THE METEOROLOGICAL EFFECT ON ICE IN THE GULF OF ST. LAWRENCE

by

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ABSTRACT

The atmospheric circulation over the Gulf of St. Lawrence, for the winter seasons 1960-61 and 1964-65, has been grouped into distinct types.

After a discussion of the equations used and the assumptions necessary in preparing the input parameters, the characteristic evaporative and sensible heat fluxes associated with each mode of circulation have been calculated.

These turbulent heat fluxes were then related to the observed changes in the ice cover, revealing the factors which govern ice formation and decay.

ACKNOWLEDGEMENTS

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CHAPTER I

INTRODUCTION

1.1 PURPOSE OF STUDY

In a continuing programme designed to investigate the factors relating to the growth and decay of ice in Canadian coastal and inland water bodies, the present study deals with the meteorological influence on ice conditions in the Gulf of St. Lawrence.

Owing to the lack of adequate data, very little has been done towards the determination of the complete energy balance of the Gulf of St. Lawrence. It is only in recent years that the data availability has approached the requirements necessary for an investigation of this type. None the less, a preliminary investigation of the heat budget of the Gulf, from mid November to mid February, was carried out by Coombs (1962).

Long-period energy budget studies such as Coombs' have provided, and will continue to provide, valuable information regarding the air-sea energy exchange. The basic concept from which this investigation will be developed has been summarized by Malkus (1962):

> While the average air-sea fluxes over long periods and large regions are computable from and constrained by planetary budget requirements, the fluctuations which build this picture and give rise to the enormous departures from it, so important to human life, are directly governed by transient atmospheric phenomena, themselves the product of circulation instabilities.

Thus, consideration will be directed towards the fluctuations in the energy exchange from the mean conditions, and the consequent control which these undulations have on the growth and decay of ice. This will be accomplished by probing the extent to which variations occur in the energy balance terms due to the variety of synoptic circulation conditions which prevail over the Gulf region.

1.2 DESCRIPTION OF THE AREA UNDER STUDY

1.2.1 GEOGRAPHICAL

The Gulf of St. Lawrence is a virtually land-locked body of saline water which receives the run-off of a well watered, 360,000 mi^2 drainage area (Lauzier, 1957). Mean annual total precipitation along the 2000 mile St. Lawrence River system ranges from 30 to 44 inches, being evenly spread over the year. The Gulf region itself has an annual precipitation range of 32 to 48 inches (Atlas of Canada). It is open to the Labrador Sea through the Strait of Belle Isle, to the Atlantic through Cabot Strait, and to the estuary of the St. Lawrence River. (See location map, Fig. 1.1). The open water area of the Gulf is about 185,000 km^2 (71,000 mi^2).

1.2.2 OCEANOGRAPHIC

1.2.2.1 CURRENTS

Two major surface-currents are in evidence in the water circulation of the Gulf (see Fig. 1.2). The first has its origin in the Labrador Current, a branch of which rounds Newfoundland in a clockwise manner, enters the Gulf through Cabot Strait and dissipates in the disorganized flow of the northern Gulf. It has also been found (Hare, 1950) that a sustained northeasterly wind stress can cause a branch of the Labrador Current to flow into the Gulf through the Strait of Belle Isle.

Fig. 1.1 LOCATION MAP FOR GULF OF ST. LAWRENCE
(see following page)

Station Number	Name	Elevation (feet)
185	Daniels Harbour, Nfld.	64
189	Fox River, Que.	134
197	St. Andrews, Nfld.	35
702	Summerside, P.E.I. (A)	78
706	Charlottetown, P.E.I. (A)	186
707	Sydney, N.S. (A)	197
709	Grindstone Island, Que.	196
717	Chatham, N.B. (A)	112
719	Miscou Island, N.B.	15
808	Cape Whittle, Que.	35
809	Belle Isle, Nfld.	426
810	Port Menier, Que.	18
811	Seven Islands, Que.	190
814	Harrington Harbour, Que.	25
815	Stephenville, Nfld. (A)	44

Legend: A denotes airport.
0 denotes grid points used for numerical computations.

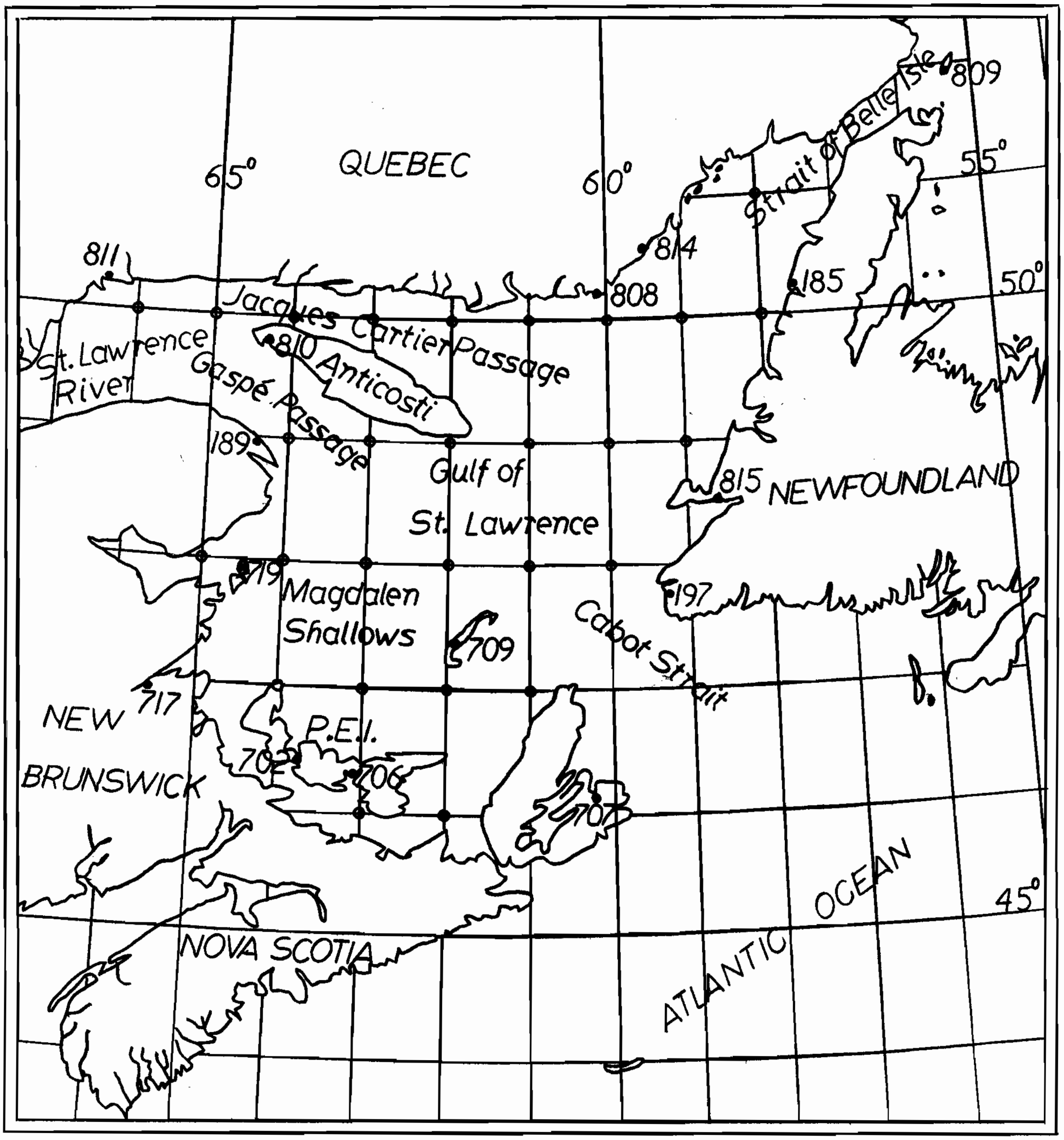

Fig. 1.1 LOCATION MAP (see preceding page)

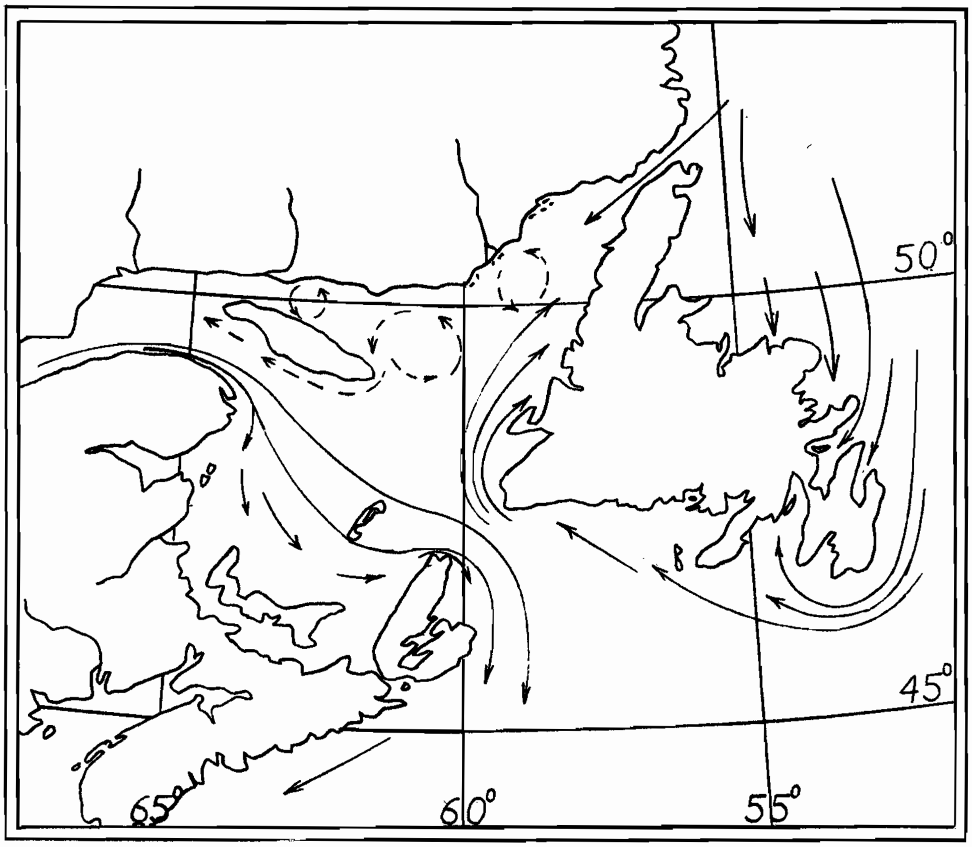

Fig. 1.2 Gulf of St. Lawrence Surface Currents
(after Sandstrom)

The second dominant current is the Gaspé Current, which is relatively warm and fresh due to the fact that it is derived from St. Lawrence River water, ex-Labrador Sea water from the northern Gulf, as well as upwelling of deeper water layers. Normally, this current hugs the Gaspé coast, thence flowing into the southwestern Gulf, turning counter-clockwise and flowing between Prince Edward Island and the Magdalen Islands. Finally it flows out to sea, via the southern portion of Cabot Strait.

1.2.2.2 MEAN ICE CONDITIONS

Based on the five years of ice cover data from 1961 to 1965, published by the Meteorological Branch, five-year mean ice concentration maps have been formulated at twelve fortnightly intervals during the ice season. These mean conditions have been set forth in Figs. 1.3 to 1.14.

It is noted from these charts that freeze-up begins in mid December with fast ice forming in most sheltered harbours, while heavy Labrador Sea pack ice blocks the Strait of Belle Isle. The Gulf freezes over almost completely by late January (Fig. 1.5), and re-opens in late April and May (Figs. 1.11 to 1.13).

During January the concentration of the ice increases rapidly. The region off the west coast of Newfoundland remains unfrozen as a result of the influx of Labrador Current water which, at this time of year, is relatively warm and saline as compared to the fresh water of the Gaspé Current. As a result, the Newfoundland coastal waters are last to freeze (Fig. 1.5) and earliest to break up (Fig. 1.10).

LEGEND

FOR ICE COVER MAPS ON FOLLOWING PAGES

Open water

1-3 Tenths ice cover
very open pack ice

4-6 Tenths ice cover
open pack ice

7-9 Tenths ice cover
close pack ice

10 Tenths ice cover
consolidated pack ice

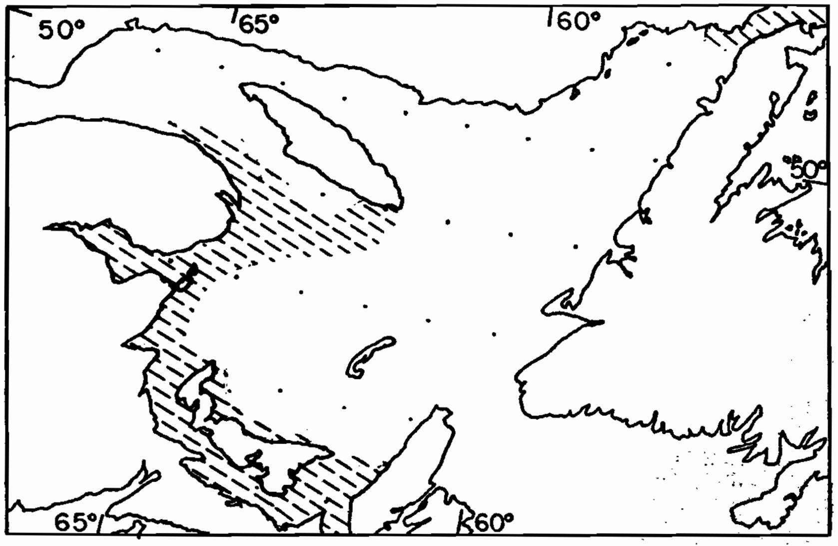

Fig. 1.3 Five-year mean ice concentration on Jan. 1

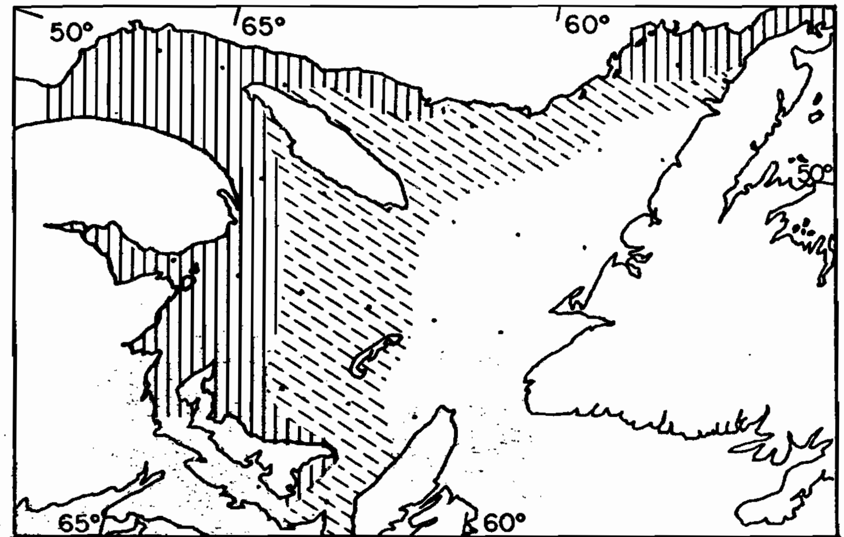

Fig. 1.4 Five-year mean ice concentration on Jan. 15

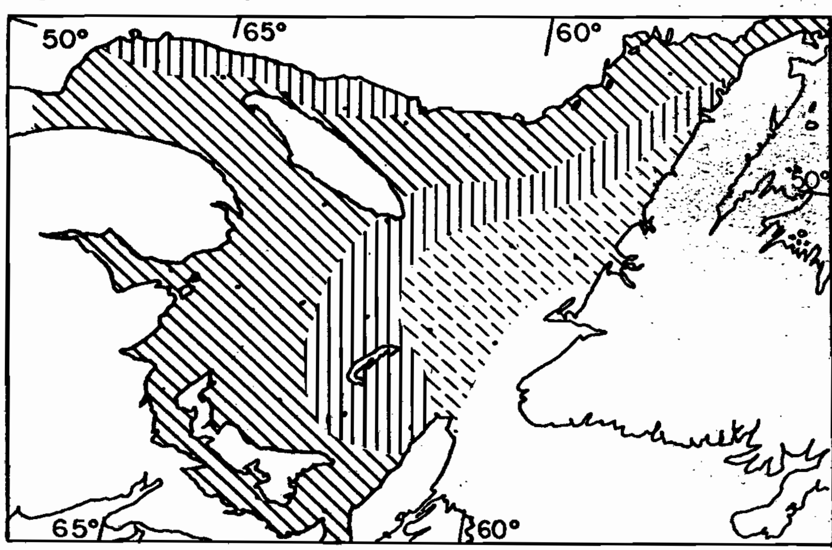

Fig. 1.5 Five-year mean ice concentration on Jan. 29

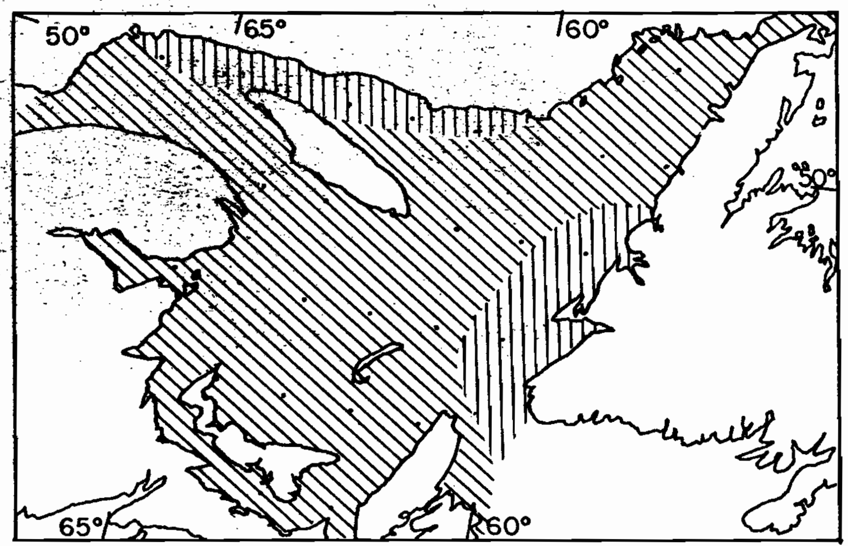

Fig. 1.6 Five-year mean ice concentration on Feb. 12

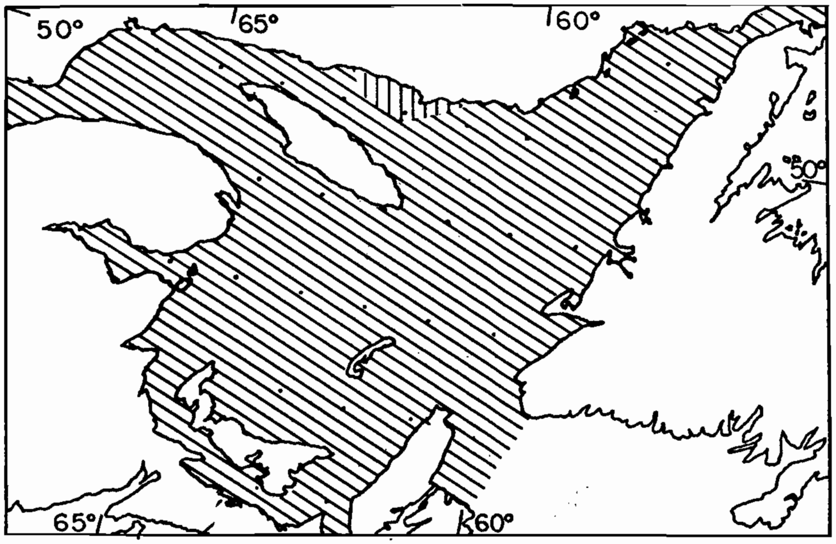

Fig. 1.7 Five-year mean ice concentration on Feb. 26

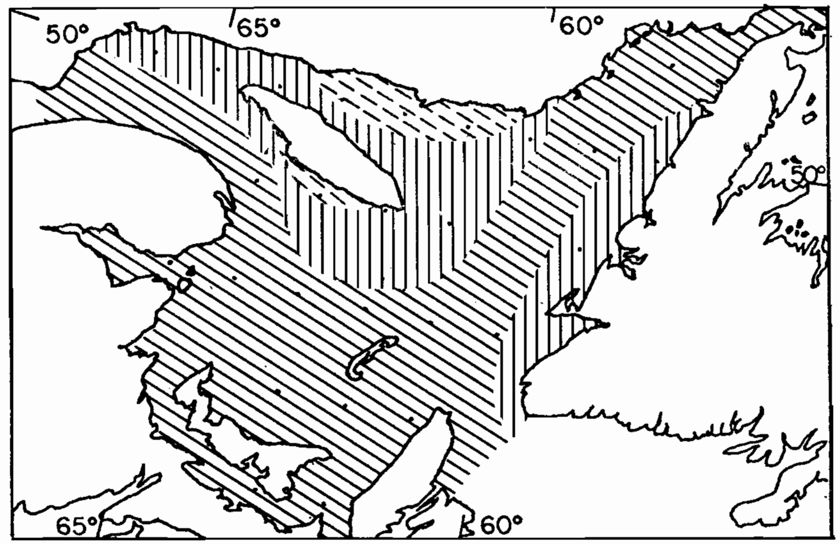

Fig. 1.8 Five-year mean ice concentration on Mar. 12

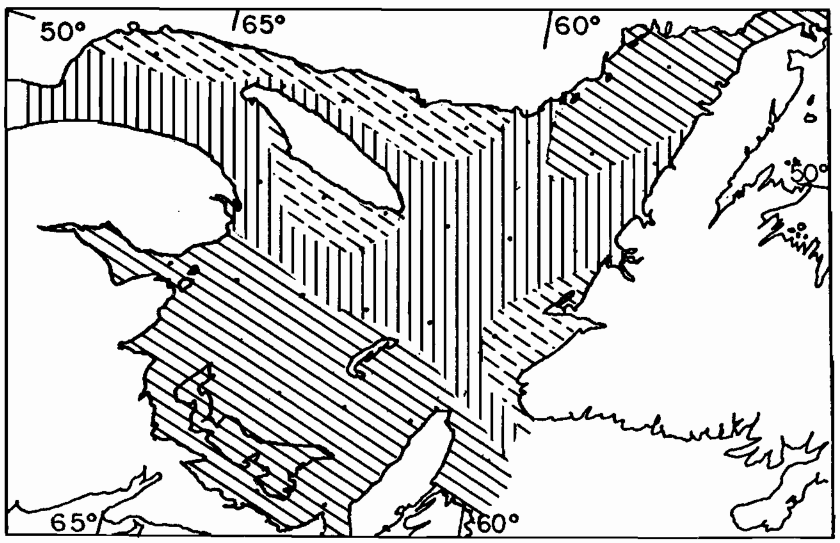

Fig. 1.9 Five-year mean ice concentration on Mar. 26

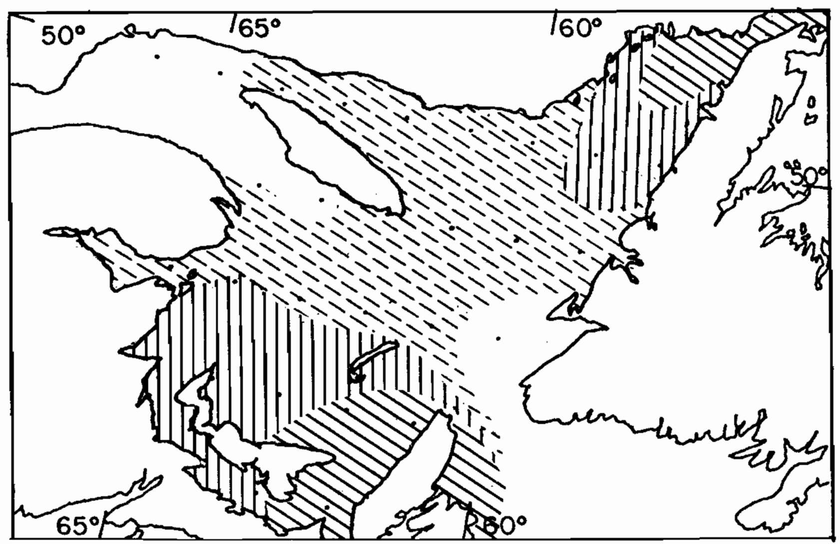

Fig. 1.10 Five-year mean ice concentration on Apr. 9

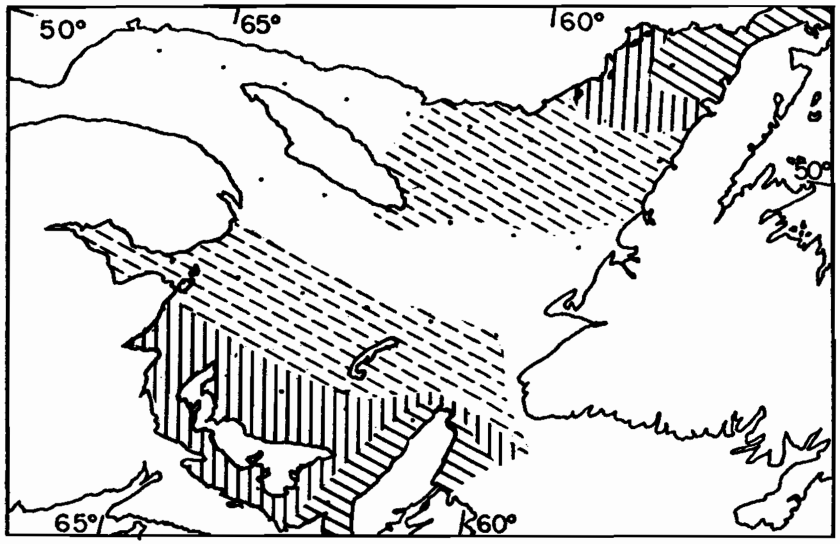

Fig. 1. 11 Five-year mean ice concentration on Apr. 23

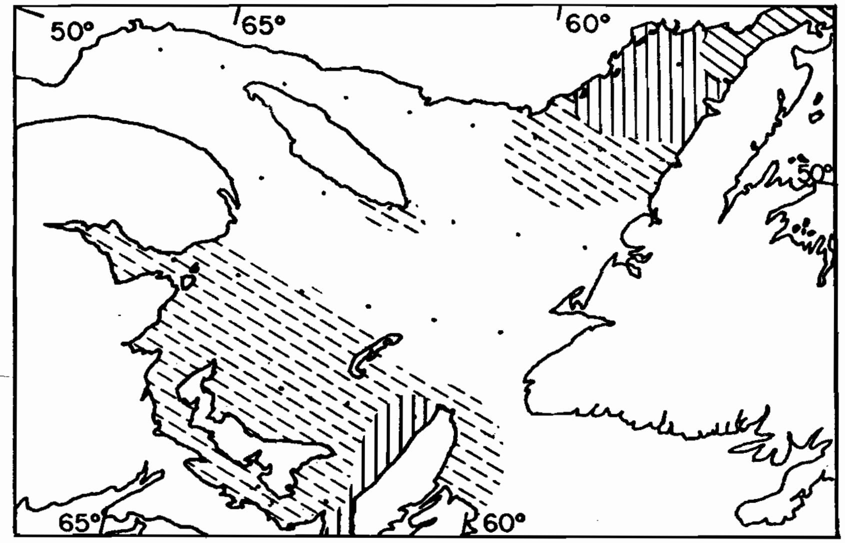

Fig. 1. 12 Five-year mean ice concentration on May 7

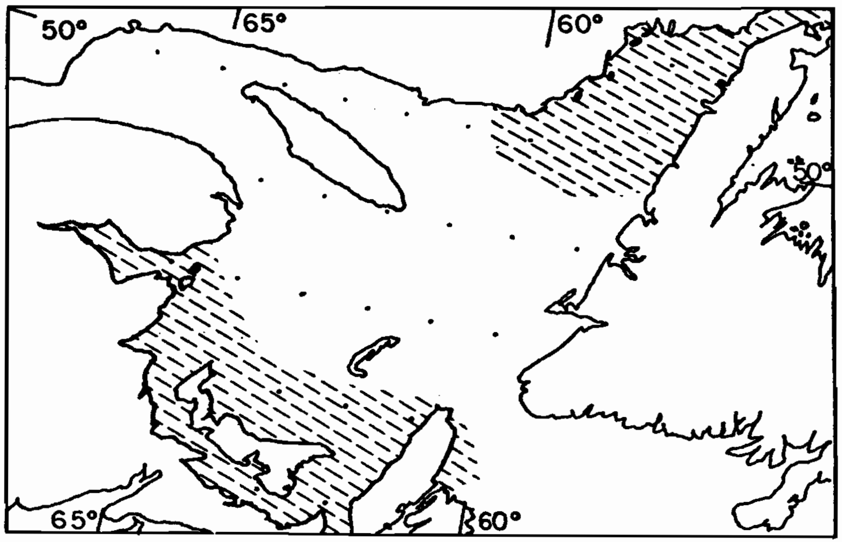

Fig. 1. 13 Five-year mean ice concentration on May 21

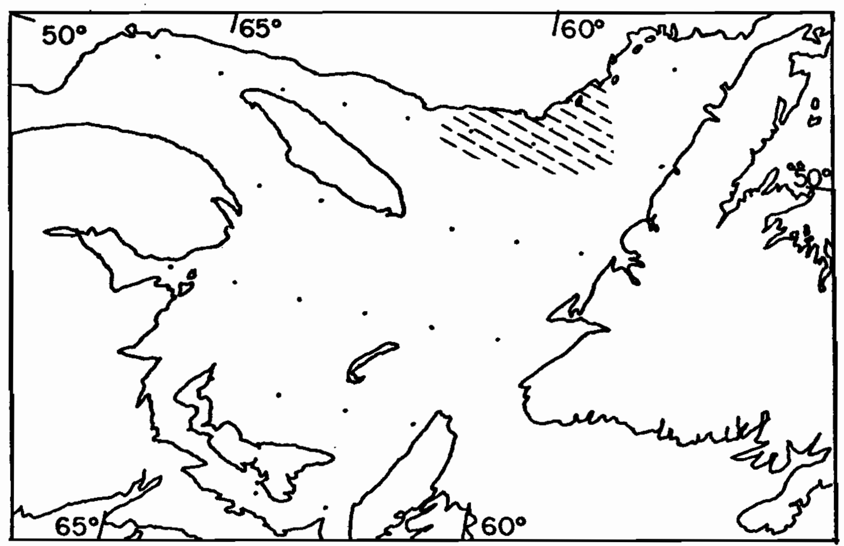

Fig. 1. 14 Five-year mean ice concentration on June 4

Ice charts show that by the last week of January, the south western and central portions of the Gulf are inundated by heavy pack ice. This is ice which has originated in the St. Lawrence River Estuary, and has drifted into the central Gulf under the influence of winds and currents. This pack ice reaches the thick winter stage sooner than ice which forms in New Brunswick and Prince Edward Island coastal waters. As winter progresses, a larger and larger volume of this pack ice is discharged through the southern portion of Cabot Strait, reaching peak amounts in April when the entire Strait may become "bridged". Rapid break-up over the entire Gulf takes place with vernal warming.

A number of the mean maps indicate a decrease in the ice concentration along the North Shore, e.g., Figs. 1.5, 1.6 and 1.7. The individual maps, from which the mean charts are compiled, show that the low densities arise from open leads which have been formed. These open leads are a direct result of the prevailing[1] off shore wind stress.

1.2.3 METEOROLOGICAL

1.2.3.1 UPPER-AIR CIRCULATION

Mean upper-air circulation charts, averaged over periods of three months, and based on the seven years from 1946 to 1952, were abstracted from Henry (1957). The 500 mb map (Fig. 1.15) for the December to February period indicates a generally westerly flow over

[1] See Section 1.2.3.2 on prevailing winds.

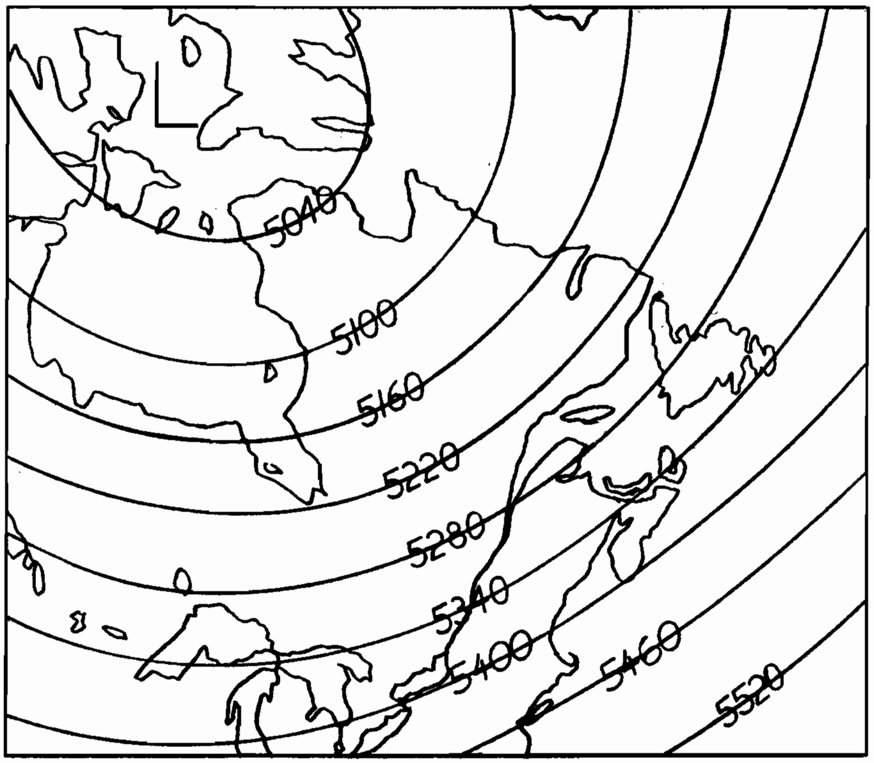

Fig. 1.15. 500 mb Mean Height (in metres) for December to February period.

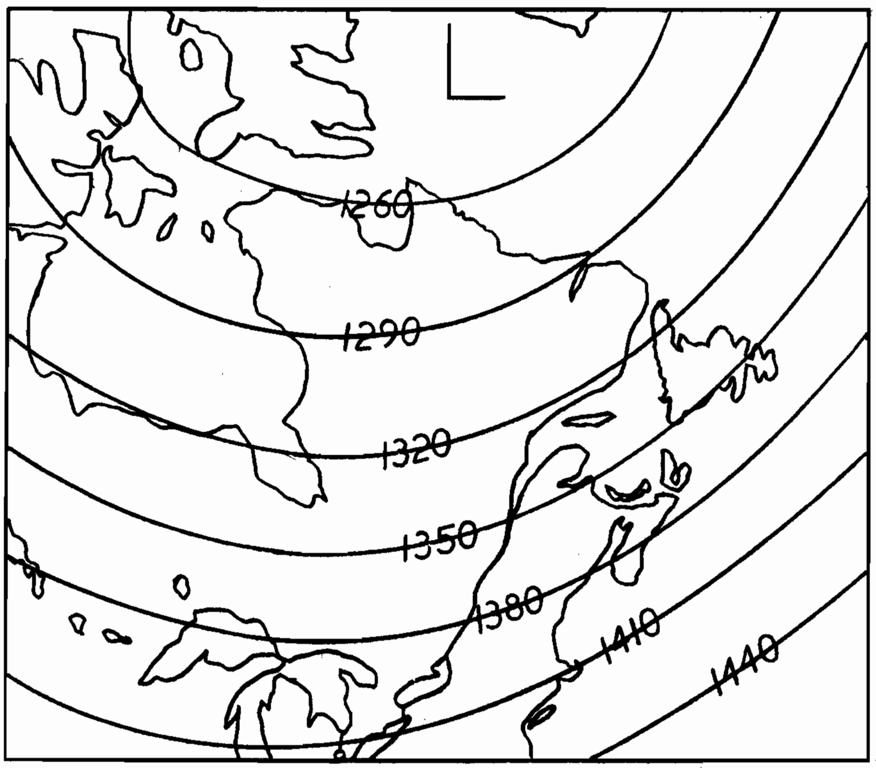

Fig. 1.16. 850 mb Mean Height (in metres) for December to February period.

the Gulf of St. Lawrence region. Lower levels are marked by cyclonic circulation around a low centred over Baffin Island or Davis Strait (see Fig. 1.16). At higher levels a deep trough pervades over the Arctic Islands.

1.2.3.2 MEAN SURFACE PRESSURE AND WIND FLOW

A study of Potter's (1955) monthly mean surface pressure charts (Fig. 1.17), shows that the winter circulation is under the influence of the western extension of the Icelandic Low. This results in a cyclonic circulation over Eastern Canada and a prevailing flow from the northwest over the Gulf. Only minor changes of the pressure pattern occur from November to April. Typical, therefore, of the pattern for the winter months would be the January chart - also the month which exhibits the strongest flow.

1.2.3.3 AIR MASSES AFFECTING THE GULF OF ST. LAWRENCE

With attention directed towards such features of an air mass as its origin, trajectory, stratification, moisture content and its associated weather, the following paragraphs will serve as a general description of the typical air masses found over the Gulf of St. Lawrence.

Continental Arctic and Maritime Arctic Air

The most frequent air masses affecting the Gulf in winter are continental Arctic (cA) or maritime Arctic (mA), the latter being the modified form of cA due to its trajectory over a cold or frozen water-surface. These two air masses have their origins over the

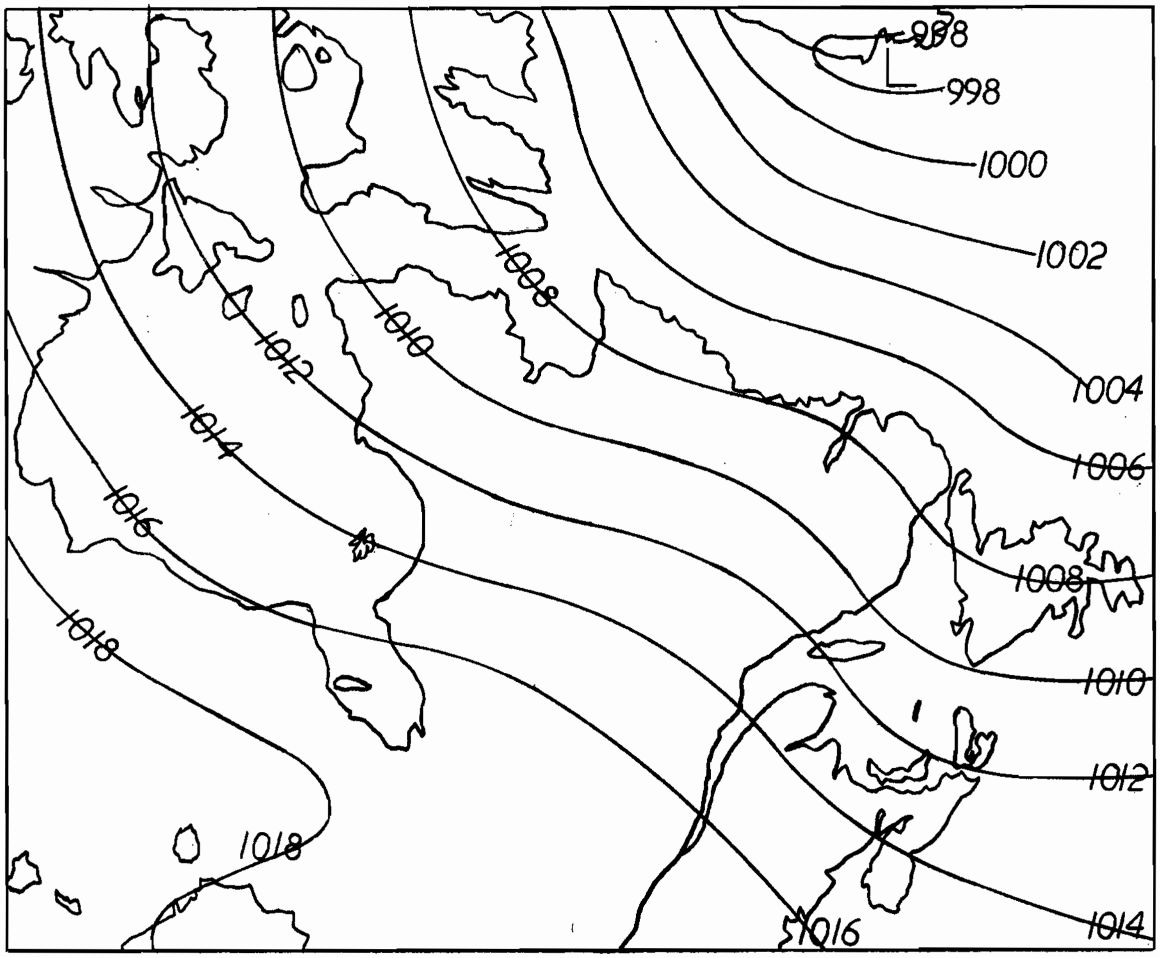

Fig. 1.17. Mean Sea-level Pressure for January.

Greenland Ice Cap, the Arctic Ocean and Archipelago, or over the Northwest Territories. Outflows of cA and mA air usually occur in the rear of deep cyclonic waves travelling along the Arctic Front. A more persistent flow results from a well developed high pressure cell over northwestern Canada, allowing a strong southward flow of cA air along its eastern flank.

The cA air mass is typified by its intense dryness and very low temperatures. At its origin cA air is very stable and skies are cloudless. However, after reaching a body of open water, the stable stratification at its lower levels is completely destroyed, and cumulus clouds develop.

It is interesting to note that the highest frequencies of occurrence belong to cA and mA air masses (Hare, 1950). This is due to the dominance of the Icelandic Low in the circulation pattern and the resulting airflow, which enables air masses which form over Greenland and the Arctic Islands to be advected towards the Gulf region.

Maritime Polar Air

The chief warm air mass affecting the Gulf is not of tropical origin but is cA or cP air which has swung southeastwards out of Western Canada and out over snow-free regions, has become moist and warm, and is returning northeastwards towards the Gulf as an mP air mass. This type of air often forms the warm sector of those cyclones which bring most of the rain and snow to the Gulf region.

As the air mass moves poleward the chilling of the surface layers results in increasing stability of these layers while upper levels remain moderately stable. The moisture content of mP air is variable, generally surface layers are moist and approach saturation, while the middle and upper troposphere are often very dry.

The prime source region of maritime air along the east coast is the Atlantic. During winter months the high latitudes of the North Atlantic remain unfrozen with surface temperatures above the freezing point and much warmer than the continent to the west. Should cP air pass over these warmer waters, the lower layers are rapidly modified to $(mP)_{Atl}$ air, i.e., an air mass which is mild and moist at the surface and is always highly unstable. This results in heavy cumuliform cloud and wintry showers, with temperatures hovering around the freezing point. $(mP)_{Atl}$ air may thus introduce notably mild, cloudy weather to the Gulf region.

Eastward travelling cyclones tend to slow down or become quasi-stationary near Newfoundland, thus inducing a northeasterly flow of mP air from the Atlantic.

1.2.3.4 FRONTS AND TRAVELLING CYCLONES

Of importance to the dynamic climatology of a region are the zones of potential frontogenesis, since it is from these zones that frontal waves develop, creating their associated weather. Two such frontal zones affect the Gulf of St. Lawrence region (Petterson, 1956):

a) the American Arctic Front,

b) the Atlantic Polar Front.

Cyclonic waves develop and ripple along these fronts and tend to converge over the Gulf area. In terms of wind, rain or snow, the depressions from the Atlantic become the most vigorous and intense storms which affect Eastern Canada. This is due to the greater horizontal temperature gradient and the abundant moisture supply from the Atlantic. Although they are no more numerous than the "Alberta" lows, they bring far more precipitation to the area.

1.2.3.5 ANTICYCLONES

The Gulf area is traversed less often by anticyclones than by migratory cyclones. These slow-moving systems, which originate in the area dominated by the Polar High, occasionally move across the Gulf, and just as cyclones tend towards the Icelandic Low to the north-east, anticyclones tend to move southeastwards to merge with the Bermuda High.

As a result then, even though the Gulf may, at times, be under the influence of arctic air masses, the anticyclones which originate in these air masses are rarely centred over the area.

Forasmuch as anticyclones are slow moving, they are persistent, producing frigid temperatures and extremely calm weather towards the centre of the cell.

CHAPTER II

ON THE TECHNIQUE OF CIRCULATION ANALYSIS

2.1 SELECTION OF THE WINTERS STUDIED

In selecting the winters to be investigated, a limitation is imposed by the availability of data. Thus, attention must be confined to the 1960's. Consideration of the ice analysis charts published by the Meteorological Branch suggested a study of the 1960-61 and 1964-65 ice seasons, since they provided a most interesting contrast in ice conditions.

The 1960-61 winter season was selected because it offered some of the heaviest ice conditions in recent years. The severe conditions appear to have been due both to below normal daily mean temperatures from December 1 to March 31, and to pile-up of the ice by wind.

From the sixteen year temperature trend at Grindstone Island (Fig. 2.1), it can be seen that prior to the 1960-61 winter, only one season exhibited a lower daily mean temperature, suggesting that low temperatures, at least, were a cause of the heavy ice conditions.

The second period selected for study was the 1964-65 ice season. The Ice Summary and Analysis charts revealed that ice formation was a little slower than in previous winters. This season also presented the interesting case of a daily mean temperature of 22.2°F at Grindstone Island which was below the 1950-65 mean of 23.7°F. Nevertheless, ice conditions were classed as "better than

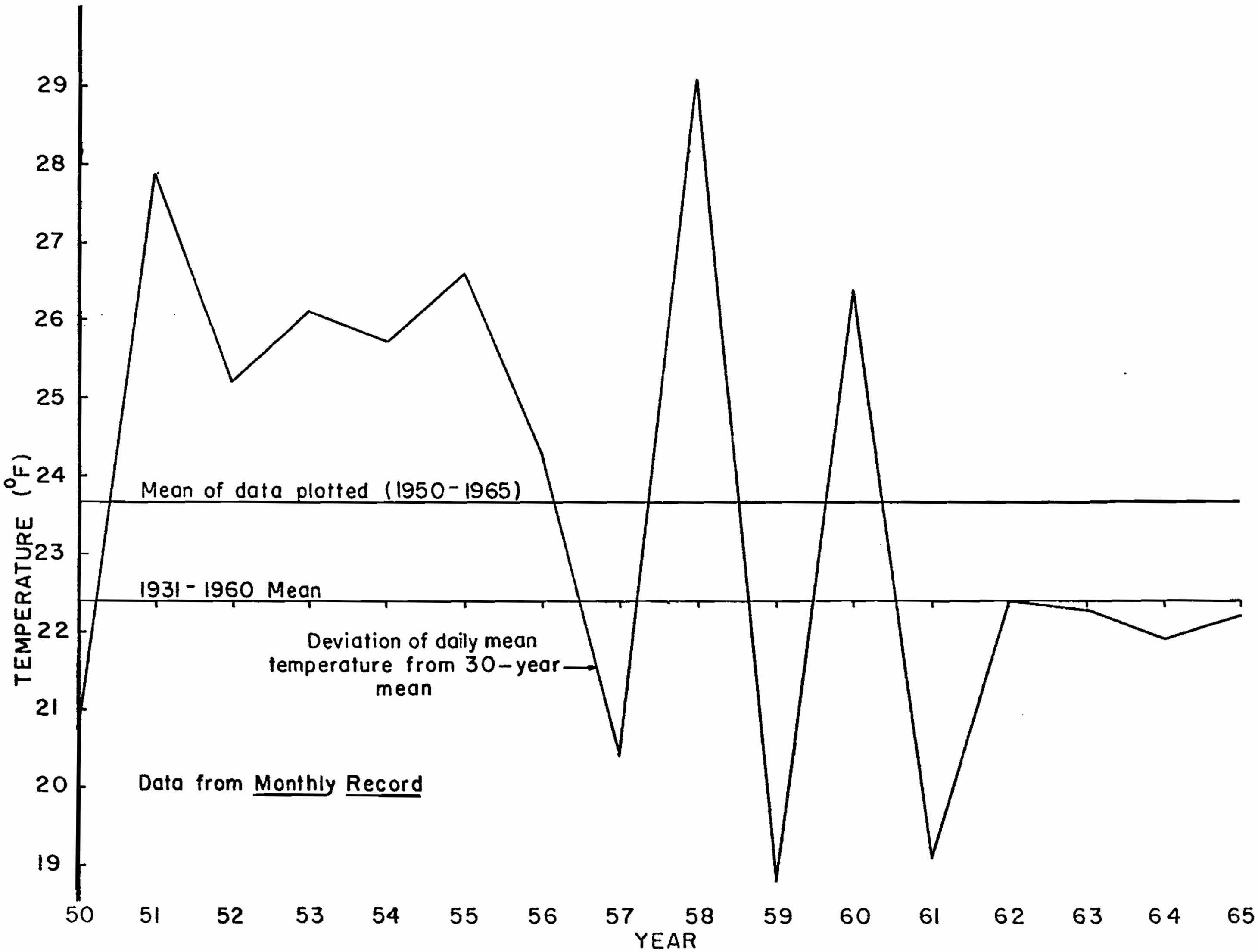

Fig. 2.1 Temperature trend at Grindstone Island for the December 1 to March 31 period.

average".[2] This indicates that daily mean temperature alone is not directly related to ice severity, and so it remains to determine what combination of factors must be employed as an indicator of ice severity.

2.2 DESCRIPTION OF AIRFLOW TYPES

Since the purpose of this study is to evaluate the effects which the various circulation types have on the local energy balance, and hence on the formation of ice in the Gulf, the two winter periods will be examined with reference to the variety of different synoptic circulation types.

It is accepted that the air mass overlying a station at any instant may be composed, at different levels, of air parcels which have had markedly different trajectories. Nevertheless, the classification of airflow types based on surface pressure patterns provides a reasonably close approximation to the complexities of synoptic situations (Barry, 1965).

In section 1.2.3.3 the three dominant air masses which influence the Gulf region during the winter were introduced and discussed. The following paragraphs will be devoted to associating these air masses with definite circulation types.

The classification is first divided as cyclonic or anticyclonic. Cyclonic cases, including transient ridges, are arbitrarily defined as having mean sea level pressures less than 1016 mb and possessing

[2] Ice Summary and Analysis 1965, Eastern Canadian Seaboard p. 40.

cyclonic curvature of the isobars. Anticyclonic types are defined as having mean sea level pressures of 1016 mb or above, with anticyclonic or neutral curvature (adopting the criterion set by Barry (1960).). Barry (1960) was able to distinguish nine different airflow types (cyclonic and anticyclonic) over Labrador. However, such a plurality is not necessary since many of his cases were composed of similar air mass types. It was found possible to reduce the number to four cyclonic and one anticyclonic type.

A description of each of the four cyclonic types, and a discussion of the anticyclonic case, follows:

Type I

This type is characterized by a flow of cA air from the northwest (see Fig. 2. 2), occurring either behind a depression travelling along the Arctic front, or flowing southeastwards along the eastern flank of a well developed high pressure cell over northwestern Canada.

Type II

This case occurs basically in situations of zonal flow at the 500 mb level, accompanied by surface airflow from the west (see Fig. 2. 3). In the situation of an Alberta low developing on the Arctic front, this westerly Type II flow is found in the warm sector of the wave. Considering the Maritime front which lies to the south, this type of airflow exists either ahead of and/or to the rear of the depression.

Type III

The air mass involved in this circulation type is mP, with the airflow either from the southwest or the south (see Fig. 2. 4). Flow from the southwest has had a trajectory over the continent and

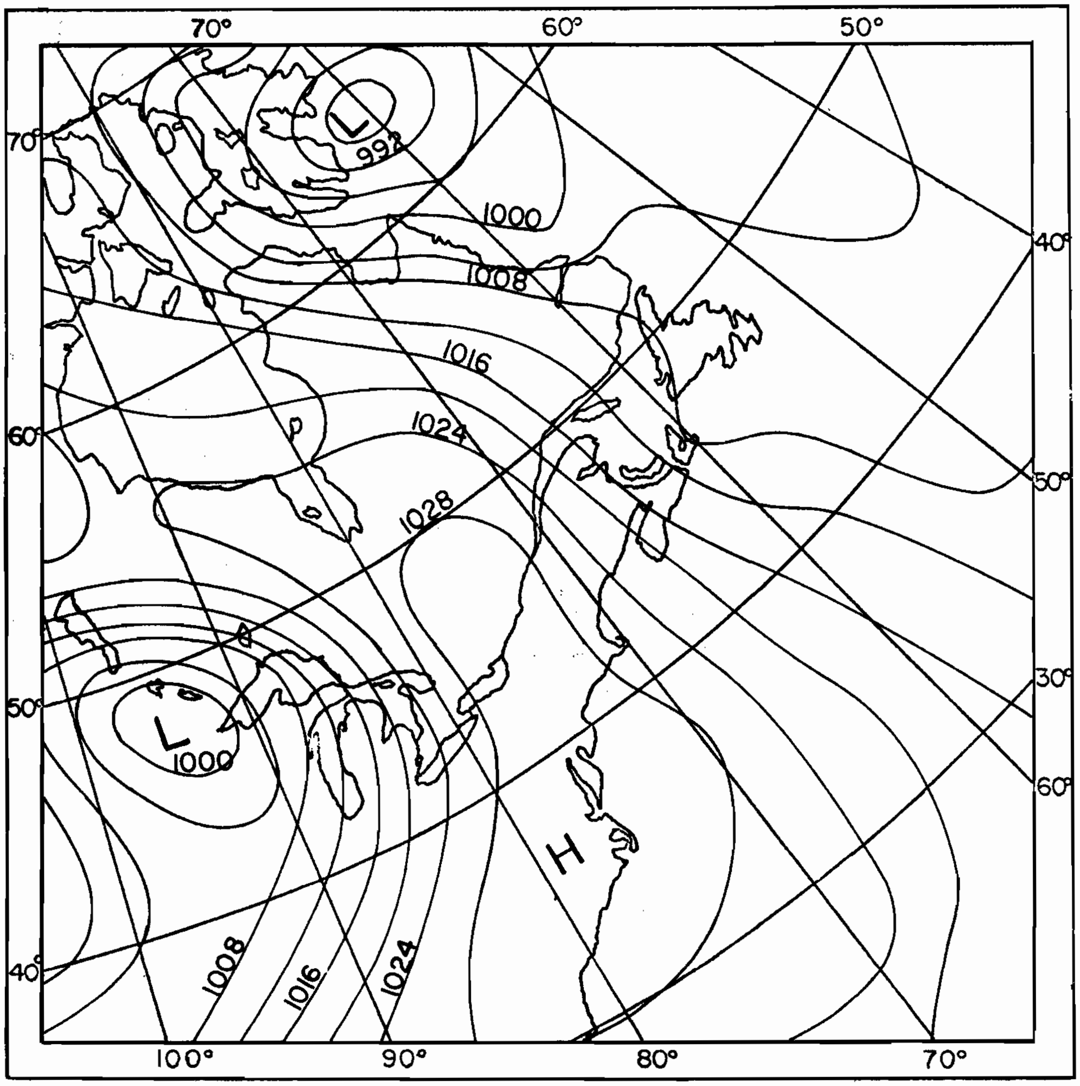

Fig. 2.2 Type I (northwesterly) surface pressure pattern.

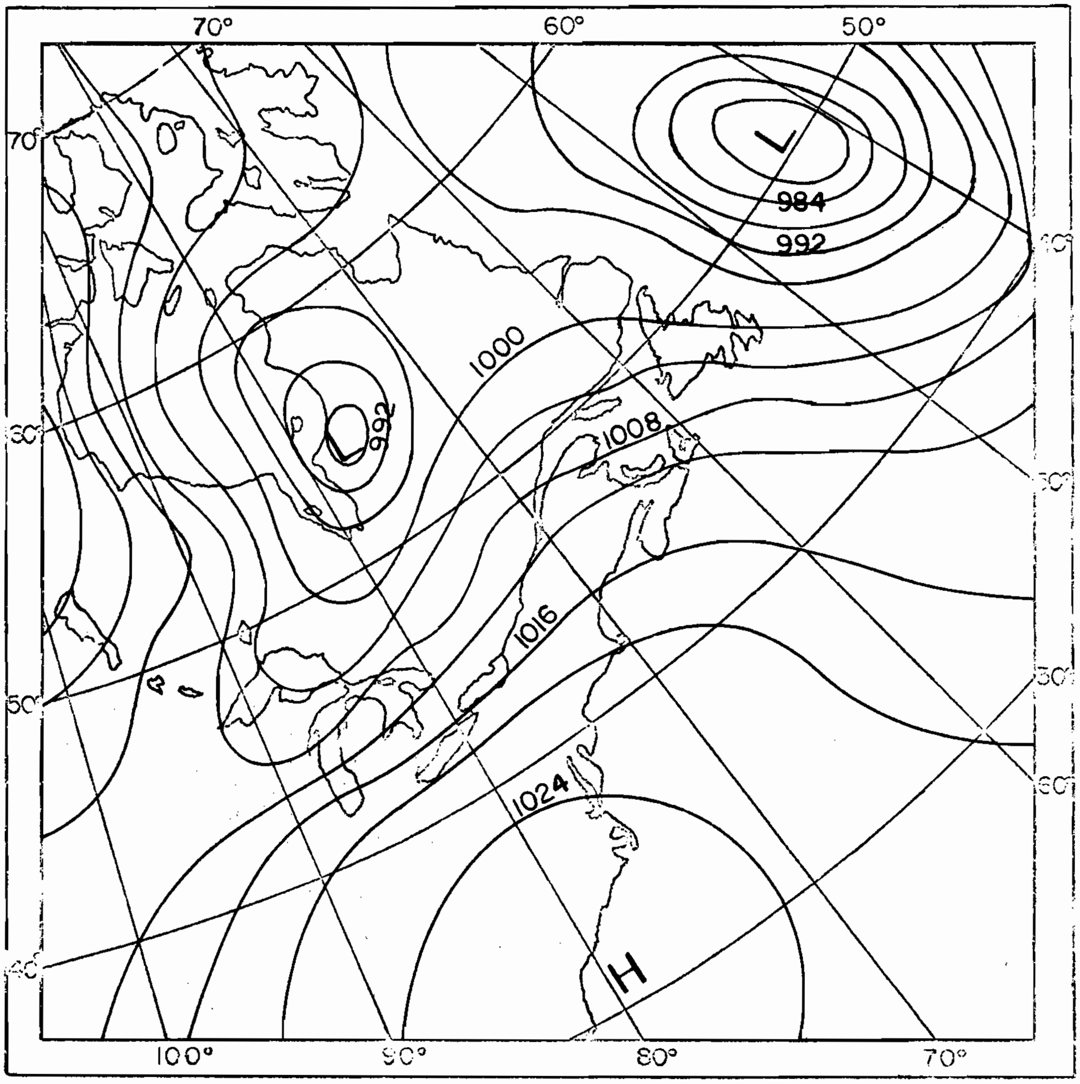

Fig. 2.3 Type II (westerly) surface pressure pattern.

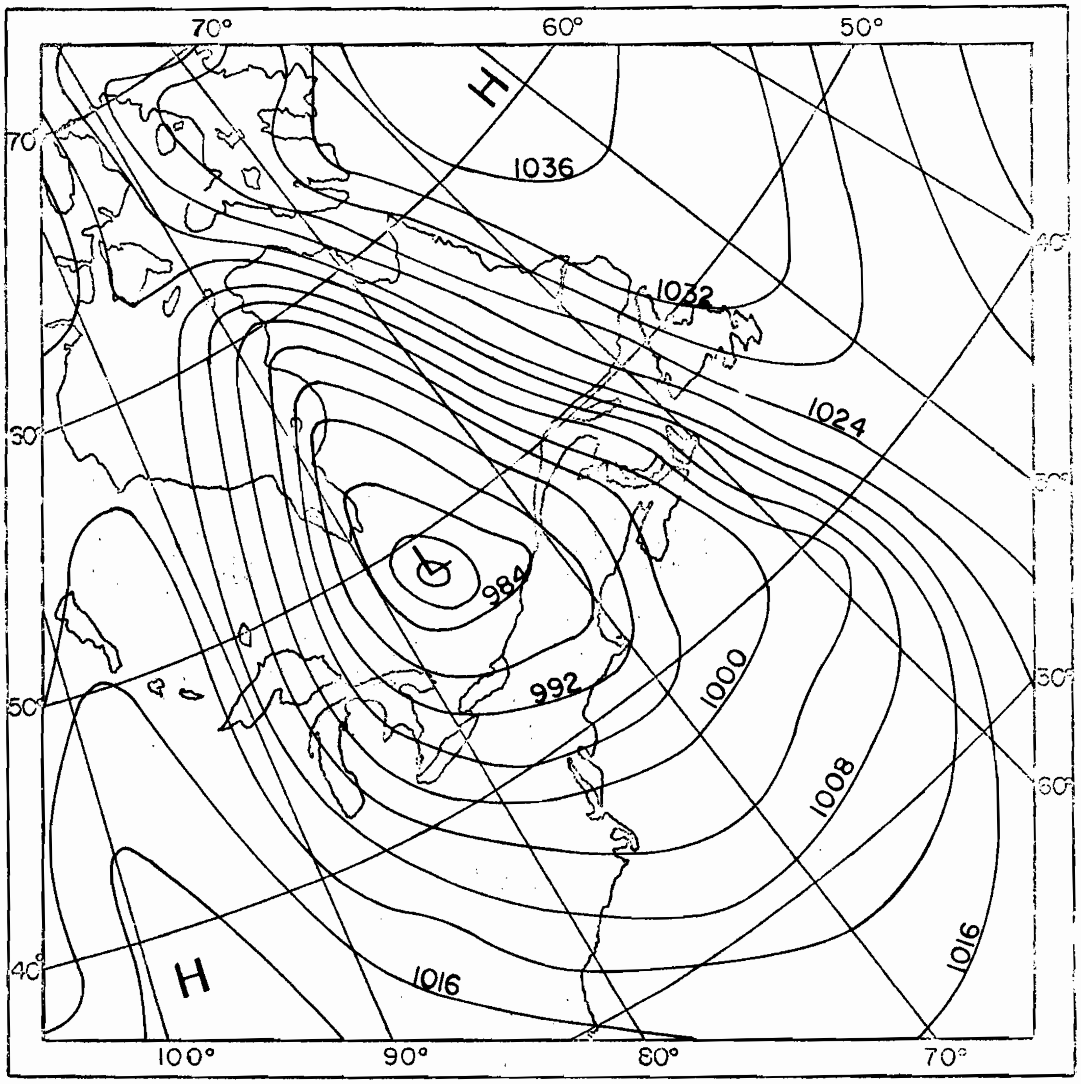

Fig. 2.4 Type III (southerly) surface pressure pattern.

occurs predominantly in the warm sector of cyclones moving northeastward from the Great Lakes region. Southerly flow, on the other hand, is associated with waves on the Atlantic Polar front moving northward along the east coast of North America.

Type IV

This circulation type also consists of mP air; however, it has its source over the North Atlantic. Should a cyclone stall south of Newfoundland - a phenomenon which arises frequently - the southerly wind discussed in Type III will back, becoming easterly as the cyclone occludes. Fig. 2.5 provides an illustration of this type. Similar to easterly flow is airflow from the northeast, which may result from either the backing of easterly flow, or from the rare occurrence of a high pressure area over the Labrador Sea.

A final, infrequent flow type, which may be classed with this group, is northerly flow, which represents the final stage of a series of circulation types beginning with southerly flow, subsequently changing to easterly and becoming northerly. The essential requirement for this series of events to occur is for a quasi-stationary upper-air wave pattern to exist, such that a surface depression may persist off the east coast of Newfoundland.

Anticyclonic.

Since anticyclonic airflow tends to be of diverse origin, and in order to avoid unnecessary complexity, subdivision of the anticyclonic cases was not carried out. Also, the number of cases of sub-types would be too few to be considered as representative samples.

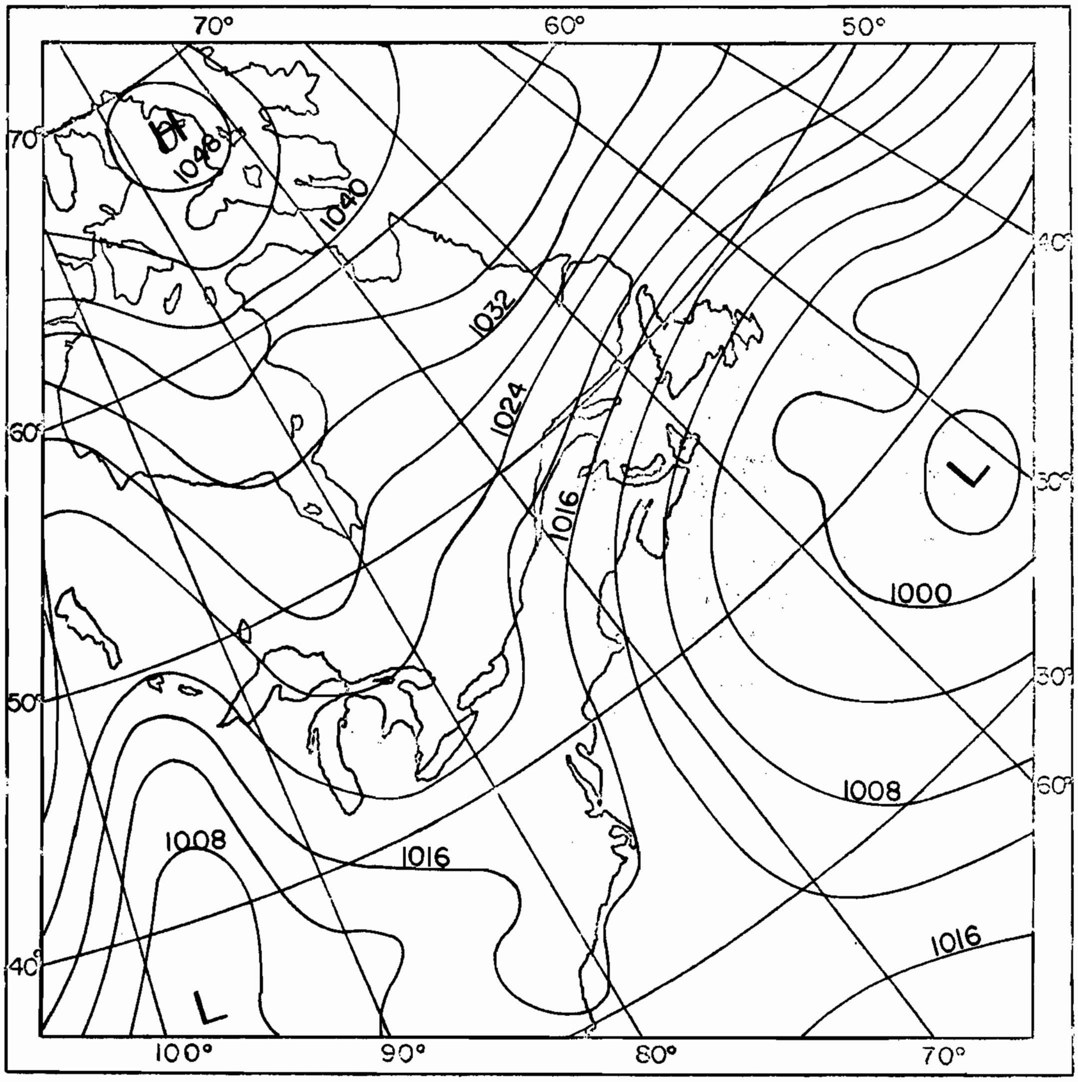

Fig. 2.5 Type IV (northeasterly) surface pressure pattern.

The isolation of the anticyclonic category was carried out merely to provide a comparison to the effects induced by airflow under non-anticyclonic conditions.

2.3 DISCUSSION OF THE CLASSIFICATION METHOD

It now remains to classify each day of the two winter periods as anticyclonic, or as one of the four cyclonic circulation types introduced in the foregoing section.

In carrying out the grouping, use was made of microfilm copies of the 6 hourly surface, and 12 hourly 850 mb and 500 mb analyses, obligingly provided by the Central Analysis Office of the Canadian Meteorological Service. These charts made it possible to determine the trajectory of the air at the surface, and thus identify the circulation type to which each day had been subjected.

On most days the circulation type was readily determined from the surface maps alone. However, for days whose surface pressure charts left the airflow in doubt, a considerable aid in determining the air current trajectory was provided by the upper-air charts.

Consideration was first given to the 500 mb analysis which provided a broad indication of the airflow tendency, i.e., whether the flow tended to be zonal or meridional. The 850 mb chart gave an even more definite suggestion of the trajectory of the air, since this map also indicates the temperature-dew point spread (moisture content) of the air mass. This was a valuable parameter in differentiating between the marine nature of northerly flow and the continental characteristics of northwesterly flow. In instances of light and variable surface winds, the 850 mb flow greatly aided the determination of the airflow direction.

It was found that the transition from one type to another took place frequently and usually quite rapidly. Distinct circulation types were found to last two to three days on the average, while some types persisted for several days, due to a stationary wave situation, or to a cut-off low at the 500 mb level.

Some transitions from type to type took place rather slowly however, and the period during which the change occurred could not be identified as a recognizable circulation type. This resulted in a number of short periods which had to be grouped as "indeterminate". Another case which had to be included in this group existed when a low pressure system remained quasi-stationary directly over the Gulf, which led to the convergence of air from several directions.

There were a number of days during which a warm or a cold front passed over the region. Generally, the cold fronts moved over the area more rapidly than did the warm fronts, and cases of fronts stalling over the Gulf were practically nil. For those days on which frontal passage occurred during the earlier or later part of the day, the passage was retarded or advanced to the beginning or the end of the day respectively, according to the support given by the upper-air data. This is permissible since in any energy balance study one is primarily concerned with the temperature and moisture structure of the atmosphere. On days during which frontal passage took place towards the middle of the 24 hour period, creating the existence of two distinct airflow trajectories, each type was given a "half-day" weight.

Certainly all the characteristics of a particular circulation type will not be found unless the air current maintains itself for an extended period of time. Arising from this fact, the criterion is chosen that the same airflow must be sustained for at least 48 hours, in order that it qualify as a distinct and representative type.

CHAPTER III

RESULTS OF THE CLASSIFICATION

3.1 PRELIMINARY OBSERVATIONS

In Fig. 3.1 are set forth the results of the allocation of a specific flow type to each day, or half day, of the two winter periods, in a notation intended to create a rapid mental picture of the airflow direction. When displayed in this form, the picture does not particularly help in a comparison of the two seasons. Nevertheless, it does indicate the high frequency of change-over from type to type; also the average duration of a specific airflow type is seen to extend over a day to a day and a half, although some cases were maintained for much longer periods.

It is also worth noting that, of the eight cases of flow duration of 5 days or more, all but one occurred during the last two months of the ice seasons, i.e., March and early April. This phenomena of increased duration of flow towards spring may be attributed to blocking action over the Atlantic. Rex (1950) has shown that there is a seasonal variation in blocking activity over the North Atlantic, and that these blocks may be as near as 40°W to the Gulf. This activity increases through the winter season to a peak in April. Indeed, perusal of the upper-air maps revealed that blocking was hindering the eastward advance of the surface systems.

The most logical feature of the classification to examine for comparative purposes would be the total occurrence of each flow type

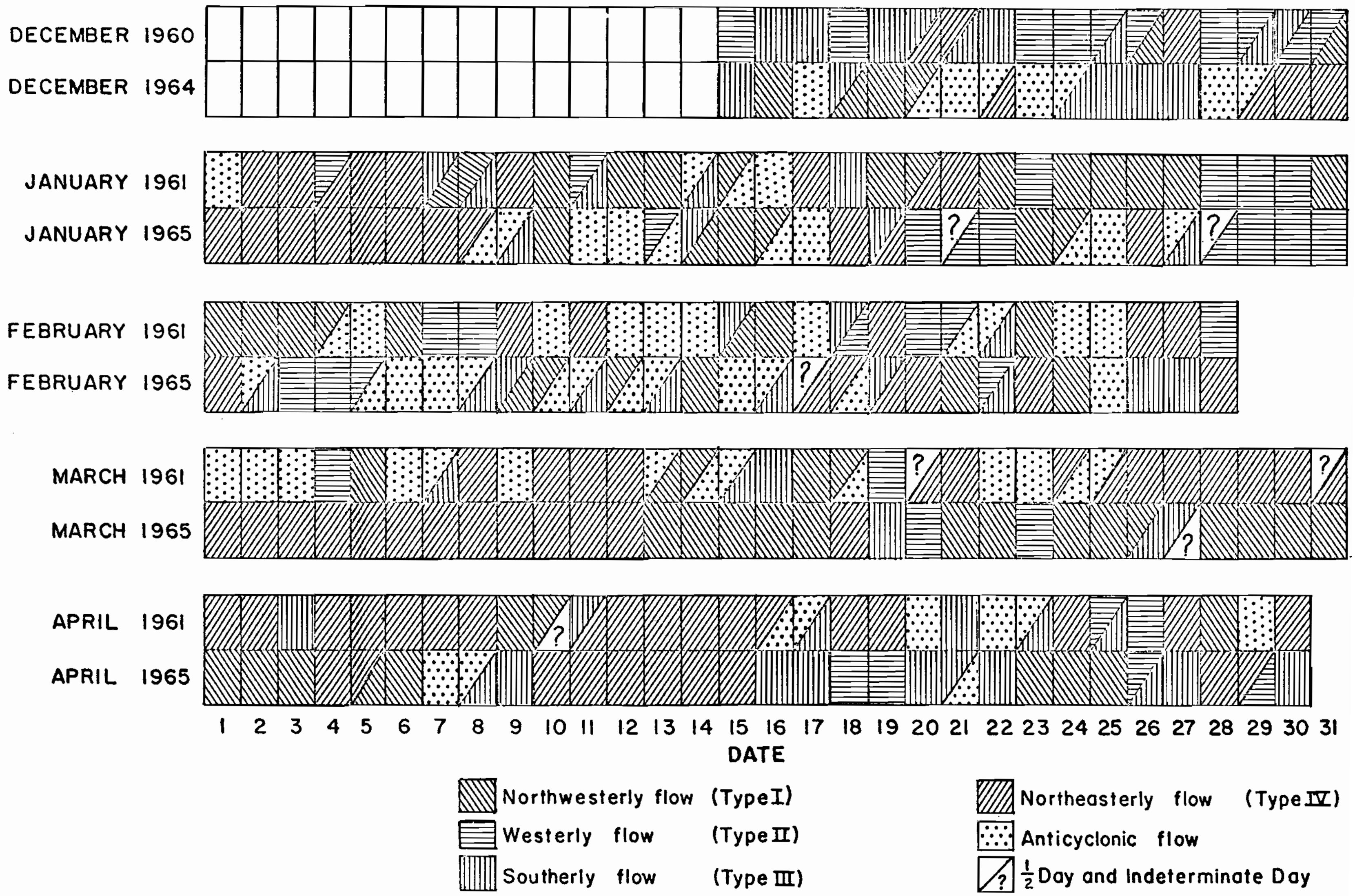

Fig. 3.1 Daily circulation types

during each winter. The results are indicated in Fig. 3.2. The relatively infrequent anticyclonic cases are obvious (one fifth as frequent as the cyclonic types). Both winters had a preponderance of airflow from the east, northeast and north, and were thus not quite like the long-term average flow from the northwest. This may be attributed to the fact that the 500 mb mean trough position was shifted towards the east. In fact, for a two month period beginning in mid February 1965, the Icelandic low was displaced from its normal position allowing a separate centre to exist east of Newfoundland.

One would thus expect the winter during which the heavier ice conditions prevailed to have been subjected to an excess of cold northwesterly (Type I) and westerly (Type II) flow, and a lesser amount of the warm southerly airflows (Type III). This assumption is supported, as Fig. 3.2 shows, for westerly and southerly flow (II and III). However, it is not supported for northwesterly (Type I) flow, for this cold air advection was lacking during the heavy ice winter (1960-61), but was abundant in the light ice year (1964-65).

By breaking down the flow types by the method just discussed, no definite conclusion can be drawn to assist in explaining the broad difference in ice conditions between the two winters. Consequently, a more refined analysis must be made in order to isolate the factors which control the ice conditions.

3.2 CHARACTERISTICS OF SYNOPTIC SCALE AIRFLOW

3.2.1 SEASONAL DISTRIBUTION OF AIRFLOW TYPES

An analysis as indicated in Fig. 3.3 will help in finding the

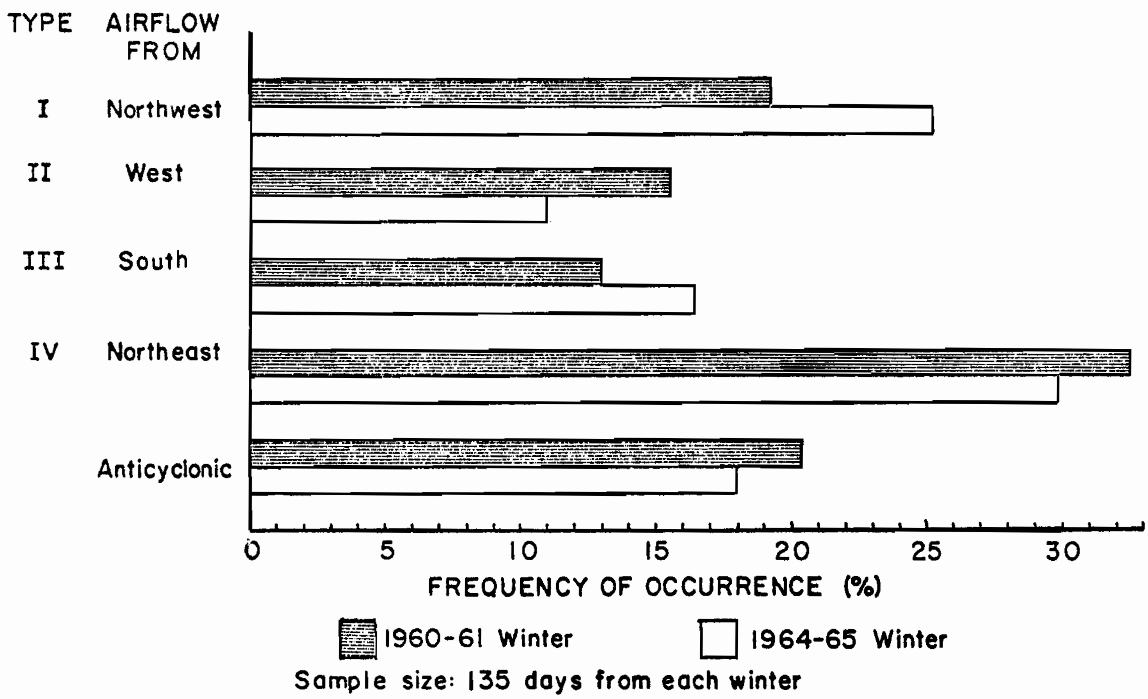

Fig. 3.2 Total occurrence of each flow type during each winter.

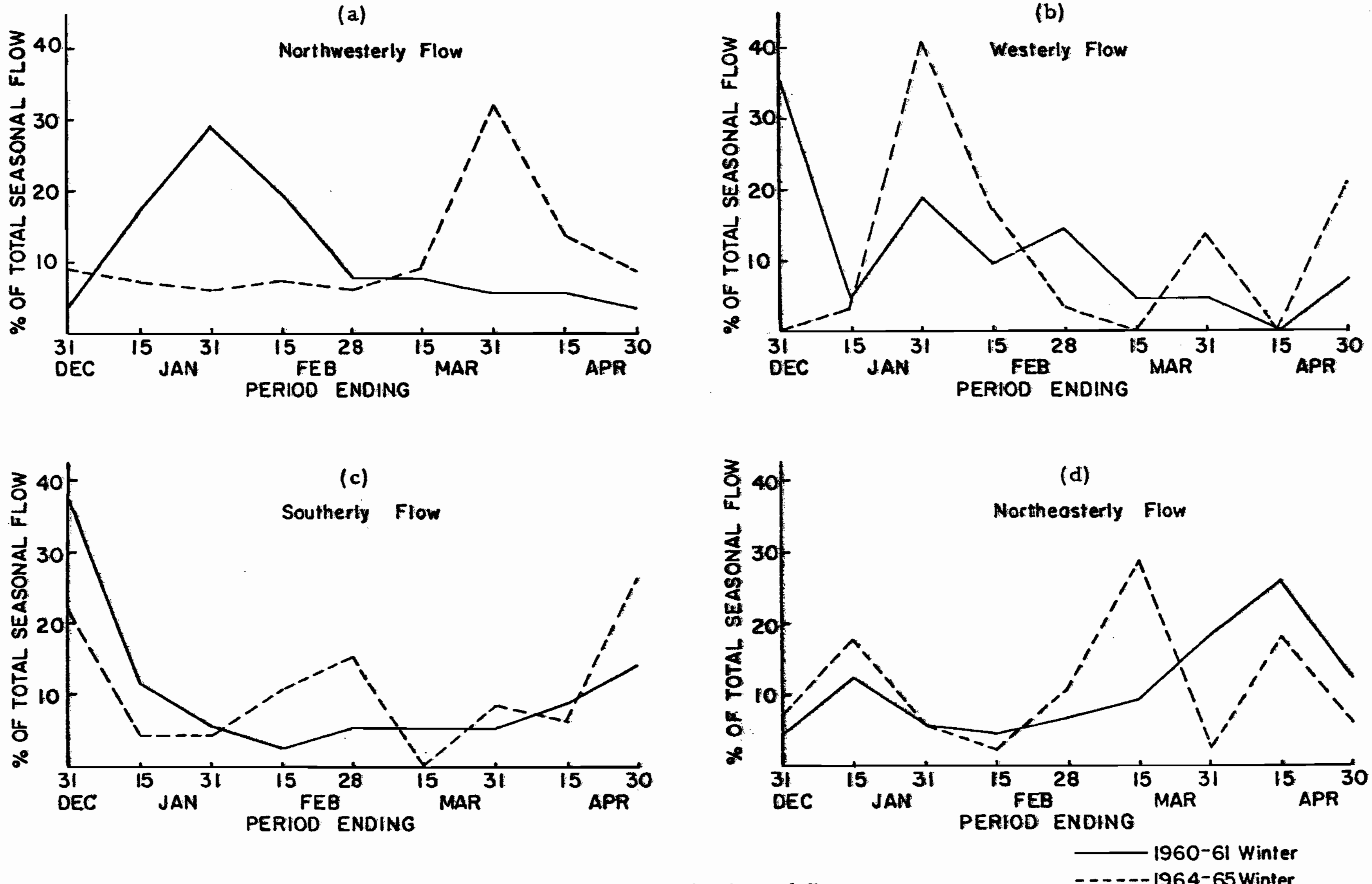

Fig. 3.3 Seasonal distribution of flow types.

reasons for the dissimilarity in ice conditions between the two winters. The four parts of this figure indicate the seasonal distribution (in two week intervals) of each of the cyclonic flow types, suggesting the times of the season during which the flow type was either dominant or weak.

Fig. 3.3(a) points out a clear disparity between the two winters. Although each year exhibited a peak period of northwesterly flow, it occurred in late January and early February during 1960-61, while it was concentrated towards the end of March in 1964-65. If other factors were equal, this would produce a tendency for rapid ice development early in the 1960-61 season. On the other hand, an increase in the amount of ice would be expected to take place towards the end of the 1964-65 season. These features were, in fact, reflected in the ice growth of both winters.

Fig. 3.3(b) shows how the rather cool Type II (westerly) airflow was apportioned over the two winters. Prominent features are the high frequency of westerly flow during late December 1960, and also the high frequency of this cool airflow late in January 1965. Considering these two facts, again with other factors equal, one would expect the high frequency of westerly flow so early in the winter of 1960-61 to have brought on an early onset in ice formation. The actual ice observations show this to have been true. Secondly, the frequent occurrence of westerly flow in late January 1965 should have been reflected in a substantial increase in ice cover during this period - this also is supported by the ice observations.

No firm conclusions from Fig. 3.3(b) can be extracted for the remaining months of the season, since the differences are too small for comparison.

Up to this point, this mode of circulation analysis has produced observations which have been related to the changes which took place in the ice conditions. However, the results of Fig. 3. 3(c) for the warm southerly airflow types (Type III), indicate a contradiction. The early weeks of the heavy ice season (1960-61) were subjected to more warm air advection than were the early weeks of the light ice year, thus the actual changes which took place in the ice conditions were opposed to those implied by Fig. 3. 3(c). This serves to point out that the method requires further refinement if an explanation is to be found. It will be shown in Section 7. 3 that the duration of the flow proves to be an important factor in the development of an ice cover.

The significant feature revealed by Fig. 3. 3(d) is the high frequency of northeasterly flow towards the end of the 1960-61 ice season, and also the two airflow peaks in 1965, one in mid March and the other in mid April. It will be seen from ice cover diagrams to be discussed in Section 7. 2 that a high occurrence of airflow from the northeast causes ice congestion in the southwest region of the Gulf and a prolongation of the ice season.

3. 2. 2 BIWEEKLY DISTRIBUTION OF AIRFLOW TYPES

Displayed in Fig. 3. 4 is the distribution of each of the four cyclonic types, and the one anticyclonic type, in fortnightly periods extending over the entire winter season. From this form of analysis it can be seen which flow type was dominant and which types were weaker during a two-week period.

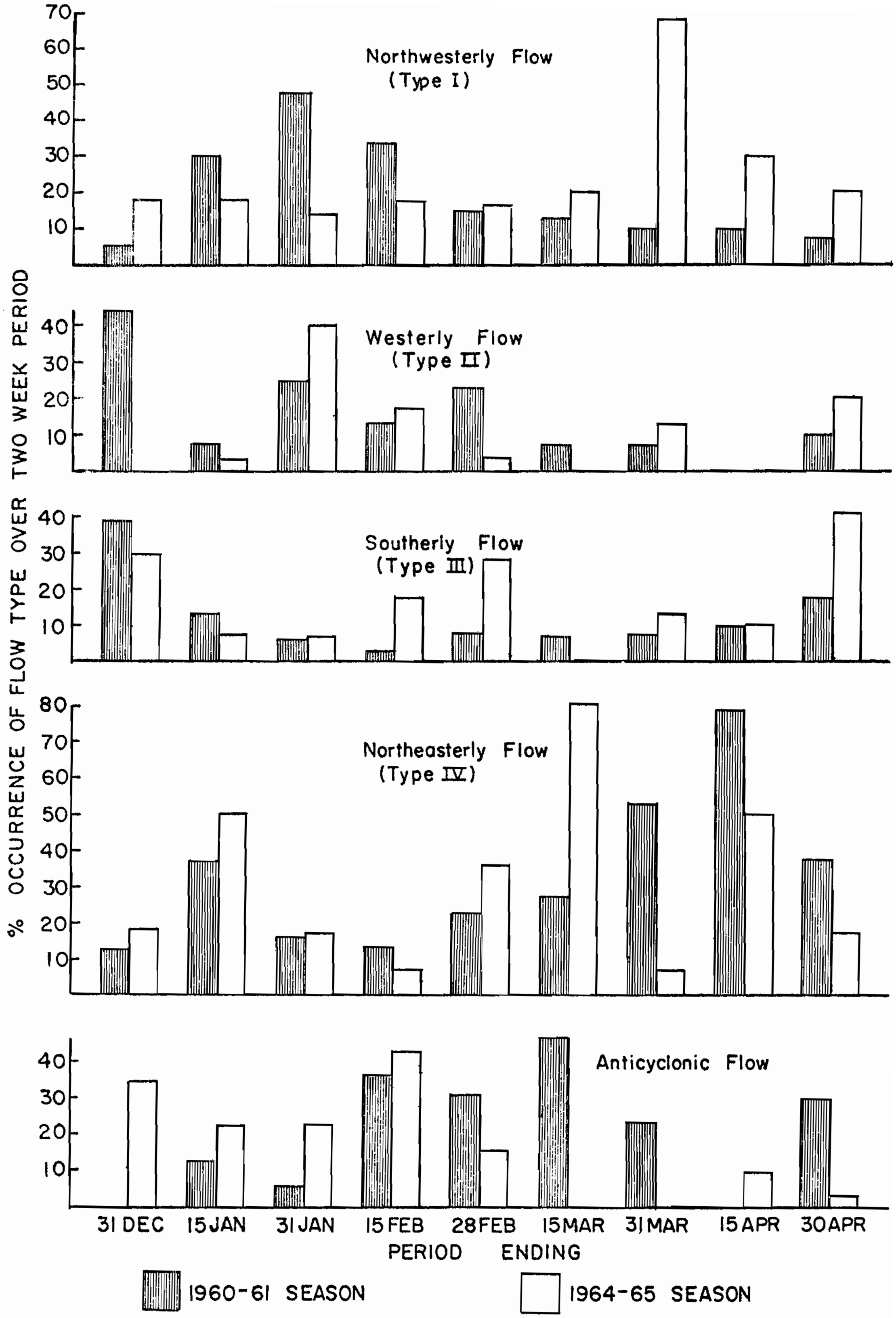

Fig. 3.4 Biweekly distribution of flow types.

In the conclusions (Section 7.3), the information provided in Fig. 3.4 will be combined with fortnightly ice cover data in order to reveal the influence which meteorological conditions have on the ice cover.

3.2.3 CHARACTERISTICS OF THE DURATION OF FLOW

A number of interesting observations may be derived from the type of analysis shown in Fig. 3.5. This diagram indicates the duration, in days, of each flow type case, and also the frequency with which each duration period was maintained.

First, it is noted that the southerly airflow cases (Type III) have the greatest tendency towards short duration. Also evident is the fact that this type has the highest frequency of airflow which lasted only one half a day. These observations arise as a result of the very situation in which southerly flow occurs, i.e., in the warm sector of frontal waves which travel rapidly through the Gulf region.

Secondly, it can be seen that westerly flow is also inclined to be of short duration (e.g., only one case lasted more than 3 days). This is a reflection of the wind pattern aloft, since westerly surface airflow occurs under zonal flow conditions at higher levels. Thus, it is implied that zonal flow aloft does not maintain itself for extended periods from December to April, but rather is very frequently interrupted by the passage of a wave superimposed upon the mean westerly flow.

Thirdly, in contrast to the westerly and southerly types

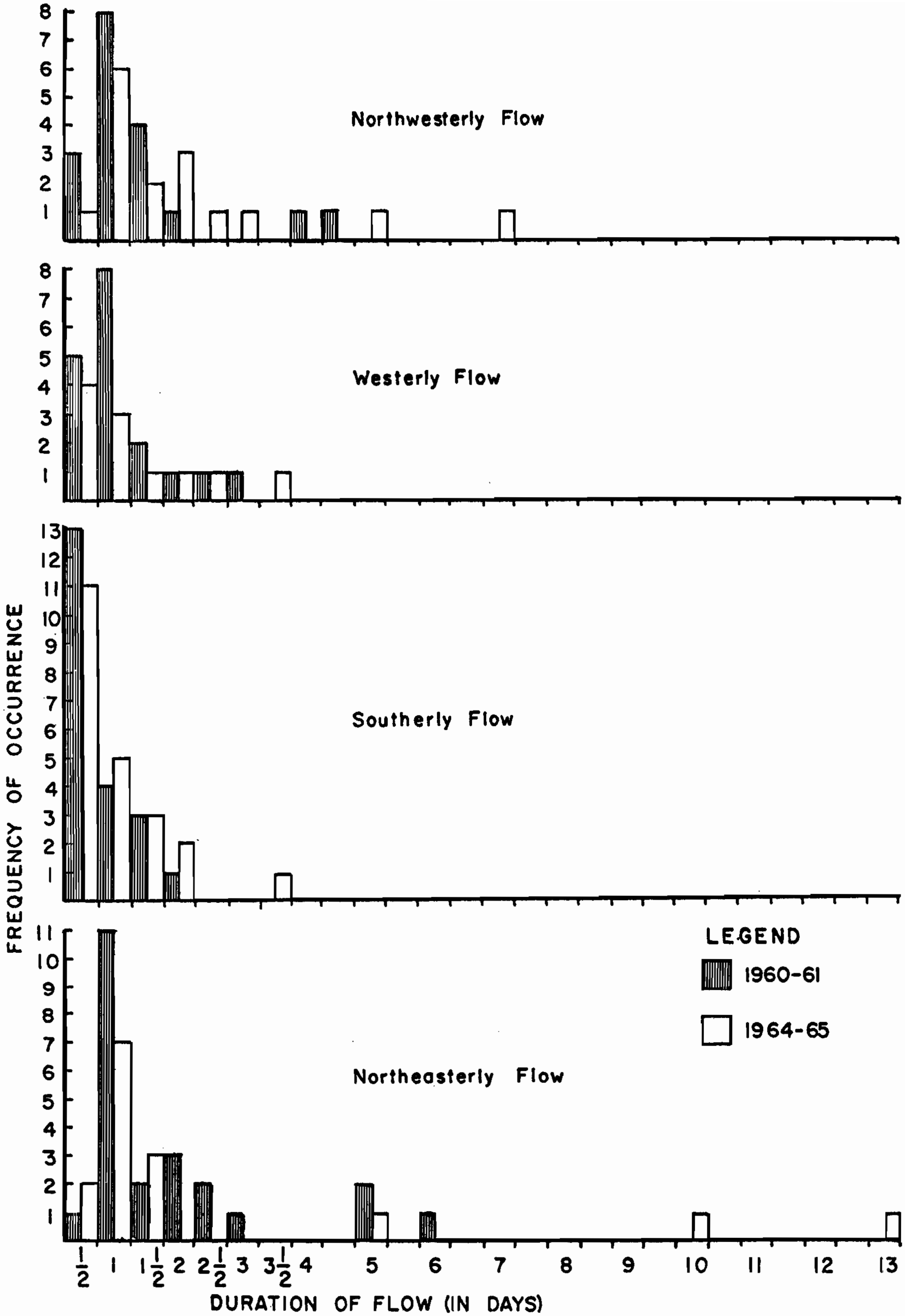

Fig. 3.5 Characteristics of duration of flow.

(Types II and III), Types I and IV show a distinct tendency towards extended duration. Type IV, in particular, shows two extreme cases of 10 and 13 days. This is a consequence of the similarity of these flow types to the normal, or mean surface pressure pattern, with the result that once the airflow from one of these directions - northeast to northwest - becomes established it tends to maintain itself over a long period of time.

CHAPTER IV

THE ENERGY BALANCE AND ITS TERMS

The foregoing chapter provided an analysis of the different types of synoptic airflow to which the Gulf was subjected during the 1960-61 and 1964-65 winter seasons. It now becomes necessary to perform a quantitative study of the actual fluxes associated with each flow type, in an attempt to uncover the physical effects of the different circulation modes.

4.1 HEAT BUDGET OF THE SEA SURFACE

From the principle of conservation of energy, one form of the energy balance equation at an evaporating surface is:

$$Q_r + Q_e + Q_h = Q_s \qquad 4.1$$

where Q_r is the net radiation (positive upwards),

Q_e is the transfer of latent heat of evaporation to the atmosphere (positive upwards),

Q_h is the transfer of sensible heat to the atmosphere (positive upwards),

Q_s is the heat conducted to the surface from below, including the heat stored in the water or ice, and the latent heat of fusion (positive when directed to the surface from below).

In a study of the total energy balance of the Gulf of St. Lawrence, the energy available from precipitation, ground water, and river flow would also require consideration.

In the following study, only the two turbulent exchange terms, Q_e and Q_h, will be examined in detail. A more detailed investigation would also require a study of the radiation and storage terms. There is an indication in Coombs' (1962) results that the sensible and evaporative fluxes make up a sufficiently large portion of the total balance to allow the heat flow trends to be observed from these two terms alone.

4.2 THE EVAPORATIVE TERM Q_e

The evaporation rate depends on the deficiency of water vapour in the air immediately above the water-surface, wind speed, and thermal convection.

The aim of many researchers has been to relate the heat fluxes to sea surface and air parameters which can be easily measured. The approach has been either via the classical "bulk aerodynamic method" or by the "profile method".

The bulk aerodynamic method requires knowledge of the coefficient K in equations of the Dalton form:

$$E = K(e_s - e_a)V \qquad 4.2$$

where E is the evaporation in mm day^{-1},

e_s, e_a are the vapour pressures at the surface and at the "anemometer height" (5-10 metres) in mb,

V is the wind speed in m sec^{-1}.

Sverdrup (1951) attempted to establish an empirical evaporation coefficient. He based his results on an area in the North Atlantic ("Meteor" observations) where accurate meteorological observations could be obtained, and where evaporation could be computed by energy balance techniques. He was able to show that K varied with the wind speed. From the graph of his results, a value of $K = 0.090$ is read for wind speeds less than 6.5 m sec^{-1}, while for speeds greater than 6.5 m sec^{-1}, Sverdrup accepted the value $K = 0.142$, a coefficient which Jacobs (1951) also had obtained by using climatological values of humidity, wind speed, and energy considerations.

These values were derived for a specific area in the North Atlantic, and there is no assurance that the same values would apply to a virtually land-locked water body such as the Gulf of St. Lawrence.

The methods of Sverdrup and Jacobs are not easily capable of refinement, since it is very difficult to evaluate all the terms of energy transfer with sufficient accuracy.

Laevastu (1960) considered the empirically derived evaporation equations of Penman, Rohwer, and two equations by Kohler (all of the Dalton type). Laevastu rejects all but Rohwer's formula. Rohwer (1931) had reviewed all the existing empirical formulae and arrived at an empirical expression which he based on a large amount of data from various conditions. Laevastu (1960) revised Rohwer's formula to allow the calculation of evaporation by using winds measured at the 8 metre level, obtaining:

$$E = (0.26 + 0.077 V_8)(0.98 e_s - e_8) \qquad 4.3$$

where E is the evaporation in mm day^{-1},

V_8 is the wind speed at the 8 metre level in m sec^{-1},

e_s, e_8 are the vapour pressures at the surface and at the 8 metre level in mb.

Laevastu opines that the extra term (0. 26) should be neglected when the vapour pressure gradient ($e_s - e_8$) is negative and when the wind speed is low.

The second approach to determining evaporation formulae is by profile observations. Deacon and Webb (1962) state "there is no doubt that the approach by profile measurements will prove fruitful".

Malkus (1962) began with the familiar heat transfer equation of the form:

$$F_p = -K_p \frac{dp}{dz} \qquad 4.4$$

where F_p is the flux of property,

$\frac{dp}{dz}$ is the vertical gradient of the property,

K_p is the "eddy transfer coefficient" for momentum, water vapour, and sensible heat fluxes.

From this form she derived equations relating each flux at the surface to the wind speed at anemometer height, and the difference between the property in question at that level and the sea surface. She assumed the equality of the three transfer coefficients K_m, K_e , and K_h of momentum, water vapour, and sensible heat respectively, and obtained the two transfer equations for evaporation and the sensible heat flux as follows:

$$E = \rho c_d (q_s - q_a) V \qquad 4.5$$

$$Q_h = \rho c_p c_d (T_s - T_a) V \qquad 4.6$$

where E is the rate of evaporation in cm day^{-1},

ρ is the density of air in gm cm^{-3},

c_d is the drag coefficient,

q_s, q_a are the specific humidities at the surface and at the anemometer height in gm kgm^{-1},

V is the wind speed in cm sec^{-1},

Q_h is the sensible heat flux in cal cm^{-2} sec^{-1},

c_p is the specific heat of dry air at constant pressure in cal gm^{-1} C deg^{-1},

T_s, T_a are the temperatures at the surface and in the air in deg C.

Since the primary interest lies in the heat loss caused by evaporation rather than the actual water loss, the two concepts can be related by:

$$Q_e = LE = \rho c_d L (q_s - q_a) V^{*} \qquad 4.7$$

---

*It is interesting to note that Bowen's expression for the ratio Q_h/Q_e (see equation 4.13), may be derived from equations 4.6 and 4.7, i.e.,

$$\frac{Q_h}{Q_e} = \frac{\rho c_p \; c_d (T_s - T_a) V}{\rho c_d \; L \cdot (q_s - q_a) V}$$

$$= \frac{c_p \rho (T_s - T_a)}{.622 L (e_s - e_a)}$$

where Q_e is the rate of heat loss due to evaporation in cal cm^{-2} day^{-1},

L is the latent heat of evaporation in cal gm^{-1}.

A brief discussion on the legitimacy of Malkus' assumption of equality of the transfer coefficients is warranted. Support is given to the assumption of equivalence of the transfer coefficients of momentum and water vapour ($K_m = K_e$) by the observations of Rider (1954) and by Deacon and Swinbank (1958) who, by using two separate techniques, arrived at the following results for the ratio K_e/K_m :

Rider	1.12 $\pm$ 0.04
Deacon and Swinbank	1.04 $\pm$ 0.09

Swinbank (1955), in a similar manner, made a comparison between K_h and K_m - the transfer coefficients for sensible heat and momentum transport. He found that K_h could exceed K_m by as much as 30% under lapse conditions, with the reverse applying under stable conditions.

In practice, therefore, the assumption of equality of K_e and K_m is not likely to lead to serious error in the calculation of evaporation. However, discrepancies can be expected in the calculation of sensible heat fluxes if the coefficients K_h and K_m are assumed equal.

The expression for the drag coefficient C_d in equation 4.5 has been taken from Deacon and Webb (1962), who collected the observations made by a number of different authors over water-surfaces, at neutral, or near neutral atmospheric conditions. Their results show a comparatively slow increase of drag coefficient with wind speed, with the simple linear relationship for near-neutral conditions being:

$$c_{10} = (1 + 0.07V_{10}) \times 10^{-3} \qquad 4.8$$

where V_{10} is the wind speed at the 10 metre level in m sec^{-1}.

Recalling the three evaporation formulae which have been introduced:

Sverdrup: $E = 0.090\,(e_s - e_a)V$ for $V < 6.5$ m sec^{-1} 4.9

$E = 0.142\,(e_s - e_a)V$ for $V \geq 6.5$ m sec^{-1} 4.10

Rohwer: $E = (0.26 + 0.077V)(0.98e_s - e_a)$ 4.11

Malkus: $E = .065\,(1 + 0.07V)\,(e_s - e_a)V$ 4.12

The vapour pressure gradient term is common and virtually equal in all three equations, thus a plot of the remaining function of the wind speed against a range of wind speeds will indicate the difference in the results produced by each equation.

It can be seen from Fig. 4.1 that although Sverdrup's empirical equation produces the largest values at wind speeds above 6.5 m sec^{-1}, the general shape does correspond to Malkus' theoretically derived equation up to speeds of 20 m sec^{-1}. On the other hand, Rohwer's equation differs greatly, and produces values within the range of Sverdrup and Malkus for wind speeds between 6.5 and 10 m sec^{-1} only.

Deacon and Webb (1962), obtained theoretical values of the evaporation coefficient at different wind speeds, through use of the drag and wind profile coefficients. The evaporation coefficients obtained

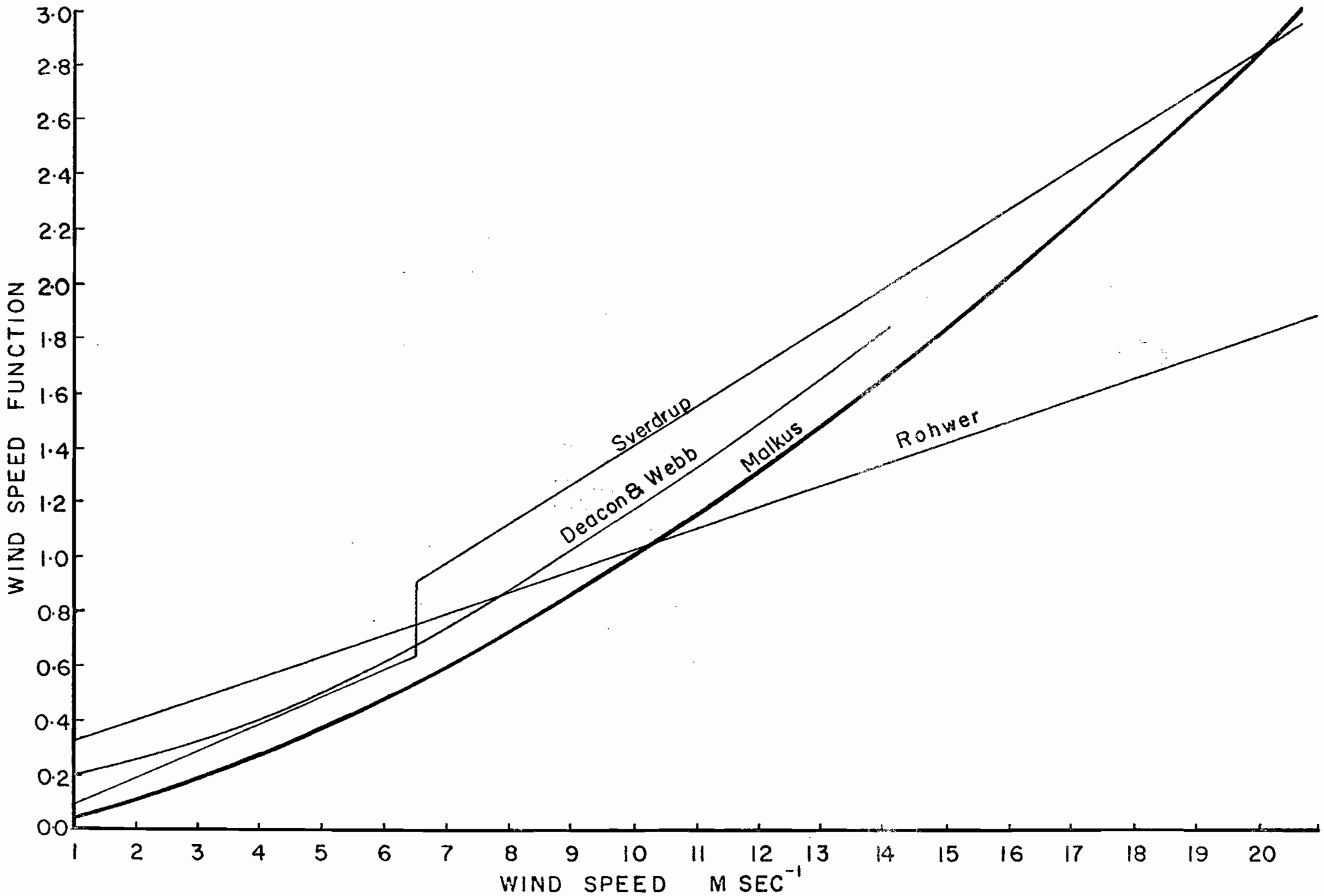

Fig. 4.1 Wind speed function versus wind speed.

from their table are also plotted in Fig. 4.1, and compared to the equations of Sverdrup, Malkus and Rohwer - the very close resemblance to Malkus' values is noted.

Deacon and Webb found that their results were comparable (within 5%) to the Lake Heffner observations of Marciano and Harbeck (1954). Also, at two other lakes, Harbeck et al (1958), and Webb (1960), produced results which compared to those of Deacon and Webb. Deacon and Webb then concluded that their value of the bulk coefficient was "applicable to water expanses of a wide range of sizes from the open sea down to lakes a mile or so across". Russian investigators Budyko, Berliand and Zubenok (1954) also indicate similar values.

Quoting Malkus (1962), "the exchange formulae are on a sufficiently sound footing at present to enable useful flux computations to be made from simply made and relatively plentiful observations". With this in mind, the empirical equations of Sverdrup and Rohwer have been neglected since the former gives larger values, and the latter gives much smaller values at high wind speed, than Malkus' theoretically derived equation.

The conclusion is to accept Malkus' equation on the grounds that it produces values which are virtually equal to the widely supported values proposed by Deacon and Webb, and because this form of the evaporative heat flux equation has been accepted by many different authors. Also, it is in a simple and useful form which allows calculation from readily available data from the area under consideration.

4.3 THE SENSIBLE HEAT TERM Q_h

Direct measurement of the convective transfer of sensible heat between a water-surface and the atmosphere presents more difficulties than the determination of evaporation. However, the sensible heat flux is an important factor in the winter energy balance of the Gulf.

An indirect method for calculating the sensible heat flux if the evaporative flux is known, is by use of the Bowen ratio - the ratio of the sensible heat flux to the latent heat flux. Bowen (1926) arrived at the following expression for the ratio:

$$B = \frac{Q_h}{Q_e} = 0.64 \frac{p}{1000} \frac{(T_s - T_a)}{(e_w - e_a)} \quad 4.13$$

where p is the atmospheric pressure in mb (p is very little different from 1000 mb, making the coefficient about equal to 0.64).

The general validity of the ratio for determining the sensible heat flux if the evaporative flux is known, has been the subject of much discussion, and caution must be exercised in its use. With regard to the use of the Bowen ratio for this purpose, Anderson (1954), states from the Lake Heffner investigation: "in exceptional conditions, for example, when evaporation rates are small and the differences in vapour pressure of the atmosphere and that of air saturated at the surface-water temperature approaches instrumental accuracy, the Bowen ratio is inadequate". Furthermore, the vapour pressure differences over snow and ice in particular are usually quite small.

Owing to the inadequacy of the Bowen ratio under the conditions experienced over the Gulf in winter, the decision has been to ignore the values of sensible heat calculated by the Bowen method, retaining them only for the purpose of comparison.

The expression given by Malkus (1962) for calculating the sensible heat flux was introduced in Section 4.2, viz:

$$Q_h = 2.55\,(1 + 0.07\,V)\,(T_s - T_a)\,V \qquad 4.14$$

where Q_h is the sensible heat flux in cal cm^{-2} day^{-1}.

Walmsley (1966), quoted values obtained by five authors for K' in the sensible heat flux equation of the form:

$$Q_h = K'\,(T_s - T_a)\,V \qquad 4.15$$

His table shows that the magnitude of K' ranges from 2.67 to 8.33 - which, in fact, are the values corresponding to the Shuleikin (1953) formulae. The corresponding limits for Malkus' equation are from 3.20 to 4.56, agreeing well with the values proposed by other authors.

Shuleikin's (1953) formula for computing the sensible heat flux neglects the wind speed if the water is warmer than the air:

$$Q_h = 30.24\,(T_s - T_a) \qquad 4.16$$

The large values obtained by this formula are to be expected since heating from below causes instability of the air, and as the turbulence increases, the vertical heat transport becomes large.

If the water is cooler than the air, the wind speed is considered:

$$Q_h = 0.432\,(T_s - T_a)\,V \qquad 4.17$$

The reasoning behind this equation, which gives small values, is that water colder than the air produces small heat transfer, since it stabilizes the air. If the vertical stability is sufficiently large, turbulence of the air and the corresponding heat flux ceases, Defant (1961).[3]

Due to the very small evaporation values over ice covered water, Shuleikin's equations, rather than the Bowen ratio technique for calculating the sensible heat flux, were chosen by Vowinckel and Taylor (1964) in their calculation of the Polar heat fluxes (their maximum daily mean evaporative heat flux was 110 cal cm^{-2} day^{-1}). Walmsley (1966) also used the same pair of equations in calculations for Baffin Bay (his maximum daily mean evaporative flux was 152 cal cm^{-2} day^{-1}. As will be seen in Chapter VI, the evaporative heat fluxes over the Gulf of St. Lawrence were often less than 160 cal cm^{-2} day^{-1}. Since evaporation values over the Gulf fall in a range in which other authors have concluded that the Bowen ratio technique was inaccurate, this method has therefore been deleted. The use of the Shuleikin formulae is prescribed for this study. Also, it is important to note that the Shuleikin formulae were originally derived over a partially ice covered water surface.

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[3] Note that Malkus' equation (4.14) would produce a large downward flux.

CHAPTER V

DATA PREPARATION

In the previous chapter, the decision was made to use Malkus' equation to calculate evaporation:

$$E = .065\,(1 + 0.07V)\,(e_s - e_a)V \qquad 5.1$$

and Shuleikin's formulae to obtain the sensible heat flux:

$$Q_h = 30.24(T_s - T_a) \qquad T_s \geq T_a \qquad 5.2$$

$$Q_h = 0.432(T_s - T_a)V \qquad T_s < T_a \qquad 5.3$$

The present chapter deals with the data preparation, and the assumptions made in calculating the evaporative and sensible heat fluxes.

5.1 VAPOUR PRESSURE

The saturated vapour pressure at the surface, and the vapour pressure at the anemometer height, may be calculated from the integrated form of the Clausius-Clapeyron equation:

$$e = 6.11 \exp\left[\frac{m_v L}{R^*}\left(\frac{1}{273} - \frac{1}{T}\right)\right] \qquad 5.4$$

where m_v is the molecular weight of water vapour in gm mole^{-1},

R^* is the universal gas constant in cal mole^{-1} deg K^{-1},

and (a) at anemometer height:

$e = e_a$ is the vapour pressure of the air in mb,

$L = L_v$ is the latent heat of vaporization in cal gm^{-1},

$T = T_d$ is the dew point temperature of the air in deg K,

(b) at an ice surface:

$e = e_i$ is the saturation vapour pressure at the ice surface,

$L = L_s$ is the latent heat of sublimation,

$T = T_i$ is the surface temperature of the ice,

(c) at a water surface (the air in contact with water is assumed to be saturated at the water surface temperature):

$e = e_w$ is the saturated vapour pressure at the surface,[4]

$L = L_v$ is the latent heat of vaporization,

$T = T_w$ is the water surface temperature.

5.2 AIR AND DEW POINT TEMPERATURES AND WIND SPEEDS

The air temperature, dew point temperature, and wind speed at anemometer height are also required. These data were obtained on punched cards for the periods from December 1, 1960 to April 30, 1961, and December 1, 1964 to April 30, 1965, from the Meteorological Branch for fifteen stations around the Gulf of St. Lawrence (see location map Fig. 1.1). Daily mean values of each of the three parameters were calculated at all stations from the 00Z, 06Z, 12Z, and 18Z observations, except for Port Menier and Daniels Harbour where the 12Z, 18Z, and

[4]The air in direct contact with the sea surface has a water vapour pressure equal to that of sea-water at the sea-surface temperature. At normal salinity (35‰), this is approximately 98% of the vapour pressure of pure water at the same temperature (Deacon and Webb, 1962).

00Z observations were used, and Miscou Island where the 12Z, 15Z, 18Z, and 21Z reports were averaged.

These data were plotted and analysed to obtain grid point values from a 28-node grid covering the water-surface area of the Gulf. The grid, of dimensions 1 degree latitude by 1 degree longitude (111 km by 75 km), is indicated in Fig. 1.1.

The extrapolation of land station data out over a water surface is a dubious practice. In an attempt to derive a relationship between land observations and conditions over the water, twenty-five days of oceanographic data were obtained from the Canadian Oceanographic Data Centre for short periods during January, February, and April 1960; January 1961; and February and April 1962. Daily mean temperatures for land stations surrounding the Gulf were derived by averaging the daily maximum and minimum temperatures extracted from the Monthly Record.

The twenty-five-day mean temperature derived from the ship observations was found to be 1.5 centigrade degrees higher than the corresponding mean of all the land stations. The day-to-day variation of this temperature difference was very large, and consequently it could not be applied to daily calculations.

A test was made using Rodgers and Anderson's (1961) equation which relates land station temperatures and water-surface temperature to give the air temperature over the open water. The equation was found to be inapplicable over the Gulf due to the presence of ice.

It is fortunate that complete data coverage is maintained at

Grindstone Island, a station located in the centre of the Gulf in rolling, nearly treeless country, and at an elevation of 192 feet. With observations available from such a prime location, a study was made to determine the relationship between the Island meteorological observations and concurrent ship reports over the Gulf. Oceanographic data were obtained for the January 14-19, 1961 cruise - the only sailing for which land data had been obtained.

The ship observations were all taken at a distance from Grindstone Island of between 25 and 60 miles. The only observations used to determine the relationship were selected such that the time delay between the land observation and ship report was never greater than one hour. Unfortunately, the Beaufort wind scale was used in the records, thus only the range of wind speed was available. A comparison of the air temperature and wind speed parameters is provided in Table 5.1.

From this small sample, it can be seen that concurrent temperature and wind observations differ greatly from ship to land, and it is not until mean values over 24 hour, or longer, periods are taken that the ship and land observations become similar. In particular, for the two days January 17 and 18, there is only a 4% temperature difference between the open water and land observations.

It can also be seen that the wind speed values registered on Grindstone Island are very close to the estimated wind speeds observed over the Gulf. Again, averages must be taken.

Humidity data were not recorded on this cruise. However,

TABLE 5.1

Concurrent Observations of Meteorological Parameters
at Grindstone Island and over the Gulf

C. O. D. C. Cruise No.381, January 1961

Date	Time Grind. Obs.	(LST) Ship Obs.	Air Temp (^{o}F)		Wind Speed (mph)		Daily Means			
			Grindstone	Ship Data	Grindstone Island	Ship (range)	Grind. Air T	Ship Air T	Grind. Wind	Ship Wind
14	17	18	35	20	24	19-24	35	20	24	19-24
15	11	11	17	14	18	19-24	17	9.5	18	16-21
15	14	14	17	5	18	13-18				
16	17	16	23	21	10	4-7	25.5	25.5	14	12-16
16	23	22	28	30	18	19-24				
17	2	2	29	19	24	32-38				
17	11	10	28	33	16	19-24				
17	17	18	28	20	14	4-7	28.0	27.2	17.2	13-18
17	20	20	27	34	14	4-7				
17	23	22	28	30	18	8-12				
18	2	2	27	28	16	19-24				
18	5	4	26	19	20	13-18				
18	11	11	30	20	16	13-18	24.0	23.0	20	21-26
18	14	13	30	33	16	8-12				
18	20	20	19	18	22	32-38				
18	23	22	12	20	30	39-46				

Coombs (1962) concluded from November, 1961 and February, 1962 data taken in the Gulf of St. Lawrence, that there was very little difference (2%) between the relative humidity measured at sea and on land.

The conclusion is to assume the Grindstone Island observations to be representative of the central portion of the Gulf.

A straightforward temperature and dew point analysis was performed, without incorporating correction factors but taking into account the airflow direction and ice distribution, in order to locate the regions of strong temperature and dew point gradients.

In carrying out the wind speed analysis, consideration was made of the fact that, under unstable conditions, the wind speed increases over the water relative to its land value, but decreases under stable conditions (Richards, Dragert and McIntyre, 1966).

5.3 SURFACE TEMPERATURES

Surface-water temperatures for extended time periods are generally not available for the Gulf of St. Lawrence, since winter oceanographic surveys are infrequent and of short duration, and also because the winter season is one of low shipping density.

For the 1964-65 season, surface temperatures extracted from merchant ship synoptic reports were kindly provided by Mr. W.B. Bailey of the Oceanographic Office, Halifax.

The water temperature was assumed to be at the freezing point for grid points at which ice concentration was greater than 5 tenths (Walmsley 1966). For water of the Gulf's salinity (31 to 32‰), the freezing temperature is -1.7°C (Defant 1961).

All available data were plotted and analysed, using as a guide the maps of average distribution of water-surface temperature for November and May compiled by Lauzier, Trites, and Hatchey (1957).

Since the variation in surface temperature is not large during the ice season (from the freezing temperature to 2°C by late April), any discrepancy in the surface temperature would not produce a serious error.

5.4 SNOW AND ICE SURFACE TEMPERATURES

Williams (1961), after measuring the evaporation from a snow cover, concluded that the average snow-surface temperature was equal to the mean air temperature.

This concept is also supported by data obtained by Pounder and Little (1959) during the winter 1956-57 at Shippegan, New Brunswick, a station on the shore of the Gulf of St. Lawrence. The snow-surface temperature was obtained by inserting thermometers 1 inch beneath the snow-surface. The following is an abbreviated form of their table:

Table 5.2

Relationship between air and surface temperatures

Date, Time	Feb. 25, 1520	Feb. 25, 1805	Feb. 26,1150
Air Temp. (°C)	-10.0	-10.5	-7.5
Wind velocity (mph)	0	E15	E10
Temp. at snow-surface (°C)	- 8.9	-10.6	-7.7

From the last two observations, it appears that the mechanical turbulence caused by the relatively high wind speed was enough to break down the vertical temperature gradient, while the observation for which there was no wind (the area was dominated by an anticyclone), a vertical temperature gradient existed. Moreover, of the 82 days being studied in this report, only four had daily mean wind speeds less than 10 m.p.h. It appears appropriate, therefore, to assume equality of the air temperature and the snow-surface temperature, with the exception that when the air temperature is above the melting temperature of snow or ice, the ice-water puddle temperature is assumed to be $0^{\circ}C$.

5.5 ICE CONCENTRATION

Ice conditions in the Gulf for the winter 1960-61 were obtained from Beaton (1962), while conditions for the winter 1964-65 were obtained from the Aerial Ice Observing and Reconnaissance maps, and from the Ice Summary and Analysis, 1965.

Ice concentrations at the 28 grid points were read from these maps. At grid points for which ice conditions were not observed, the concentration was interpolated between days for which data were available.

5.6 ACTUAL HEAT FLUXES

It was noted in Chapter IV that the evaporative heat flux is related to the evaporation by:

$$Q_e = 0.1 \rho L E \tag{5.5}$$

where ρ is the density of water in gm cm^{-3}, and (a) over water:

$$Q_e = Q_{ew}$$
$$L = L_v$$

(b) over ice:

$$Q_e = Q_{ei}$$
$$L = L_s$$

For calculation of the sensible heat flux, the Shuleikin formulae are used:

$$Q_h = 30.24\,(T_s - T_a) \qquad T_s \geq T_a \qquad 4.16$$
$$Q_h = 0.432\,(T_s - T_a)V \qquad T_s < T_a \qquad 4.17$$

where (a) over water:

$$Q_h = Q_{hw}$$
$$T_s = T_w$$

(b) over ice:

$$Q_h = Q_{hi}$$
$$T_s = T_i$$

Having calculated the evaporative and sensible heat fluxes over open water and over an ice surface, it remains to weight each of these fluxes according to the concentration of ice at each grid point.

Thus for the evaporative heat flux:

$$Q_e = Q_{ew}\,(10 - CONC) + Q_{ei}\,(CONC)$$

where CONC is the ice concentration in tenths, while for the sensible heat flux:

$$Q_h = Q_{hw}(10 - CONC) + Q_{hi}(CONC)$$

CHAPTER VI

THE TURBULENT HEAT FLUX TERMS

With quantitative observations of the turbulent exchange terms available for the Gulf of St. Lawrence, the relative magnitudes and typical heat flux patterns associated with each flow type will be discussed. Subsequently, the calculated heat fluxes will be related to the changes which occurred in the ice cover.

Rather than computing fluxes for each day during the two seasons studied, it was decided to restrict attention to days which had been subjected to the same flow type for at least three consecutive days. A discussion of the factors considered in arriving at this judgement will be presented in Section 6. 5. This criterion made it possible to include at least one case of each flow type. Sixteen such cases existed, nine during the 1960-61 winter, and seven during the 1964-65 season.

6. 1 DISCUSSION OF THE TURBULENT FLUX TERMS

A comparison of the heat flux magnitudes associated with each flow type was achieved as follows:

Evaporative and sensible heat fluxes were calculated at each grid point from grid values of temperature, wind speed, and ice concentration which had been averaged over the duration period of each case. An area weighted mean flux, which would represent the average flux for the entire Gulf, was calculated by weighting the fluxes at each grid point by the area represented by the point.

These fluxes and their contributory parameters have been entered in Table 6.1, and the evaporative and sensible heat fluxes have been plotted in Fig. 6.1.

Values of the evaporative and sensible heat losses computed over Bering Sea (Batalin, 1960), and over Baffin Bay (Walmsley, 1966), for the period January to April, range from 50 to 180 cal cm^{-2} day^{-1}. Jacobs' (1951a) northern hemisphere maps of mean wintertime evaporative and sensible heat fluxes indicate flux magnitudes in the Gulf of St. Lawrence region of about 100 cal cm^{-2} day^{-1} and 50 cal cm^{-2} day^{-1} respectively. Since the magnitudes of the Gulf of St. Lawrence fluxes fall within these limits, it is reasonable to accept the values which have been computed in this paper. Coombs (1962) used the Bowen ratio technique in his calculations. The failure of this method for determining the sensible heat flux from the evaporative flux has already been indicated, and in consequence no comparison to his seemingly high estimates can be made.

It can be seen in Fig. 6.1 that, of the days being studied, the winter season transport of heat by eddy fluxes is, on the average, directed away from the surface. It is also noted that the evaporative heat flux is very little different from that of sensible heat.

The fluxes of identical flow type cases in Table 6.1 show that the values of individual cases vary over a moderate range. Such direct comparisons are not permissible, however, because they are being made between periods of radically different ice conditions, wind speed and air-to-water temperature gradients.

TABLE 6.1

CHARACTERISTIC HEAT FLUXES OF THE DIFFERENT FLOW TYPES

Date	Type	Flow	$\overline{Q_e}$ Ly/day	$\overline{Q_h}$ Ly/day	$\overline{\lvert Q_e + Q_h \rvert}$ Ly/day	$\overline{\text{CONC}}$ tenths	$\overline{T_a}$ °F	$\overline{T_d}$ °F	$\overline{V}$ MPH	$\overline{(T_w - T_a)}$ °F	$\overline{Q_e}$ Ly/day	$\overline{Q_h}$ Ly/day
J24-J27, 61	I	NW	62	130	192	7	1.8	-3.7	18	27.1	84	75
J31-F3, 61	I	NW	28	61	89	9	2.7	-2.9	15	26.0		
M13-M17, 65	I	NW	105	56	161	7	19.8	13.4	24	9.5		
M28-A3, 65	I	NW	141	53	194	5	24.4	16.9	24	5.4		
J28-J30, 61	II	W	44	72	116	8	7.0	1.5	17	21.9	64	55
J29-J31, 65	II	W	84	103	187	7	10.6	5.3	20	18.9		
D25-D27, 64	III	S	-95	-14	-109	0	39.5	38.2	24	-5.4	-95	-14
M10-M12, 61	IV	NE/N	30	9	39	9	24.5	21.8	23	4.4	29*	2*
M26-A2, 61	IV	NE/N	13	0	13	5	31.5	29.4	17	-1.9		
A4 -A8, 61	IV	NE/N	28	-1	27	5	32.0	28.6	16	-2.1		
A12-A15, 61	IV	E/NE	26	3	29	4	32.4	30.0	21	-1.9		
D30, 64-J7, 65	IV	N/NE	160	128	288	0	25.5	22.8	23	7.8		
F28-M12, 65	IV	E/NE	13	-2	11	5	31.9	29.3	18	-2.2		
A11-A15, 65	IV	NE	66	5	71	2	31.1	26.0	22	-0.2		
F12-F14, 61		A	22	44	66	9	9.7	4.3	12	19.3	19	32
M1-M3, 61		A	16	21	37	9	15.6	11.1	12	13.3		

* Excluding case from December 30, 1964 to January 7, 1965.
Bar indicates area weighted mean.

1 Ly = 1 cal cm^{-2}

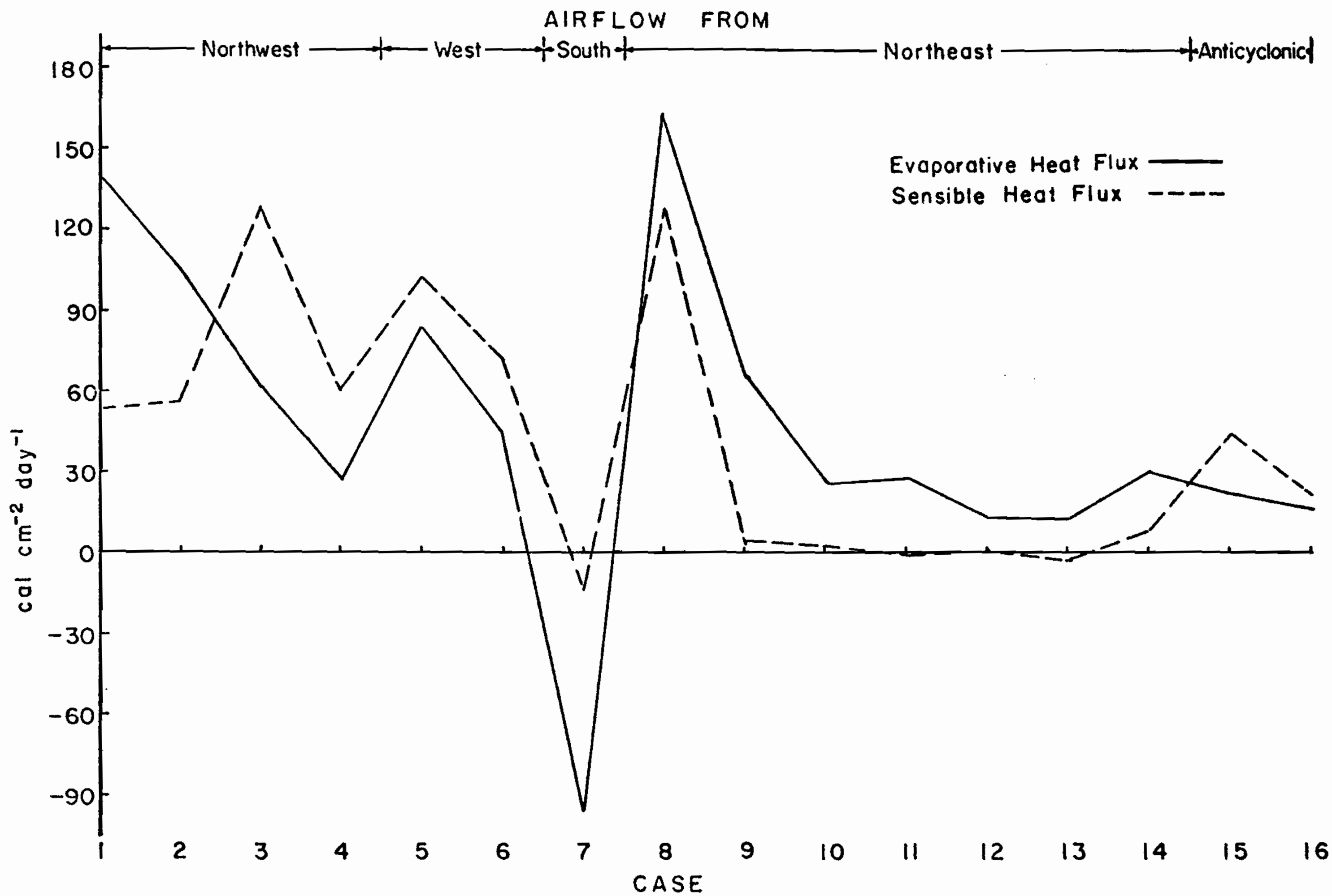

Fig. 6.1 Turbulent heat fluxes of the different flow types

None the less, it can be seen from the table that the highest upward fluxes are caused by northwesterly and westerly airflow - as a result of the intense temperature gradient at the surface. Correspondingly steep gradients also exist during anticyclonic flow. However, wind speeds are commonly low, thus decreasing the rate of heat transport.

In sharp contrast to the two flow types mentioned previously, periods of southerly flow are characterized by an exchange of heat from the atmosphere to the surface. The energy exchange process takes place through the condensation of water vapour and the release of its latent heat of evaporation. With such an ample supply of heat[5], it is possible to conceive the quantity of ice which could be melted.

Compared to the northwesterly and westerly types, airflow from the north, northeast or east is typified by significantly lower heat losses. The low and often negligible sensible heat losses are a product of the very small difference between the temperatures of the lowest layers of an air mass which has had a trajectory over the Atlantic, and the water-surface temperature of the Gulf. This type of air mass also tends to be saturated in its lowest levels, thus reducing the evaporative heat flux.

One marked exception, however, was the December 30, 1964 to January 7, 1965 case of northerly and northeasterly flow, which resulted in a very high heat loss. This is explained by the complete lack of ice in the Gulf at the time, combined with a wide difference

[5] Almost 600 cal gm^{-1} are released in the formation of water fog, increased, on rare occasions, by 80 cal gm^{-1} if sublimation takes place.

between the air dew point and water-surface temperatures.

Although no pre-freeze-up cases of long duration of northwesterly or westerly flow were studied, from the magnitude of the heat losses which occurred under northeasterly flow one gains a feeling for the much larger heat loss which would occur under colder and drier air advection from the west or northwest. Indeed, relative to the other airflow directions, even a very short duration of either northwesterly or westerly flow would result in a considerable liberation of heat.

6.2 CHARACTERISTIC HEAT FLUX PATTERNS OF THE DIFFERENT FLOW TYPES

6.2.1 NORTHWESTERLY FLOW, TYPE I

The analysis of the evaporative and sensible heat fluxes (Figs. 6.2 and 6.3) for the five days from March 13 to 17, 1965 show relative maxima to the south of Anticosti Island and over a region off the west coast of Newfoundland. An investigation of the parameters which contributed to these areas of high energy exchange revealed that the flux pattern bore a marked resemblance to the ice concentration (Fig. 6.4) rather than to the temperature or wind parameters. One would expect to observe large upward fluxes along the north shore of the Gulf for cases of advection of cold, dry air over the water. However, the insulating effect of the ice, which was of greater than 6 tenths concentration, decreased the amount of heat released.

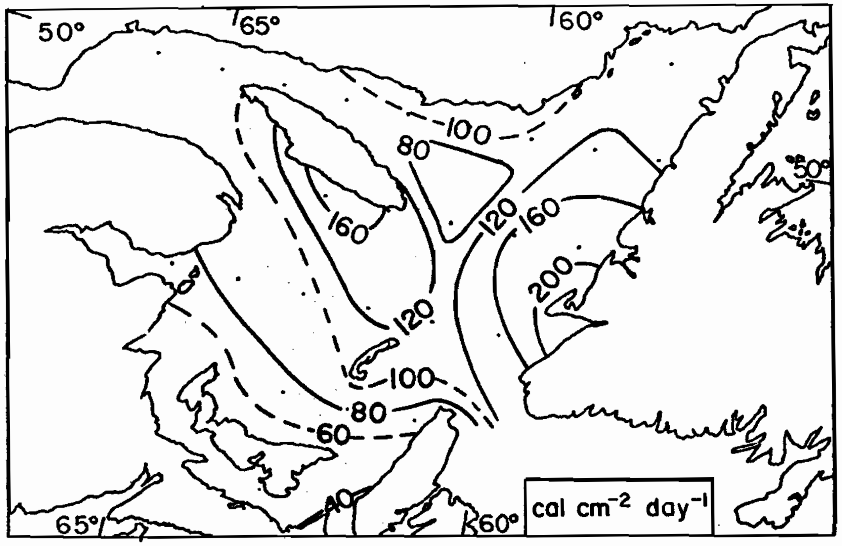

Fig. 6.2 Northwesterly flow evaporative heat flux

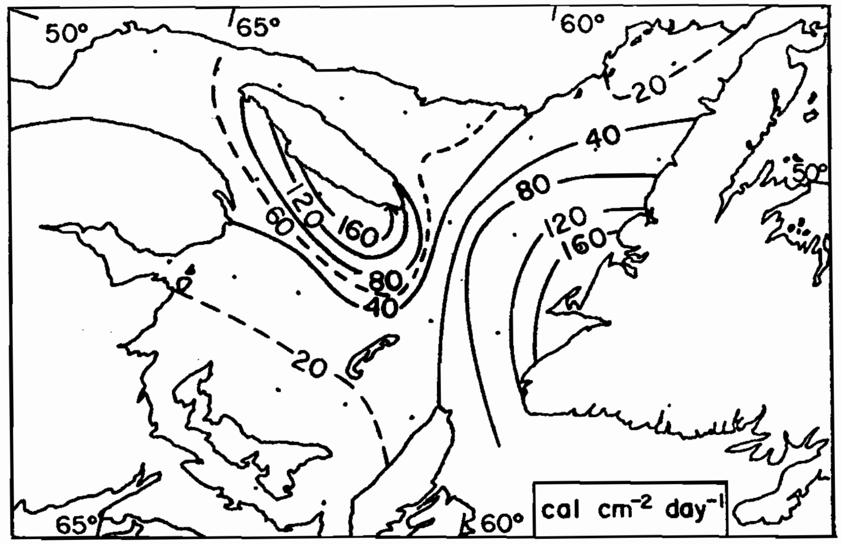

Fig 6.3 Northwesterly flow sensible heat flux

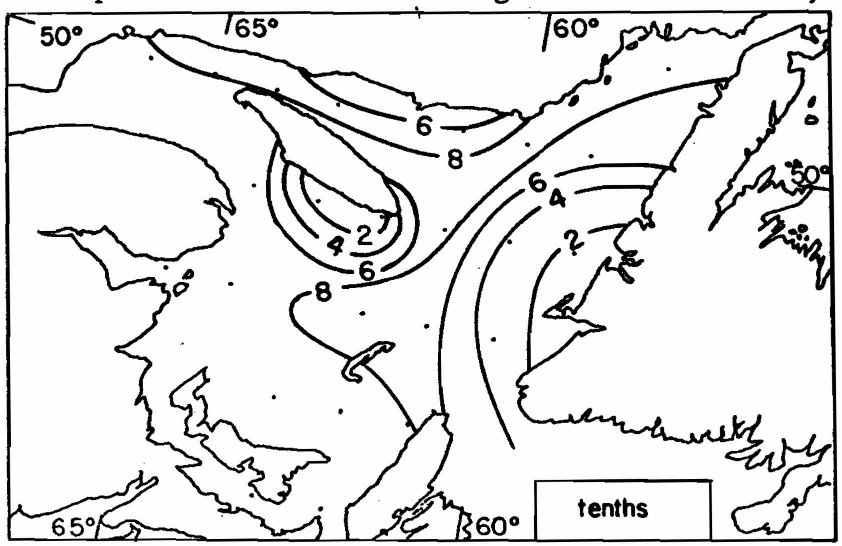

Fig. 6.4 Ice concentration

6.2.2 WESTERLY FLOW, TYPE II

High values of upward flux of evaporative and sensible heat arose during the three days from January 29 to 31, 1965 (Figs. 6.5 and 6.6). Once again the ice concentration (Fig. 6.7) was reflected strongly in both the heat flux patterns.

The high evaporation rate in the northeast arm of the Gulf was due to the high wind speed (Fig. 6.8). On the other hand, the relative magnitude of the sensible heat flux did not increase over the same area because the wind speed does not enter the calculation when the water temperature is higher than the air temperature (equation 4.16).

Figures 6.5 and 6.6 also indicate clearly how little the air is modified as it moves across ice covered water. Once the air reaches a region of lower ice concentration, however, and the blanketing effect of the ice is removed, the air mass modifies rapidly in consequence of the release of large quantities of heat from the surface-water layer.

6.2.3 SOUTHERLY FLOW, TYPE III

Although it rarely occurs for extended periods, a situation of southerly flow continued for the three days December 25 to 27, 1964. The ice concentration factor was absent in all calculations since ice had not begun to form in the Gulf.

An examination of the evaporative and sensible heat fluxes for this period, (Figs. 6.9 and 6.10) reveals relative maxima of downward fluxes over the Magdalen Shallows and along the west coast of Newfoundland. These maxima were due to the advection of relatively

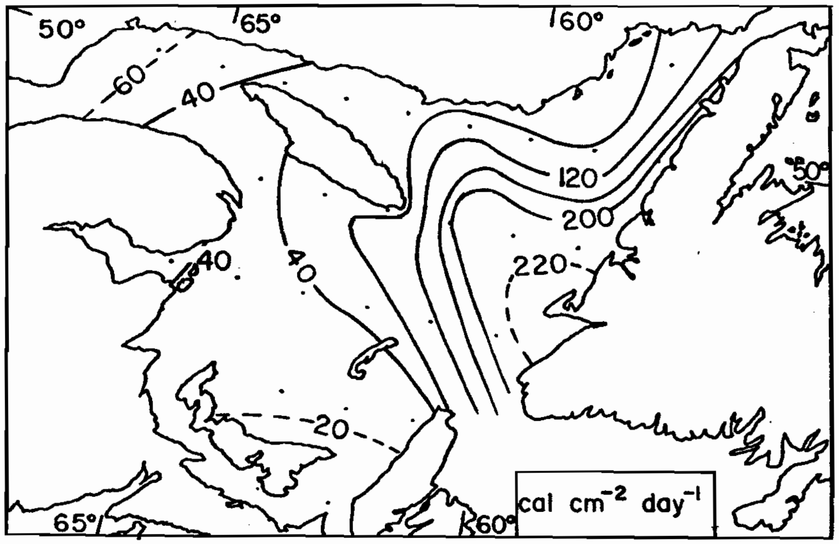

Fig. 6.5 Westerly flow evaporative heat flux

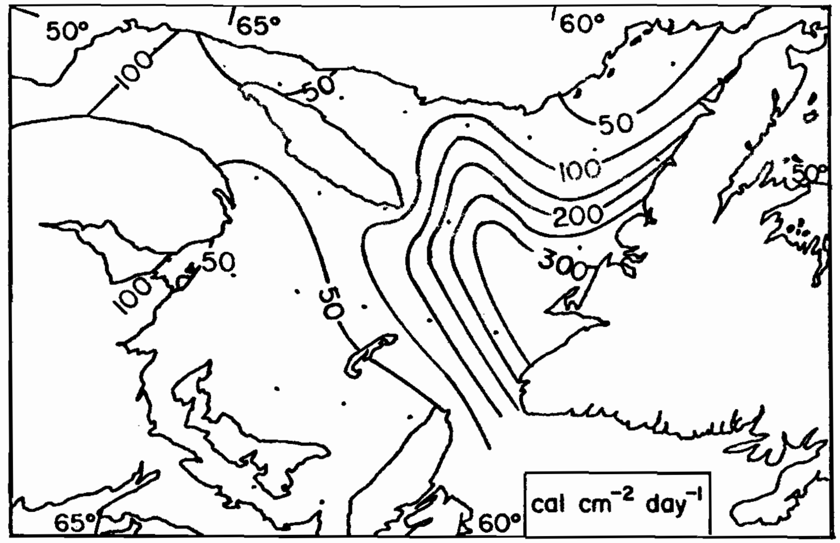

Fig. 6.6 Westerly flow sensible heat flux

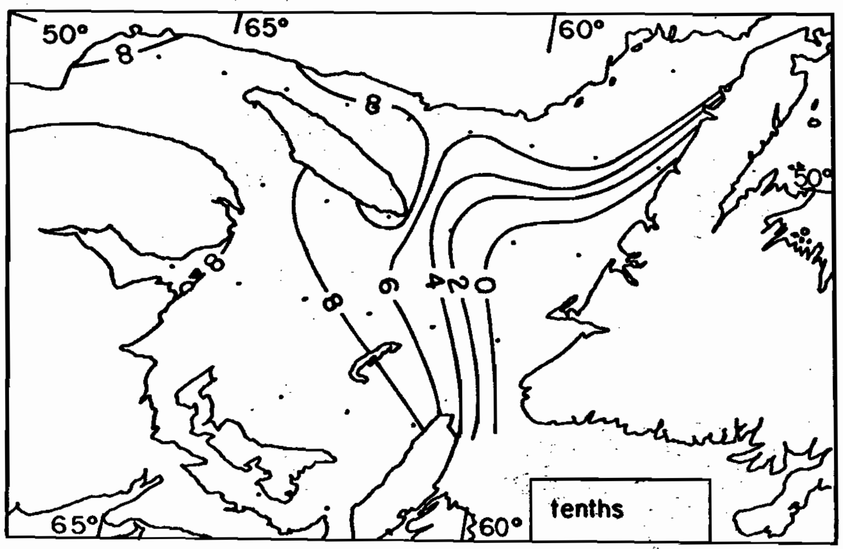

Fig. 6.7 Ice concentration

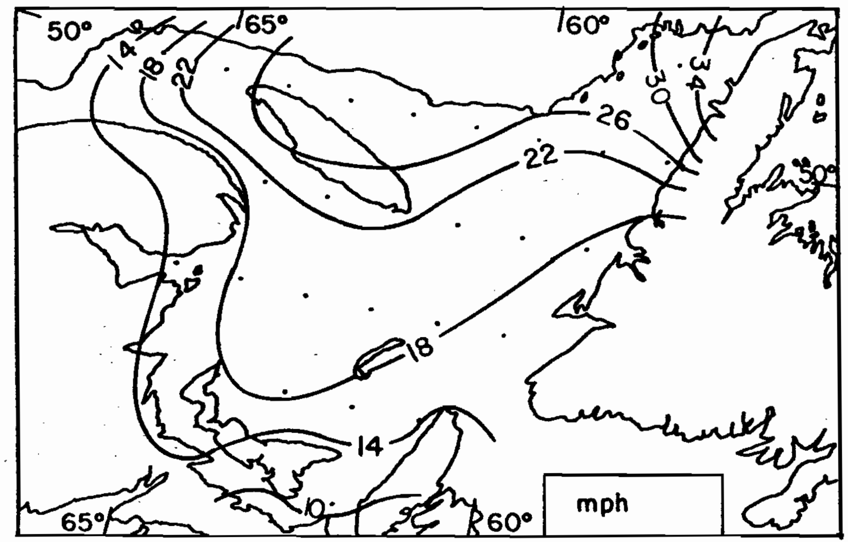

Fig. 6.8 Surface winds

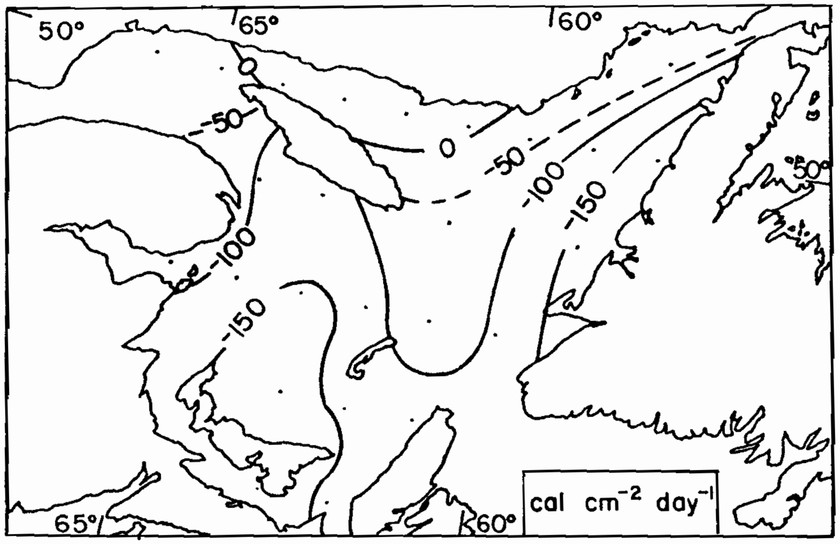

Fig. 6.9 Southerly flow evaporative heat flux

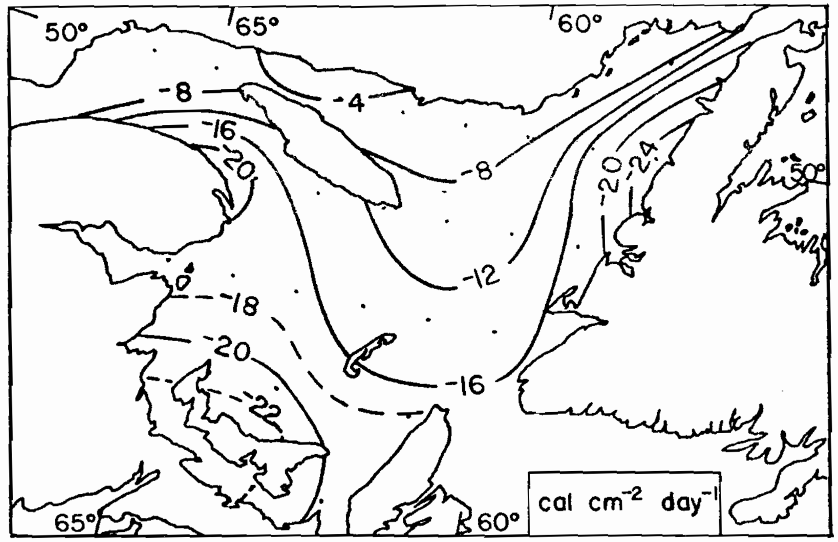

Fig. 6.10 Southerly flow sensible heat flux

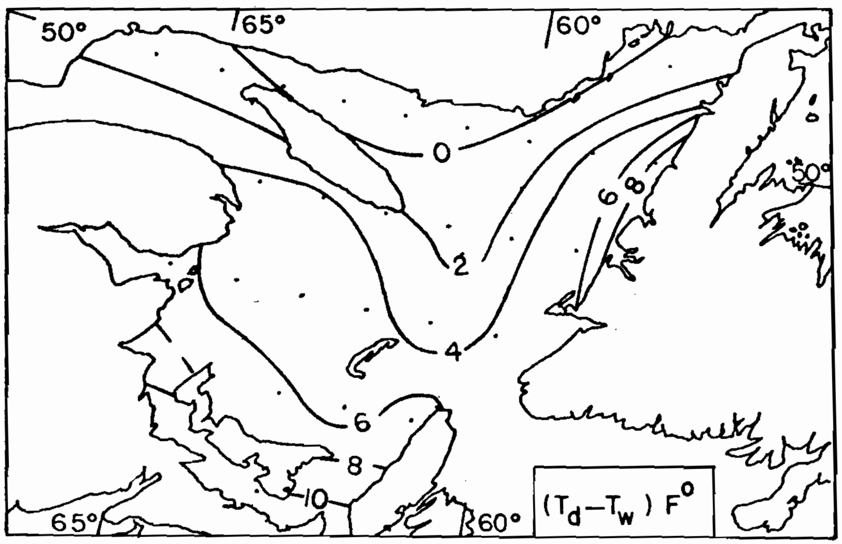

Fig. 6.11 Air dew point-to-water temperature difference

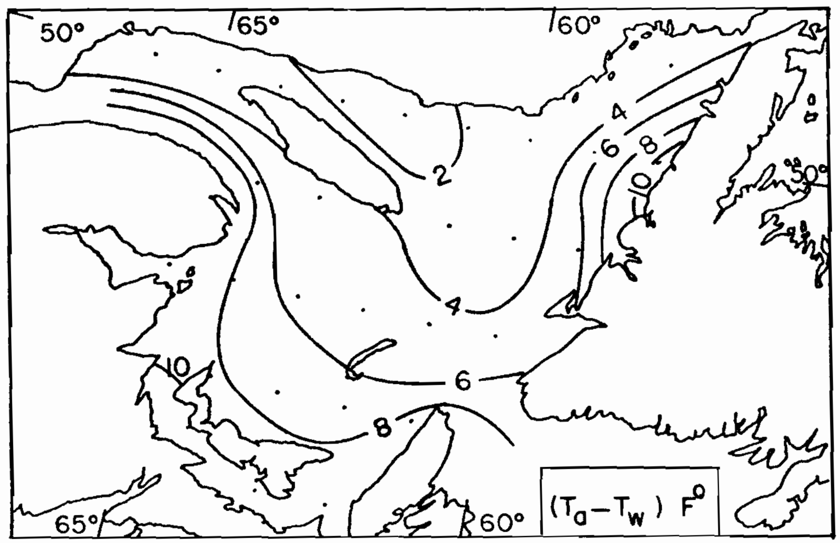

Fig. 6.12 Air-to-water temperature difference

warm air over the cooler water surface. Fig. 6.11 shows the air dew point-to-water temperature difference, and Fig. 6.12 shows the air-to-water temperature difference, approximately representing the evaporative and sensible heat fluxes respectively.

A typical flux pattern for southerly flow would be made up of large downward fluxes immediately adjacent to land masses, but decreasing in absolute magnitude towards the north as the air at the surface becomes modified.

6.2.4 NORTHEASTERLY FLOW, TYPE IV

The situation of northeasterly flow was studied for the period from April 11 to 15, 1965.

Two relative maxima of upward evaporative heat flux are indicated in Fig. 6.13. The maximum in the Magdalen Shallows was caused by high wind speeds (Fig. 6.14). Such high speeds may ensue in this area of the Gulf due to the 200 mile fetch which exists during northeasterly flow. The relative maximum of evaporative flux along the north shore of the Gulf resulted from a combination of increased wind speed, and a large temperature-dew point spread (Fig. 6.15) of the dry air being advected over the water.

The decrease in windspeed and temperature-dew point spread towards the Newfoundland west coast are responsible for the decrease in the evaporation rate towards this coast.

The sensible heat pattern (Fig. 6.16) shows a maximum heat loss region over the central portion of the Gulf. Since the water

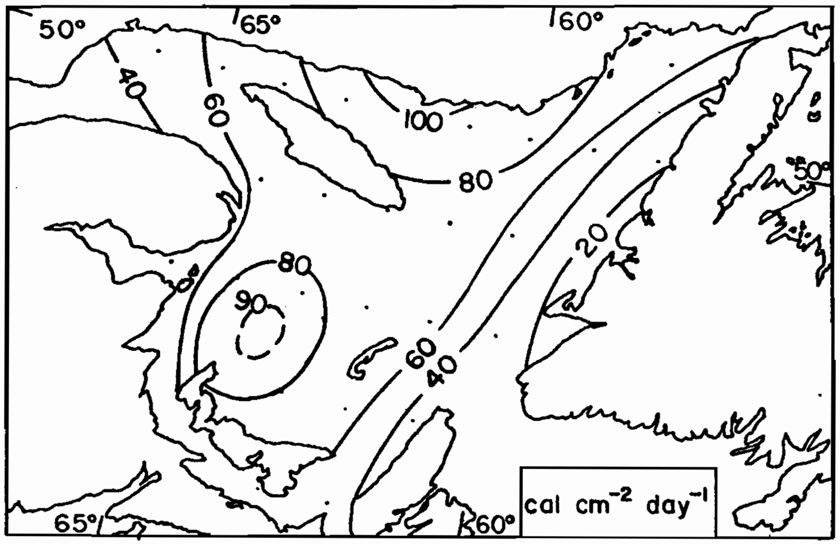

Fig. 6.13 Northeasterly flow evaporative heat flux

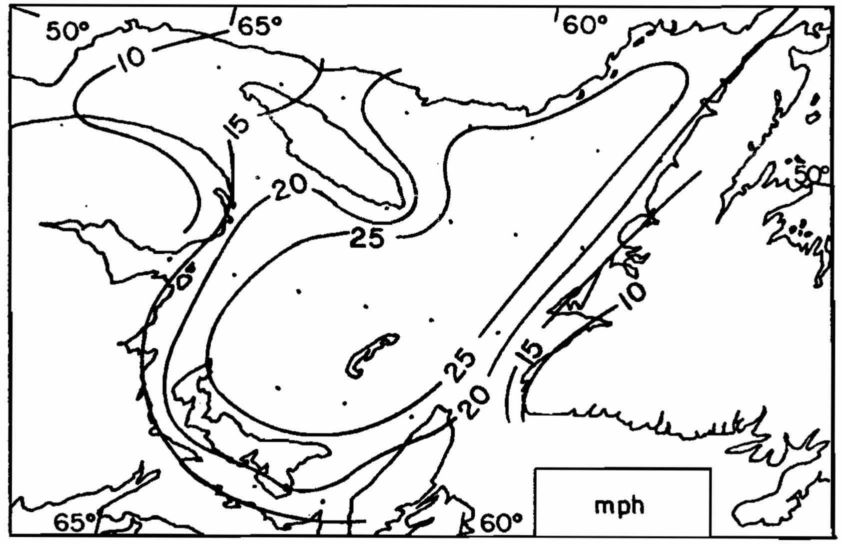

Fig. 6.14 Surface Winds

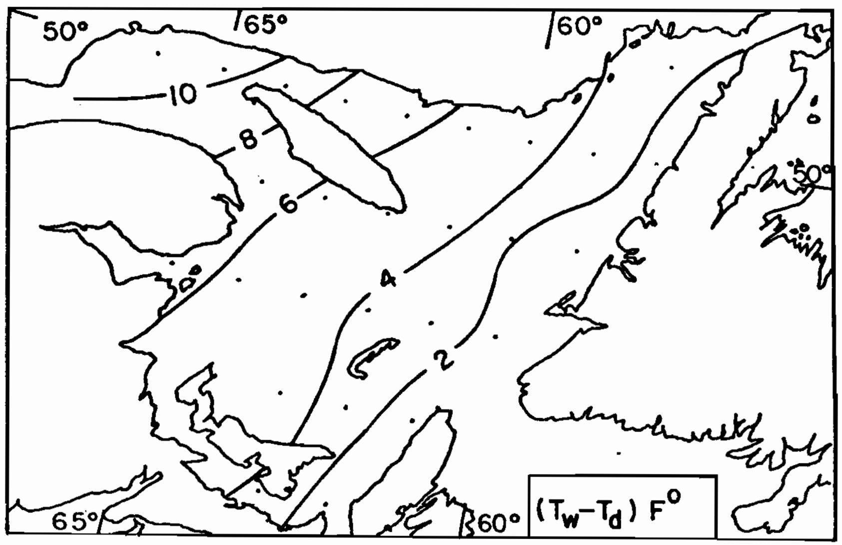

Fig. 6.15 Air dew point-to-water temperature difference

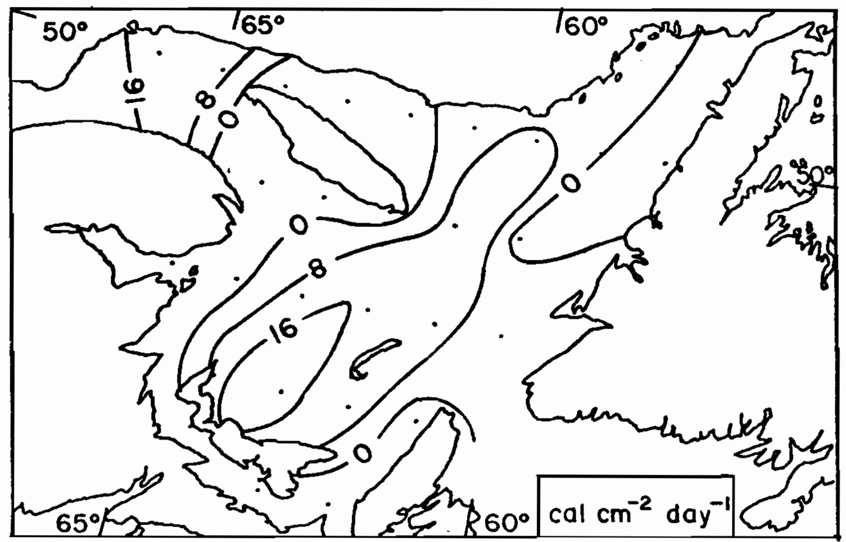

Fig. 6.16 Northeasterly flow sensible heat flux

temperatures were higher than the air temperatures throughout the Gulf region, the wind speed does not enter the calculation (see equation 4.16). Thus, it could only have been the increased air-to-water temperature gradient which could have caused the higher sensible heat losses in the central region of the Gulf.

6.2.5 ANTICYCLONIC FLOW

Selected for study was the February 12 to 14, 1961 case of an anticyclonic surface pressure pattern, typified by its low wind speeds. The maxima of evaporative and sensible heat flux (Figs. 6.17 and 6.18) which arose in the Jacques Cartier Passage, were a consequence of the low ice concentration (Fig. 6.19).

6.3 ANALYSIS OF THE DAILY TURBULENT HEAT FLUX TERMS

The daily mean evaporative and sensible heat components of the total heat flux, for those days which made up the 3-day, or longer, periods of a unique flow type, have been plotted in Figs. 6.20 and 6.21 for the 1960-61 and 1964-65 ice seasons.

A comparison of the two figures indicates the higher heat losses which occurred during the 1964-65 season. Indeed, of the days which were studied, the mean evaporative flux was more than 35% higher, while the mean sensible heat loss was 7% higher than the 1960-61 winter. Recalling that 1964-65 was the lighter ice season, it may be concluded that high heat losses alone, occurring throughout the ice season, do not necessarily result in severe ice conditions.

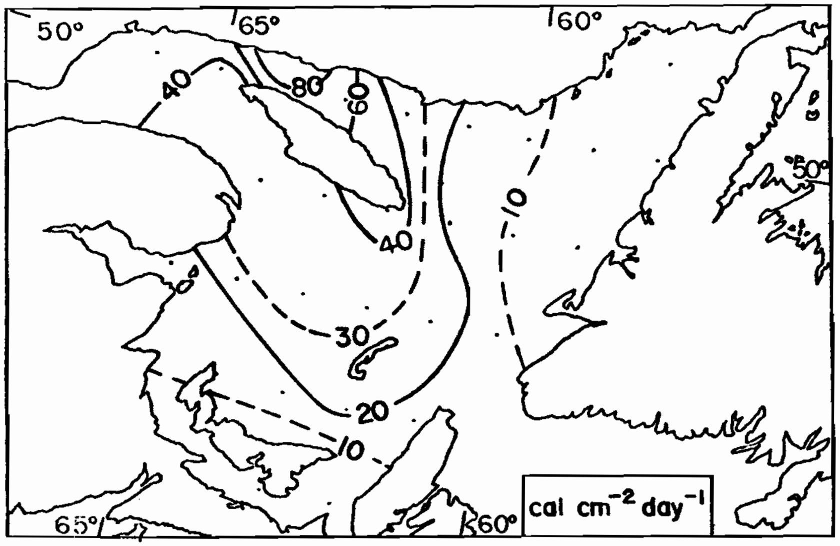

Fig. 6.17 Anticyclonic flow evaporative heat flux

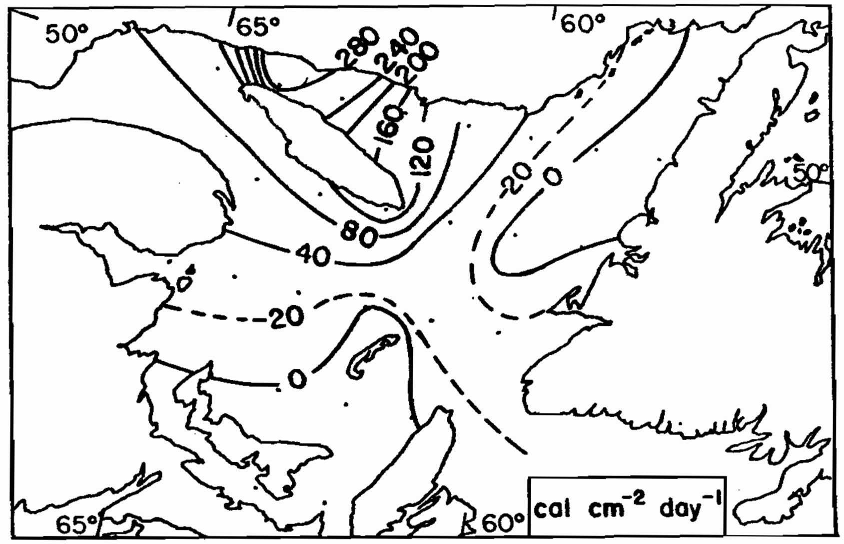

Fig 6.18 Anticyclonic flow sensible heat flux

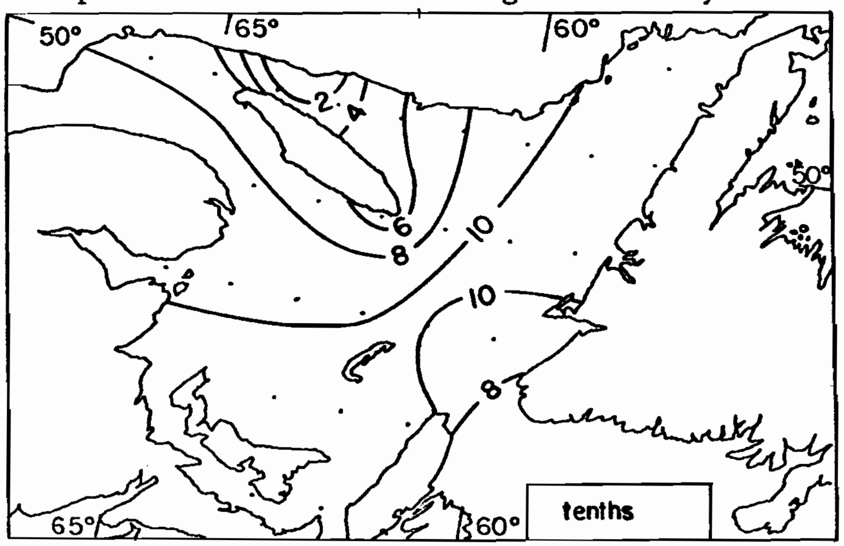

Fig. 6.19 Ice concentration

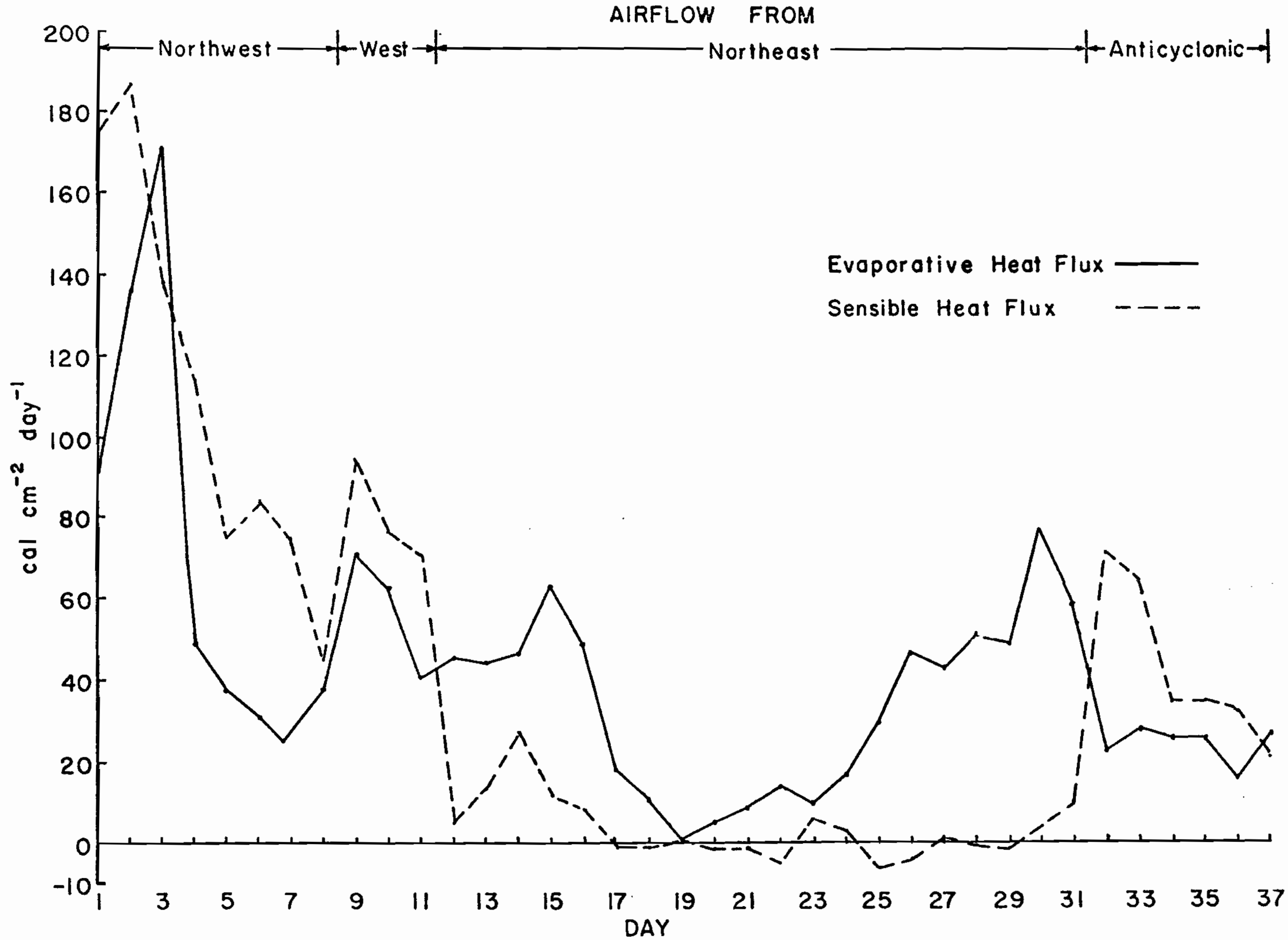

Fig. 6.20 Daily turbulent heat flux terms, 1960-61

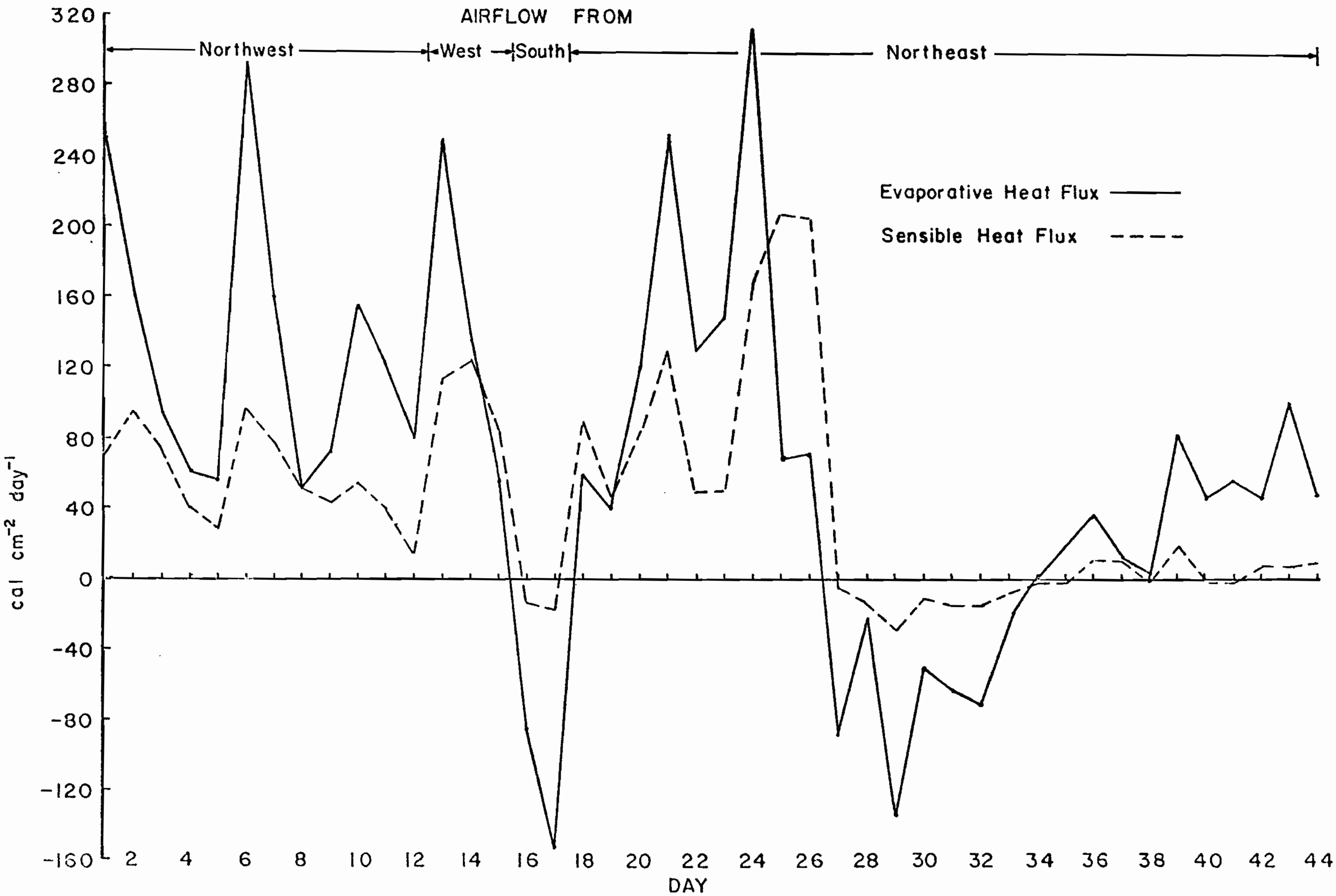

Fig. 6.21 Daily turbulent heat flux terms, 1964-65

The daily classification into flow types was a rather subjective process. However, with the exception of one extraneous period of high fluxes during northeasterly flow during 1964-65[6], the practice of classification into unique flow types was successful, in that it produced groupings which were indeed different by virtue of the relative magnitude of their heat fluxes.

With reference to the high evaporative fluxes during northeasterly flow in 1960-61 (Fig. 6. 20), cognizance must be paid to the sensible heat flux as well. For as much as the evaporative heat loss was tantamount to the losses during outbreaks of northwesterly and westerly air, the sensible heat was of a magnitude which would argue for its retention in a class by itself, i. e. , northeasterly.

6.4 SEASONAL TREND OF THE DAILY TURBULENT HEAT FLUXES

Daily heat gains or losses were obtained by summing the area weighted daily means of the evaporative and sensible heat fluxes. The fluctuations in the heat loss rates during the winter seasons 1960-61 and 1964-65 are plotted in Figs. 6. 22 and 6. 23. In order to observe the cause-effect relationship between eddy heat transport and ice modification, each day's weighted mean ice concentration has been plotted, and to provide a basis for comparison the area weighted quinquennial mean ice concentration has also been plotted. (Where necessary, the ice concentration has been interpolated over the days

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[6] The cause of the high fluxes was discussed in Section 6. 1

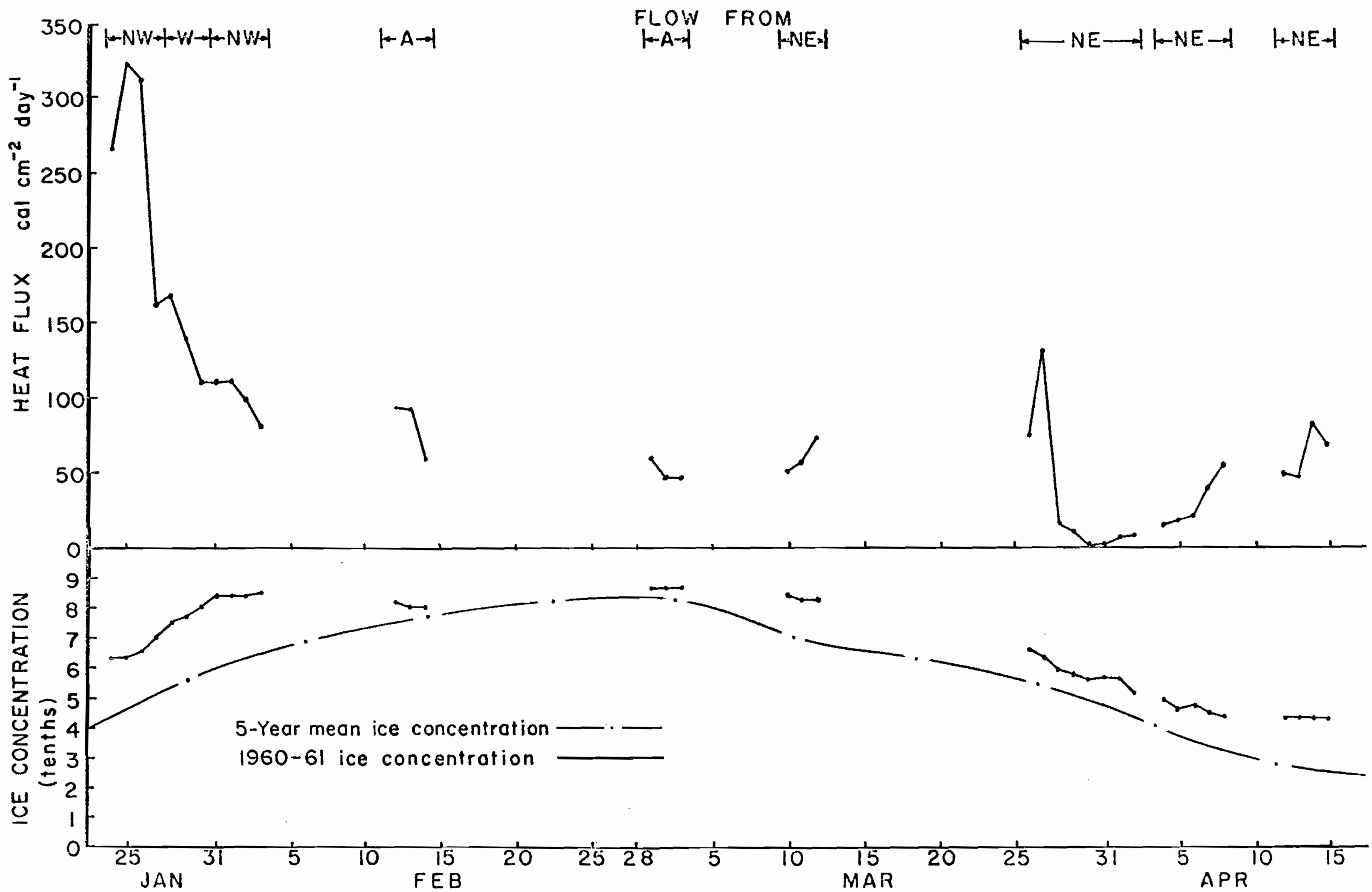

Fig. 6.22 Turbulent heat fluxes and ice concentration, 1960-61

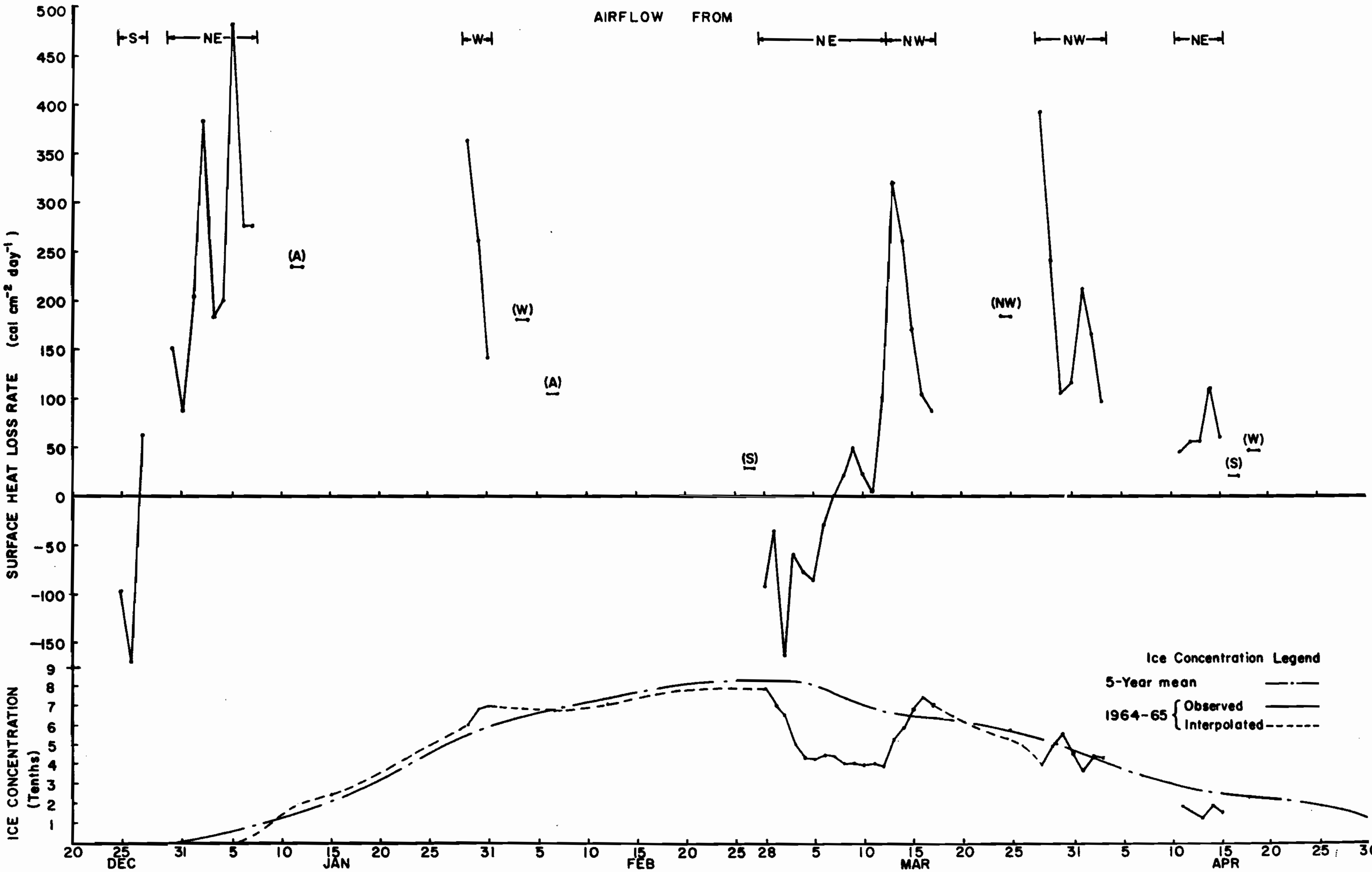

Fig. 6.23 Turbulent heat fluxes and ice concentration, 1964-65

for which the heat exchanges were not calculated). Note that the parenthesized airflow types in Fig. 6. 23 are for 2-day duration periods, to be discussed in Section 6. 5.

The basic interest is in the manner in which the atmospheric circulation affects the growth and decay of ice. Directing attention to Fig. 6. 23, it is apparent that the most rapid fluctuations in ice concentration show a marked correspondence to the periods of high heat exchange. During the intervening periods - which were subjected to a variety of flow types, none of which lasted for more than two days - no significant changes occurred in the ice cover. Admittedly, there were changes in ice concentration during the intervals for which the fluxes were not computed which were equal in magnitude to the variations which took place during high heat exchange. However, these changes took place over a much longer period. Thus, the existence of days for which heat fluxes were not calculated is not a serious drawback. By confining attention to the periods of long duration of a unique flow type, the significant perturbations imposed upon the mean ice conditions have been isolated for investigation; while the changes during the intervening periods are merely following the seasonal ice cover trend.

An excellent example of the control imposed upon ice development by the heat fluxes occurred in the course of a spring thaw during the first two weeks of March 1965 (see Fig. 6. 23). Within four days the mean ice concentration over the Gulf decreased by almost 4 tenths as a consequence of the heat recovered from the atmosphere. Subsequently, under the influence of an extended spell of northwesterly flow, and the associated surface heat loss, ice reformed to the intensity it possessed prior to the thaw.

Although not as prominent, the creation of ice which evolved from the period of cold westerly air advection during the latter days of January 1965 (Fig. 6. 23) can also be distinguished. Similar patterns exist in late March and mid April.

In marked contrast to the season discussed previously, the 1960-61 ice season differed greatly by virtue of the diminished seasonal heat exchange. The fact is that extended periods of high flux-inducing northwesterly or westerly flow did not exist after the end of January 1961 (Fig. 6. 22).

The greatest heat losses, which arose during a 10-day spell of cold, westerly and northwesterly flow in late January 1961, led to the creation of an ice cover which was concentrated enough to reduce considerably further heat losses during the winter. On the other hand, the 1964-65 winter lacked the opportunity to release, early in the season, the heat stored in the surface layers, and a well developed ice cover never did form. The low ice density was responsible for the large upward fluxes which were observed during the season.

6.5 AN INVESTIGATION OF TWO-DAY FLOW DURATION PERIODS

As mentioned in the introductory paragraph to this chapter, the decision was made to study the effect on ice cover resulting from flow types which had persisted for 3 days or longer only. This criterion was set after considering the information provided in

Table 6.2 - made up of days for which the same flow persisted for only two days.

The most outstanding feature revealed by this table is the very high fluxes for some cases of northwesterly flow - notably those which occurred while the ice concentration was low. These values are not at all unrealistic, for Craddock (1951) determined the total rate of gain of heat by a typical Arctic air mass passing over the warm Eastern North Atlantic to be 47 cal cm^{-2} hr^{-1}, or 1128 cal cm^{-2} day^{-1}, a very large value compared with everyday experience. Craddock admits that "a cold air mass can be modified far more rapidly in temperate latitudes by passage over warm sea than by summer insolation".

Considering the mixed surface-water layer of the Gulf to extend to a depth of 40 metres (Lauzier, Trites and Hatchey, 1957), and assuming complete mixing, a total heat loss of 1000 cal cm^{-2} day^{-1} from the surface would cause the entire surface layer to cool by 0.25C^{o}. This provides a feeling for the rapid autumnal cooling which is possible in the Gulf.

By comparing the magnitudes of the heat flux components in Table 6.2 to those in Table 6.1, the following points are noted: Apart from the cases of high flux during low ice concentration, the values for the two-day durations of northwesterly and westerly flow fall within the range of the fluxes calculated for the long period flows. The cases of southerly flow, however, differ widely, while the differences in ice concentration do not permit a comparison of the anticyclonic cases. Although the very small sample size does not

TABLE 6.2

SHORT FLOW DURATION HEAT FLUXES

Date	Flow	$\overline{Q_e}$ cal $cm^{-2}day^{-1}$	$\overline{Q_h}$ cal $cm^{-2}day^{-1}$	$\overline{\lvert Q_e+Q_h \rvert}$ cal $cm^{-2}day^{-1}$	$\overline{CONC}$ tenths
Dec. 19, 64	NW	530	391	921	0
Jan. 15, 65	NW	523	414	937	2
Mar. 22-23, 65	NW	161	107	268	6
Mar. 24-25, 65	NW	121	62	183	5
Feb. 3-4, 65	W	71	89	160	7
Apr. 18-19, 65	W	45	1	46	2
Feb. 26-27, 65	S	29	0	29	8
Apr. 16-17, 65	S	18	0	18	2
Dec. 21, 64	A	206	445	651	0
Jan. 11-12, 65	A	135	233	368	2
Feb. 6-7, 65	A	20	84	104	7

Bar indicates area weighted mean.

permit generalities, there is a suggestion that the surface characteristics (and so the typical heat fluxes) of northwesterly and westerly flow become established more rapidly than those for southerly flow.

Finally, the total evaporative and sensible heat fluxes for the two-day periods were plotted in Fig. 6.23. It is made quite obvious that these short duration periods had very little effect on the growth or decay of ice. For even though the fluxes were high, they were of such short duration that the total heat loss was not sufficient to influence the ice cover significantly.

For these reasons, it was decided to consider only the flow types which had continued for three days or longer.

CHAPTER VII

CONCLUSIONS ON METEOROLOGICAL INFLUENCE ON THE GROWTH AND DECAY OF ICE

7.1 PRE-WINTER INFLUENCES ON ICE COVER

7.1.1 OCEANOGRAPHICAL INFLUENCE

It is quite possible for a large variation to exist in the date of freeze-up, due to the variability of oceanographic parameters from year to year. The two most important parameters are temperature and salinity and their change with depth, which determine the water density at each level, and hence its stability.

Once an ice cover is established, however, the salinity becomes the only important oceanographic factor since it determines the temperature of the lower face of the ice cover. The Gulf can then be considered as an infinite heat source - the heat being released as latent heat during ice formation - its loss being determined solely by the thermal properties of the ice and the atmospheric conditions (Pounder, 1965).

None the less, meteorological conditions during the weeks prior to complete freeze-up are of great importance, and the effects of the different types of airflow can be such as to give ice conditions quite different from those expected from an examination of the autumnal oceanographical observations. As was seen in the previous chapter, the meteorological effect is also clearly evidenced during the weeks following the late January freeze-up.

Autumnal heat losses establish a convective cooling process which penetrates deeper and deeper as the water temperature decreases towards the freezing point. The depth to which this mixing must penetrate may be calculated from oceanographic soundings, and so provides a method of determining the amount of heat which must be extracted before an ice cover will begin to form.

Pre-freeze-up oceanographic data for the 1960-61 season are not available, so the influence of the oceanographic parameters cannot be compared to the 1964-65 data. Consequently, the only option available is to investigate the climatological factors which dominated the late summer and fall preceding each winter, and which therefore determined the temperature and salinity structure of the Gulf.

7.1.2 CLIMATOLOGICAL INFLUENCE

The following discussion is based on the Monthly Weather Maps, and the Canadian Weather Review, for the July to November period preceding each winter season.

During the summer and autumn of 1960, monthly mean temperatures along the Eastern Canadian Seaboard were all close to, or well above normal, while the corresponding months in 1964 were all colder than normal.

Precipitation along the St. Lawrence River watershed in 1960 was normal, with only slight deficiencies; although in sharp contrast, the Gulf region was subjected to record low precipitation amounts. During 1964, the St. Lawrence River watershed had deficiencies greater than in 1960, while the Gulf was close to, or above, normal.

In a computation of the monthly mean values of total daily solar radiation over the two summer-fall periods (data taken from the Monthly Record for five stations around the Gulf) it was found that the monthly mean insolation was 13% greater in 1960 than in 1964.

Also, from the Monthly Record, wind summaries indicate that the mean wind speed during 1960 was slightly higher (8%) than in 1964.

It is now possible to postulate the combined effect which these four factors should have on the establishment of the oceanographic conditions which existed prior to each of the winter seasons.

Over the Gulf, mean air temperatures during 1960 were higher than in 1964, producing a warmer surface-water layer. Also, during 1960, there was a rainfall deficiency over the St. Lawrence River and Gulf watersheds, reducing the amount of fresh water input, leaving the Gulf waters more saline. The higher flux of solar radiation in 1960, coupled with the higher wind speeds, produced, due to turbulent mixing, extra deep penetration of surface heating.

In conclusion, a body of water which is warmer both at the surface and to a greater depth, as well as being more saline, will take longer to freeze - provided the meteorological conditions are comparable. It could be expected, therefore, that ice formation would commence later in 1960 than in 1964. None the less, in spite of the unfavourable oceanographic conditions, ice formation did begin earlier in 1960, pointing out the very dominant effect which the atmospheric situation during December and early January must have had on overcoming the much greater convective depth of 1960.

7.2 ICE CONDITIONS DURING THE TWO WINTERS

Making use of the ice observations published by Beaton (1962) and in the Ice Summary and Analysis, 1965, the areas covered by each concentration of young, winter, and fast ice[7] were measured. Since the concern is more with the total amount of ice present, it was found more beneficial to form the cumulative product of each ice concentration and its areal extent. This provides the area of the Gulf which would be covered if all the ice were compressed to 10 tenths concentration.

Fig. 7.1 shows the seasonal trend, in intervals of two weeks, of this hypothetical ice cover. Owing to its relatively small magnitude, the area of the Gulf covered by fast ice was not plotted. It may be noted that the area covered by fast ice, which reached a peak by mid February, amounted to 6% and 3% of the total Gulf area in 1961 and 1965 respectively.

It can be seen in Fig. 7.1 that throughout the two winter seasons, the ice during 1960-61 was much heavier than the 1964-65 conditions. The figure also indicates the rapid growth to winter ice

[7] Sea ice is classified by age into young ice, which is from 2 to 6 inches thick and has usually formed within the past few weeks, and winter ice, which originates from young ice, is thicker, and of not more than one winter's growth. Fast ice is found along the coasts and is ice which remains fast, generally in the position where it originally formed.

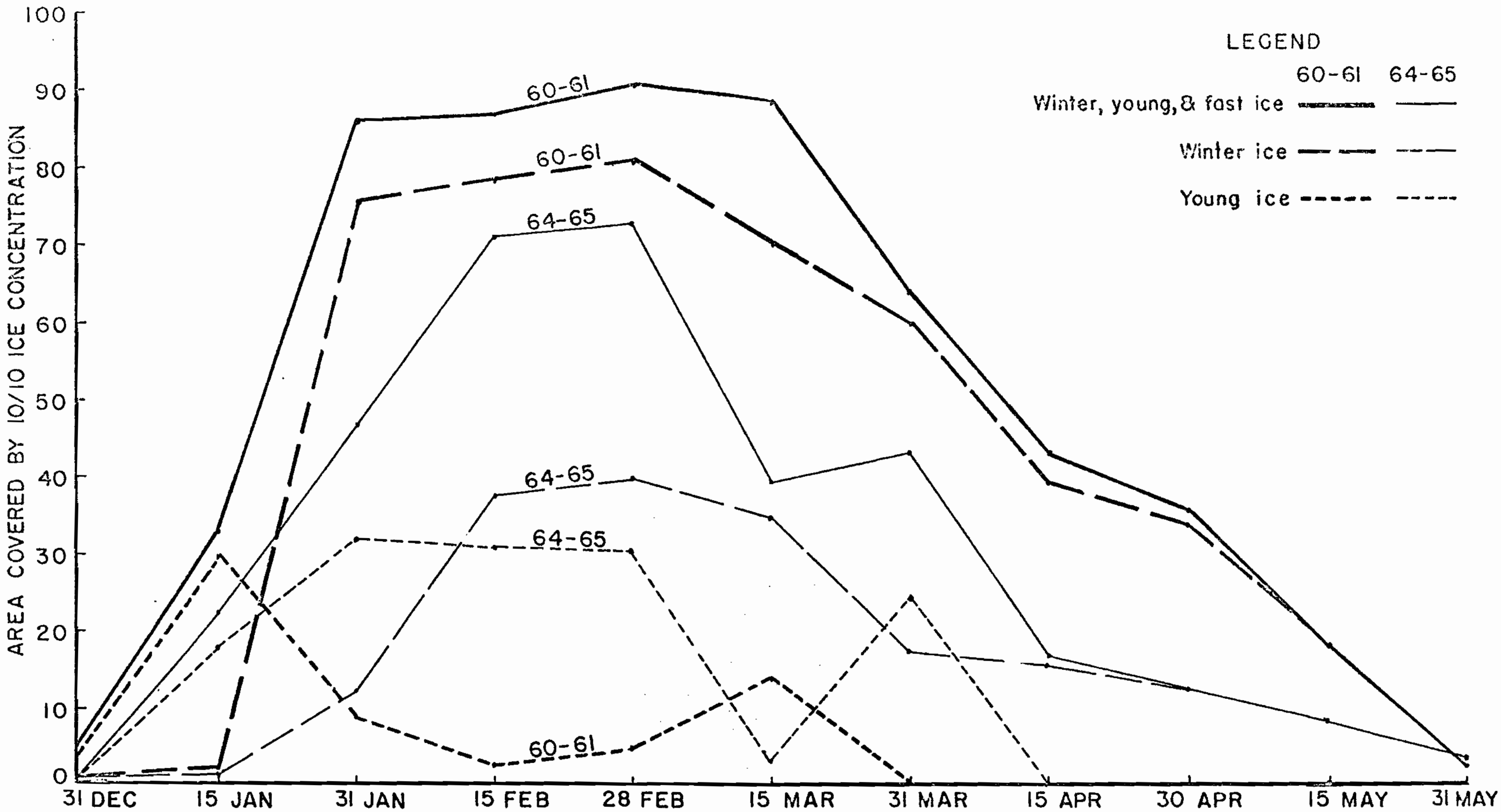

Fig. 7.1 Area of the Gulf covered by 10 tenths ice

conditions which occurred during 1960-61. The ice during the 1964-65 winter shows a decrease in mid March, but an increase took place later in the month. During April in both seasons, the rate of clearing of the Gulf was quite slow, as ice became congested in the southwestern region, the pile-up being more intense in 1960-61 than in 1964-65.

One drawback of Fig. 7.1 is the fact that the ice observations do not include the thickness of the ice, so that winter ice volume could change considerably, and not show up in the diagrams.

A discussion of the changes in the ice cover will be presented in the following section, with their relation to the circulation patterns.

7.3 THE RELATIONSHIP BETWEEN AIRFLOW AND THE GROWTH AND DECAY OF ICE

The intention is to explain how the ice formation and decay processes in the Gulf of St. Lawrence are influenced by the variety of synoptic airflow patterns which affect the region. This will be accomplished by means of the circulation data discussed in Chapter III, and the heat fluxes associated with the different flow types considered in Chapter VI.

In particular, the problem is to explain the biweekly changes in the actual ice cover observations indicated in Fig. 7.1 by means of the flow distribution over the two-week periods indicated in Fig. 3.4, by the duration of each flow case as revealed in Fig. 3.5, and also by the total heat fluxes plotted in Figs. 6.22 and 6.23. As an aid to the explanation, a combined form of the information provided by Figs. 3.4 and 3.5 has been produced, viz. Fig. 7.2. This diagram provides a comparison between the two winters, in two-week periods, of: (a) the percentage frequency of occurrence of each flow type over the two weeks (vertical axis), and (b) the duration periods, in days, of the flow types (figures in body of diagram), and their frequency of occurrence (horizontal axis). As an example, for westerly flow, during the two-week period ending December 31, 1961, there was one occurrence which lasted for 2 1/2 days, and three occurrences which lasted only 1 day. Note that flow types which were maintained for only half a day were omitted, since the characteristics of the flow could not become effective in such a short period.

Period ending December 31

It is first noted in Fig. 7.1, that by the end of December 1960 more ice had formed than by the end of December 1964. It has been argued, by a comparison of the oceanographic conditions preceding each winter, that the ice situation should have been reversed, i.e., heavier in 1964 than in 1961, the difference then, must be attributed to atmospheric control.

During the last two weeks of December, the major contrast between the two seasons, as Fig. 7.2 indicates, was the domination of cool westerly flow in 1960. It also shows that of the five separate occasions on which westerly flow occurred, it lasted for 2 1/2 and 1 1/2 days on two of these occasions. These relatively long duration periods of airflow permitted the influences of the westerly flow to take

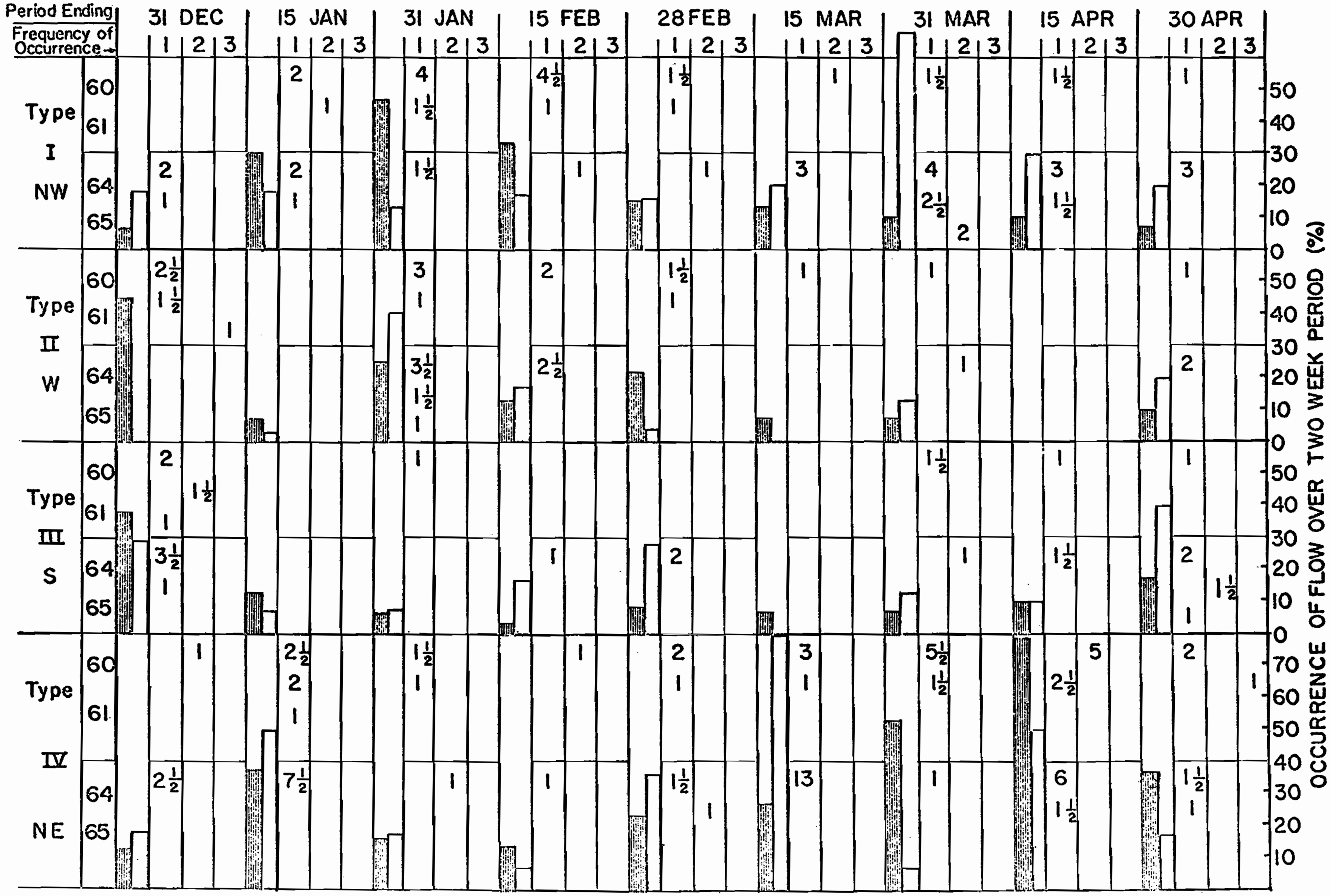

Fig. 7. 2 Biweekly distribution of flow types and duration of flow 1960-61 1964-65

effect - for if the duration had been only a day or less, the influence would not have had time to take effect.

It will be recalled from Section 3.2.1 that Fig. 3.3(c) revealed that there was more warm air advection early in 1960 than in 1964, yet there was more ice formed early in 1960. The explanation to this observation is as follows:

Fig. 7.2 shows that in late December, both winters were subjected to a high frequency of warm Type III (southerly) circulation. However, one duration period extended over $3^1/2$ days in 1964 (the magnitude of the surface heat gain is shown in Fig. 6.23), while the longest period in 1960 was only 2 days. It appears, therefore, that the long duration of warm advection in December 1964 did much in hindering ice formation, while little effect was felt by the short duration periods during late December 1960.

During the final two weeks of December 1964, Type I (northwesterly) airflow existed for almost 20% of the time and lasted for one 2-day period. It is to this cold flow that the small amount of ice which was present owed its existence.

Type IV (northwesterly) airflow was almost as frequent in 1960 as in 1964, so little can be expected in the way of contrast.

Period ending January 15

Through the first two weeks of January, the rate of new ice growth in both winters was almost identical, so that in comparing the two years, no large contrasts in flow type frequencies should be expected. This notion is upheld by Fig. 7.2, which indicates only a slight excess of the cooler northwesterly and westerly airflows in January 1961. More important, however, is the fact that there was very little difference in the duration periods of these flow types.

A comparison of the Type IV airflows produces the interesting observation that during early January 1961 the total occurrence was divided into separate intervals, (1, 2 and 2 1/2 days), while in 1965 the total occurrence was made up of one continuous 7 1/2-day stretch. Fig. 6.23 shows the high magnitudes of the surface heat loss for the 7 1/2-day case. As the changes in ice concentration for the two years were almost identical, it appears that the long duration period produced the same effect on the ice as did the short flow intervals. This suggests that the characteristics of northeasterly flow can become established in a short time period.

Period ending January 31

The last two weeks of January 1961 saw a very rapid increase in the amount of ice, a very large portion of it being winter ice, while the same period in 1965 was characterized by a less rapid increase, with the larger portion of the Gulf being covered by young ice (see Fig. 7.1).

The explanation for this contrast is made clear in Fig. 7.2, for in 1961 airflow from the northwest was dominant. Furthermore, the flow was maintained for a continuous 4-day period - resulting in a much stronger effect on the ice cover than the 1 1/2-day interval in 1965. Fig. 6.22 also shows the very high heat losses which were involved.

Not only was Type II (westerly) airflow more frequent in

1965, but it lasted for longer time periods as well. Thus, the young ice which did form owed its existence mostly to this flow. The heat losses, and the increase which took place in the ice concentration, are indicated in Fig. 6.23.

It must also be added that the comparatively greater total heat loss in late January 1961, during the series of westerly and north-westerly flows, explains why a larger amount of heavy winter ice was formed rapidly in 1961, while only young ice was created under the influence of westerly flow in 1965.

Very little contrast in Type III and IV flows can be observed between the two winters.

February

Consolidation of the ice cover took place during February of both winters, and other than the increase in thickness, no marked changes in the ice cover were observed. For this reason, a detailed discussion of the meteorological influence has not been included in this investigation.

Period ending March 15

The first half of March 1961 brought the first signs of break-up with the development of open leads. Winter ice decreased, but a considerable amount of young ice was formed during the period.

In early March 1961 the two cases of one-day duration of northwesterly flow and the single case of one-day's duration of westerly flow were enough to support the growth of new young ice in the Estuary, but not enough to maintain the winter ice amount over the entire Gulf. Also, the high frequency of Type IV (northeasterly) flow certainly could not serve to increase the amount of winter ice. Fig. 6.22 indicates the rather low heat loss that occurred during the northeasterly flow.

A significant feature to note is that the domination of northeasterly flow during the early weeks of March 1961 marks the beginning of a trend towards the domination of this flow type right up to the end of the 1961 ice season. The wind stress on the pack ice under northeasterly flow causes the ice to drift into the southwestern region of the Gulf.

During March 1965 extensive melting of the young ice in the Gulf occurred. Concurrently, an influx of winter ice took place through the Strait of Belle Isle.

The major flow condition in early March 1965 was the continuous 13-day period of northeasterly flow. Its effect was to advect warm air completely around a cyclonic centre which had remained stationary to the southeast of the Gulf region (the surface heat gains are revealed in Fig. 6.23). This definitely inhibited the formation of new ice and melted vast amounts of any ice present. A compound effect of the northeast wind stress was to cause the advection of ice into the Gulf through the Strait of Belle Isle, explaining why such a large amount of winter ice still remained at the end of the relatively warm, two week period.

It must also be stressed that temperature changes are not rapidly reflected in changes of winter ice amounts, since a very thick winter ice cover can decrease in thickness down to six inches and yet still be recorded as winter ice. None the less, the ice advection into the Gulf during early March was considerable and cannot be neglected.

Period ending March 31

During the second half of March 1961 there was a complete absence of new young ice growth, coupled with a decrease in the amount of winter ice.

The trend towards domination of northeasterly flow, which started early in March 1961, continued through the remaining weeks of March, with continued congestion of ice in the Magdalen Shallows. The advection of this moderately warm air, which lasted for periods of $5^1/2$ and $1^1/2$ days - certainly incapable of supporting ice growth at this time of the year - melted all the young ice which had been present in the Gulf, as well as decreasing the amount of winter ice.

The cooler flow types (Types I and II) during March 1961 were so infrequent as to be ineffectual in producing lasting amount of young ice, and any ice which might have formed was soon melted by the warmer flow types (Types III and IV).

Atmospheric effects during the last two weeks of March 1965 produced a radical change in the ice cover, as suggested in Fig. 6.23 by the extensive formation of new young ice. It should also be noted that there was a concurrent decrease in the amount of winter ice during the same period (see also Fig. 7.1).

The striking feature of the second half of March 1965 was the very high frequency and long duration periods - 4 days, $2^1/2$ days, and two 2 day periods - of northwest flow. This resulted in a significant growth of new young ice over the areas of open water, particularly in the Estuarial region of the Gulf. A second effect of the persistent northwesterly flow was the discharge of ice which ensued through Cabot Strait, thus explaining the decrease which took place in the amount of winter ice, but still allowing the formation of young ice in the Estuary.

The occurrences of the three remaining flow types were so infrequent that they had little effect on the ice cover.

Period ending April 15

During the first two weeks of April 1961 the amount of winter ice continued to decrease at a fairly rapid rate. The trend towards domination of northeasterly flow late in the 1960-61 winter continued through early April, and ice congestion in the Magdalen Island-Prince Edward Island region of the Gulf was maintained. The decrease in the amount of ice was due to energy transfer terms other than the turbulent terms, i.e., either radiational or advective. This must be so since Fig. 6.22 shows a surface energy loss over a number of days in early April 1961. Also, the virtual absence of cold air advection disallowed any further growth of winter ice.

By the end of the first two weeks of April 1965, all the young ice in the Gulf had melted, while the amount of winter ice decreased only slightly. The 3-day and $1^1/2$-day stretches of northwesterly flow occurred early in the two week period, and any young ice which might have formed had disintegrated by mid April. It appears that the winter ice was thinning and rotting gradually, but maintaining its concentration.

Period ending April 30

In the two winters the decrease in ice continued at almost equal rates, although at a slower pace than during the first two weeks of April.

Northeasterly flow was again dominant in April 1961 and resulted in the retention of a significant amount of ice, due to wind pile-up, in the southwestern area of the Gulf. The remaining flow types in 1961 were of such short frequencies and durations as to be ineffectual.

During late April 1965 the 3-day stretch of cold northwesterly flow combined with the 2-day stretch of cool westerly flow were sufficient to balance the frequent, and reasonably long, periods of warm southerly flow, in such a manner as to allow only a slight decrease in the total quantity of winter ice.

May

The month of May 1961 was marked by a steady and rapid decrease of winter ice, while during May 1965 the melting process was very slow, and congestion persisted in the northeast arm of the Gulf.

Final clearing in both winters had taken place by the early weeks of June.

7.4 SUMMARY

It has been shown that the rather subjective qualitative analysis of the surface pressure charts was capable of revealing the agents which prompt or hinder ice formation or decay. These controlling factors have been established to be:

(a) the distribution of the flow types over the entire winter season,

(b) the distribution of the flow types over shorter two-week periods,

(c) and, most importantly, the duration of each flow type case.

High heat losses during the season do not necessarily imply severe ice conditions. Also, it is not sufficient simply to compare the total occurrence of each flow type in order to assess the ice severity, for the controlling influence on the ice cover changes is determined by the temporal distribution of each flow type.

The significance of the type of flow was revealed by the quantitative investigation of the upward or downward turbulent heat exchange associated with each airflow situation. Computation of the heat fluxes also revealed the very high heat gains or losses which may ensue if a flow type is maintained over a long period. It is imperative that these high heat loss periods continue for at least 3 consecutive days in order to produce a significant effect on the ice cover. Short durations of flow lasting for less than 3 days do not impose marked changes on the ice cover since the total heat loss is not sufficient.

There is an indication, however, that the characteristic heat fluxes associated with airflow from the northwest or west are higher, and tend to become established more rapidly (within $1^1/2$ to 2 days) than the characteristics of southerly flow. This can be explained by the relative heat capacities of air and water, since, for equal masses of air and water, more energy is required to heat the water. This would produce a damping effect on the downward heat transfer associated with southerly flow, and also tend to keep the magnitude of the flux small. The upward fluxes during northwesterly and westerly flow would be higher, and the heat could be released more freely.

The basic reason for the contrast in ice severity observed between the two winter seasons may be analysed as follows:

It must be realized that any heat which might be stored, or could be advected into the Gulf, must be removed before ice can begin to form. During the early months of the 1960-61 ice season, the atmospheric circulation was such that it possessed the capability to remove the heat stored in the Gulf. Further capacity was also available to cool the Gulf surface to the point which forced an extensive ice cover to form over the Gulf, creating severe ice conditions by the end of January.

The atmospheric circulation during the early months of the 1964-65 season completely lacked the capacity to absorb heat from the underlying surface. As a result, only a meagre amount of ice was created. The high upward fluxes during early January merely decreased the temperature of the surface-water layer. It was not until late in the ice season that the essential long duration periods of cold, dry air advection were available to transport heat away from the surface. At this time of year, however, vernal warming, together with the more frequent interspacing of warm air advection, were rapidly bringing about the decay of ice. The result was that the ice density during the course of the 1964-65 winter remained relatively low.

Thus, a severe winter will develop if the synoptic airflow is such that it possesses the capability to remove sufficient heat from the surface water layers to create an extensive ice cover early in the season. Provided thaws are absent, the heavy ice conditions will persist throughout the winter, and will only deteriorate with the vernal warming.

On the other hand, if the atmospheric circulation is such that the heat removing potential is absent early in the season, ice will not form. High upward heat fluxes will be possible during the winter as long as there is open water, and stored energy is available from the surface-water layer. If the stored energy is not sufficient, the ice concentration will increase as the latent heat of fusion is taken up by the atmosphere.

On the average (see Fig. 6.23), the ice tends to be most concentrated (greater than 5 tenths) from late January to late March. A sustained interval of cold air advection during this period would increase the concentration to a level which would warrant the ice year being classified as "severe". If, however, the cold spell had occurred after the end of March, the ice concentration would increase, but not to the point where it would seriously hinder navigation. The ice season would then be considered as having been "prolonged", but certainly not "severe". This appears to have been the case for the 1964-65 season.

No doubt, the autumnal bathymetric soundings provide valuable information on the heat content of the Gulf preceding each winter. However, it would be presumptious to attempt to form a legitimate ice forecast from these data alone, for such soundings merely provide one of the input parameters necessary in predicting future ice conditions. It has been shown that the ultimate control is imposed upon the water body by the nature of the atmospheric circulation in late autumn and throughout the winter.

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