of Greenland (i.e., the Denmark Strait, East Greenland and Norwegian Seas) is also quite variable in wintertime sea ice area, as found by WJ-1, but the atmospheric differences were nonetheless relatively small and insignificant. This indicates different time scales and/or different physical processes are involved between atmosphere and ice in this region, but as these results were not statistically significant, they will not be shown here.

Fig. 9 shows the sea level pressure differences using the Davis Strait region alone in the heavy and light ice analyses. The most significant atmospheric differences are in the same region, so that northerly flow through Baffin Bay and Davis Strait is significantly associated with simultaneous above-normal sea ice in that region. In this case, it is fairly clear that we have cold air advection causing more sea ice to form and/or sea ice being pushed southward. In other words, atmospheric forcing sea ice at zero time lag is the simplest and most obvious interpretation here.

Actually, the atmospheric "simultaneous" fields are an average of the atmospheric state two weeks before the ice extremes and six weeks after. Since the latter result points to atmosphere forcing sea ice at “zero time lag”, it may be that the atmospheric state two weeks before the ice anomalies are weighting the two-month averages. The lagged correlations between the dominant modes of sea ice and surface pressure variability computed by WJ-II support this contention. They found that the correlations of surface pressure two weeks before the ice well above the 95% significance level, and for the surface pressure six weeks after, the correlations were considerably smaller and relatively insignificant.

From Table 2, it can be seen that sea ice in the
Davis Strait varies considerably more than in the Bering Sea. Its wintertime variance is also greater than any of the other 30° sectors. Walsh (1978) found that the predominant mode of variability in high-latitude surface geostrophic wind includes a strong meridional component only in the Davis Strait. This could be additional evidence of atmosphere driving ice in that region, i.e., high variability in the meridional surface wind is associated with high variability in sea ice.

Some results of van Loon and Rogers (1978, Fig. 3) are remarkably similar to those in Fig. 9. In an examination of the seesaw in winter temperatures between Greenland and northern Europe, they defined “Greenland below” winters as winters when Greenland was below normal in temperature and northern Europe above. For these Greenland below winters, they found lower than normal sea level pressure centered between Iceland and Greenland and higher pressure over central Europe extending westward into the Atlantic. These anomalies are not only in geographic proximity and of the same sign, but also of comparable magnitude to those in Fig. 9.

The main area in the Pacific of sea ice variability during wintertime is in the Bering Sea (WJ-I). The atmospheric differences at the surface are again quite marked for this region (Fig. 10). During winters with more than normal sea ice in the Bering Sea, the Aleutian low is stronger and displaced somewhat downstream, while the Icelandic low is also stronger, although the latter is not as significant. More importantly, it is again clear that we have northerly flow associated with more than normal sea ice. In addition, these surface results indicate that sea level pressures are more closely associated with local extremes in sea ice than with extremes over the entire cap.

The heavy and light analyses of the Davis Strait and the Bering Sea ice areas were also carried out for the 700 mb level. Fig. 11 shows the 700 mb height differences for the Davis Strait area. In those winters when there was more than normal ice in the Davis Strait, the eastern North American trough is displaced about 30° downstream, while the Aleutian trough is weaker. It can be seen that over the North American continent, the westerlies are much stronger for icy Davis Strait winters.

Comparing these differences with the corresponding sea level pressure differences (Fig. 9), it should be noted that baroclinicity increases for heavy ice over much of Canada while the atmosphere becomes more barotropic over northern Europe. Nevertheless, the patterns at the two levels show strong similarities.

The 700 mb height differences for heavy minus light January Bering Sea ice are not so large (Fig. 12). The only large significant differences are centered over western Alaska where the height differences reach 8 dam. There is a strong similarity between 700 mb height differences for ice extremes in the two regions: each heavy ice region is associated with lower than normal heights slightly to the east. The 700 mb height differences for extremes in Bering Sea ice are also markedly similar to the sea level pressure differences (Fig. 10).
This regional analysis has shown that more than normal sea ice in both the Bering Sea and Davis Strait are strongly associated with intensification of the closest major surface low pressure center and the overlying 700 mb trough. In addition, large departures in the surface meridional wind over an ocean area play an important role in sea ice extent in that ocean area.

4. Conclusions and discussion

In examining extremes in sea ice over the entire polar cap, it was evident that the surface differences were not nearly as significant as were those at the 700 mb level. The regional approach in the last section indicated that this may be true because sea ice is related to surface flow in an advective way, i.e., surface northerly flow through an ocean area is significantly associated with more than normal sea ice in that region. This type of flow regime would have been difficult to ascertain by examining extremes in sea ice over the entire cap. Nevertheless, the sea level pressure and 700 mb height fields were in each case well correlated.

Regional extremes in sea ice were also associated with intensification of the closest major low-pressure centers and troughs (for heavy ice; the reverse is true for light ice). In the case of the Davis Strait, the Icelandic low and the eastern North American trough strengthened and in the case of the Bering Sea, the Aleutian low and trough (although they both moved eastward). It is also important to note that extremes in sea ice between Greenland and Scandinavia were less significantly related to changes in the atmospheric circulation, although there is considerable wintertime sea ice variability in that region (WJ-I).

When ice was greater over the entire cap, the surface Aleutian low was weaker and the Icelandic low stronger, while the associated 700 mb troughs showed consistent changes. The main polar areas, which were found to be cooler during heavy ice, are the Bering Sea and the area around Greenland—the two main areas where sea ice can grow in wintertime. Since the results from the regional analysis most strongly support atmosphere driving ice, the latter result indicates that atmospheric heat transports in a few key regions may play an important role in determining wintertime sea ice extent. van Loon and Williams (1977) suggest that the rising surface temperatures over the Barents Sea may be due to a decrease in the stability there, associated with a northward movement of the storm tracks in the north Atlantic.

While the results of Herman and Johnson (1978) showed the 700 mb Pacific and Atlantic troughs strengthened in response to maximum sea ice conditions, the observational results indicate a weakened Pacific circulation and only slightly strengthened north Atlantic circulation. The key words here are in response, i.e., their model simulated changes in the atmospheric circulation with fixed ice boundaries and with all other factors prescribed and constant (e.g., sea surface temperatures, snow cover). The observational results, in contrast, point out the atmospheric conditions which prevail during extremes in sea ice conditions. In addition, it must be reiterated that Arctic sea ice tends to vary in a fashion where it is shifted from east to west as opposed to uniformly around the cap. That thereby makes not only the low significances obtained in the first part of this analysis understandable, but also the poor comparison with the model results. These empirical results may actually prove to be more useful for future model simulations where coupled ice-atmosphere-ocean interactions are included.

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