JAMES BAY

MANUSCRIPT REPORT SERIES

No. 24

On the oceanography of James Bay

F.G. Barber

The tides of James Bay

G. Godin

Circulation in James Bay

T.S. Murty

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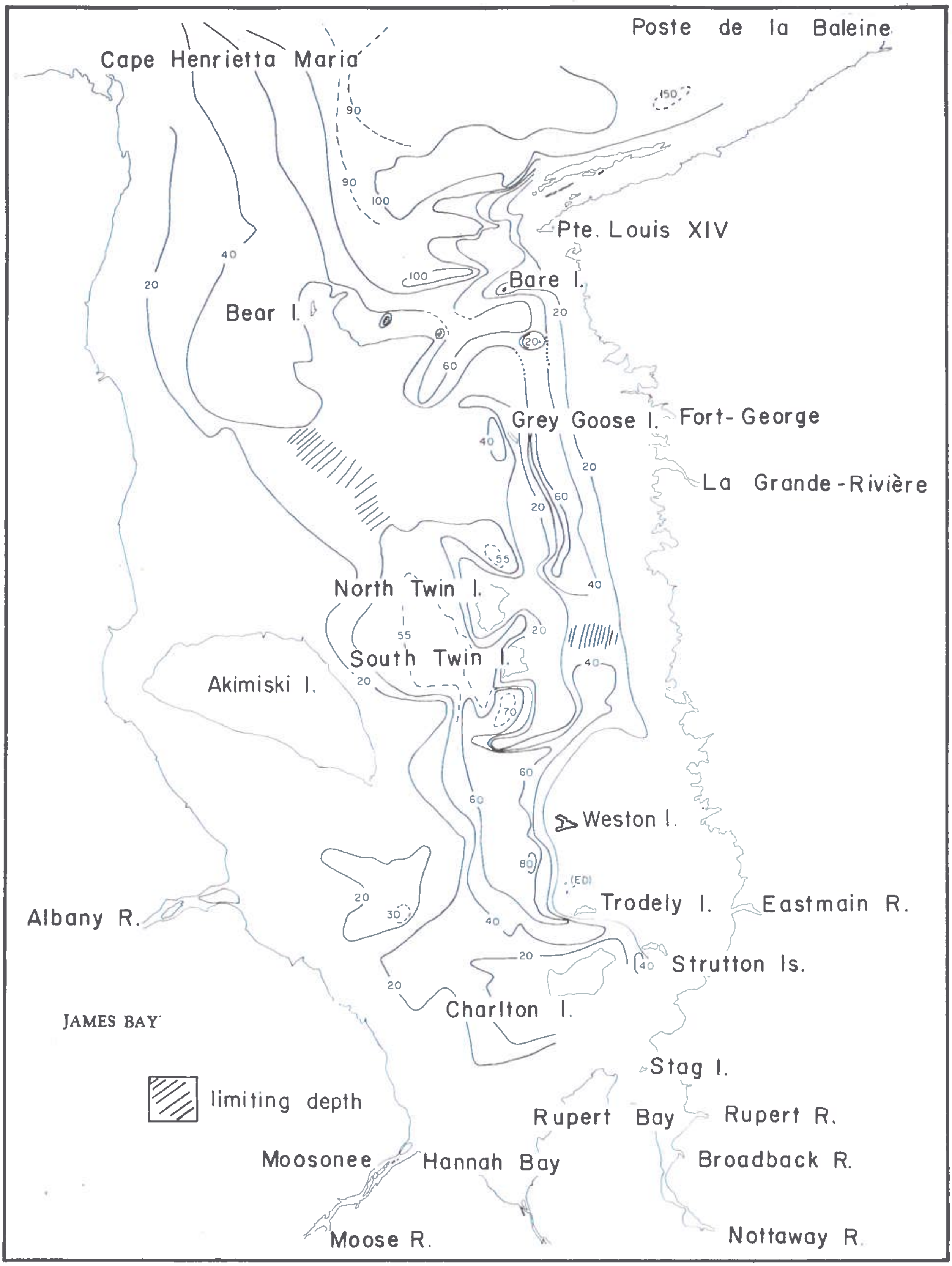
Marine Sciences Branch
Department of the Environment, Ottawa



Frontispiece Bathymetry (metres) of James Bay and some place names. An interpretation of bathymetric data for James Bay from Canadian Hydrographic Service Chart 5800, edition 1971, and from topographic maps of the Surveys and Mapping Branch.

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Foreword

The reports contained in this number of our manuscript report series were prepared in response to a request to Federal authorities from the government of the Province of Quebec for an initial consideration of the possible influence of the James Bay Project on the environment; the three reports are directed to the marine environment and comprise but a portion of the total activity. Specifically, the study was to be directed to a question concerning the area to be developed first, i.e. whether construction should begin on the southern rivers or on the northern rivers. It was realized from the outset that we could not, for a number of reasons, contribute significantly to the resolution of the question; nevertheless, it was deemed useful that we attempt to understand the oceanography of James Bay as well as available data would permit.

A.E. Collin,
Director,
Marine Sciences Branch.

December 2, 1971
Ottawa

Contents

On the Oceanography

of

James Bay

F.G. Barber

Contents

1. Introduction

The examination of certain aspects of the oceanography of James Bay (frontispiece) contained in the following was carried out in a consideration of the possible influence on the water of James Bay of the development of the hydroelectric potential there. The work constitutes more of a review or reappraisal of knowledge of the region than it does an impact study, but was believed necessary in our initial approach to the problem. The review aspect emphasizes the importance to James Bay of processes operating in adjacent waters (Figure 1) and the fact that all of these waters generally reflect to similar degree the influence of similar internal and external factors; in consideration of this, the word "system" has been used to include all the region (Barber, 1967, p20).

It was not surprising that an assessment of the influence of the development would prove almost intractable for, while data on the region are few, our understanding of both the qualitative and quantitative changes which may occur through oceanographic processes is limited. The consideration applies to the system as a whole as well as to James Bay, for not only are certain major features not understood but also much of the "understanding" is not well-founded; we have a great deal to learn about our northern waters generally. Part of the understanding could result through a study of the changes actually wrought on the system by the project. It is recommended therefore

Figure 1. Some place names within the system.

that consideration be given this so as to ensure that an adequate description of the system as it existed prior to development is available. Of course this will not ensure that post-development changes will be related in a causal way to the development, for if a measurable change does occur over a decade or two, or longer, other influences may appear by that time to be of equal or greater significance (Anon., 1971a, p45).

It will be evident in the following that the data requirement is rather formidable, not only in oceanography (Hamill, 1969, p37) but also in hydrography and perhaps in geography. Robinson (1968) reviewed knowledge of the geography of the surrounding land areas where two sub-regions could be delineated (p202). One, "the South Coast Lowland", extends southeast of Churchill to west and south of James Bay where along the coast (p217),

> ...there is a flat strip five to ten miles wide, with the widest parts generally being to the north. The coastal zone is treeless, but grass or marshes are common. Storm beaches, a few feet high, are the only topographic features. Tidal flats may be exposed for one to six miles; even at high tide shallow water extends far offshore. Much of the monotonous coast has been unapproachable from the sea even by small ships; this being one of the main reasons for the lack of settlement. Deeper water may be found at the drowned river mouths, but shifting sand bars and minor deltaic deposits are navigation hazards.

The other, the "East Coast Upland" (p222) extends from east of James Bay to the north coast of Hudson Bay a north-south distance of 750 miles. James Bay constitutes almost a third of this, i.e. 230 miles (389 km), and is up to about 100 miles

(185 km) wide with an average depth of about 32 m. It seems likely that given adequate hydrographic data, distinct sub-regions would be recognizable within the bay. The project will likely influence the priority for such data so that the required knowledge of the bathymetry so important to an understanding of the oceanography may become available reasonably soon.

As yet only limited information is available concerning the project beyond that reported in newspapers, eg. Anonymous, 1971b, which in the main have been about the development of hydroelectric potential on the three southern rivers, Broadback, Nottaway and Rupert. A parallel development is proposed for certain of the northern rivers with diversion of water into La Grande-Rivière, including part of Grande-Rivière de la Baleine (Anon., 1971c). Considerable interest in possible long-term climatic effects of the project arose subsequent to newspaper, radio and television commentary based on expressions of two well-known Canadian scientists (R.W. Stewart and L.M. Dickie), which in turn led some to the conclusion that marked change was to be expected, eg. Time Magazine for September 27 (Anon., 1971d). In a subsequent letter to the magazine, Stewart and Dickie (1971) clarified their views somewhat and seemed to suggest that although it was not possible to predict the influence of the project, it was possible to predict that it will have one. Apparently the influence as they view it will be only on the local climate, or microclimate, and could occur in a number of ways, some rather subtle, requiring detailed and extensive data in order to recognize the change and process (Landsberg, 1970, p1269). While it is by no means certain whether man's activities have yet influenced global climate (Frisken, 1971), it is the experience that this has not yet occurred (Landsberg, 1970) and may not be likely (Anon. 1970a, p97).

In the following it is assumed that a major feature of the oceanography of the region, the annual ice cover, is due to the global atmospheric circulation and hence not likely influenced by man. The ice cover limits the exchange of heat with the atmosphere so that both the gain of heat during the open season and the loss of heat during the period of ice cover are much smaller than were the area without an ice cover. It is believed that for the region, i.e. Hudson Bay and James Bay, the gain is about equal to the loss although smaller, so that a small deficit occurs which is balanced by an advection from the ocean. In this regard the main parts of the system may be even more uncoupled from the world ocean than is the Arctic Ocean, where a perennial ice cover is general and the balance is maintained by advection from the world ocean and an excess of export over import of ice. It is believed that the export and import of ice from and to our system are small and about equal, so that the amount of ice which melts is equivalent to that which forms there, although considerable redistribution of the ice occurs within the system.

It is possible that the project could influence this redistribution of ice (Murty, 1971) and some attention is given this aspect later. The treatment generally is quite inadequate and does not provide significant insight to those factors of the circulation which might influence ice distribution. This is of particular significance when we realize that not only is the ice cover unstable but also in some years a portion very nearly survives through a summer, eg. in 1969 (see section 2.3). If the project were to lower the average air temperature in summer through an impact on local climate, the ice cover would persist for a longer time causing a further decline in average temperature. It may be that through observation and study it could be determined whether the project will constitute a "major environmental disturbance" (Kershaw, 1971) or whether our considerations which suggest little impact are, in fact, "miscalculations" (Heyneman, 1971).

1.1 Climate

> In contrast to the broad expanses of Arctic tundra that surround Hudson Bay the sub-Arctic lands bordering James Bay are partially forested and thus protected from strong winds. As a result two of the most distinguishing features of Hudson Bay winter climate - wind chill and blowing snow - are not nearly as evident near James Bay. With this important exception and the fact that James Bay is several degrees of latitude farther south the factors that influence the climate of James Bay are essentially the same as those outlined for Hudson Bay.
>
> Along southern coastlines more than 100 inches of winter snowfall are evenly distributed during the months November through March. Fort George, on the east coast, receives almost as much snow but approximately one half of the annual total falls during November and December. Lesser amounts are measured in the mid-winter months

when almost solid ice cover over James Bay cuts off the supply of moisture to the air moving over it. Throughout the winter the snow is soft and deep in the woods but drifted and hard packed over the exposed Bay ice.

In general the lands bordering James Bay are free of snow from late in May until the middle of October. During this period precipitation, mostly in the form of rain, accounts for more than one-half of the 25 to 30-inch annual total. At Moosonee, in fact, rain or drizzle are reported on approximately 15 days of each month in summer. As would be expected from these figures, cloudy days are frequent. While Moosonee and Fort George receive less than 240 hours of bright sunshine in July, stations at approximately the same latitudes in the prairie provinces record close to 320 hours. Moosonee ranks above Churchill in the number of thunderstorms that occur in summer with an average of three or four per month. The frequency of thunderstorms decreases markedly to the north and east however, where the cold waters of Hudson and James Bays tend to inhibit shower development.

Although reports from the weather stations at Moosonee and Fort George do not fully substantiate it, fog is probably quite prevalent over James Bay in June and July when sea ice is still present.

As is the case with Hudson Bay, the cold waters are more influential than latitude in determining summer air temperatures along the shores of James Bay. July temperatures at Moosonee and Moose Factory average about 60°F while daily maxima are near 70°F. Corresponding temperatures for Fort George are about 5 degrees lower. Along the east coast of James Bay daytime high temperatures over 85°F are unusual. At Moosonee, on the other hand, where southwesterly winds bring warm air from the heart of the continent, 90°F temperatures have been recorded in all months from May to August. Sharp and rather frequent temperature fluctuations may be expected during these months depending on whether the winds are off the land or off the water.

Freezing temperatures have been reported in every month of the year at weather stations on the shores of James Bay. While there are wide local variations in the incidence of frost, depending on the proximity of the Bay or on the presence of lakes or muskeg, the first frosts of autumn usually occur early in September. Average daily temperatures generally remain above 32°F, however, until late October.

> January is the coldest month in the James Bay region with average temperatures in the neighbourhood of -5° to -10°F. An extreme minimum temperature of -52°F has been recorded at Moosonee and temperatures of -40°F are not uncommon. Readings as low as -20°F may be expected on one day in four at the height of the winter season. In contrast to the Hudson Bay area where above freezing temperatures are rare in winter, James Bay has the occasional mild spell in January and February. An example of such mild conditions occurred in February 1954, when maximum temperatures at Moosonee exceeded 40°F on five successive days, culminating in a reading of 51°F on the 19th.
>
> In summary, the climate of James Bay is not as severe as that of Hudson Bay. Winters are long and cold however, and summers cool, and on an annual basis the lands surrounding James Bay are colder than most in Canada at similar latitudes.

That the "lands surrounding James Bay are colder than most in Canada at similar latitude" is probably due to the "deep southward penetration of Arctic climate in eastern Canada in winter" caused by alteration of "global wind patterns" by geographic features. The exerpt and the quotes are from Thompson (1968) who also provided climatological data for a number of locations around Hudson Bay and James Bay. Stressed in the article (p267) is the effect of ice and relatively cold water on the local climate whereby winter conditions become quite continental and those of summer quite maritime. Earlier Burbidge (1951) discussed the influence of the open water of Hudson Bay on continental polar air moving over the region. Apparently the influence is greatest over the eastern portion in mid-summer and in early winter prior to freeze-up; although the tempering influence does persist past freeze-up into January (Hagglund and Thompson, 1964). It is clear that a persistent change in the average extent of ice cover would be followed by a change in climate there. As noted, it is generally believed that ice extent in the northern hemisphere is determined largely by global atmospheric patterns (see also Fletcher, 1969); however, in Hudson Bay the persistence of ice is particularly sensitive to warm air from the south (Mackay, 1952) or, as we shall see, to cold air from the north.

1.2 Ice cover and a possible consequence

The "deep southward penetration of Arctic climate into eastern Canada in winter" is such that ice thickness over much of Hudson Bay approaches that for Arctic regions generally, although for James Bay it is considerably less (for example see Bilello and Bates, 1971).

> The greater part of James Bay is frozen over in the winter. By the end of January, local inhabitants have been known to go across the ice from the eastern shore to almost every island lying in the middle of the bay. At the head of the bay, in mid-winter, frequent journeys are made across the ice in a direct line between Charlton Island (Lat. 52°00'N., Long. 79°25'W.) and Moose River and between Charlton Island and Rupert River. The movement of ice in spring is greatly affected by the direction of the wind. If southerly winds predominate, the wind and outflow of the rivers acting together clear the southern part of the bay by the middle of June. But if northerly winds predominate, the ice will remain in the bay and seriously obstruct navigation till the sun becomes strong enough to melt it.
>
> Coasting vessels are put into commission during the last days in June, and make their passage from one post to another in lanes of open water between the coast and the ice in the centre of the bay. (Anon., 1965, p441).

Larnder (1968) stressed the variability of the ice cover in Hudson Bay, both in its formation and breakup, which she noted differed "widely from year to year and from one locality to another" (p319) and, "Although, as already noted, the ice in James Bay generally melts or moves out of the southern two-thirds of the bay by early July, the ice that moves into its northern part from Hudson Bay may be so heavy, close-packed and persistent" (p335) that local transportation may be limited. Danielson (1971, p102) also considered that the northern "parts of James Bay receive ice from Hudson Bay" and implied that like southwest Hudson Bay much more ice melts there than forms there (p98). From the charts of ice conditions provided annually by the Meteorological Service it is possible to obtain an appreciation of the extent to which breakup can vary. For example, observations on July 9 of each of 1968 and 1969 (Figure 2) indicated quite a wide divergence in extent of ice cover. This may be due to variations in direction of persistent winds early in the summer which can accumulate or disperse ice. This in turn can influence the surface albedo and hence the amount of insolation eventually absorbed. A pattern of circulation associated with mixed freshwater moving seaward at the surface probably also influences the distribution of ice, but not likely to the same extent as can this early wind.

As James Bay is a region of annual, rather than perennial ice cover and as open water generally occurs at a relatively early time in summer, a seasonal variation in temperature and salinity may be anticipated. In the approach to James Bay the annual sequence of events is, in the main features, likely similar to that suggested

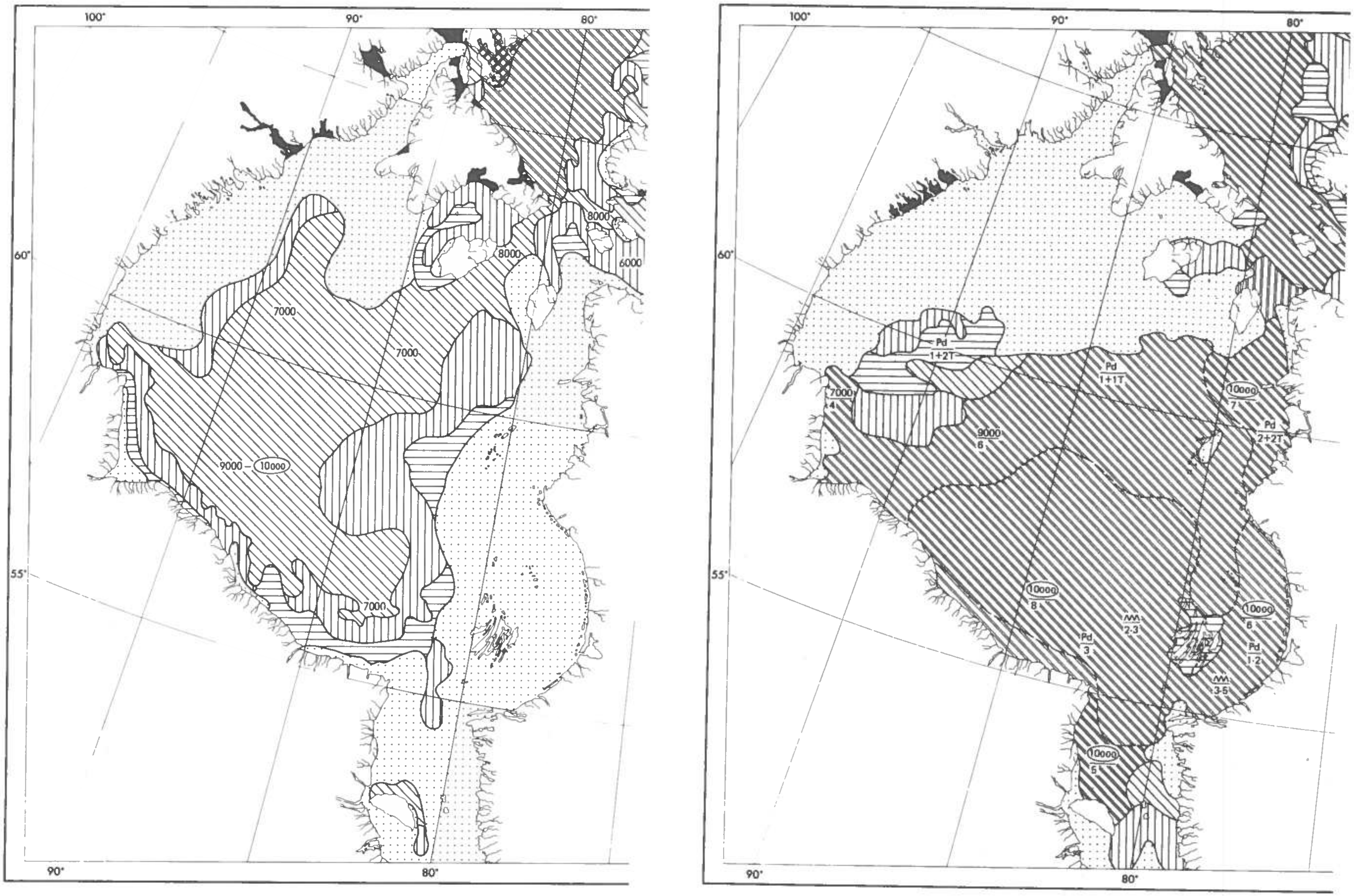

Figure 2. The extent of ice cover on July 9 of 1968 and 1969 (from Anon., 1970e pl3; 1971e pl3). (a) 1968. (b) 1969.

for a more northerly location in Hudson Bay (Figure 3), where the variation of salinity within a perennial surface layer is largely related to change in the ice cover. Within James Bay, the salinity structure is likely strongly influenced by the amount of freshwater from runoff and in the southern portion it may be that in winter a layer of freshwater exists under the ice (see also section 2.2.2). The existence of such a layer would be a consequence of the marked decrease of mixing at the surface due to a reduction in the transport of energy, from wind and exchange processes, across the sea surface resulting from the existence of an ice cover. Tidal currents would provide some energy for mixing but as these are not likely much above about 1 knot away from shore (Godin, 1971), it is possible the layer persists. Thus, in the absence of observations, the possible late winter distribution of salinity and temperature (Table 1) for the deep water west of Trodely Island is highly speculative, perhaps even fanciful. The small maximum temperature in a halocline is characteristic of arctic seas, as is the halocline itself. It seems almost certain that the halocline exists in winter, so that the water below a surface layer may be effectively uncoupled from local surface processes, i.e. convective processes are limited to depths above the halocline. This situation is believed to exist throughout the year over most of James Bay, so that the deeper water probably is from Hudson Bay with relatively little change.

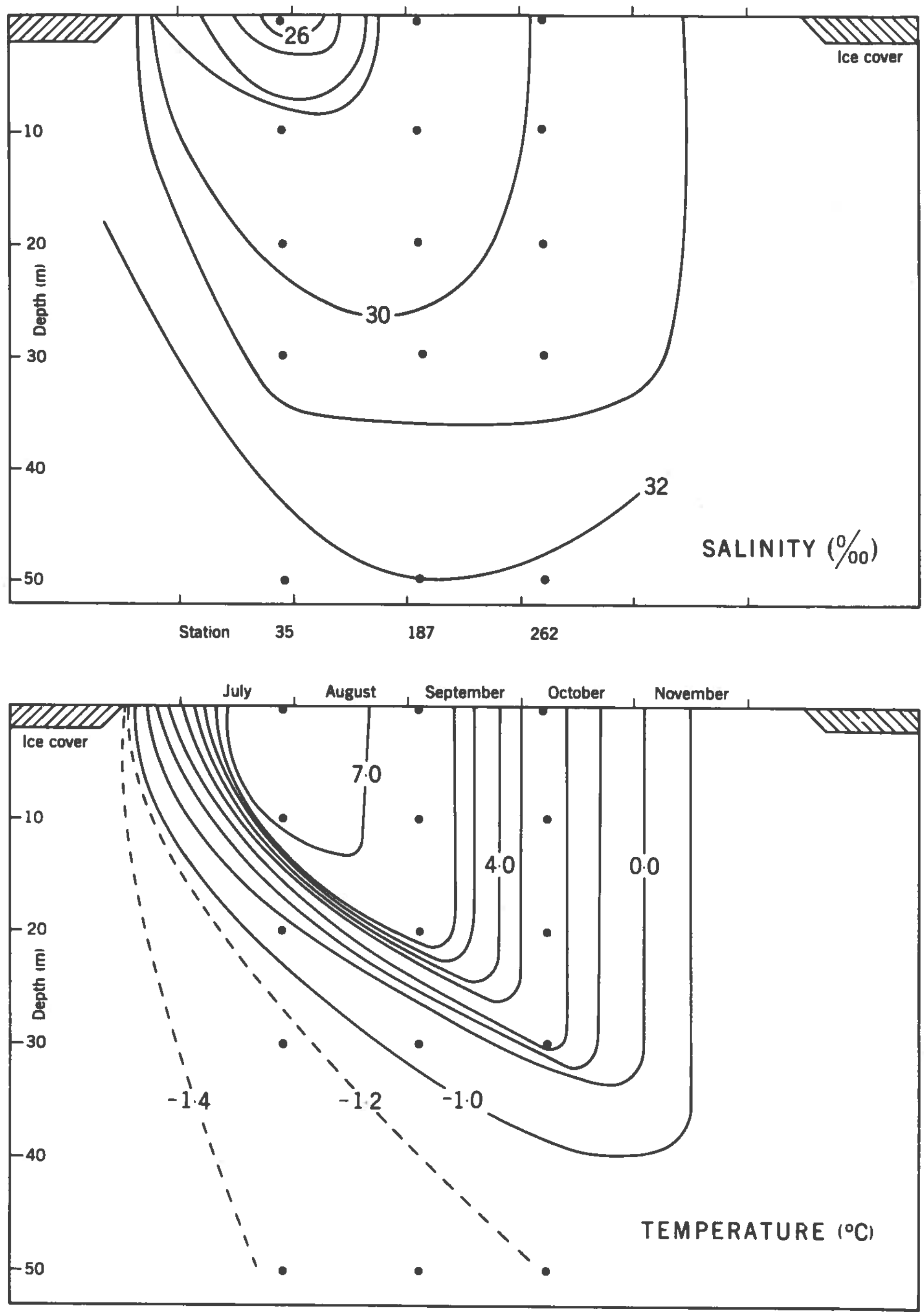

Figure 3. The annual sequence in the near-surface at a location in Hudson Bay (from Barber, 1967, p29). (a) Salinity. (b) Temperature.

Table 1 A possible distribution of salinity and temperature under the ice cover at a point just west of Trodely Island.

Depth (m)	Temp. (°C)	Sal. (°/oo)
2	0.0	0
4	0.0	0
10	0.4	15
20	-0.8	25
30	-1.3	27
50	-1.3	27
75	-1.3	27

1.3 Bathymetry

Most knowledge of the bathymetry (frontispiece) is contained in Canadian Hydrographic Service chart number 5800 and pilotage information is available in another publication of the Service (Anon., 1965) from which the following was extracted (p441):

> The low, flat shores of the bay are fringed with wide mud flats. The bay is filled with numerous islands, rocks and shoals, with no harbours for large vessels. In the fairway which, in the outer part, lies close westward of the middle of the bay, there are general depths of from 10 to 40 fathoms ($18^{m}3$ to $73^{m}2$) to within about 20 miles of the head of the bay.
>
> As in the case of Hudson Bay itself, the eastern shore of James Bay is very irregular in outline, with many islets and rocks lying close off it, while the western shore is even and almost free of islands. There is, however, Akimiski Island lying close off the middle of the western shore. In the eastern half of the bay are many islands and shoals. The western shore of the bay affords no harbours, but there are several fairly good harbours on the eastern shore for vessels drawing up to 12 feet ($3^{m}7$) of water.

Limiting depths exist in certain parts of the bay so that the water in some areas may be isolated from that at similar depth in Hudson Bay. For example, it is thought that for a station occupied in the deep water off

Trodely Island a limiting depth of 40 m exists so that a salinity of about 27 ‰ could be expected there. The actual situation is quite uncertain and it is possible to interpret available bathymetric data as though there were shallows, less than 20m, extending continuously off the east shore from Cape Jones (Pte. Louis XIV) to Bare Island southward through Grey Goose Island, the Twin Islands, Weston and Charlton Islands to the east coast in the vicinity of Rupert Bay, so that a depth of 60 m would be isolated to the eastward. However, the interpretation here (frontispiece) indicates just one such area eastward of the shallows between Grey Goose Island and the Twin Islands, but with deeper areas south of Bare Island and north of Weston Island. Depths of 80 m occur almost as far south as Trodely Island; a depth which it was noted is not likely continuous to Hudson Bay. The areas of limiting depth of about 40 m are believed to exist between Bear Island and Akimiski Island and to the east of South Twin Island.

With regard to the water which may occur in the northern approach to James Bay, oceanographic data indicate that a limiting depth of between 50 to 75 m occurs between the Belcher Islands and Cape Henrietta Maria (Barber, 1967, p15), thus the deeper water north of Pte. Louis XIV would be slightly isolated from Hudson Bay (an interpretation of more recent, but still incomplete, data on Canadian Hydrographic Service field sheet 025A suggests a limiting depth of 80 m) so that the range of salinity observed at depth in the approach, i.e. 31.7 to 32.0 $^{o}/_{oo}$, is somewhat less than is observed in the open areas of Hudson Bay generally, i.e. up to 33.0 $^{o}/_{oo}$ at 100 m (Figure 4).

1.4 Surface wind

Knowledge of surface wind within the system is based largely on observations at coastal stations, which indicate that winds are strong in all but the summer months when they are:

> ...generally lighter and are variable in direction with a higher proportion of onshore winds at coastal stations, the effect of local sea-breeze circulations.
>
> (Anon., 1965, p395).

Danielson (1971, p97) remarked on the influence of persistent north winds which move ice southward and evidence provided by Archibald (1969) suggests that wind strength may decrease from north to south, being less in southern James Bay than elsewhere in the system. It is the author's (limited) experience that storms can occur for short periods at any time during the summer with winds from the southeast to southwest. At Churchill in late September strong northwest winds may persist for 4 to 5 days and in early October in 1961 in western Hudson Strait strong wind (290°, 50 knots) was experienced. Occasionally it is possible to see in the oceanographic distributions, correlations with surface wind. Of particular interest here is the influence of wind during the summer, i.e. when it is weak and variable with a "high proportion of onshore winds". As cloud and fog are frequent over the water such

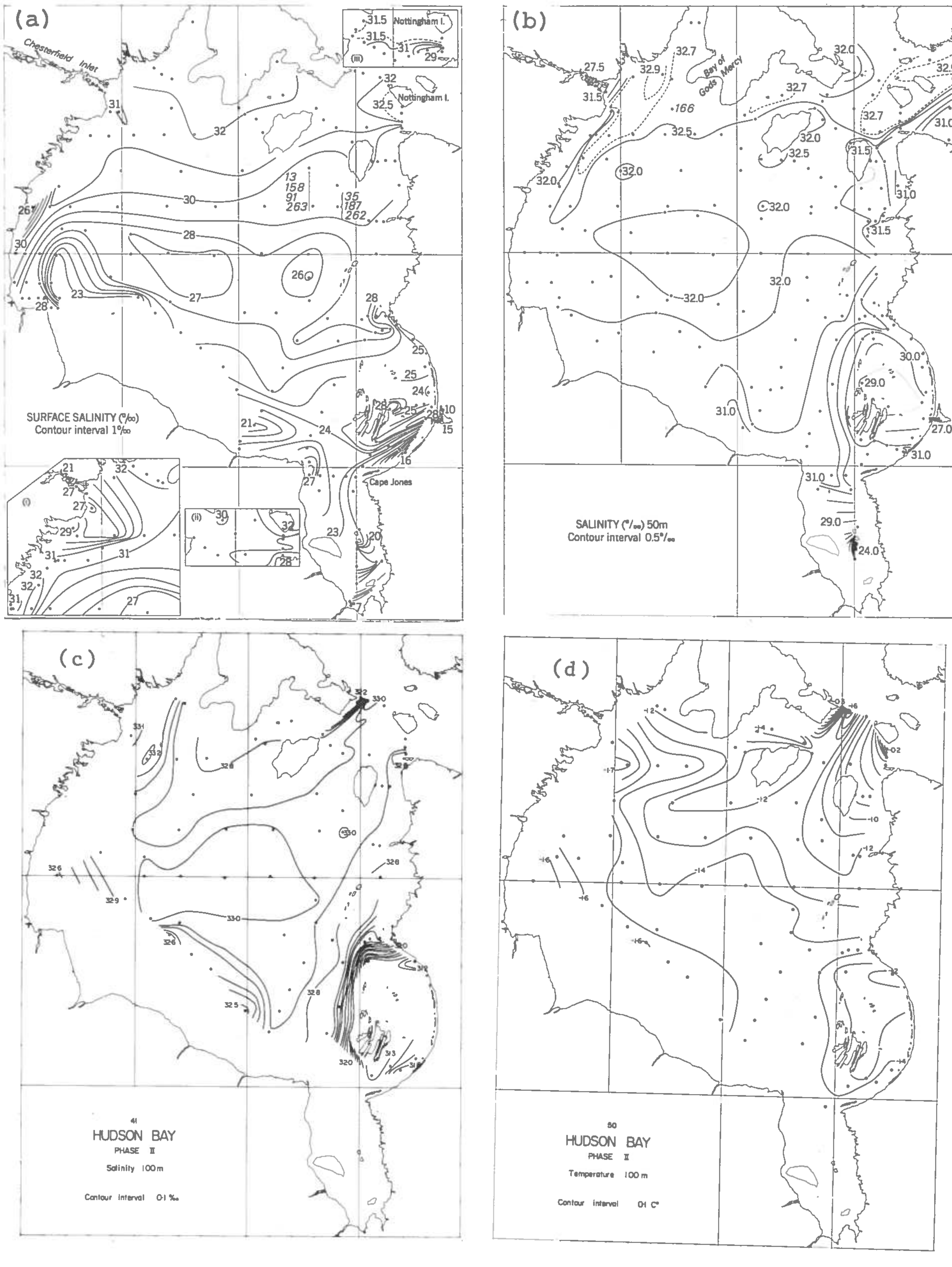

Figure 4. The distribution of salinity at the surface (a) and 50 m (b) in Hudson Bay and James Bay (from Barber, 1967) and the distribution in Hudson Bay of salinity (c) and temperature (d) at 100 m (from Barber and Glennie, 1964).

a wind would tend to increase the frequency of cloud and fog at coastal stations; an increase in each has been observed in both Hudson Bay and James Bay (Thompson, 1968). However, the main movement of air at the surface is from west to east so that a relatively warm air mass is moved over the southern-most parts of the system. It is useful to consider the likely surface temperature toward the end of July based on the 1961 experience (Figure 5) by which time considerable warming of the water had occurred over northern Hudson Bay, but with little or none in the southern part due to the ice cover there. Some warming has also occurred on land by this time, for example to the southwest (Figure 6). A warm air moving west to east from the land over the ice covered area would likely become relatively stable with which would be associated a decrease in surface wind. The implication is that this small-scale region of atmospheric stability through a decrease of surface wind, increase of cloud and fog and with the high albedo of ice cover, contributes to the stagnation of the ice in the southwest usually observed (Anon., 1970e). Also implied is a lack of water movement due to other factors; however, this may result as a consequence of the James Bay circulation (Murty, 1971).

1.5 Freshwater from runoff

It seems likely that the pattern of runoff to James Bay will continue to be influenced by man, either through the increasing development of hydroelectric potential as has already occurred to some extent on the

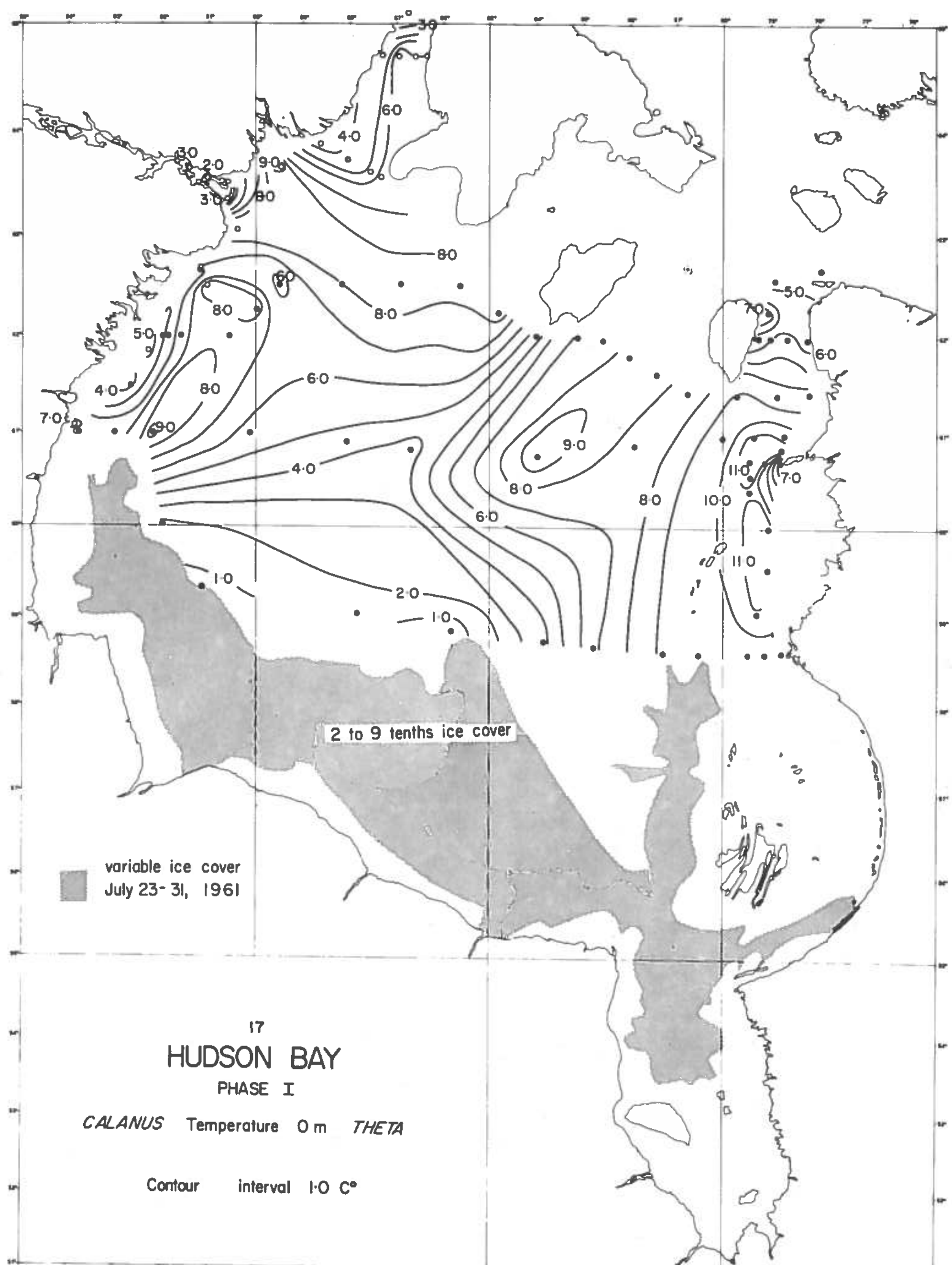

Figure 5. The likely distribution of surface temperature in Hudson Bay toward the end of July based on temperature data (Barber and Glennie, 1964 Figure 17) and ice data (Anon., 1962) in 1961.

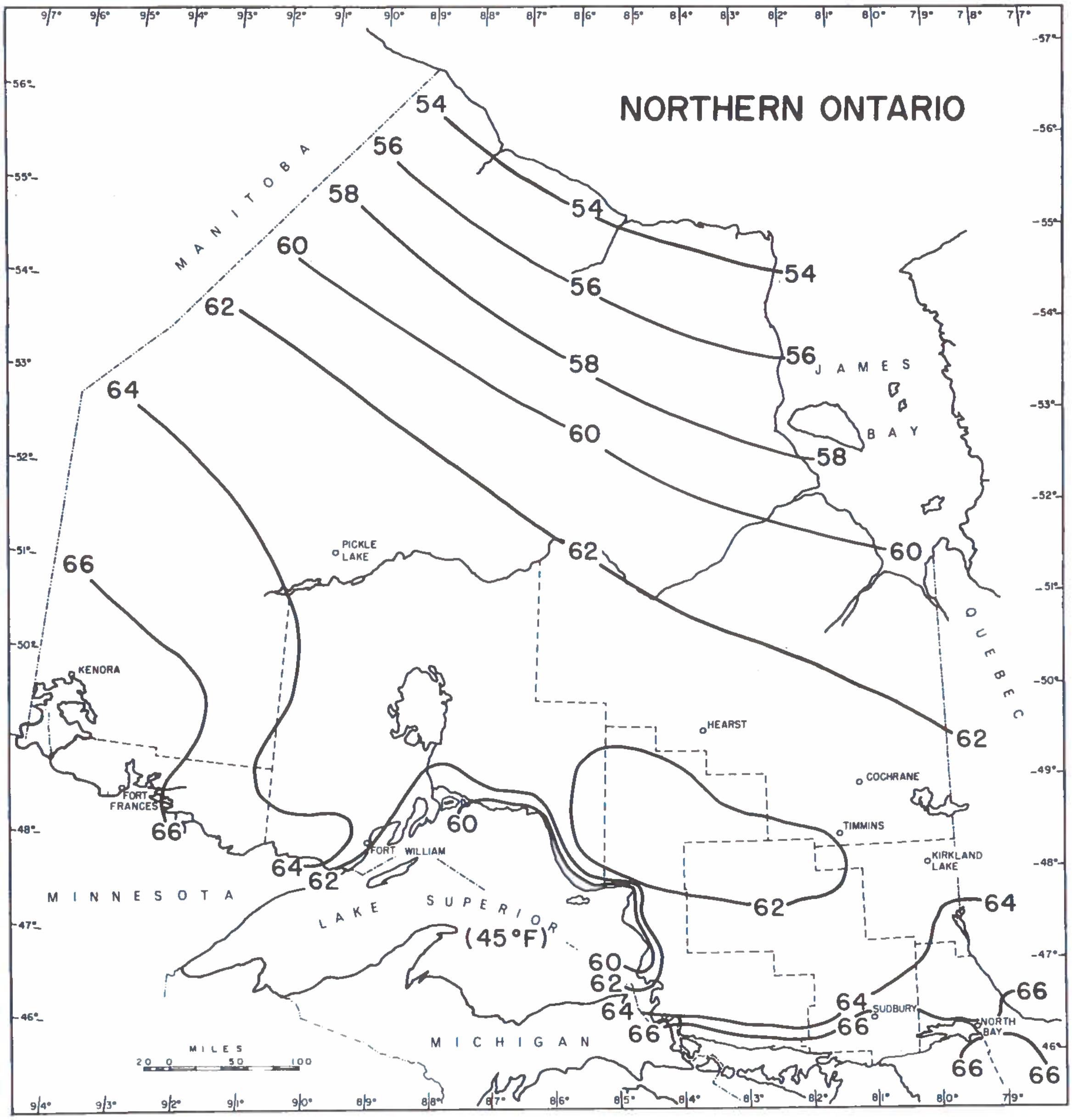

Figure 6. Mean daily temperatures (°F) in northern Ontario for July (Chapman and Thomas, 1968 their figure 8).

Moose River (Robinson, 1968, p219), or perhaps through diversion of water southward (p219). At the present time the annual runoff volume appears to be about $3 \times 10^{11} m^3$, that is if there were no diversions. The value was obtained by extrapolation of the data on runoff from the Québec rivers (Figure 7) over the total drainage basin including James Bay. A check was obtained through the estimation of a value for the annual excess of precipitation over evaporation (P-E) over the total region. Hydroelectric development, in the absence of diversion, would not be expected to influence the annual value appreciably (except that it modified P-E), but would markedly influence the present pattern of runoff (Figure 8), which varies from very high levels in spring to extreme low levels in March and occasionally in summer. Black remarked (1968, p843):

> The rivers become quite shallow by July, even during wet weather. At this time the low water exposes numerous bars, bouldery shoals and rocky outcrops which render stretches of river difficult to navigate, ... A short period of navigation is possible before freeze-up when a second high-water period occurs during heavy autumn rains, ...

Presumably the rivers are not used for navigation during the period of low runoff, i.e. in March, at which time the flow can vary between a fifth to more than an eighth of the much more variable spring volumes. No doubt this large spring runoff has a significant influence on conditions in James Bay, including the distribution of ice, salinity and currents, particularly as it occurs when the ice cover is still extensive. However, even though

Table 2 A tabulation of 1968 surface water data from six stations indicating the name, size of basin, mean flow and the page reference to the source (Anon., 1971f). Other data are available (Anon., 1967a; 1970b; c; d).

Name (St. No.)	Area (Sq. Miles)	Mean Flow (cfs)	Page
La Grande-Rivière (092704)	37,600	60,000	363
Eastmain (090601)	17,100	31,400	352
Grande-Rivière de la Baleine (093803)	16,500	23,700	379
sub-total	71,200	115,100	
Rupert (081002)	15,800	30,300	345
Broadback (080801)	6,610	11,500	343
Nottaway (080701)	22,200	36,600	333
sub-total	44,610	78,400	
TOTAL	115,810	193,500	

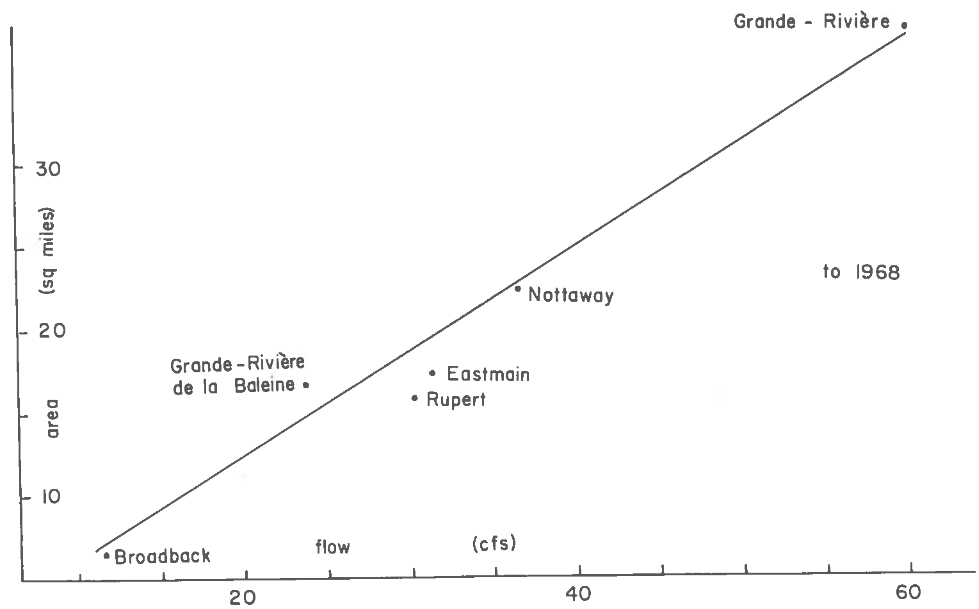

Figure 7. A presentation of the data of Table 2. The extrapolation referred to in the text is based on the "best fit" of the straight line.

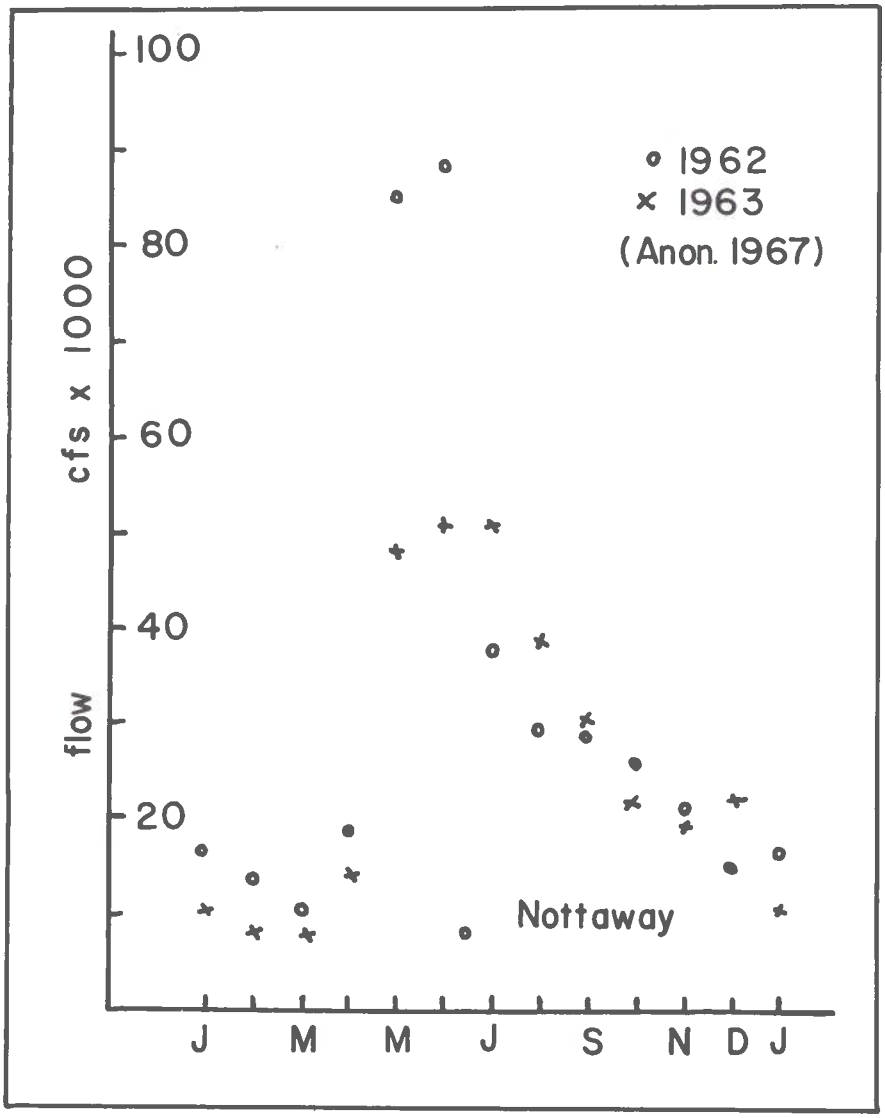

Figure 8. A presentation of the monthly values of runoff for the Nottaway River for 1962 and 1963 (Anon., 1967a).

data on the flow from the Québec rivers are quite adequate, lack of oceanographic data limit their application. For the purpose here the significant annual values appear to be:

1)	amount from rivers and excess of P-E over the bay	$3x10^{11}m^3$
	but not including Grande-Rivière de la Baleine	$0.02x10^{11}m^3$
2)	amount from rivers to be developed	$1.88x10^{11}m^3$
	i.e. the sum of the southern rivers	$0.78x10^{11}m^3$
	and of the northern rivers	$1.1x10^{11}m^3$

Thus the development will increase the total annual value slightly and markedly influence the pattern for about half the runoff to the bay. With the potential for development of the southern and west coast rivers the runoff to the bay could be almost completely smoothed or at least reflect a demand for energy, which may be maximum in winter, i.e. at a time when the runoff prior to development, was at a minimum.

2. The system

2.1 Data in James Bay

It is curious that James Bay has not been examined by oceanographers to nearly the same extent as have areas much further north. Apparently it is considered a difficult area, particularly for vessels suited to ocean survey, so that during the development of the 1961 oceanographic programme for Hudson Bay the work in James Bay was limited to two sections across the entrance. Earlier, the negative results of an investigation into the fisheries potential of Hudson Bay by Hachey (1931a) considerably deflated interest in the region generally (Hunter, 1968, p373). The 1961 material and the 1959 work of "Calanus" (Table 3) constitute the only significant data yet observed there, although temperature and salinity data were observed in the Moose River with tidal observations (Langford, 1963, p90), but these have not been located. All of the available material were observed during the period of open water, though some of the "Calanus" stations were occupied at a time (Table 4) when some ice cover occurred (see data report for existence of ice nearby at the time the station was occupied). Distributions based on the "Calanus" data were presented by Grainger (1960) and these were also utilized in a description of Hudson Bay (Barber, 1967; 1968a). The recent compilation of information about the region (Beals, 1968) proved useful in a number of ways, for it is "an impressive assemblage of facts not readily found elsewhere" (Jackson, 1970, p841).

2.2 Review

James Bay is part of a general system (Figure 1), comprising it, Hudson Bay and Foxe Basin, connected to the world ocean through Fury and Hecla Strait and Hudson Strait. Fury and Hecla Strait is likely much less important than is Hudson Strait to the character of the water within the system, for not only is Hudson Strait relatively wide

Table 3 A tabulation of the data available indicating the year obtained, name of ship, cruise reference number and a reference.

Year	Ship	CRN	Reference
1958	"Calanus"	320	Grainger, 1960
1959	do	321	do
1961	"Theta"	337	(Anon., 1964a; b)

Table 4 A tabulation of the "Calanus" 1959 data in James Bay indicating the station number and day. The latitude of station 59-4 has been taken as 52°30' and that of 59-15 to be 53°56'.

Number	Day	Number	Day
1	June 20	57	Aug. 26
2	22	58	26
4	22	57*	29
6	23	59	29
7	25	60	29
8	26	61	30
9	27	62	30
11	30	63	30
13	July 1		
15	10		

*Apparently a re-occupation.

and deep, but also strong mixing due to tides occurs there. Thus the water reflects to large extent an influence of the Atlantic Ocean and an influence of a process occurring within the system, and even though James Bay is 600 to 1,000 miles away from these influences they are still strongly evident there. In the present context however, distance need not be particularly significant, for factors such as the general location and the associated climate and pattern of runoff in such a uniquely shallow region as James Bay can be paramount.

The foregoing assumes certain physical properties of water which are of prime significance. For example, at salinities generally encountered in the ocean a temperature of maximum density does not exist at any temperature warmer than the freezing point (Figure 9), as exists for freshwater at 4°C. It will be shown that much of the water is less than oceanic salinity, specifically less than 24.7 $^{o}/oo$, so that a point of maximum density can occur at temperatures warmer than that of ice formation at the surface. At such low salinities, surface cooling at temperatures close to freezing would lead to stability. Furthermore, at those locations where the surface layer is fresh or nearly so, the amount of salt made available to the surface layer through freezing is negligible. Thus, the evolution of a surface mixed layer through the winter would not relate to the formation of denser water at the ice-water interface (Barber, 1968b). Also of utmost significance, here as elsewhere, is the

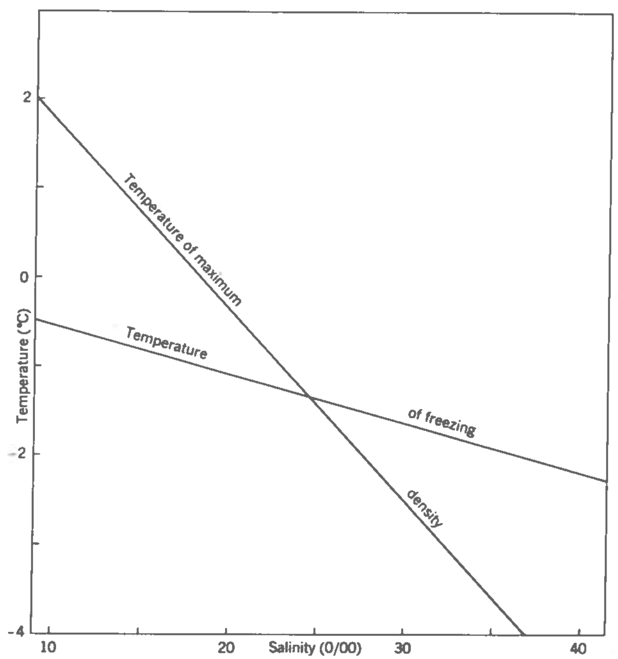

Figure 9. Temperature of the freezing point and temperature of the maximum density as a function of salinity (from Barber, 1967b).

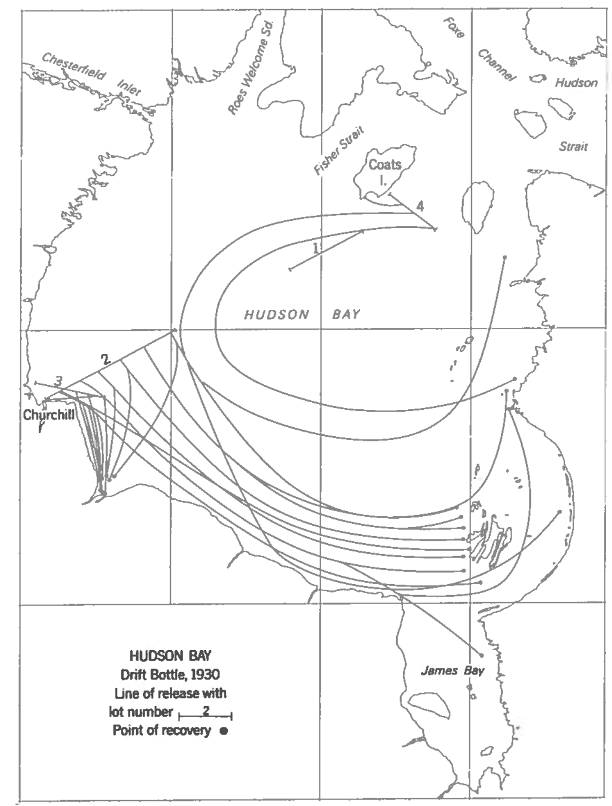

Figure 10. A re-presentation of the drift bottle data of Hachey (1935), (from Barber, 1967).

fact that ice formed at the surface remains there, as it is less dense than water, providing a very different boundary than is the air-sea interface.

2.2.1 Currents

It may be significant that only one of the drift bottles released in 1930 (Hachey, 1935) was returned from James Bay (Figure 10). If it is significant, then it is envisaged that the inflow from Hudson Bay occurs at depth rather than the surface and comprises a cold, relatively saline water. On the other hand, Grainger (1960) inferred an anti-clockwise surface movement with water entering from Hudson Bay along the west coast. Data observed in the autumn (October, 1961) suggest the development of a well-defined movement from James Bay north along the east coast of Hudson Bay and to considerable depth (for example see salinity at 100 m, Figure 72 of Barber and Glennie, 1964), which was partly due to an increase in the amount of freshwater there in October over August (Figure 11). Thus the pattern of circulation in these areas may exhibit a time-dependence in association with the distribution of freshwater from runoff.

Presumably a portion of this runoff would have absorbed a significant amount of shortwave radiation and would enter the bay at a temperature above that of the surface water there, so that it would constitute a gain of sensible heat. Presumably too, much of the outflow at the surface would have, in the absence of ice, absorbed heat and would constitute a loss. The heat transported

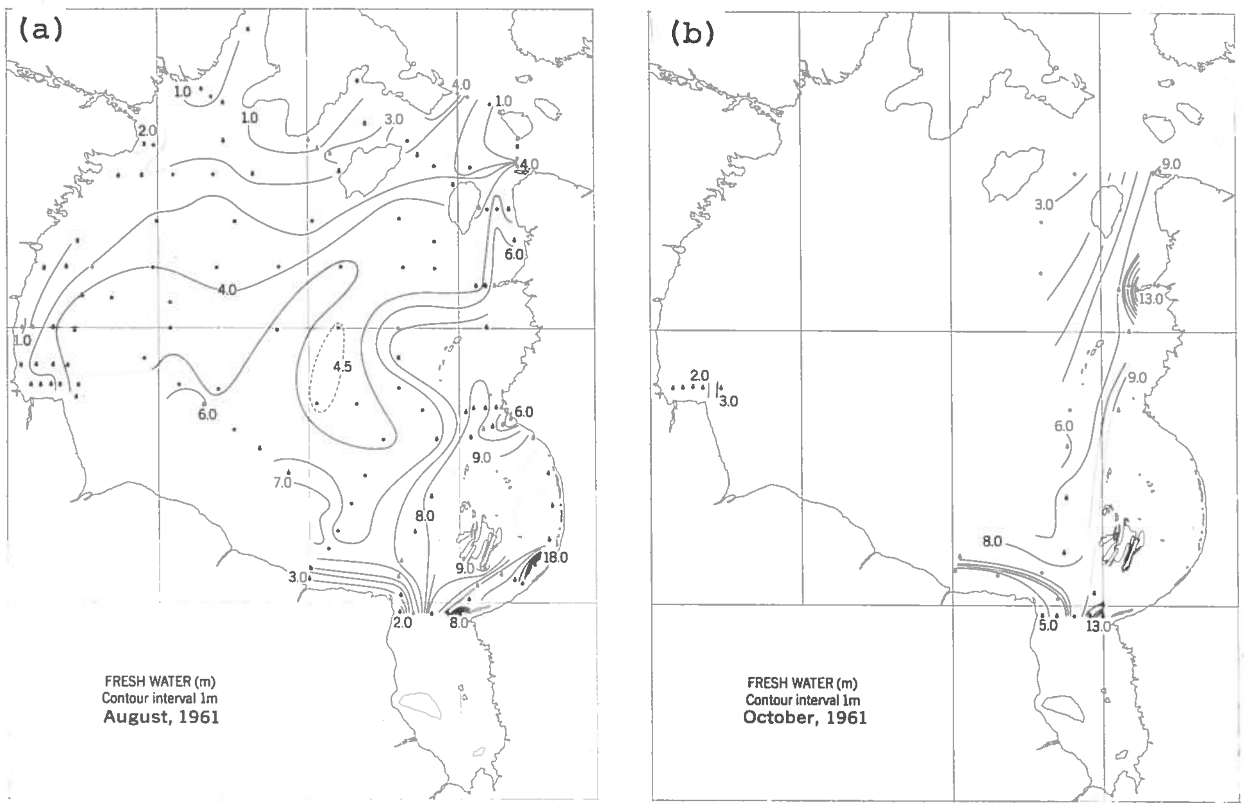

Figure 11. The distribution of the depth of freshwater (m) in August and October, 1961 (from Barber, 1967). (a) August. (b) October.

with the inflowing deep water would be very much less so that in the section across the entrance a loss of heat occurs due to transport. The loss may be balanced by an excess of export over import of ice, but there is no direct evidence that either occurs, and it may be that it is balanced by the input of heat with the runoff. In this circumstance the annual sum of the flux terms would on the average be zero. Danielson (1969, p162) considered that a variation in the flux occurred over the bay ranging from a deficit of about 100 g cal cm^{-2} $year^{-1}$ in the north to about 6000 in the south and leading to an average gain of about 2500 g cal cm^{-2} $year^{-1}$. A rough heat budget for a position in the northern part of the bay was attempted and although a value which might be considered better than Danielson's was not achieved, a comparison of the actual with the potential heat storage for two periods was made from existing data. Between August 11 and 29 (1959) an increase in heat storage of 6500 g cal cm^{-2} occurred, which is about twice the amount to be expected in the absence of advection. This suggests a movement out of the bay of the warmed water. During the period August 29 to October 2 the mixed layer depth increased to 25 m and the heat storage was unchanged. This would have been expected in the absence of advection. The data therefore suggest a variation in a movement of surface water from the bay, again suggesting a time-dependence in the outflow.

Such a time-dependence could be expected to have a distinct influence on the pattern of surface movement seaward of James Bay, but direct evidence is not available. Some secondary evidence has been mentioned, for example drift bottle data and the distribution of freshwater, but there is one other which should be noted even though quite speculative. This relates to the distribution of ice within Hudson Bay during the summer which exhibits lingering ice, or a pattern of last ice, in the southwest (Anon., 1970e; Danielson, 1971) due in part to movement of ice from the north during the early part of the breakup period. Later the ice appears to stagnate in the southwest with little or no tendency to move and, in particular, no tendency to move with anti-clockwise movement around the coasts presumed to exist.

2.2.2 Salinity

Various distributions based mainly on the 1959 data are included in the final group of figures (Figures 21 to 24). Generally, the material was difficult to interpret and in some distributions certain of the data are not included.

The distribution of salinity at the surface and at 50 m (Figure 4) within Hudson Bay and James Bay is based on both the 1959 and 1961 data. Surface salinity was generally low and ranged from 32.5 ‰ in northernmost Hudson Bay, to 27 ‰ in the approach to James Bay and to less than 10 ‰ at the head of the bay, most of the decrease occurring in the southern part. Salinity values in excess of 31 ‰ were observed at 50 m in the approach to the bay and a marked north-south gradient extended well into the bay.

The 1959 data are utilized in the illustration of the longitudinal distribution of surface and bottom salinity (Figure 12) which emphasizes both the strong gradient at the surface toward the head and the marked vertical stratification or structure. It seems likely that stratification is a permanent feature of the oceanography of the bay; a feature which may become more pronounced seaward of the major rivers in winter to the extent that freshwater occurs under the ice. Such a circumstance has been observed in Kugmallit Bay (Barber, 1968b) just east of the delta of the Mackenzie River and in Tuktoyaktuk Harbour. The spatial distribution of the freshwater layer seaward of Kugmallit Bay was not observed, but it was visualized that the freshwater moved eastward close to the coast within a surface layer and with little mixing with deeper salt water. As entrainment of salt was minimal the normal estuarine type of circulation did not exist and it was possible to discern an exchange process due to tides. In James Bay, however, the rms value of the tidal current appears somewhat larger than in Kugmallit Bay so that tidal mixing may be considerable. This would tend to remove stratification so that the layer of freshwater suggested here may exist only in the immediate vicinity of the various estuaries. Nevertheless, in order to emphasize the lack of data and the considerations in

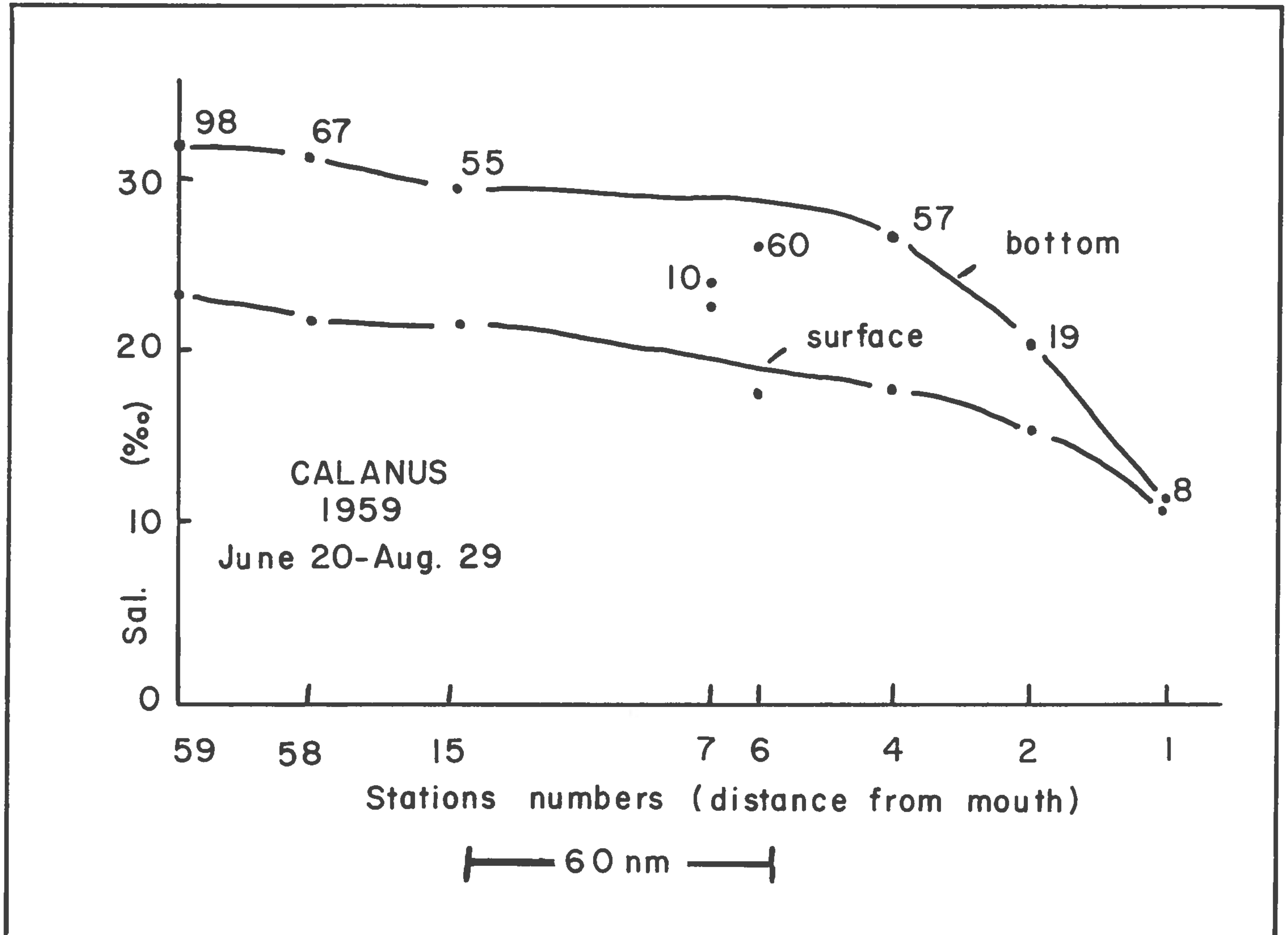

Figure 12. A longitudinal distribution of surface and bottom salinity based on "Calanus" data during the period June 2 to August 29, 1959. The depth in metres of the near-bottom sample at each station is indicated.

other sections about the coupling to Hudson Bay, the layer is shown (Table 1) as extending to Trodely Island.

It was suggested earlier (section 1.2) that in the approach to James Bay a surface layer probably existed within which seasonal changes of temperature and salinity could be recognized; the salinity change was due to variations in the annual ice cover. Observations at about one location in a section across the mouth of the bay (Figure 13) support the view that such a change occurs, but the extent to which it reflects the influence of the annual ice cover as opposed to annual runoff is not known, expect that at positions in the section close to the west and east coasts, a variation in salinity occurred which is believed due to runoff (Figure 24).

2.2.3 Secchi disk depth

The distribution of Secchi disk depth in James Bay in 1959 and Hudson Bay in 1961 during the period of navigation of those years is shown in Figure 14. Least values occurred in James Bay where depths less than 2 m were observed in the south and where in general the depth appears to have been less than 5 m. In the approach to James Bay values to 10 and 15 m occurred. Over much of Hudson Bay readings were close to 15 m with a significant number to 20 m and an occasional (2) reading to 25 m.

The relatively low values in James Bay may be due to a sediment load associated with the large inflow of freshwater from runoff, although other factors could be of equal or greater significance. Langford (1963)

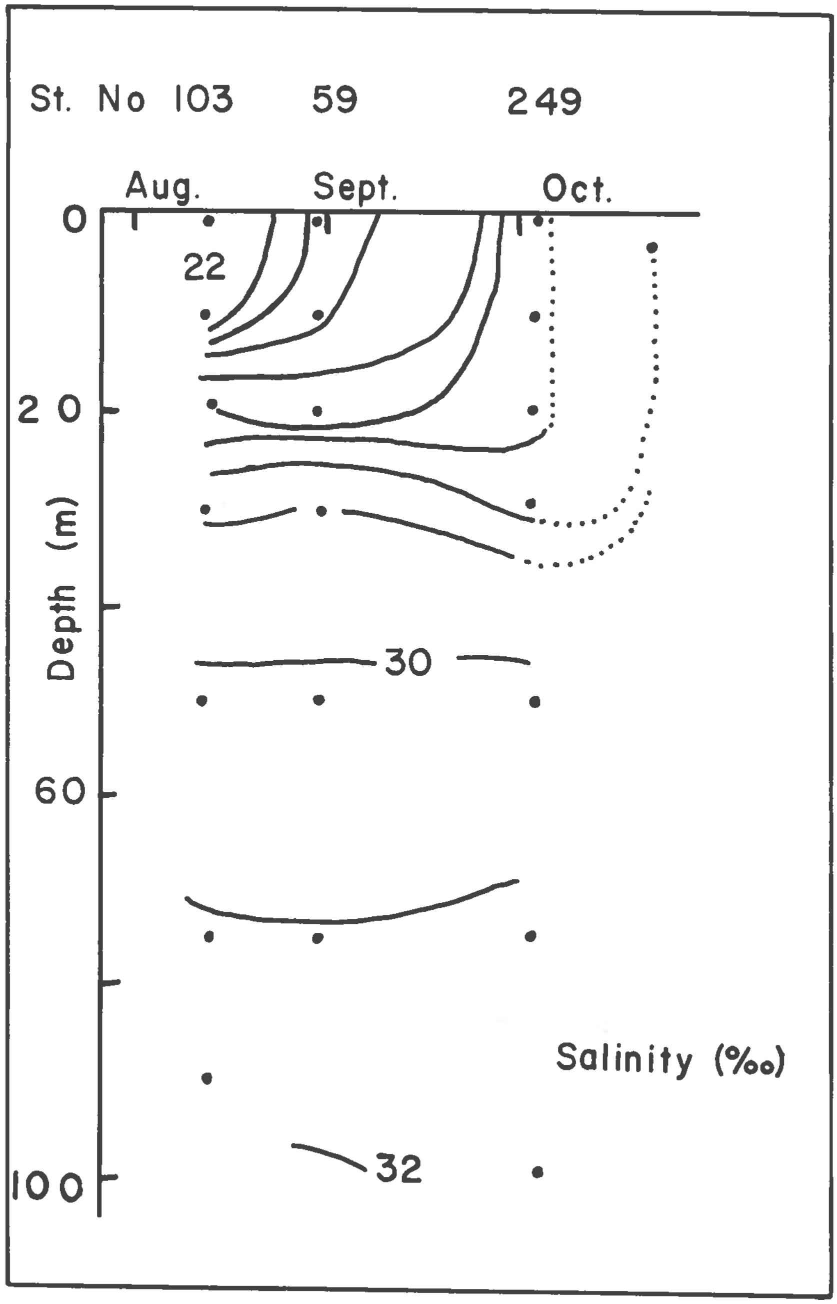

Figure 13. The distribution of salinity at about one location in the mouth of James Bay derived from "Calanus" station 59 in 1959 and "Theta" stations 103 and 249 in 1961.

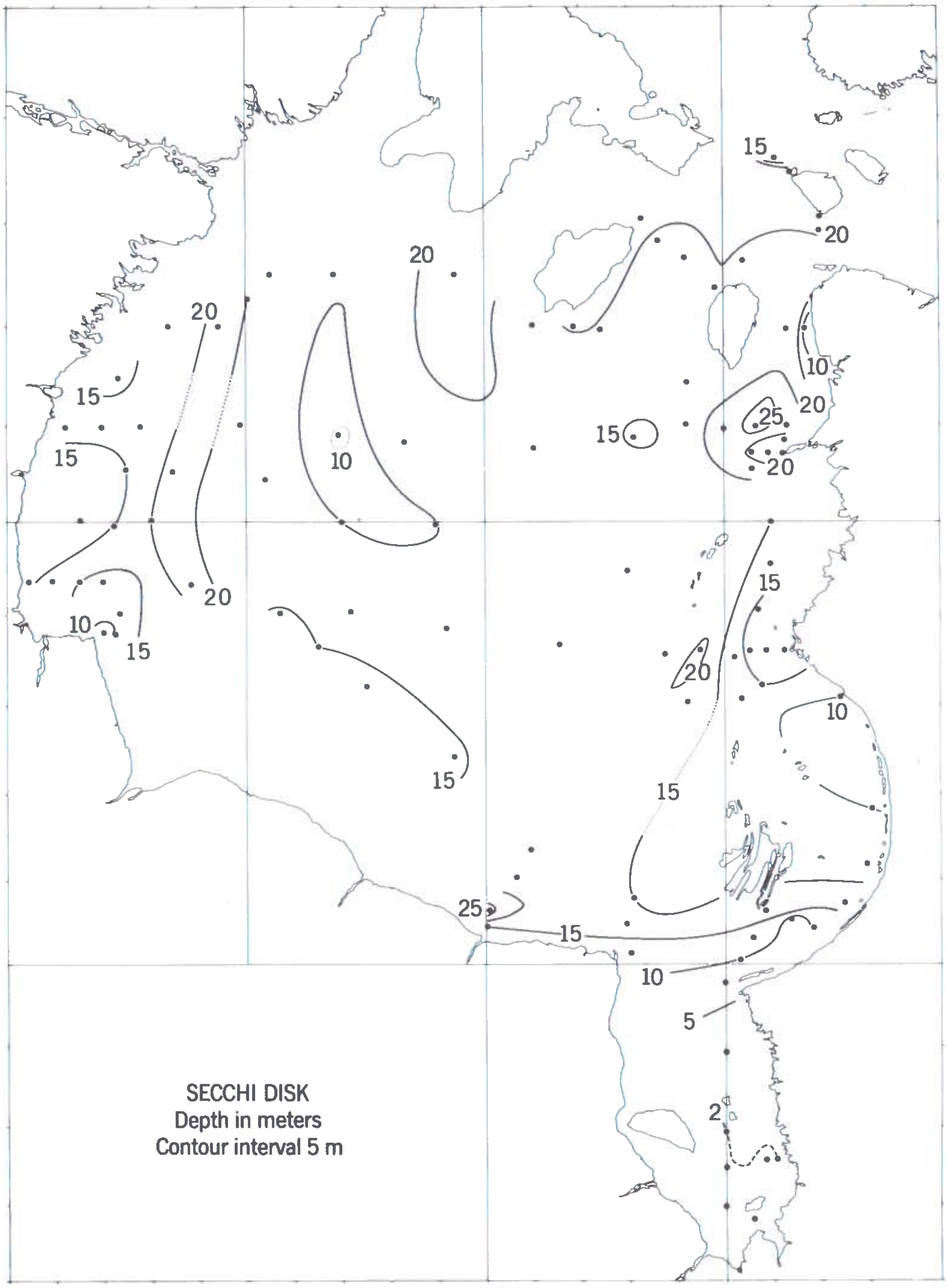

Figure 14. The distribution of Secchi disk observations in James Bay and Hudson Bay based on data observed in 1959 (Grainger, 1960) and in 1961 (Anon., 1964a). The solid circles indicate the location of the observations. A dashed contour indicates one additional to the regular contour interval; a dotted contour indicates a doubtful interpretation.

considered that as the sediment load of the Moose River appeared small in summer months it must be large during the spring freshet. The north-south gradient at the mouth of James Bay does not appear to be reflected in the distribution of either the surface salinity (Figure 4) or the freshwater content (Figure 11) as might be anticipated. A tentative conclusion is that the relatively low Secchi disk readings in James Bay represent the influence of that portion of the input of freshwater derived from runoff*, as opposed to that derived from the ice cover of the previous winter; the gradient across the mouth would represent the "front" of the seaward moving freshwater from runoff. This suggests that the stored volume of freshwater within James Bay derived from rivers entering there may be distinctly time-dependent. Secchi disk depth would then be time-dependent, and so might sediment deposition.

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*It was anticipated that the optical difference of the sea surface caused by the sediment laden runoff into James Bay during the spring and early summer might be visible in APT optical photography. While the available data are not sufficient to allow a study of this, it does seem that the contrast is so slight that a muddy water would not be recognized. On the other hand, Taggart *et al.*, (1965, p190) were able to detect tone variations in the APT photography of the bay in September which they related to the shallow water there.

The latter feature could lead to the existence of annual varves in the sub-bottom vertical sediment structure. Leslie (1965, p136) studied the sediment core obtained at "Theta" station 104 in the approach to James Bay but did not indicate the existence of varves. However, he did suggest that the bay is an important source of finer material in the bottom sediments of Hudson Bay. He remarked (p18), "North and east of James Bay the bottom sediments consist mainly of medium grey silty clay," and (p20), "James Bay is the source of medium grey sediment along the southeast coast and around the Belcher Islands. Rivers flowing into James Bay drain the region to the south which is underlain by soft Mesozoic shales and siltstones. Much of this fine detritus is carried into James Bay and thence into Hudson Bay". This distribution of sediment finer than 2mm in diameter as deduced by Leslie (1964) is shown in Figure 15. The area of silty clay, which apparently originates in James Bay, extends to the north and east and the westward in the northernmost part of its distribution. The distribution is compatible with present knowledge of the circulation; the main feature of which is believed to be an anticlockwise movement around the bay with relatively strong northerly currents along the east shore. The westward extension may reflect the

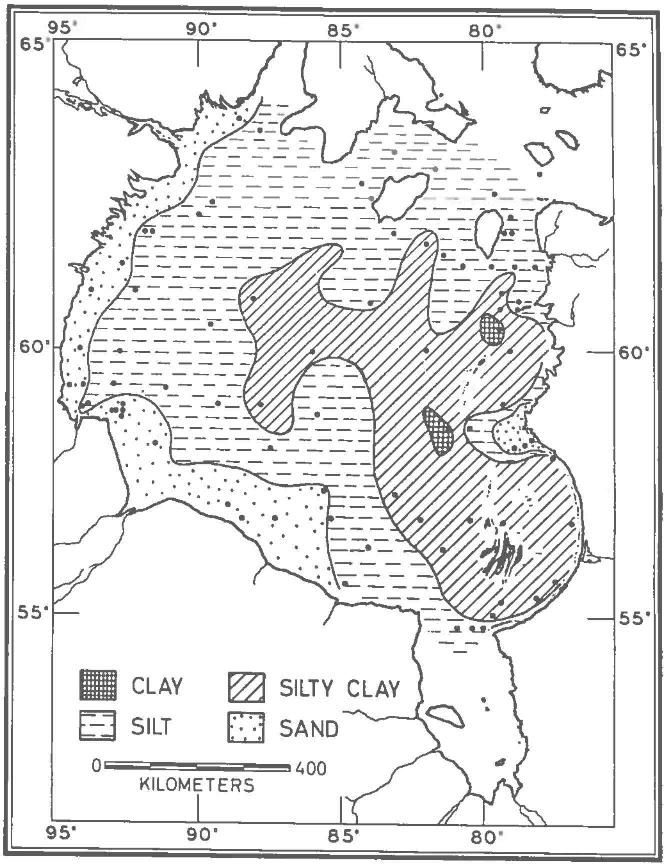

Figure 15. A re-presentation of Figure 8 of Leslie (1964) the caption to which read, "Distribution of bottom sediment finer than 2mm in diameter, or the predominately water-deposited material".

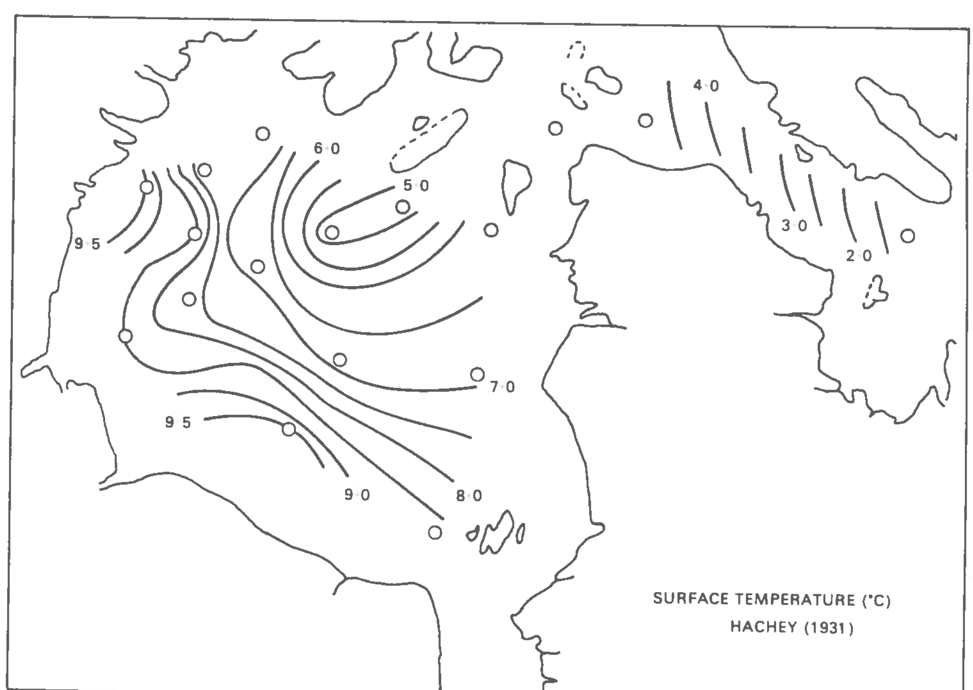

Figure 16. A reproduction of the distribution of surface temperature in Hudson Bay in 1930 (from Hachey, 1931b his Figure 8).

influence of a pattern of water re-circulation* within Hudson Bay.

2.3 Some recent observations within the system

Wendland and Bryson (1966; 1967) obtained surface temperature data in Hudson Bay using airborne radiation thermometry and concluded, "A signature of last ice of Hudson Bay apparently can be located throughout the remainder of the ice free season". They attributed the persistence of the feature to the "Stratification of the upper layer of the Bay"; a stratification in salinity due to melting ice. It seems that such stratification could have quite the opposite effect and also it seems that other factors could be important including the ice cover itself. Nevertheless, the existence of such a "signature" in the temperature distribution at the surface appears to be entirely possible as other summer data reflect such an influence and it is known that each year a characteristic pattern of ice dissipation, from north to south, can be expected (Markham, 1962, p6; Larnder, 1968, p335; Danielson, 1971). The 1930 data of Hachey (1931b) are an exception, for not only is a signature not apparent (Figure 16) but also the surface temperatures, particularly in the west and

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*A (drift) bottle set adrift in 1952 at a position off the northeast coast of Hudson Bay, 10 miles west of Povungnituk, was found in 1968 on the southwest shore, 4 miles northwest of the Kaskattama River (Richard H. Russell, personal communication).

in the south, are relatively high. Of the 1930 data Barber (1967, p7) remarked, "...There does not appear to be any effect due to a recent ice cover or accumulation of ice as in the 1961 season". He also remarked (p7) that 1930 may have been a light ice year, but this is quite speculative as the data (Anon., 1931) on the distribution of ice are few. Of these the second voyage of the tug "Ocean Eagle" during July 10 to 18, 1930 suggests a scattering of the ice (p14 and 15) eastward of Churchill rather than a concentration. If a subsequent movement did not concentrate* the ice in 1930 it is possible that a condition similar to the unusually open season of 1962 occurred. In this, the main portion of the bay seems to have been effectively clear of ice by August 1, (Anon., 1963) so that there would have existed sufficient time and open water for the surface waters to have been warmed by insolation in 1930 to the extent indicated by Hachey.

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*Ice was observed off Port Harrison on August 8 and 9 in SS "Nascopie" (Anon., 1931, p19) and "loose ice" was observed in SS "Ungava" (p21) on August 2 at a position (57°06', 82°58') west of the northernmost Belcher Islands and north of Cape Henrietta Maria on a voyage to Charlton Island. Both support the considerations that the ice condition in 1930 was similar to that of 1962 and not similar to either 1967 (Anon., 1969, Figure 16) or 1969 (Anon., 1971e, Figure 16).

It is a tentative conclusion that in some years the "signature of last ice" may not be apparent.

That 1967 would not likely be such a year was predicted from study of satellite imagery and of ice forecasts (Anon., 1967b) which indicated that an open water situation was well-defined in northwest Hudson Bay and close along the east coast by about May 11. Subsequent imagery indicated increasing open water in the northwest and little or no open water along the east coast where it had been earlier observed. The latter change suggested a movement or "pressure" from the west to east. As well, there was little evidence of open water in the eastern half of the bay prior to the end of June; indeed, some ice persisted in the area south of Coats and Mansel Islands to the end of July (Figure 17). An open water condition along the west coast toward the Bay of Gods Mercy was interpreted from the imagery of April 28; a condition which has been described (Dunbar and Greenaway, 1956, p418; Bowley, 1969, p13). Another feature was the persistence of an ice boundary in Roes Welcome Sound throughout the period April 28 to July 12 (Figure 20a). In 1961, aerial reconnaissance indicated open water throughout Roes Welcome Sound on July 13 (Archibald *et al.*, 1962, their Figure 8) and the relatively high (25 kg cal cm^{-2}) seasonal heat storage to about the latitude of Wager Bay (Barber, 1967) indicated that open water had occurred there early in the season and persisted up to the time of the temperature observations. Clearly, the heat storage in the area in

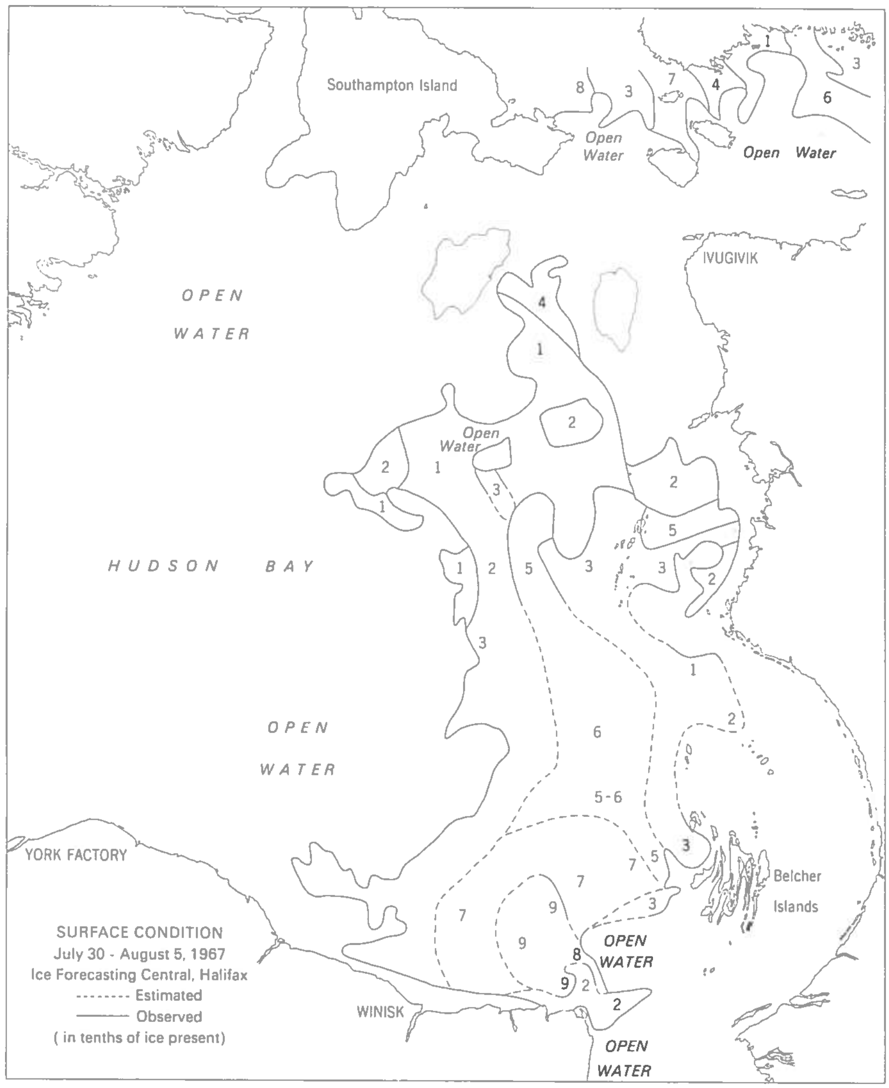

Figure 17. The distribution of open water in Hudson Bay about the end of July, 1967 from Ice Forecasting Central, Department of Transport, Halifax.

1967, if observed, would have proved much less than in 1961. Similarly the slow development of open water in the east suggested that the absorbed incoming shortwave radiation, and hence the surface temperature and seasonal heat storage of the water, would be much reduced. This prompted a request to the Department of Transport for bathythermograph observations in the northeast. These were obtained by CCGS "Labrador" (Anon., 1968) and revealed that the seasonal heat storage was low (Figure 18), certainly much lower than occurred at about the same time in 1961 (Barber, 1967). The subsequent airborne survey in late August (Wendland and Bryson, 1967, their Figure 3) indicated a relatively low surface temperature, which confirmed that in the eastern portion of Hudson Bay the peak of the seasonal heat storage was less than average. Hence their conclusion concerning the "signature of last ice".

On the other hand surface temperature data observed between the Bay of Gods Mercy and Churchill about mid-July in 1967 (Figure 19) indicated a level of temperature close to that observed about the end of July in 1961 (Barber and Glennie, 1964, their Figure 17). If it is assumed that the development of the surface layer was similar in the two years, then it is a tentative conclusion that the peak value of the seasonal heat storage there was at least as great in 1967 as in 1961.

More recently, the 1969 season in terms of development of open water also appears to have been

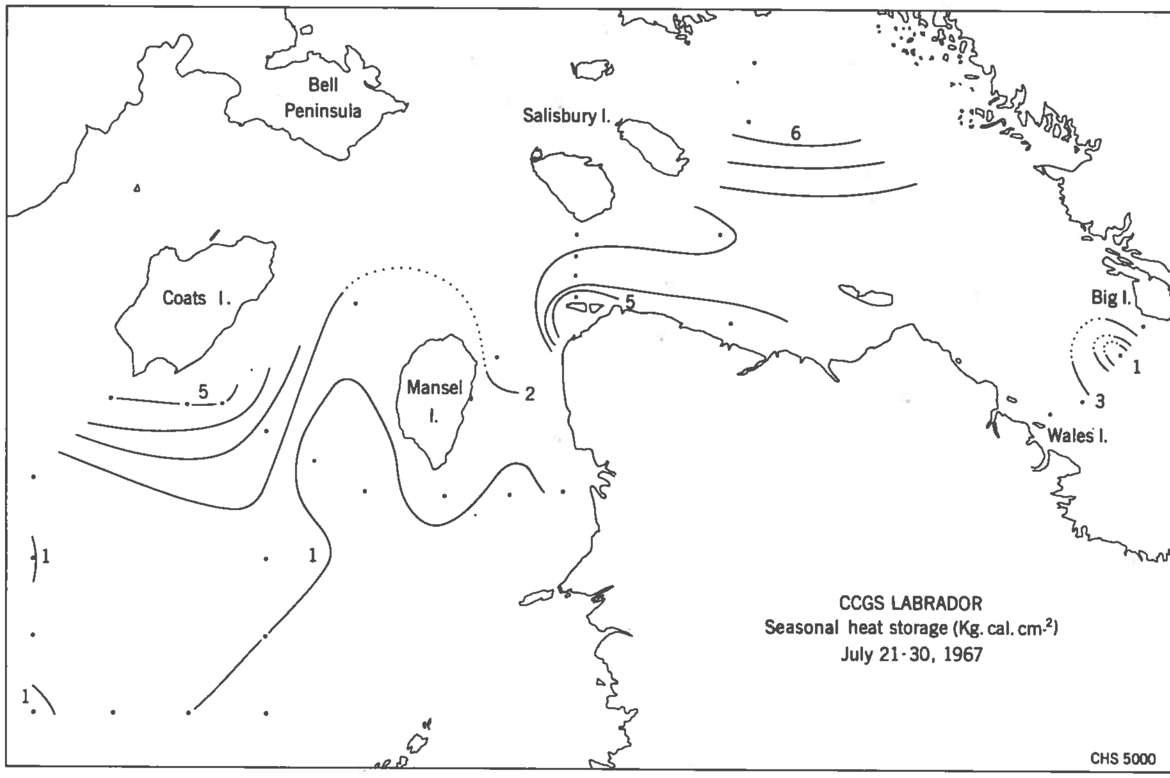

Figure 18. The distribution of the seasonal heat storage in late July, 1967 as interpreted from bathythermograms obtained in the C.C.G.S. "Labrador".

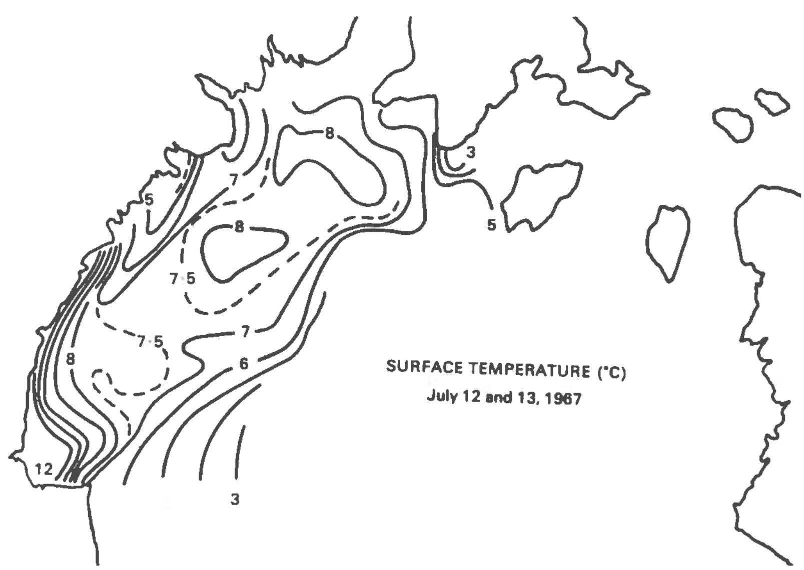

Figure 19. The distribution of sea surface temperature as observed July 12 and 13, 1967 using an airborne radiation thermometer (from Wendland and Bryson, 1967, their Figure 1).

anomalous for, "in southern Hudson Bay final clearing after mid-September was the latest on record" (Anon., 1971e, pl); the persistence was related to the occurrence of "some heavier than usual ice" (p42). At the same time the existence of, "Broad areas of open to close pack ice," west of Belcher Islands to August 20 (p42) was also considered unusual as was the, "Break-up and disintegration of ice which was earlier than normal in northern Hudson Bay..." (pl). The distribution of ice in May and June strongly suggested the influence of advection due to wind and examination of the data indicated that winds from the west through to north occurred 75 percent of the time at Chesterfield, i.e. about 25 percent greater than normal. This pattern of wind no doubt contributed to the earlier open water in the north and to the observed accumulation of ice toward the south and east. It is suggested that the accumulation here was such as to maintain a high average albedo well into the normal melting season and was a factor in the persistence of the ice cover there.

The deeper temperature data of 1967 (Figure 20a), while limited, suggest the existence of a slighly warmer water, to -0.9°C, in a tongue-like distribution southwest of Coats Island similar to that observed in September 1962 (Figure 20b). Such warming is thought to be due to downward mixing of heat absorbed at the surface in that season, but was not considered to occur as early as observed in 1967, i.e. as early as the end of July. It seems possible that the extensive open water in 1967 in the northwest

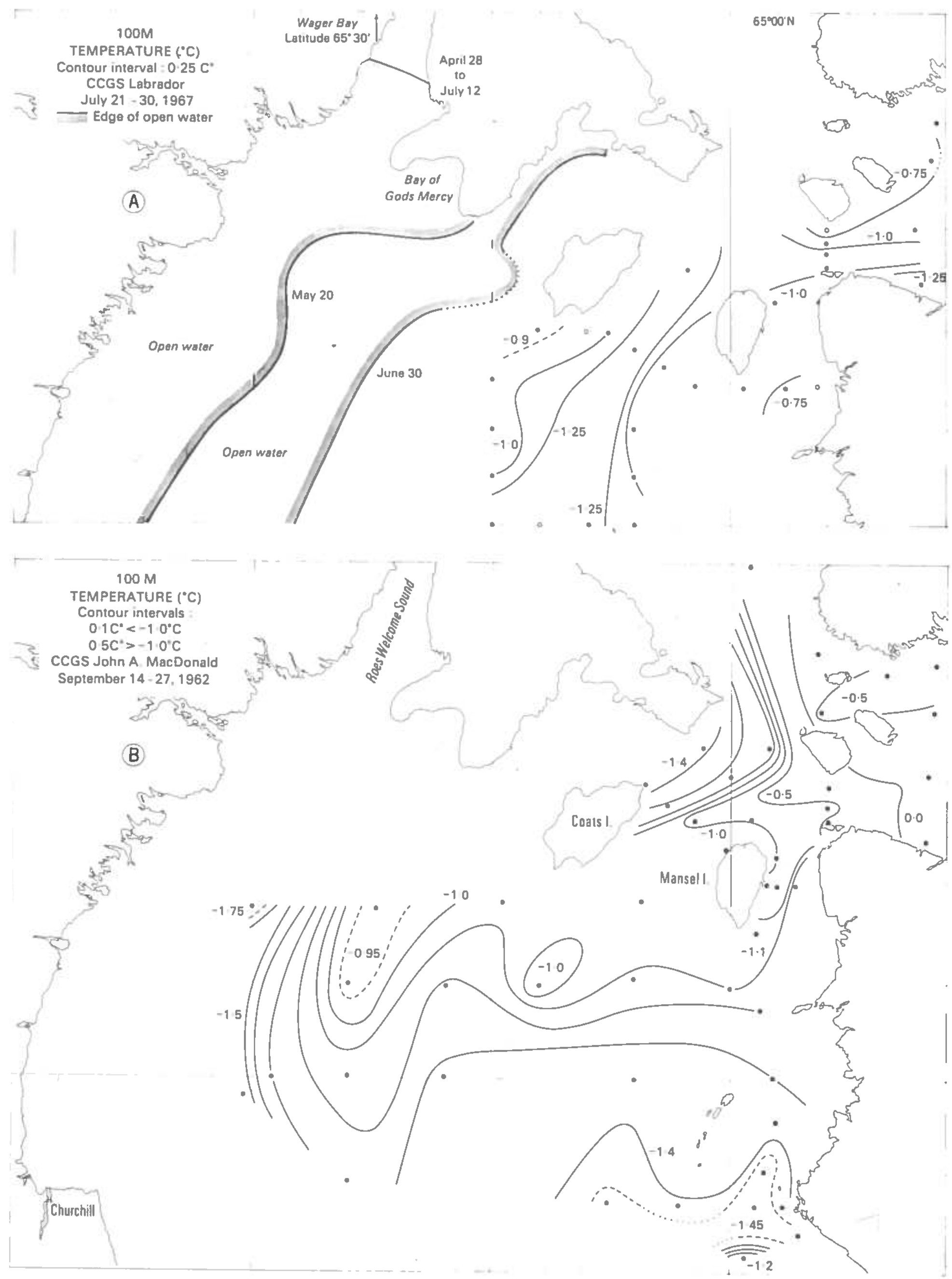

Figure 20. The temperature distribution at 100m depth in 1967 and 1962. (a) From the 1967 "Labrador" bathythermograms. The edges of open water at the dates indicated are an interpretation by the author of satellite (APT) imagery. (b) From the 1962 "John A. Macdonald" serial data (Anon., 1966).

may have contributed to this result which, due to the similarity in open water, probably occurred again in the 1969 season.

There is one major caveat here concerning the 1967 data and that is they are the result of but one instrument, i.e. one bathythermograph, so that both the values and pattern (Figure 20a) are suspect. The chances are, however, that the bathythermograph data are appropriately used here relative to the assessment of heat storage. It is realized that this may not be the case for all data observed within the system, for certain of the apparent differences are disconcerting. The most recent example is that of Pelletier *et al.* (1968) who, even though aware of earlier interpretations, could report for Hudson Bay, temperatures to -2.0°C (their Figure 4 and p565) and salinities in excess of 34.0 °/oo (their Figure 6* and p565) without particular comment, either to the precision of the observations or their relevance. Such extreme values were reported on earlier occasions, but at a time when data were relatively few. For example Campbell (1964, p49) reported temperatures to -1.97°C and salinities to 34.07 °/oo at unusually shallow depth in Foxe Basin.

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*Their figure 6 indicates a value of 35.56 °/oo which may be an error in the preparation of the illustration.

The existence of such extreme value generally leads to conjecture about the process which forms the water; a process which may be a major feature of the oceanography of the region. A consideration of all the data led to the definition of the probable winter surface salinity (Barber, 1967) and to a simple model (p54) of the Hudson Bay system. The model also served as a review of earlier concepts and emphasized that these were hampered through lack of winter data. Hachey (1931b) realized this and developed the view that, "the waters of Hudson Bay differ markedly from the waters of Hudson Strait and the waters of the open ocean". Of course he was not aware that conditions in Hudson Strait exhibit marked seasonal variation. He was led (Hachey, 1954) therefore to consider the evolution of Hudson Bay water alone and offered two conjectures. In one was visualized removal of the observed summer low salinity surface layer prior to freeze-up, so that water cooled at the surface would sink and become the deep water. In a variation, he visualized that water initially at the surface in winter in Hudson Strait moved into Hudson Bay replacing a less-dense seaward flowing water. In the other he suggested a movement "of deep, cold, saline water" into Hudson Bay from Foxe Basin; the origin of such water remained a question. Campbell (1964) expressed his accord with the latter hypothesis, describing observations made in 1955 and 1956 in support, and developed a theory wherein the cold high-salinity water in Foxe Basin was directly related to the production of winter ice in the extensive inter-tidal zone (tide flats) of Foxe Basin. Apparently he considered the "exposure" of the tide flats the significant feature in the process and visualized the production of a "concentrated brine solution". No primary evidence which might lend credence either to the existence of such a brine solution, or to the main hypothesis, was presented. However, the described data are important as they demonstrate the existence below sill depth in Foxe Channel in both years of a high salinity water (about 33.7 $^{o}/oo$) close to the freezing point. This conditioning must have occurred at the surface in association with cooling during the winter months. The water so conditioned would sink and move away, if not depth limited, from the source area eastward into Hudson Strait, and into Hudson Bay, enhanced perhaps through increased heat losses in areas of open water, including tide flats. The extent of open water in the system at this time is not known; however, amounts likely significant in the formation of the water are believed to occur in northern Foxe Basin, throughout Hudson Strait, Foxe Channel, and Frozen Strait to Roes Welcome Sound.

Looking beyond the system, it is suggested that the process would be favoured by an extension of the Baffin Current into Hudson Strait along the north shore and into the area of Foxe Channel. The evidence for such a current during the winter is limited, but it is the hypothesis that it exists and in association with a process similar to that described by Mosby (1934) for the formation of a cold, saline water on the Antarctic continental shelf, particularly in the Weddell Sea. In this, he elaborated the conclusion of Brennecke (1921), that the deep and bottom water of the Antarctic is derived in part from an extremely cold and moderately saline water formed in winter through cooling at the surface and freezing on the shallow shelf. Mosby emphasized the importance of horizontal movement whereby the surface water sinks and moves away from the source area, and off the shelf. The similarity of the situation in the Antarctic with that observed in the Foxe Basin - Hudson Bay system is heightened for according to Mosby, a portion of the winter-formed water is prevented from mixing into the Antarctic deep through the existence of a limiting depth. Thus, it retains its main characteristics, i.e. like the deep water in Foxe Channel a temperature very close to the surface freezing point. It was these data, specifically those at "Deutschland" station 125, that led Brennecke to the hypothesis concerning the formation of the winter shelf water, and its subsequent contribution to the deep water. Deacon (1937) and Fofonoff (1956) have contributed further to the understanding of the process.

In the situation for Hudson Bay it is thought that the observed cold and saline water at depth in Foxe Channel is the result of the cumulative effects of cooling and ice formation at the surface of the current throughout its movement along the east Baffin Island coast, into Hudson Strait, and into Foxe Channel and Foxe Basin. It is not recognized as a distinct water outside the system in the Labrador Sea because it is too dense to remain at the surface and is not sufficiently dense to participate in the formation of a distinct bottom water. It must then contribute to an intermediate water there.

An assessment of this contribution will not be attempted here, although it is of major interest, for it seems that much more needs to be known of the influence of Hudson Strait, where intense mixing due to tides leads to a further modification of water characteristics. These subsequently became recognizable within Hudson Bay through an inward movement associated with the estuarine circulation, which in effect results in considerable recirculation, i.e. the coupling of the system to the Atlantic through the estuarine circulation is limited by mixing in Hudson Strait.

Assessments of plankton data by Grainger (1961) and Bursa (1961) showed that both Atlantic and Arctic plankton types were found in Hudson Bay, and Grainger (1962) made a similar observation with regard to Foxe Basin. In a study of the distribution of three species of copepod, Grainger (1963) showed that while two Arctic species were widespread in Hudson Bay, the Atlantic was not. Huntsman (1954) in a discussion of the production of life in Hudson Bay outlined a number of reasons for the apparent low productivity as compared to areas at similar or higher latitude and, while appearing to emphasize the "lack of heat," concluded that as yet the data are too few "on which to base proper judgement". Other factors which might influence the "apparent low production" (Grainger, 1968, p357) include "the effects of long periods of ice cover" (p358) and consequences related thereto. Dunbar (1970) on the other hand apparently considers nutrient levels to be important. He suggested that the oceanographic regime of Hudson Bay would be altered if the "supply were cut off" of Arctic water moving eastward through Fury and Hecla Strait. He considered the significant alteration would occur in the stability which would become less and as a consequence of mixing processes, more of a nutrient of the deeper water would become available to the surface layer, where productivity would increase. Of a number of questions which might be raised, that concerning the influence of Hudson Strait seems the most important. It is visualized above that mixing processes in the strait determine to considerable degree the distributions within Hudson Bay, such that a change in a surface water characteristic, e.g. to a higher salinity, would be reflected in a change in the same direction in the deeper water. Thus, a decrease in the contribution to the system of low-salinity surface water brought about by damming Fury and Hecla Strait would lead to an increase in salinity throughout, perhaps with little change in stability.

It seems that Dunbar assumed that the nutrient level in the deeper water is high, but this may not be. Should the uncoupling and recirculation suggested here be significant it is possible that the level of nutrient, or of a nutrient, may be generally low throughout the system (it may be that the limited occurrence of Atlantic copepod there may also be a partial result of this uncoupling).

Nevertheless, it would be of interest to demonstrate that Dunbar's secondary consideration is indeed possible, i.e. that productivity would increase in a region of annual ice cover were the level of nutrients increased. As noted, data are not available for Hudson Bay, but the evidence for some other areas with ice cover indicate that a depletion of at least one nutrient occurs by mid-summer (Apollonio, MS undated; McLaren, 1969). If this is so, it seems that an experiment (McLaren, 1969) in which nutrient is applied to a relatively isolated (uncoupled) body of water in order to avoid the development there of a period of nutrient depletion might provide a useful result. Omarolluk Sound is suggested as a suitable site should it be determined that the nutrient depletion occurs there.

A further consideration here relative to Hudson Bay is that of the dissolved oxygen. It seems that it is characteristic of the water in Hudson Strait to be near saturation levels, from which it follows that the moderate depletion of oxygen observed in the deeper water of Hudson Bay occurs entirely within the bay. If the consumption there were known, it would be possible to estimate an age for the oxygen depleted deep water. This is apparently not known, but is likely about the 0.21 ml/l/year suggested for the North Atlantic by Riley (1951). Assuming about 80 percent is retained after leaving the surface and subsequent oxidation of surface nutrients (Redfield *et al.*, 1963), then the age is in the range 5 to 7 years.

2.4 Longer term change within the system

Bailey and Hachey (1951) recognized that the general level of salinity observed by Hachey (1931b) in 1930 in Hudson Bay was low. They compared the data to observations made in 1948 and suggested that the observed difference was due to an increased Atlantic influence. It was suggested (Barber, 1967, p55) that it is not possible through study of other data to reject their hypothesis. In the latter work and in Bailey and Hachey (1951) it appears to have been assumed that the tabulated depth and salinity values (Hachey, 1931b, p96) are without more than the usual error.

It is not difficult to accept the assumption as regards the salinity data for, although they are generally low, it is possible to see in the observed distributions similarities with more recent data. For example, the distribution in Hudson Bay at 50m (Hachey, 1931b, his Figure 4) can be interpreted so that the pattern is compatible with a 50m distribution shown here (Figure 4) based on much more data. However, there are two salinity values in Hudson Strait, each at 200m depth at stations 57 and 48 (Hachey, 1931b, p96), which on the basis of all the data are 0.5 $^{o}/oo$ lower than is to be expected at the depth. At nearby station 58 the salinity values within the surface layer, 32.5 $^{o}/oo$, are appropriately high. Thus, the data in Hudson Strait and in Hudson Bay suggest that if a persistent error exists it is not in the salinity and could be in the depth.

Depth data presented in the report of the 1930 observations (Hachey, 1931b) are of two kinds. One is the depth to the bottom at each station (p95) which when plotted appear to fit current information. The other data are the tabulated value of the serial samples (p96). A feature of the tabulation is that each is at a "nominal" depth. This could, of course, be easily achieved, particularly in Hudson Bay where water movements are generally small. In Hudson Strait, however, strong water movements are known to exist so that extraordinary consideration would have had to be given in order to achieve a desired sampling depth, particularly in the circumstance that a cast comprised a lowering of only one reversing bottle (Hachey, 1931b, p95). It is suggested that this did not occur and consequently the depth data are liable to more than the usual error. Consideration of the validity of the hypothesis of Bailey and Hachey (1951) must therefore include an evaluation of the precision of the depth data.

It does appear therefore that there is little good physical evidence on which to base hypothesis for recent changes of marine climate, i.e. beginning about 1930. It seems that such hypotheses were of much interest in about the late 1940's and early 1950's, e.g. Dunbar's

(1951) extensive work, so that even limited data in rather complicated fiords were used to suggest the existence of change (Nutt and Coachman, 1952); more recently such data were said to "document" a change in the water of the fiord (Coachman, 1969, p215).

3. Discussion

The region is ice and snow covered for much of the year so the extent that the relative proportion of land and water areas are altered by the project would not have a major influence during the winter because the ice and snow largely serve to uncouple the underlying ground or water and the atmosphere. This uncoupling and the fact that the winter climate is due to air mass movement of global scale suggests that if the project is to have an influence on the water it would occur at other seasons. We know that the world ocean is largely buffered against change and that the James Bay Project would not have significant impact were the project not located in a system so uncoupled from the world ocean. The limitation to the coupling is due largely to processes within the system including tidal mixing in Hudson Strait. In James Bay the coupling to Hudson Bay is also limited, but through a weak convective circulation. It seems that the project could influence this convective circulation.

It is characteristic that the freshwater from runoff entering James Bay would lead to the observed layer of low salinity water in the surface and to a distinctive pattern of circulation. The freshwater moves seaward in this surface layer, entraining salt from below and eventually leaves the bay as salt water. The outward movement of salt is balanced, during some time interval, through a sub-surface inflow of water of relatively high salinity so that:

$$T_i \; S_i = T_o \; S_o;$$

As the freshwater moves seaward at a rate equal to the supply then:

$$T_o = T_i + R.$$

If these be applied in the section across the entrance where S_i and S_o may be 29 $^o/oo$ and 22 $^o/oo$ respectively, then the inflow is three times the runoff; thus the net transport is a significant part of the total circulation.

Present evidence indicates that the strength of the outflow is strongly time-dependent, presumably in association with seasonal changes in the volume of runoff. The outflow would probably be a maximum at some time after the peak in runoff in early summer. Earlier, in late winter, a minimum in the runoff is indicated about which time a minimum in the outflow would occur. As well, the runoff would occur to a region covered with ice and may not entrain the deeper seawater to the same extent as when not ice covered. The energy for mixing is believed input to the system from tides, wind and surface processes of heating and cooling. An ice cover would reduce the influence of wind considerably as well as of the other surface processes, so that mixing would be less under an ice cover and the volume of the inflow would be even closer to that of the runoff than it is in the late summer.

The significance of the foregoing is that the runoff does not provide a strong coupling to the water of Hudson Bay, i.e. the estuarine circulation is relatively weak, so that changes in the pattern of runoff may not change the coupling significantly. Should changes in the extent of the ice cover occur, say to decreasing cover, then an increase in the coupling could be expected and a greater forced circulation would result. However, the return flow comprises a relatively cold water, -1.4° to -1.0°C, so that an increased coupling during the summer (period of heating) would tend to a colder surface layer. Conversely, uncoupling would tend toward a warmer surface layer.

For example, if James Bay were completely uncoupled from Hudson Bay through a physical barrier across the entrance, except that an outflow occurred to balance the inflow as in a lake, then a direct influence of Hudson Bay water would not occur. Eventually the bay would likely contain freshwater only, which would be warmer than now, probably close of 4°C in the deeper water, and would likely undergo a wider range of temperature in the surface as a greater storage of seasonal heat would take place. An ice cover would still form and would likely be similar to that which occurs now. The sum of the radiative and turbulent flux terms in the annual heat budget would be close to zero, as it is now believed to be (assuming the average temperature of the inflow is the same as the outflow), but the average temperature would be higher throughout.

If the coupling were somehow steadily increased the water in the bay would tend toward that of Hudson Bay, in particular toward that comprising the inflow, i.e. toward a water of salinity 29 °/oo and temperature about -1.4° to -1.0°C, but eventually without an ice cover. Of course this situation would not likely be achieved, for processes within James Bay would, at some stage, begin to influence distributions within Hudson Bay, i.e. the radiative and turbulent flux terms of the heat budget would show a large deficit which could only be balanced by advection of water from Hudson Bay. It seems, therefore, than an increased coupling to Hudson Bay would tend to a decrease of water temperature, while a decrease in the coupling would have the opposite result. An increase in the runoff to James Bay, for example by diversion from Grande Rivière de la Baleine, would increase the coupling, while the smoothing of the runoff so that a greater portion occurs under an ice cover would tend to reduce the coupling. It does not seem possible now to provide a quantitative estimate of the results which may be anticipated, except that the influence of the smoothed runoff would be greater. This would decrease the coupling in summer so that the water in summer would tend to be warmer, or at least would tend to store more heat. Most of this heat would be given off during the period of net heat loss prior to ice formation. During the latter part of this period rather massive losses occur, perhaps as large as 500 - 600 g cal cm^{-2} day^{-1}, so that in order for the time of formation of first ice to be influenced, the stored heat would have to be appropriately large. An estimate of the difference has not been possible. Neither has it been possible to determine the changes which might be brought about in the average pattern of ice cover.

In this simple analysis it is assumed that changes in the surface condition in Hudson Bay, mainly that of the ice and snow cover, will not occur. There is, however, the possibility, described in earlier section (2.2.1), that the present distribution of ice in early summer there may reflect the inferred strong time-dependence in the outflow at the surface from James Bay. Some conclusions might be possible were we to achieve an understanding of the factors which determine the present average situation.

Winds, which are generally from the northwest and north early in the breakup, have generally become light and variable by mid-July and continue so into August. Thus, after creating an accumulation of ice toward the south of Hudson Bay by about the end of July, the wind has little or no influence so that no subsequent ice movement occurs due to wind. Furthermore, there is evidence that during this period of stagnation the frequency of both cloud cover and fog is increased so that less insolation reaches the ice surface and melting is retarded;

complete clearance of the ice may not occur till about the end of August on the average.

In conclusion, it seems likely that advection is not a significant factor in the total heat balance of the system and although a real understanding of the relative influence of inflows and outflows and of processes is not achieved, the system within Hudson Strait appears to be closed rather than open. This tentative result emphasizes the importance of the radiative and turbulent flux terms, i.e. the mainly climatic factors, which in turn reflect the influence of a process global in scale, at least in winter through to early summer, which determines the nature of that important variable the surface condition, at this time either ice or snow. Within the system the water of James Bay is not strongly coupled to that of Hudson Bay but as similar climatic conditions prevail so the surface condition is one of ice or snow for a significant period. It does not seem likely that the project could measureably influence the coupling of the system to the world ocean. It could, however, influence the coupling of James Bay to Hudson Bay to the extent that a greater storage of heat may occur within James Bay during summer after breakup and influence the distribution of ice in the approach to the bay during breakup. Thus the impact of the James Bay Project on the system will be relatively small and it may be necessary, in order to assess the impact, to acquire a particularly extensive body of data; a significant portion of which should be obtained during the period of ice cover.

At other times it may be possible to utilize remote sensing techniques to considerable advantage; although the fact that the peak of the present runoff occurs when there is still considerable ice cover emphasizes the difficulty of the field problem. Of particular value now would be information on the water structure over a period of a year at one postion within the bay, although data in sections seaward of Rupert Bay, Fort-George and Poste de la Baleine and eventually at Inoucdjouac would be preferred. Part of the field programme could be based on relatively simple oceanographic instruments with a frequency of observations about once a week and perhaps sustained by residents there, at least during the period of ice cover, as occurred at Tuktoyaktuk (Kelly, 1967, p8).

4. Acknowledgements

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6. List of figure captions

7. Appendix

Figures 21, 22, 23 and 24 based mainly on the "Calanus" 1959 data (Grainger, 1960), but including some of the "Theta" data (Anon., 1964a).

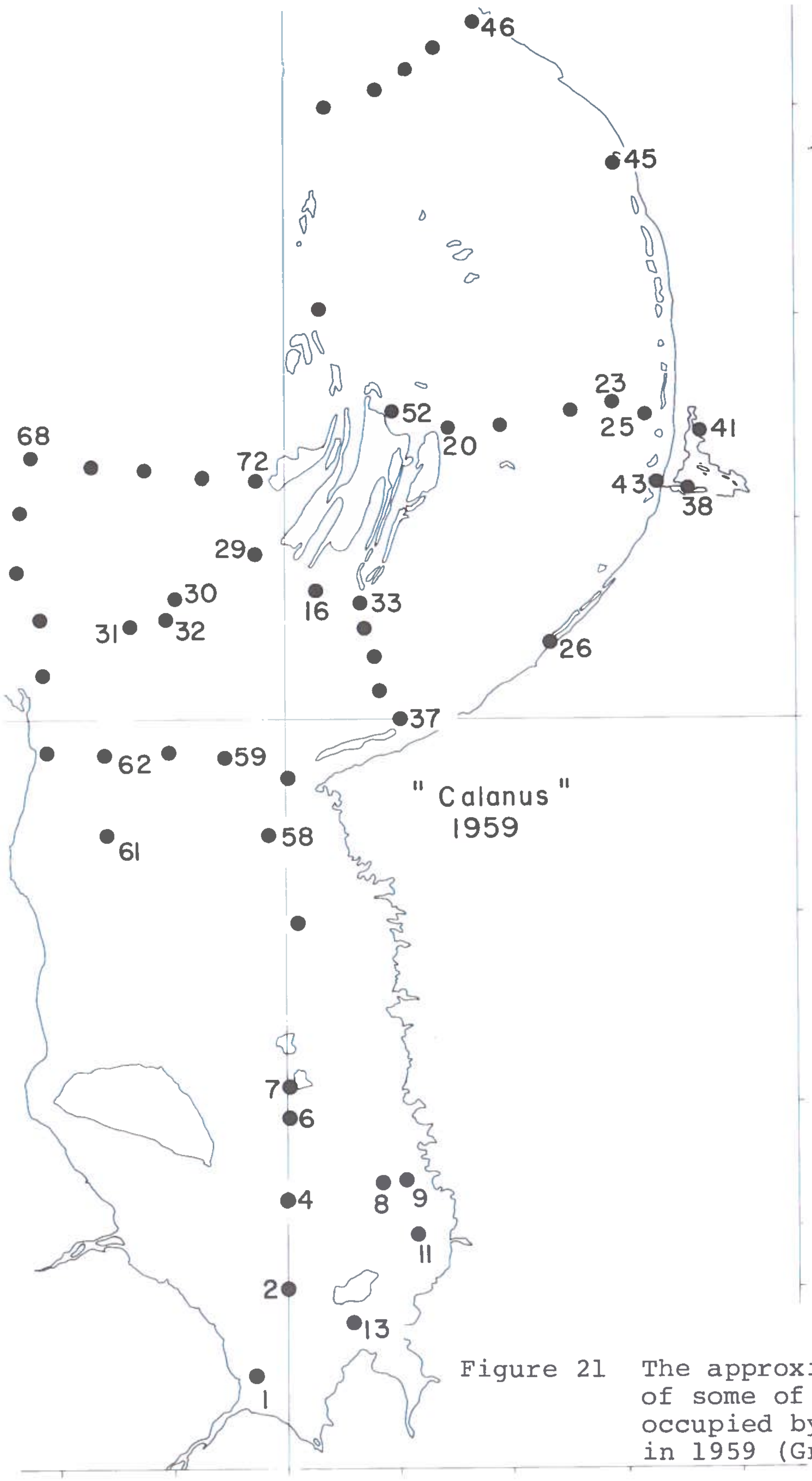


Figure 21 The approximate location of some of the stations occupied by "Calanus" in 1959 (Grainger, 1960).

Figure 22. Distribution at various depths of salinity (o/oo) and temperature (°C) as interpreted from the "Calanus" data of 1959. The positions of the stations are shown in Figure 21 and the date of sampling in Table 2 where it may be seen that the interval between occupations was relatively long.
(a) Surface. (b) 10m. (c) 25m.
(d) 30m. (e) 50m. (f) Deepest.

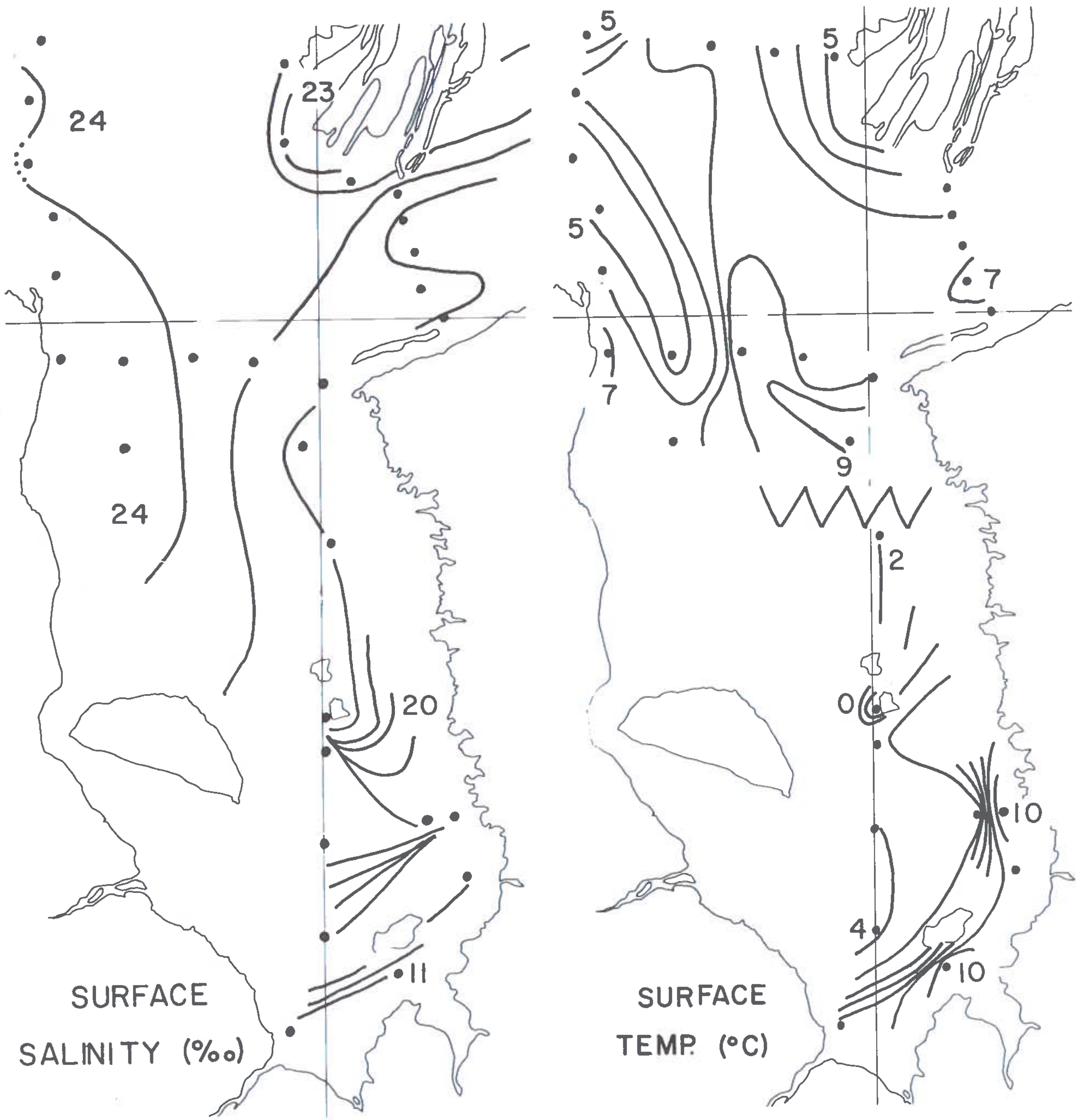

24
23
24
20
11
SURFACE
SALINITY (‰)
5
5
5
7
7
9
2
0
10
4
10
SURFACE
TEMP. (°C)

Figure 22(a)

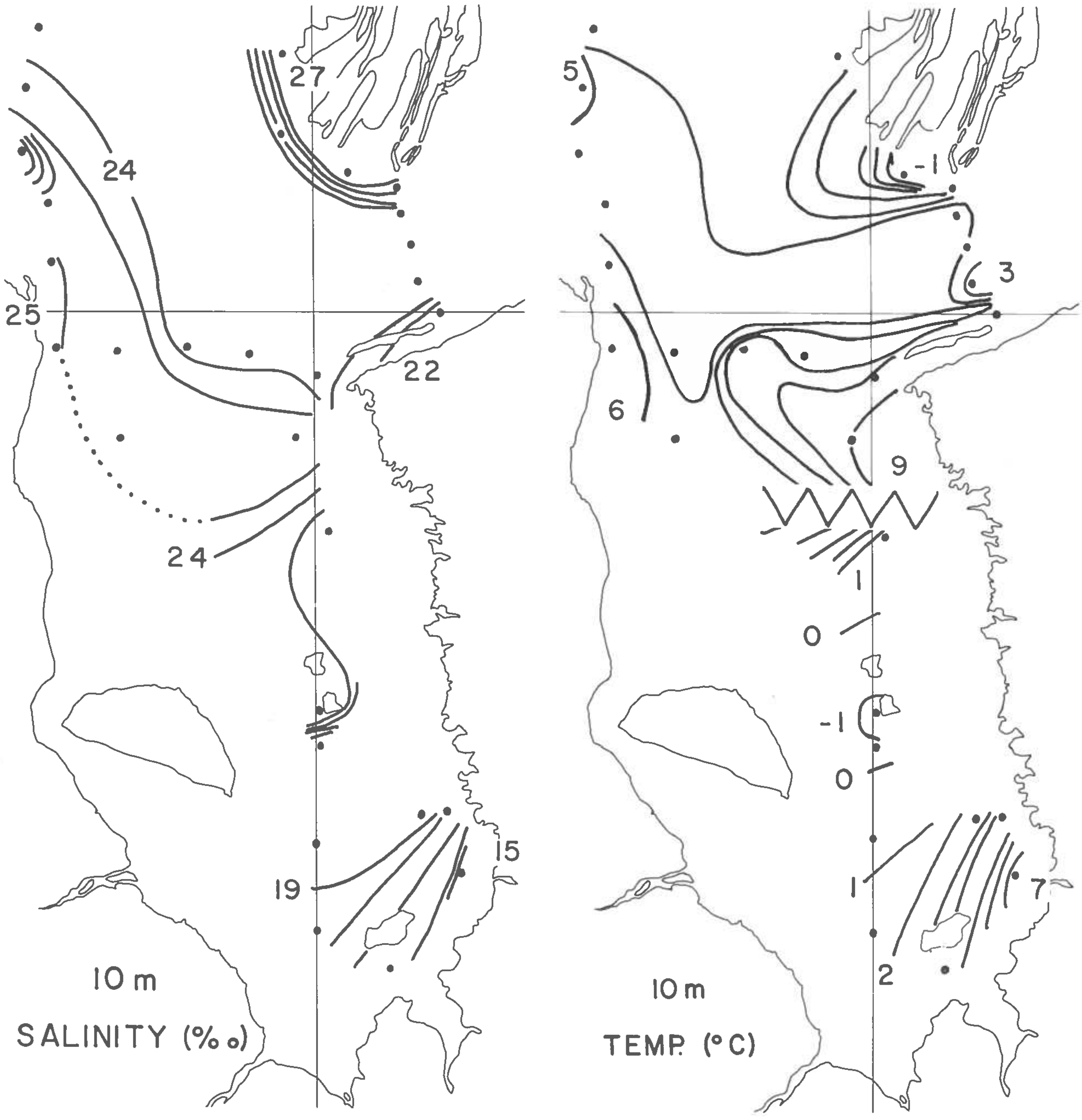


Figure 22(b)

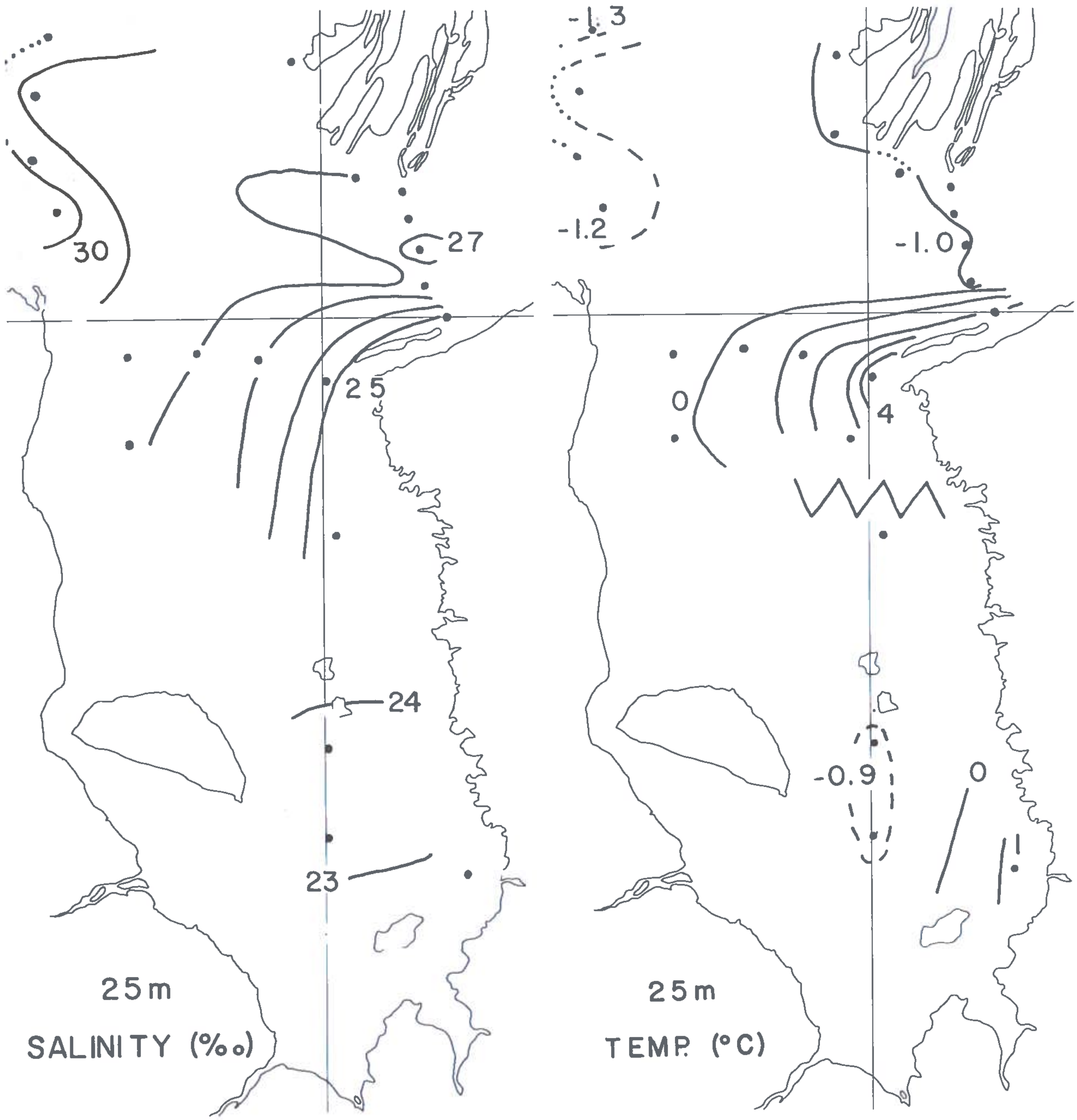


Figure 22(c)

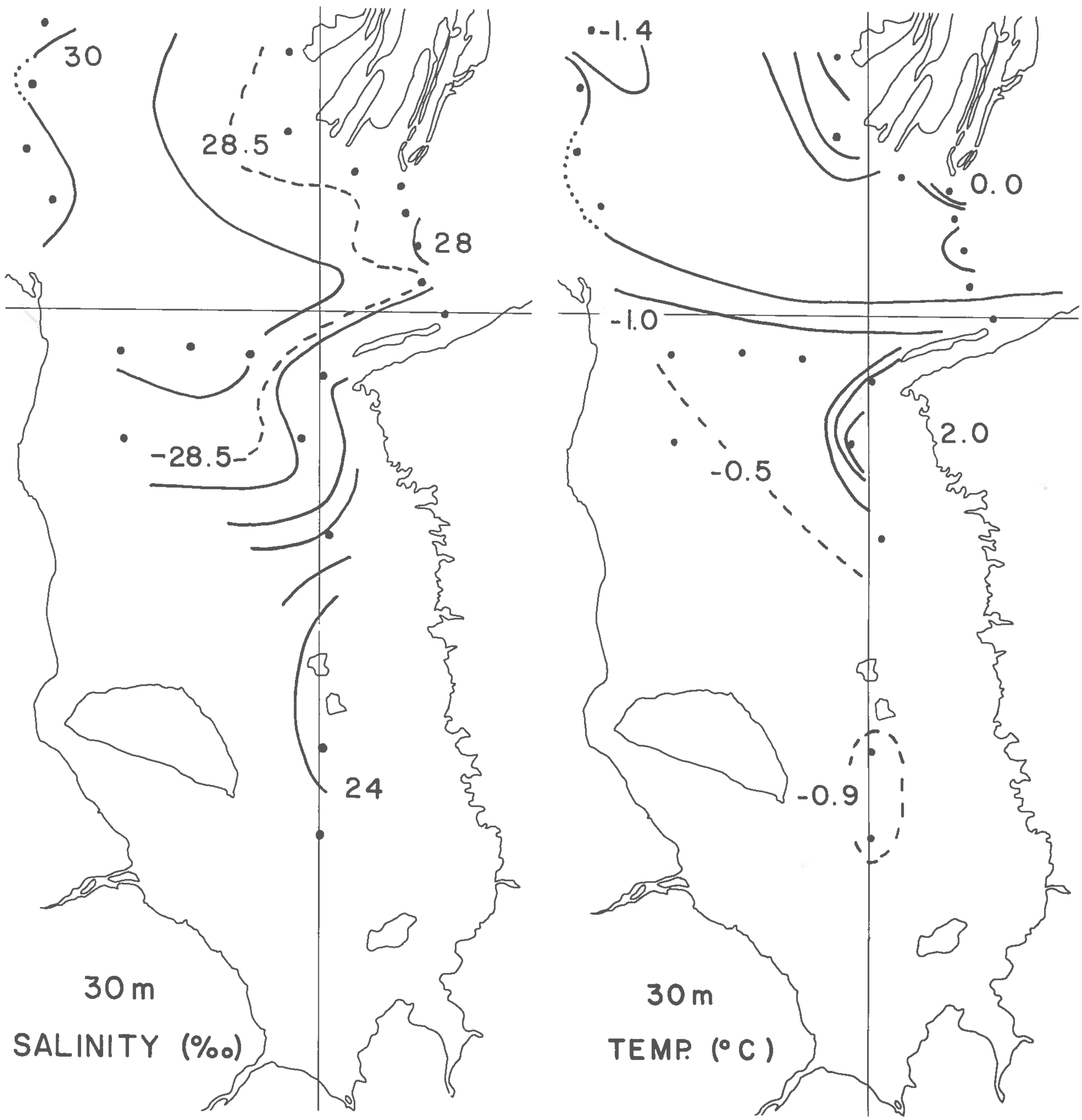


Figure 22(d)

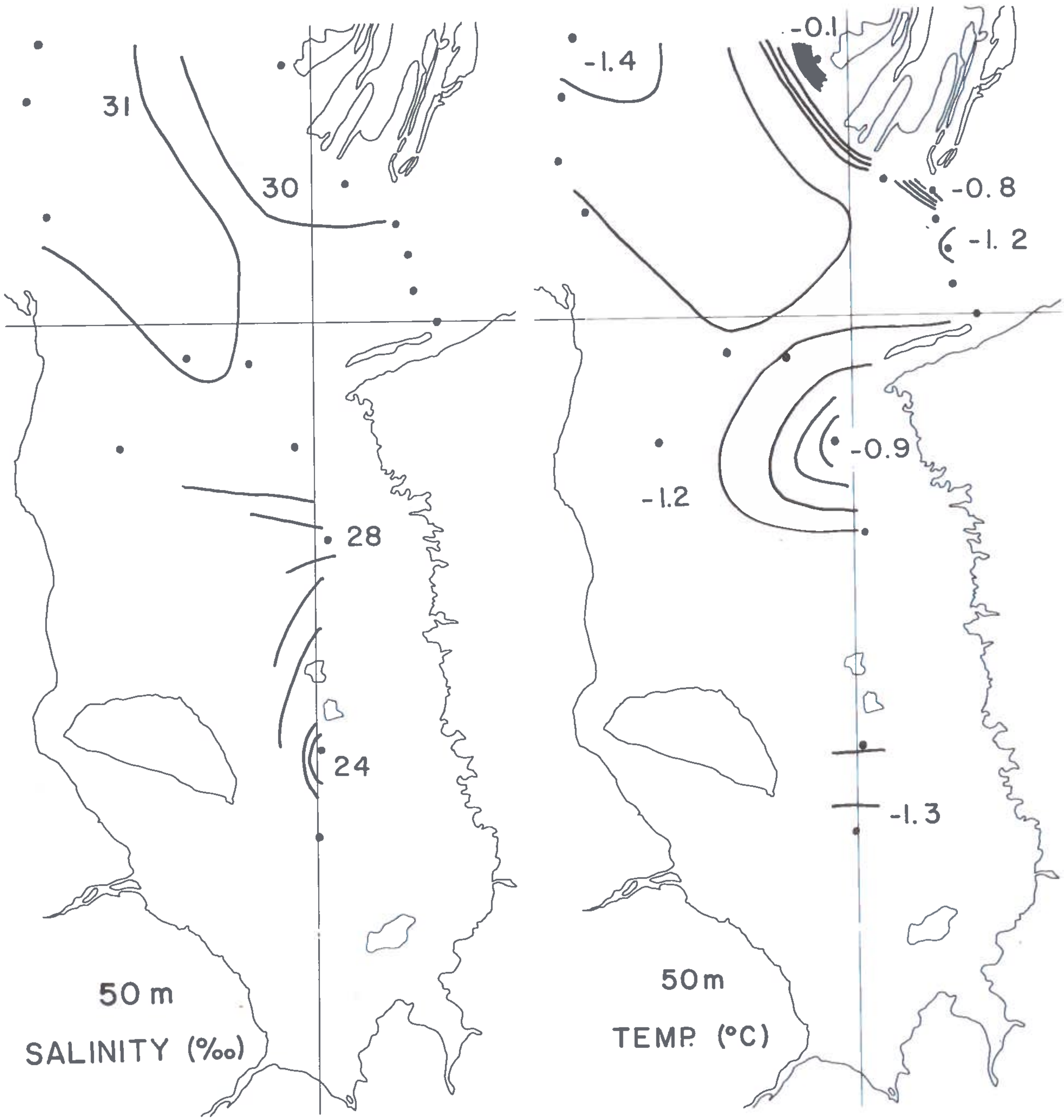


Figure 22(e)

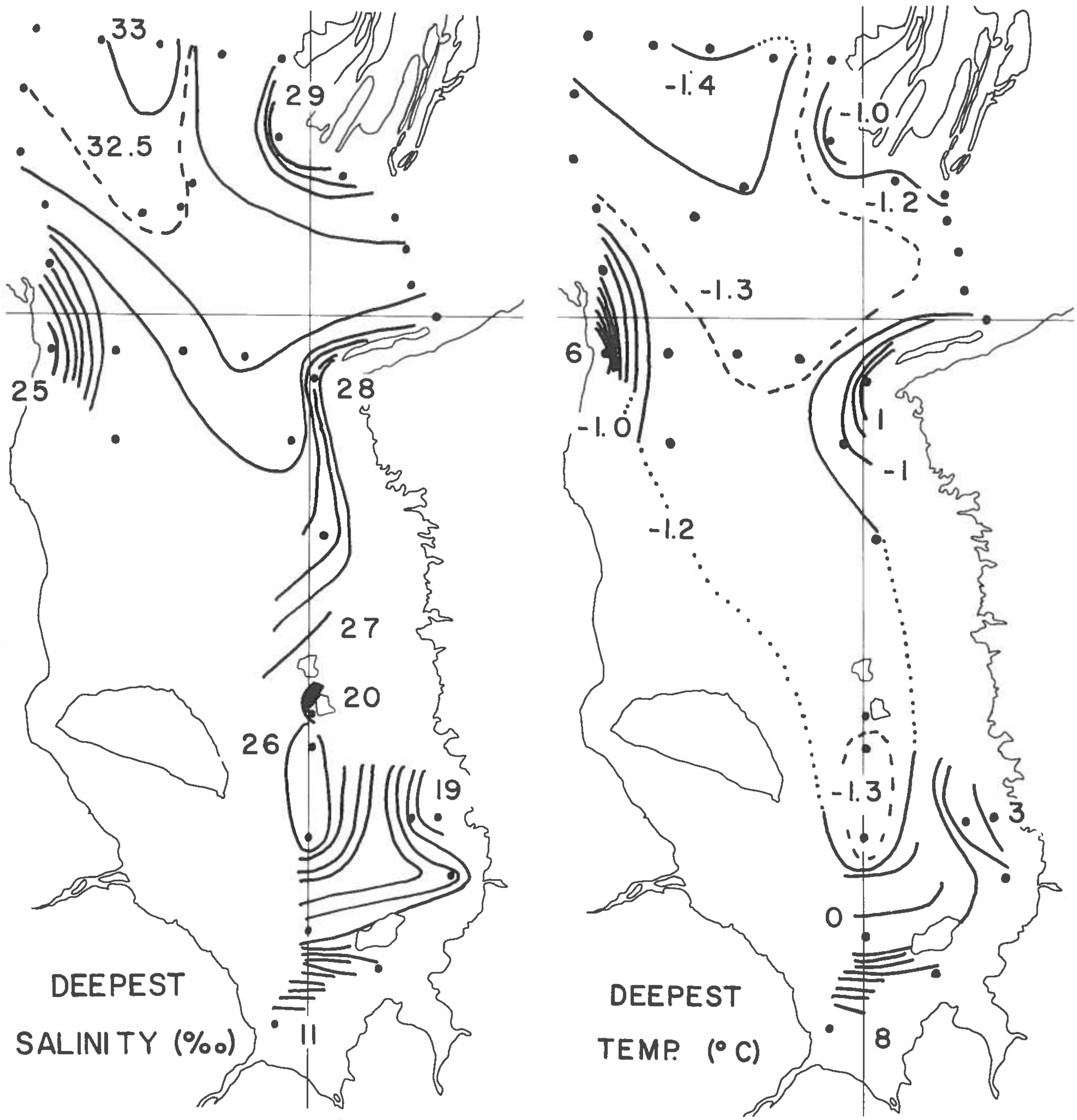

33
29
32.5
25
28
27
20
26
19
11
DEEPEST
SALINITY (‰)
-1.4
-1.0
-1.2
-1.3
6
-1.0
1
-1
-1.2
-1.3
3
0
8
DEEPEST
TEMP. (°C)

Figure 22(f)

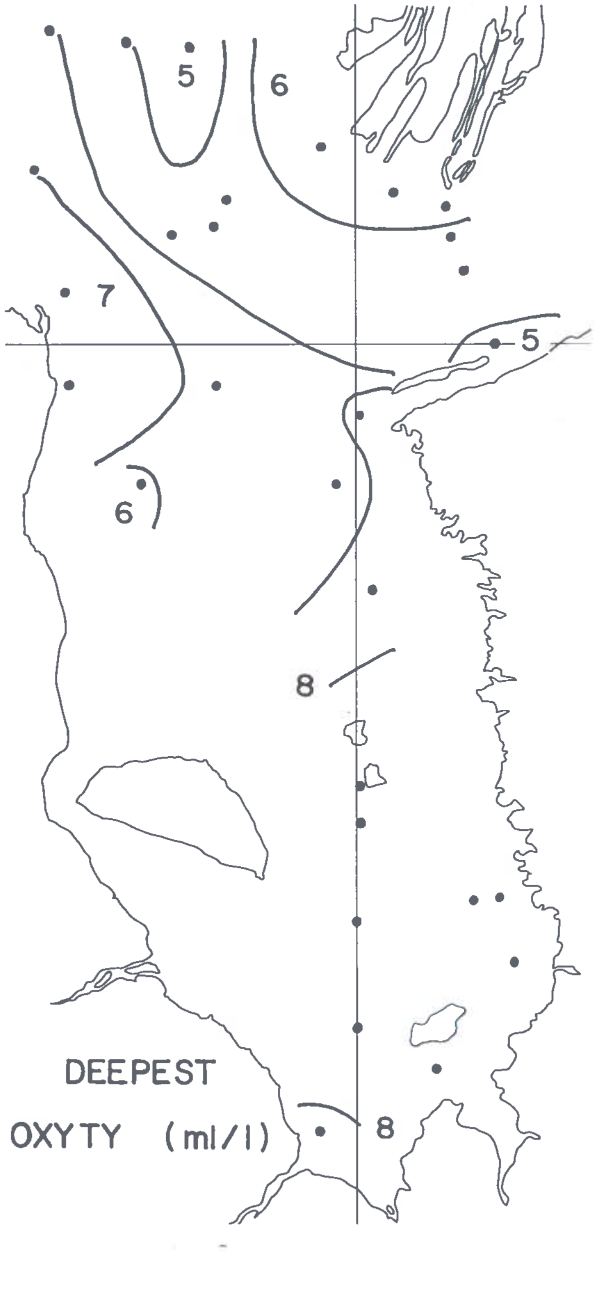
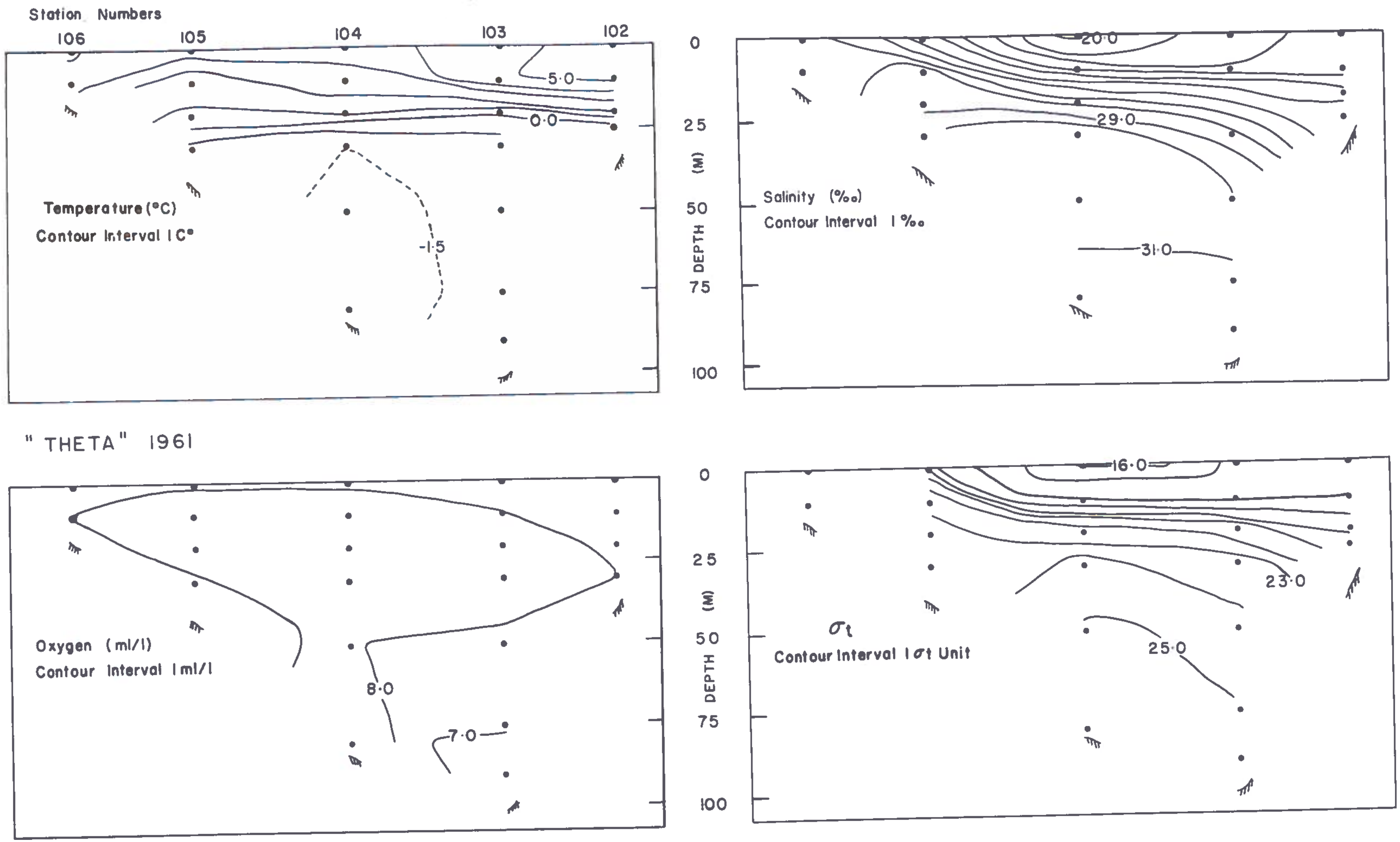


Figure 23. The distribution of oxyty (ml/l) from the "Calanus" data of 1959 at the depth of deepest observation.

Figure 24. The distribution of temperature, salinity, dissolved oxygen and σ_t in a section across the mouth of James Bay. (a) From "Theta" stations 102 to 196. (b) From "Calanus" stations 57, 59, 60, 62 and 63. (c) From "Theta" stations 246 to 250.



Station Numbers
106
105
104
103
102
5·0
0·0
-1·5
Temperature (°C)
Contour Interval 1 C°
0
25
50
75
100
DEPTH (M)
20·0
29·0
31·0
Salinity (‰)
Contour Interval 1 ‰
" THETA" 1961
Oxygen (ml/l)
Contour Interval 1 ml/l
8·0
7·0
16·0
23·0
25·0
σ_t
Contour Interval 1 σ_t Unit

Figure 24(a)

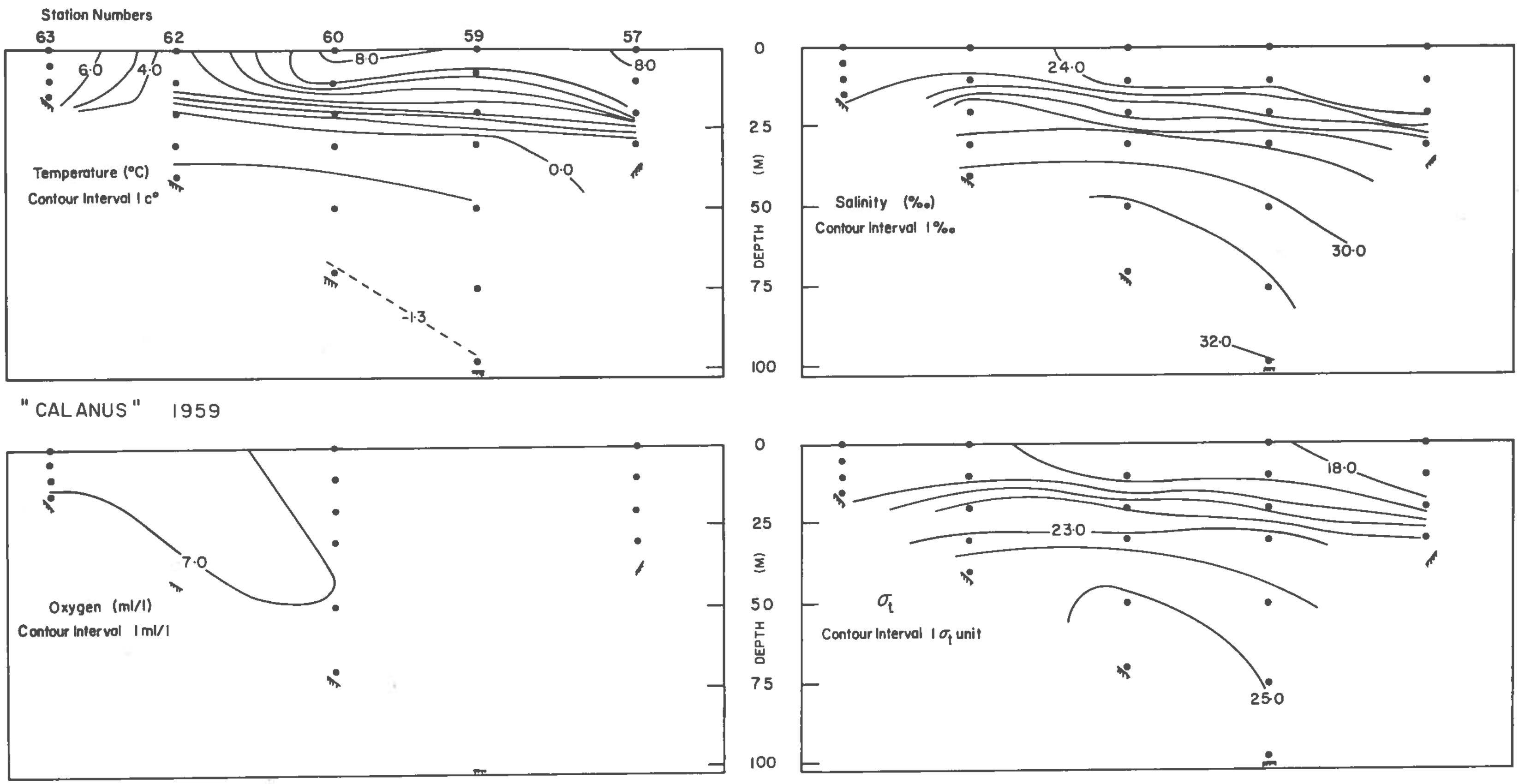

Station Numbers
63
62
60
59
57
6·0
4·0
8·0
8·0
0·0
-1·3
Temperature (°C)
Contour Interval 1 c°
0
25
50
75
100
DEPTH (M)
24·0
30·0
32·0
Salinity (‰)
Contour Interval 1 ‰
"CALANUS" 1959
7·0
Oxygen (ml/l)
Contour Interval 1 ml/l
18·0
23·0
25·0
σt
Contour Interval 1 σt unit

Figure 24(b)

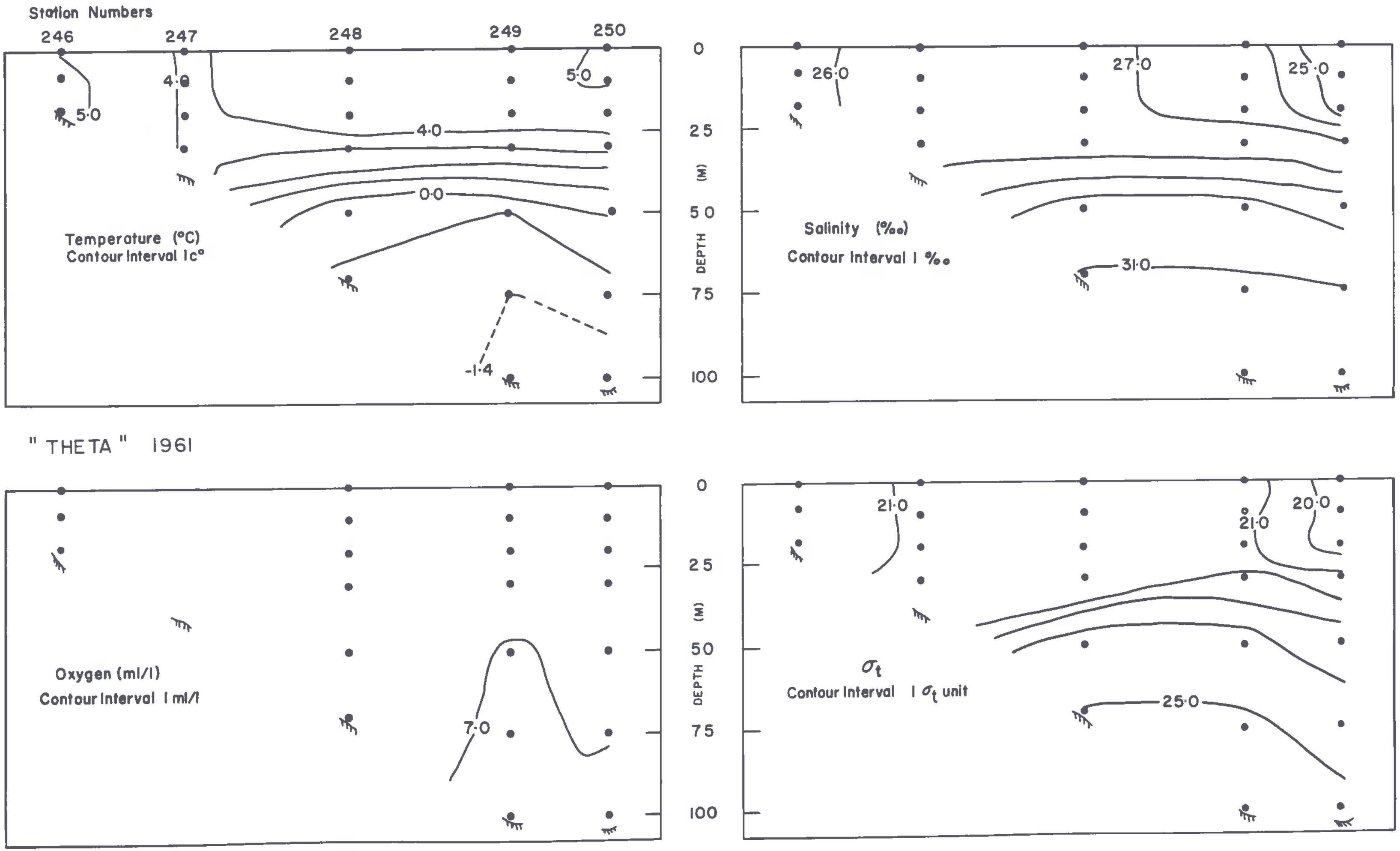

Station Numbers
246
247
248
249
250
5·0
4·0
5·0
4·0
0·0
-1·4
Temperature (°C)
Contour Interval 1c°
0
25
50
75
100
DEPTH (M)
26·0
27·0
25·0
31·0
Salinity (‰)
Contour Interval 1 ‰
"THETA" 1961
Oxygen (ml/l)
Contour Interval 1 ml/l
7·0
21·0
20·0
21·0
25·0
σt
Contour Interval 1 σt unit

Figure 24(c)

The tides in James Bay

by

Gabriel Godin

Contents

1. The data available

James Bay (frontispiece) because of its subarctic climate and its inhospitable shores has not attracted a large permanent population around its rim. Scientific information about its geography, bathymetry and oceanographic characteristics has been acquired in a succession of hesitant steps. Its extensive flats, its hundreds of islands and reefs make most of it virtually inaccessible to any hydrographic vessel. This severe limitation on ship traffic inhibits any motivation for acquiring more extensive soundings with the help of launches, an expensive and time-consuming project. As a consequence the bathymetric information about the western and eastern sides of James Bay consists mostly of blanks. The extent of soundings in the central portion of the bay however, suffice to delineate a relatively deep channel of depths exceeding 25 m, extending from its mouth down to the west side of Charlton Island, while the remainder of the bay remains very shallow.

Temporary tide gauges have been installed at odd stations at irregular time intervals and records of short duration have been accumulated. The majority of stations were established on islands such as Bear, Strutton, Charlton and Stag Islands, as well as inside sand and sediment choked rivers such as La Grande-Rivière, Eastmain, Moose and Albany. None of these observations can be considered of good quality with the exception of those carried out in the Moose River (Langford, 1963). As a matter of fact, the careful observations in the Moose River, which monitored currents and water levels from well outside the estuary up to the Government Wharf in Moosonee at the foot of rapids, help assess the representativeness of the observations gathered at the other stations. For instance the tide takes 2 1/2 hours to travel the 13 nautical miles from Sand Head to Moosonee; therefore, one must expect a lag of the same order of magnitude between the tide recorded at other trading posts located at the head of rivers and the tide present in their respective estuaries, which makes such observations quite unrepresentative of what is happening in James Bay proper. The observations gathered at the island stations should represent the actual tide in James Bay more adequately. Even at that, there exists a peculiar phase difference between the M_2 tide observed at Charlton Island and Strutton Island which might be accounted for by the very sheltered position of Charlton Depot.

2. Cotidal charts using the gauge observations and a one dimensional model

With the help of these rather sparse observations and the use of a simplified one dimensional model of James Bay, it is still possible to gain a fair idea of the progression of the tide inside the bay and of its change in amplitude.

The tidal wave which progresses along the southern rim of Hudson Bay is strongly refracted around Cape Henrietta Maria and enters James Bay as a damped progressive wave. Its advance appears to be considerably retarded in the vicinity of Akimiski Island and it eventually reaches Hannah Bay and Rupert Bay seven hours after it has rounded Cape Henrietta Maria. Its amplitude is larger on the western and southern portions of the bay due to the presence of a degenerate node of the semidiurnal tide on the eastern side of the bay. The mean amplitude of the tide varies from over 90 cm in the western and southern portions to less than 40 cm in the eastern portions near Akimiski Island. Table 1 lists the amplitude and phase of the major constituents of the tide observed at various gauge stations established in James Bay; we include Poste de la Baleine and Winisk to relate these observations to Hudson Bay proper. M_2 is the major lunar semidiurnal constituent; it represents the mean tide. S_2, the solar semidiurnal constituent, adds or subtracts its contribution to that of M_2, creating spring and neap tides. N_2 is of lunar origin and reflects the variable distance of the moon from the earth. We have kept it in parentheses because most likely it has not been properly separated from the other semidiurnal constituents. K_1 and O_1 are diurnal constituents and are related to the declination of the orbits of the moon and of the earth. Their amplitude and phase are irregular and O_1 appears to be twice as small as K_1 throughout the bay. Normally O_1 should be about two thirds of K_1. If there were no friction in James Bay, both of these constituents should have a node just at the mouth of the bay, the node of K_1 being positioned a little further inside the bay. In practice the node is degenerate and somehow O_1 enters the bay considerably weakened compared to K_1. All in all, the diurnal constituents are very much smaller than the semidiurnal constituents thus indicating that the tide throughout James Bay is truly semidiurnal at all times.

Table 1 Amplitude and phase of various tidal constituents observed at thirteen stations around James Bay.

Station	M_2 amp cm	M_2 phase deg	S_2 amp cm	S_2 phase deg	(N_2) amp cm	(N_2) phase deg	K_1 amp cm	K_1 phase deg	O_1 amp cm	O_1 phase deg
Poste de la Baleine	63	232	15	302	9	184	4	25	0	334
Pte. Louis XIV	64	222	20	281	20	191	6	27	1	311
Fort George	66	238	15	329	10	211	8	116	5	51
Eastmain	34	49	6	134	5	349	10	145	2	110
Strutton	55	14	5	124	5	355	10	141	6	90
Charlton	63	48	16	128	12	3	12	128	3	78
Stag	91	80	20	164	16	38	12	145	9	110
Sand Head	94	44	22	126	18	357	16	128	7	63
Ship Sands	71	71	16	155	12	31	15	152	3	64
Moosonee	62	111	14	200	11	68	12	171	2	60
Fort Albany	88	40	22	120	18	0	15	139	4	90
Bear Island	97	219	19	302	19	182	8	62	3	349
Winisk	109	83	29	158	19	45	5	322	3	280

We may use the data presented in Table 1 to draw cotidal charts of the constituents over James Bay. In these charts the lines of equal phase may be considered as delineating the front of the tidal wave as it progresses up James Bay. A phase difference of 60° between two such lines represents about two hours for a semidiurnal tide and four hours for a diurnal tide. The lines of equal amplitudes denoted by dotted lines define areas within which the amplitude exceeds or is less than the value indicated on the rim.

Such isopleths may be drawn in an almost infinite number of ways because of the scarcity and problematic value of the actual observations. It seems wise at this stage to solve the equations of hydrodynamics over a simplified one dimensional model of James Bay. In this way we may predict the mean vertical tide and tidal currents at various sections of the bay assuming that the tidal wave is perfectly reflected at the head. The results of such calculations may be combined with the coastal observations to draw the most plausible system of cotidal and coamplitude lines, besides yielding values of the tidal currents which have not been observed in the bay except around the Moose River. The equations of hydrodynamics in one dimension are

$$\frac{\partial}{\partial x}(Au) + B\frac{\partial Z}{\partial t} = 0 \qquad (1)$$

$$\frac{1}{g}\frac{\partial u}{\partial t} + \frac{|u|}{C^2 H}u + \frac{\partial Z}{\partial x} = 0 \qquad (2)$$

where x is the distance along the bay moving northward,

A is the area of a vertical section of the bay drawn across its width,

B is the width of such a section,

H is the mean depth of such a section,

Z is the amplitude of the tide and

u is the tidal current,

while C stands for the Chézy coefficient which measures the intensity of friction. Since the friction all over the bay must be rather high we take the value of C as $C = 45\ m^{\frac{1}{2}}/sec$.

Because of the friction term $(|u|/C^2H)u$, the phase difference between Z and u varies from section to section and for a sinusoidal oscillation of frequency σ (a given tidal constituent), we may write a solution to (1) and (2) in the form

$$Z=Z_1(x)\ \cos\ \sigma t + Z_2(x)\ \sin\ \sigma t \tag{3}$$

$$u=u_1(x)\sin\sigma t - u_2(x)\cos\sigma t, \tag{4}$$

so that we may write for the x dependence of Z and u:

$$\frac{\partial}{\partial x}(Au_1) = B\sigma Z_1 \tag{5}$$

$$\frac{\partial}{\partial x}(Au_2) = B\sigma Z_2 \tag{6}$$

$$\frac{\partial Z_1}{\partial x} = -\left[\frac{\sigma}{g}\ u_1 - \frac{|u|}{C^2H}\ u_2\right] \tag{7}$$

$$\frac{\partial Z_2}{\partial x} = - \left[\frac{|u|}{C^2 H} u_1 + \frac{\sigma}{g} u_2 \right] \tag{8}$$

u_1, u_2, Z_1, Z_2 are inextricably linked because of the friction. (5) to (8) become useful to us in difference form:

$$(Au_1)_{j+2} = (Au_1)_j + \Delta x \sigma \; B_{j+1} (Z_1)_{j+1} \tag{9}$$

$$(Au_2)_{j+2} = (Au_2)_j + \Delta x \sigma \; B_{j+1} (Z_2)_{j+1} \tag{10}$$

$$(Z_1)_{j+1} = (Z_1)_{j-1} - \Delta x \left[(\sigma/g)(u_1)_j - (|u|_j/C^2 H_j)(u_2)_j \right] \tag{11}$$

$$(Z_2)_{j+1} = (Z_2)_{j-1} - \Delta x \left[(|u|_j/C^2 H_j)(u_1)_j + (\sigma/g)(u_2)_j \right] \tag{12}$$

By dividing the bay into segments of length Δx, (9) to (12) allow us to evaluate Z and u at section $2\Delta x$ apart from the values set at the boundaries. We have in fact, subdivided James Bay into 23 segments of length Δx=10 nautical miles extending from 51°10'N to 55°00'N. The integration has to be performed for each constituent. We restrict ourselves to M_2 and K_1 only, since calculating S_2, N_2 and O_1 would simply be repetitious. We use for M_2 (σ=28.98°/hour) as boundary conditions

$$u=0 \quad \text{at} \quad j=0 \quad (51°10'\text{N}) \tag{13}$$

$$Z=91(\cos \sigma t - 40°) \quad \text{cm} \quad \text{at} \quad j=1 \; (51°20'\text{N}),$$

which is equivalent to taking

$$Z_1=70 \text{ cm and } Z_2=59 \text{ cm at } j=1 \qquad (14)$$

For K_1 (σ=15.04°/hour) we take

$$u=0 \quad \text{at} \quad j=0 \qquad (15)$$

$$Z=15\cos(\sigma t-125°) \text{ cm} \quad \text{at} \quad j=1$$

$$\text{or} \quad Z_1=-9 \text{ cm} \quad Z_2=12 \text{ cm} \quad \text{at} \quad j=1 \qquad (16)$$

Table 2 lists values of B, A and H which have been derived from our schematization of the bay into the 23 subsections; the actual profiles are presented in the appendix. Table 3 contains the results of the integration of (9) to (12) using (13) to (16). These values cannot describe any two dimensional features of the tidal motion and represent averages over a whole section. They are plotted at the points indicated by x and ⊙ in the cotidal charts and they supply considerable assistance in the drawing of plausible and consistent coamplitudes and cotidal lines. We draw similar charts for S_2, N_2 and O_1 using the features that had been delineated for M_2 and K_1 using the one dimensional model. All these charts are shown in Figures 1 to 5.

A side product of the numerical calculations is that they yield values of the mean current at the various sections; these have been noted on the charts of M_2 and K_1. Actual

Table 2 A one dimensional schematization of James Bay indicating the section j, (equation 13) and the width B, the area A and the depth H (equations 1 and 2).

Section j	Width B (km)	Vertical Area A ($\times 10^6 m^2$)	Depth H (m)
0			
1	38		
2		.68	11
3	109		
4		2.07	16
5	118		
6		2.91	19
7	203		
8		3.68	18
9	126		
10		3.09	24
11	125		
12		3.50	44
13	195		
14		3.42	33
15	200		
16		5.80	29
17	175		
18		7.21	37
19	194		
20		9.31	51
21	170		
22		11.44	69
23	245		

Table 3 Values of M2 and K_1 deduced from the one dimensional model of James Bay.

j	M_2 u amp cm/sec 0	M_2 u phase deg	M_2 z amp cm	M_2 z phase deg	K_1 u amp cm/sec 0	K_1 u phase deg	K_1 z amp cm	K_1 z phase deg
1			91	40			15	125
2	26	310			2	35		
3			78	31			15	124
4	31	304			3	34		
5			64	22			14	123
6	37	299			5	33		
7			49	5			13	22
8	45	291			6	33		
9			39	332			11	120
10	61	285			8	33		
11			49	285			9	116
12	56	274			8	32		
13			61	258			7	112
14	58	252			9	31		
15			75	238			4	100
16	36	224			6	29		
17			78	226			3	87
18	32	200			5	27		
19			76	216			2	62
20	30	180			4	25		
21			71	207			2	29
22	28	167			3	24		
23			65	199			2	359

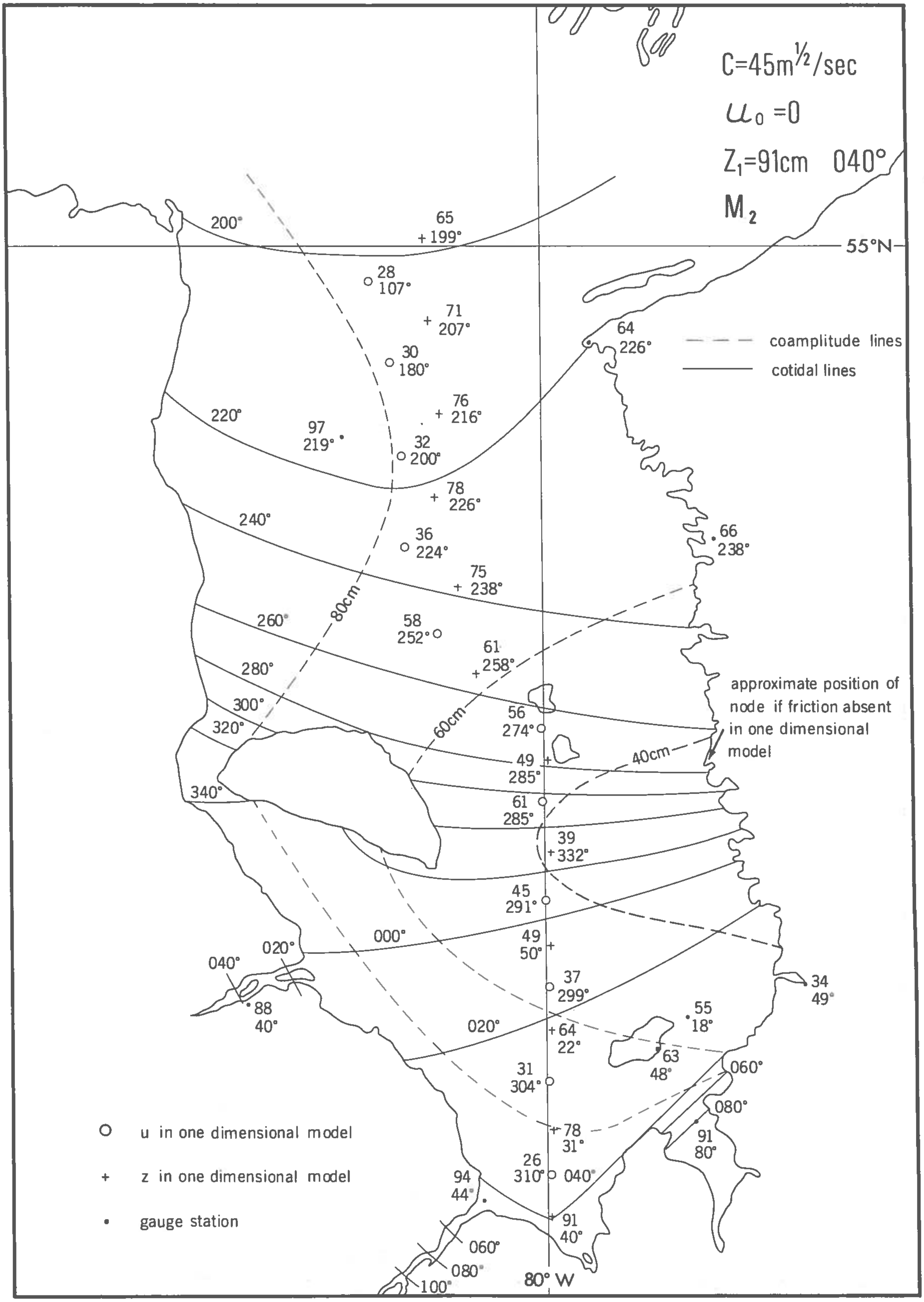


Figure 1 Cotidal and coamplitude lines for M_2. The values observed are squared. x and ⊙ indicate the values of Z and n deduced in the one dimensional model.

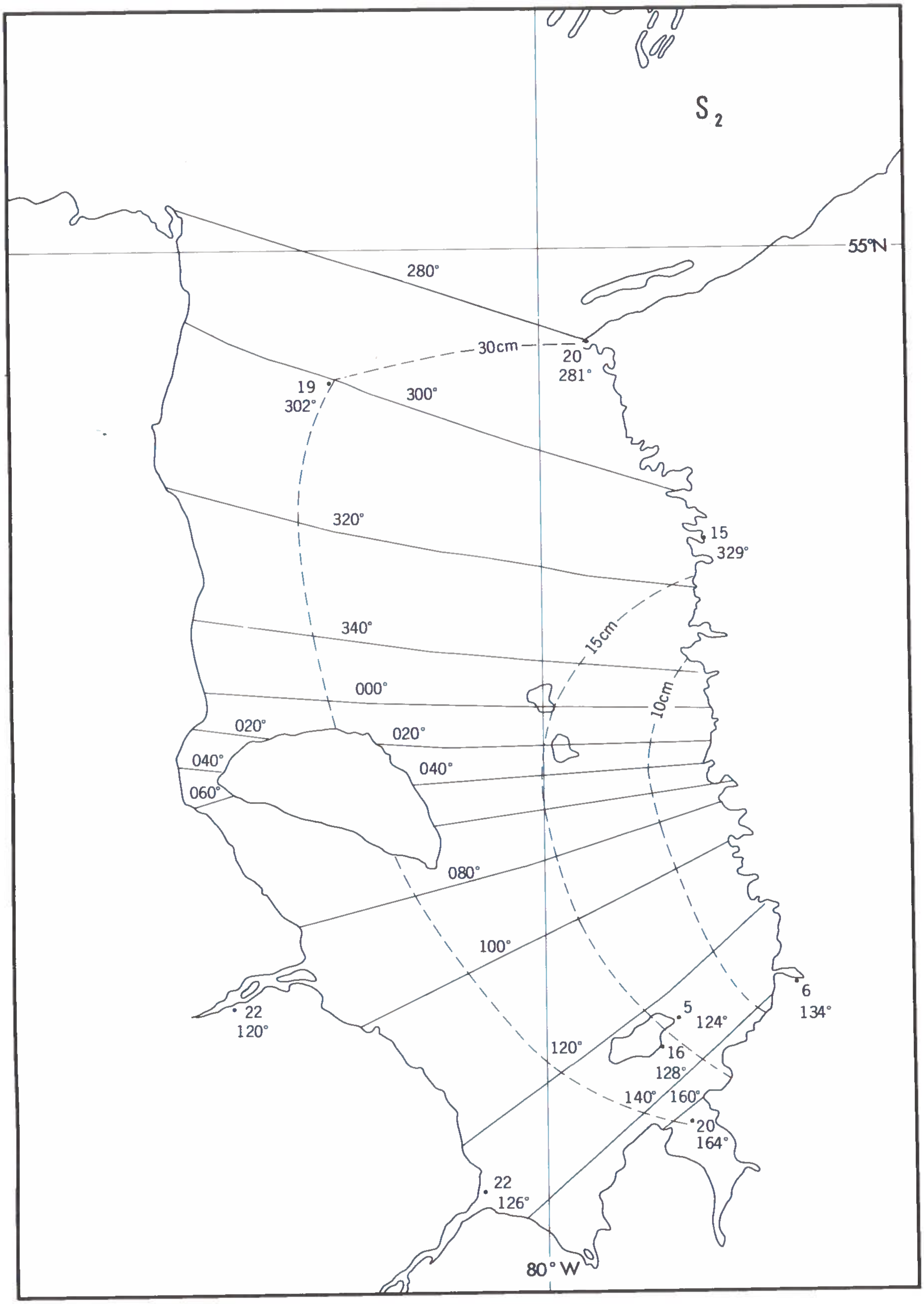


Figure 2 Cotidal and coamplitude lines for S_2.

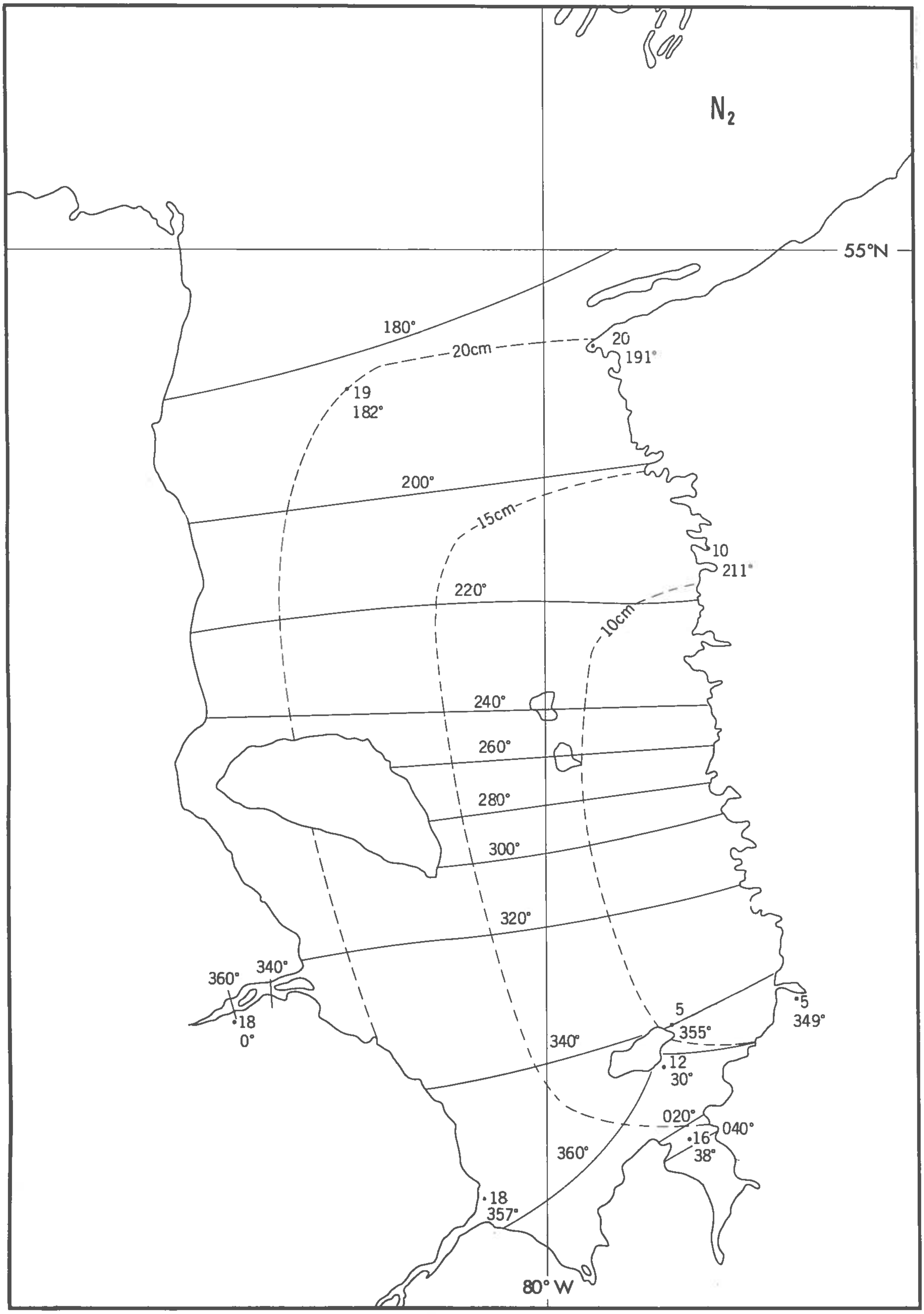


Figure 3 Cotidal and coamplitude lines for N_2.

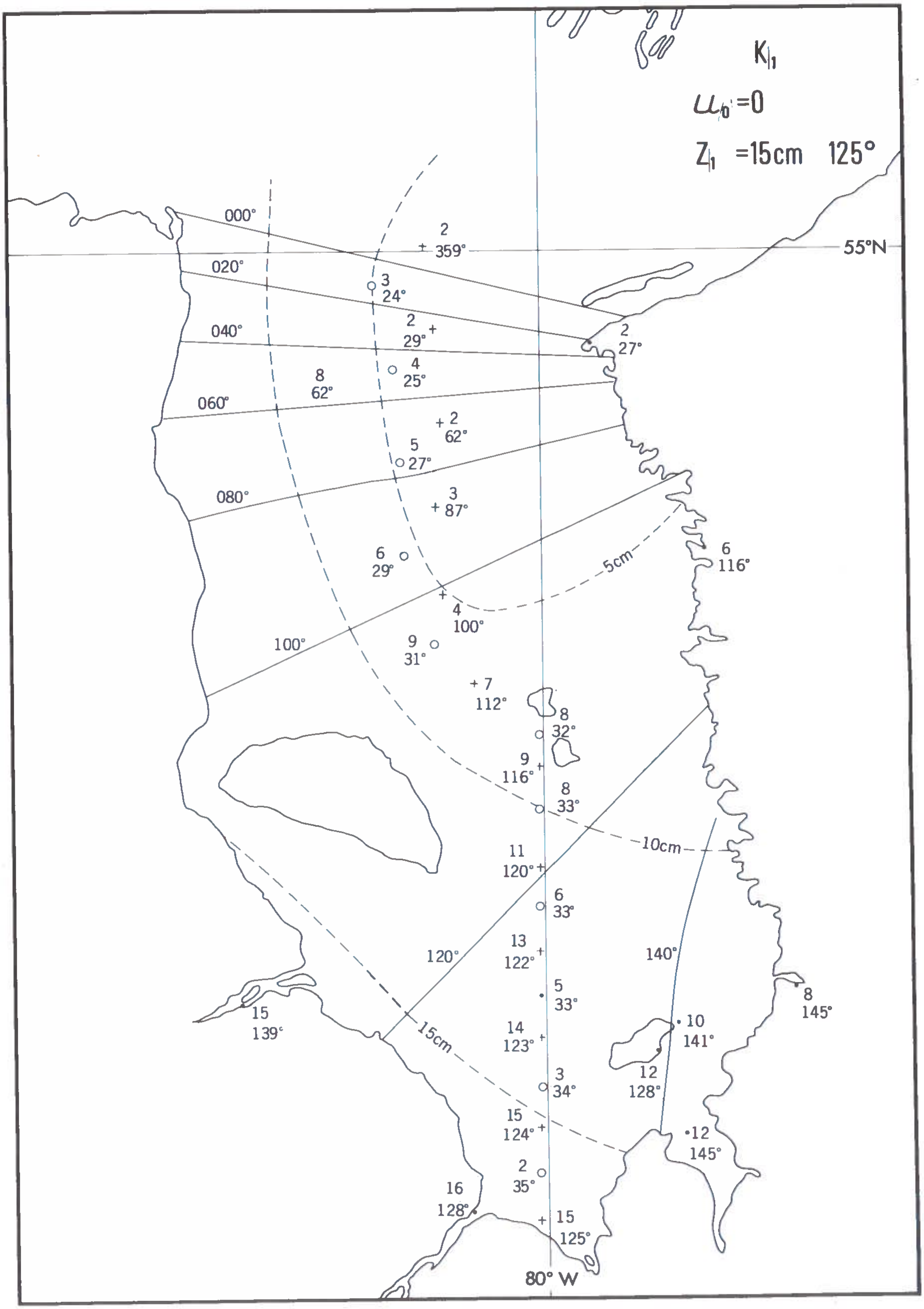


Figure 4 Cotidal and coamplitude lines for K_1.

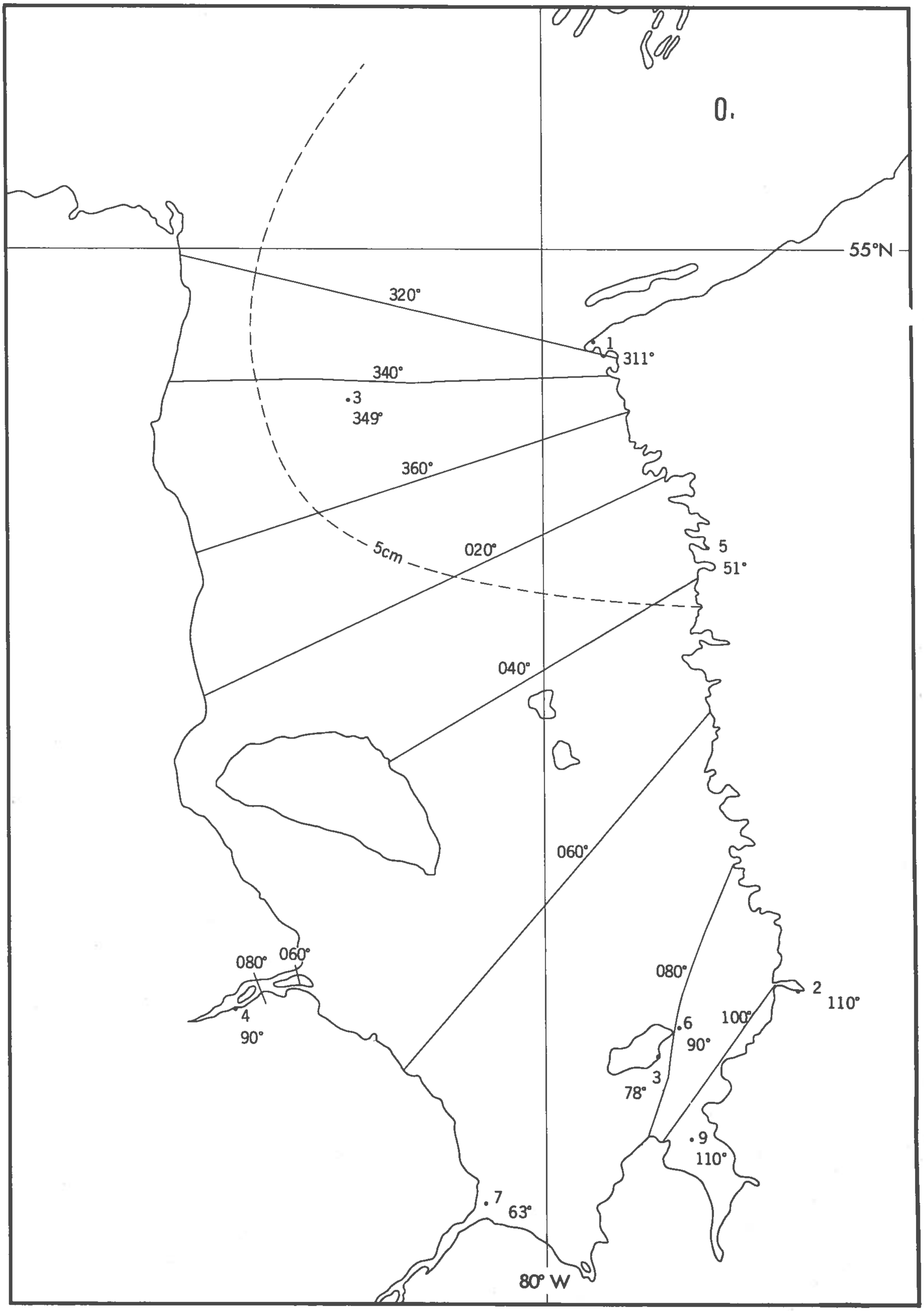


Figure 5 Cotidal and coamplitude lines for O_1.

observations of the currents in James Bay are not available but some currents were measured in the estuary of the Moose River.

Our solution to (1) and (2) depends on our assumption of perfect reflection at the head of the bay; in practice the reflection is rather diffuse and we cannot expect that the current observed in that area should lead the vertical tide by 90° of phase as our calculations suggest.

In the vicinity of Sand Head in the estuary of the Moose River, the vertical tide was observed on the tidal flat while the current was monitored in two neighbouring channels. At station B (Figure 6) the current was exactly 90° of phase ahead of the vertical tide as predicted by the model, but at Station E the current was more like 45° of phase ahead; the current at E is more representative of the current in the bay proper because the channel there is deeper and the flow is much less influenced by friction as it is at Station B. The observed currents are also larger than those predicted by the model, but it must be kept in mind that the model predicts the average current over a whole section while the current observed is present at a given point which happens to be more shallow than the remainder of the section. Moving away from the boundary the currents predicted from the model should be more representative and we notice that they reach their maximum of 61 cm/sec off Akimiski Island for M_2 and 9 cm/sec for K_1. Taking the contribution of S_2 and N_2 into account, the currents caused by the semidiurnal tide can therefore reach a magnitude of 80 cm/sec at spring tide when the moon is in perigee. This gives an

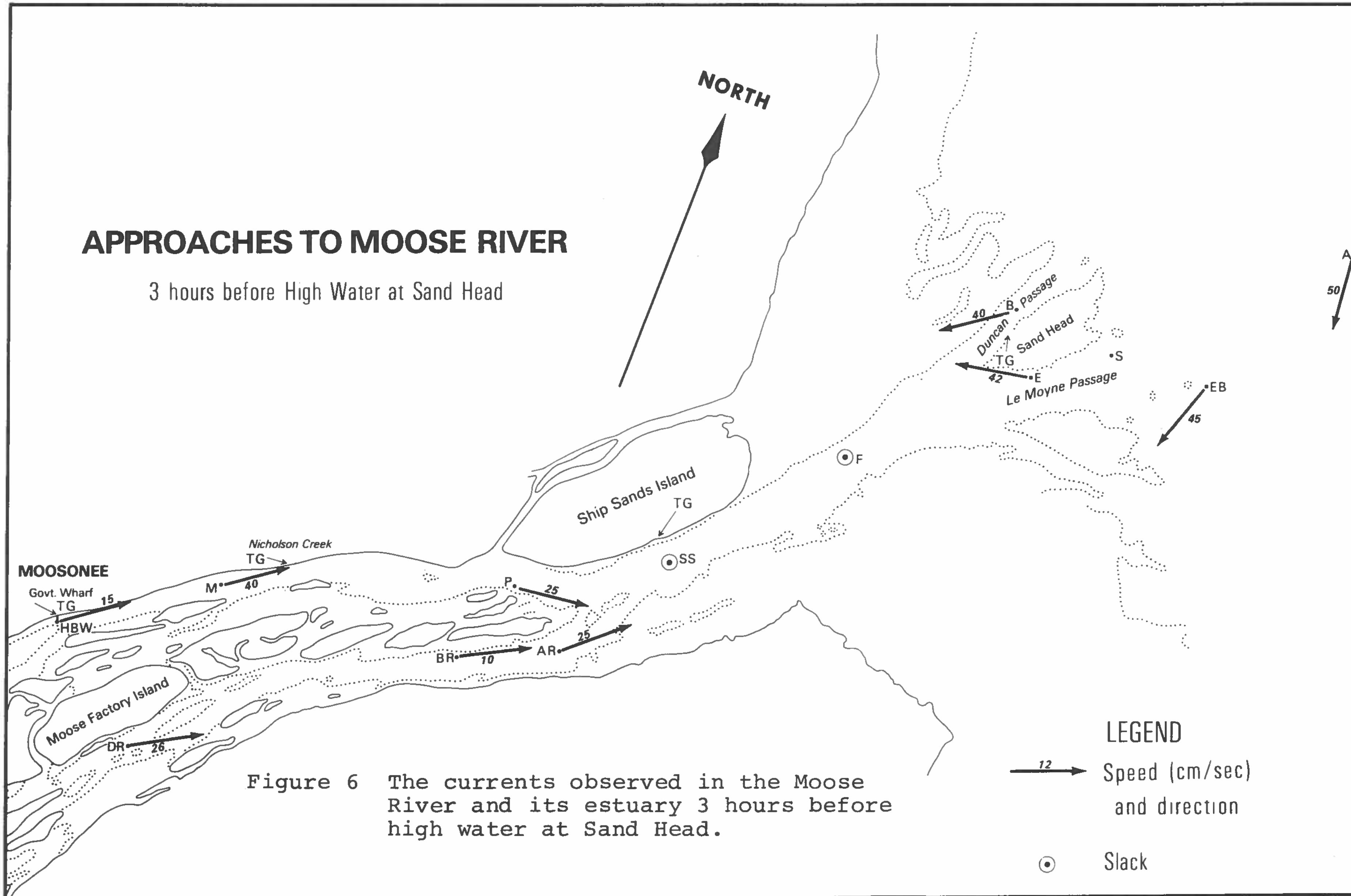


Figure 6 The currents observed in the Moose River and its estuary 3 hours before high water at Sand Head.

indication that the tidal currents over the body of James Bay in the vicinity of Akimiski Island can reach quite a respectable intensity.

3. A detailed view of the flow caused by the tide in Moose River

Up to now we have struggled with a handful of data and a coarse model in order to obtain some idea of the tidal motion in James Bay. It comes as a relief to this bleak situation to have in our possession the results of the detailed survey carried out in the Moose River and its estuary by the Canadian Hydrographic Service (Langford, 1963). In this fashion we can actually witness the flow of water in and out of a tributary of James Bay which most likely is typical of what is happening inside the other rivers emptying into James Bay.

The bed of the Moose River is quite undefined, shallow and cursed with numerous islands and drying flats of mud, sand and boulders. Sand flats nearly block its estuary and only the pressure of its impounded water manages to keep a gully open towards the sea (the Le Moyne Passage). The shores of the Moose River consist in many places of soft materials which are undermined by the ice and rushing waters during the spring freshet; later in the season, these mined areas collapse. Rapids effectively block the river upstream. The Moose River appears as rather typical of the other rivers around the bay and a study of the intricacies of the water movements inside it during a tidal cycle will help give an idea of the actual complexity of the patterns of flow in the vicinity of any of the other tributaries.

In Figures 6 to 9, the solid line delineates the shores and the islands and the dotted line the flats that dry at lower low water. The tide gauge stations indicated by "T.G." were set up at four locations. The first station at Sand Head measures with little distortion the tide that comes in from the main body of the bay. The distance between the first and last gauge amounts to approximately 13 nautical miles. Current meter stations were established at points indicated by the origin of the arrows; they are labelled by letters. We have shown 11 such stations. A circled point at one station indicates that at the moment of observation, no detectable current was noticed (slack water, turn). The arrow shows the orientation and the velocity in cm/sec is indicated. The origin of time is chosen as the moment when it is high water at Sand Head, the station located furthest out in the estuary.

In Figure 6, three hours before high water at Sand Head, which corresponds to nearly 1/4 period of the semidiurnal tide (and therefore approximately to mean water), flood has been established in the estuary while the water is still ebbing in the upper regions of the river; as a consequence we find a region of no motion in the vicinity of Ship Sands Island. This front moves gradually upstream as the tide rises and by the time it is high water at Sand Head (Figure 7) the river is in flood as far as Moosonee. We may notice that the directions of flow in individual channels depends very sensitively on the bottom configuration and that it may change rapidly at times. In the estuary, slack water is already reached in Duncan Passage and in the shallower portions

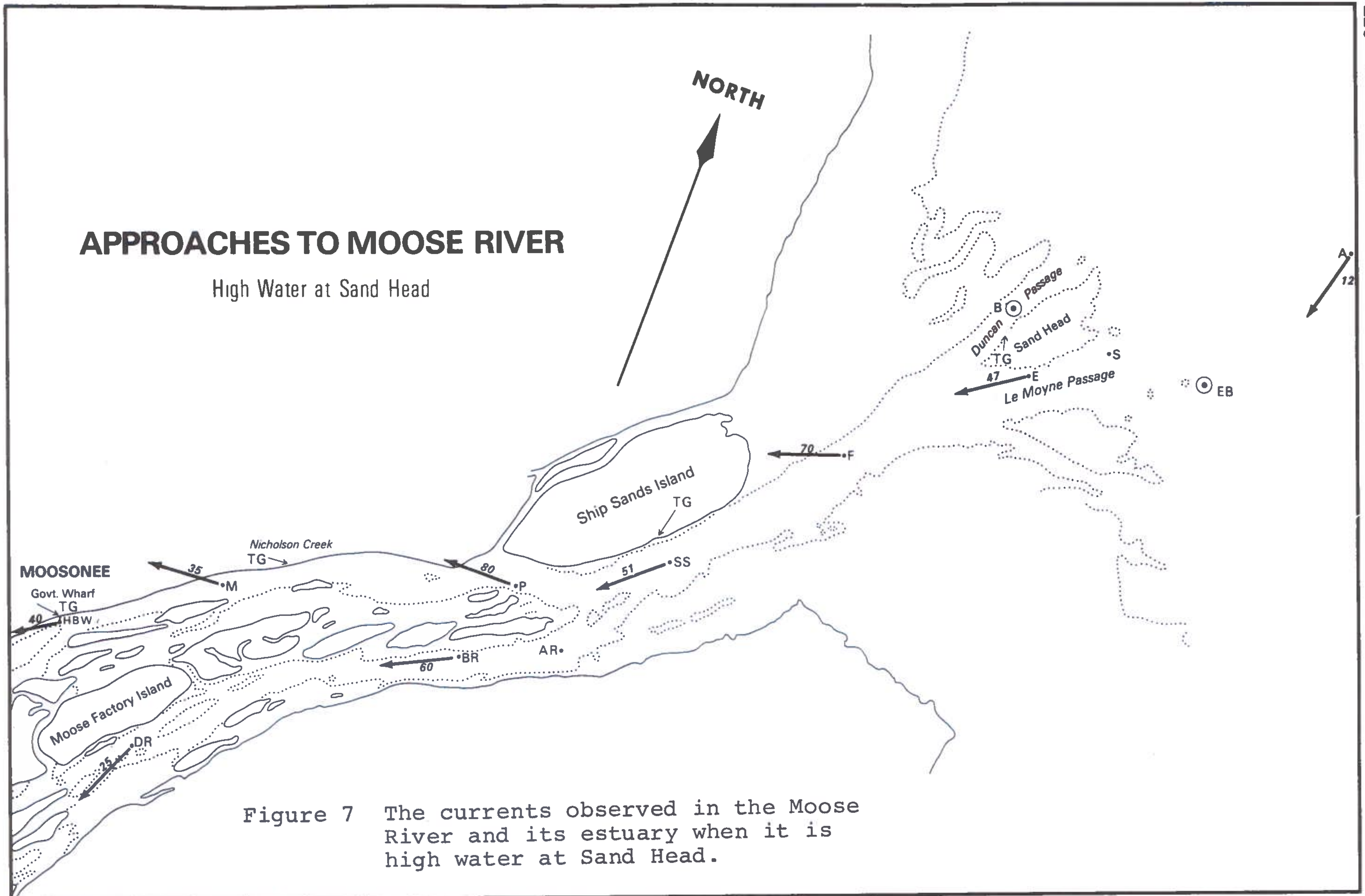


Figure 7 The currents observed in the Moose River and its estuary when it is high water at Sand Head.

of the estuary while it is still flood over the deeper portions. Slack water occurs earlier in shallow areas because of the friction and it will take a further 1 1/2 hours before the current turns in the deeper portions of the estuary. The currents turn clockwise there, while in Le Moyne Passage it slackens suddenly and turns to flood 1 1/2 hours after high water. This zone of slack water progresses upstream as the water level starts lowering, reaching the northern portions of the river first and the southern portions later. Three hours after high water (Figure 8) the river is in ebb. The water level is decreasing, high water having occurred at Ship Sands one hour after Sand Head while the water was still in flood. High water occurred at Nicholson Creek two hours after Sand Head and then it was slack water in its vicinity. Finally high water at Moosonee occurred 2 1/2 hours after Sand Head while the ebb started to take place in its vicinity. Six hours after high water at Sand Head (Figure 9) it is low water at Sand Head and the entire river continues ebbing.

Figure 10 shows the relationship between the water level at the four gauge stations and the current in their immediate vicinity. The time interval between high water and slack is noted since it gives some idea of the phase difference between the current and the vertical tide. Because of the net outflow from the river, the actual time difference between flood and high water is actually longer than the time noted on the graphs.

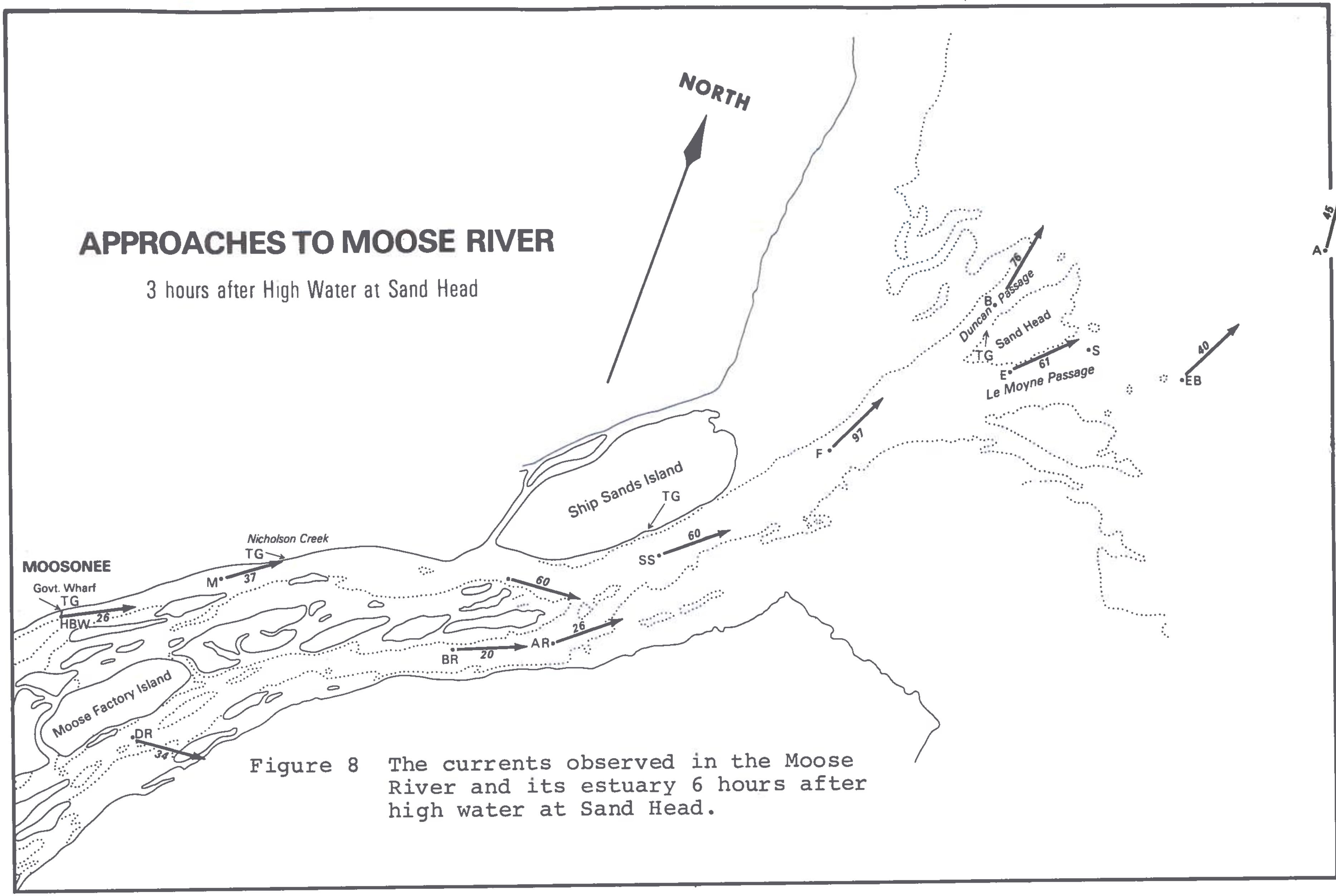


Figure 8 The currents observed in the Moose River and its estuary 6 hours after high water at Sand Head.

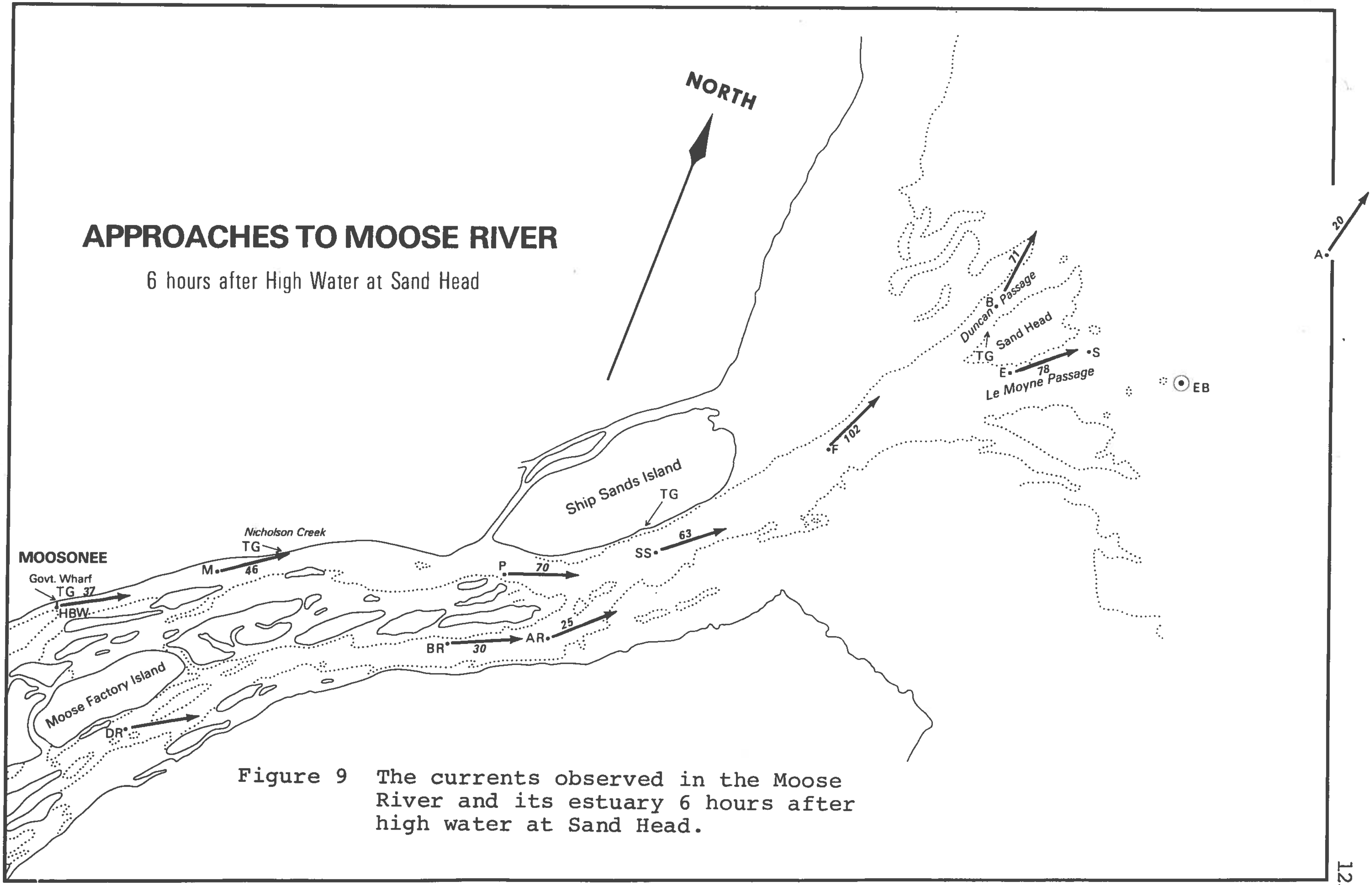
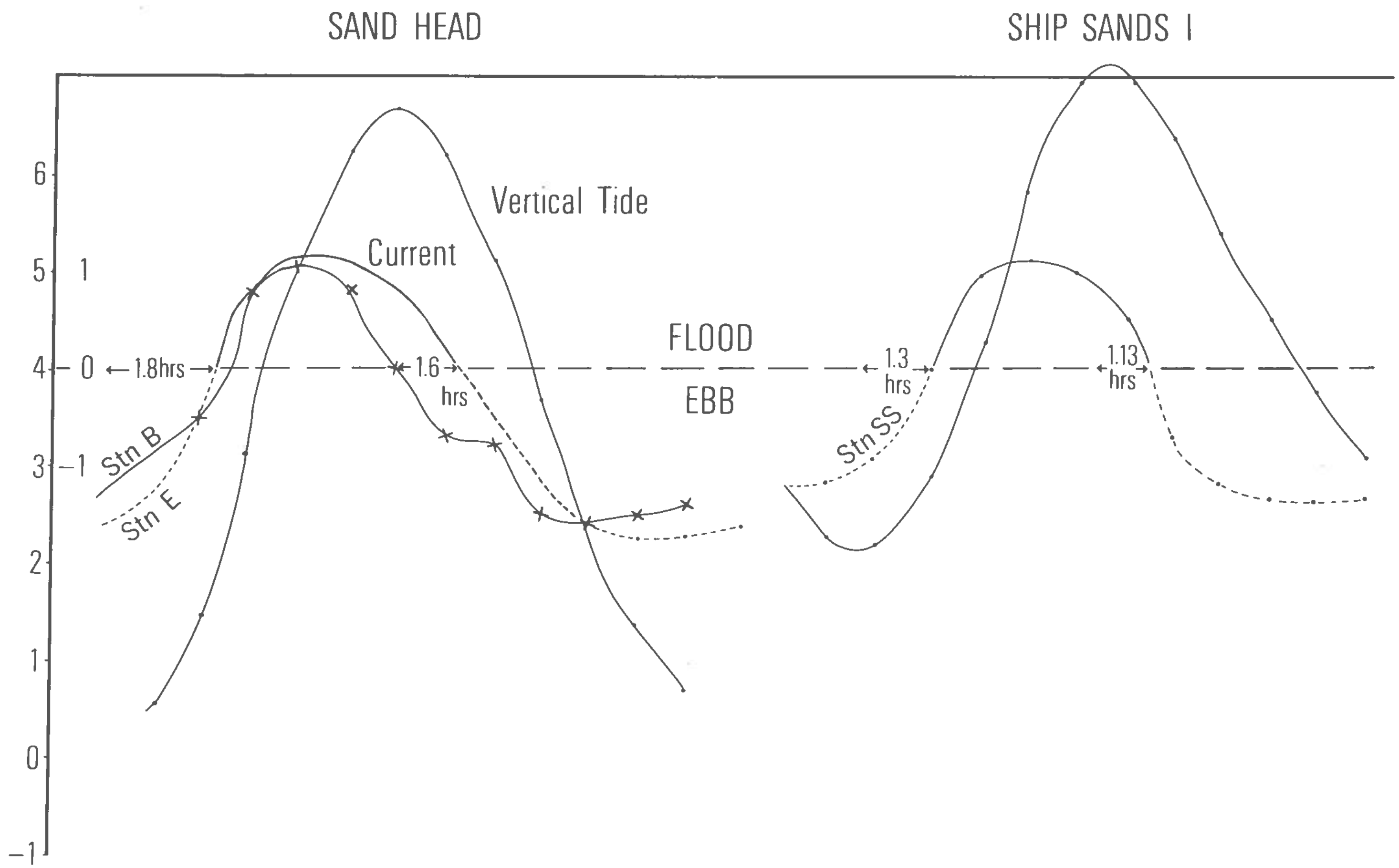


Figure 9 The currents observed in the Moose River and its estuary 6 hours after high water at Sand Head.

Figure 10 The relationship between the vertical tide and the current observed in the vicinity of the observing station.



SAND HEAD
SHIP SANDS I
Vertical Tide
Current
FLOOD
EBB
1.8hrs
1.6
hrs
1.3
hrs
1.13
hrs
Stn B
Stn E
Stn SS
6
5
4
3
2
1
0
−1
1
0
−1

Figure 10(a)

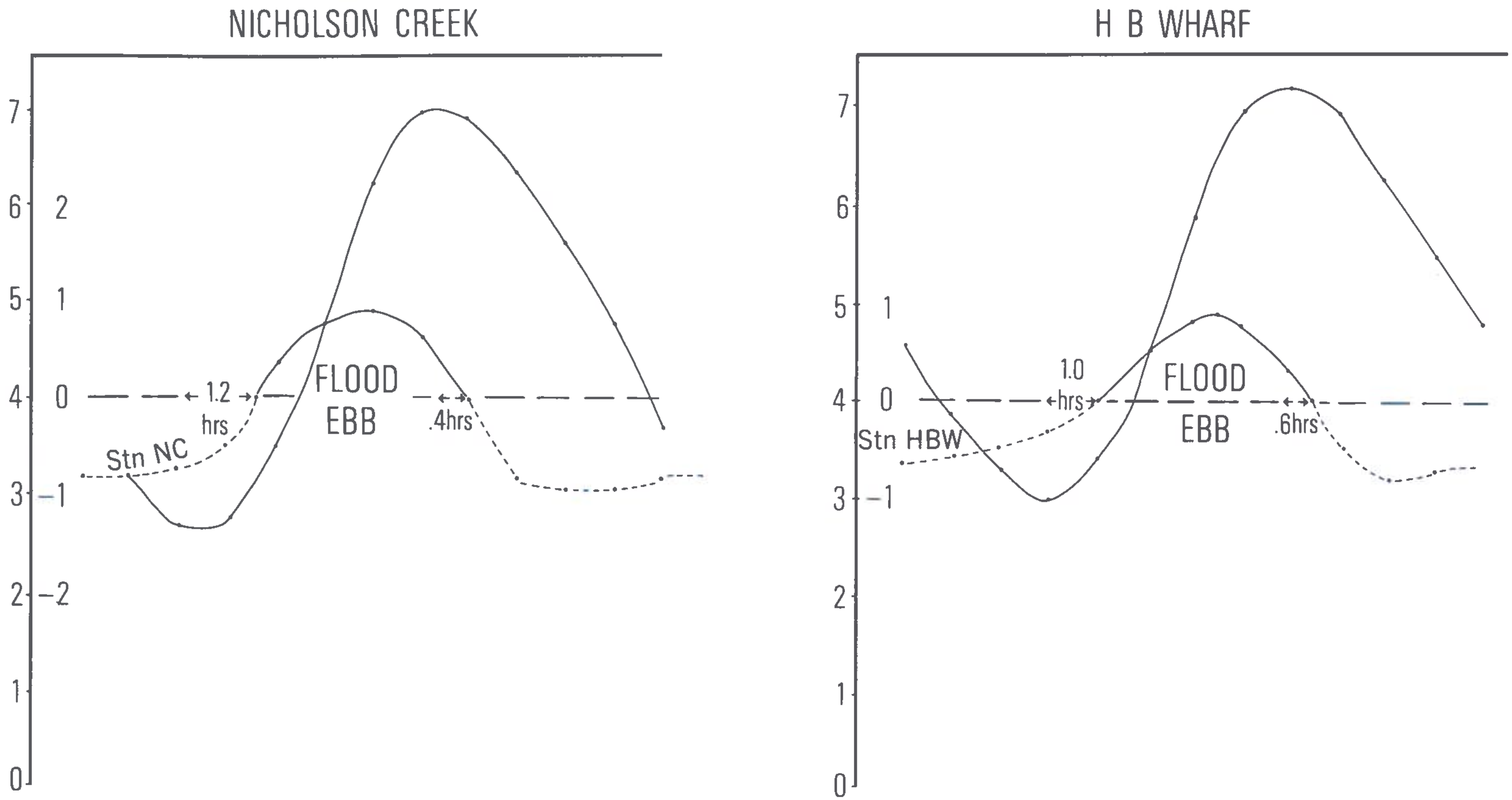

NICHOLSON CREEK
H B WHARF
FLOOD
EBB
1.2
hrs
.4hrs
Stn NC
1.0
hrs
.6hrs
Stn HBW
0
1
2
3
4
5
6
7
-1
-2

Figure 10(b)

4. Irregularities in the tidal regime

Two factors may affect the regularity of the tides: the spring freshet and weather disturbances. During April-May ice breaks up in the rivers surrounding James Bay and their discharge increases abruptly. An increase in discharge inhibits the penetration of the tide into the river and a wall of fresh water may extend a fair distance into James Bay where the tide will be reduced in range and where the mixing of salt and fresh water will take place at an intensified rate. This damping of the tide by a strong current is indicated by the solution of the equations of hydrodynamics (1) and (2) if we represent the current by

$$u = u_o + u_1 \cos (\sigma t - \alpha), \tag{17}$$

where u_o represents the steady current due to the discharge of fresh water and the second term represents the oscillatory term contributed by the tide. Assuming that the tidal current is never strong enough to cause any flood current,

$$|u|u = u^2 \simeq u_o^2 + \tfrac{1}{2}u_1^2 + 2u_o u_1 \cos(\sigma t - \alpha) \tag{18}$$

for which (1) and (2) have a solution of the form

$$u,h \sim \exp i\sigma t \exp \pm \frac{i\sigma}{\sqrt{gD}} \sqrt{1 - \frac{2igu_o}{C^2 \sigma D}}\, x \tag{19}$$

There is a damping term in x of the form

$$\exp - \left[\frac{\sqrt{2u_o\sigma}}{CD} \sin\left(\frac{1}{2} \text{ arc tan } \frac{2u_o g}{C^2 \sigma D} \right) \right] x \tag{20}$$

which would vanish if u_o were zero.

This situation must occur with various degrees of intensity depending on the strength of the freshet. At that time the water will be fresh in the rivers, the overall salinity of James Bay in the vicinity of their estuaries will be significantly reduced and processes of mixing will be intensified but will take place further away from the shores. The stronger currents will carry sediments into the bay at an accelerated rate, particles of a larger size will be dragged along and most probably during this time the major portion of the nutrients fed into James Bay over the course of one year will be carried into it. Observations on the Moose River support this conjecture.

Weather disturbances must also have a marked effect on the motion of water on account of the general shallowness of the area and the presence of modes of resonance in the bay of a period of only a few hours. James Bay might respond dramatically to the passage of some of the cold fronts which are frequent and most severe during early summer and the fall and to the passage of depressions. Rupert Bay, Hannah Bay and the portion of James Bay south of Akimiski Island have minimodes of their own which could be excited by specific patterns of wind and pressure. During the course of a weather surge, the level of water will undergo irregular oscillation and transient currents will be created. Masses of salt water could invade the rivers. The non-linear interaction of the surge with the tide may enhance or decrease the water level depending on fortuitous combinations of pressure gradients and wind stress. Surges are most likely to occur during the storm seasons, namely between September to December and April to June. A surge, a mild one, was observed on the Moose River while tidal observations were being carried out; it occurred between October 12 and October 14, 1963 and had a height of 120 cm with a period of 30 hours. The accidental detection of such a surge indicates that these phenomena are probably quite frequent in the James Bay area (Figure 11).

5. Changes caused by the regulation of the Nottaway, Broadback and Rupert Rivers

Once the NBR project is completed, the tide in Rupert Bay will not be inhibited during the spring freshet as it normally is at that time of the year; this implies that a more regular tidal regime will tend to prevail in and around Rupert Bay throughout the year. The zone of mixing of fresh water and salt water which must wander appreciably inside Rupert Bay at the spring freshet will no longer suffer such fluctuations and will be restricted to an area defined by the intensity of spring and neap tides and the mean value of the regulated outflow from the NBR complex. Consequently, there will be fewer localized short-range fluctuations in density, salt and sediments. The net amount of nutrients brought into Rupert Bay will be reduced since the larger currents necessary for their intensified transport will no longer have any occasion to prevail: the migratory birds

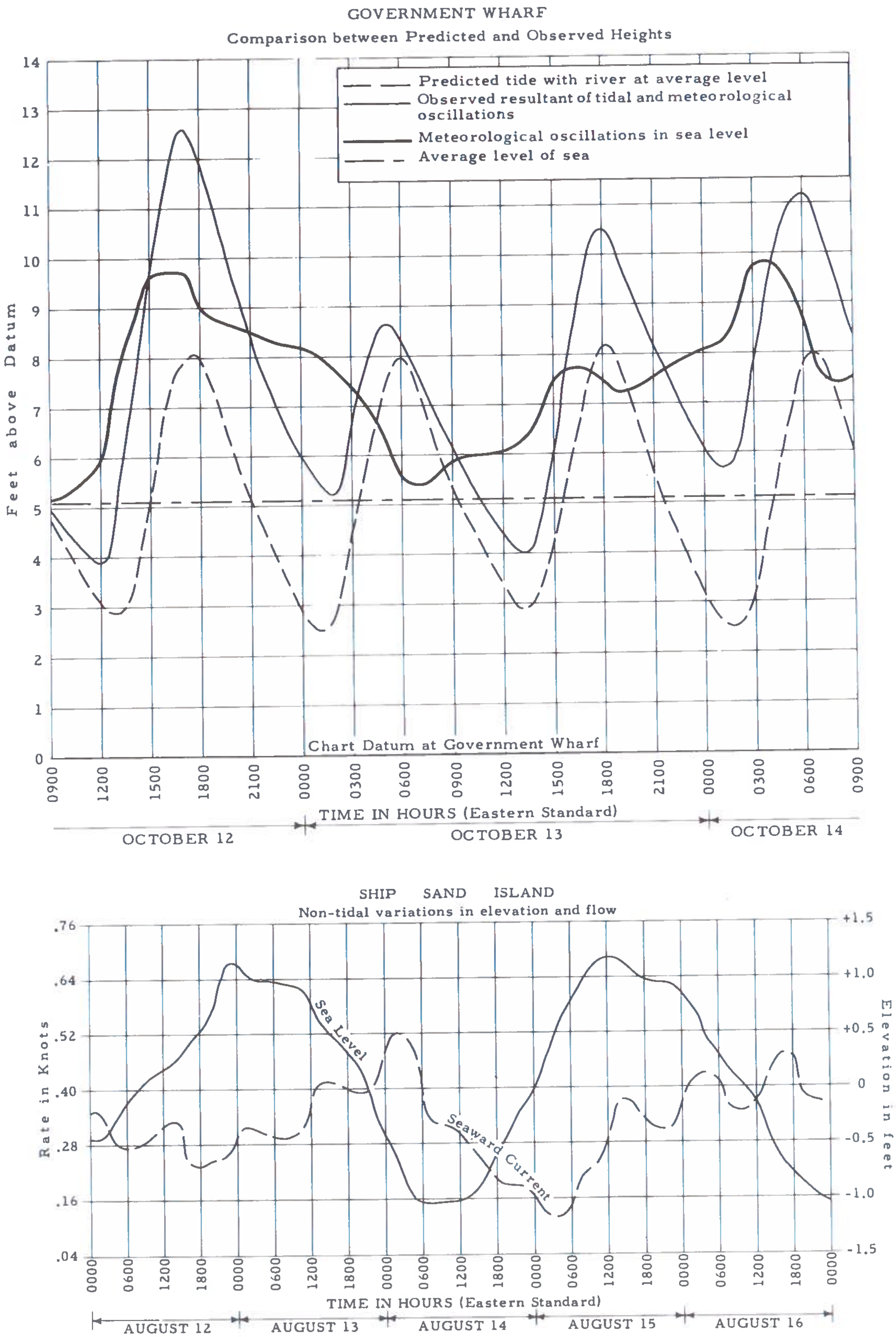


Figure 11 A surge observed in the Moose River during October, 1963.

which feed in the bay for a short time might choose to move to better feeding grounds while the local sea life which is already at a low level will be further reduced.

The probability of occurrence of surges will in no way be affected by the more regular outflow from the NBR complex. The primary factors controlling surges are wind strength, pressure patterns and pressure fluctuations; changes in the density of the air and of the water affect them too but imperceptibly compared to the other factors.

6. Other sources of energy in Hudson Bay and Hudson Strait

Besides the obvious sources of hydroelectric power which it is now intended to tap around James Bay, another potential source of energy could be exploited in the New Quebec with a minimum of ecological upheaval: the very large tide in Ungava Bay. It is a little known fact that the tidal range in the vicinity of Leaf Basin in Ungava Bay is as large as in Minas Basin in the Bay of Fundy and that at some times of the year the tide in Ungava Bay is larger than in the Bay of Fundy. In Ungava Bay the tidal range increases southwestward and reaches its maximum in the vicinity of Leaf Basin and the Koksoak River. Not only is the tide very large but numerous fjords and basins exist in that area which would be most suitable as reservoirs for tidal energy. Not only do they have an appreciable storage capacity but the width of their mouth is relatively small so that construction of a dam across them would not be at all prohibitive. As well, the depths in the entrances are smaller than those which would have to be faced in the Bay of Fundy.

The exploitation of the tidal energy in these basins would not preclude in any way the conventional exploitation of the upper waters of the rivers emptying into some of them for the production of hydroelectric power. As a matter of fact the two modes of exploitation would rather complement each other since the construction of hydroelectric plants upstream would reduce the flow of sediments into the tidal basin and this would prolong the useful life of the tidal plant. Such an arrangement is not possible in James Bay because of the reduced range of the tide there.

The tidal power which can be extracted from some basins in Ungava Bay may be estimated using the formula (Godin, 1969)

$$P = \frac{\rho g S_o \; [(2h_o)^2 + (1/3)e^2]}{T} \qquad (21)$$

where ρ = density of the water, S_o = the area of the basin, h_o = the amplitude of M_2, $1/2e$ = the amplitude of $N_2 + S_2 + K_2$ and T = the tidal period, 12.4 hours. The equation (21) is an upper limit of the value of the power; in practice a tidal plant can extract at most 30 percent of this power. Table 4 lists the potential power output of some basins in Ungava Bay.

Table 4

Potential tidal power for some sites in Ungava Bay.

Basin	Surface area S_o	M_2	$S_2+N_2+K_2$	30 percent of potential tidal power
	$10^7 m^2$	m	m	Megawatts
Payne River	11.4	4.12	2.20	617
Leaf Basin	45.2	4.33	2.28	2450
Koksoak River	5.7	4.09	2.12	274
George River	8.0	3.37	1.68	258
Abloviak Fjord	4.4	2.14	1.73	57

The total assessed power amounts to 3650 MW and compares with the whole power potential of the Grande-Rivière. In addition, the tidal power output is highly reliable on a yearly basis (Bernshtein, 1965) while fluctuating from day-to-day and its yearly regularity could be used to compensate the yearly variabilit of the output of the rivers exploited around James Bay. This would involve integrating the network of the Ungava Bay project with the NBR, Grande-Rivière and Eastmain River projects; this would be most natural to do in any case, simply to take advantage of the networks already established.

7. References

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Godin, G. 1969. Theory of the exploitation of tidal energy and its application to the Bay of Fundy. J. Fish. Res. Bd. Canada 26:2887-2957.

Langford, C.J. 1963. Moose River and approaches. Survey of tides, currents, density and silt. Canadian Hydrographic Service. Ottawa, Canada.

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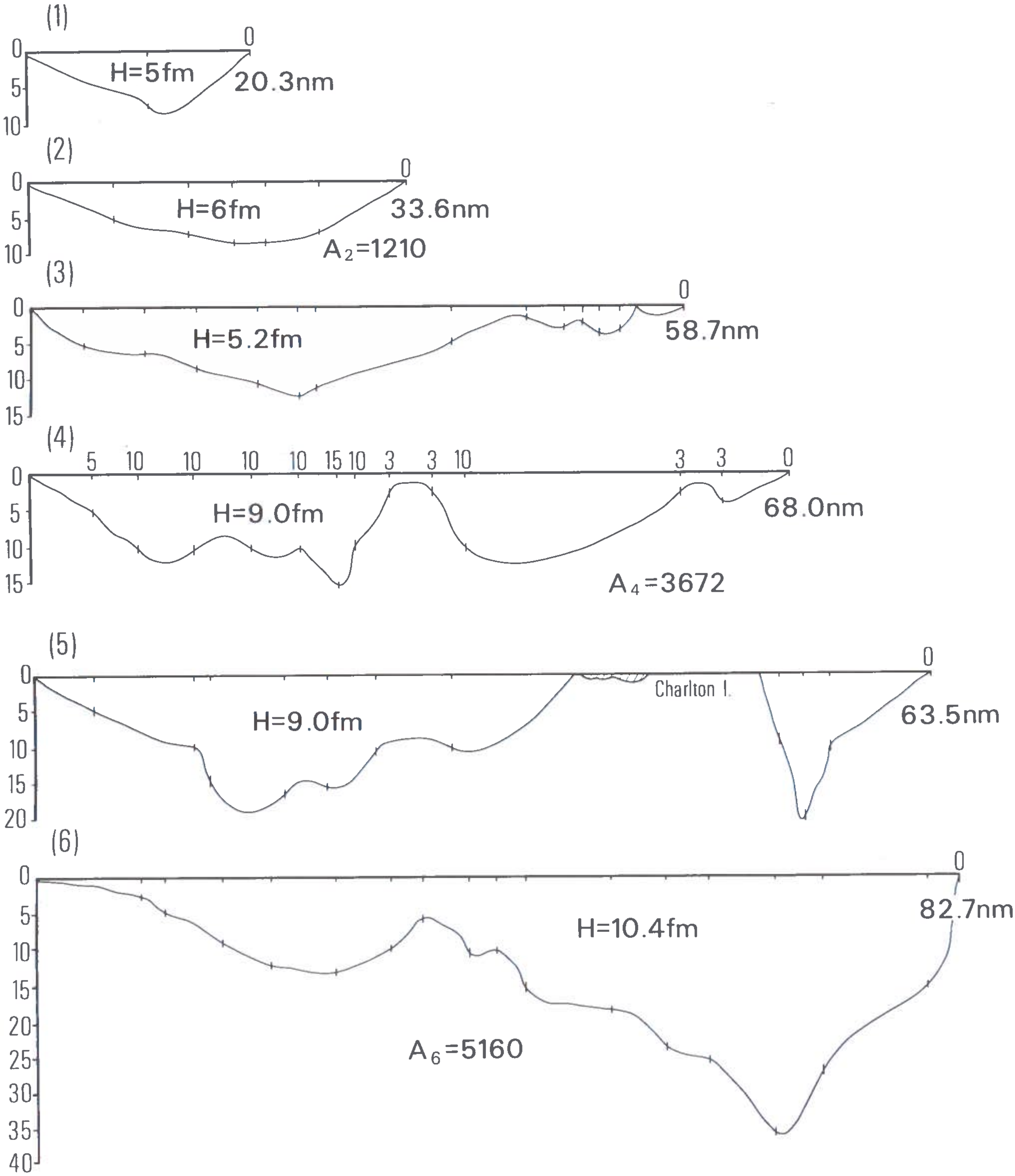
8. List of figures and tables

9. Appendix

Pictorial presentation of the profiles derived from the schematization of the bay (Table 2).



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(2)
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33.6nm
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(3)
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(4)
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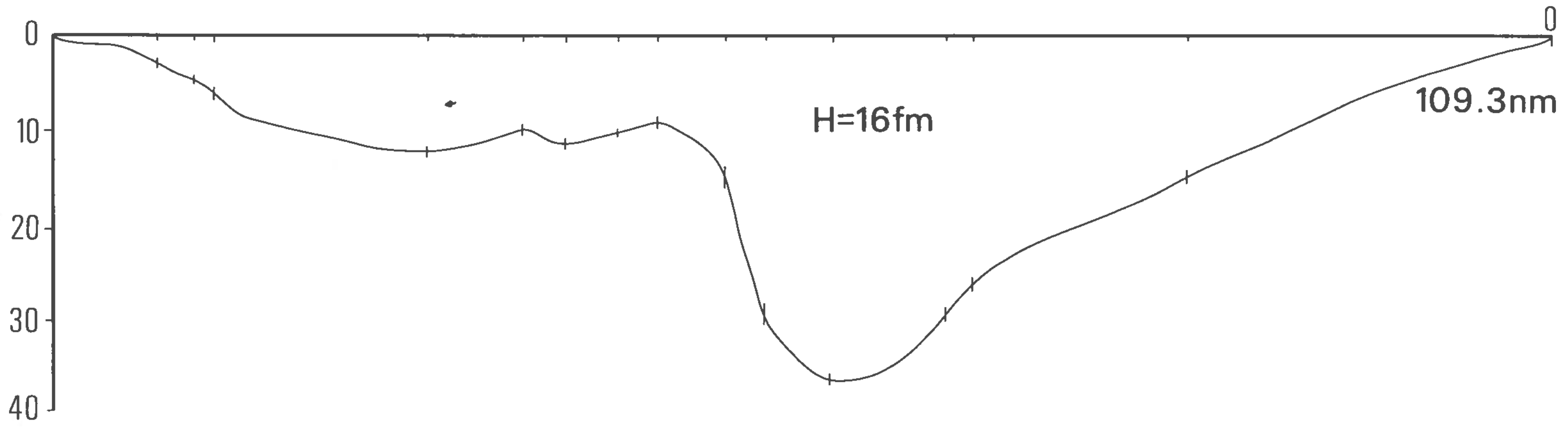

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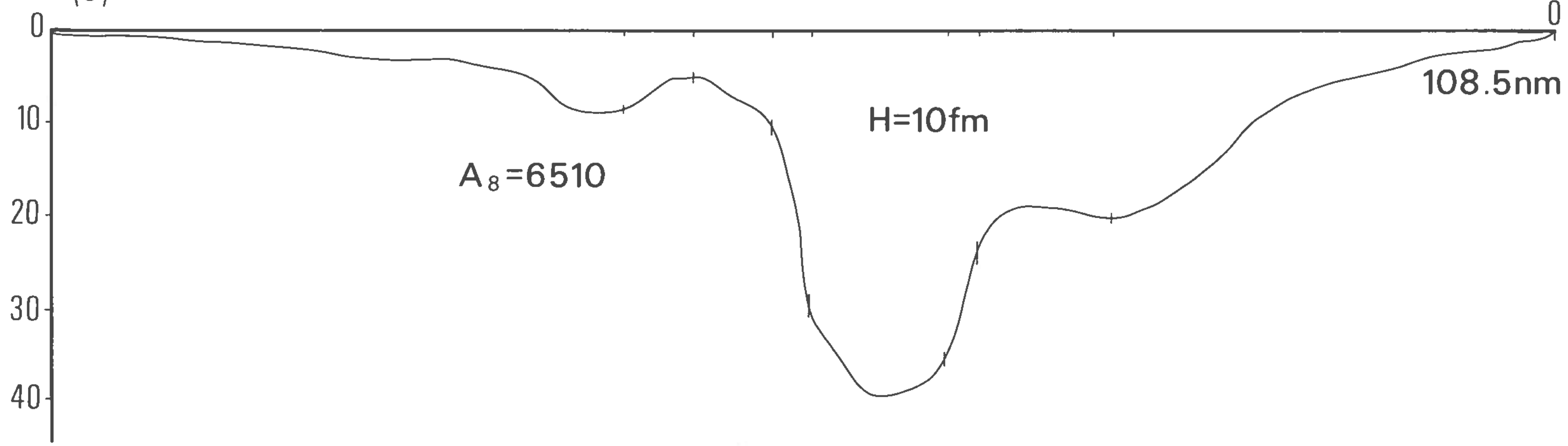

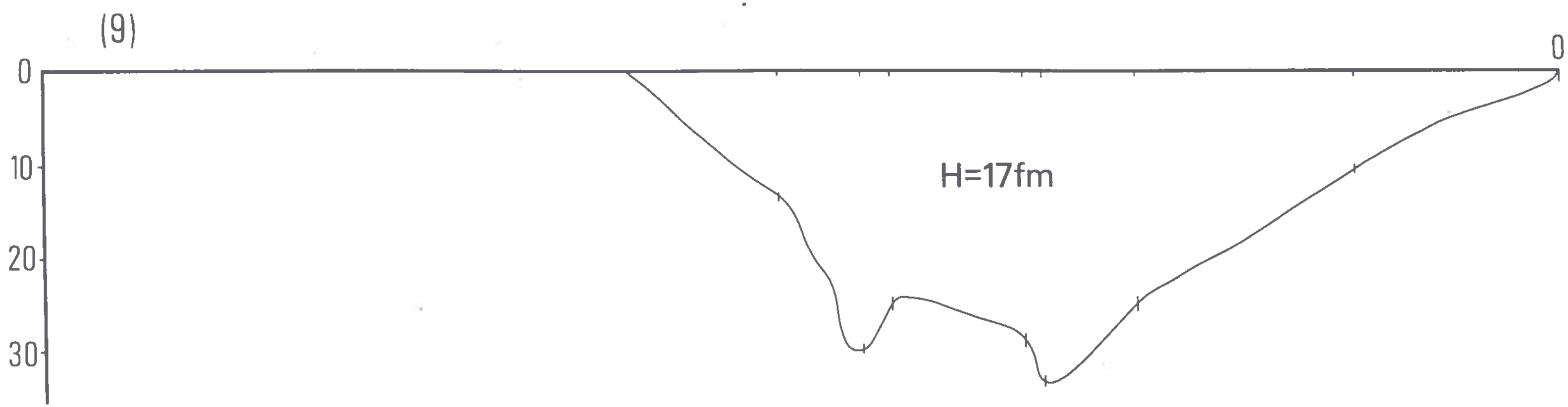
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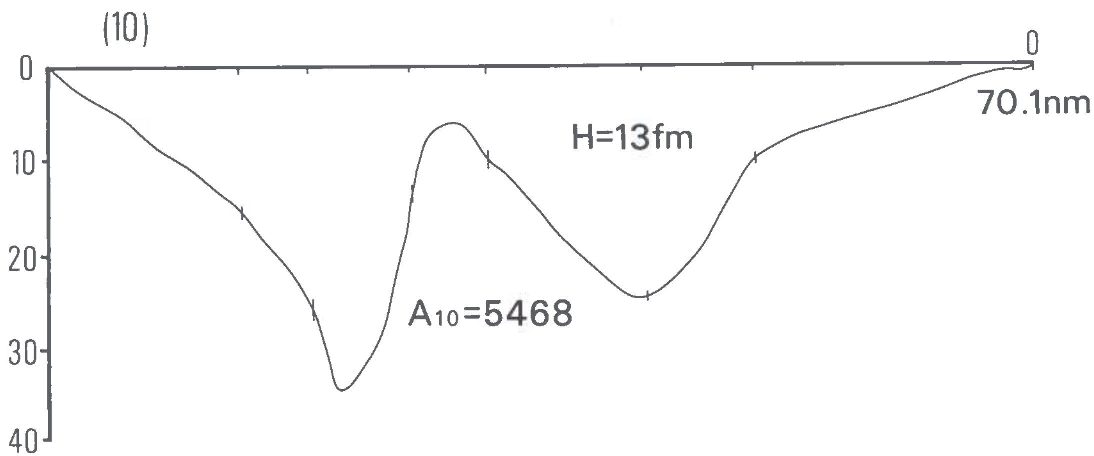
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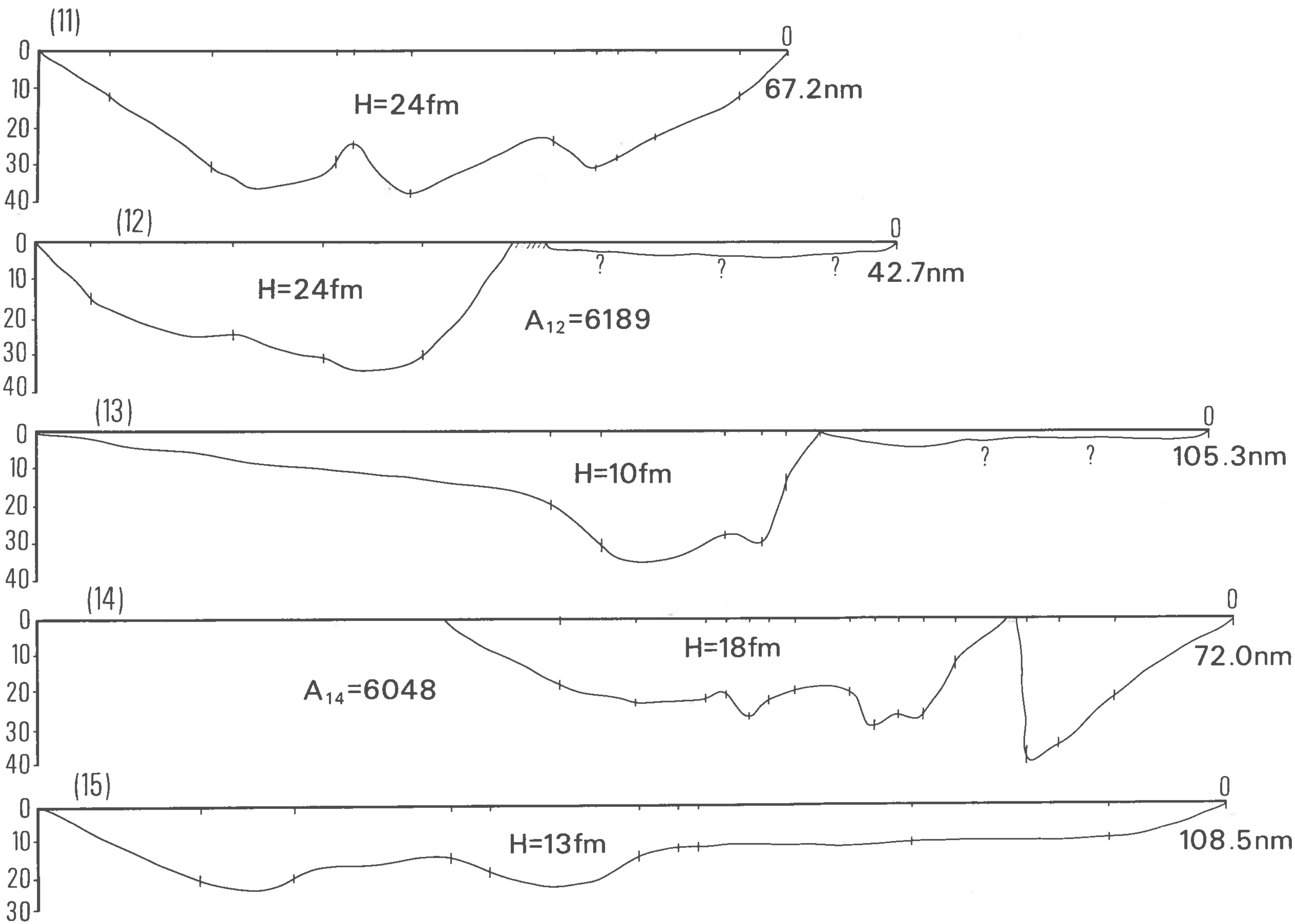
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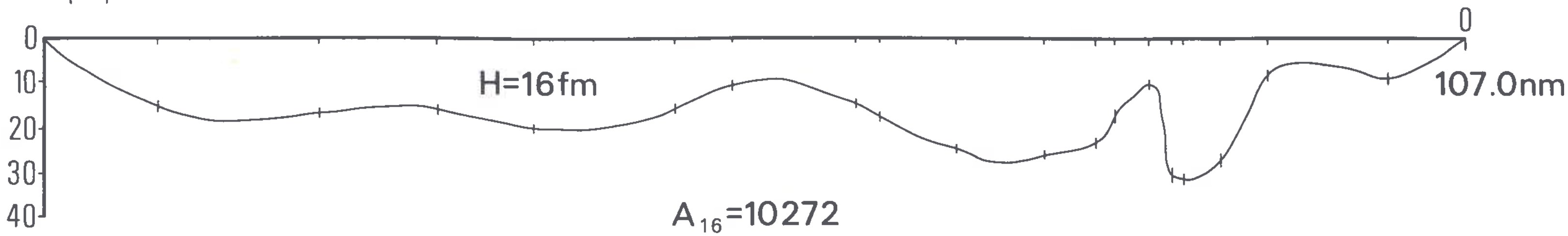
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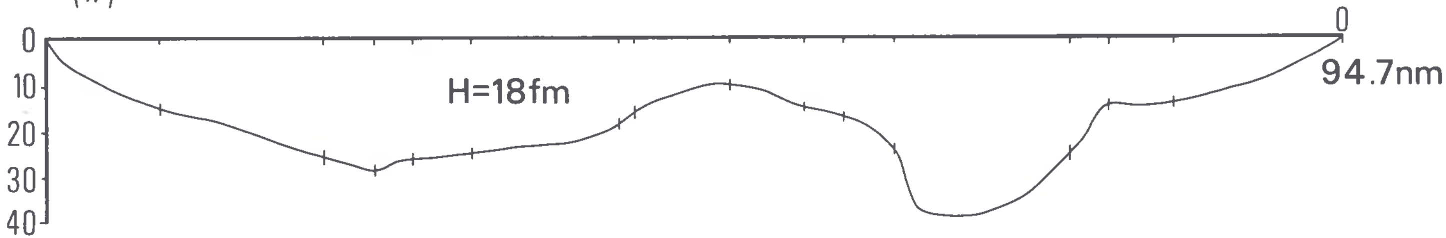
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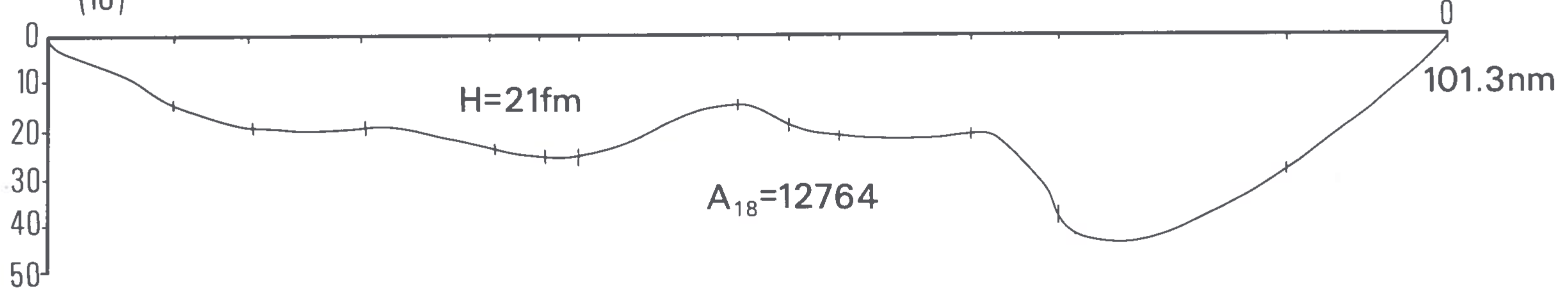
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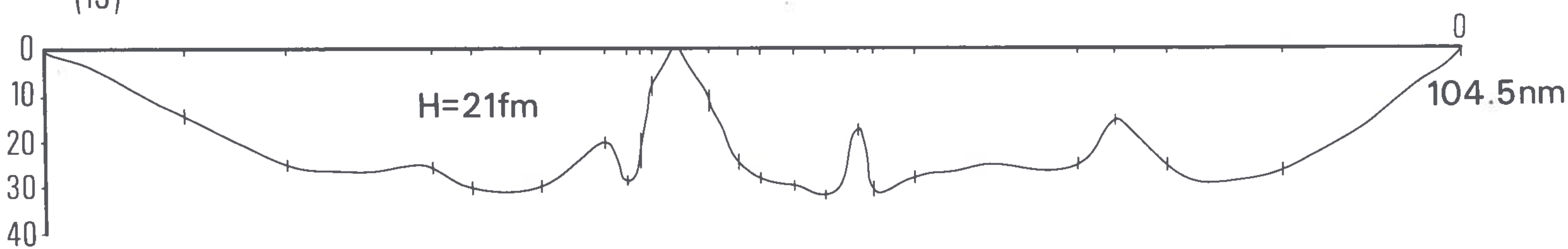
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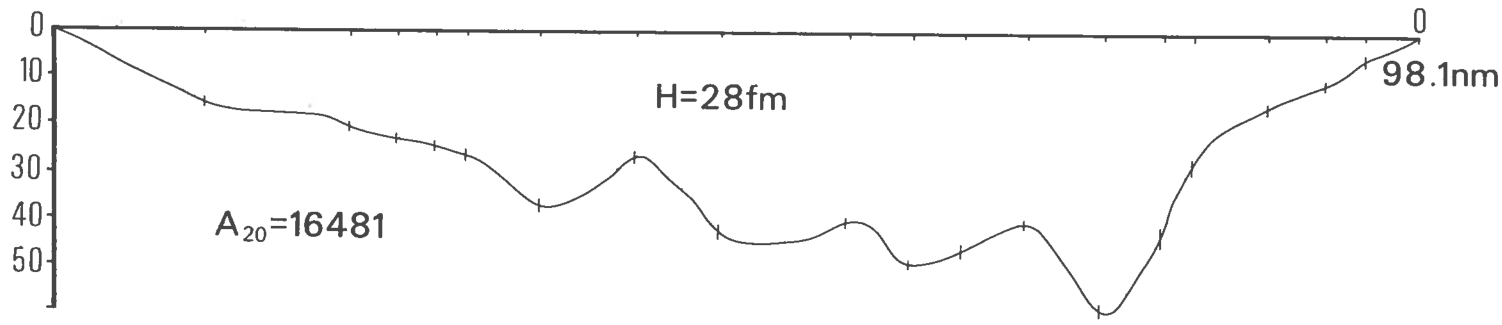
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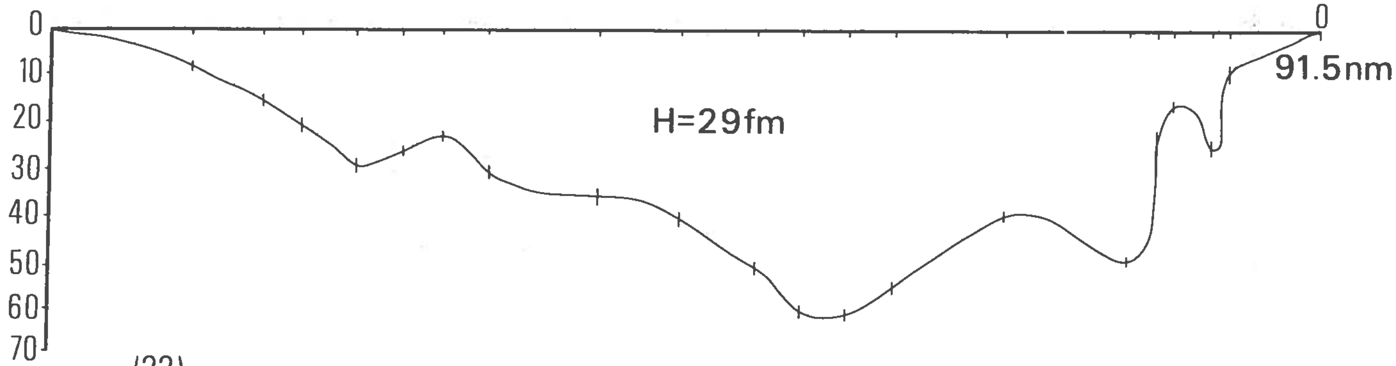
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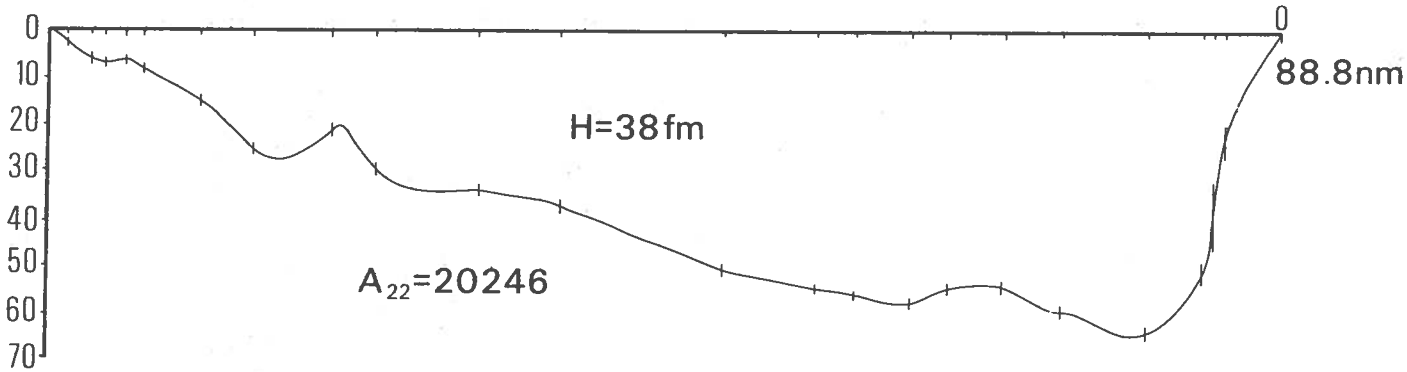
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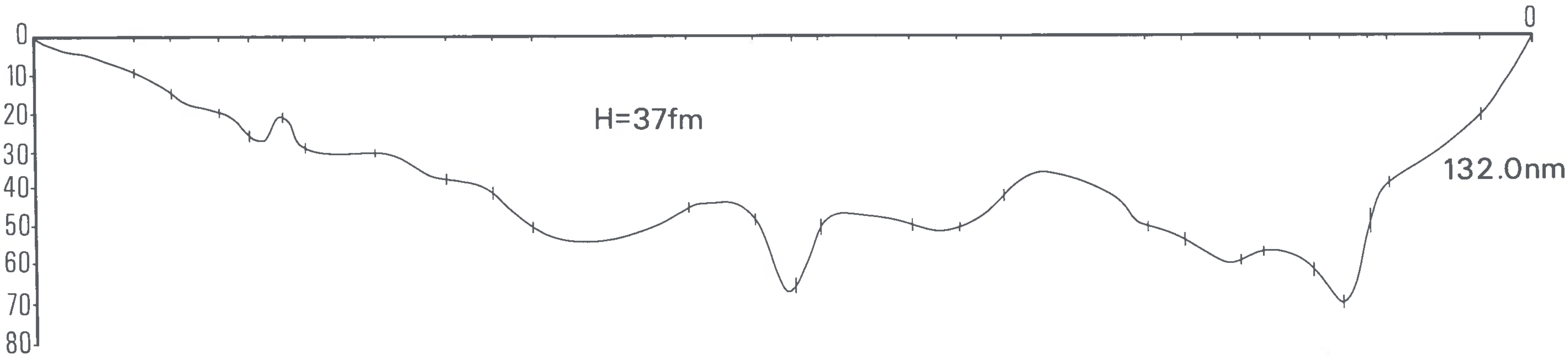

Circulation in James Bay

by

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Marine Sciences Branch

Department of the Environment

Ottawa, Ontario

Contents

0. Abstract

This is a preliminary study on some physical oceanographic problems of James Bay. The interest in this study arose in connection with the proposed hydroelectric power project on some rivers draining into James Bay. The normal modes of response of James Bay are calculated using a one dimensional topographic model. Using the method of characteristics it has been shown that at the southern shores of James Bay the storm surge amplitude could reach up to 19 feet. It was shown that the thermohaline circulation in James Bay could be as intense as the wind generated circulation and this may explain some of the anomalous ice drift patterns. Although no barotropic coastal jets appear to be possible in James Bay, baroclinic coastal jets with a width of 7 kilometers (km) can occur with the core of the jet being at about 7 km seaward from the position where the thermocline intersects the bottom. The circulation in some rivers draining into James Bay is examined theoretically using the concepts of the three modes of estuarine circulation, namely the river-discharge mode, wind-stress mode and gravitational-convection mode. The question of atmospheric water balance over these river basins is briefly examined and some possible consequences of the man-made changes have been speculated.

1. Introduction

This is a preliminary study of some physical oceanographic problems in James Bay. Because of the deadline imposed on the time available for this study no rigorous numerical modelling has been attempted. The present interest in this study arose from the recent announcement by the government of Quebec to divert the waters of certain rivers flowing into James Bay for hydroelectric power purposes.

Barber (1971) summarized some of the physical oceanographic problems in James Bay and its connected waters, namely Hudson Bay. He also pointed out the considerations that led to this present study. Godin (1971) studied the propagation of the tide into James Bay using a one dimensional topographic model. The topographic data used in the normal mode study has been kindly supplied to me by G. Godin and the input data on the estimates for the circulation problem has been supplied by F.G. Barber.

2. The normal modes of James Bay

A knowledge of the normal modes of response of James Bay is useful in understanding the phenomena of tides, storm surges and circulation and for this reason these have been computed using a one dimensional topographic model. The linearized momentum and continuity equations for flow in a water body of varying cross-section are given by the so-called channel equations (Rao, 1968),

$$\frac{\partial \hat{M}}{\partial t} = -gA \frac{\partial \hat{\eta}}{\partial x} \tag{1}$$

$$\frac{\partial \hat{\eta}}{\partial t} = - \frac{1}{B} \frac{\partial \hat{M}}{\partial x} \tag{2}$$

where the x axis is locally tangent to the principal axis of the channel, t is time and g is gravity. The other parameters have the following meanings:

$\hat{M}(x,t)$ = volume transport through a vertical section,

$\hat{\eta}(x,t)$ = water level deviation from the undisturbed level,

$A(x)$ = area of cross-section and

$B(x)$ = surface width of the section.

In the above equations, friction and terms due to the earth's rotation are ignored. Later we will estimate the possible effects of these terms on the frequencies of the normal modes. For free oscillations of the systems assume,

$$\hat{M}(x,t) = M(x) \cdot \sin(\sigma t) \text{ and} \tag{3}$$

$$\hat{\eta}(x,t) = \eta(x) \cdot \cos(\sigma t)$$

Here $M(x)$ and $\eta(x)$ are the space dependent normal mode functions of James Bay and σ is the frequency. Substitution of (3) into (1) and (2) gives:

$$\sigma M = -gA \frac{d\eta}{dx} \tag{4}$$

$$\sigma\eta = \frac{1}{B}\frac{dM}{dx} \tag{5}$$

The boundary conditions are the following:

M is arbitrary and $\eta=0$ at the mouth of James Bay, (6)

M=0 at the head of James Bay. (7)

Equations (4) to (7) form an eigenvalue problem of the frequencies σ.

Define, following Rao (1968)

$$C_i \equiv \frac{\Delta x}{gA_i} \quad \text{and} \quad D_i \equiv -B_i \Delta x \tag{8}$$

Then equations (4) and (5) become:

$$\eta_{i+1} = \eta_{i-1} - \sigma C_i M_i, \ i=2(2)22 \tag{9}$$

$$M_{i+1} = M_{i-1} - \sigma D_i \eta_i, \ i=3(2)23 \tag{10}$$

These finite difference forms are written with respect to a staggered grid in which M and η are computed at alternate points. The James Bay topography used here is the same as the one used by Godin (1971) in his calculation of the propagation of the tide. The grid distance Δx is 20 nautical miles, this being the distance either between two successive M sections or two successive η

sections. In our model the grid numbering is such that, section one is the mouth and section twenty-four is the head of James Bay, opposite to the scheme of Godin (1971).

The equations (9) and (10) were solved by prescribing $\eta_1=0$ and $M_2=10^8$ cm^3 sec^{-1} and alternately solving these equations. Since these are linear equations, any arbitrary value can be assigned to M_2. A trial value is provided for σ for the first mode and if M_{24} is not very close to zero, the value of σ is changed slightly and the computation is repeated. This is continued until that value of σ is found which makes M_{24} equal to zero for practical purposes. This value of σ gives the frequency of the fundamental mode. The same procedure can be used to calculate the frequencies of the higher modes.

The trial values for σ for each mode can be calculated from the Merian formula:

$$T_m = \frac{4L}{m\sqrt{gH}} \tag{11}$$

where T_m is the period of the m^{th} longitudinal mode, L is the total length of James Bay and H is the average depth of James Bay. The values used were L=230 nautical miles and H=32 metres. Table 1 shows the periods of the normal modes of James Bay calculated both from the Merian formula and from the topographic model.

Table 1 Periods of the normal modes of James Bay.

Modal Number	Period Calculated from Merian Formula Hours	Min	Period Calculated from Topographic Model Hours	Min
1	26	45	22	42
2	13	23	8	54
3	8	55	6	00
4	6	41	4	24
5	5	21	3	48
6	4	28	3	06

Figures 1(a) to 1(f) show the structures of these six longitudinal modes in terms of the modal functions $M(x)$ and $\eta(x)$. In each, the abscissa denotes the grid point number with 1 denoting the mouth and 24 the head. The ordinate scale on the left side shows η and on the right side shows M. Figure 2 shows the positions of the nodes in James Bay computed from the topographic model.

In the above calculation the effect of the earth's rotation is ignored, which is a serious drawback considering the size of James Bay. The inclusion of rotation not only changes the frequency of the mode but also destroys the standing nature of the oscillation by introducing amphidromic systems into the modal structures. James Bay may be treated as a rectangular bay with length about three times the average width for purposes of estimating the effect of rotation on the frequency of a given mode. Rao (1968) mentions that Van Dantzig and Lauwerier (1960) gave the following formula for a rectangular bay with length twice the width,

$$\sigma = \sigma_0 + 0.504 \frac{f^2 L}{2\pi C} + O(f^4) \tag{12}$$

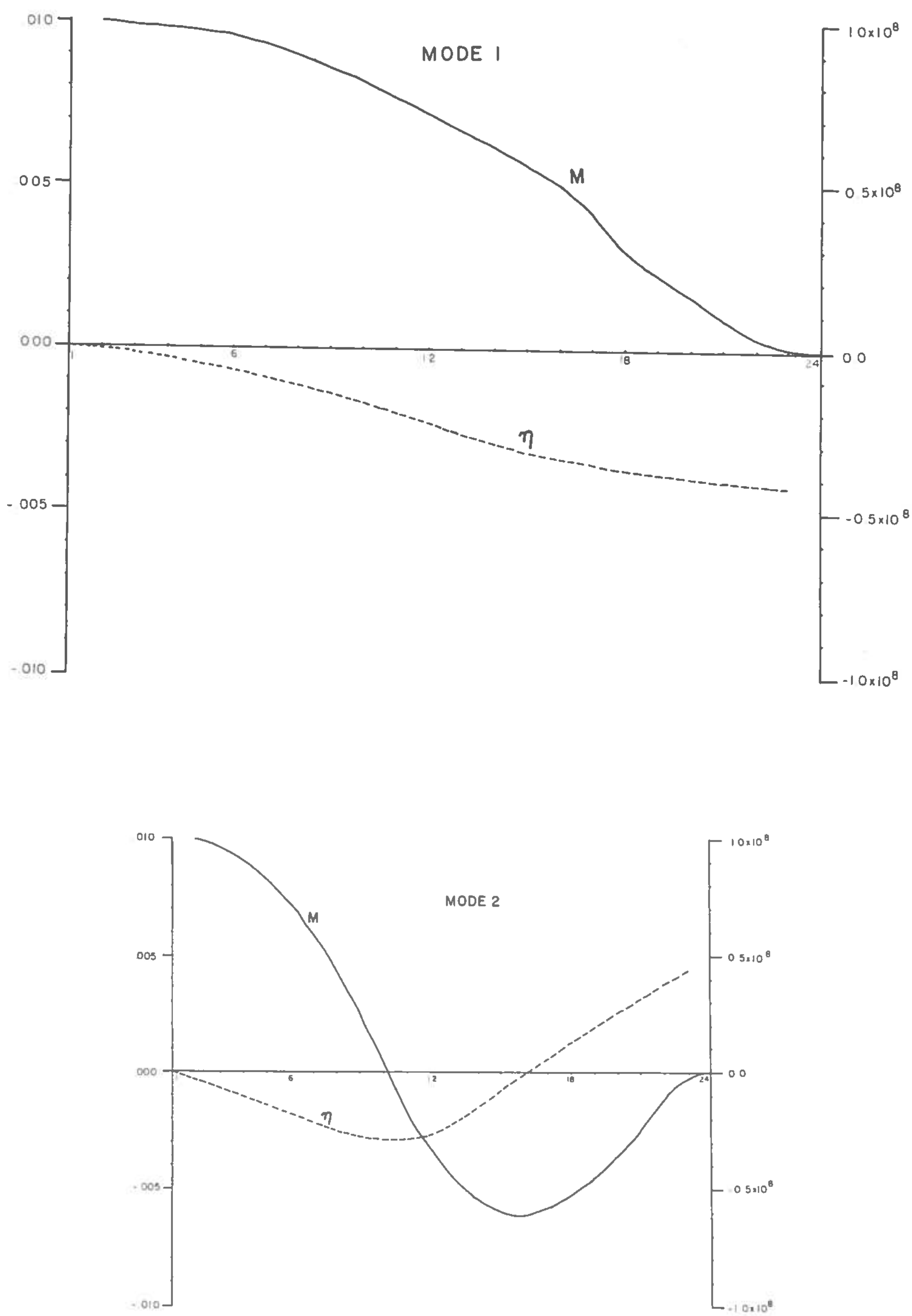
where σ_0 is the lowest nonrotating frequency, given by,

$$\sigma_0 = \frac{\pi C}{2L} \quad \text{and}$$

$$C = \sqrt{gH} \tag{13}$$

and where the other symbols have the same meaning as before.

Figure 1 Structure of the first six longitudinal modes of James Bay. The abscissa shows the grid number (1 is the mouth and 24 is the head). The ordinate scales on the left and right sides are for the water level η and the volume transport M respectively. Since this is a linear problem the actual units are arbitrary.


MODE 1
M
η
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.005
0.00
-.005
-.010
1
6
12
18
24
1.0x10^8
0.5x10^8
0.0
-0.5x10^8
-1.0x10^8
MODE 2
M
η
.010
.005
.000
-.005
-.010
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12
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24
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Figure 1

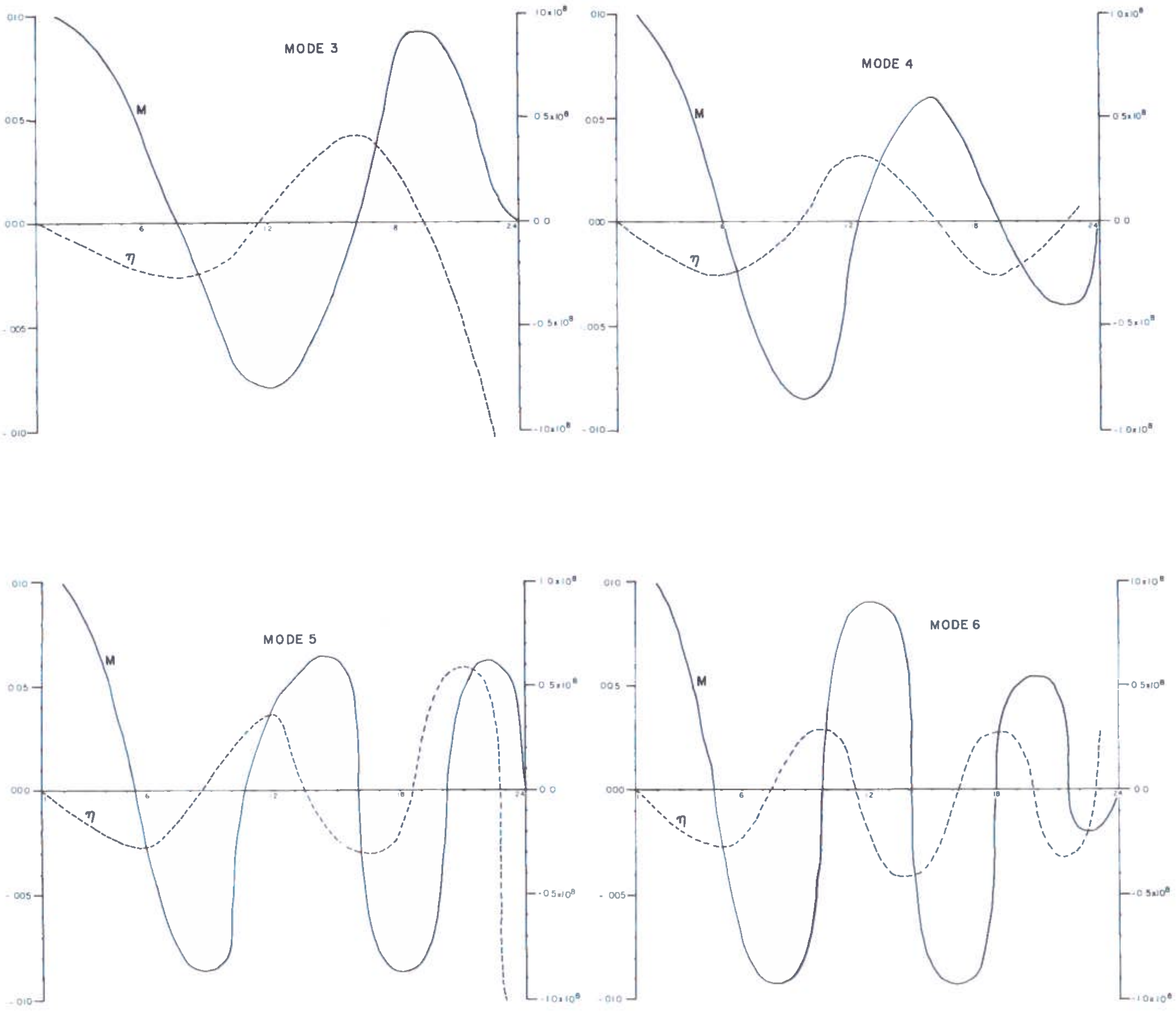


Figure 1 (con't)

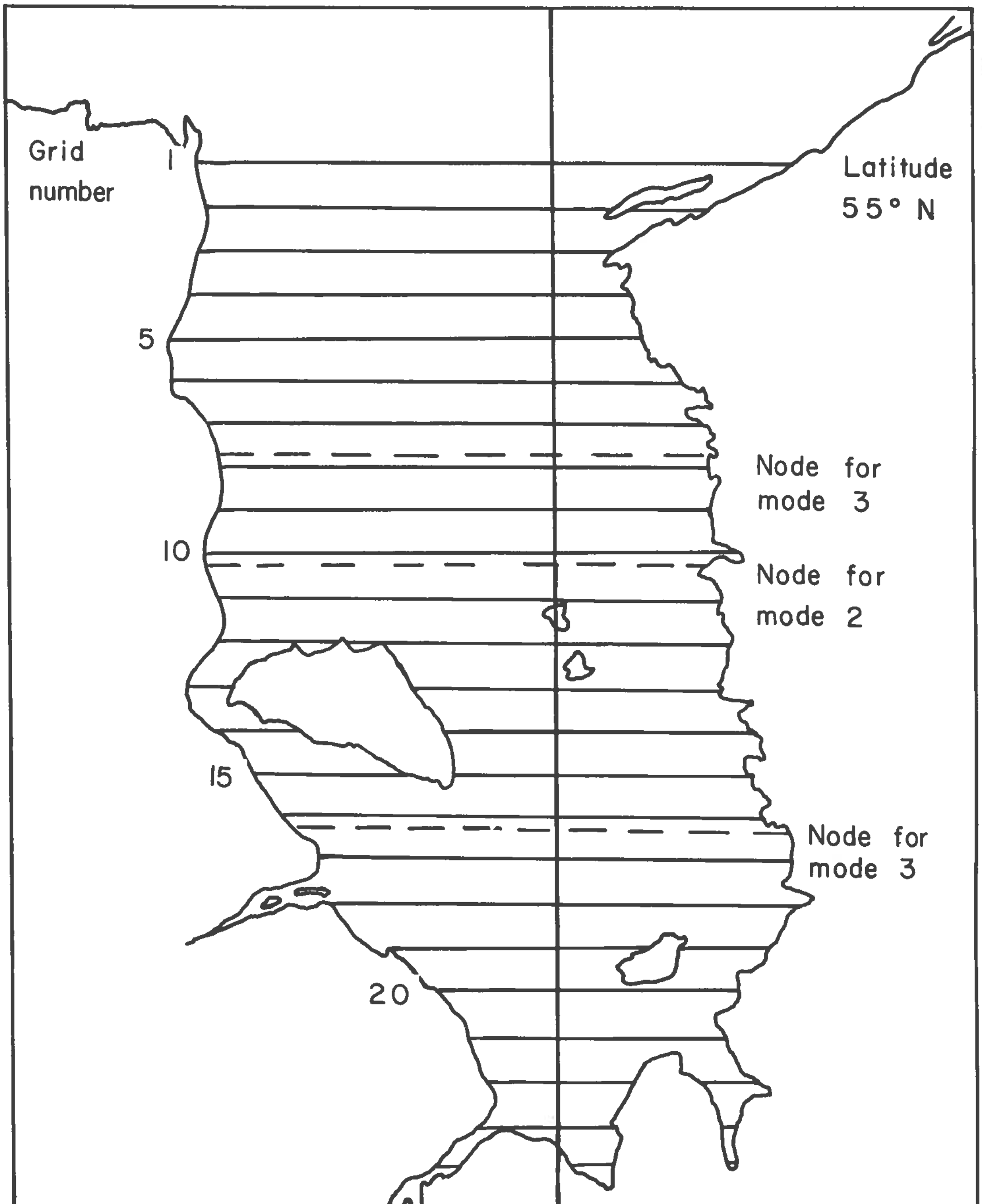


Figure 2 Nodal positions in James Bay calculated from the topographic model.

The equation gives a result that is exactly opposite to what happens in a completely closed basin. That is, in the case of a rectangular bay, rotation increases the frequency of the lowest mode. Since in our case (Table 1) the period of the fundamental mode is 22 hours 42 minutes then,

$$\sigma_o = \frac{2\pi}{22.7 \times 3600} \ \sec^{-1} \tag{14}$$

$$\text{Since } \frac{L}{C} = \frac{\pi}{2\sigma_o} = \frac{22.7 \times 3600}{4}$$

$$= 20{,}430 \ \sec$$

From (12), (13)

$$\frac{\sigma L}{C} = \sigma_o \frac{L}{C} + 0.504 \frac{f^2}{2\pi} \left(\frac{L}{C}\right)^2$$

$$= \frac{\pi}{2} + 0.335$$

The nondimensional period, $T = 2\pi/(\sigma L/C)$ is therefore 3.3, while in the nonrotating case it is 4. Thus rotation can reduce the period by 17.5 percent for a rectangular bay with length and fundamental mode period of James Bay and with width equal to half the length of James Bay. Since, in reality, James Bay is about three times as long as it is wide, the reduction of the period due to rotation would be less than 17.5 percent.

3. Storm surge estimation for James Bay

Thompson (1968) summarized the climate of Hudson Bay. He stated that the upper air circulation over Hudson Bay is the persistent counter-clockwise air flow around a low pressure vortex over northern Baffin Island in winter. This gives rise to a general transport of cold air in a northwest to southeast direction in winter. Thompson further stated that since many of the storm centers lack sufficient moisture to cause heavy precipitation, their principal effect is in inducing strong north winds over Hudson Bay as they travel across it. He further stated:

> "In contrast to the broad expanses of Arctic tundra that surround Hudson Bay the sub-Arctic lands bordering James Bay are partially forested and thus protected from strong winds. As a result two of the most distinguishing features of Hudson Bay winter climate - wind chill and blowing snow - are not nearly as evident near James Bay. With this important exception and the fact that James Bay is several degrees of latitude farther south,the factors that influence the climate of James Bay are essentially the same as those outlined for Hudson Bay".

Archibald (1969) studied the storm tracks over Hudson Bay and eastern Canada. As can be seen from his diagrams the storms move from different directions over James Bay. Since my present interest is in the maximum possible amplitudes of storm surges in James Bay, I will concern myself mainly with storms moving from north to south over James Bay thereby piling up water on the southern shore and causing large surges. Since the magnitude of the surge is strongly dependent upon the nature of the coast of James Bay I will briefly summarize some relevant facts from the article by Robinson (1968).

It can be seen from Figure 1 of Robinson's article that the eastern half of James Bay falls into the classification of "East Coast Uplands", while the western half falls into the category "South Coast Lowlands". Regarding the south coast lowland, I quote from Robinson (1968):

> "Along the whole coast there is a flat strip five to ten miles wide, with the widest parts generally being to the north. The coastal zone is treeless, but grass or marshes are common. Storm beaches, a few feet high, are the only topographic features. Tidal flats may be exposed for one to six miles; even at high tide shallow water extends far offshore. ...Deeper water may be found at the drowned river mouths, but shifting sand bars and minor deltaic deposits are navigation hazards".

Regarding the east coast upland I again quote Robinson (1968):

> "The east coast and adjoining interior have a combination of landform features which are different from those west of Hudson and James bays, but in some characteristics they have regional similarities. The central and northern sections of the east coast are high and rugged, with rocky hills and drift-covered uplands inland, as is characteristic of the Northwest Hills; the southern section along James Bay is a poorly-drained lowland, although not as wide as that on the west side."

Because of the time limitation of this study, no numerical modelling with topography taken into account has been attempted. Instead, use was made of the analytical solutions developed by Rao (1969) for a rectangular bay of uniform depth, making use of the method of characteristics. In this study

James Bay will be treated as a rectangular bay of length 230 nautical miles, of width up to 100 nautical miles and of average depth 32 meters, connected to Hudson Bay at the mouth. James Bay can be considered narrow because its length is about three times its average width and a one dimensional model is used here and the effect of earth's rotation is suppressed. Since we are interested in the transient response of the bay to a time and space-dependent wind field, the bottom friction is not included. It can be shown by order of magnitude considerations that the nonlinear term in the momentum equation can be omitted. Hydrostatic assumption has been made for the pressure field. For convenience, the relevant equations and the solutions will be summarized here.

Let $\bar{x}$, $\bar{z}$ represent a cartesian coordinate system such that the $\bar{x}$-axis is oriented along the length of James Bay with $\bar{x}=0$ at the mouth and $\bar{x}=\bar{L}$ at the southern shore. The $\bar{z}$-axis is positive upwards and is measured from the undisturbed position of the water surface. Then the bottom is at $\bar{z}=-\bar{h}$. Here, superior bar denotes dimensional quantities. The vertically integrated forms of the momentum and continuity equations are:

$$\frac{\partial \bar{M}}{\partial \bar{t}} + \bar{g}\,\bar{h}\,\frac{\partial \bar{\eta}}{\partial \bar{x}} = \bar{R} \tag{15}$$

$$\frac{\partial \bar{\eta}}{\partial \bar{t}} + \frac{\partial \bar{M}}{\partial \bar{x}} = 0 \tag{16}$$

where

$$\bar{M} = \int_{-\bar{h}}^{0} \bar{U} \, d\bar{z} \tag{17}$$

is the volume transport through a vertical section, $\bar{U}$ being the horizontal velocity in the $\bar{x}$ direction and $\bar{\eta}$ is the deviation of the water level from its mean position, $\bar{g}$ is gravity, $\bar{t}$ is time and $\bar{R}$ is the force due to wind stress given by:

$$\bar{R} \equiv \frac{\bar{\tau}}{\bar{\rho}} \tag{18}$$

The boundary conditions are the following:

$$\bar{M}=0 \text{ at the head of James Bay, i.e. at } \bar{x}=\bar{L}, \tag{19}$$

$$\bar{\eta}=0 \text{ at the mouth, i.e. at } \bar{x}=0. \tag{20}$$

Initially, $\bar{M}=0$, $\bar{\eta}=0$ at $\bar{t}=0$ for all $\bar{x}$. (21)

Rao (1969) introduced the following nondimensionalization:

$$x \equiv \frac{\bar{x}}{\bar{L}} \qquad \eta \equiv \frac{\bar{c}^2}{\bar{R}_0\bar{L}}\bar{\eta} \qquad R \equiv \frac{\bar{R}}{\bar{R}_0} \tag{22}$$

$$t \equiv \bar{t}\,\frac{\bar{c}}{\bar{L}} \qquad M \equiv \frac{\bar{c}}{\bar{R}_0\bar{L}}\bar{M}$$

where $\bar{c} \equiv \sqrt{\bar{g}\,\bar{h}}$ (23)

and $\bar{R}_0$ is a scale value of the wind stress. Using this scheme, equations (15) and (16) can be nondimensionalized. From addition and subtraction of these we get:

$$\frac{d}{dt}(M \pm \eta) = R \text{ for } \frac{dx}{dt} = \pm 1 \tag{24}$$

This states that the quantity $M \pm \eta$ is conserved along the positive and negative characteristics $dx/dt = \pm 1$, respectively. Since the bay is assumed to have a uniform depth $\bar{h}$, the positive and negative characteristics are straight lines given by $x = \pm t + \text{const}$.

I will consider two different types of stress bands and assume that the wind stress force R is in the form of a step function and moves with a constant speed $\bar{V}$ starting from x=0 at t=0. The nondimensional form of this translational speed is

$$V \equiv \frac{\bar{V}}{\bar{c}} \tag{25}$$

The value of R at any given point in the bay is given by

$$R = \begin{cases} 1 \text{ for } t \geq \dfrac{x}{V} \\ 0 \text{ for } t \leq \dfrac{x}{V} \end{cases} \tag{26}$$

This stress band crosses the head of the bay (x=1) at t = 1/B and in the case of the semi-infinite band, for t > 1/V the entire bay is under the influence of the wind stress. In the case of the finite stress band we assume a pulse of a square wave shape with zero forcing ahead and behind the pulse and the width of the pulse being $V(t-t') = VT = X$ where T is the time taken by the pulse to travel past a given point in the bay. The forcing function for the finite stress band can be obtained by superimposing on (26) another semi-infinite band $R'(x',t')$ where

$$R'(x',t') = \begin{cases} 0 \text{ for } t' \leq \frac{x}{V} \\ -1 \text{ for } t' \geq \frac{x}{V} \end{cases} \qquad (27)$$

and $t' \leq t$. Hence, the total forcing function for the case of finite band width is

$$R+R' = \begin{cases} 0 \text{ for } t \leq \frac{x}{V} \\ 1 \text{ for } t \geq \frac{x}{V} \geq t' \\ 0 \text{ for } t' \geq \frac{x}{V} \end{cases} \qquad (28)$$

Since our equations are linear, the solutions due to forcing from R' can be obtained from the solutions due to forcing from R by changing the sign of η and replacing t with $t' = t - T$.

Rao showed that the maximum possible elevation at the head is given by:

$$\bar{\eta}_{max} = \frac{2}{\bar{g}} \frac{\bar{L}}{\bar{h}} \bar{R}_o \qquad (29)$$

He used the following representation for $\bar{R}_o$

$$\bar{R}_o = 4 \times 10^{-6} \ |W|W \tag{30}$$

where W is the wind speed (ft sec^{-1}) along the axis of the bay. If we take W=75 ft sec^{-1}, corresponding to 51 mph we get

$$\bar{\eta}_{max} = \frac{2 \times 230 \times 6080}{32 \times 32 \times 3.281} \times 4 \times 10^{-6} \times 75^2$$

$$= 18.8 \text{ feet}$$

For higher winds speeds the amplitude of the surge is even higher. The reason for the possibility of such large storm surges in James Bay is the long fetch and the relatively shallow water in the bay.

Thus the storm surges appear to attain very large amplitudes at the southern end of James Bay when intense storms pass over it. This result coupled with the fact that the shores are very flat could give rise to very serious storm surge inundation.

4. Circulation in James Bay

The circulation in James Bay cannot be studied completely independent of the circulation in Hudson Bay to which it is connected at its widest portion. As is expected for any large water body in the northern hemisphere, the circulation in Hudson Bay is dominated by a large counterclockwise cell (Murty and Yuen, 1970). Figure 3 shows the circulation in September (left side) and in May (right side) calculated by these authors using wind stress computed through the balance equation. Figure 4 shows the corresponding water level deviations. These calculations were made using a steady state topographic model and pressure data averaged over several years has been used in the wind stress computation. It can be seen from Figures 3 and 4 that while the circulation itself is extremely weak and not too well defined in the southern portion of Hudson Bay and James Bay, the water level gradients in the equilibrium state are strong in James Bay (relative to Hudson Bay).

The Ice Central Office at Halifax which is responsible for predicting the movement of ice has faced some difficulty in their prediction because of anomalous drift patterns,

> "...The summer of 1968 provided another such situation. The general water circulation pattern is believed to be a cyclonic drift around the Bay. This would normally be expected to carry ice from the Cape Churchill area southeastward toward James Bay. A west to northwest wind accompanying such drift should move the ice southeastward rather quickly...the combination of wind and water current drifts is almost negligible.
>
> It must be assumed, therefore, that during the melting period thermohaline processes occur which tend to disrupt the normally weak circulation in the southern bay. The addition of fresh water through runoff from the rivers along the west and south coasts also tends to disrupt the salinity balance of the southern section of the Bay. However, the failure of ice to drift southeastward under the influence of persistently favourable winds leads again to the idea of a counter-current in the southern part of the Bay, especially during the period when melting ice is present.

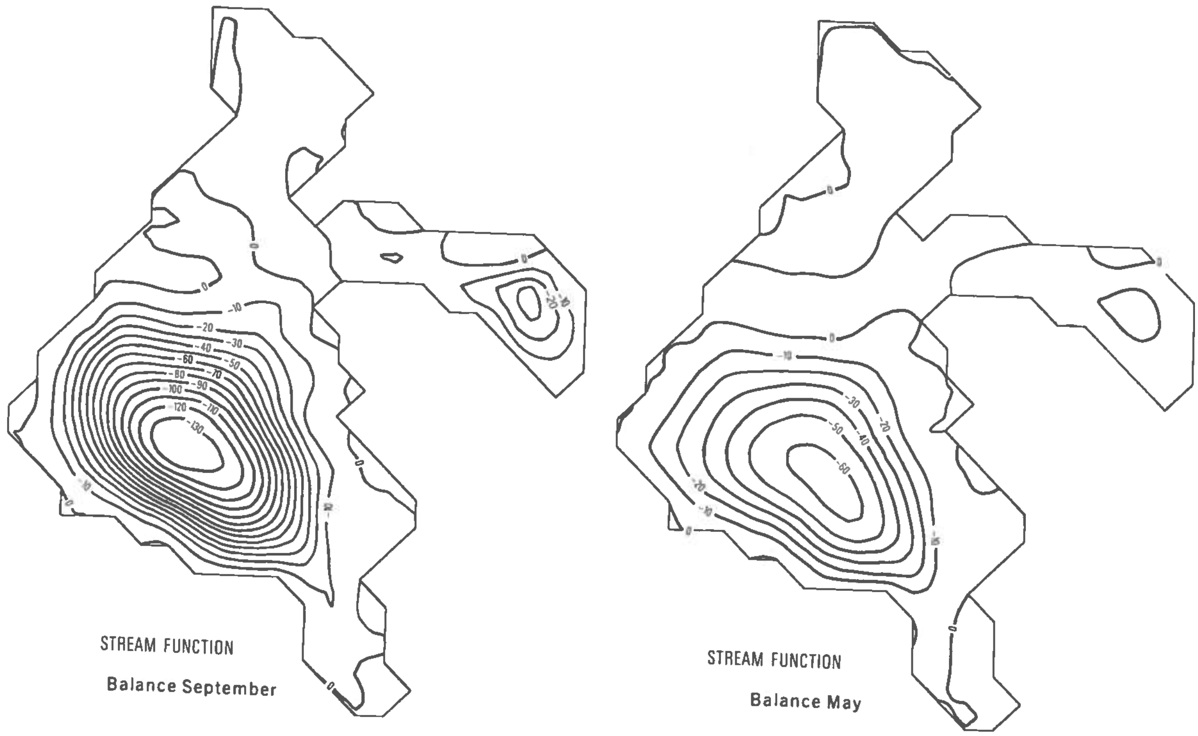


Figure 3 Stream function (for the volume transport) in units of 10^{12} cm^3 sec^{-1} for September (left) and May (right) using data averaged over 30 years.

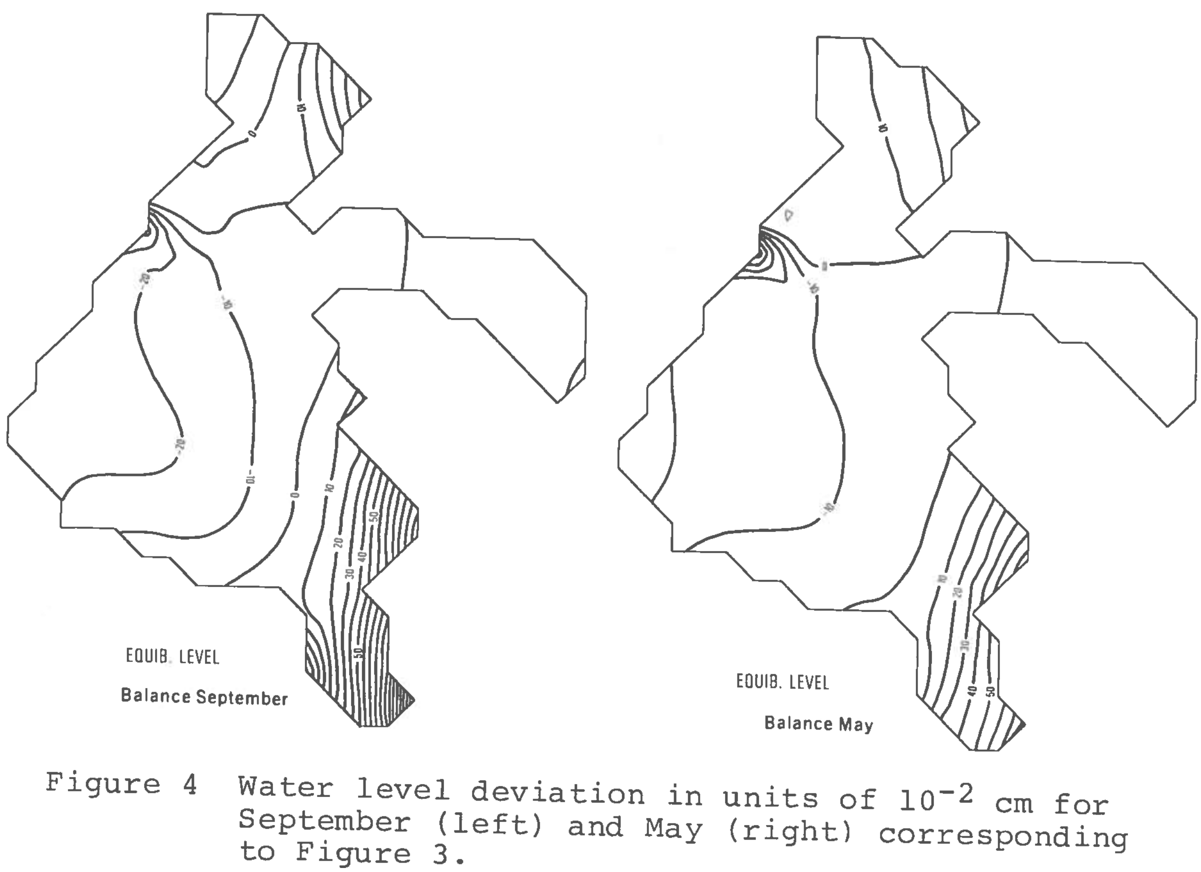


Figure 4 Water level deviation in units of 10^{-2} cm for September (left) and May (right) corresponding to Figure 3.

> It is obvious that a serious study of thermohaline relationships to water circulation in this area is necessary before useful forecasts can be provided for southern Hudson Bay". (Anon, 1970).

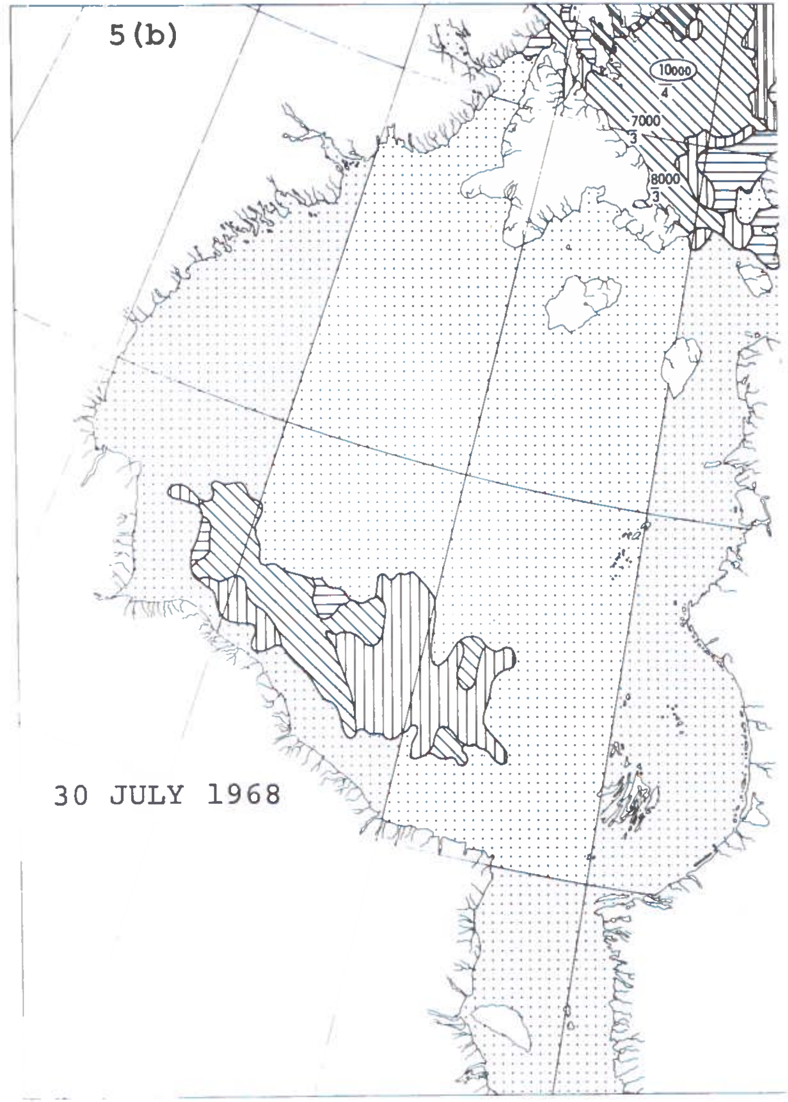
Figures 5(a) to 5(f) show a sequence of diagrams in which the ice conditions in Hudson Bay are shown on different days. These figures are respectively the figures 12, 14, 16, 18, 20 and 22 in the Ice Central Report for 1968. It can be seen from these that the ice simply melted away and did not move with the water current pattern.

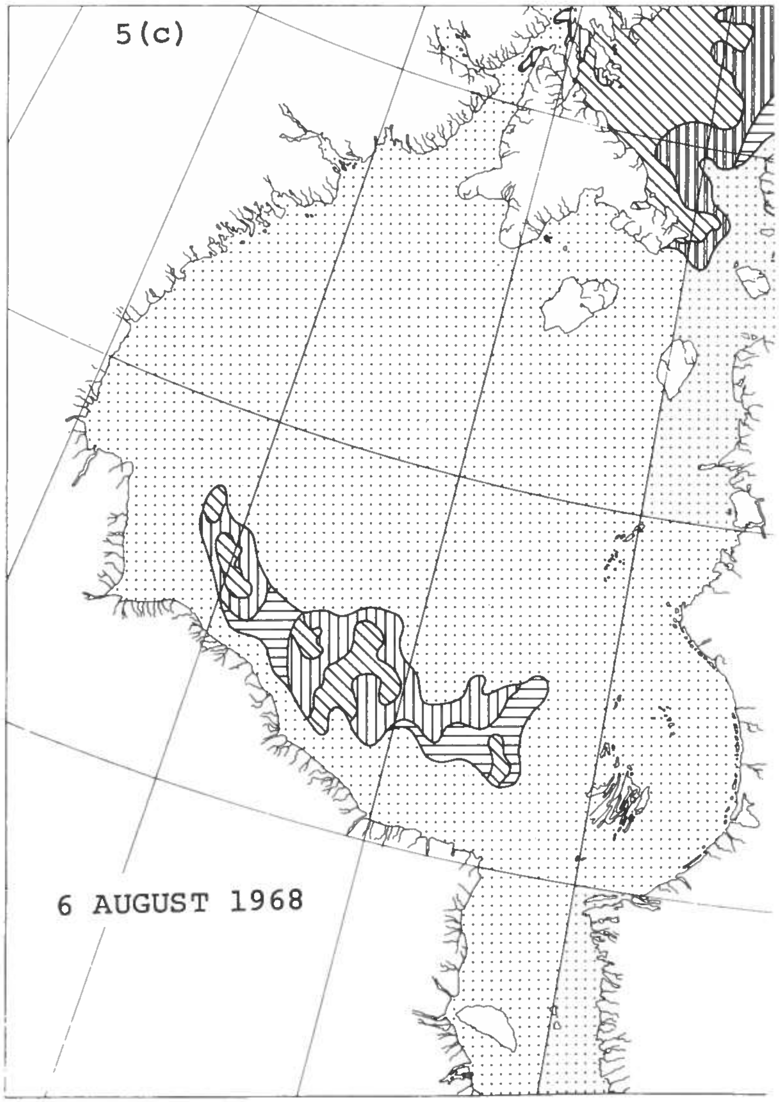
It is clear that an estimation of the intensity of the thermohaline circulation is essential, especially to answer the question whether this could be strong enough to offset the wind-generated circulation. Another consideration is that because James Bay really is part of the complex consisting of Hudson Bay, Hudson Strait and Foxe Basin, this system should be studied together and any estimates made should involve the length and time scales representative of this system as a whole. The north-south extent of this system is large enough for β, the variation of the coriolis parameter with latitude to play a significant role. Because of the time limitations on this study no numerical model has been developed. Instead, estimates have been made using an analytical study made by Gates (1968). For convenience I will summarize Gates' theory.

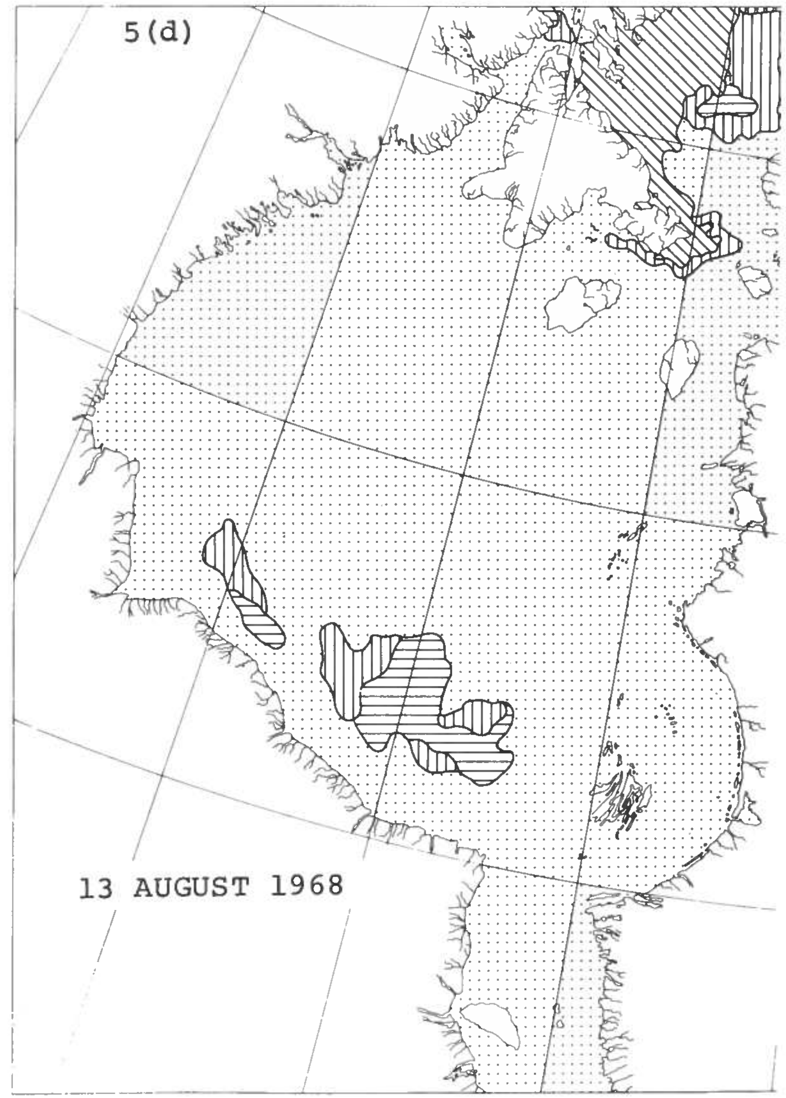
> "The action of surface wind stress on an underlying ocean has been extensively studied since the pioneering work of Ekman (1905). The characteristic depth

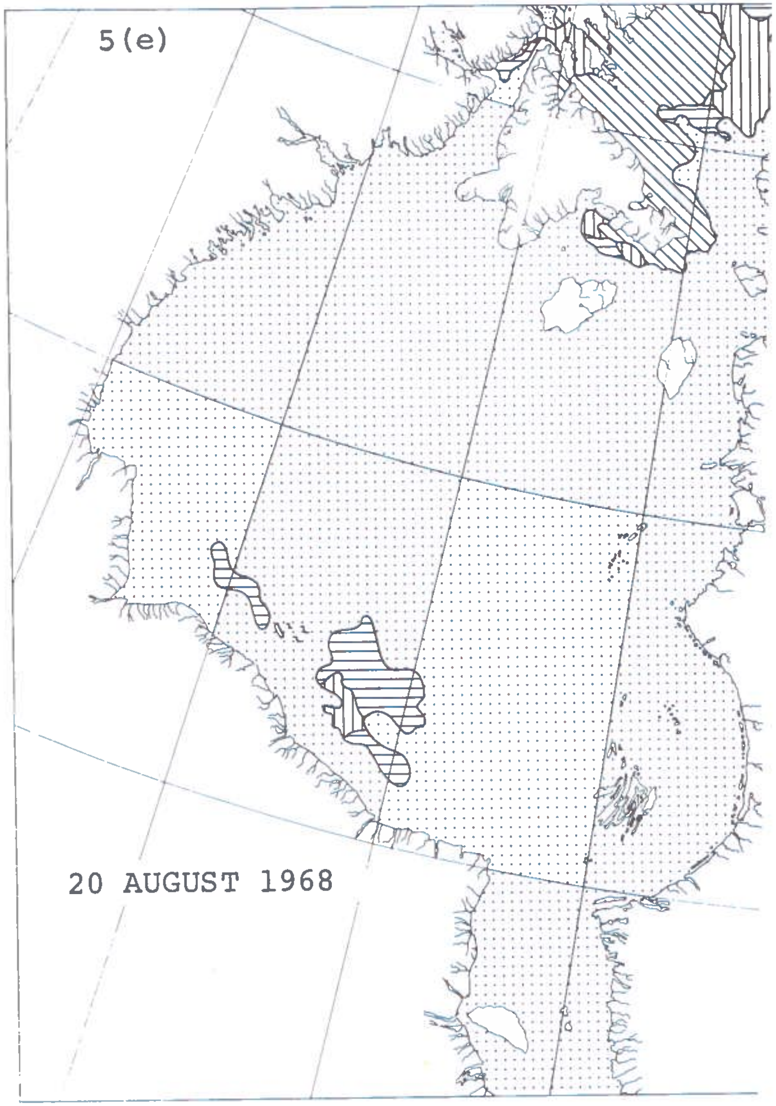
Figure 5 Ice conditions in Hudson Bay for the winter period of 1968-69 on six different dates.

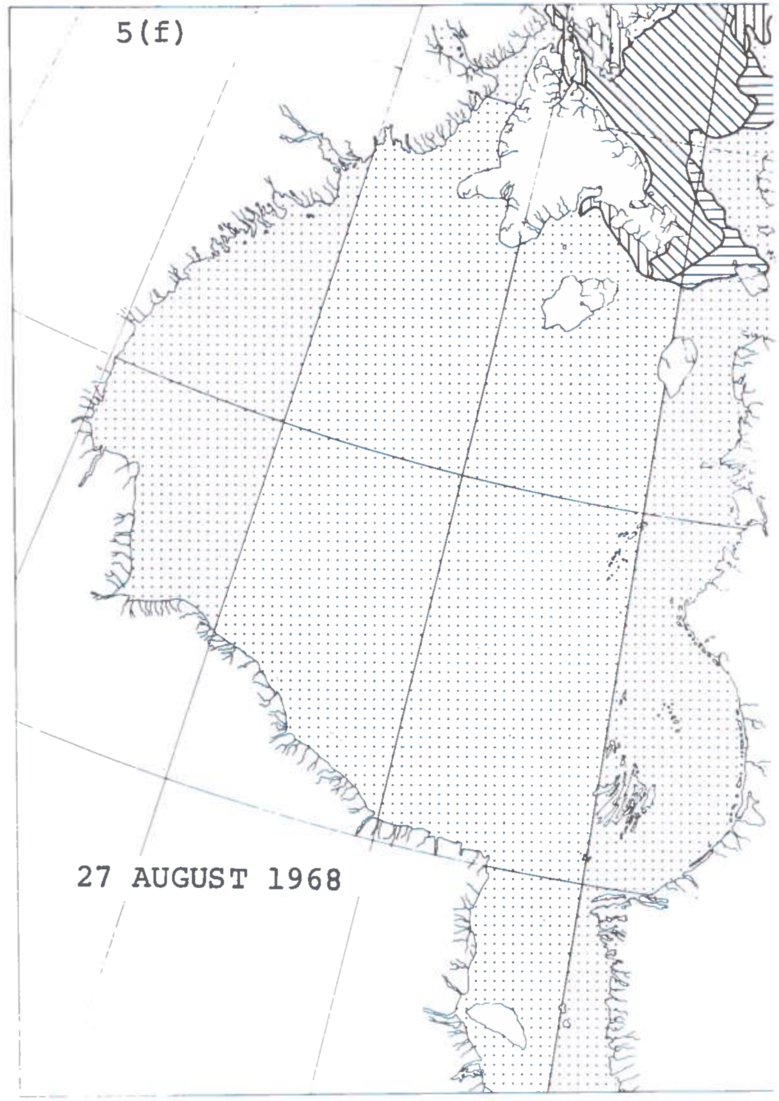

5(a)
23 JULY 1968


5(b)
10000
4
7000
3
8000
3
30 JULY 1968


5(c)
6 AUGUST 1968


5(d)
13 AUGUST 1968


5(e)
20 AUGUST 1968


5(f)
27 AUGUST 1968

> of penetration of wind-induced motion (Ekman depth) depends upon both the vertical eddy diffusion and the coriolis parameter, and the net transport in deep waters is normal to the surface wind stress with the horizontal velocity displaying the well known spiral structure. The modifications of these features in shallow water and near coast lines were also explored by Ekman by the introduction of a bottom boundary layer whose transport complements that of the surface layer. In addition to describing the horizontal transport in terms of the wind stress, the Ekman theory provides an estimate of the net vertical flux required beneath the surface boundary layer by mass continuity. This is the so-called Ekman vertical velocity."

Gates (1968) studied at first, a homogeneous ocean of uniform depth on a β-plane and then considered stratification. His assumptions include steady-state, neglect of inertial terms and horizontal friction. The homogeneous ocean is divided into an interior region in between two boundary layers, one at the top and one at the bottom. It should be cautioned that because of the shallow nature of the Hudson Bay System it is debatable whether there really is an interior region where the vertical eddy viscosity is negligible.

The characteristic thickness E of the Ekman boundary layer is:

$$E = \sqrt{\frac{2\nu_z}{f}} \tag{31}$$

where f is the coriolis parameter and ν_z is the vertical eddy viscosity. Gates defined E slightly differently from the conventional definition of Ekman depth, the difference being, here E is thickness at which the speed is reduced by the factor e^{-1} and this is $1/\pi$ times the conventional definition. Since f increases with increasing latitude, E decreases with increasing latitude provided ν_z does not change. Another assumption involved in Gates' study is that the thicknesses of the boundary layers are small compared to the total depth H (assumed uniform).

The equilibrium pressure field P is determined from the following relation:

$$E\nabla^2 P + \frac{\beta}{f}(2H-E)\frac{\partial P}{\partial x} - \frac{\beta E}{f}\frac{\partial P}{\partial y}$$

$$= 2\,\mathrm{curl}_z\,\tau + \frac{\beta}{f}\tau_{sx} \tag{32}$$

where τ is the vector wind stress with x and y components τ_{sx} and τ_{sy}. It can be seen that, when $\beta=0$ this reduces to

$$\nabla^2 P = \frac{2}{E}\,\mathrm{curl}_z\,\tau \tag{33}$$

which was originally derived by Ekman (1923). The vertical velocity at the bottom of the surface boundary layer is

$$W_1 = \frac{1}{\rho_o f_o}\,\mathrm{curl}_z\,\tau \tag{34}$$

which agrees with the Ekman vertical velocity. Here, ρ_o is the density of the homogeneous fluid and f_o is the coriolis parameter at the standard latitude such that

$$f = f_o + \beta y \tag{35}$$

The vertical velocity in the interior region is given by

$$W_I = W_K + \left(\frac{\beta}{f^2 \rho_o} \frac{\partial P}{\partial x} \right) z \tag{36}$$

where $W_K = W_K(x,y)$ is an arbitrary function to be determined knowing the wind stress. In the absence of β, this is uniform in depth.

Next, Gates introduced a surface temperature distribution and considered its effects on the Ekman vertical velocity. This temperature distribution is assumed to be maintained by surface heat exchanges and Gates assumed that the horizontal temperature gradient decreases linearly with depth, from a maximum at the surface, z=H, to zero at the bottom, z=0. Thus if T is the temperature field, then

$$\frac{\partial T}{\partial x} = \frac{z}{H} \left(\frac{\partial T}{\partial x} \right)_o \tag{37}$$

$$\frac{\partial T}{\partial y} = \frac{z}{H} \left(\frac{\partial T}{\partial y} \right)_o$$

where the subscript o denotes the value at the surface. Let the equation of state be

$$\rho = \rho_o (1 - \alpha T) \tag{38}$$

where α is the coefficient of thermal expansion of water. That is, here we ignored the effect of salinity variations on the density. However, qualitatively speaking it will add one more term in equations (39) and (40). In (38), T is the temperature relative to a reference value corresponding to $\rho=\rho_o$.

Assuming that $\eta \ll H$ (where η is the perturbation of the free surface) and $\rho \approx \rho_o$ when η and ρ are undifferentiated we can write the following expressions for the interior geostrophic velocity components

$$U_I(z) = -\frac{g}{f}\frac{\partial \eta}{\partial y} + \frac{\alpha g(H^2 - z^2)}{2fH}\left(\frac{\partial T}{\partial y}\right)_o \tag{39}$$

$$V_I(z) = \frac{g}{f}\frac{\partial \eta}{\partial x} - \frac{\alpha g(H^2 - z^2)}{2fH}\left(\frac{\partial T}{\partial x}\right)_o \tag{40}$$

Thus the presence of a temperature gradient permits the reversal of the current with depth.

The surface boundary conditions are now

$$\frac{\partial U}{\partial z} = \frac{\tau_{sx}}{\rho_o \nu_z} + \frac{\alpha g}{g}\left(\frac{\partial T}{\partial y}\right)_o \tag{41}$$

$$\frac{\partial V}{\partial z} = \frac{\tau_{sy}}{\rho_o \nu_z} - \frac{\alpha g}{f}\left(\frac{\partial T}{\partial x}\right)_o \tag{42}$$

both at z=H. The vertical velocity in the interior is given by

$$W_I(z) = \frac{\beta g}{f^2}\frac{\partial \eta}{\partial x} z - \frac{\beta \alpha g}{2f^2 H}\left(\frac{\partial T}{\partial x}\right)_o \left(H^2 z - \frac{z^3}{3}\right) + W_K \tag{43}$$

Thus, in the non-homogeneous case the vertical velocity varies as the cube of the depth in the interior while in the homogeneous case it varies linearly.

The equation to determine η for the non-homogeneous case is given by

$$E\nabla^2\eta + \frac{\beta}{f}(2H-E)\frac{\partial\eta}{\partial x} - \frac{\beta E}{f}\frac{\partial\eta}{\partial y} = \frac{2}{g\rho_o}\,\mathrm{curl}_z\,\tau + \frac{\beta\tau_{sx}}{gf\rho_o} + \alpha E\left(\frac{H}{2} - E\right)\nabla^2 T_o$$

$$+ \frac{\alpha\beta H}{f}\left(\frac{2H}{3} - \frac{E}{2}\right)\left(\frac{\partial T}{\partial x}\right)_o + \frac{\alpha\beta E}{f}\left(\frac{3E}{2} - \frac{H}{2}\right)\left(\frac{\partial T}{\partial y}\right)_o \tag{44}$$

NOTE: equation (32) gives a corresponding equation for the homogeneous case.

It can be seen that when $\beta=0$ this reduces to

$$E\nabla^2\eta = \frac{2}{g\rho_o}\,\mathrm{curl}_z\,\tau + \alpha E\left(\frac{H}{2} - E\right)\nabla^2 T_o \tag{45}$$

Thus the effect of a local excess of temperature (for example, $\nabla^2 T_o < 0$) is analogous to that of the curl of the anticyclonic wind stress ($\mathrm{curl}_z\,\tau < 0$).

Without giving any more details we will simply state that starting with (45) Gates showed that the thermally forced contribution could be comparable to the wind-induced portion.

5. Possibility of coastal jets in James Bay

The dimensions of James Bay (length 230 nautical miles and average width of 100 nautical miles) are comparable to the Great Lakes. James Bay is very shallow (with an average depth of 32 meters) like Lake Erie. The main difference between James Bay and the Great Lakes, besides the saline water in the former and fresh in the latter, is that James Bay has a very wide mouth and is connected to Hudson Bay while the Great Lakes are essentially closed systems with only some interconnecting rivers. Csanady (1967, 1968a, 1968b) studied the coastal jets in the Great Lakes and his concepts (Csanady, 1971) could be applied here to study the possibility of coastal jets in James Bay:

> "Theoretical studies of some simple model Great Lakes (Csanady, 1967, 1968a,b) have suggested that shoreward wind drift may set up concentrated boundary currents of essentially zero potential vorticity in the Great Lakes by the mechanism of vortex stretching. Along an infinite vertical shore, as shown by Charney (1955) in an extension of Rossby's (1938) work on geostrophic adjustment, some coastal jets may be generated in this manner, although these would possess practical significance only in the *baroclinic* modes. In a closed basin of constant depth, the implusive application of wind stress leads, by the same mechanism, to the generation of slow baroclinic Kelvin waves (Csanady, 1968b), the profile of which, perpendicular to shore, is very similar to the steady-state coastal jets.
>
> The application of these previous results to the real Great Lakes or other similar bodies of water is not immediate for at least two reasons: 1) depth variations in the shore zone are certain to modify the vortex stretching mechanism; and 2) a linear, frictionless theory (from which the above results were obtained) can hardly be expected to describe water movements accurately, particularly in the shallow coastal zones."

Csanady (1971) at first considered barotropic boundary currents near a sloping shore. He postulates that a boundary current is generated by the arrival of water in the shore zone, either driven by wind or due to the passage of a long slow wave. Csanady shows that no narrow boundary currents are possible in the barotropic case. Then Csanady considered a two-layer system and showed that the analysis for the barotropic case with a modified scale factor can be used for the baroclinic case. Let ρ and ρ' be the densities of the top and bottom layers and h and h' be the depths. Then $\varepsilon = (\rho' - \rho)/\rho'$ is the fractional density defect. The top layer depth h is assumed to be uniform while the depth h' of the bottom layer is variable and is assumed to be

$$h' = \begin{cases} sy \text{ for } y \leq y_0 \\ h'_0 \text{ for } y \geq y_0 \end{cases} \tag{46}$$

where y denotes a coordinate normal to the shore and it is assumed that conditions are uniform along the shore (for example, $\partial/\partial x = 0$). Here, s is the beach slope.

Csanady showed that for this problem there are two important parameters, K and L where

$$L \equiv \left(\frac{g\varepsilon h}{f^2}\right)^{1/2} \tag{47}$$

and

$$K \equiv \frac{f}{2s}\left(\frac{h}{g\varepsilon}\right)^{1/2} \tag{48}$$

Next he defines

$$S \equiv \frac{sg\varepsilon}{f} \tag{49}$$

Here S is the slope length scale for effective gravity $g\varepsilon$ in the baroclinic case. Thus for the baroclinic case, there are two length scales S and L. A key parameter of this problem is κ defined in (48) which is also the same as

$$\kappa = \frac{L}{2S} \tag{50}$$

For James Bay, we can take

$$\begin{aligned} s &= 7 \times 10^{-4} \\ \varepsilon &= 10^{-2} \\ h &= 5 \text{ meters} \\ f &\sim 10^{-4} \text{ sec}^{-1} \end{aligned} \tag{51}$$

From (48) we get

$$\begin{aligned} \kappa &= 1/2 \\ L &= S = 7 \text{ km} \end{aligned} \tag{52}$$

Thus, the width scale of the boundary current L for James Bay is 7 km. Also the peak current occurs at a distance of order L from the point where the thermocline intersects the bottom.

6. Some miscellaneous topics

Here I will discuss the estuarine circulation in the rivers that drain into James Bay. Because of strong tidal action the saline water intrudes into the rivers and this might give rise to the type of estuarine circulation discussed by Hansen and Rattray (1965). They derived a set of coupled partial differential equations to study the circulation and salt flux processes for estuaries in which turbulent mixing results primarily from tidal currents and they separated the circulation into modes analogous to the barotropic, baroclinic and Ekman modes of oceanic circulation. These authors state that their solutions (although derived for the Delaware river estuary) hold well for sea straits and narrows having strong tidal currents and well-defined water and density budgets.

The present analysis may be of some use in understanding the dispersion processes and circulation in those regions of the rivers that are affected by strong tidal action from James Bay.

Hansen and Rattray divided the estuary into the three regimes, namely the outer, central and inner regimes, the outer being the closest to the connecting sea. For the circulation in the central regime they derived the following relation using similarity theory.

$$\phi(\eta) = \frac{1}{2}\underbrace{(2 - 3\eta + \eta^3)}_{\mathrm{I}} - \frac{T}{4}\underbrace{(\eta - 2\eta^2 + \eta^3)}_{\mathrm{II}} - \frac{\nu R_a}{48}\underbrace{(\eta - 3\eta^3 + 2\eta^4)}_{\mathrm{III}} \qquad (53)$$

where ϕ is a stream function, η is a non-dimensional vertical coordinate, T is a non-dimensional wind-stress, ν is a constant and R_a is an estuarine Rayleigh number. Thus, equation (53) expresses the circulation in the central regime as the sum of three modes: the river discharge mode (term I), the wind-stress mode (term II) and the gravitational-convection mode (term III). The important result is that only the river-discharge mode allows a net transport of water. In deriving equation (53) it has been assumed that, in the central regime the width and depth of the river as well as the river discharge are constant. This assumption would be more questionable both in the outer and inner regimes. In the next section (Section 7) we will speculate on the relative influences of these three modes in some of the rivers and how the situation may change due to the proposed hydroelectric power project.

Next we consider the question of atmospheric water balance. A study incorporating this along with the hydrology of the various river basins before and after the man-made changes will help to evaluate the effects of the man-made changes. At present, although in principle, the study could be carried out for the present situation, because of the necessity to handle a large amount of data and because of the time limitations on this study, this part of the study is not carried out. However, I will outline a procedure described by Rasmussen (1970) and speculate on this in the next section. Rasmussen has developed the following model to study the atmospheric water balance and hydrology of the upper Colorado River Basin. Since any climatic effects are intimately connected with the atmospheric water balance, the following procedure developed by Rasmussen to study the hydrometeorology of a given river basin using data on precipitation and evaporation will be especially useful. I quote from Rasmussen:

"Traditionally, studies of the hydrologic balance of river basins have been approached from the viewpoint of the terrestrial part of the hydrologic cycle. The factors determining the runoff from an area are precipitation, evaporation, change in ground-water storage, and underground seepage from the basin. Such an approach to the study of hydrologic problems is often plagued by measurement deficiencies... If a large mountainous region is studied, the measurement problem is maximized because not only is the density of observations low, but the observations are biased toward the lower elevations where precipitation is lower.

Alternately, the atmospheric part of the hydrologic cycle may be studied to evaluate the net deposition of water over an area. A budget parallel to that of the terrestrial part of the hydrologic cycle must be observed. The atmospheric water balance may be expressed as the evaporation minus precipitation occurring at the ground over an area balanced by the net transfer of water mass through the atmospheric volume over the area and by the change in storage of water mass within the atmospheric volume. In theory then, given a continuous distribution in time and space of the atmospheric water mass, an accounting can be done to determine as a residual the quantity evaporation minus precipitation. In practice, however, the distribution of water in the atmosphere is not continuously known; rather only water in the vapour state is sampled and only at time intervals of 12 hours and over distances of hundreds of kilometers. The problem is to approximate the water balance from this imperfect sampling procedure, realizing that the computation is only meaningful over sufficiently large areas and for sufficiently large weather systems."

In a coordinate system with pressure as the vertical coordinate, the time rate of change of water and water vapour can be written as:

$$\frac{d}{dt}(q+r) = \frac{\partial}{\partial t}(q+r) + \mathbf{V}_2 \nabla_2 (q+r) + \omega \frac{\partial}{\partial p}(q+r) \tag{54}$$

where q is the specific humidity, r is the ratio of the mass of water (either in liquid or ice form) to the mass of air, $\mathbf{V}_2$ is the velocity vector on a pressure surface and $\omega = dp/dt$ is the vertical velocity in the isobaric coordinate system. Equation (54) can be integrated from the earth's surface to some pressure level at which the amount of water in any form is negligible. Then, making use of the continuity equation and the divergence theorem due to Gauss one obtains the so-called atmospheric water balance equation:

$$E - P = \frac{1}{g} \int_{dp} \int_{dA} \frac{\partial}{\partial t}(q+r)\, dA\, dp$$
$$+ \frac{1}{g} \int_{dp} \oint_{dt} (c_n q + c_n r)\, dl\, dp \tag{55}$$

where P and E are the rates of precipitation and evaporation at the earth's surface, g is gravity, dA is an area increment on a pressure surface, dl is a line increment on the vertical boundary and c_n is the component of the wind vector $\mathbf{V}_2$ normal to the walls of the volume such that it is positive upward. If the integrals in (55) can be properly evaluated through observational data, the exchange of water and water vapour at the earth's surface, given by E - P can be determined as a residual.

One should be able to obtain the same exchange of water at the earth's surface from the hydrological balance provided we confine ourselves to surface waters. The hydrologic balance equation for a river basin is

$$P - E = R_0 + \Delta W + L \tag{56}$$

where R_0 is the runoff from the entire basin, ΔW is the total change in the surface and subsurface water storage and L is the depletion from the basin due to use within the basin itself or due to man-made diversion from the basin.

Thus the results from (55) and (56) should agree and this provides a check on the validity of the computation. Equation (56) is straightforward enough for computation but (55) has to be replaced by the following finite-difference form that is convenient for computation

$$P - E = -\frac{1}{g}\frac{\Delta}{\Delta t}\sum_{j=1}^{N}\sum_{i=1}^{M} q_{ij}\ \Delta A_{ij}\ \Delta P_{ij}$$

$$-\frac{1}{g}\sum_{j=1}^{N}\sum_{i=1}^{M} c_{nij}\ q_{ij}\ \Delta l_{ij}\ \Delta P_{y} \tag{57}$$

Here i denotes the grid point number (with maximum M) and j denotes the pressure level with N being the top most one.

7. Speculation on the possible effects due to man-made changes

It has been shown in Section 4 that the thermohaline processes could give rise to circulation as strong as that due to wind-generation in James Bay and the southern part of Hudson Bay. The discharge from the rivers flowing into James Bay will be effected when the project proposed by the Quebec government is carried out and it appears that towards the end of the winter season, the thermohaline processes could become stronger and thus further impede the movement of ice. Though this is highly speculative it appears as though the movement of ice in the southern part of Hudson Bay may be even more restricted towards the end of the winter season.

Another speculation could be made on the ice pressure. It appears from Russian work (Kagan, 1967a, 1967b) that because of tidal movement, ice could concentrate in the regions near amphidromic points. To illustrate this we reproduced Figures 2 from Kagan's reports of 1967a and 1967b to produce our Figure 6. The left side shows the cotidal lines for the M_2 tide in the Okhotsk Sea while the right side shows the lines of convergence and divergence in the ice pattern. Figure 7 shows the semi-diurnal tide in the James Bay-Hudson Bay-Hudson Strait complex (Dohler, 1964). This figure shows that there are no amphidromic points in James Bay while there are two in Hudson Bay. Hence it is possible that while strong and well defined zones of convergence and divergence may not occur in James Bay, these could occur in Hudson Bay and thus create ice pressure situation. However, the proposed hydroelectric power project may not effect the tidal effects on ice.

Although the total water discharged from the rivers into James Bay annually will be essentially unaffected by the hydroelectric power project, it is very likely that the amount of spring discharge into James Bay will be reduced. Because of the

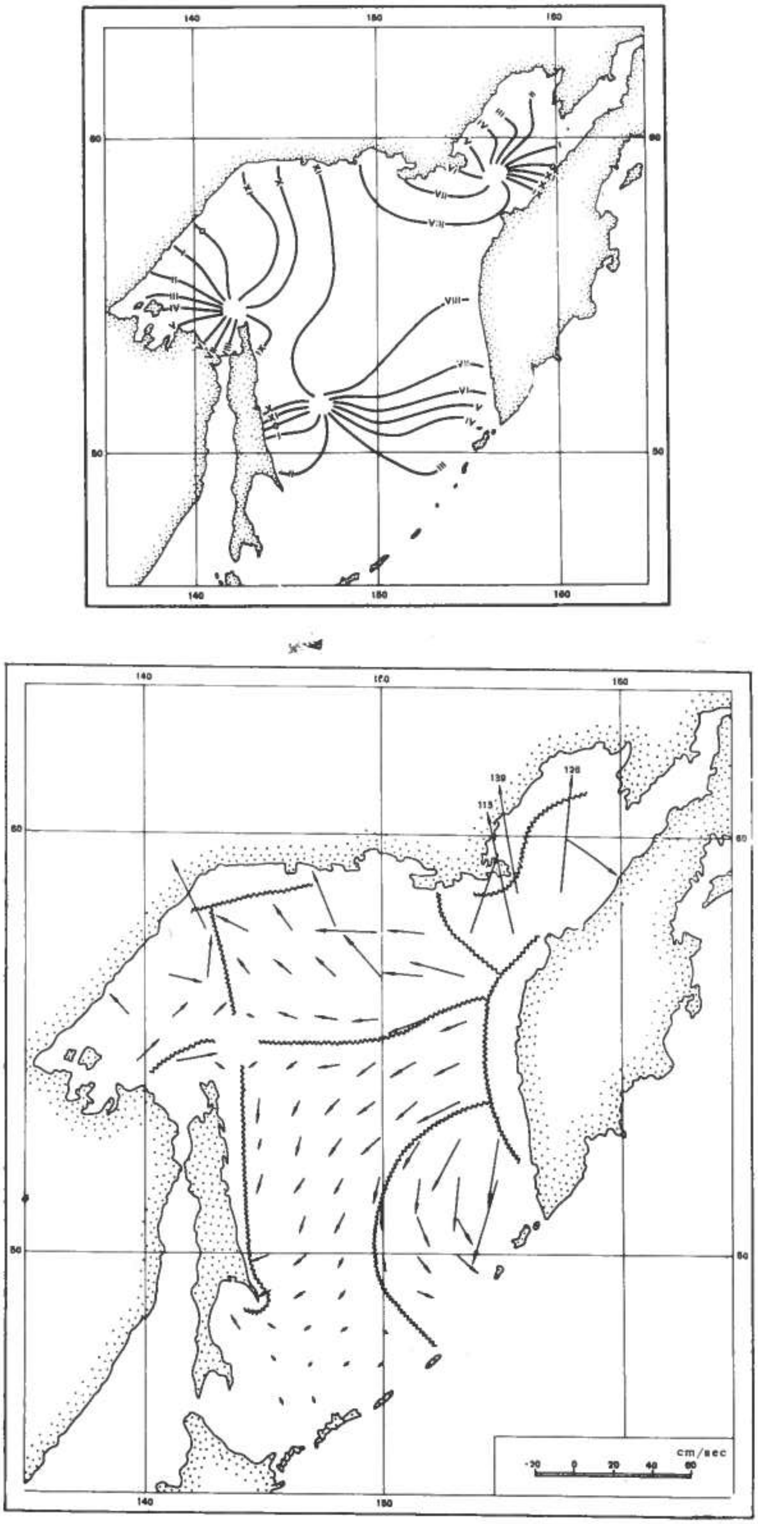


Figure 6 Cotidal lines for the M_2 tide (left side) and convergence-divergence pattern in the ice (right side) in the Okhotsk Sea.

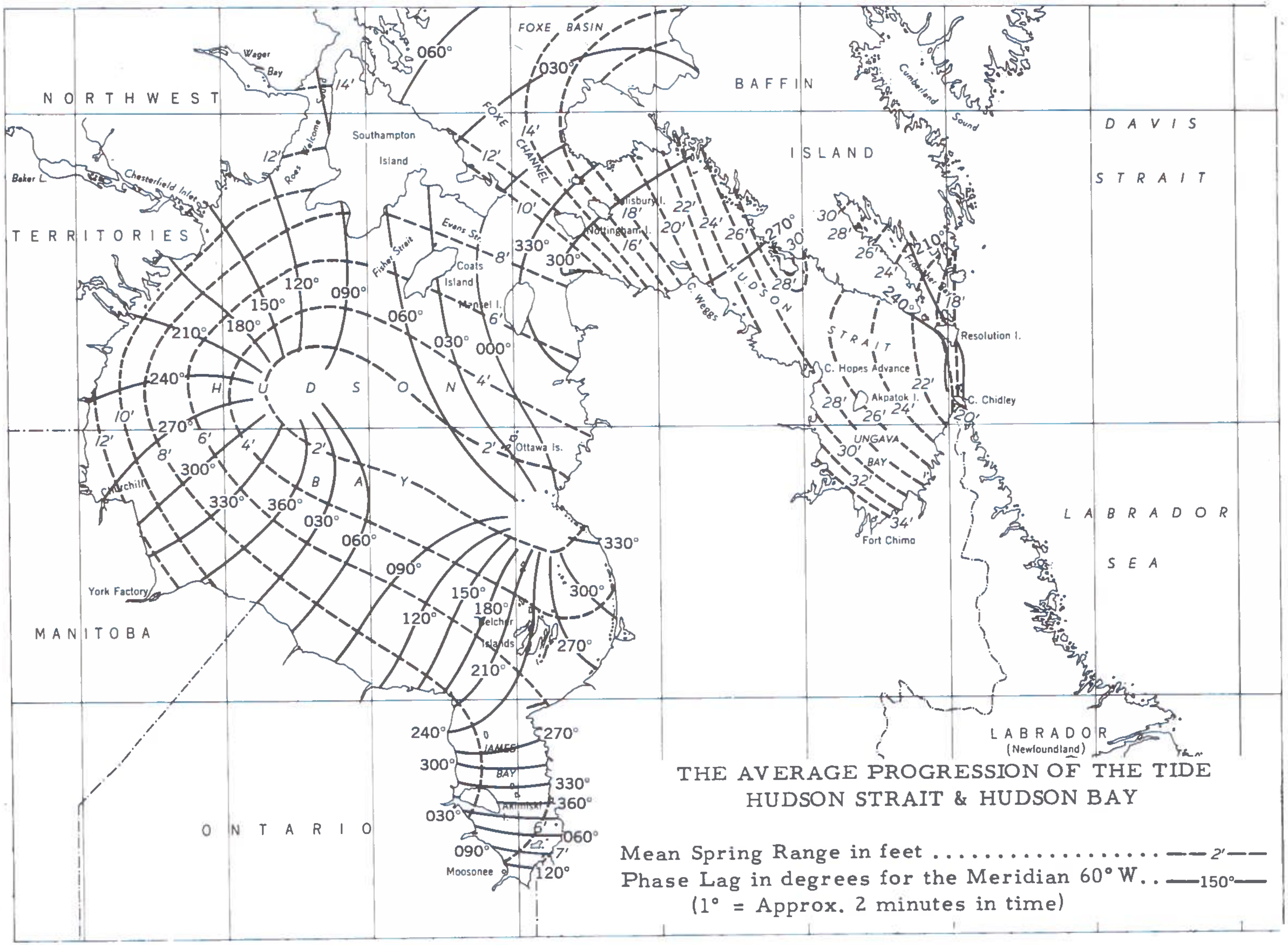


Figure 7 Cotidal lines for the semidiurnal tide in Hudson Bay-Hudson Strait-James Bay complex.

reduction in the freshwater content at the surface, the stability will be reduced and this may give rise to increased upwelling.

The storm surge activity is mainly due to wind stress effects and the effects of baroclinicity usually are negligible. Thus we do not expect any significant changes in the storm surge amplitudes after the construction of the hydroelectric power project.

It appears that the construction of the project may have noticeable effects on the baroclinic coastal jet in James Bay. Because of the reduced freshwater content, the density difference between the bottom and top layers may be reduced. Suppose ε is of the order of 5×10^{-3}, then the coastal jet will be somewhat weakened in the sense that it may have only a width of about 3.5 km.

Next we consider the effect of the project on the estuarine circulation in the central regime of the rivers. As was explained in Section 6 through equation (53), the circulation could be visualized as the combination of a river discharge mode, wind stress mode, and a gravitational convection mode. It is unlikely that the wind stress mode will be affected by the project. However, it is very likely that the dominant term will be the river discharge mode. Naturally the relative influence of this term varies not only with season but also from one river to the other. The gravitational convection mode may play an important role in determining the salinity pattern, noting that the strong tidal action will make some saline water intrude into the rivers.

One of the intractable problems a priori is the effect of the project on the local climate. Our approach to this is to calculate the atmospheric water balance and compare this with the hydrologic water balance. In the equation for the hydrologic water balance given by (56) allowance is made for the change in water content due to man-made projects (L). However, in the atmospheric water balance equation given by (55) any consequence of human interference has to be felt only indirectly. Thus the only meaningful way to attempt to answer this is to calculate the atmospheric water balance over the river basin system using data before and after the power project construction.

8. Acknowledgements

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10. List of Figures