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ABSTRACT

An active role in modifying the earth's climate is suggested for low cloudiness over the circumarctic oceans. Such cloudiness is linked to the spatial differences in ocean surface temperatures. The temporal variations from year to year of ocean temperature patterns can be pronounced and therefore, the low cloudiness over this region should also show strong temporal variations, affecting the albedo of the Earth and therefore the climate.

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CLIMATIC CHANGE BY CLOUDINESS LINKED TO THE SPATIAL VARIABILITY OF SEA SURFACE TEMPERATURES

"Behold, there is a little cloud arising out of the sea . . . ."

I Kings, Chapter XVIII, verse 44.

INTRODUCTION

Machta and Telegadas (1974) discuss the global trends in the surface temperature between 1870 and 1968, and note that the temperature has generally risen between 1880 and about 1940, after which it has fallen. The amplitude of the changes is larger in the Northern than in the Southern Hemisphere. The warming from 1880 to 1940 was almost twice as large at the temperate and high latitudes (25° to 60°) as in the tropics; and the entire cooling at the surface in the Northern Hemisphere in the decade of the 1960's, compared to the period of 1931-1960, occurred north of about 50°N (Machta and Telegadas, 1974).

More detailed annual mean air temperatures for various latitudinal zones have been compiled for the three decades 1920-1950 vs 1890-1920, by Willett (1950) and by Callendar (1961). Callendar used a different method of data selection [illegible] different approach to the analysis than Willett, yet their findings, tabulated by Landsberg and Mitchell (Willett's data were recalculated and updated by Mitchell), show a remarkable agreement that the greatest warming occurred between the latitudes 25°N to 60°N (see Table 1, reproduced from Landsberg and Mitchell, 1961). Mitchell discusses more fully the zonal effect and points out that latitudinal zones of the most rapid warming prior to 1940 tend to coincide with those of the most rapid cooling since 1940 (Mitchell, 1963).

Looking farther backwards, during the last million years (the Quaternary era), the climate has several times undergone changes between the great ice ages and the interglacial climates as warm as now. The changes of climate between the ice ages and the interglacials were greatest in the lands on either side of the North Atlantic (Lamb, 1969, p. 177). Throughout the Quaternary, the shifts of ice cover in all the glaciations were according to Budyko and Vasischeva much larger in the Northern Hemisphere than in the Southern Hemisphere (see Table 2).

The recent investigations of possible causes for these climatic fluctuations, are concentrated on factors external to the atmosphere-oceans-lands-ice system, namely solar constant variability and increased loading of both volcanic and man made aerosols (MacCracken and Potter 1975; Schneider, 1975; Wetherald and Manabe, 1975). A study of a possible internal cause, of cloudiness fluctuations, is presented in this paper.

The possibility of an important role for cloudiness in changing the climate has been presented but questioned both by Lamb (1964, pp. 345 and 346) and by H. Wexler (1953, p. 98). Wexler states: "Apparently, changes in the earth's albedo, no matter how brought about, can be an important factor in changing the mean temperature of the earth and its atmosphere, and consequently world climate.

If one ascribes changes in albedo to changes in cloud amount, it is difficult to see how, keeping constant the land-forms and the radiation received by the earth, significant changes in cloudiness can arise from internal (atmospheric) considerations alone. A possibly important terrestrial influence on cloud amount will be mentioned in the section on atmosphere turbidity. It is recognized that variations in albedo may be caused by changes in reflectivity of clouds, rather than their amount, created by large changes in the prevailing type and thickness of clouds. But again it would be difficult to explain these changes without recourse to external influences." Flohn (1963) endorsed this viewpoint, stating that there were no reasons for a global change in cloudiness in the recent centuries, but such effects cannot be excluded in the geological past.

It is suggested here that a fluctuating, cloudiness-influencing factor might exist, external to the atmosphere itself but internal to the atmosphere-ocean system, in the spatial variability of the sea surface temperatures (SVOSST). The mechanism is discussed in detail, but in the introduction here the suggested time-varying aspect of the mechanism can be described as follows: It is reasonable to expect that if Idaho were to be plucked from the middle of the North American Continent and deposited to float around in the oceans, an increase in the cloudiness around the new island would result. Now, as Wexler (1953) says, the landforms are constant. The point is made here that even though Idaho is not budging from its place, and it is not floating in the Atlantic, sea surface temperature anomalies, nearly equivalent to land-sea temperature differences, with linear dimensions comparable to the length of Idaho, can occur in the oceans, and actually have been reported (see Fig. 1, reproduced from Dickson and Lamb, 1971; Bjerknes, 1963, Figs. 6-14, pp. 304-312; also Lamb, 1972, Fig. 10.6, p. 396).

The spatial variability of the sea surface temperatures (SVOSST) is especially likely to be subject to temporal changes (both in the magnitude of the SST gradients and the orientation relative to the prevailing surface winds) in the North Atlantic and the North Pacific. Linked to this temporal variability in SVOSST, the cloudiness over the arctic and circumarctic waters, at the approaches to the sills linking the Arctic Ocean to the Atlantic and the Pacific, might be an important climatogenic factor. The low cloudiness is stressed, of convection cumulus and advection fog, arising from the local meteorological processes. Such low cloudiness is quite prevalent over circumarctic waters, especially in the summer, and is seasonally highly variable.

Before discussing details of this mechanism and how it can actively alter the hemispheric radiation balance, we shall discuss in somewhat general terms the question of the climatic variability.

Table 1
Thirty-Year Change of Annual Mean Temperature

Callendar[1]		Willett, revised by Mitchell[2]	
Latitude Limits	°C.	Latitude Limits	°C.
60° N. - 25° N.	0.39	60° N. - 20° N.	0.32
25° N. - 25° S.	0.17	30° N. - 30° S.	0.19
		20° N. - 50° S.	0.06
25° N. - 50° S.	0.14		
		20° N. - 60° S.	
		60° N. - 60° S.	0.21
60° N. - 50° S.	0.23		
		80° N. - 60° S.	0.27

[1] 1921-50 minus 1891-1920
[2] 1920-49 minus 1890-1919.

Table 2
Climatic Change in Glaciation Epochs
(Source Budyko and Vasischeva, 1971, quoted in
Study of Man's Impact on Climate, SMIC, MIT, 1971)

Time in Thousands of Years before 1800 A.D.	Δ°N	Δ°S	Δ°T
22.1	- 8	- 5	-5.2
71.9	-10	- 3	-5.9
116.1	-11	- 2	-6.5
187.5	-11	0	-6.4
232.4	-12	- 4	-7.1

Notes: Δ°N-displacement of the mean latitudinal limit of ice cover in the Northern Hemisphere
Δ°S-displacement of the mean latitudinal limit of ice cover in the Southern Hemisphere
ΔT-temperature change at 65°N

CLIMATIC VARIABILITY AND TRANSITION-PRONE CLIMATIC REGIONS

Climate is the average state of the land-ocean-atmosphere-system and for the purposes of this discussion all the oceans can be considered internal to the system. Climate or climatic parameters can be defined as long term averages of monthly* averages of meteorological variables. For a given region, the set of 12 monthly points in the space of meteorological variables represent the yearly component of climate information. The next year, the yearly component is represented by a different 12 point set in the meteorological variables space. The question is how different is the next year set going to be, and with what probabilities. What kind of "random-walk" is this set of 12 points likely to take in the next decade, century or millenium?

The basic discussions of the problems underlying such questions were presented by Lorenz (1971). He raised and analyzed the possibility that what we interpret as climatic change might be nothing more than natural fluctuations arising solely from the complex nonlinear interactions between the four ingredients of the system: land, atmosphere, ocean and ice/snow canopy (especially polar ice). Not all of these ingredients of the system are found in every region.

Lorenz outlines and discusses such distinct climatic regimes as glacial on one hand and interglacial on the other—and asks the question, whether the two regimes represent different manifestations of the same global climate. If the

*It is realized that seasonal average, with length of seasons itself a variable, might be more appropriate.

answer is yes, i.e., when such changes can take place back and forth just because of fluctuations, the system is defined by Lorenz as transitive. In terms of our 12 point sets in the meteorological variables space, the yearly sets for a few thousand years are located in the subspace representing the glacial conditions. However, there is no confining or separating boundary around this subspace, and subsequently the path moves to the subspace representing the interglacial. If such a movement or transition is impossible in the system, and glacial and interglacial subspaces are really separated by a boundary, (i.e., the system, in Lorenz's definition, is intransitive) the climate change has to be ascribed to external causes, which created "a new world."

Lorenz goes on to an important suggestion that the system might be indeed transitive, but behaving apparently stably in one or the other of two different regimes. Lorenz calls such a system almost intransitive. The differences between the two regimes (such as glacial and interglacial) can be so pronounced that one would hesitate to interpret both as manifestations of one transitive system.

Assuming the system as transitive, one can gain in understanding of climatic change, by attaching some transition probabilities to the changes. And the point brought up here is that such transition probabilities should be studied first of all on a regional basis. Rather obviously, some regions can be transition prone, i.e., more likely than others to act as setting up a cooling or a warming trend for the whole globe. Such regions are characterized by more processes at hand, more complex processes, or some special processes, a special dependence of their radiation balance on the variable effects emanating from the neighboring regions. As an example, consider the radiation balance of the equatorial Atlantic Region: it varies considerably depending whether it is or it is not covered by a dust cloud from the Sahara. Such clouds can be thousands of kilometers in extent.

The possibility of a region acting as a sensitive touch stone for a climatic change in a passive sense has been discussed by Wolbach (1953). Discussing the hypothesis that the ice ages are preceded by reduced solar radiation, Wolbach suggests that under the postulated geographical conditions of the Permian, with a reduction in the solar radiation, the extended Antarctic Continent would be the first region to suffer a severe drop in temperature. The Arctic Sea fed by tropical ocean currents might remain unfrozen even while pack ice began to form in the Southern Sea around a snow-covered Antarctica (Wolbach, 1953, p. 113). Arguing that evidence exists that this actually happened, Wolbach presents a case for a change in the solar constant as causing the events. Here we follow a similar line of argument, suggesting an active regional mechanism as a cause for the climatic fluctuations. The suggested mechanism, internal to the climate system, would cause more pronounced fluctuations in the Northern Hemisphere than in the Southern, and especially strong variability of the circumarctic regions, trends that were actually observed.

LOW CLOUDINESS OVER THE OCEANS: LINK TO THE SPATIAL VARIABILITY OF SEA SURFACE TEMPERATURE (SVOSST) AND STATISTICS

There are two mechanisms that promote cloudiness when air over the oceans flows across a surface temperature gradient. When crossing from cold to warm regions, the air is heated from below, uplift occurs, and if the air contains sufficient moisture, convective cumulus clouds form. This qualifying remark about moisture contents is especially important in the polar regions, in which the air can be quite dry. The second mechanism is that the moist air coming from the warm water regions, is cooled when entering cooler waters, and mist or stratus cloud formation occur.

The first mechanism has been studied extensively by Joanne Malkus. Her attention was drawn to the effect of a heated surface on the cloudiness through observations of clouds forming on the shore of the Nantucket Island (Malkus and Bunker, 1952). She and Stern continued the studies of the subject analytically (Malkus and Stern, 1953; Stern and Malkus, 1953). J. Malkus subsequently carried out an observational program of trade cumulus over the oceans (Malkus, 1957), in which she found very good correlation between cloud formation and existence of elevated water temperatures possibly by as little as 0.1°C-0.3°C.

It is difficult to postulate generally cloud formation due to such small temperature differences, but with differences of a few degrees in the surface temperatures, strong cloudiness effects can be expected. Such differences occur at boundaries of ocean currents. To illustrate the effect, a cumulus cloud street formed due to heat island effect of an actual island is shown (Fig. 2, after Pitts et al., 1974, island of Fernando de Novonha, Skylab 4 image SL-4-138-3874). In Figure 3 thermal mapping of the Western North Atlantic, we can discern large scale cumulus forming over the Gulf Stream. The cloud streets of the forming cumulus correspond to the outline of the Northern Wall of the Gulf Stream.

Temperature anomalies of 1°C-3°C occur because of the shifts in currents, but such anomalies are especially common with shorter spatial dimensions of up to 300 km. Scully-Power and Twitchell (1975) report a series of circular cumulus clouds of approximately 150 km in diameter, off the east coast of Australia. They link the formation of these clouds with eddies of waters warmer by 2 or 3°, at the edge of the East Australian current. Effects of Gulf Stream eddies on clouds were discussed by Rao et al., 1971.

The formation of fogs and low clouds when warmer air flows over cooler water is a well recognized phenomenon, with the prime examples being the region of the North Pole in the summer, with its many seasonally open leads among the ice, as well as waters off Labrador and Newfoundland (Lamb, 1972, p. 309). In this process a stable stratification of density is induced in both air and sea, inhibiting convection, (Lamb, 1972, p. 330). Fogs are quite common near the coasts of Greenland, about 50-60 days/year, and in general the appearance of fog is associated with specific wind directions of easterly and southeasterly winds (Putnins, 1970, p. 98), i.e., regions that are generally warmer than the coastal waters. Prevalence of fog in the North Polar Basin is discussed by Vowinckel and Orvig (1970). Deacon (1969) discusses fog over the sea, stating: "More purely advection-type fog occurs over the sea where radiation fog is not found owing to the smallness of the diurnal variation in sea surface temperature. Air travelling over warm sea has a high humidity, so that when its trajectory takes it over cooler water, fog readily forms if the wind is not too strong. Advection fogs are therefore common where the cold Labrador Current meets the Gulf Stream in the vicinity of Newfoundland: advection fog there occurs on more than 100 days per annum. A more generally useful idea of magnitudes may however be gained from the examination of climatic charts of the oceans: for example, those for the North Atlantic in July show that regions where the climatic sea temperature gradient is 1°C per 100 km, have typically a 10% frequency of days with fog." As an illustrative case of fog formation, consider Figure 4, an ERTS image of the Pacific offshore area near San Francisco. The fog formed over the ocean, hugging the coast. A similar image (interpreted by P. L. Szereszewski) showed that fog is absent from two areas of warmer waters close to the shore, where two rivers flow into the Pacific.

A definite correlation exists between prevalence of cloudiness and the location of ocean temperature gradients. Consider the very pronounced belt of cloudiness extending from the shores of Georgia and Carolinas, USA, to the Bay of Biscay and Great Britain, slanting across the Atlantic through the latitudes of

35°N to 50°N (Miller and Feddes, 1971; Winston, 1969). This band, prevalent through most of the year, coincides with the northern boundary of the Gulf Stream, with its strong sea surface temperature gradients (Lamb, 1972, p. 309, Fig. 8.2 adopted from Dietrich). Figure 3 constitutes a good example of cumulus formation along this band.

In this cross-Atlantic band of cloudiness, obvious high and low clouds occur. Currently, statistics of cloudiness in which low cloudiness is separated from high cloudiness are limited. Telegadas and London (1954) complied incidence of various types of cloudiness vs latitude bands from the Atlas of Climatic Charts of the Oceans (McDonald, 1938). The statistics of low cloudiness from 40°N to 75°N are reproduced in Table 3. Maps of fog (Guttman, 1971) are shown in Figs. 5-8. Additional information of this type will become available from the current analysis of satellite imagery by Becker (1975), who in a preliminary report indicated pronounced low cloudiness at the higher latitudes.

It can be seen from Table 3 that the average summer cloudiness in the latitude belt 40-75°N is about 34% and that fog (20%) predominates, constituting six tenths of low cloudiness. Month-to-month variability of fog can be assessed by comparing Figures 5-8. Seasonal variability of cloudiness is very pronounced at the high northern latitudes and increases with increasing latitude (see Fig. 9, after Vowinckel and Orvig, 1970).

Table 3
Latitudinal Distribution of Average Observable Amount of Low Clouds by Type in Percentage of Sky Cover
-Summer- (from Telegadas and London, 1954)

Latitude	St; Sc; Fc.	Cu	Cb
40-45	15.5	8.8	4.3
45-50	18.3	9.0	4.5
50-55	19.6	9.0	4.9
55-60	19.6	9.4	4.9
60-65	19.5	9.6	4.8
65-70	20.0	10.0	4.9
70-75	21.0	10.2	5.0

Simulation of the global climate for the month of January by the GISS general circulation model, produced pronounced low cloudiness in the North Atlantic (see Fig. 10 reproduced from Somerville et al., 1974).

Charts of the mean SST for the month of July are presented for the North Atlantic in Fig. 11 and for the North Pacific in Fig. 12. From the examination of these charts, and those of fog, Figs. 5-8, the close correspondence of region of high incidence of fog and pronounced SVOSST can be seen.

WHY CIRCUMARCTIC OCEANS?

Climate is statistics of the weather, and here we actually discuss a climate change as due to a cloudiness change linked to a change in SVOSST, i.e., a train of events in changing weather. The point made is that where the weather processes are more complex involving more interactions, more pronounced temporal variability of the weather statistics, i.e., of the regional climate, can be expected.

One can visualize the climate as a position (as defined by the means and distributions of the coordinates) of a small sphere, bombarded by molecules, whose individual motions are beyond our capability to track. But the higher is the number of the molecules in a position to impact the sphere, the larger is the "random walk" of the sphere in its Brownian motion and the more variable is its position.

In a region where all of the major ingredients of the weather systems, i.e., atmosphere, ocean, ice and land co-exist, there are $\frac{4 \times 3}{2} = 6$ interactive processes; in the regions where ice or land are absent, there are $\frac{3 \times 2}{2} = 3$, only half as many interactive processes, and the regional system is simpler. Thus, if we set to look for a region where the weather can be expected to be more complex and, therefore, more variable, we should give first attention to regions with all the four major system ingredients, i.e., circumarctic and circumantarctic. However, the circumantarctic region is a thin band in latitude, whereas in the circumarctic region, the land and sea are so formed, at least in the Quaternary period, that the land interaction with the other three ingredients comes to a full play. Indeed, this is the region of the sills which link the Arctic Ocean with the Pacific and the Atlantic. The flow through the sills and their approaches can be expected to be variable, since it reflects in a high leverage the fluctuating conditions in areas much larger than the narrow sills. Land can interact all along the sills (see, for instance, the discussion of fog in Greenland offshore areas).

The strong thermal gradients in circumarctic waters are due to (a) the sea ice on the fringe of the Arctic, which constitutes a surface temperature discontinuity and (b) existence of both North and South currents in each of the three passages between the North Atlantic and Arctic, as listed below. These thermal gradients due to both (a) and (b) above are subject to pronounced temporal fluctuations.

In the case of the sea ice, the changes can be local, on a small scale, caused by icebergs breaking off and moving away from the pack ice. Since an iceberg or rather an ice island can be up to 200 km in length, the location of just two or three ice islands can have an impact on SVOSST in the region.

On a larger scale, the extent of pack ice can vary by 3-4 degrees in latitude especially around Iceland (Lamb, 1972, Fig. 8.12, p. 338). The changes or anomalies can be in opposite direction in various locations (see Fig. 1). The pronounced effects on the atmospheric circulation by such shifts have been stressed by Lamb (1972, p. 338).

The shifts are a sensitive measure of changes in the budget of available radiation (Lamb, 1972, p. 339); thus, the cloudiness fluctuations discussed here as arising from fluctuations in SVOSST would obviously have a feedback effect on the boundary of the pack ice. This is further discussed below.

The SST variability due to changes in the ocean currents, at the shortest dimensions is due to the eddy-like protrusions of the major currents with dimensions of up to 300 km. "The Gulf-Stream Labrador current boundary itself becomes contorted into meanders up to 100-300 km across, and great "globules" (or pools) of the warm or cold water of this size are liable to be cut off and surrounded by the other water-mass" (Lamb, 1972, p. 324). It should be recalled that, in the vicinity of Australia, low clouds associated with such eddies were discussed by Scully-Power and Twitchell (1975).

The main currents over the sills that connect the Arctic to the Atlantic, (Labrador current and West Greenland current between Canada and Greenland; East Greenland current and Irwinger current between Greenland and Iceland; branch of East Greenland current and North Atlantic drift between Iceland and Norway) are topographically controlled, and major shifts are not expected, and indeed not postulated here. However, it should be stressed that even the

Antarctic convergence, which in part is topographically controlled in the Drake Passage, and which is regarded as maintaining a constant position from year to year and from era to era, in reality exhibits week-to-week changes of position, strength, and sometimes development of double or triple structure of its thermal boundary (Lamb, 1972, p. 336), and a variability of the same nature occurs in the six currents listed above. "The surface currents in the ocean respond even to week-by-week and day-to-day changes of wind. The sea surface temperature anomalies of the order ±2°, and more, which commonly persist for months at a time in the region of one Gulf Stream-Labrador current boundary south of Newfoundland probably mark lateral displacements of that boundary over a range of 200-300 km" (Lamb, 1972; p. 333).

One fluctuating condition deserves a special mention -- the fresh water inflow into the Arctic. Of the 160 x $10^3 m^3$/sec inflow of the fresh water, 17 x $10^3 m^3$/sec to 20 x $10^3 m^3$/sec m^3 comes seasonally from a single river, Yenisei (Lamb, 1972, p. 324). Year to year fluctuations in runoff in the Yenisei basin introduce fluctuations in the salinity of the top layer of the Arctic -- and affects the density differences between these waters and the North Atlantic waters, known for their high salinity (Lamb, 1972, p. 340). This fluctuation possibly could be regarded as an interaction between two oceanic ingredients: an ocean of low salinity and an ocean of high salinity. The climatic importance of this interaction was stressed by Weyl (1968).

Year to year changes in the mean annual sea surface temperatures (SST) for the North Atlantic have been studied by Bjerknes (1959; 1963) and his many graphs reveal a pronounced year to year variability. Quite often, the SST's vary from year to year in a different sense in various locations. Bjerknes points out that a maximum rate of rise of SST was to be found from the 1890s to the 1950s along the Gulf Stream from Cape Hatteras to the edges of the Newfoundland Banks; a belt from Ireland to the edge of Labrador was not warmed up during the same period, but north of that belt there was also a warming. It follows that SVOSST must have been different at different times.

Actual observations show that two areas have been particularly likely to show surface temperature anomalies in the present era. "The one between 40° and 50°N, south of Newfoundland and Nova Scotia, concerns the limit of the warm Gulf Stream water. The other, between 30° and 45°N about 160-180W in the Pacific, similarly concerns the limit of the warm Kuro Shiwo water." (Lamb, 1972, p. 342). In these two locations of SST anomalies, one south of the sill currents between the Atlantic and the Arctic and the other south of the Aleutian current, the sill current from the Pacific to the Arctic, the SST gradients are strong and low cloudiness pronounced in the summer.

There is apparently some correlation between the cooling of the region and the intensity of the flow over the sills. The pronounced ice spell, i.e., colder conditions, in the Greenland-Iceland-Spitsbergen in the 1960's was shown to be associated with an increased volume of the flow, and a further southward penetration of the cold polar water in the east Iceland branch of the East Greenland current. This current had been strong in the nineteenth century but it apparently weakened in the warm decades between 1920 and 1950 (Lamb, 1972, p. 331, and also see the discussion on pages that follow).

This increased flow can readily explain the regional cooling, in a passive sense. The cooling in the Atlantic can be explained by a stronger coupling between the two adjoining regions, the Arctic and the Atlantic. Under this interpretation, the meteorological averages for the whole Northern Hemisphere (for temporal averages of a decade or longer) would remain invariant. Here, it is suggested that SVOSST is increasing at the time of the increased flow, and results in an increased cloudiness and a changed radiation balance. A cooling trend for the region and for the whole Northern Hemisphere follows.

A point should be brought up here, that Ewing and Donn (1956) linked the changes between the glacial and interglacial periods to the question of flow over the sill passages between the Arctic and the adjoining oceans. On their mechanism, cause and effect lie in the feedback of the depth over the sill, the flow, and the ice formation (Ewing and Donn, 1956; Don and Ewing, 1968). However, as Lamb (1969, p. 242) points out, estimation of the average temperature anomaly for the whole earth around 1800 A.D. and calculation of the increased area of ice and snow surface as compared with 1900-1939, indicate changes that were about a tenth part of the corresponding changes between, say, 1950 and the maximum development of the Pleistocene ice age proper. By contrast, the increase of mass of ice accumulated in ice sheets and glaciers on land in the Little Ice Age, as indicated by sea level changes, cannot have amounted to more than a few thousandths of the major ice age position. Thus, change in depth over the sills must have been insignificant, and this is not the crux of the mechanism suggested here, even though it might have played a role in the true ice ages.

RADIATION REDUCTION BY LOW CLOUDINESS OVER THE CIRCUMARTIC OCEANS

The effects of clouds on the Earth radiation fluxes are of a dual nature: because of a high albedo, clouds reduce the incoming solar radiation; and because they absorb a large fraction of the Earth's thermal infrared radiation, and re-radiate at a level of their own lower temperatures, clouds reduce the total outgoing thermal radiation, constituting a thermal blanket over the Earth.

Kra is (1973) suggested a significant climatic cooling effect for the low level cloudiness. He stressed this role over the tropical oceans, stating that above 60° latitude the blanketing effect on the infrared radiation predominates and a net warning results. Here a strong cooling role for low cloudiness of two types, fog and convective cumulus, above 40°N* is postulated. The discussion is limited to cloudiness above the open oceans, which, in the absence of clouds, are characterized by a very low albedo.

The albedo of the atmosphere - ocean system in the absence of clouds is of the order of 10%. The exact figure depends on the solar elevation, state of the atmosphere and the sea-state. In actual satellite measurements large regions of atmosphere ocean albedo less than 10% were observed, apparently where cloudiness was practically zero at the time of observations. Considering that we are discussing regions of low sun elevations, a relatively high value of

*At this latitude, both in the Northwestern Atlantic and the Northwestern Pacific low cloudiness increases sharply from South to North, the occurrence of summer fog specifically increasing from insignificant at 35°N to about 30% at 45°N.

albedo of 13% might be appropriate. This value was reported by Kondratyev et al. (1974) for the atmosphere-ocean above the Pacific, Atlantic and Indian Oceans. The albedo of clouds ranges quite widely, depending on type, fractional cloud cover and the nature of the underlying terrain (Mironova, 1973; Budyko, 1974). Here a value of 0.64 was selected, reported by Conover (1965) for dense clouds at an altitude of 500m over the ocean.

Using those two selected albedos, cloudiness effects an increase of albedo from 0.13 to 0.64, or by 0.51. The mean planetary albedo of the Northern Hemisphere, including cloudiness, is 35% (Mironova, 1973). In terms of the radiation balance and the average hemispheric absorptivity of 1-0.35 = 0.65, cloudiness introduces a drastic change in the absorbed solar radiation, by $\frac{0.51}{0.65} \approx 80\%$, multiplied by the fraction of the solar irradiation on the hemisphere that the cloudiness intercepts.

The blanketing effect is due to clouds radiating at a lower temperature than the surface, and thus the blanketing effect depends on the product of cloud top heights and the lapse rate. For a 500m cloud tops, and a high lapse rate of $11^{\circ}C/km$, the clouds radiate at a temperature by $5.5^{\circ}C$ lower than the surface. With the surface at $275^{\circ}K$, this is a reduction in the radiation temperature by 2%, and a reduction in the radiation by 8%.

Schneider (1972) discussed cloudiness as a global climatic feedback mechanism. He stressed the cloud top heights as an important parameter. Extrapolating data in his study, 1.5 km cloud tops can produce a blanketing effect in the infrared flux of only 0.25 cal/cm^2/min, i.e., 6% of the outgoing flux. This is a global average, with a globally averaged lapse rate. When convective cumulus form, lower than average lapse rate prevail, and the blanketing effect is still lower.

The blanketing effect for the type of clouds discussed is thus only a small fraction, smaller even than the uncertainties in the cloud albedo introduced into the albedo reduction calculations. Neglecting the blanketing effect, or rather the difference between the blanketing effect of low clouds and that of a clear atmosphere and treating such low level cloudiness as an albedo adjustment only, is therefore justifiable.

In the polar regions, when average yearly effects by cloudiness, of blanketing and albedo increase are compared, the albedo increase effect has to be multiplied by a fraction: absorbed yearly solar irradiation divided by the sum of the absorbed yearly solar irradiation <u>and</u> the energy transported from the lower latitudes. This enhances the importance of the blanketing, but this question was not examined in detail. It is also quite obvious that at the high latitudes it is the summer cloudiness, with the high values of the solar irradiation prevailing in the region, that is important as a cooling agent, and not the winter cloudiness. Detailed examination of this seasonally variable climatogenic importance of cloudiness will produce results at 40°N sharply varying from those at 75°N, and it is not undertaken here. Rather, we use, in a different connotation, calculations of the effect on hemisphere temperatures of a surface albedo change due to a shift of the ice boundary.

Recently, Flohn (1974), using the Manabe and Wetherald model (1967), computed the hemispheric temperature changes that would result from the changes in the hemispheric surface albedo corresponding to the actually observed or estimated changes in the areal extent of sea ice. For the Northern Hemisphere, with a surface albedo of 0.1373 for the year 1890 as compared with 0.1294 recent, he obtained the temperatures at the end of the warming period higher by 0.95°K, somewhat larger effect than the observed value of 0.6°K. Based on Flohn's work, a numerical example can be presented of the sensitivity of the climate of the Northern Hemisphere to a fractional increase in low cloudiness over all the open oceans north of 40°N. We ask the question, what cloudiness change will reproduce the albedo change discussed by Flohn of $\Delta a = 0.1373 - 0.1294 = 0.0079 \approx 0.008$. Since the area of the circumarctic water comprises slightly more than 1/7 of the hemisphere surface, one has to postulate a change in the cloud cover by 11% to reproduce the effect: $0.11 \times 0.51/7 = 0.008$. Interpolating linearly, a change of only 6.5% in the low cloudiness annual average* would produce the actually observed hemispheric temperature difference of 0.6°K.

Thus, a variation in regional low level cloudiness from about 40% in 1890 to the 34% recent could produce the observed temperature difference. Such a change in cloudiness would most likely go undetected. It is postulated here that a decrease in cloudiness smaller than by 6.5% would trigger a warming trend and an accompanying shift of the ice boundaries, such as actually was observed.

*Since a strong summer enhancement of cloudiness occurs, coinciding with the time of peak solar irradiation, a smaller percentage change is actually required.

DISCUSSION AND CONCLUSIONS

Both Flohn (1974) and Kukla and Kukla (1974) discussed the climatogenic effects of changes in the surface albedo of the circumarctic region. Flohn, as discussed above, computed the hemispheric warming between 1890 and recent, that could be attributed to the reduction in the albedo, due to the shrinkage of the ice areas. Kukla and Kukla linked the climatic anomalies of 1972/73 to the abnormal snow/ice extent in the winter 1971/72. In this paper albedo changes due to fluctuations in low cloudiness, linked to SVOSST which is especially pronounced and subject to fluctuations in circumarctic water, are suggested as a potent climatogenic mechanism.

No proofs are brought up here, about the actual correlation of SVOSST related low cloudiness occurrence and the cooling/warming trends. It can be postulated that such small fluctuation of cloudiness are quite likely. The discussion is very much based on Lamb's studies--and the conclusions reached here are not too different from Lamb's own statements: "The equatorial rainbelt generally . . . was somewhat more active in the 19th century than since and it is possible that slightly greater world cloud cover contributed to lessening the intake of solar radiation, though it does not seem likely that this change was as great as resulting from the greater extent of snow and ice. In addition to these two effects, therefore, reduction of the radiation supply itself by something of the order of 0.5-1% may have to be involved to account for the coldness of the climate around 1800 — whether this was due to a variation in the output of the sun or the frequent volcanic dust veils in the stratosphere or both." (Lamb, 1964, pp. 345-346.)

The possibility of climatogenic action by SVOSST related cloudiness is most likely over the circumarctic oceans above 40°N. As discussed in the Introduction, this is the region where largest temperature fluctuations occurred in the climatic changes observed in the past. It is calculated that a regional increase in low cloudiness by only 6.5% can explain, in the absence of any feedback mechanism, a cooling of the Northern Hemisphere by 0.6°K. This constitutes a potent mechanism, but it should be stressed that in the same region a strong positive feedback exists in an increase of the ice areas, which inevitably has to follow a cooling brought about by an increased cloudiness. The actual cloudiness change required to trigger this positive feedback might be only 1 to 2%, which indicates the extreme fragility of our climate. Two foci of especially strong climatogenic activity might exist, both at latitude 45°N, one in the Northwest Pacific, latitudes 150-180°W, and the other in the Northwest Atlantic, latitudes 40°-65° E. In these regions low cloudiness is commonly occurring, the SST's show strong gradients and also have been particularly likely to show anomalies (Lamb, 1972).

A cooling due to cloudiness reinforced by the positive feedback of ice formation might proceed in oscillations of the order of 1° until interrupted by a period of reduced SVOSST. Such a period of reduced SVOSST can be well expected to occur after some years due to the SVOSST temporal variability, possibly linked to the strength of the flow across the sills. In the absence of knowledge of specific causes of SVOSST fluctuations, we would have to regard the climatic fluctuations as random.

The ice ages could be also attributed to cloud cover fluctuations. If the cooling proceeds to be nearly by an order of magnitude higher, than 1°C to 5°C or more (see Table 2), the reduced average temperatures of the ocean reduce the available moisture for cloud formation; cloudiness reduction, a reduced albedo and a warming trend follow. Butzer and Twidale (1966) indicate that first parts of glacials were moist, last parts dry. The suggested feedback would be consistent with their observations in the Mediterranean. However, since regional Mediterranean weather shifts can also explain their observation, not too much weight can be attached to it.

This suggested negative feedback appears to be in an opposite direction to the results obtained by Schneider and Washington (1973). They varied the SST in the National Center for Atmospheric Research general circulation model and found that a ±2°C change in SST produced a positive feedback, i.e., increased SST produced reduced cloud cover which would cause the SST to increase further. However, their model dealt with high level non-convective clouds and the results are irrelevant to cloudiness discussed here. Actual observations indicate an increase of the cloud cover in summer (Sasomori, et al., 1972; Telegadas and London, 1954), which is especially pronounced in northern latitudes (see Fig. 9, after Vowinckel and Orvig, 1970).

The SVOSST linked cloudiness mechanism is presented as a possible cause of climatic fluctuations. In spite of lack of evidence for actual correlation of increased cloudiness with a cooling trend, conclusions can be reached of the need to simulate such processes by general circulation modeling on the one hand and to monitor cloudiness and SVOSST on the other.

There is no established evidence that the solar constant varies*. However, just because it is possible, it was stated by Landsberg (1958) that measurements from satellites of the solar constant (and spectral distribution of the incoming energy) will be one of the most valuable contributions of space research to climatology. Along the same lines, monitoring the albedo of the Earth with an accuracy better than 1%, especially over the ocean in the North Atlantic and North Pacific regions should be the goal of the climatological satellite programs. Monitoring of the cloud temperatures (top heights) and the sea surface temperatures in these regions also should be carried out. The need for monitoring ocean temperature, for long range weather prediction and climatic studies has been forcefully advocated by Namias (1970; 1971), citing in part Namias' own work on the link between ocean temperatures and the general circulation. Here, the desirability to monitor the relatively fine structure is stressed, i.e., sea surface temperatures anomalies even only 100 km in extent. Such monitoring

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*Opik (1968) attributed the rise in mean temperature between 1880 and 1940 (amounting to 0.86°C in the zone 40 to 70° northern latitude, and to 0.47°C in the tropics 30°N to 30°S, for a global average rise of about 0.60°C) to a rise of 0.28 ± 0.06 per cent in the solar constant between 1921-1952. The rise in the solar constant was calculated by Opik from analysis of the Smithsonian observations, but unpublished. Schneider (1975) obtained a striking agreement for the climatic fluctuations with postulated effects of the sunspots on the solar radiation and with the volcanic dust in the atmosphere produced by the volcanic eruptions. On the other hand, Fritz and Angell (1975) could not substantiate any effects of variations in the solar constant linked to sunspots or solar cycle in their studies of stratospheric temperatures in the tropics.

of the sea at the present cannot be completely carried out from satellites (in the thermal IR) because of the cloud cover, especially since the correlation with cloudiness is the important aspect of study. Ground observations should supplement satellite observations — and it is actually suggested that of the numerous buoys being programmed for oceanic research an appreciable fraction should be located in those critical regions of the North Atlantic and the North Pacific.

Even though there are severe limitations on the capability to model the climatic change (Stone, 1974; Smagorinsky, 1974) considerable insight can be derived from such simulation. Wetherald and Manabe (1975) used a three-dimensional dynamical hemispheric model to simulate the effects of a solar constant increase by 2 percent. The statistical equilibrium of this general circulation simulation produced surface temperatures higher by 2 to 4°C at low latitudes and a 4 to 10°C increase poleward of 60°C, for a 3.1°C hemispheric average increase. This model represents only one hemisphere with a completely idealized geography. MacCracken and Potter (1975), using the zonal climate model of the Lawrence Livermore Laboratory, University of California, simulated a uniform global increase of aerosol in the stratosphere, and also a 3% reduction in the solar constant. This two-dimensional model (latitude and the vertical) extends to both hemispheres. Solar radiation reduction produced a surface temperature decrease about 4°K at the equator and 8 to 15°K in both polar regions; above 60-70°. The sensitivity of the polar regions, in Weatherald-Manabe model is apparently due to the ice albedo feedback and to the stable stratification of the atmosphere in these regions; in the MacCracken-Potter simulation the sensitivity apparently lies in the reduced capability for the atmosphere to transport energy poleward as latent heat due to a sharp reduction in surface evaporation and precipitable water. The MacCracken-Potter simulation did not reproduce the assymetry between the hemispheres that the actual climatic past indicates (see Table 2), and which low cloudiness fluctuations are likely to reproduce. Neither of these two models reproduces the seasonal changes, which is an important test of simulations. Simulation of the solar constant changes and Earth's albedo changes by more detailed models seems to be highly worthwhile.

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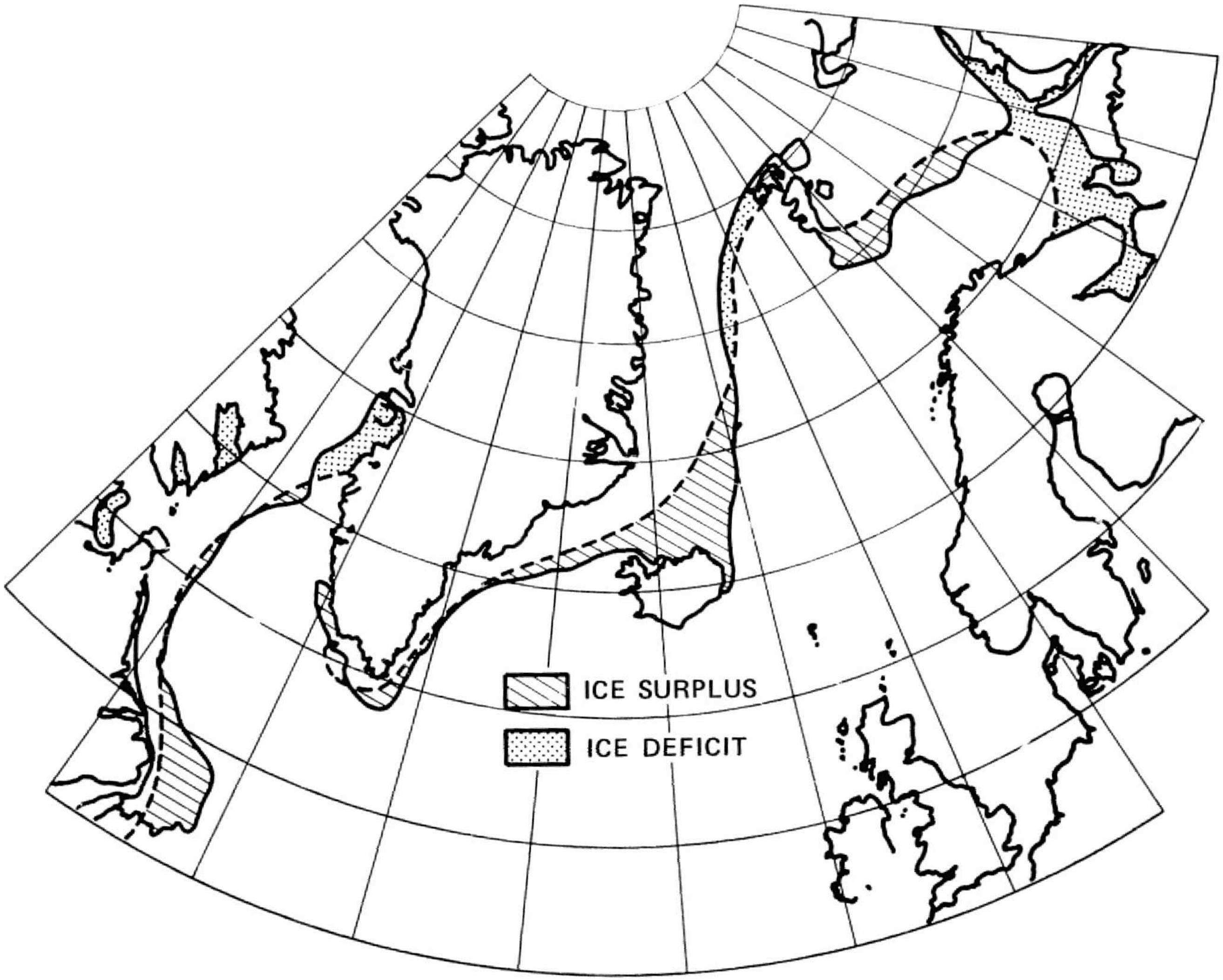

Figure 1. Comparison of Ice Extent at May 8, 1968 (Solid Line) vs Average for the Year 1911-50 (Broken Line), from Dickson and Lamb (1971)

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Figure 2. Cumulus Formation in the Sea of Fernando de Novonha, Western North Atlantic,
Skylab 4 Image SL-4-138-3874

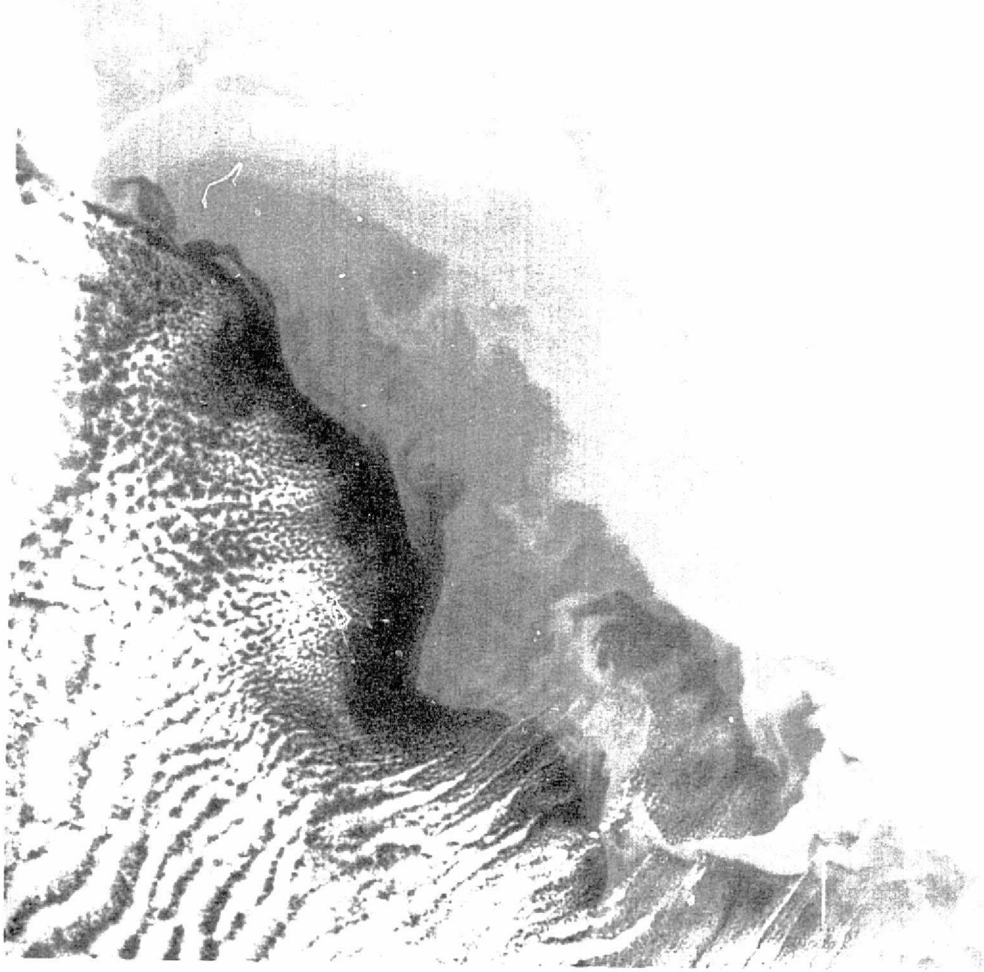


Figure 3. Thermal Image of North Atlantic Showing Cumulus Formation and the North Wall of the Gulf Stream - Coast of USA from Massachusetts to North Carolina Visible at Left, Feb. 22, 1975

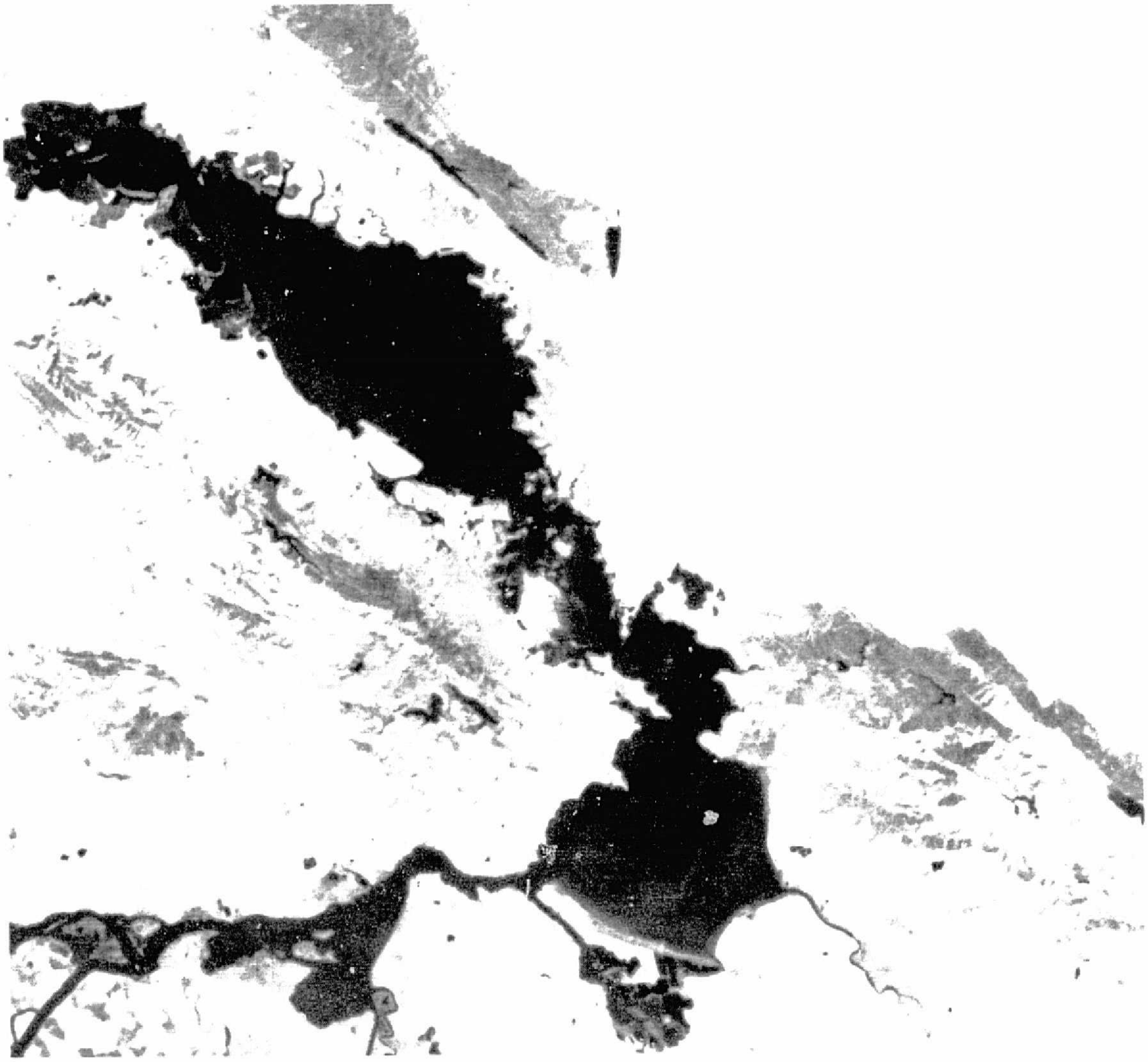


Figure 4. Summer Morning Fog Over the Pacific, Hugging the Shoreline, Lower Left LANDSAT Image E-1003-18175, 26 July 1972

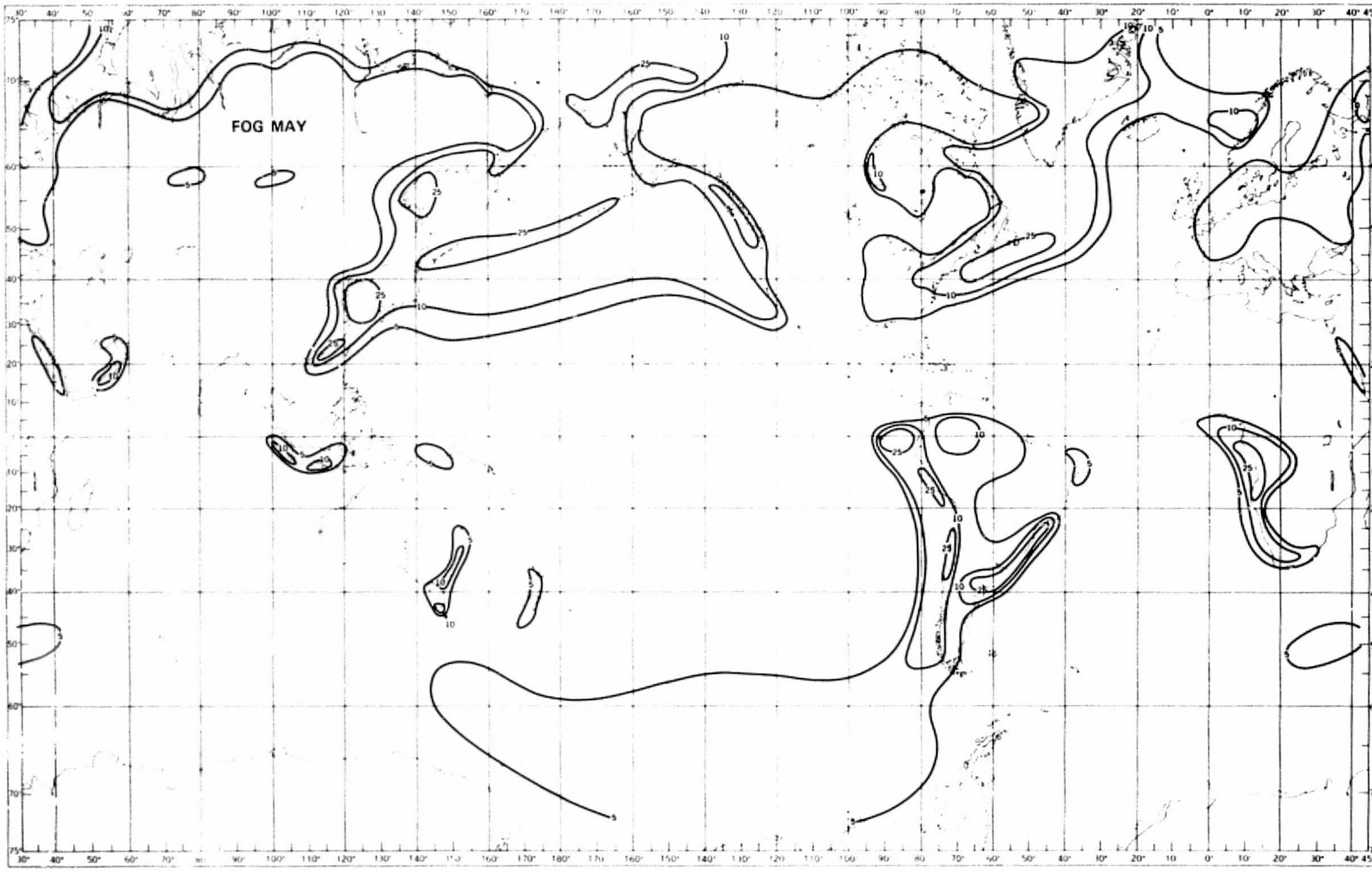


Figure 5. Percentage Frequency of Occurrence of Fog, May, after Guttman (1971)

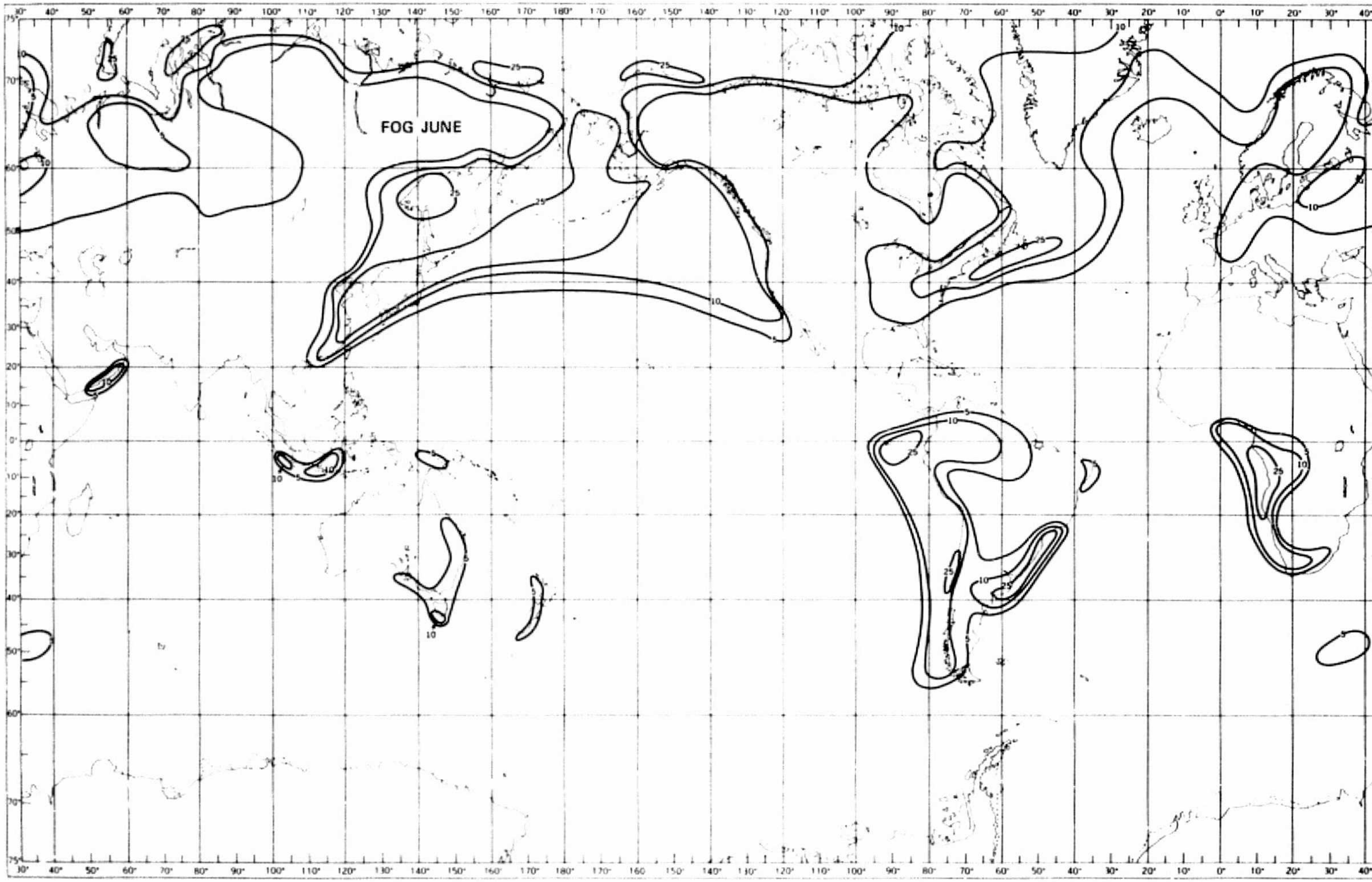


Figure 6. Percentage Frequency of Occurrence of Fog, June, after Guttman (1971)

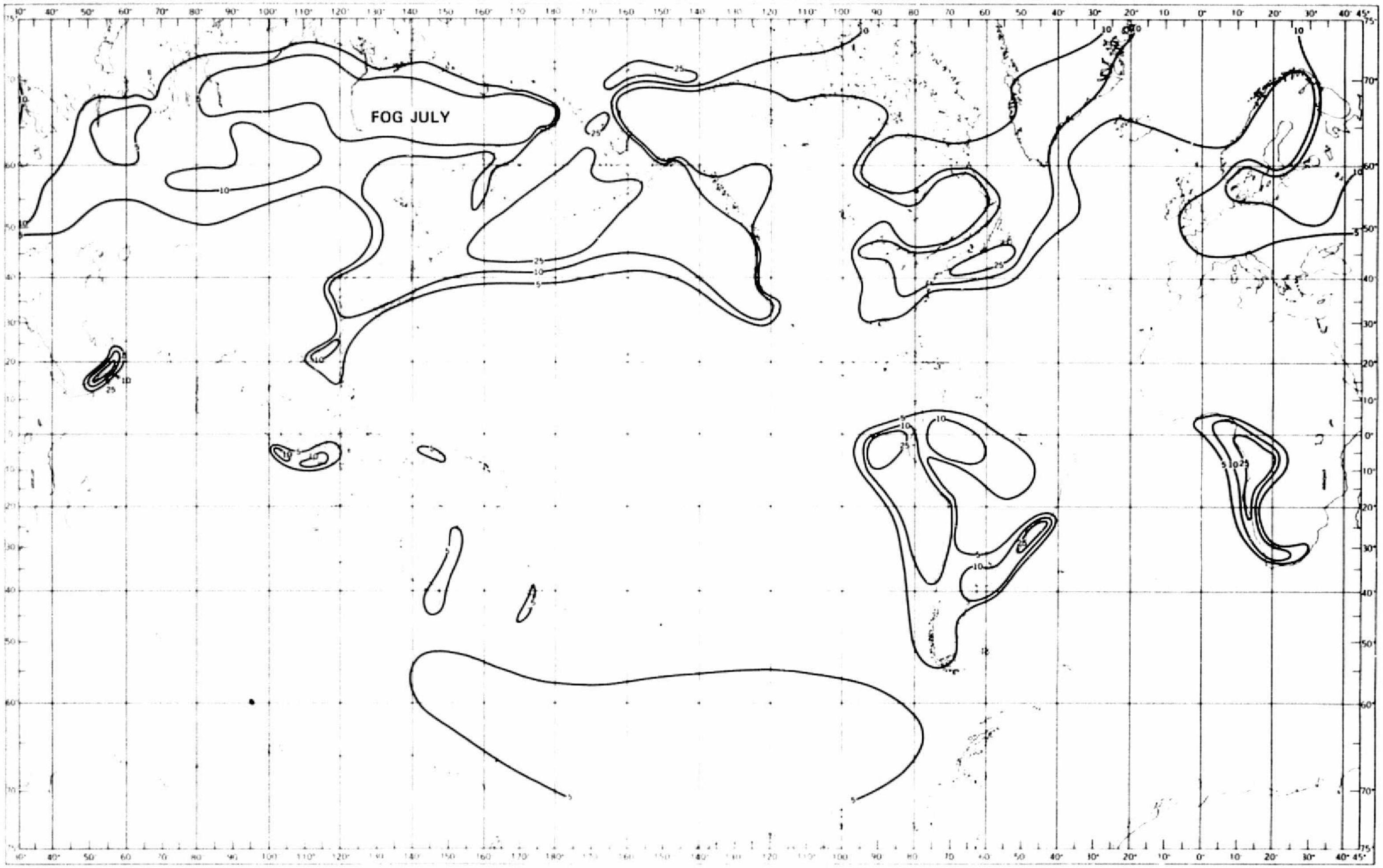


Figure 7. Percentage Frequency of Occurrence of Fog, July, after Guttman (1971)

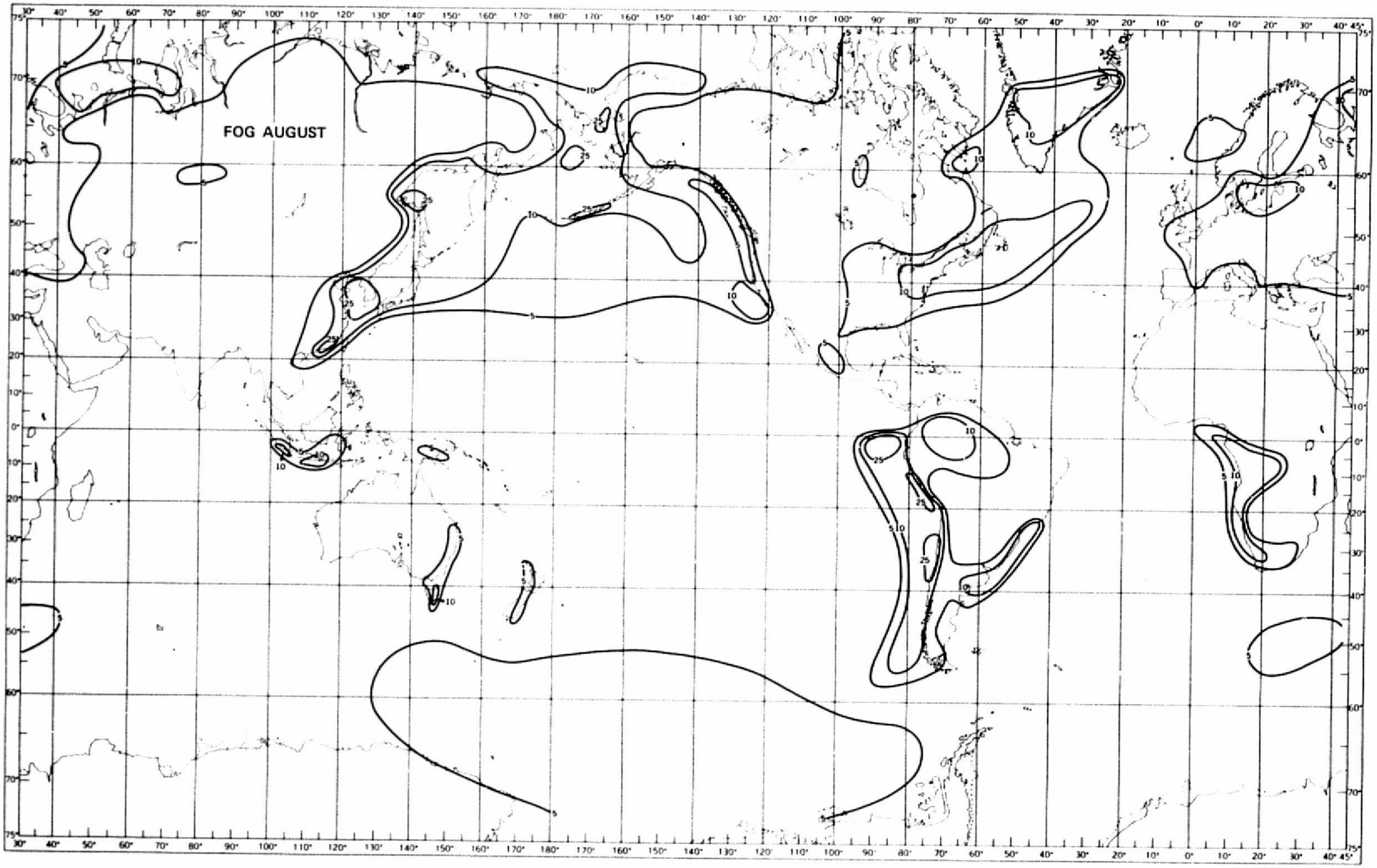


Figure 8. Percentage Frequency of Occurrence of Fog, August, after Guttman (1971)

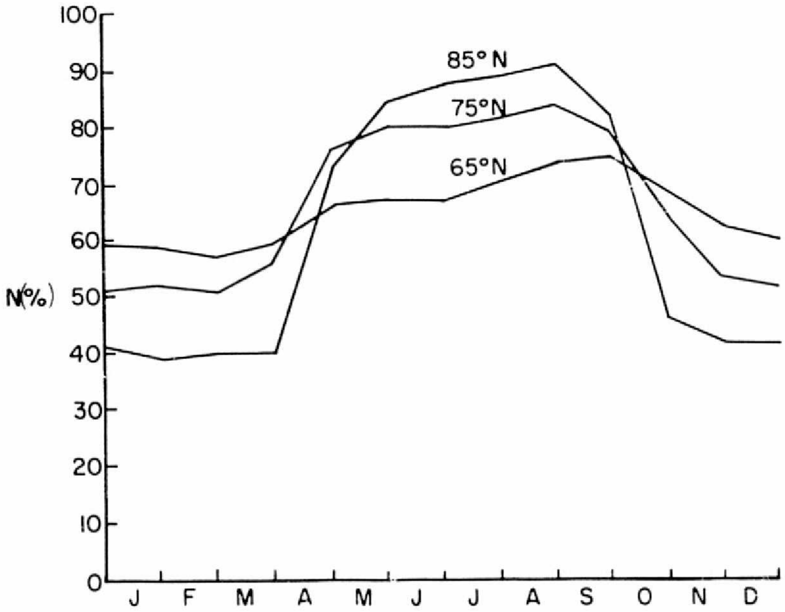
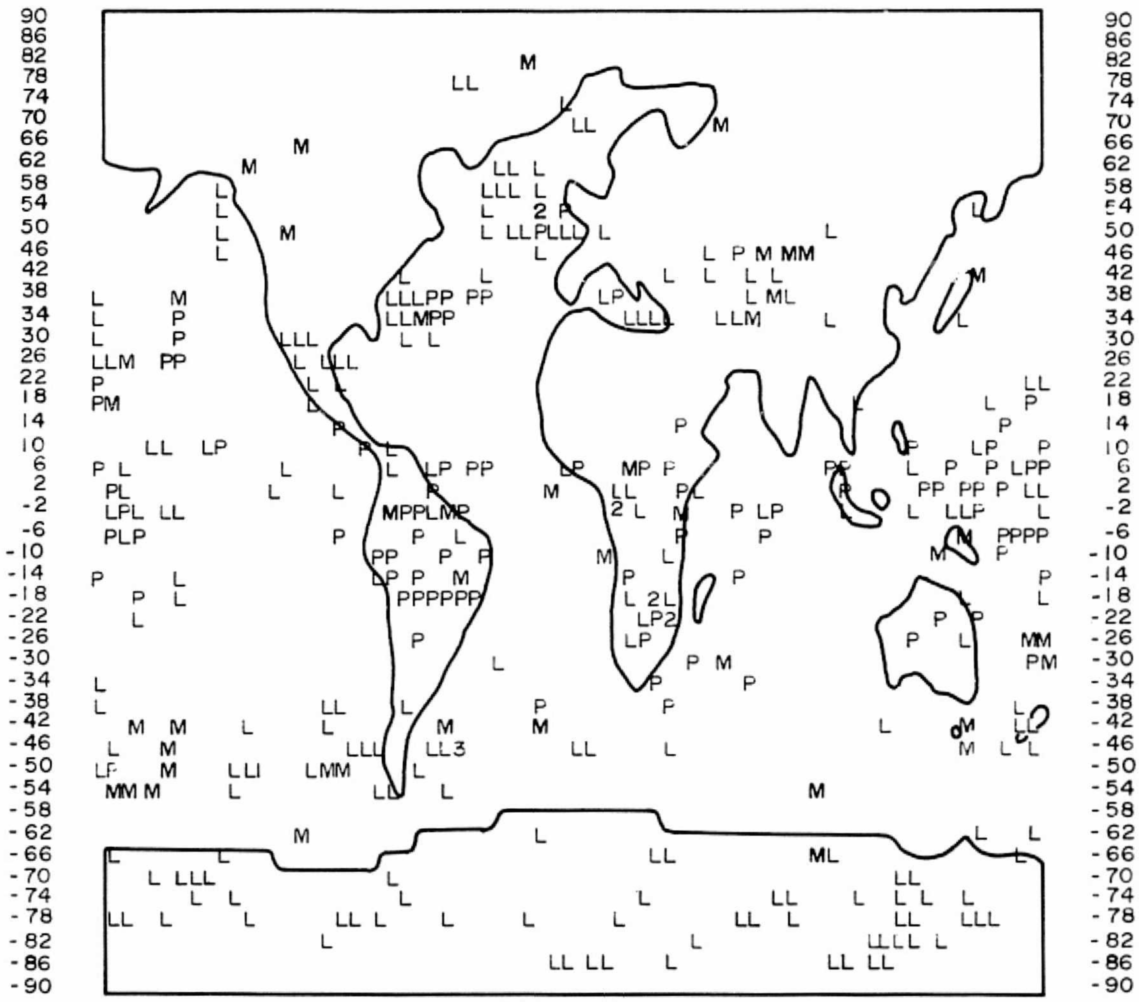
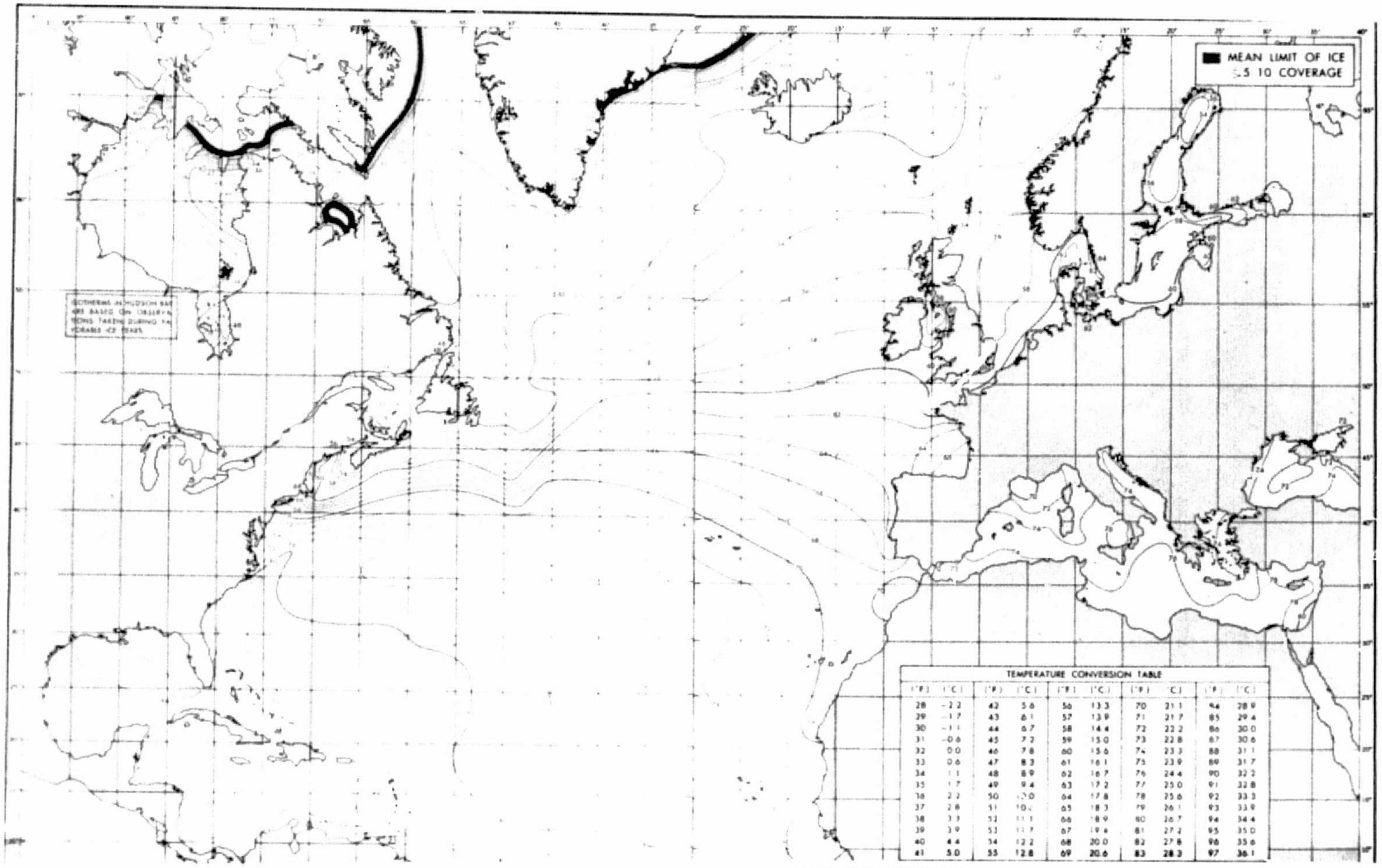


Figure 9. Monthly Latitudinal Means of Cloudiness, after Vowinkel and Orvig (1970)



Legend: L, Low Convection; M, Middle Convection; P, Penetrating Convection; 2, Low and Penetrating Convection; 3, Low, Middle and Penetrating Convection.

Figure 10. Convective Cloudiness, Simulation Results, after Somerville et al. (1974).



(°F)	(°C)	(°F)	(°C)	(°F)	(°C)	(°F)	(°C)	(°F)	(°C)
28	-2.2	42	5.6	56	13.3	70	21.1	84	28.9
29	-1.7	43	6.1	57	13.9	71	21.7	85	29.4
30	-1.1	44	6.7	58	14.4	72	22.2	86	30.0
31	-0.6	45	7.2	59	15.0	73	22.8	87	30.6
32	0.0	46	7.8	60	15.6	74	23.3	88	31.1
33	0.6	47	8.3	61	16.1	75	23.9	89	31.7
34	1.1	48	8.9	62	16.7	76	24.4	90	32.2
35	1.7	49	9.4	63	17.2	77	25.0	91	32.8
36	2.2	50	10.0	64	17.8	78	25.6	92	33.3
37	2.8	51	10.6	65	18.3	79	26.1	93	33.9
38	3.3	52	11.1	66	18.9	80	26.7	94	34.4
39	3.9	53	11.7	67	19.4	81	27.2	95	35.0
40	4.4	54	12.2	68	20.0	82	27.8	96	35.6
41	5.0	55	12.8	69	20.6	83	28.3	97	36.1

Figure 11. Mean Sea Surface Temperatures for the North Atlantic, July, after U.S. Naval Oceanographic Office (1967)

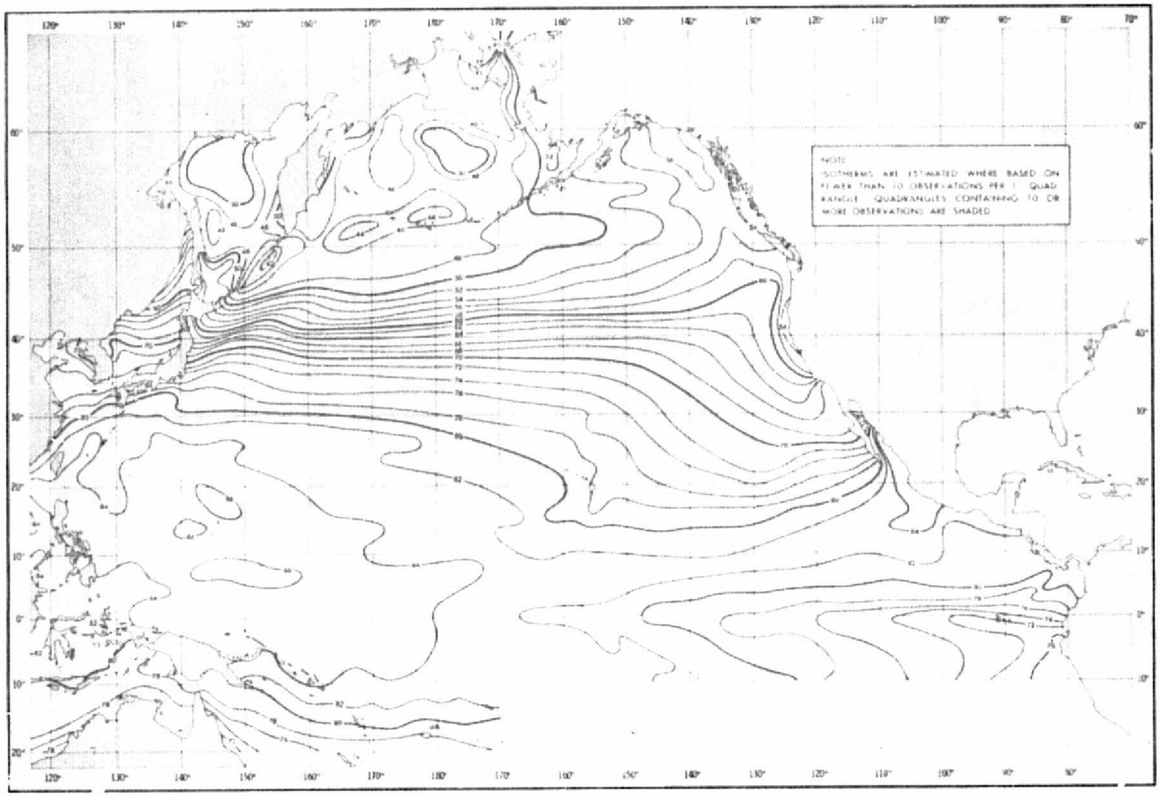


Figure 12. Mean Sea Surface Temperatures for the North Pacific, July, after Laviolette and Seim (1969)