IOS Note - 3

ASPECTS OF OCEANOGRAPHY IN THE ARCHIPELAGO

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PREFACE

This review of aspects of physical oceanography of the Canadian Arctic Archipelago, including some relevant meteorology and hydrology, is intended as a primer for those coming to work here. Since many will not be oceanographers an attempt has been made to minimize jargon.

Hopefully the note will be re-written in the not-too-distant future when results of work presently underway are analyzed and incorporated into our body of knowledge.

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INTRODUCTION

This note began as background material for studies of the Institute of Ocean Sciences, Patricia Bay, Canada, in the Queen Elizabeth Islands. However, like Topsy, it just grew, resulting in the present heterogenous form. However it has a plan. Chapter 2 gives a brief description of the Arctic Ocean whence most water in channels of the archipelago comes. Next is a brief description of the geography of the archipelago followed in Chapter 4 by a discussion of those aspects of meteorology most relevant to oceanography, temperature, precipitation, and the most important components of the surface heat budget, the radiation terms. In Chapter 5 aspects of freshwater runoff affecting water characteristics and locally, movement are noted. The tidal and water level material in Chapter 6 stresses longer period fluctuations, daily tide data being widely available. In Chapters 7 and 8 sea ice statistics and water currents are briefly noted. Water structure as characterized by temperature and salinity profiles is discussed in Chapter 9 while results of studies in three fiords or bays are outlined in Chapter 10. Budgets and principles of pollutant dispersion are next noted. Brief appendices outline equations governing water motion, give common conversion factors, and include a short gazetteer.

Contents largely concern the northern and western archipelago. The areas between the archipelago and Greenland have been recently studied by Sadler (1976) and Muench (1971). Barber (1967) has dealt with Hudson Bay. Chemical, biological, geological oceanography are not included in this note.

In Figure 1-1 is a map of the archipelago with place names. The early history of exploration of the area has been discussed by, amongst others, Dunbar and Greenway (1956), and Anon. (1959). The major objective for those first entering the area was to find the northwest passage, although Collin (1958) mentions that Parry obtained underwater temperatures as early as 1821. After conclusion of the search for Franklin, when it became evident that no practical northwest passage existed, emphasis shifted to the race to the north pole, the latter largely centered in the Ellesmere - west Greenland parts of the area we are discussing. These activities were tidied up by the time of the first world war, and exploration since has been small scale (and frequently by "amateurs"). From the mid-nineteenth century on, sporadically, expeditions with appreciable scientific intentions filtered into the archipelago. The first International Polar Year (1882-1883) was the initial large scale scientific effort with stations being occupied from Labrador to northeastern Ellesmere. During the twentieth century scientific effort increased slowly and perhaps pre-world war II effort can be considered to culminate in the Danish and American expeditions in the Baffin Bay area in 1928 (Riis-Carstensen, 1936).

Canadian efforts in arctic oceanography, which might have been considered to begin with Bernier from 1906, were stepped up in the 1930's particularly in the southeast arctic, the Hudson Bay, Foxe Basin area. During the second world war oceanography in the Canadian Archipelago effectively ceased. However shortly after the war a renaissance occurred. It may have been because of the establishment of the Fisheries Research Board's Eastern Arctic Investigations in 1947. At the same time weather stations in the Queen Elizabeth Islands were established jointly by the Canadian Meteorological

1-1 A map of the Canadian Arctic Archipelago with place names

Service and the U.S. Weather Bureau. This led to an opening of the archipelago to a good deal of work by the American agencies, the U.S. Coast Guard and Navy. Although most of the work was done in the eastern archipelago a certain amount took place in the Beaufort Sea area. American studies out on the perennial ice of the arctic ocean (in which Canada has never been involved to any extent) began at about this time. In the mid-fifties the entrance of Canadian vessels into the archipelago in a serious way, was signalled by the passage, on 1954, of H.M.C.S. Labrador through the northwest passage. In 1958, the year of the I.G.Y., the Polar Continental Shelf Project began its vital work of filling in details of bathymetry and oceanography in the northwest archipelago.

The spate of activity in the post war years established the gross outlines of oceanography of the archipelago and led to a number of review papers which appeared in the early 1960's (Collin and Dunbar, 1964). Since that time oceanography activity has in desultory fashion filled in details. Large projects have included the North Water Project which resulted in the studies of Muench (1971) and Sadler (1976) but lacked the continuing organization to obtain much needed year-round oceanographic measurements, although the meteorological program and measurements were more of a success. The other large project which included oceanography was the Beaufort Sea Project of 1974-1976. Apart from these, other studies have been smaller, sea ice work by groups from the Institute of Ocean Sciences, Patricia Bay, Marine Sciences, Ottawa, Defence Research Establishments, Pacific and Ottawa, McGill University, physical oceanography studies by groups from the same establishments, geological work by the group from AOL, a quick dash through the northwest passage by the Hudson in 1970, continuing accumulation of ice cover statics by the Polar Continental Shelf Project and the Atmospheric Environment Service, some acoustics work by groups from Defence Research Establishment Pacific. These projects have undoubtedly filled in some of the details left vague by the post-war studies.

To anticipate, after undertaking this review of physical oceanography in the archipelago, the author is of the opinion that the largest gaps occur in (a) knowledge of the currents over appreciable periods and areas, and (b) in details of local oceanography in areas of interest. From the view point of physical oceanography the instrumentation to fill these gaps would appear to be chains measuring current velocity (frequently where no magnetic direction sensors are practical), salinity, temperature and depth (or profilers if they turn out to be more practical). The effort of developing, acquiring and deploying such instrumentation in useful quantities is beyond the capabilities of any of the groups mentioned just above, so collaboration is obviously essential. Chemical and biological oceanography should be blended in more closely. The other gap quite noticeable in this review are the gaps or errors in budgets of various quantities, heat budgets and their components, sea ice, water flow and so on. The measurements mentioned above will also fill some of these gaps but combined meteorological/oceanographic studies are also necessary. Again, the effort is too large for any single group so again strong and stern compulsion to collaboration will be necessary to obtain efficient action.

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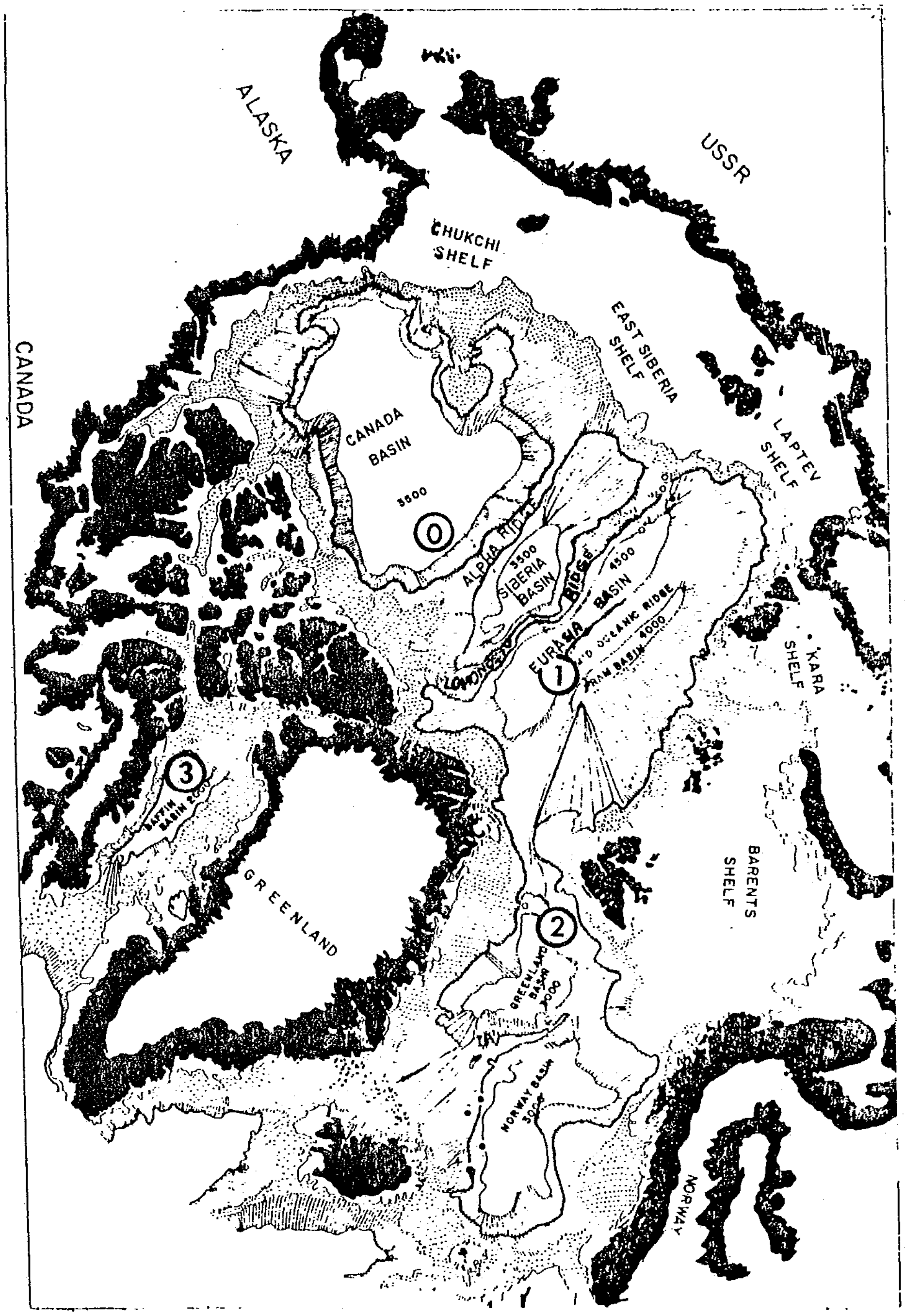
THE ARCTIC OCEAN

The Arctic Ocean must affect water flow and water structure in the Canadian Archipelago which the ocean borders on the northwest. Most comprehensive (and accessible) studies of the Arctic Ocean include those of Timofeyev (1960), Coachman and Aagaard (1974), and Treshnikov and Baranov (1972). The results of numerical modelling described by Semtner (1973, 1976) are very useful in supplementing the sometimes inadequate observation material upon which these surveys are based. Meteorological conditions and air-sea interactions have been reviewed by Doronin (1969) and Vowinckel and Orvig (1970) while the ocean's sea ice cover has been treated by Zubov (1943) and Koerner (1973).

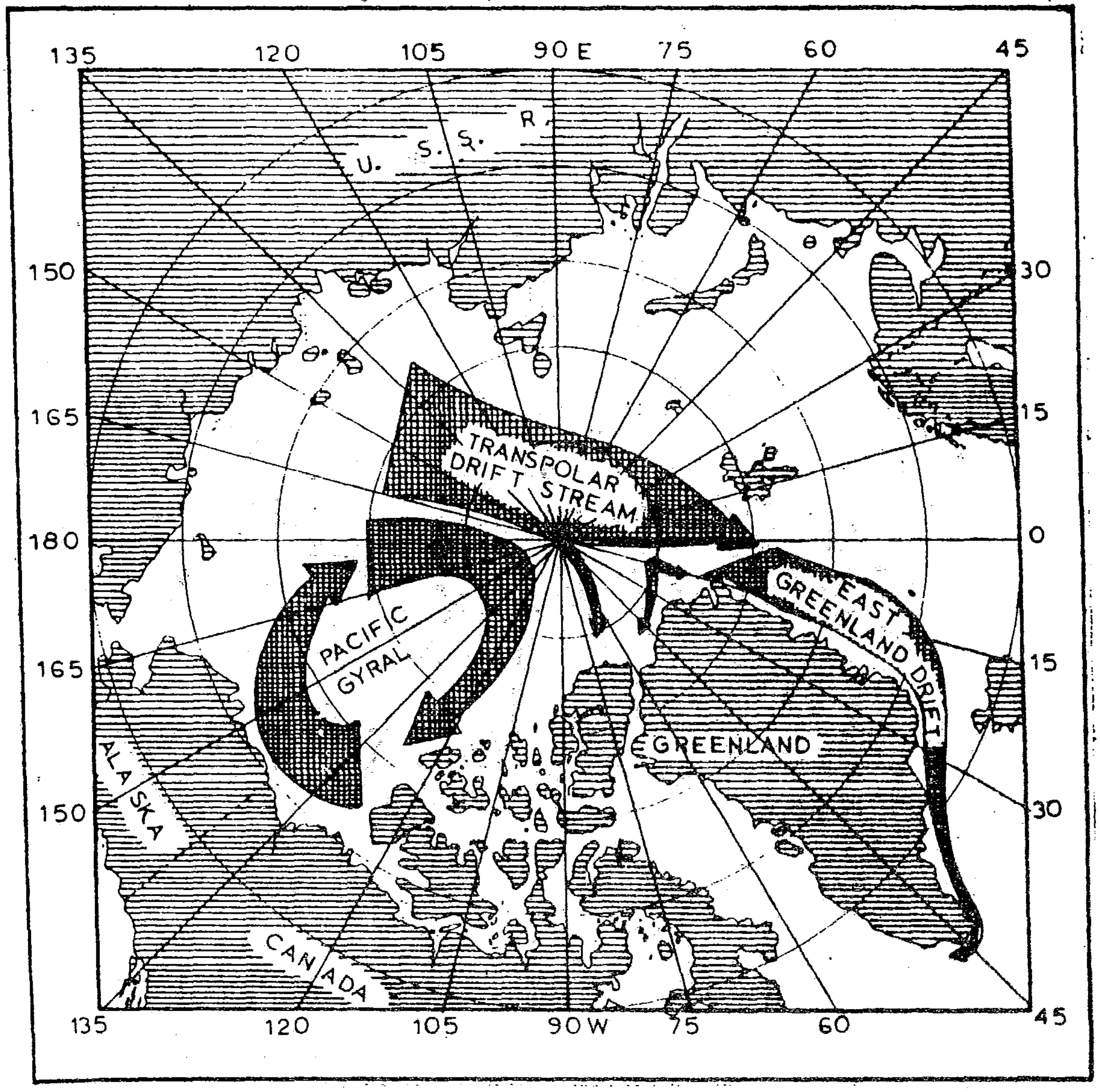
The bathymetry of the Arctic Ocean is shown in rather graphic detail in Figure 2-1 after Menzies (1963). The large scale features in Figure 2-1 are fairly well defined although nomenclature varies slightly between different authorities. The main basins are the Canada Basin separated by the Lomonosov ridge (depth about 1500 m) from the Eurasian Basin. Maximum depths are about 4000 m in the Canada Basin, and about 4500 m in the Eurasian Basin. Noticeable is the extensive continental shelf off the USSR, the shelf off the Canadian Arctic Archipelago being narrow. The main connection with the world ocean is that to the North Atlantic, the deepest section (2800 m) lying between Spitzbergen and Greenland. The connection to the Pacific through the Bering Strait is much smaller, width about 85 km, depth 45 m. The connecting passages through the Canadian Arctic Archipelago total (according to Collin, 1963) about 12 times the cross sectional area of Bering Strait and represent 6 percent of the total area through which exchange of the water can occur between the Arctic Ocean and the World Ocean.

The circulation of the surface water layers and its floating ice cover, generally conceded to be largely wind driven, is shown schematically in Figure 2-2 (after Dunbar and Wittman, 1963). All studies agree that the long term average circulation features which dominate surface water movements are the Pacific (or Beaufort) gyre and the transpolar drift. At depths below the surface layers the circulation is largely of water believed to come in through the north Atlantic connection (Timofeyev, 1960). The mid-depth circulation inferred from Atlantic water structural changes is shown in Figure 2-3 (after Coachman and Aagaard, 1974). The circulation in Figure 2-3 is much less certain than that in the surface layers. Numerical models of flow in the Arctic Ocean suggest that the flow at the depth of the Atlantic water may be clockwise in the Beaufort gyre region (Semtner, 1973). Measurements of currents are insufficient at the present time to resolve this question.

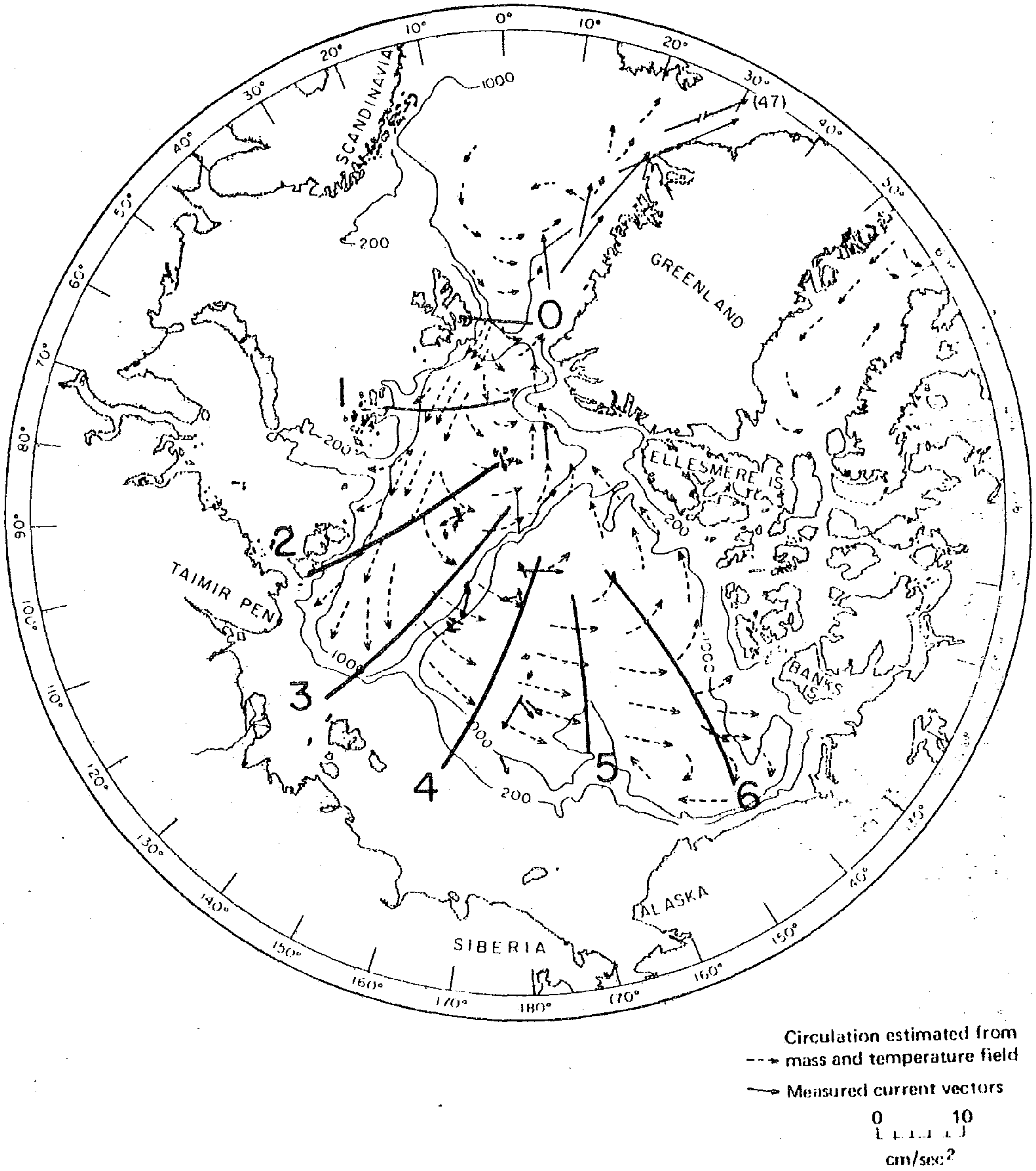
Measured and inferred flows to and from the Arctic Ocean have been used to draw up budgets of heat, water, salt content, and so on. Budgets drawn up from an oceanographers' viewpoint include those of Timofeyev (1960), Vowinckel and Orvig (1962), Mosby (1963), and Aagaard and Greisman (1975). The latters' budget, given in Table 2-1, estimates an annual volume outflow through the channels of the Canadian Archipelago of 2.1 Sverdrups (one Sverdrup



2-1 Schematic bathymetry of Arctic Ocean and surrounding seas. (after Menzies, 1963). The locations of the oceanographic stations in Figure 2-4 are shown.



2-2 Schematic of ice movements and surface water currents in the Arctic Ocean. (after Dunbar and Wittman, 1963).



2-3 Composite of mid-depth current measurements and circulation of Atlantic water in the Arctic Ocean, inferred from the temperature field (from Galt, 1967). Isochrones (years) after Timofeyev (1960).

equals $10^6 m^3 sec^{-1}$ or 31560 $km^3 year^{-1}$) based upon work by Muench (1971). The current measurements from the Greenland Spitsbergen passage (Aagaard, Darnall and Greisman, 1973) upon which the Aagaard and Greismans' budget was based indicated an annual variation in the rate of inflow of Atlantic Water. This had been remarked upon earlier, and some of the estimates of Timofeyev quoted in Panov and Shpaikher (1963) are shown in Table 2-2 with some year to year variations in Table 2-3.

TABLE 2-1

Source	Volume Transport, Sv	Heat Transport,* 10^9 kcal s^{-1}	Mean Temperature, °C	Salt Transport, 10^3 metric tons s^{-1}	Mean Salinity, ‰
Bering Strait					
Water	1.5	0.9	0.5	48.6	32.4
Ice	negligible	−0.4		negligible	
Arctic Archipelago	−2.1	1.3	−0.7	−71.8	34.2
East Greenland Current					
Polar water	−1.8†‡	2.0	−1.2	−61.2§	34.0
Atlantic water	−5.3†‡	−3.2	0.5	−185.0§	34.9
Ice	−0.1	8.0		−0.3	3.0
West Spitsbergen Current	7.1	16.3	2.2	248.9	35.0
Spitsbergen–Franz Josef Land	−0.1	−0.3	2.7	−3.5	34.9
Franz Josef Land–Novaya Zemlya¶	0.7‖	0.7	0.9	24.3	34.7
Runoff	0.1	0.5	5.0	0	
Total inflow	9.4		1.8**	321.8	34.6**
Total outflow	−9.4		−0.1**	−321.8	34.6**
Total advective heat gain		29.7			
Total advective heat loss		−3.9			
Net exchange	0.0	25.8		0.0	

Positive values are inflows, or heat gains. Negative values are outflows, or heat losses.
*Heat transport is relative to −0.1°C.
†Calculated from required combined water and salt transports.
‡The total value −7.1 is assumed in order to balance West Spitsbergen Current inflow.
§The total value −246.2 is calculated from salt continuity.
¶Karskiye Vorota Strait is included in this source.
‖Calculated from water continuity.
**Excluding runoff and ice transport.

Water, Salt and Heat Budgets for the Arctic Ocean (annual mean) after Aagaard and Greisman (1975).

Examination of the budgets in Table 2-1 indicates that in a gross sense the Arctic Ocean can be regarded as a large estuary. Runoff, 75 percent from Eurasia, and particularly the inflow through the Bering Strait which enters the Beaufort Gyre and mixes with the (relatively) fresh surface water downwelling there may be regarded as the input to surface transport out through Canadian Arctic Archipelago and by the Transpolar Drift to the East Greenland current. The estuarine "return flow" may be considered to occur mainly in the West Spitsbergen current inflowing between Spitsbergen and Greenland.

TABLE 2-2

Months	A	B	C	D
January	11.2	7.8	1.6	5.3
February	6.3	7.6	7.3	4.4
March	5.7	7.9	13.0	4.0
April	8.0	8.8	10.4	5.2
May	9.1	6.9	18.6	7.4
June	8.4	6.4	7.1	11.0
July	6.6	7.2	9.8	13.9
August	8.0	8.2	8.7	14.4
September	7.7	9.0	5.7	12.0
October	9.4	9.7	5.7	8.8
November	9.9	9.8	6.2	7.1
December	10.0	10.1	5.5	6.0

Long term average of variation of flow rates into the Arctic Ocean expressed as a percentage of annual inflow, (A) All Atlantic inflow (1902-1959), (B) inflow between Norway and Bear Island (1930-1961) (C) inflow between Bear Island and Spitsbergen (1931-1960), (D) inflow through Bering Strait (no dates) (after Timofeyev, 1963).

TABLE 2-3

Year	Influx of Water	Influx of Heat
1933	80	87
1934	92	125
1935	96	98
1936	122	150
1939	126	152
1952	89	125
1953	104	147
1954	125	122
1956	120	140
1957	83	88
1958	104	116
1959	90	94
1960	100	100
1961	120	127

Inter-annual variation of inflow of water and heat into the Arctic Basin expressed at percentage of 1960 inflow, (after Timofeyev quoted in Panov and Shpaikher, 1963).

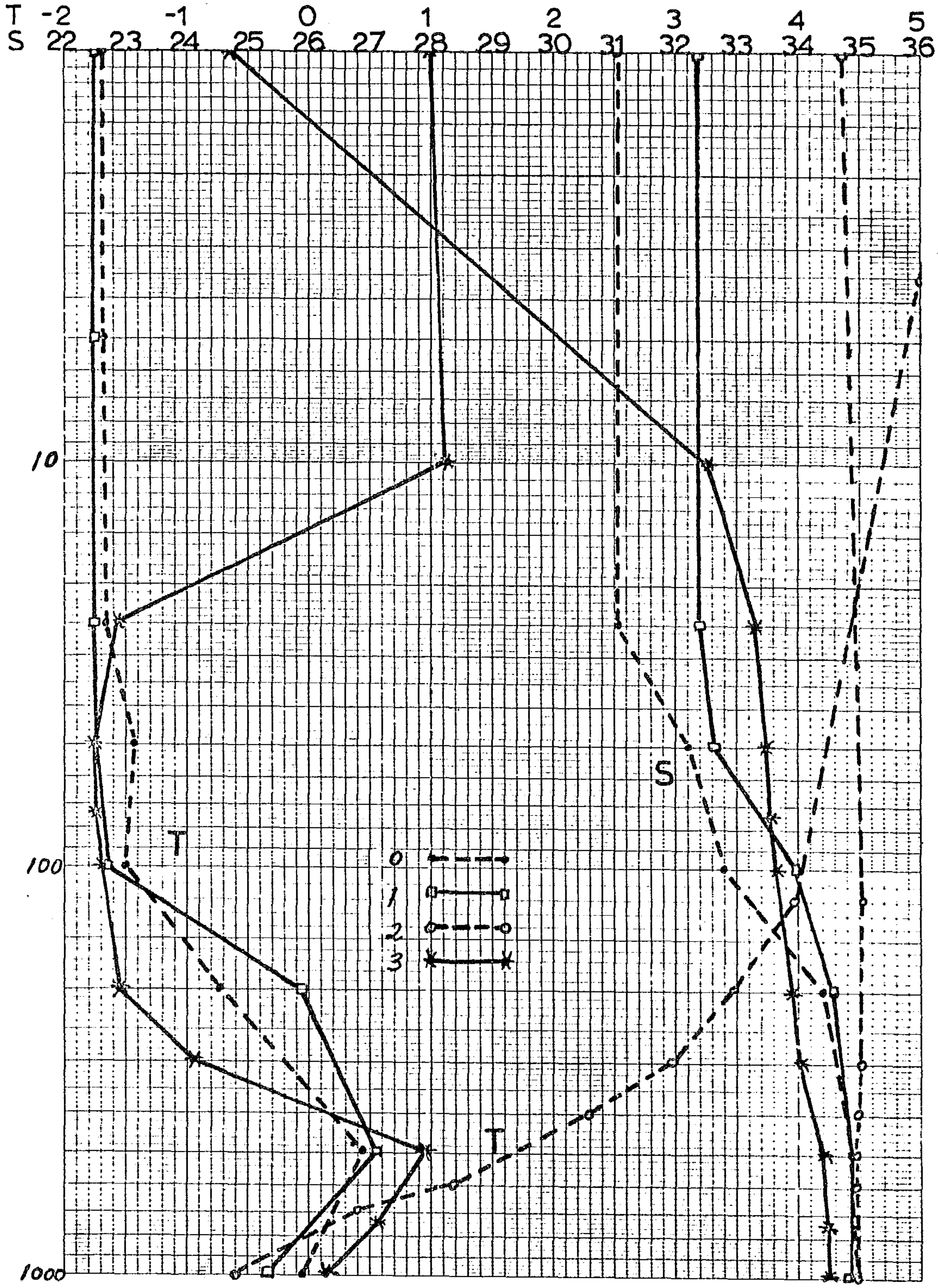
The circulations mentioned above, the characteristics of the water flowing into and out from the Arctic Basin, and the processes acting on the water within the basin result in water characteristics broadly outlined in Figure 2-4. In the diagram the values of temperature (°C) and salinity (g/kg) for the water column are plotted against the logarithm of water depth. Profile #0 whose position is shown in Figure 2-1 is typical of water in the Canada Basin. Profile #1 represents water from the Eurasia Basin. Profile #2 represents warm, salty, inflowing Atlantic water which passes below Arctic upper layers while circulating as shown in Figure 2-3. This Atlantic water forms the temperature maxima at around 500 m. shown in profiles #0 and #1. Profile #3 is a summer profile from Baffin Bay, and is similar to Arctic Ocean water although less saline at depth.

The water structures may be defined for the central Arctic as surface water, down to about 100 to 200 m, Atlantic water from perhaps 200 to 900 m and below that the bottom water which differs slightly from basin to basin. Bering Sea water from summer inflow can be identified at depths of about 50 - 100 m in the western part of the Beaufort gyre (Coachman and Barnes, 1961). The Arctic Ocean surface water is subject to the same influences as are the surface layers of water in the archipelago. The influences due to runoff are described in Chapter 5, those due to sea ice formation in Chapter 6 and some indication of the types of interactions with the atmosphere are given in Chapter 10. Profiles of Arctic Ocean water just offshore the Canadian Archipelago are presented in Chapter 9.

Tidal data and water level data for the Arctic Ocean is scanty. According to Coachman and Aagaard (1974) tidal waves come from the Atlantic and sweep counter clockwise around the Arctic Basin. Tidal ranges along the Norway coast are typically 2 to 3 m, in Spitsbergen 1 to 1.5 m, and at Point Barrow 0.25 to 0.35 m. The largest variations in sea level are probably aperiodic and occur on time scales of a few days to weeks, Matthews (1970), Hume and Schalk (1967). The annual variation of sea level has been discussed by Lisitzin (1961, 1964) and Beal (1968) who found lowest levels in February-March, the highest levels in the "Pacific" sector in August-September, and the highest levels in the Atlantic, December-January. The annual range was about 0.25 to 0.30 m for these monthly mean values of sea level, and are explained only in part by annual variations in atmospheric pressure. Variations of other periods have been considered for the Arctic by Borisov (1975).

The central Arctic Basin is covered by a permanent ice sheet, though of course, there are many leads constantly opening and closing even in winter. Because of the persistence of this ice cover (Koerner, [1973] estimated 83 percent of ice in the Beaufort gyre, and 73 percent of ice in the Transpolar drift is ice over one year old), the surface heat balance allows growth to an extent that Koerner (1973) estimated average ice thickness of 3.7 m in the Transpolar drift, 4.3 m in the gyre, at the end of winter growth period. The thermodynamics of such ice growth have been discussed by Maykut and Untersteiner (1971).

In the references given there are brief discussions of variability of conditions within the Arctic Basin over periods extending from hours to the eons of climatic time. Elements discussed include currents in the Arctic Ocean (Coachman, 1969), water levels (Lisitzin 1961, 1964) ice



2-4 Vertical profiles of temperature and salinity at the locations shown in Figure 2-1. Profile #0 is in Canada Basin, Profile #1 in the Eurasia Basin, Profile #2 inflowing Atlantic water, and Profile #3 in Baffin Bay.

cover Gudkovich et al (1970) inflow (Timofeyev, see Tables 2-2, 2-3), and atmospheric heat (Oort, 1973). As is the case for similar studies elsewhere the data base is inadequate for a really useful synthesis of these fluctuations, over the range of time scales involved.

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GEOGRAPHY OF THE ARCHIPELAGO

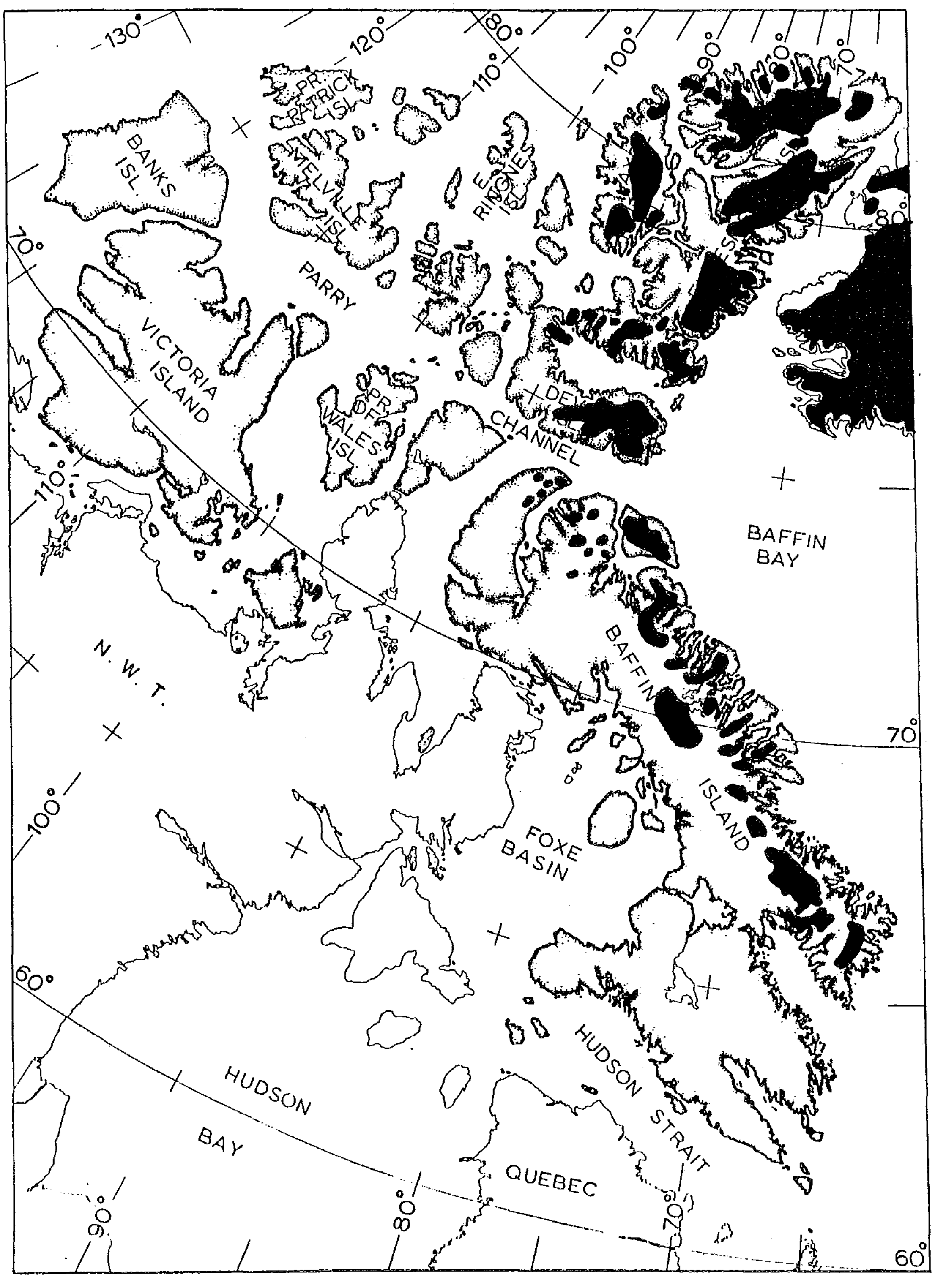
The Canadian Arctic Archipelago is shown in Figure 3-1. The area, within headlands, north of Parry Channel is about $7 \times 10^5 km^2$ of which about $4 \times 10^5 km^2$ is land. The surface area of Parry Channel itself is about $2 \times 10^5 km^2$. The surface area of the archipelago, south of Parry Channel, north and west of Baffin Island, is about $8 \times 10^5 km^2$ half of which is land. The area of Baffin Island itself is about $5 \times 10^5 km^2$.

Topography east of 100°W is mountainous and extensive glaciation exists, as shown in Figure 3-1. Fiords, as also shown in Figure 3-1 are common in the eastern Arctic Islands. West of 100°W the terrain is much flatter, and without glaciers.

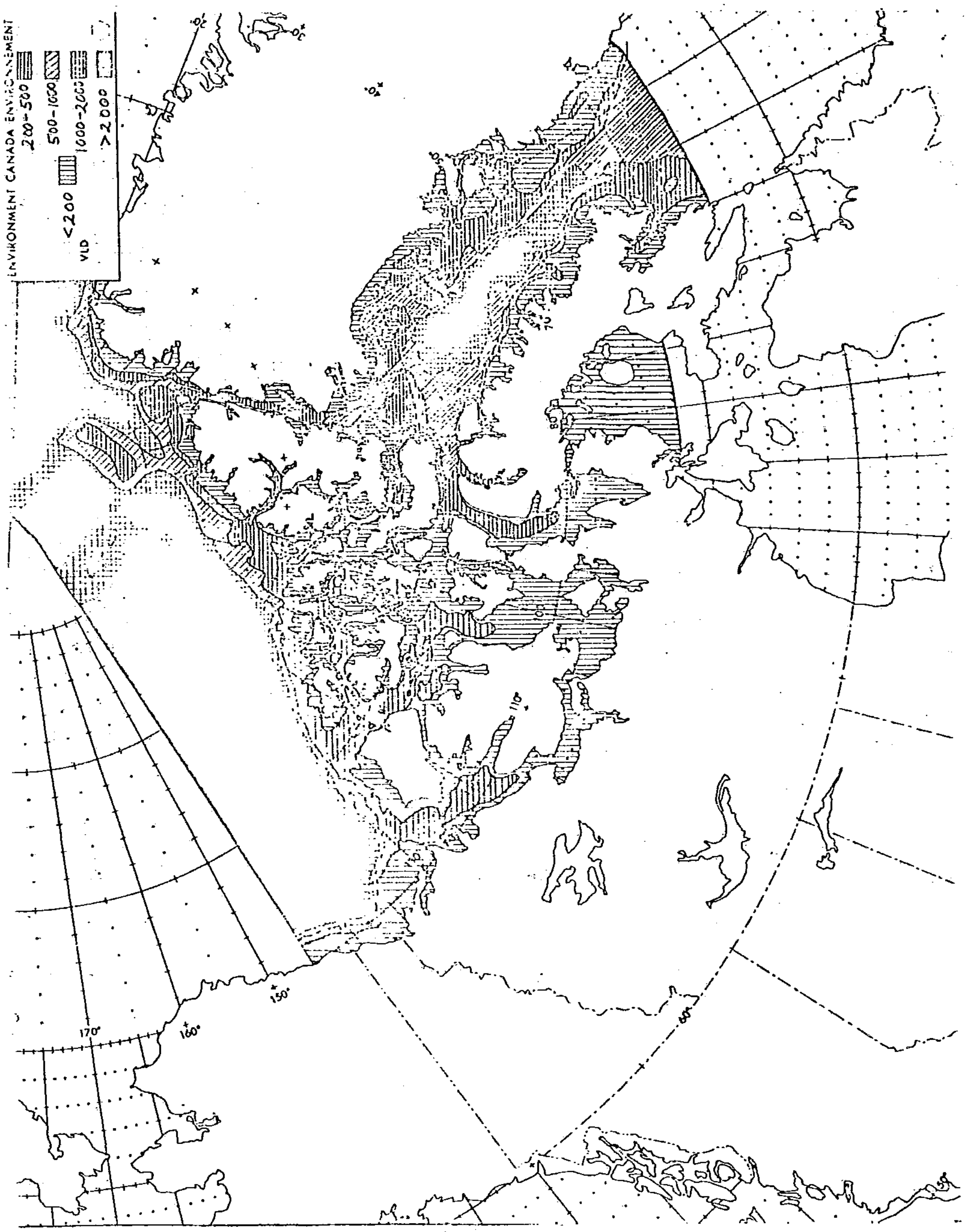
The bathymetry of channels in the archipelago is shown in Figure 3-2, (after Chart 800-A of the Canadian Hydrographic Service). The typical depth of the continental shelf bordering the western Arctic Islands is about 500 m and the width ranges from about 200 km in the Beaufort Sea to only a few km off northern Ellesmere Island. A canyon through the continental shelf runs north eastward from Nares Strait. Water depth in most of the channels within the archipelago is shallower than the continental shelf, but depressions occur, notably to about 1000 m in the Nansen Sound - Greely Fiord system, to 700 m in the Sverdrup Islands, to 600 m in the Prince Gustaf Adolf Sea, to 600 m in McClure Strait, to 700 m in Viscount Melville Sound, presumably due to glacial overdeepening. The depression of over 2000 m depth in Baffin Bay has a canyon of depth over 500 m leading westward into Lancaster Sound. A 900 m depression, apparently not connected at such depth to Baffin Bay, occurs in Jones Sound.

It is evident that flow through the archipelago between the Atlantic off southern Greenland and the Arctic Ocean is restricted to a few channels. These are Fury and Hecla Straits, Lancaster Sound, Jones Sound, and the Nares Strait passage between Ellesmere Island and Greenland. The combined area of these straits has been compared by Collin (1963) to a single waterway 120 km wide and 340 m deep. However, Chart 800-A indicates there are a number of sills, across the channels upstream from the straits mentioned earlier, that will restrict water flow to depths shallower than that suggested by Collin. These sills, with maximum depths estimated from the appropriate hydrographic charts are detailed in Table 3-1. The values there would suggest the only connection between the Arctic Ocean and the Atlantic Ocean deeper than 200 m occurs in Nares Strait, the name given to the collection of passages between Ellesmere Island and Greenland, as Parry Channel is frequently used for the passages from Lancaster Sound to M'Clure Strait. Elsewhere depths at Hell Gate, Penny Strait, Barrow Strait, Bellot Strait and Fury and Hecla Strait appear to be less than 200 m.

Figure 3-2 indicates multiple connections from the Arctic Ocean into the northwestern archipelago. The large areas of very shallow water in the southern archipelago are noteworthy.



3-1 The Canadian Arctic Archipelago showing glaciated areas in black, and the indented coastlines where fiords occur.



3-2 Bathymetry of the channels of the Canadian Arctic Archipelago. Depth in m.
(after Canadian Hydrographic Service, Bathymetric Chart 800-A).

3-2 Bathymetry of the channels of the Canadian Arctic Archipelago. Depth in m. (after Canadian Hydrographic Service, Bathymetric Chart 800-A).

TABLE 3-1

Area	Lat. (°N) (approx)	Long. (°W)	Estimated Sill Depth (m)
S.E. Victoria Island	71	101	50
Prince of Wales Strait	73	118	40
Barrow Strait	74	94	140
Bellot Strait	72	95	40
Fury and Hecla Strait	70	84	50
Nares Strait	80	70	250
Hell Gate	75	91	100
Eureka Sound	79	85	60
Nansen Sound Entrance	82	93	450
Sverdrup-Peary Channels	80	102	400
Prince Gustaf Adolf Sea Entrance	79	107	400
Ballentyne Strait	78	115	100
Wilkins Strait	78	112	140
Penny Strait	77	97	100
Byam Strait	75	105	100
Austin Strait	75	103	120
FitzWilliam Strait	77	116	250
Continental Shelf	-----		200-600
Davis Strait	66	58	700

Sill depths in the Canadian Arctic Archipelago (m) as estimated from appropriate charts of the Canadian Hydrographic Service.

METEOROLOGY OF THE ARCHIPELAGO

Before 1930 knowledge of the meteorology of the archipelago was derived from explorers' journals. About 1930 several weather reporting stations were established on the shores of Hudson Bay, on Baffin Island and along the mainland Arctic coast. In the late 1940's five weather stations were established in the Queen Elizabeth Islands. In the late 1950's weather reports from the Distant Early Warning (DEW) Line of radar stations extended across the Arctic at about 70°N. After 1960 radiation measurements were taken at several weather reporting stations in the Arctic. The present network of weather observing stations in the archipelago is shown in Figure 4-1.

Meteorological data from these stations are published by the Atmospheric Environment Service, Environment Canada, Downsview, Ontario. Their publications are too numerous to list, but useful summaries include temperature and precipitation normals 1941-1970, wind normals 1955-1972 and monthly radiation bulletins. It is understood that a definitive climatological study is presently underway (M. Berry, personal communication), which will summarize useful meteorological data for the archipelago.

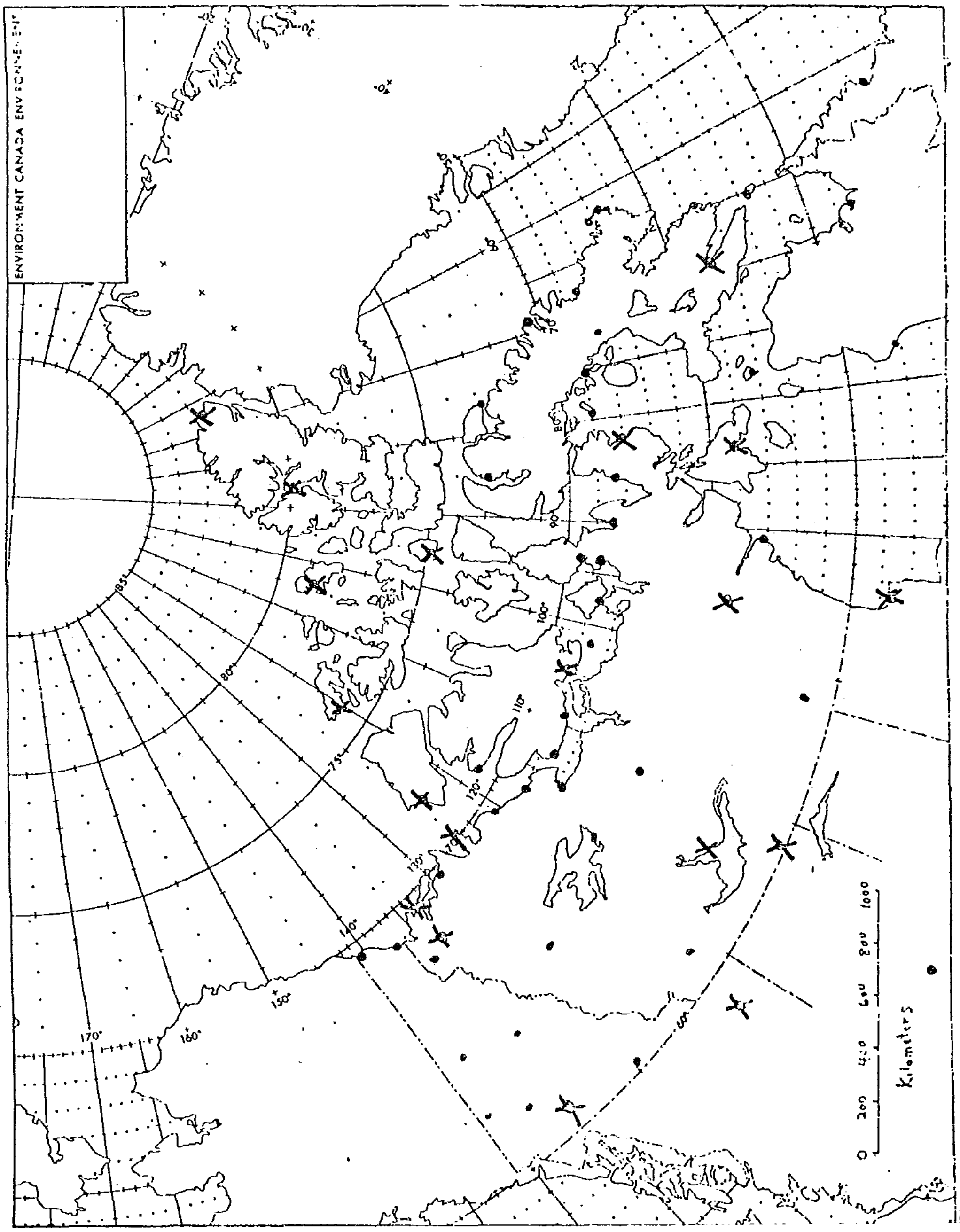
Meanwhile until this definitive study is available, the brief outline of those aspects of meteorology relevant to the oceanography of the archipelago is rather reluctantly presented below.

Atmospheric Pressure and Winds

The balance of forces in the momentum conservation equation A-1, for the atmosphere, is such that to a crude approximation winds tend to blow along isobars of atmospheric pressure, with high pressure to the right of the direction towards which the wind is blowing. The large scale distribution of atmospheric pressure at sea level is shown in Figure 4-2, in (a) for the winter season and in (b) for the summer season. These charts indicate in some sense a mean northerly flow across most of the archipelago, stronger in winter than summer. Low pressure in Baffin Bay is shown in the January map. Superimposed on the large scale mean pressure fields shown in Figure 4-2 are the travelling cyclones and anticyclones with their wind systems. An indication of the frequency of these travelling disturbances in winter and summer is given in Figure 4-3, (Anon, 1970). The presence of these travelling weather systems results in wind roses, of directions from which winds blow as shown in Figure 4-4. This figure showing variability only in direction must be supplemented by data giving both wind direction and speed as shown for Resolute in Table 4-1, (Anon., 1975). The mean wind speed over the year for Resolute is 13.7 mph or 22 km hr^{-1}. Wind records, perhaps more than most weather records are influenced by conditions at the recording station (vide wind directions at Frobisher Bay in the southeastern corner of Figure 4-4).

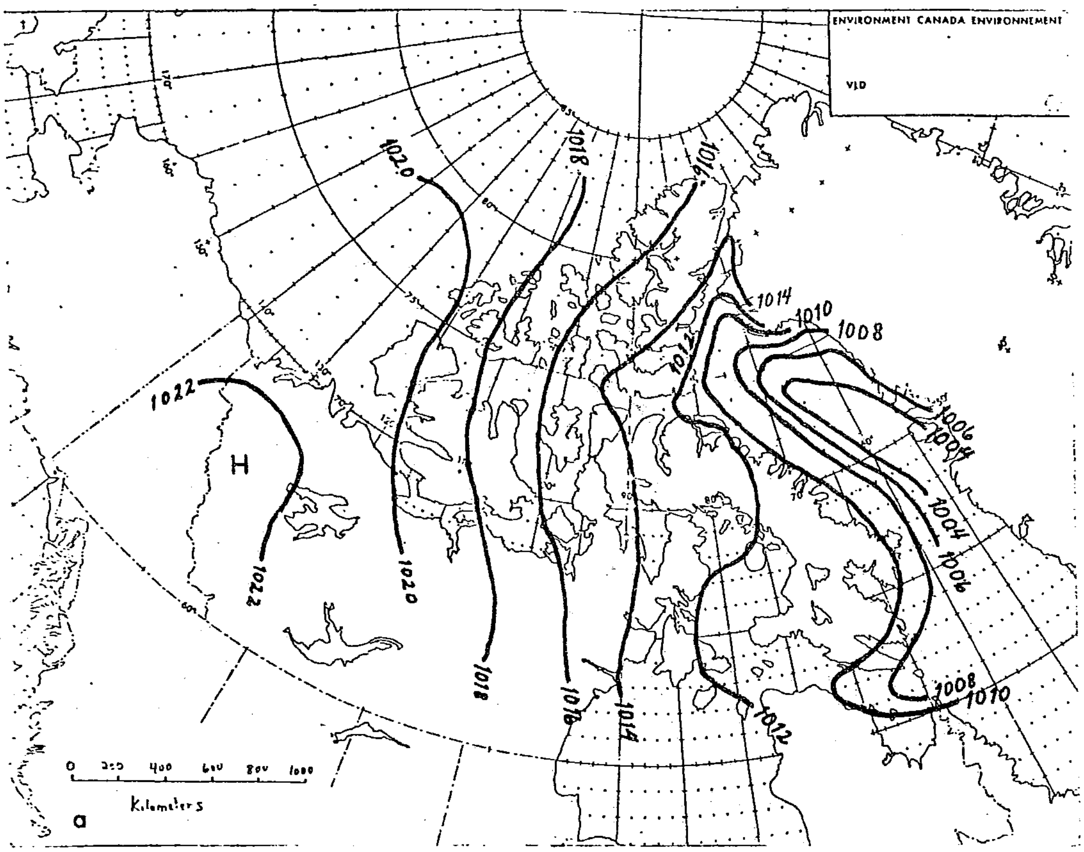
Temperature and Precipitation

The heat balance at the earth's surface and in the lower atmosphere is such that winter temperatures <u>average</u> -20 to -30°C over the archipelago and

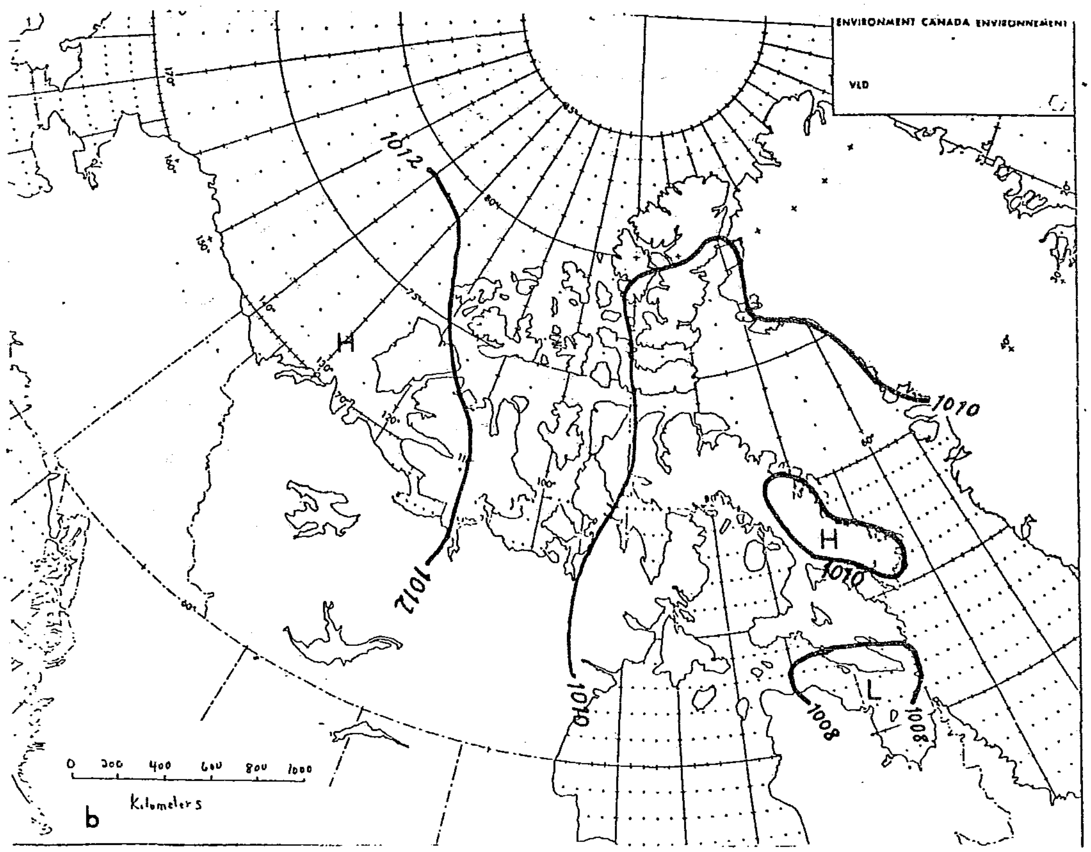


4-1 Atmospheric Environment Service weather reporting stations in and around the Canadian Arctic Archipelago. Radiation measuring stations marked with a cross.

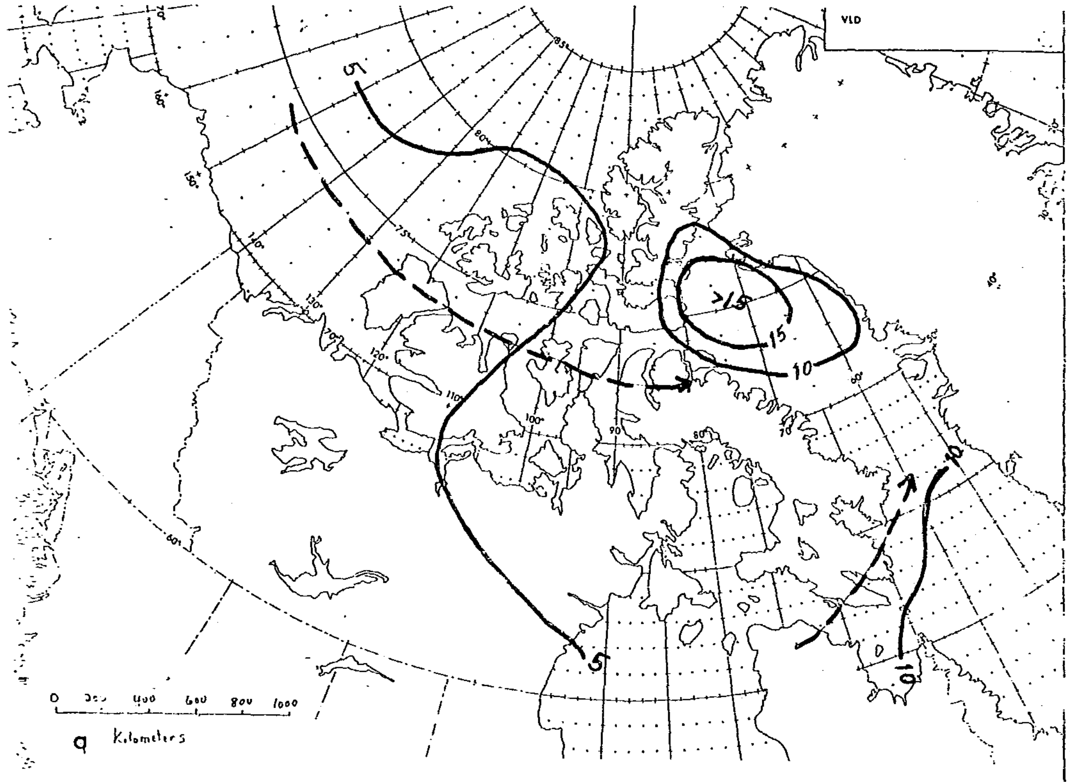
4-1 Atmospheric Environment Service weather reporting stations in and around the Canadian Arctic Archipelago. Radiation measuring stations marked with a cross.



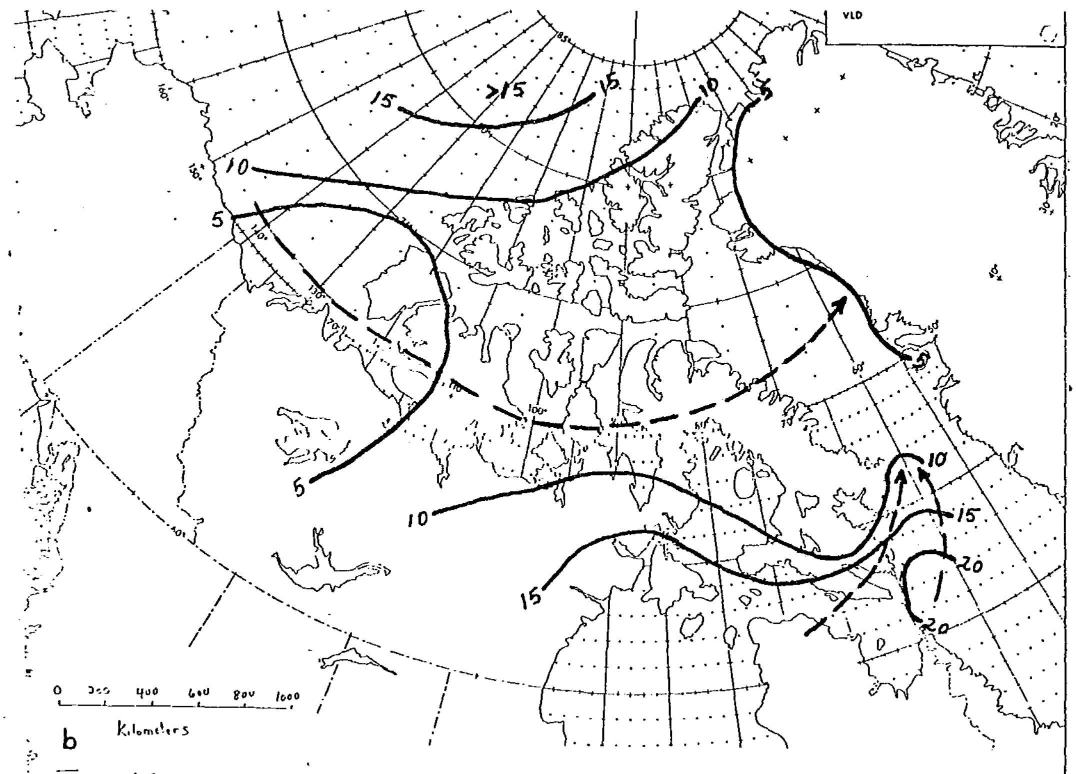
4-2 Mean sea level atmospheric pressure (mb). (a) January 1931-1960, (after AES data).



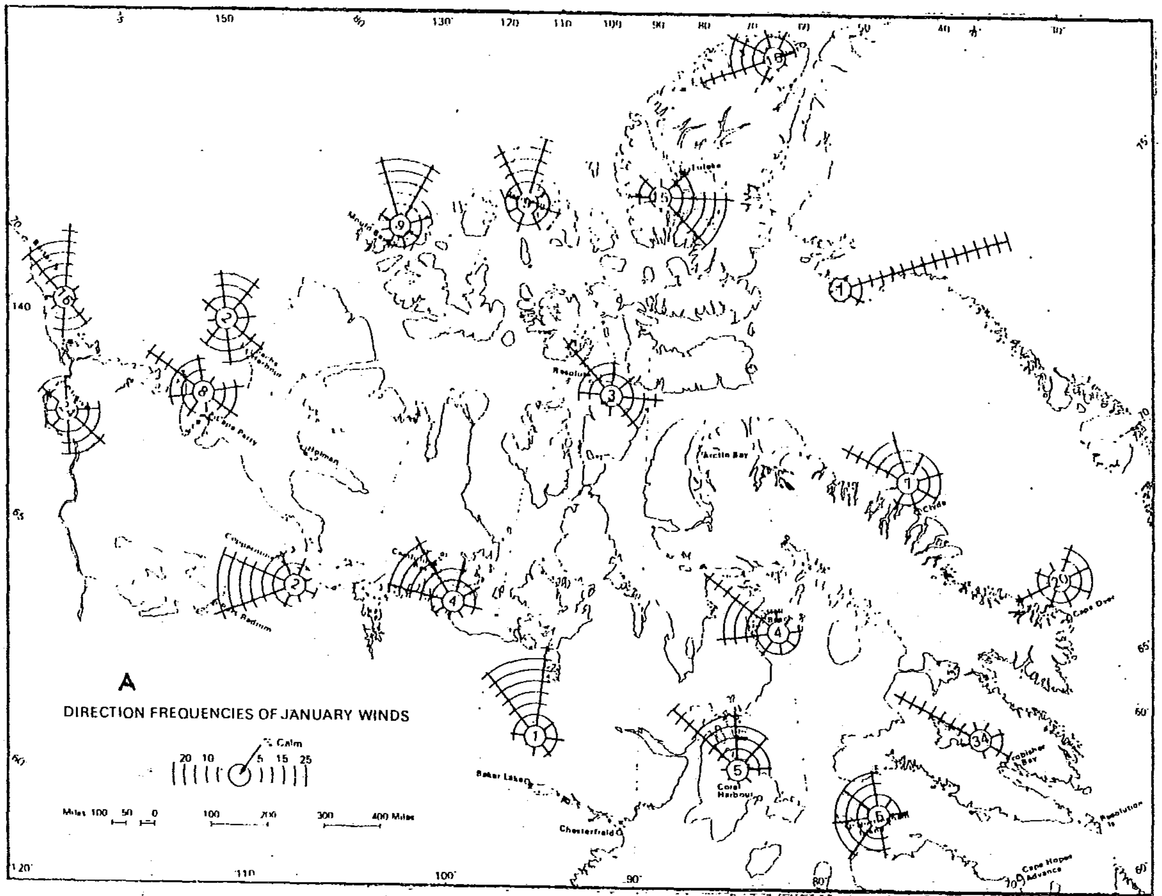
4-2 Mean sea level atmospheric pressure (mb). (b) July 1931-1960. (after AES data).



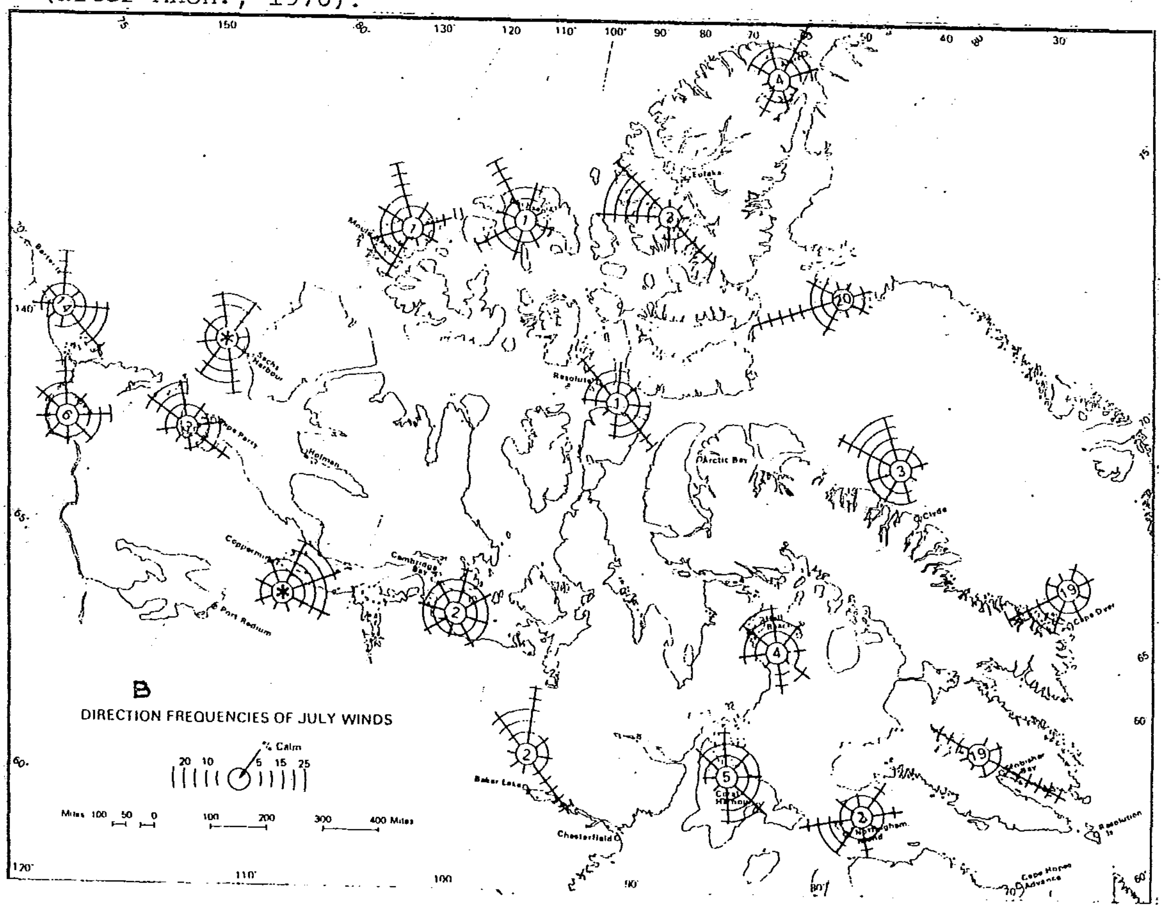
4-3 Percentage frequency of cyclone centers per 250,000 mile squares and principal storm tracks. (a) February, (after Anon., 1970).



4-3 Percentage frequency of cyclone centers per 250,000 mile squares and principal storm tracks. (b) August. (after Anon., 1970).



4-4 Wind roses, giving direction frequencies of winds. (a) January, (b) July. (after Anon., 1970).



RESOLUTE (A), N.W.T.

PERIOD 1963-72 PÉRIODE — HEIGHT OF ANEMOMETER 40' HAUTEUR DE L'ANÉMOMÈTRE

PERCENTAGE FREQUENCY — FRÉQUENCE EN %

	JAN JANV	FEB FEV	MAR MARS	APR AVR	MAY MAI	JUN JUIN	JUL JUIL	AUG AOUT	SEP SEPT	OCT OCT	NOV NOV	DEC DEC	YEAR ANNUEL	
N	8	9	8	7	9	10	6	8	18	11	9	8	9	N
NNE	5	4	3	4	6	6	5	4	8	5	5	3	5	NNE
NE	8	8	5	5	5	6	6	4	10	8	10	5	7	NE
ENE	3	3	4	3	3	2	4	2	3	4	5	3	3	ENE
E	5	5	5	6	5	4	6	9	6	6	9	6	6	E
ESE	8	10	10	10	7	7	6	11	6	9	8	9	8	ESE
SE	8	7	11	10	11	8	9	11	5	9	7	11	9	SE
SSE	3	2	4	6	5	4	7	7	2	3	2	4	4	SSE
S	3	3	3	3	4	3	4	5	4	3	3	4	4	S
SSW	1	1	1	1	3	1	1	1	2	3	1	2	1	SSO
SW	1	1	1	1	2	1	1	1	2	3	1	1	1	SO
WSW	*	1	1	1	2	2	3	2	1	3	1	1	1	OSO
W	3	4	2	3	3	8	11	6	3	5	2	3	4	O
WNW	3	5	4	4	5	9	8	8	3	2	2	3	5	ONO
NW	11	11	12	9	9	12	8	8	9	9	9	9	10	NO
NNW	15	15	12	12	13	13	9	8	15	12	12	15	13	NNO
Calm	15	12	14	15	8	4	6	5	3	5	14	13	10	Calme

AVERAGE WIND SPEED IN MILES PER HOUR — VITESSE MOYENNE DES VENTS EN MILLES/HEURE

	JAN JANV	FEB FEV	MAR MARS	APR AVR	MAY MAI	JUN JUIN	JUL JUIL	AUG AOUT	SEP SEPT	OCT OCT	NOV NOV	DEC DEC	YEAR ANNUEL	
N	14.0	16.7	13.5	11.9	14.6	14.9	16.0	14.1	16.6	13.4	10.0	11.9	14.0	N
NNE	31.0	25.0	20.7	15.7	19.9	16.2	16.6	17.6	17.8	21.2	16.4	19.0	19.8	NNE
NE	26.8	25.4	20.7	20.1	19.3	17.7	16.4	16.8	18.8	22.1	22.0	23.8	20.8	NE
ENE	20.2	19.9	20.1	16.9	15.8	16.9	18.1	16.3	17.2	18.6	16.2	19.2	18.0	ENE
E	19.5	21.9	23.5	24.2	19.5	18.5	23.9	25.6	21.9	20.9	24.5	24.5	22.4	E
ESE	15.4	19.5	22.0	22.1	20.0	22.6	18.6	21.9	18.7	21.5	18.9	16.5	19.8	ESE
SE	13.6	15.4	15.9	13.4	16.9	17.4	12.7	12.2	14.1	18.0	15.2	15.3	15.0	SE
SSE	14.2	12.2	13.1	12.5	12.6	11.2	9.5	10.1	12.3	14.0	12.0	13.9	12.3	SSE
S	13.3	12.8	14.5	12.1	13.1	9.6	7.6	9.2	13.7	13.5	13.3	16.1	12.4	S
SSW	13.4	12.9	16.4	11.5	12.6	11.0	9.0	11.1	14.3	13.5	14.1	13.6	12.8	SSO
SW	10.7	7.1	11.4	9.4	10.4	12.7	8.3	11.7	11.3	12.3	9.6	11.5	10.5	SO
WSW	7.9	9.3	8.6	11.2	9.9	9.9	10.2	12.3	12.7	13.5	9.2	11.9	10.6	OSO
W	8.8	8.3	8.5	8.8	7.9	9.3	9.6	10.0	11.0	12.8	7.7	9.3	9.3	O
WNW	9.5	8.3	9.3	9.1	8.2	11.2	8.7	10.4	12.6	10.2	8.4	10.3	9.7	ONO
NW	11.3	11.5	11.0	9.3	11.0	12.4	10.7	10.6	15.0	13.3	11.7	11.6	11.6	NO
NNW	12.6	13.9	12.0	10.5	12.9	13.8	16.2	13.9	16.0	14.2	11.0	12.0	13.3	NNO
All Directions / Toutes directions	13.5	14.1	13.4	12.3	13.5	13.8	12.6	14.0	15.9	15.6	13.1	13.0	13.7	

Maximum Observed Hourly Speed	88 E	Vitesse horaire maximale observée
Maximum Observed Gust Speed	98	Vitesse maximale observée des rafales
Probable Maximum Gust for Maximum Hourly Speed	119	Rafale maximale en rapport avec vitesse maximale des vents horaires

4-1 Mean wind data over 1963-1972 at Resolute Airport weather station. (a) Percentage frequency by direction, (b) Average wind speed (mph) by direction. (from Anon., 1975).

summer temperatures rise only to 5 to 10°C (Figure 4-5). The annual course of average temperatures at selected stations over the archipelago is shown in Figure 4-6. The shortness of the warm season shown by these curves is emphasized by the dates in Figure 4-7 on which the air temperatures pass above freezing. The mean numbers of degree days (below 0°C) and of thawing degree days (above 0°C) are shown in Figure 4-8.

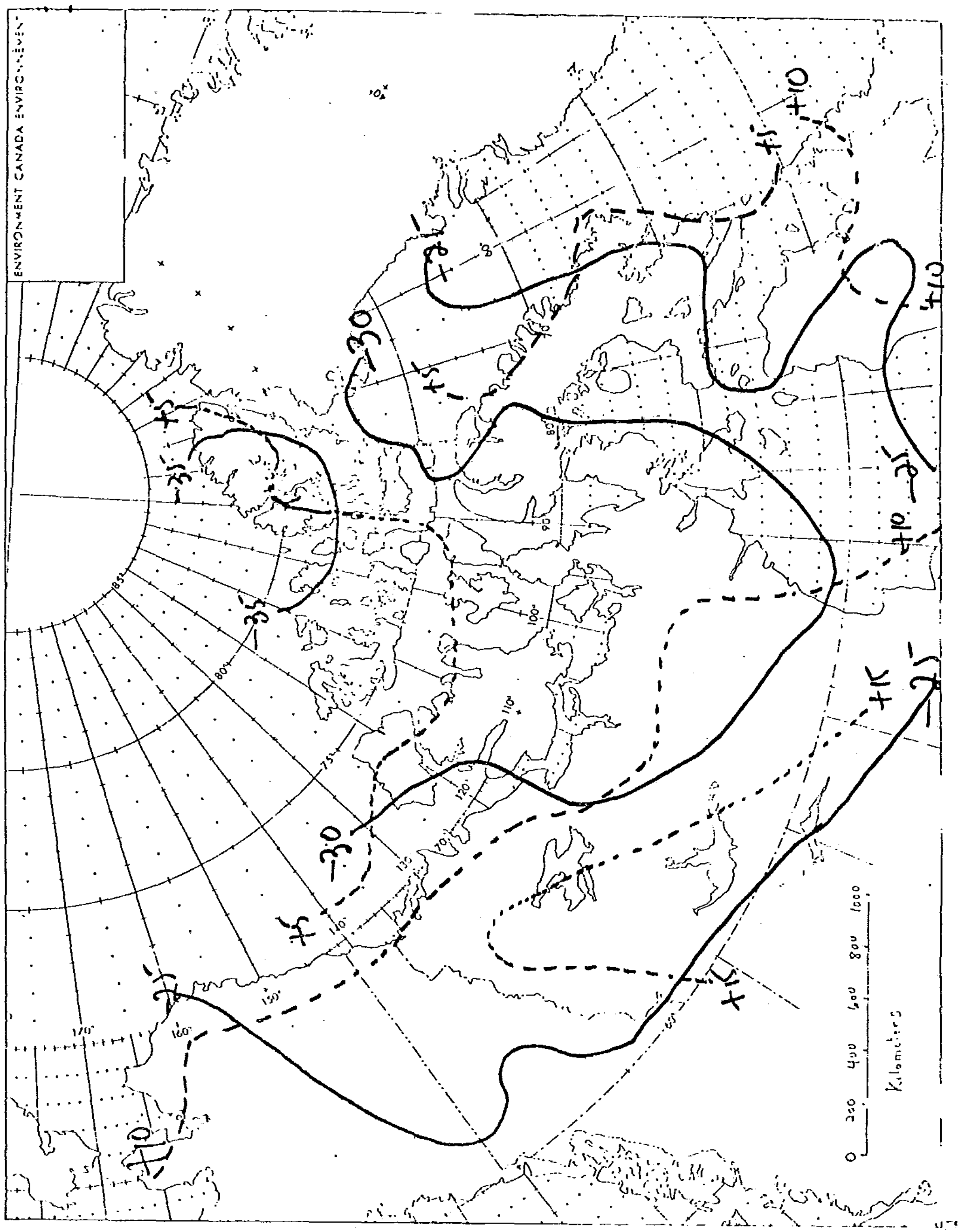
Apart from the cold, the most noteworthy feature of the winter temperature distribution is the much milder weather over Baffin Bay which relates to the cyclone frequency there, the frequent incursions of milder air in winter accompanying these cyclones, and the warmer inflow water flowing northward along the west coast of Greenland into Baffin Bay, and, in the northern part, the open water, or thin ice, in Northwater.

Precipitation is light over the Canadian Arctic Archipelago. The normal annual precipitation (1941-1970) after the Atmospheric Environment Service (cm of water equivalent), is shown in Figure 4-9. The percentage of average annual precipitation falling as snow shown in Figure 4-10 (after Walker and Lake, 1973) is above 50 percent over most of the Archipelago. The distribution of snow and rain at selected stations in the Archipelago is shown in Figure 4-11. While heaviest precipitation occurs in the brief summer, precipitation in smaller amounts accumulates as snow cover during winter months and contributes to the sudden spring runoff.

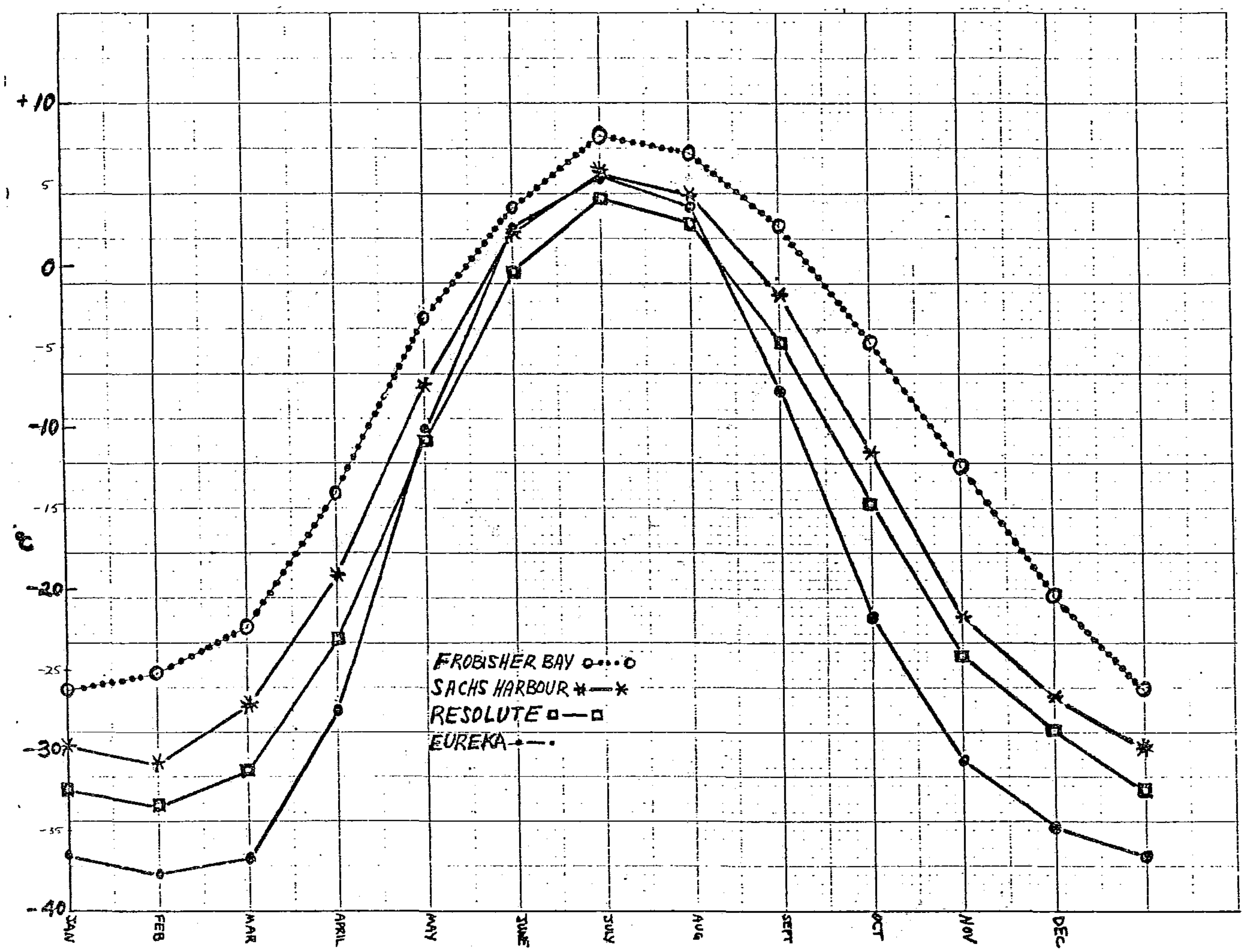
The reliability of precipitation amounts measured at the weather stations in the archipelago is perhaps lower than elsewhere in Canada because of the great difficulty of accurately measuring snowfall which could be underestimated by 30-40% (Walker and Lake, 1973; Hare and Hay, 1971). Allowance for the precipitation increase expected with higher elevations in the mountainous areas of the eastern archipelago could double, in the areas, amounts depicted in Figure 4-9.

The snow cover records over the archipelago have been reviewed by Potter (1965). His median dates of first and last snow cover are shown in Figure 4-12 and median depths of maximum annual snow cover at selected weather stations are shown in Fig. 4-13. The values of depth of snow cover, and snow density hence water-equivalent of the snowcover are subject to sampling errors as discussed by Walker and Lake (1973), Longley (1960) although estimated errors in depth of snow cover are probably less than in measurements of snowfall.

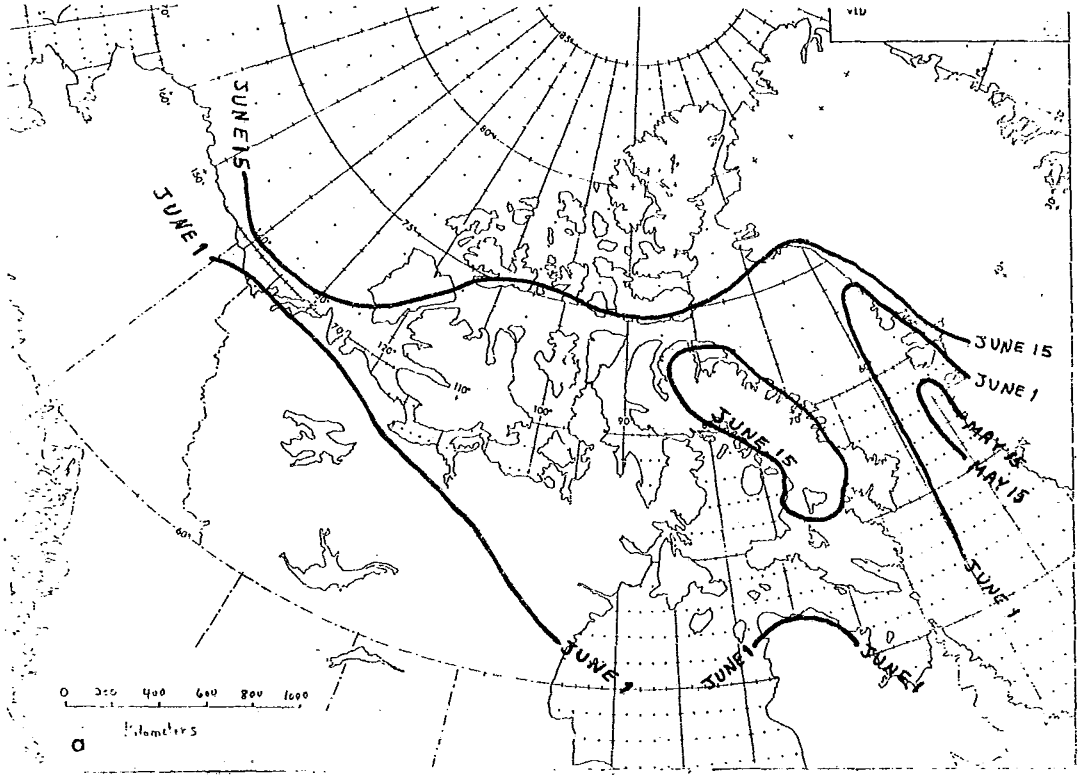
Evaporation is low in the archipelago. Estimates of average annual lake evaporation are given by Ferguson et al (1970). Crude estimates of the evaporating power of the net radiation given by Hare and Hay (1971), agree fairly well with those of Ferguson et al. Both estimates are shown in Figure 4-14. Estimates of evaporation in high latitudes are probably unreliable (Hare and Hay, 1971; Vowinckel and Orvig, 1976). In addition, the smaller scale variability caused by differences in heat budgets over sea and over different types of terrain is always present. Over the sea local studies Gade et al (1974), Huyer and Barber (1970) and others indicate maximum evaporation from an open sea surface in autumn when the water is warm and the air is



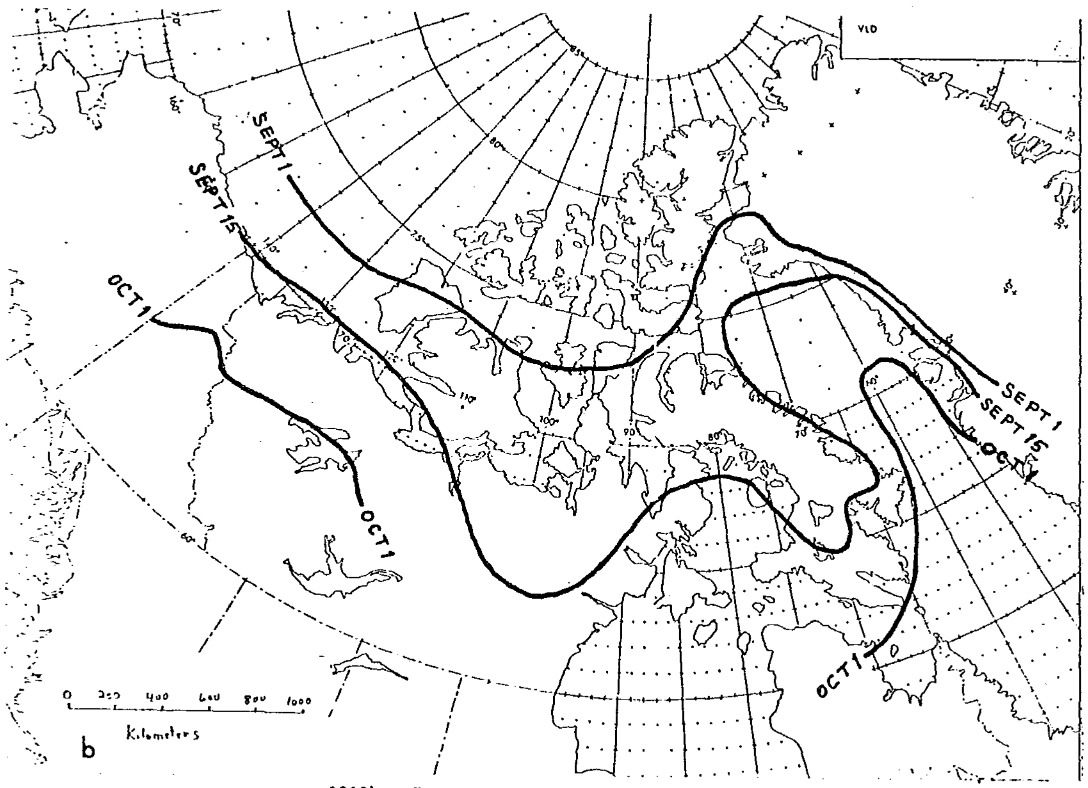
4-5 Mean air temperatures (1941-1970), (°C), January (solid curve) and July, (dashed curve). (data from AES, Canada).



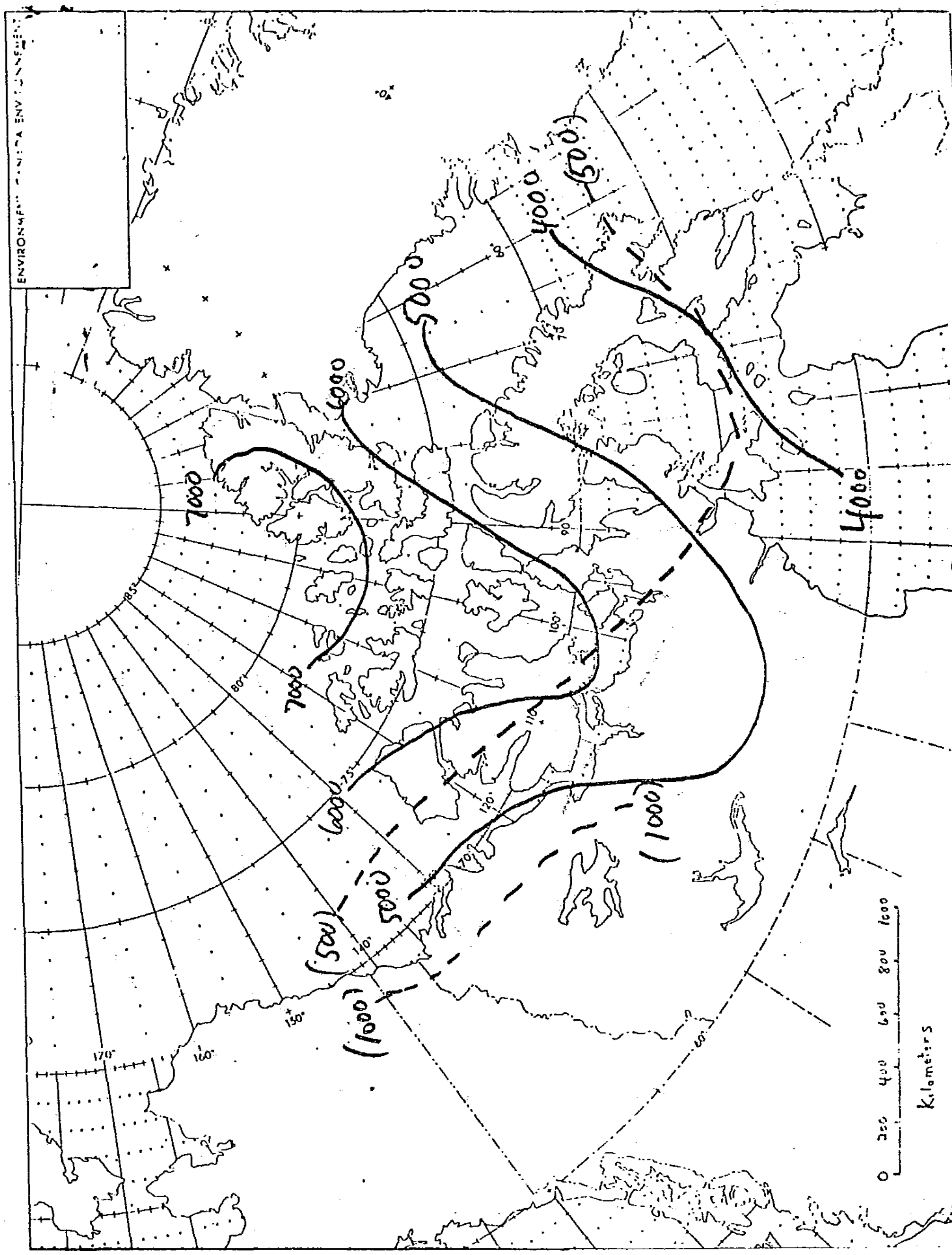
4-6 Normal (1941-1970) monthly mean air temperatures (°C) for Frobisher Bay, Sachs Harbour, Resolute and Eureka. (data from AES, Canada).



4-7 (a) Date on which mean daily temperature rises above 0°C (1951-1960). (after Anon., 1970).

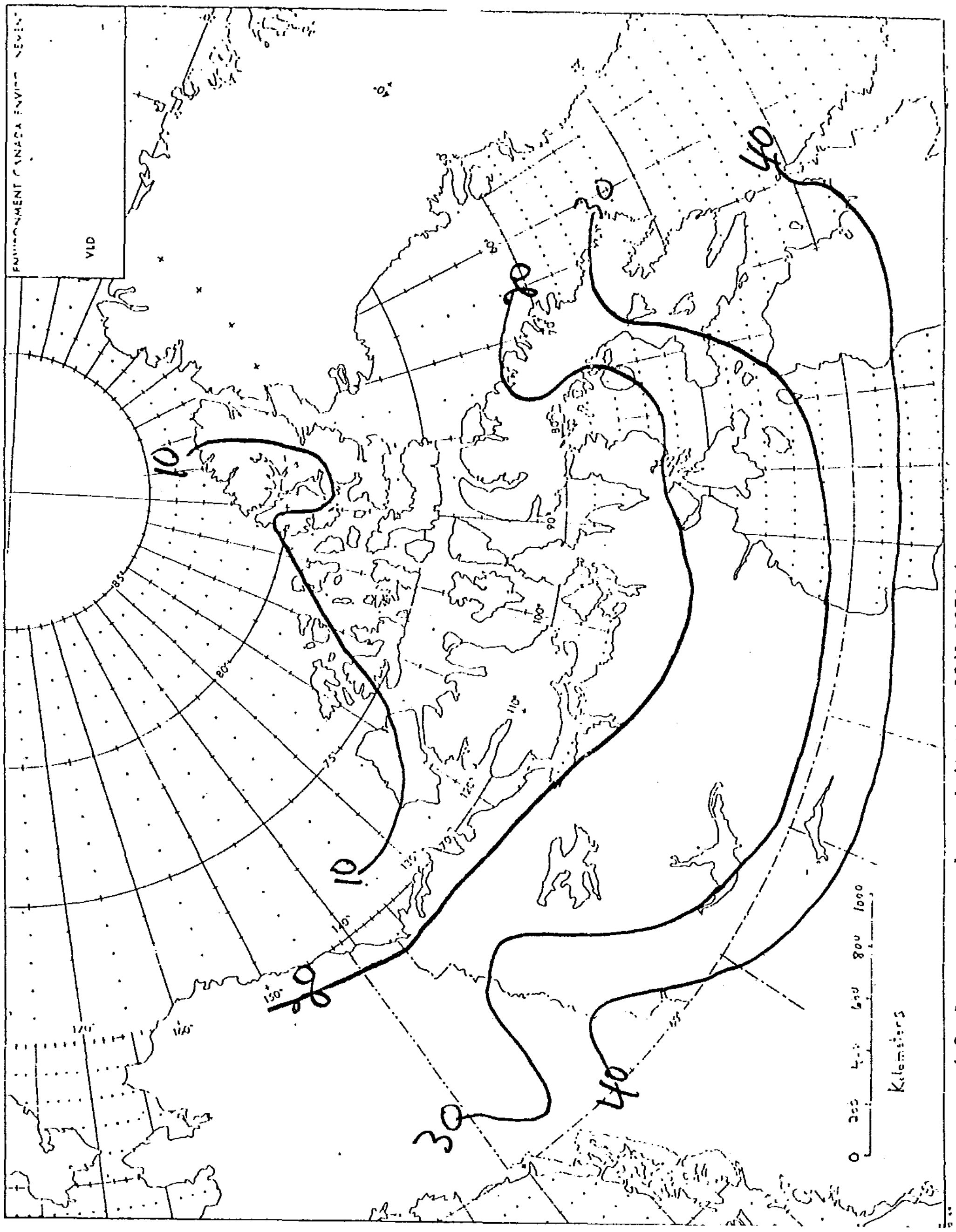


4-7 1960). (b) Date on which mean daily temperature falls below 0°C. (after Anon., 1970).



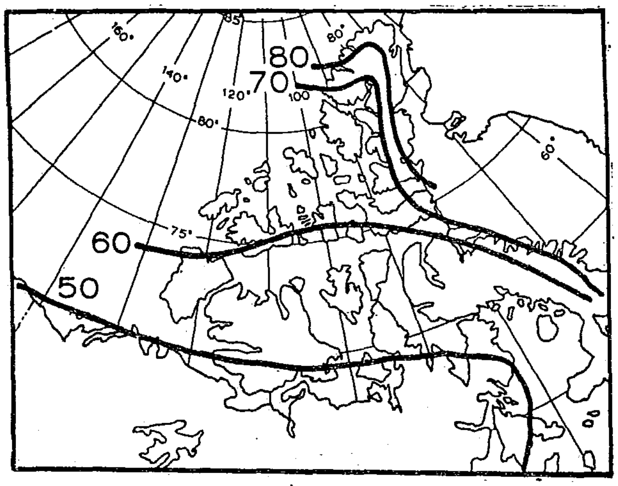
4-8 Number of freezing degree days (solid curve) and thawing degree days (dashed curve), (°C). (data from Anon., 1967).

4-8 Number of freezing degree days (solid curve) and thawing degree days (dashed curve), (°C). (data from Anon., 1967).

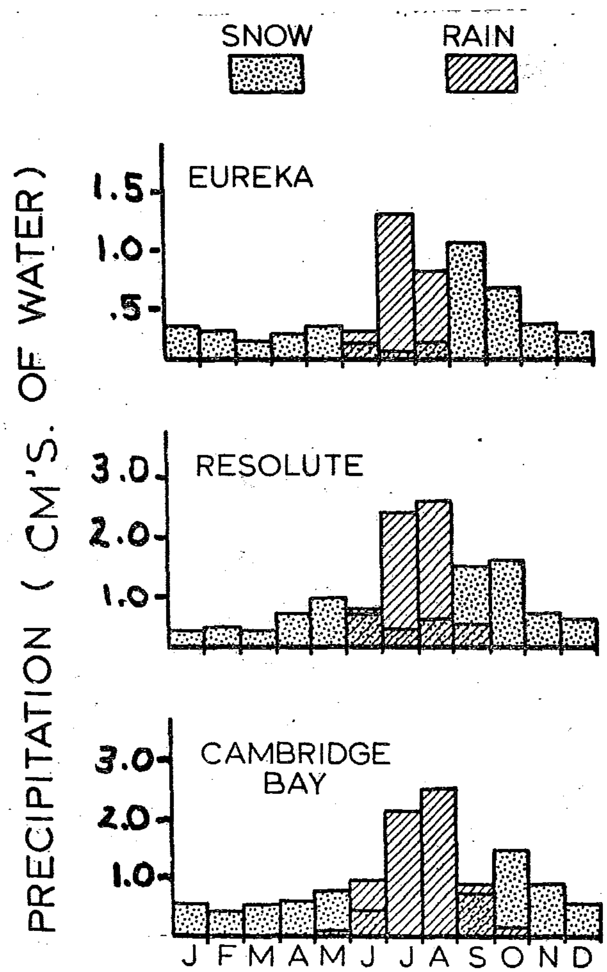


4-9 Average annual precipitation 1941-1970 (cm water equivalent). (data from AES, Canada).

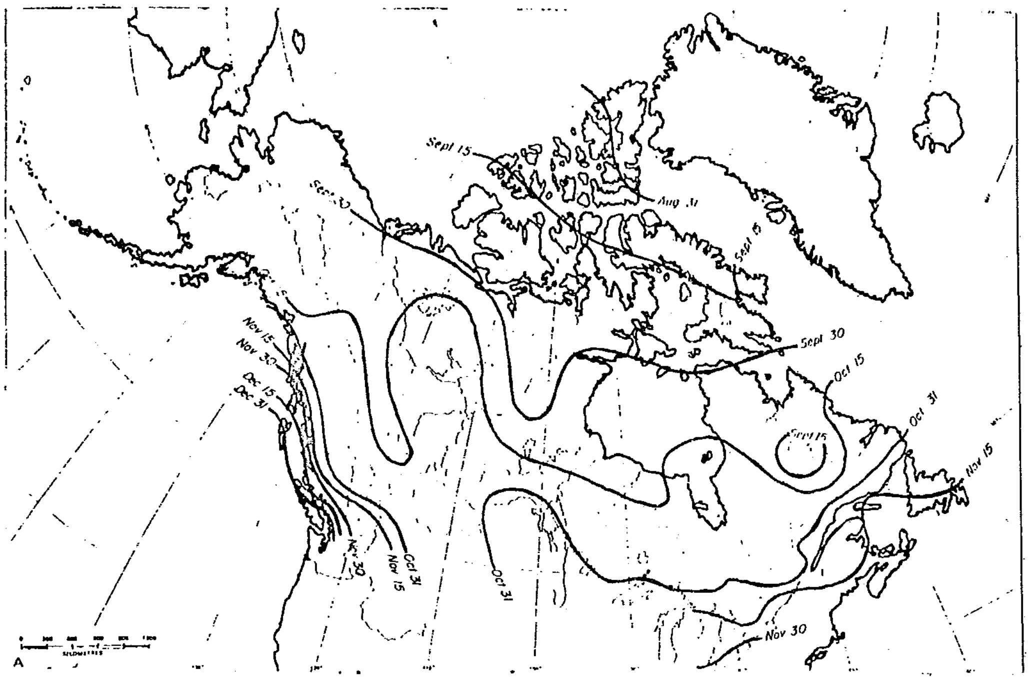
4-9 Average annual precipitation 1941-1970 (cm water equivalent). (data from AES, Canada).



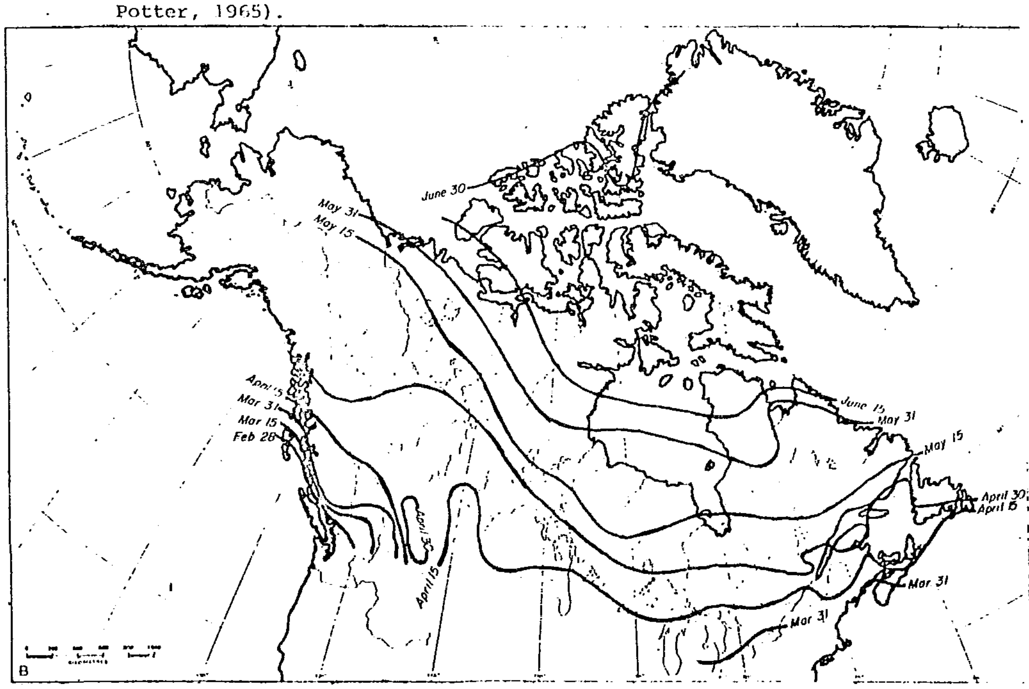
4-10 Percentage of normal (1941-1970) annual precipitation falling as snow over the Canadian Arctic Archipelago. (data from AES, Canada).

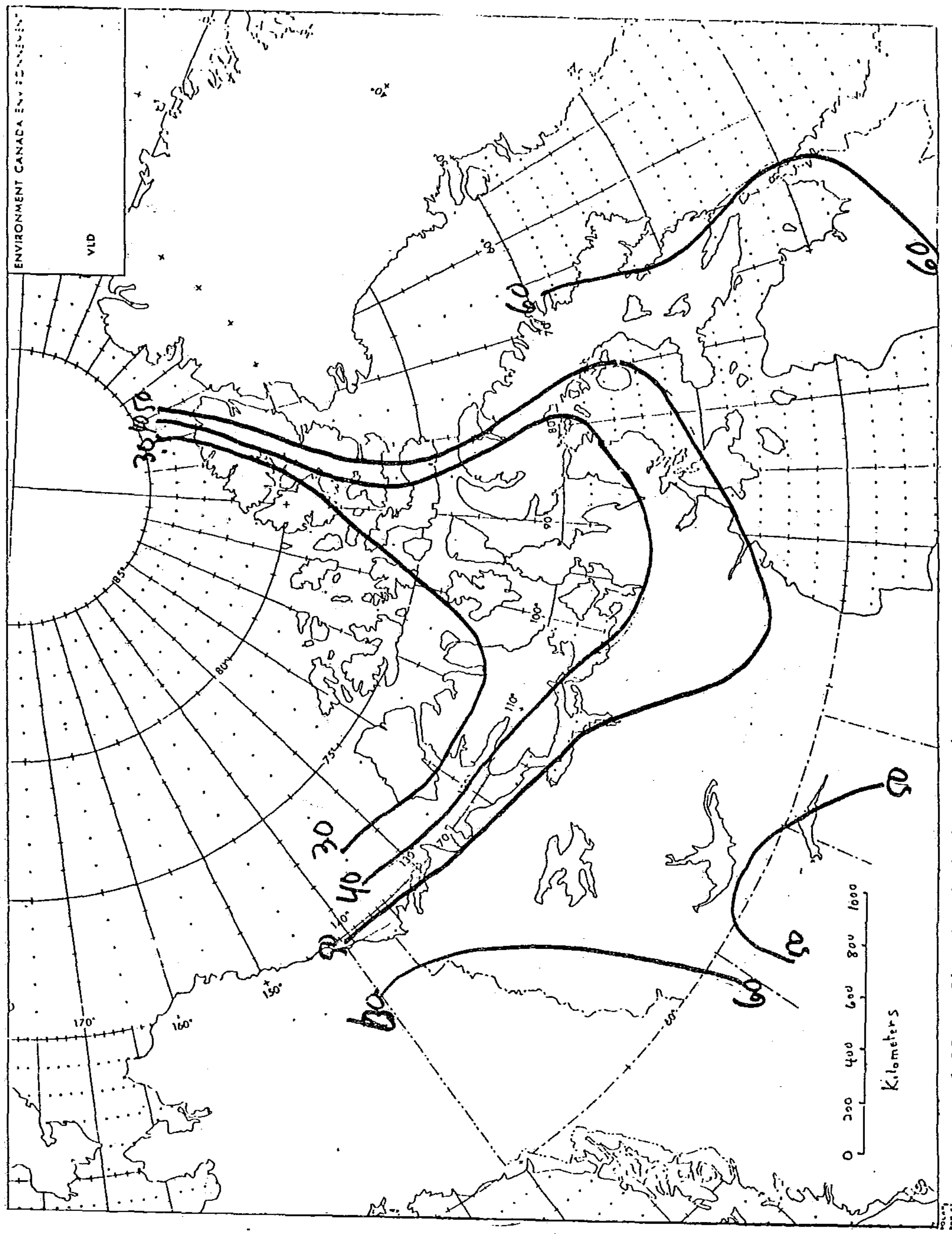


4-11 Monthly normals (1941-1970) of rainfall and snowfall water equivalent at selected stations in the Canadian Arctic Archipelago (cm of water). (data from AES, Canada).



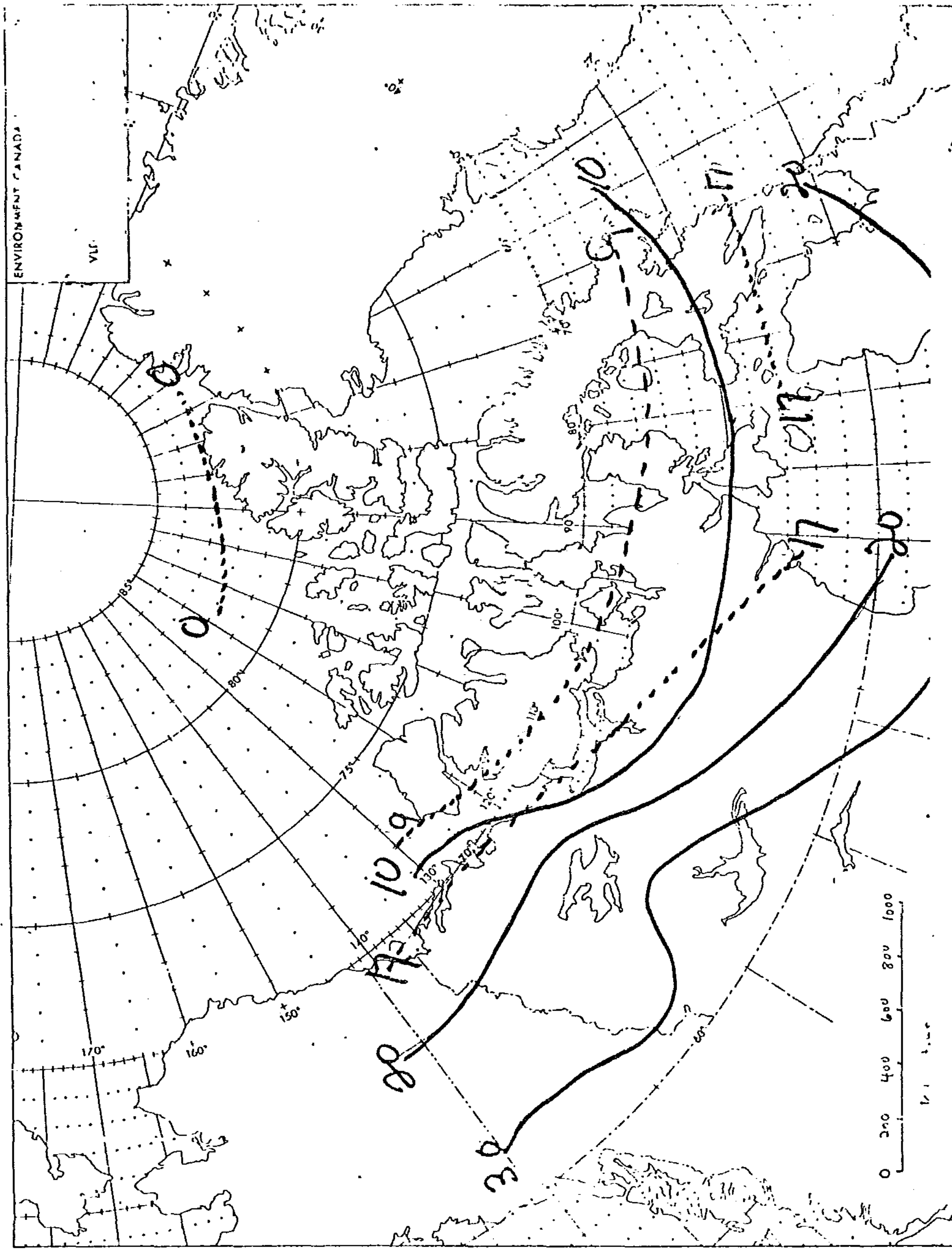
4-12 Median dates of (a) first and (b) last snow cover greater than 2.5 cm. (after Potter, 1965).





4-13 Median depth (cm) of maximum snow covers over 20 winters. (after Potter, 1965).

4-13 Median depth (cm) of maximum snow covers over 20 winters. (after Potter, 1965).



4-14 Mean annual evaporation from small lakes (cm) (solid curve) after Ferguson et al, (1970) and evaporating power (cm) of net radiation budget (dashed curve) after Hare and Hay, (1974) for the high Arctic only.

4-14 Mean annual evaporation from small lakes (cm) (solid curve) after Ferguson et al, (1970) and evaporating power (cm) of net radiation budget (dashed curve) after Hare and Hay, (1974) for the high Arctic only.

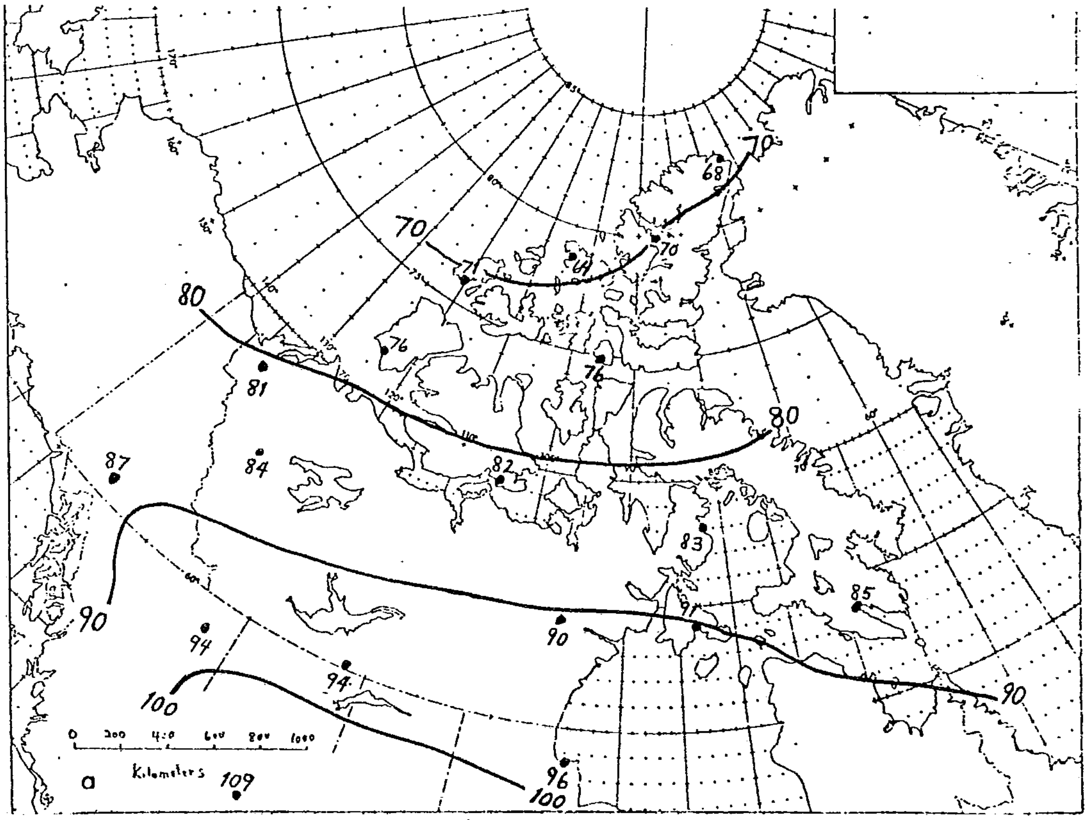
cold, with a secondary minimum in spring. The evaporation from narrow open water leads at high latitudes has been treated by Badgley (1966) who estimated evaporation at rates up to 6 cm per month in mid winter. However, such evaporation rates are accompanied by very large losses of sensible heat from the surface of the water in the lead so that usually freezing occurs quickly. Evaporation from larger open water areas, which are able to remain open for an appreciable time in cold weather has been discussed by Danielson (1969), Walmsley (1966), Vowickel and Orvig (1973) Muench (1971), Sadler (1976) and others.

Radiation

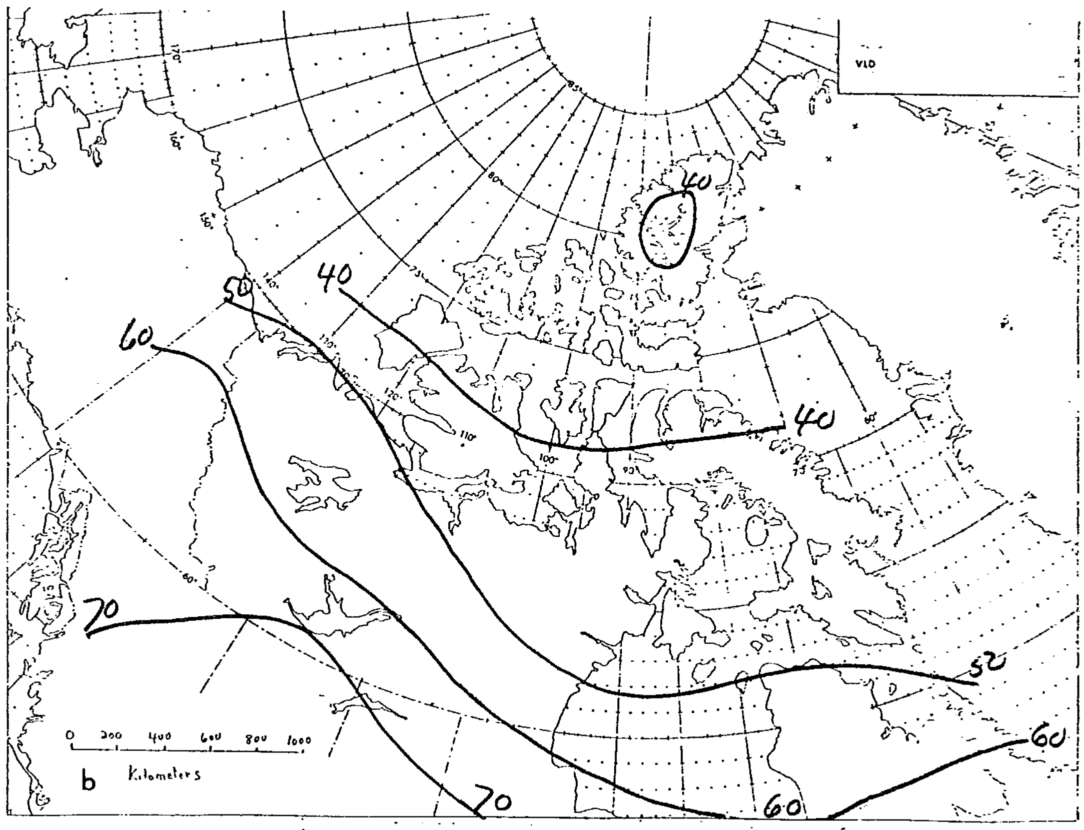
The radiation components of heat balance equations mentioned in Appendix A and Chapter 10 are very important in Arctic regions. Therefore radiation measurements have been made since 1960-65 at the meteorological stations shown in Figure 4-1. Measurements at these stations of global solar radiation (downward direct and diffuse solar radiation) received on a horizontal surface agree with calculations by Titus and Truhlar (1969). Mean annual values of global solar radiation are shown in Figure 4-15(a). During those months in which the surface in the Canadian Arctic Archipelago is covered by snow (Figure 4-12) surface albedos are high, as shown in Table 4-2. Therefore most of the global solar radiation is reflected rather than absorbed. Estimates of solar radiation absorbed annually are shown in Figure 4-15(b). Albedos measured at weather stations, in winter over snow by and large a trifle dirtier than average, in summer over bare ground on the shores of an ice-choked sea, may not be representative of large areas. Attempts to obtain albedos representative of large areas have been made as noted above by Larsson and Orvig (1961) (Figure 4-16) and by Hay (1970) in an unpublished thesis at London University. Hay's albedo estimates for Resolute Bay are shown in Table 4-3. Radiation quantities recorded at weather stations are affected by values of surface albedo at the measuring site (even global solar radiation to some extent [see Hay, 1976]). Estimates are given in Figure 4-17 of average absorption of solar energy in a summer in the Eureka area, by land, sea or ice surface. The differences between the top of the histogram curve representing global solar radiation and the amount absorbed on different surfaces are large.

Infrared terrestrial radiation fluxes are not routinely measured. Upward terrestrial radiation can be estimated from surface temperatures and other surface characteristics; downward terrestrial radiation can be measured, estimated from radiosonde measurements of atmospheric quantities, or very crudely estimated from surface observations and cloudiness (see Sellers, 1965, p. 52). Few estimates of fields of terrestrial radiation fluxes over Canada have been published. In Figure 4-18 however, are Hay's (1970) estimated upward and downward terrestrial radiation fluxes for January and July averaged over 1957-64. The annual variation related to absolute (Kelvin) temperatures varies by less than a factor of two over the year.

The net radiation, or radiation balance, the sum of downward and reflected solar radiation, downward and reflected terrestrial radiation and upward terrestrial radiation is measured at weather stations noted in Figure



4-15 (a) Annual global solar radiation (1969-1975) (kcal cm^{-2}) (data from AES, Canada).

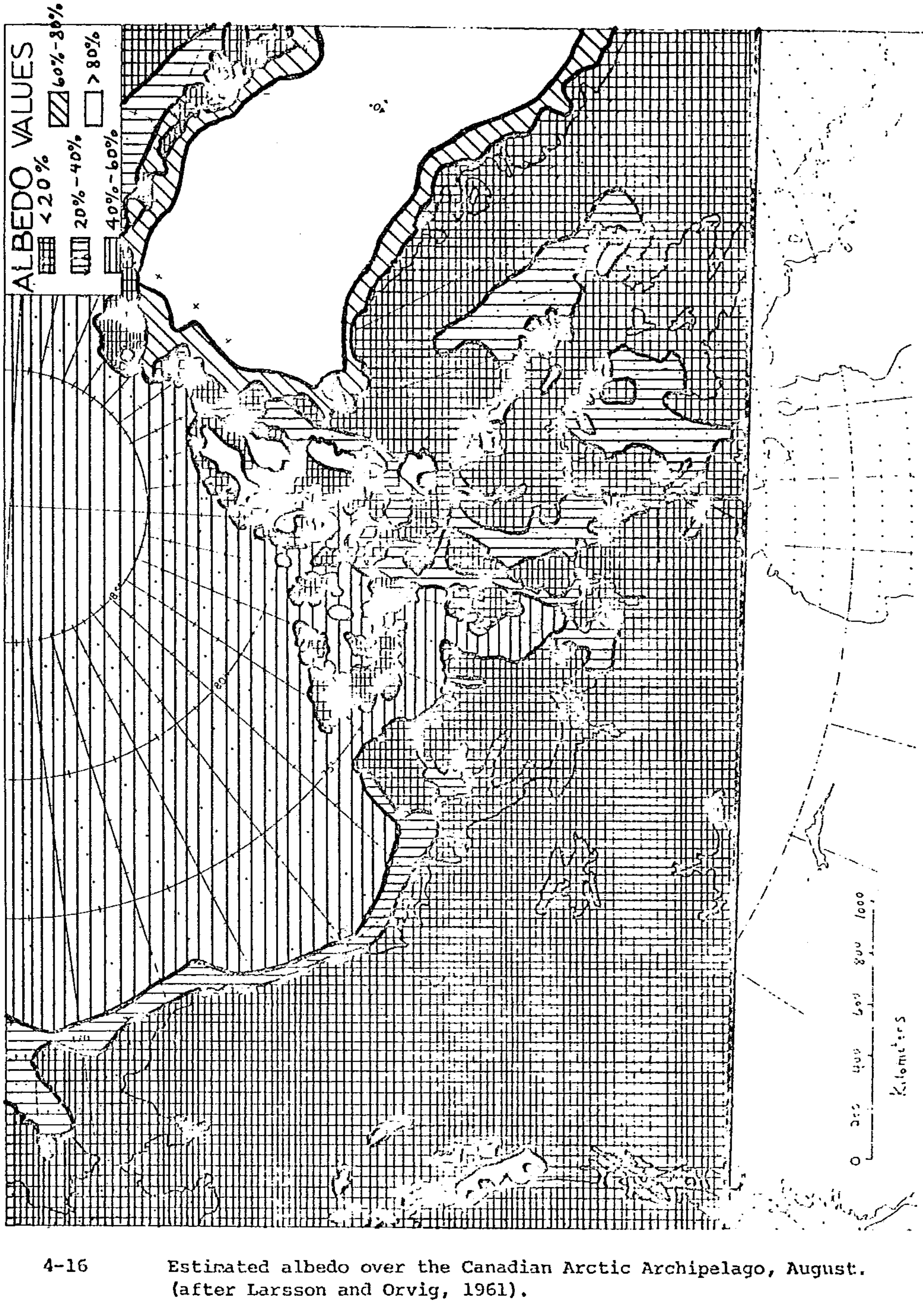


4-15 (b) Estimated absorbed solar radiation (1957-1964) (after Hay, 1970). (kcal cm^{-2}).

TABLE 4-2

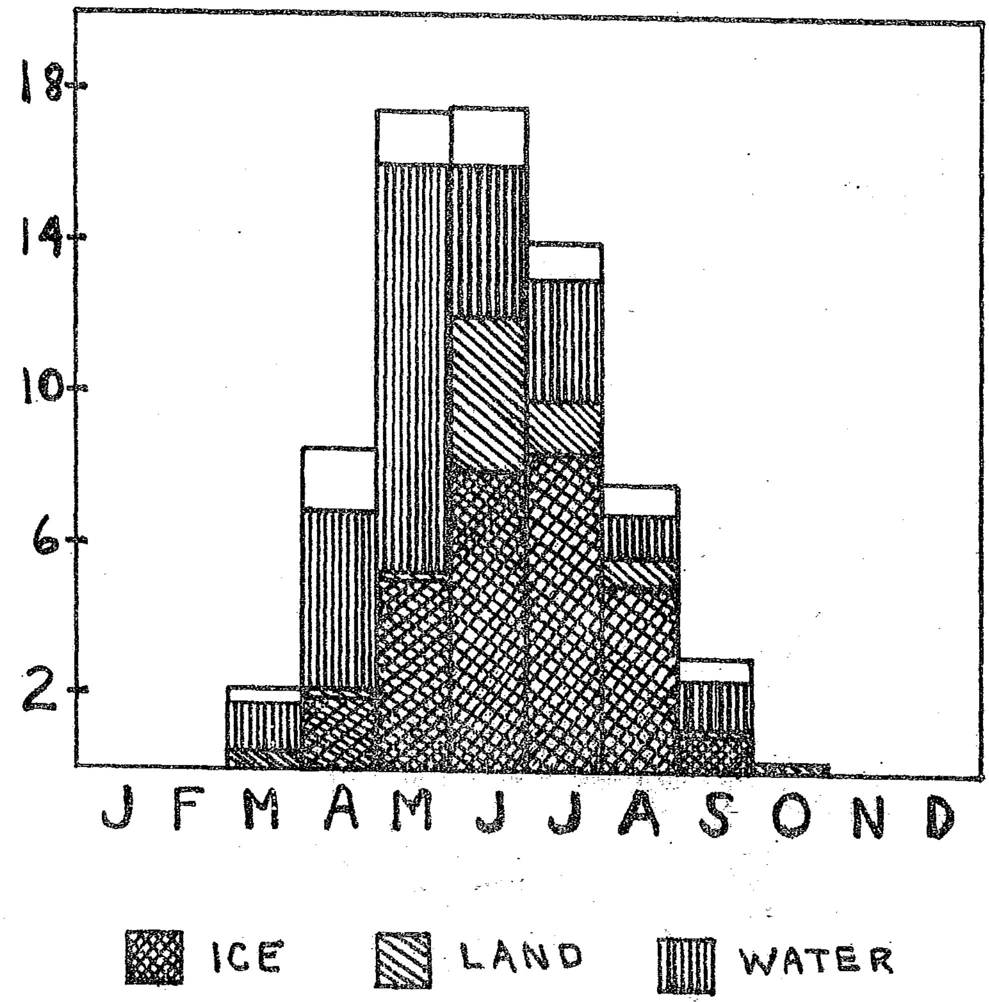
Structure	Water Content and Colour	Albedo %		
		Average	Max.	Min.
Freshly fallen snow	dry bright-white clean	88	98	72
Freshly fallen snow	wet bright-white	80	85	75
Freshly drifted snow	dry clean loosely packed	85	96	70
Freshly drifted snow	moist grey-white	77	81	59
Snow, fallen or drifted 2-5 days ago	dry clean	80	86	75
Snow, fallen or drifted 2-5 days ago	moist grey-white	75	80	56
Dense snow	dry clean	77	80	66
Dense snow	wet grey-white	70	75	61
Snow and ice	dry grey-white	65	70	58
Melting ice	wet grey	60	70	40
Melting ice	moist dirty grey	55	65	36
Snow, saturated with water (snow during intense thawing)	light green	35	--	28
Melt puddles in first period of thawing	light blue water	27	36	24
Melt puddles, 30 100 cm deep	green water	20	26	13
Melt puddles, 30 100 cm deep	blue water	22	28	18
Melt puddles covered with ice	smooth grey-green ice	25	30	18
Melt puddles covered with ice	smooth ice, covered with icy white hoar frost	33	37	21
Water surfaces	Winter 60°N	21	--	--
Water surfaces	Summer 60°N	7	--	--
Soil	dark	10	5	15
Soil	moist grey	15	10	20
Soil	dry sand	35	25	45
Cloud	overcast cumuli-form	80	70	90
Cloud	stratus	70	59	84
Cloud	Altostratus	49	39	54
Cloud	Cirrostratus	47	44	50

Typical values of albedos over various surfaces (after Vowinckel and Orvig [1970] and Sellers [1965]).

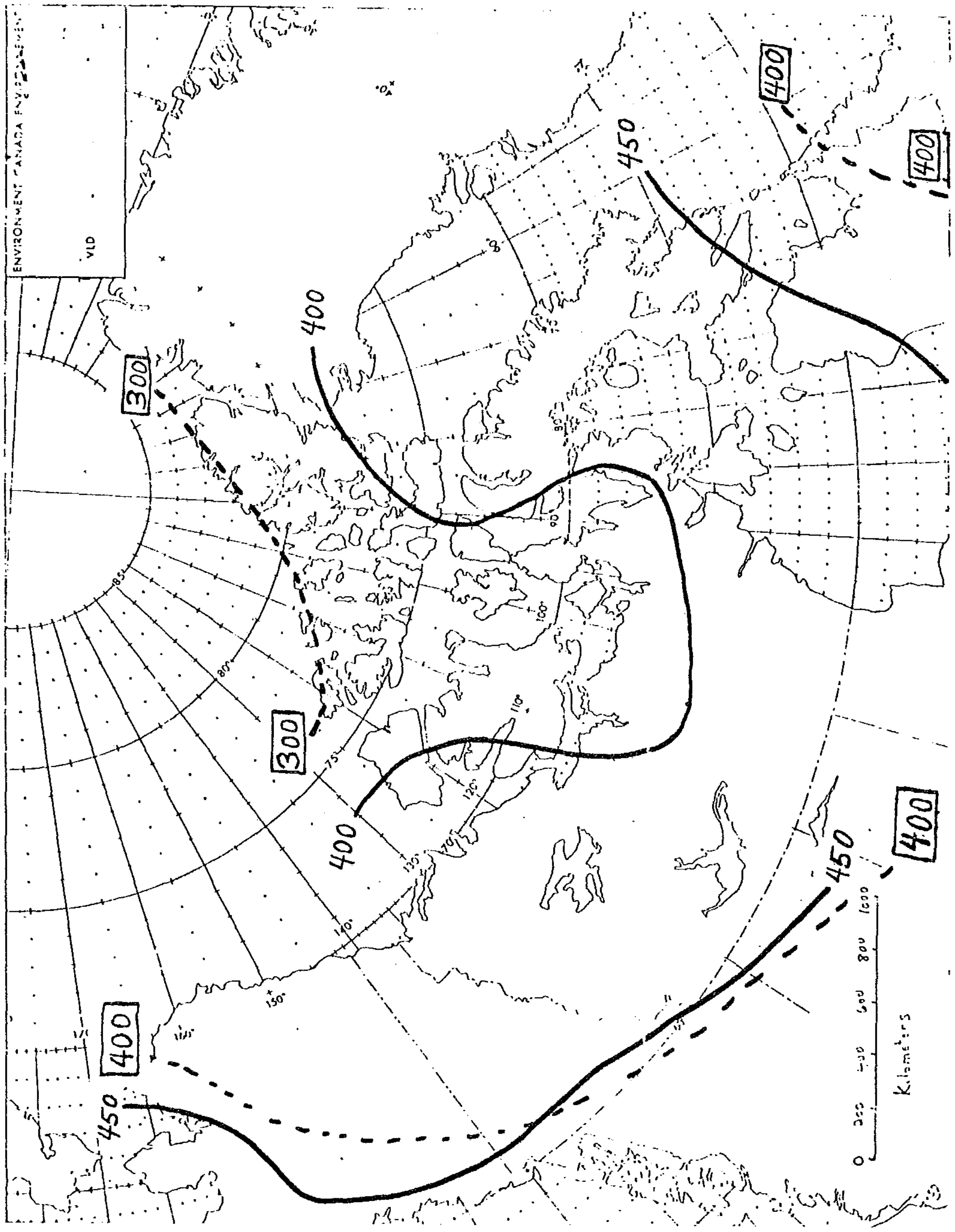


4-16 Estimated albedo over the Canadian Arctic Archipelago, August. (after Larsson and Orvig, 1961).

4-16 Estimated albedo over the Canadian Arctic Archipelago, August. (after Larsson and Orvig, 1961).

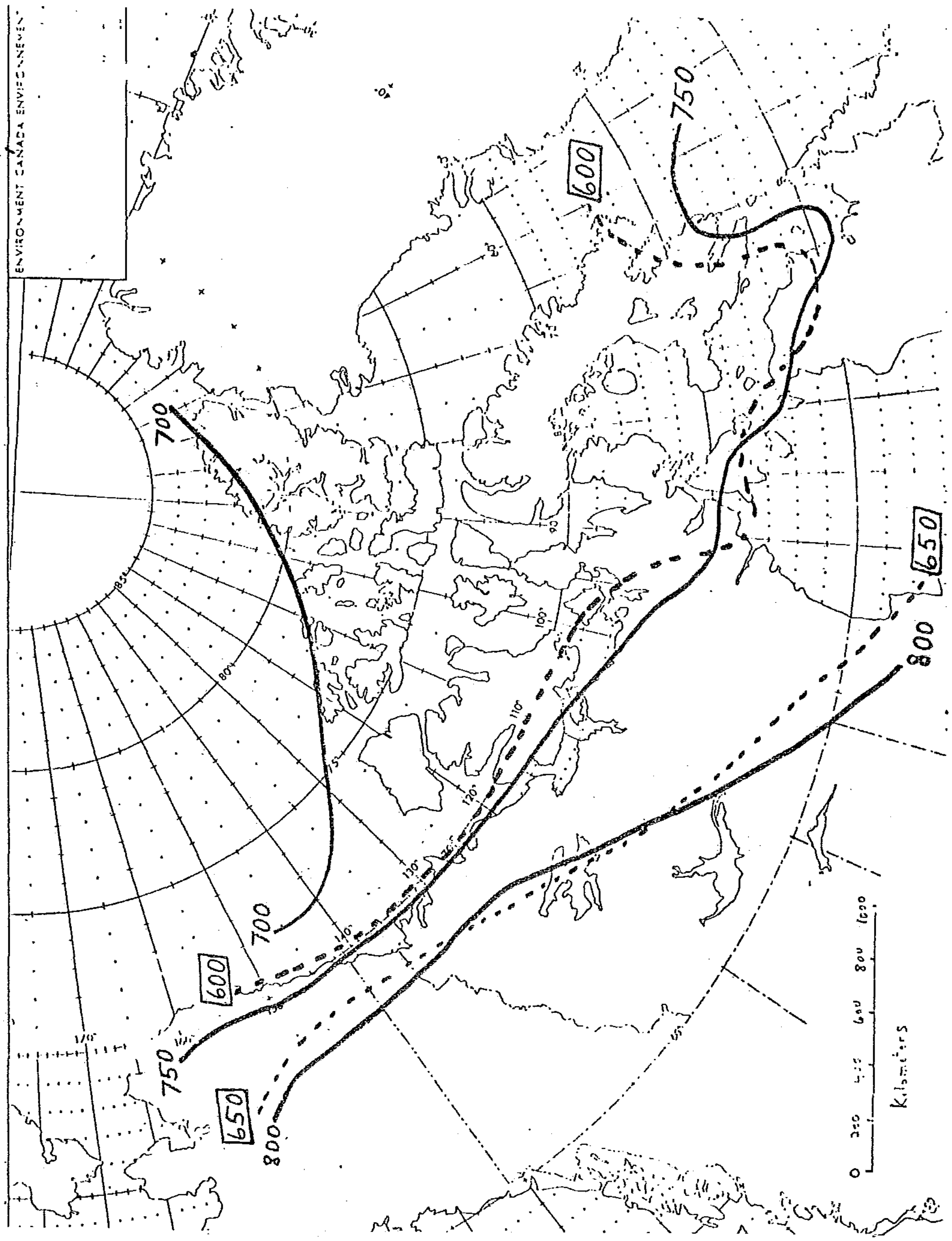


4-17 Solar radiation absorbed each month near Eureka weather station over (a) Ice surface (b) Land surface (c) Open water surface, (kcal cm^{-2}month). Albedos as in Table 4-2.



4-18 Estimated infrared radiation 1957-1964. Solid lines are radiation emitted by earth's surface (cal $cm^{-2}day^{-1}$) dashed lines - infrared radiation received at the surface from the atmosphere (cal $cm^{-2}day^{-1}$).
(a) January, (after Hay, 1970).

4-18 Estimated infrared radiation 1957-1964. Solid lines are radiation emitted by earth's surface (cal $cm^{-2}day^{-1}$) dashed lines - infrared radiation received at the surface from the atmosphere (cal $cm^{-2}day^{-1}$).
(a) January, (after Hay, 1970).



4-18 Estimated infrared radiation 1957-1964. Solid lines are radiation emitted by earth's surface (cal $cm^{-2}day^{-1}$) dashed lines - infrared radiation received at the surface from the atmosphere (cal $cm^{-2}day^{-1}$). (b) July. (after Hay, 1970).

4-18 Estimated infrared radiation 1957-1964. Solid lines are radiation emitted by earth's surface (cal $cm^{-2}day^{-1}$) dashed lines - infrared radiation received at the surface from the atmosphere (cal $cm^{-2}day^{-1}$). (b) July. (after Hay, 1970).

TABLE 4-3

	J	F	M	A	M	J	J	A	S	O	N	D
A	80	80	80	80	70	70	40	10	50	80	80	80
B	--	--	20	10	10	10	10	10	10	20	--	--
C	80	80	80	80	70	50	50	50	50	70	80	80
D	--	--	--	--	30	30	30	30	--	--	--	--
E	80	--	--	70+	--	--	25	--	--	60	--	--
F	--	80	80	80	70	50	30	20	40	60	--	--
G	--	76	77	75	70	44	25	26	51	75	--	--

Estimated values of monthly mean albedos over various surfaces at about 75°N in the Archipelago. (A) Land surface albedo (Larsson and Orvig 1961), (B) Water surface albedo (Larsson and Orvig 1961), (C) Ice surface albedo (Larsson and Orvig 1961), (D) Melting ice and water surface (Larsson and Orvig 1961), (E) Generalized albedos (Hay 1970), (F) Resolute generalized albedo (Hay 1970), (G) Albedo measured at Resolute weather station 1969-1975 (AES Canada).

4-1. In Table 4-4 values of daily global solar and net radiation, averaged over the years 1969-1975 measured at Canadian Arctic weather stations are shown (data from AES). In winter of course the net radiation is the balance between large (about 300-600 cal $cm^{-2}day^{-1}$) values of upward and downward infrared radiation. In summer the net downward solar radiation is added. This net solar radiation depends on the percentage reflected upward solar radiation. The fraction of downward solar radiation reflected upwards is so dependent on surface characteristics that the pattern of net radiation values will be far more closely related to patterns of the surface characteristics than is global solar radiation. Net radiation values measured at land stations will in summer almost certain not represent conditions over nearby ice choked water areas, and may indeed not well represent an average value of over any extensive area.

Vowinckel and Orvig (1967) for the Arctic Basin, and Hay (1970), for Canada, have attempted to estimate regionally representative values of terms in the radiation balance. In Figure 4-19 (Hay, 1970) estimates of the annual total of net radiation based on data 1957-1964 is shown. Values of net radiation measured by Canadian Arctic weather stations are over 1969-1975 and are plotted on the Figure. Reason for the wide discrepancies between observed annual totals, and generalized estimates is to be found in part in the generalized albedos used. However many aspects of the surface heat budgets in the archipelago including radiation components, cry out to be put on a firmer footing.

TABLE 4-4(a)

Global Solar Radiation - Arctic Stations

(1969 - 1975)

RF #1

STATION	J	F	M	A	M	J	J	A	S	O	N	D	YEAR
	Cal $cm^{-2} day^{-1}$												kcal cm^{-2}
Alert	0	0	47	280	533	587	434	243	85	5	0	0	68
Baker Lake	18	77	242	432	539	514	475	325	176	85	30	8	90
Cambridge Bay	2	45	177	402	524	562	458	301	150	68	9	0	82
Churchill	56	139	296	433	472	506	469	362	221	104	51	37	96
Coral Harbour	18	77	236	430	552	560	444	332	189	91	30	8	91
Edmonton	88	169	304	417	492	514	524	444	290	185	93	64	109
Eureka	0	1	67	284	561	582	450	245	101	10	0	0	70
Fort Nelson	41	115	232	382	484	479	488	383	248	124	52	26	94
Fort Smith	39	111	248	406	491	513	492	387	232	106	40	24	94
Frobisher	21	86	227	418	542	503	386	309	182	76	26	10	85
Hall Beach	3	47	185	397	533	602	469	324	145	61	10	0	83
Inuvik	4	47	188	381	497	536	464	283	162	61	12	0	81
Isachsen	0	2	76	287	549	572	382	220	99	12	0	0	67
Mould Bay	0	8	102	333	550	545	414	234	114	23	0	0	71
Norman Wells	12	64	203	364	480	540	471	333	198	71	20	4	84
Resolute	0	13	127	355	562	591	413	267	128	31	1	0	76
Sachs Harbour	0	27	160	362	513	512	441	264	125	41	3	0	76
Whitehorse	32	95	226	359	468	507	450	338	213	100	40	17	87

Monthly means (1969-1975) of daily global solar radiation (cal cm^{-2}) measured at Arctic weather stations (data after AES Canada), and annual total (kcal cm^{-2})

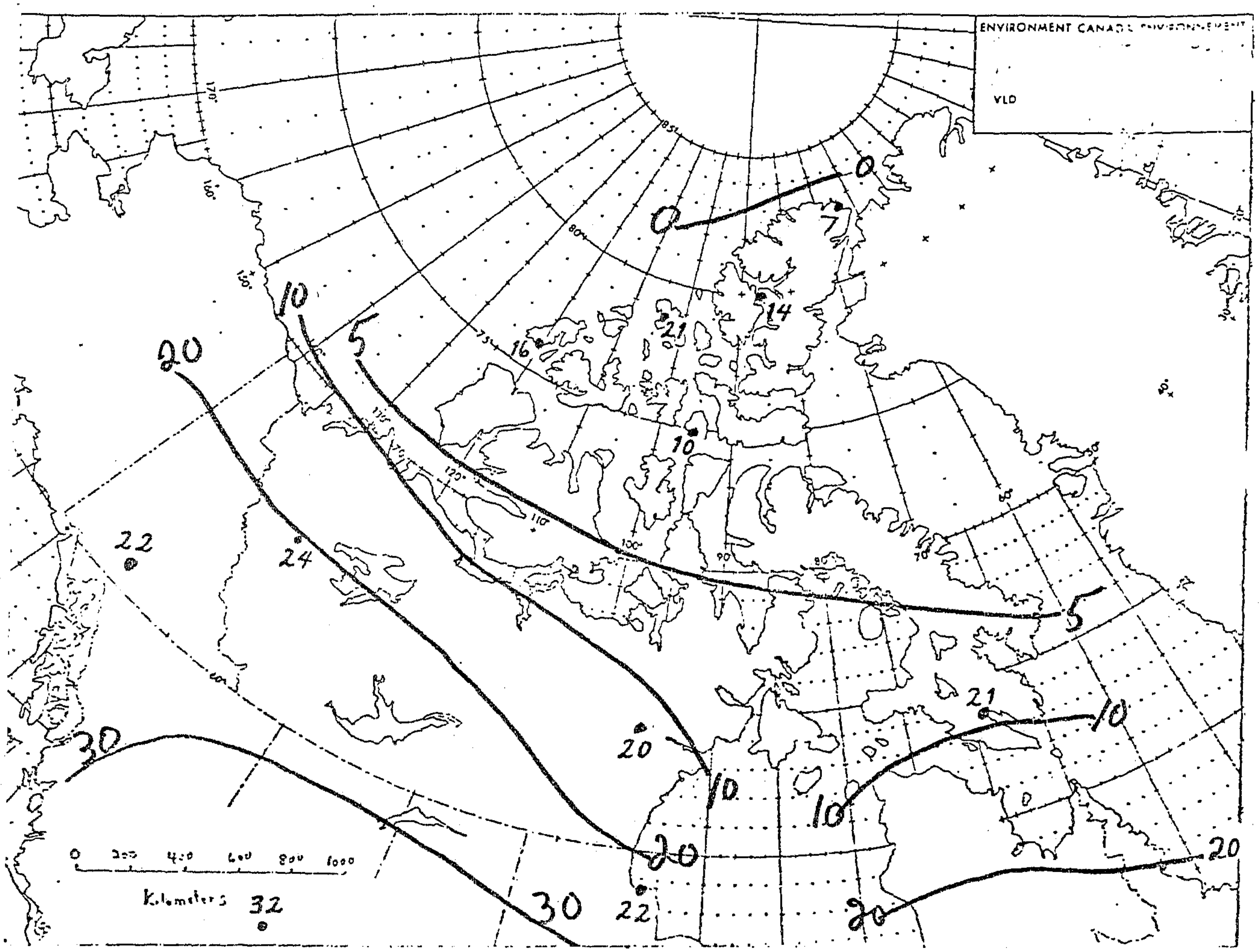
TABLE 4-4(b)

Net Radiation - Arctic Stations

RF #4 (1969 - 1975)

STATION	J	F	M	A	M	J	J	A	S	O	N	D	YEAR
	Cal $cm^{-2}day^{-1}$												kcal cm^{-2}
Alert	-31	-32	-26	- 31	6	131	257	96	-33	-42	-41	-38	7
Baker Lake	-41	-42	-43	- 16	153	262	224	142	59	-25	-37	-37	20
Churchill	-62	-50	-30	36	192	241	208	165	74	1	-33	-49	22
Edmonton	-46	-38	11	152	227	248	251	200	100	16	-41	-45	32
Eureka	-32	-25	-39	- 17	95	306	233	109	-21	-39	-40	-37	14
Isachsen	-27	-29	-27	2	143	320	233	135	7	-23	-34	-30	21
Mould Bay	-29	-29	-17	0	45	246	245	127	14	-18	-34	-30	16
Resolute	-40	-44	-47	- 20	38	179	221	127	10	-35	-38	-40	10
Whitehorse	-26	-36	-36	71	211	216	194	136	62	-16	-32	-29	22
Frobisher Bay	-60	-61	-29	34	203	290	218	152	66	-17	-53	-54	21
Norman Wells	-35	-25	-25	42	222	249	211	147	60	-14	-24	-20	24

Monthly means (1969-1975) of daily net radiation (cal cm^{-2}) measured at Atctic weather stations (data after AES, Canada), and annual total (kcal cm^{-2}).



4-19 Annual net radiation (1957 - 1964) using generalized albedo estimates (after Hare and Hay, 1974). Annual net radiation (1969-1971) measured by Canadian Arctic weather stations is plotted where available, (kcal cm^{-2}).

4-19 Annual net radiation (1957 - 1964) using generalized albedo estimates (after Hare and Hay, 1974). Annual net radiation (1969-1971) measured by Canadian Arctic weather stations is plotted where available, (kcal cm^{-2}).

Other Weather Elements

Many other weather elements, such as fog, blowing snow and other visibility restrictions, and so on certainly have large effects upon operations in the Arctic, but their direct effect upon the oceans is perhaps not so great. In any event they will not be mentioned here but along with all other weather elements should be treated fully in the aforementioned AES volumes upon climate in the archipelago.

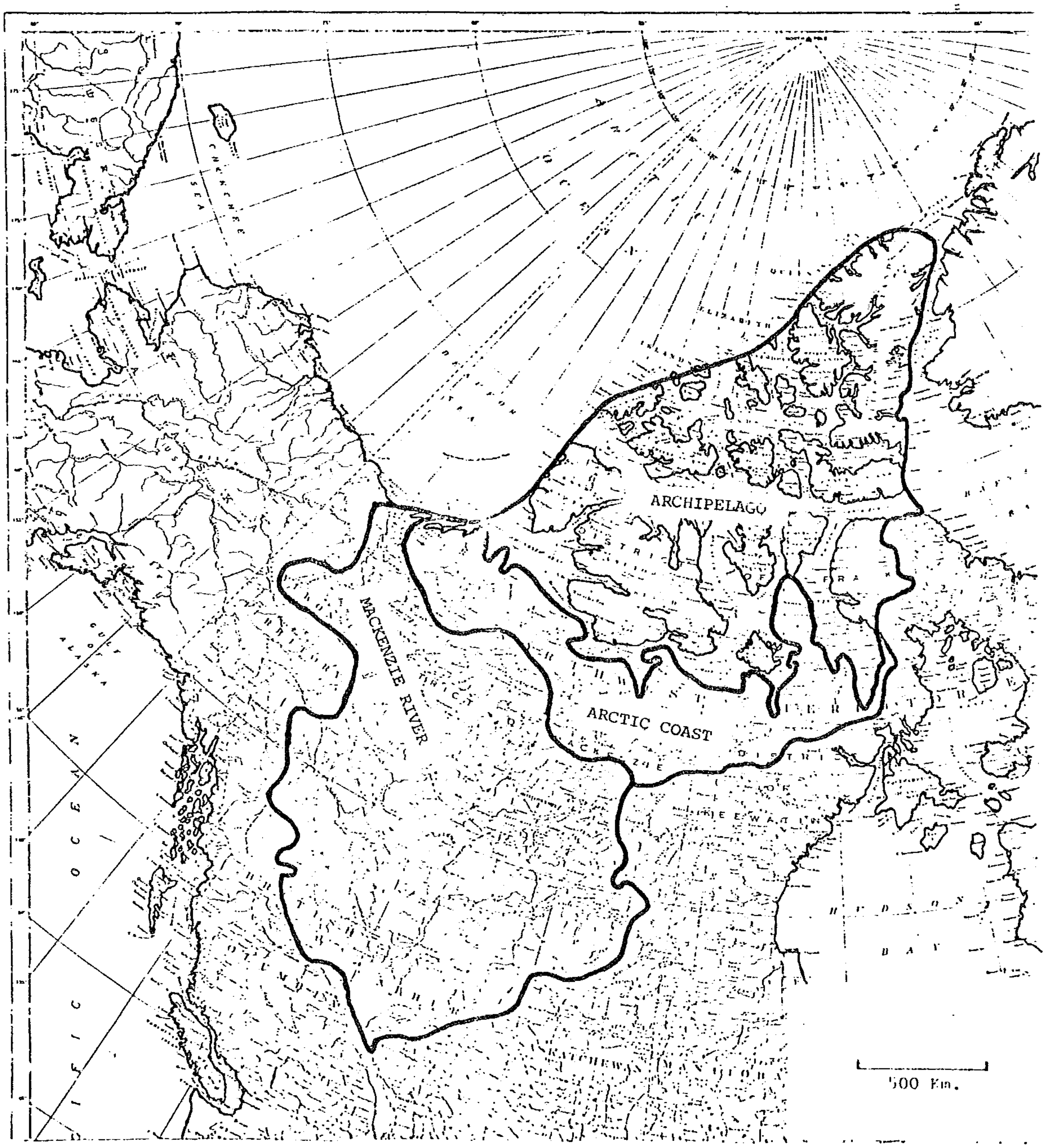
FRESHWATER INPUT

The freshwater input to the central Arctic Ocean has been mentioned above. Freshwater of the surface water layers reduces vertical mixing and helps maintain the sea ice cover of the central oceanic basin. The same situation is present in the channels of the Canadian Arctic Archipelago. The sources of freshwater input to these channels include advection or mixing of relatively fresh surface layers from the Arctic Ocean, local runoff of freshwater from land, and freshwater from melting sea ice (although the growing sea ice in autumn and winter has earlier removed the same amount of freshwater). Estimates of the first process have been made by Huyer and Barber (1970) and Barber and Huyer (1971). The last process may be less important dynamically in that, very approximately, ice forms (and later melts) at the same time and at the same rate over large areas. The local runoff, although small in volume because of the relatively scanty precipitation over the Canadian Arctic Archipelago, may be dynamically important in some areas at least. Differing land areas drain into water bodies of various sizes so freshwater layers of different thicknesses may be formed.

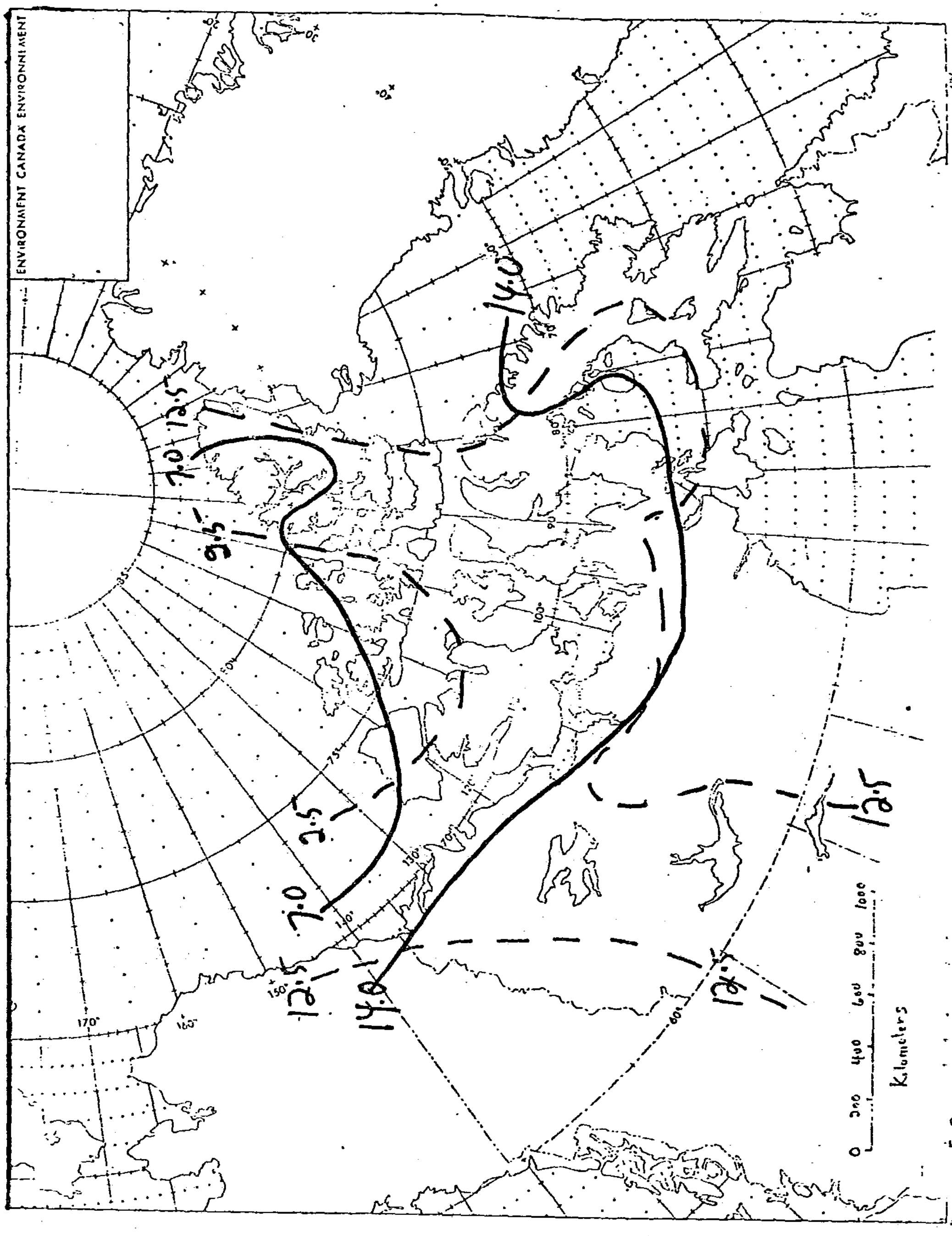
An attempt was made to estimate runoff amounts in the Archipelago. The technique used was to measure, by planimeter, the chart of the drainage areas over which the precipitation could be considered uniform. After multiplication of area by precipitation amount and appropriate map scale factor, the products were summed to give total volume of precipitation (in water equivalent) falling over the complete drainage area. The large scale areas considered are shown on Figure 5-1. The three drainage areas are the Mackenzie basin, the Arctic coast drainage east of the Mackenzie to Fury and Hecla Strait, and the archipelago west of Fury and Hecla Strait, as shown.

The precipitation values obtained from publications of the Canadian Atmospheric Environment Service were charted as in Figure 4-9. The volumes of precipitation resulting are shown in Table 5-1. The errors in the table are not easy to estimate. The errors in precipitation amounts probably far outweigh other errors. In the archipelago these precipitation amount errors may range up to 50 percent as noted in section 4. In the southern parts of the continental drainage, precipitation errors are less, perhaps of the order of 10-20 percent at most (Hare and Hay, 1971). Increases of these amounts should be applied to the precipitation volume estimates in Table 5-1, when firmer estimates of precipitation amount errors are available. From Table 5-1 the volume of precipitation falling on the Mackenzie drainage area is three times that falling over the archipelago, and about six times that falling over other arctic coast drainage in Canada.

The ratio of runoff to precipitation is quoted by Sellers (1965 p90) as 30-70 percent for forested land and 70 percent or above for tundra. Probably for the continental drainage an estimate of 50 percent of precipitation volume running off is adequate. For the archipelago 70 percent of the precipitation depths in Figure 4-9 give the runoff amounts (or rather depths) shown by the solid curves in Figure 5-2. However, the best estimates of runoff depths (quoted by Hare and Hay, 1974) are shown by the dashed curves of Figure 5-2. They differ widely from runoff amounts shown by the solid curves. The best estimates obviously take into account the types of terrain, the implication being of swampy terrain with high evaporation in the western archipelago, but



5-1 Drainage basins of the Mackenzie River, the Arctic coast drainage east of the Mackenzie River to Fury and Hecla Strait, and the Canadian Arctic Archipelago.



5-2 Estimated average annual runoff depths over the archipelago from precipitation chart 4-9 with a runoff/precipitation factor of 0.7 (solid line) and best runoff estimates quoted by Hare and Hay (1974), (dashed line), cm.

5-2 Estimated average annual runoff depths over the archipelago from precipitation chart 4-9 with a runoff/precipitation factor of 0.7 (solid line) and best runoff estimates quoted by Hare and Hay (1974), (dashed line), cm.

TABLE 5-1(a)

The Mackenzie River System

Mackenzie River Drainage

Average Annual Precipitation	Area of Basin (km^2) in Precipitation Ranges	Volume of Precipitation $10^9 m^3$ Water
cm		
16	4000	0.7
20	14700	3.0
25	252900	64.0
30	216300	66.0
35	702500	249.0
41	380100	114.0
46	120300	55.0
51	48200	24.0
56	31200	17.0
61	17400	11.0
68	18300	12.0
76	26800	12.0

Area Whole Drainage Basin $1.7 \times 10^6 km^2$

Volume of Whole Basin Annual Rainfall $6.4 \times 10^{11} m^3$ of water

TABLE 5-1(b)

Arctic Coast Drainage East of the Mackenzie River to Fury and Hecla Strait

Precipitation on Arctic Coast Drainage Excluding Mackenzie

Average Annual Precipitation	Area of Region in Precipitation Range (km^2)	Volume of Precipitation in $10^9 m^3$ water
cm		
14	93000	13
16	170000	280
20	239000	485
24	101000	238

Total Area $6 \times 10^5 km^2$

Total Volume of Precipitation $1.13 \times 10^{11} m^3$ of water

TABLE 5-1(c)

The Arctic Archipelago as in Figure 5-1

Average Annual Precipitation	Area of Region in Precipitation Range (km^2)	Volume of Precipitation $10^9 m^3$ water
cm		
6	2230	.1
9	395340	49.4
11	425980	48.7
14	637860	89.1
16	73600	11.6

Total Area $1.7 \times 10^6 km^2$

Land Areas $.9 \times 10^6 km^2$

Water Areas $.8 \times 10^6 km^2$

Volume of Annual Precipitation $2.0 \times 10^{11} m^3$ of water

perhaps of a higher percentage of runoff from the rocky sloping terrain of the eastern arctic, where precipitation is higher at higher elevations.

From the material in Figure 5-2 and Table 5-1 we can estimate crudely the amount of freshwater reaching channels. Over the whole archipelago, or that area of it shown in Figure 5-1 the depth of freshwater laid down by local precipitation and total runoff from land would, allowing for evaporation, be about 0.2 m. The freshwater from precipitation falling only on the water surface would be about 0.1 m. This is of course the average over the whole archipelago area shown in Figure 5-1. From the precipitation and runoff maps the freshwater laydown would be less than average in the Queen Elizabeth Islands. In addition in the channels of the southern archipelago appreciable quantities of the runoff from the Mackenzie River and other Arctic coast drainage would be added. If, for example, all Arctic coast drainage east of the Mackenzie River was added to the waters south of Parry Channel (area $4 \times 10^5 km^2$) the depth of freshwater additional to those mentioned above for local precipitation would be 0.2 m with the runoff factor of Sellers. The very great inflow of freshwater from the Mackenzie River does not discharge directly into archipelago water but water flow observations, see section 8, suggest that much of this Mackenzie freshwater would find its way into the channels around Banks Island and southern Victoria Island.

The technique of estimating runoff by multiplying precipitation amounts by drainage areas has been used by the Frozen Sea Research Group to make estimates of runoff in areas in which our oceanographic work was being carried out. These areas include the Greely Fiord, d'Iberville Fiord and the

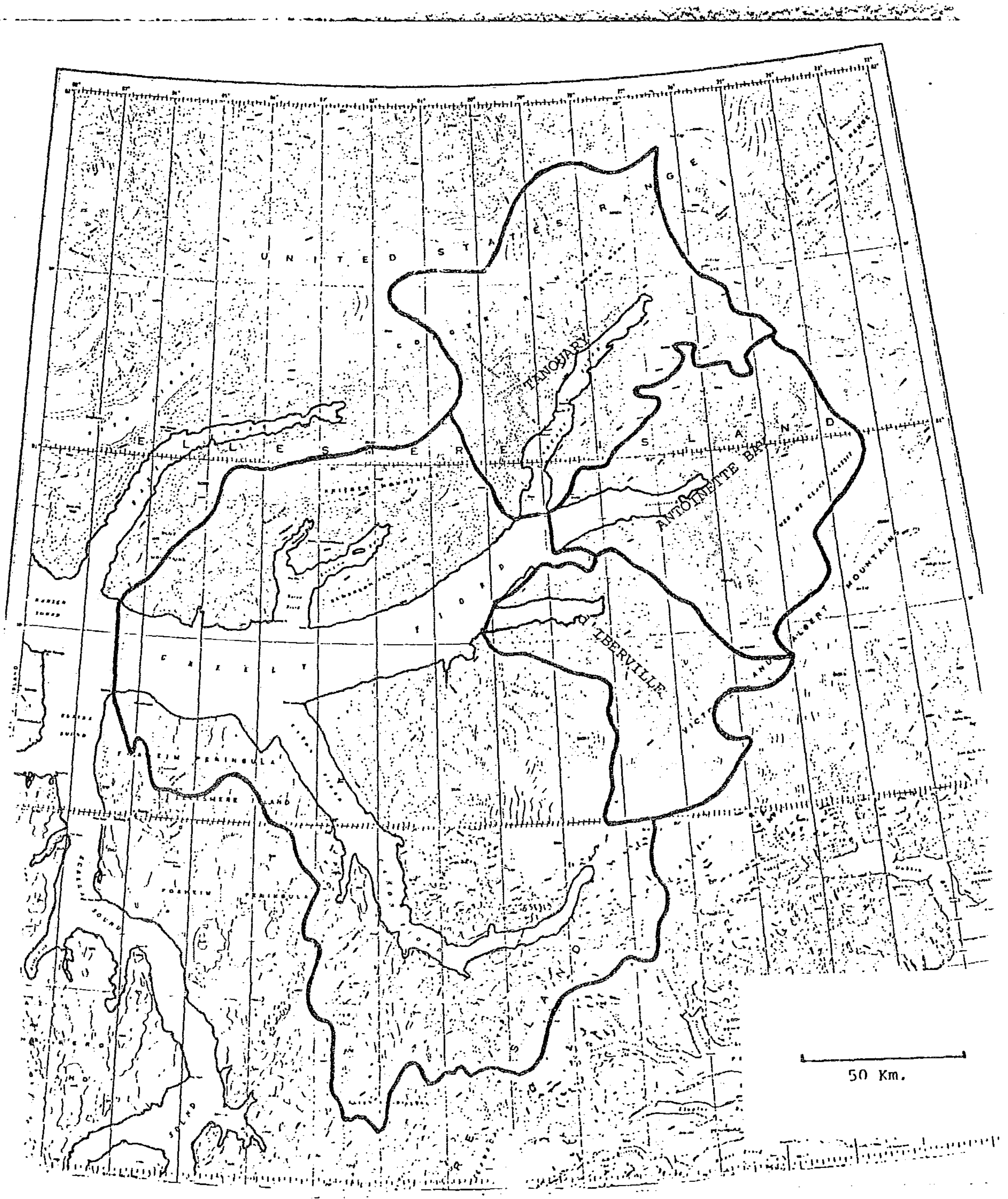
Cambridge Bay areas. The sources of error are largely in the precipitation amounts, and also in the evaporation rates, which are also poorly known. An example of our work for the Greely Fiord, d'Iberville Fiord area is given below.

For the Greely Fiord system the areas involved are shown in Figure 5-3. The areas of the drainage basins are shown in Table 5-2. The large ratio of the area of the drainage basins of the smaller fiords to the water area is evident. A large portion of the drainage areas is occupied by glaciers particularly in the north and east parts of the system. The amounts of precipitation over the area are very poorly known, to say the least. As a first approximation we may say the precipitation amounts measured at the only weather station in the area, Eureka, (6-7 cm), apply to the whole drainage area of the Greely Fiord system. This value is considered an underestimate. The areas in Table 5-2, multiplied by Eureka average annual precipitation give the volumes of precipitation shown in the table. These volumes of precipitation if laid down on the local water areas would give a freshwater layer 0.9 m deep in Tanquary Fiord and in Antoinette Bay, 1.3 m deep in d'Iberville Fiord, and 0.3 m deep in Greely-Canon Fiords. If evaporation is important or runoff incomplete depths of freshwater will be less, although the relative depths will probably be similar.

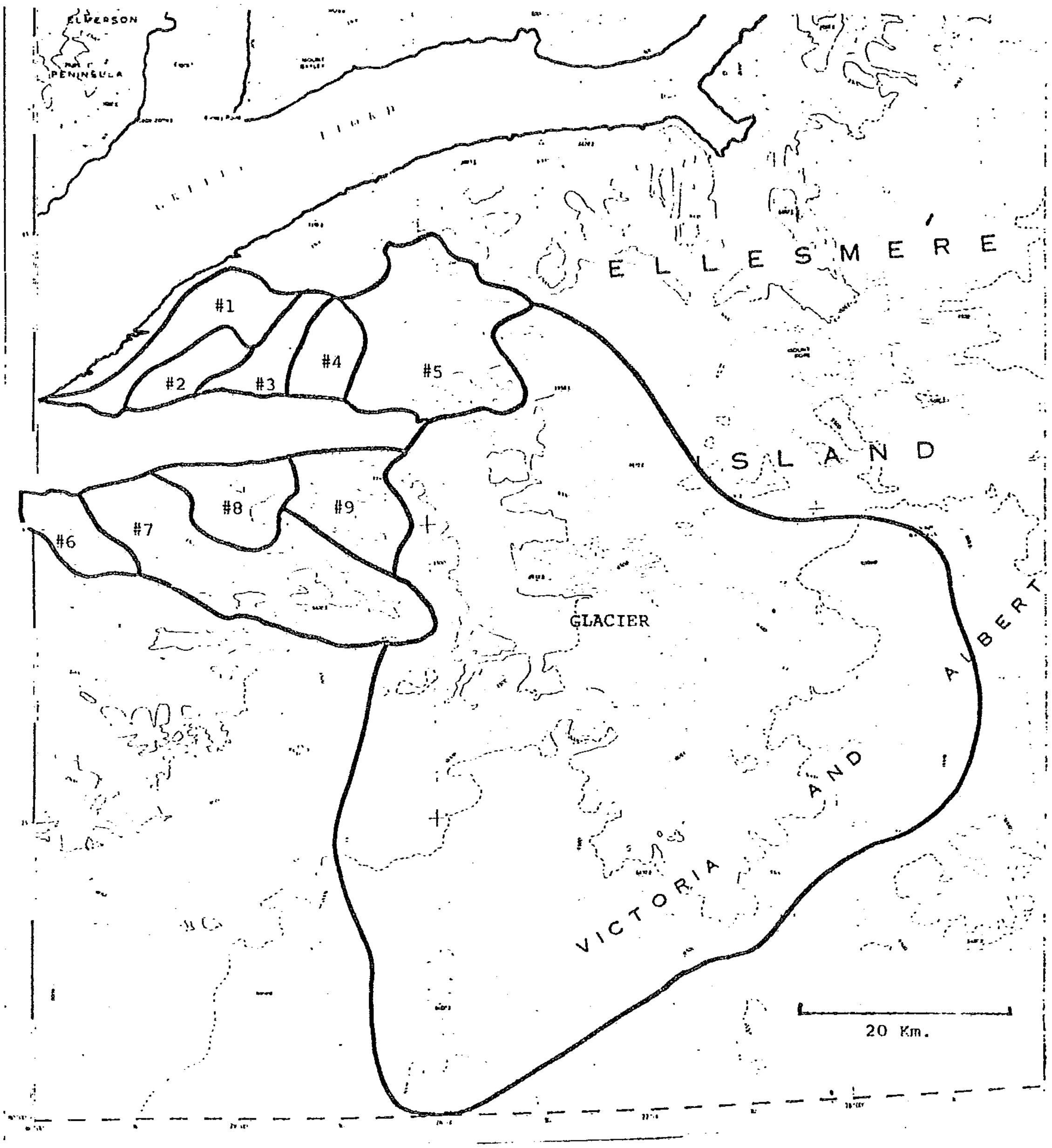
The estimates of runoff from the dashed curves of Figure 5-2 are shown in the last column of Table 5-2. The runoff estimates are 140 percent of precipitation volumes based upon Eureka precipitation. The timing of the runoff is not dealt with here but presumably will be similar to that suggested below for d'Iberville Fiord, with due allowance given to percent of drainage basin occupied by glaciers from which the runout timing differs from that of snow cover on bare ground.

Following an early review of d'Iberville Fiord oceanography (Lake and Walker, 1973) it became obvious that accurate estimates of the freshwater run off into the fiord must be made before modelling fiord water circulation could be realistically attempted. In these estimates the runoff amounts from techniques noted above were "corrected" by checking against measured river runoff.

The major river drainages into d'Iberville Fiord are shown in Figure 5-4. Basin areas planimetered from 1:250,000 scale charts are shown in Table 5-3. Probably the most striking feature of the drainage area is that the larger portion consists of glacier and ice cap at the east end of d'Iberville Fiord. Drainage basins of rivers #1 through #7 are unglaciated while the "Glacier" River basin is almost completely glaciated. In 1973 runoff was measured in #2 River (Ambler, 1974). In 1974 the flow in #2 River was again measured, and flows in basins of Rivers #1, #8 and the "Glacier" River were measured semi-quantitatively. In all basins vegetation is sparse and the terrain rocky. Terrain heights range up to about 760 m in basins of #1, #2, and #8 Rivers while the ice cap which forms most of the drainage area of the "Glacier" River ranges up to about 1370 m. Snow pack was measured before melt, and meteorological records were also made during June through August in both years, from the Frozen Sea Research Group base in d'Iberville Fiord.



5-3 Drainage basins of the Greely Fiord system, including Tanquary Fiord, Antoinette Bay, d'Iberville Fiord, and Greely-Canon Fiords.



5-4 Drainage basins of d'Iberville Fiord.

TABLE 5-2

Drainage System	Area (10 km^2)			Precipitation Volumes $10^6 m^3$	Runoff $10^6 m^3$
	Water	Land	Total	Eureka Precipitation	
Tanquary Fiord	0.5	5.8	6.3	424	729
Antoinette Bay	0.4	4.8	5.2	346	660
d'Iberville Fiord	0.2	3.7	3.9	264	456
Greely F.-Canon F.	4.8	17.6	22.4	1496	1809
Whole system	6.0	31.8	37.8	2530	3654

Computations of precipitation volumes on the Greely Fiord system, based upon Eureka weather station average annual precipitation. Runoff is estimated from dashed curves of Figure 5-2.

However, since our meteorological records were not complete we compared the runoff in d'Iberville Fiord with the 1973 and 1974 precipitation at the Eureka weather station. This was done so an average annual freshwater input to d'Iberville Fiord could be synthesized. We began with some drastic assumptions. We assumed precipitation amounts were equal over the whole drainage area of d'Iberville Fiord. This means we believe that the evaporation in the longer lasting snow fields compensates for the differences in precipitation amounts with elevation. The runoff from rivers #1 and #2 in 1973 and 1974 were related to the Eureka precipitation for the appropriate year. The runoff averages 1.9 times Eureka precipitation. This factor was then applied to the Eureka 1947-1970 average annual precipitation to give the average annual basin runoff amounts shown in Table 5-3. The glacier amounts are a balance between precipitation area amounts, observed iceberg output and estimates of undersnout melt.

These estimates for d'Iberville Fiord, when compared with the uncorrected precipitation in Table 5-2 indicate the magnitude of the corrections which may be necessary. Such corrections are particularly large in this area since the Eureka weather station is at sea level and is in the precipitation shadows of the mountains of Axel Heiberg and of Ellesmere Islands. Elsewhere and particularly in the western archipelago precipitation observations should be more representative although the problem of estimating snowfall remains.

Estimates of timing were made for modelling purposes. The runoff comes from melting snow in spring, and precipitation during summer. In d'Iberville Fiord fastest runoff was usual in spring as the snow melted, although in August, 1974 an exceptionally heavy rain gave briefly very high levels of runoff. We found snow on the sea ice melted about the time the spring flow pulse on the unglaciated rivers began, which is very soon after the mean daily air temperature rose above freezing. In d'Iberville Fiord we estimated about 1.5 m of sea ice melts in July and August. The rate of runoff from the glacier was more uncertain, but most should occur from June through September. A small

TABLE 5-3

Drainage Basin Areas for d'Iberville Fiord, N.W.T.,
With Average Annual Runoff Estimated as Described in the Text

Drainage Basin	Area (km^2)	Average Annual Runoff ($10^6 m^2$)
1	90	10
2	41	5
3	46	5
4	50	6
5	209	23
6	47	5
7	259	29
8	76	8
9	97	11
Fjord Surface	213	24
Glacier	2800	314
Icebergs		14
Runoff		277
Bottom Melt		23
Sum.	3930	440

input, from melting of the submerged portion of the ice shelf, and from melting of iceberg bottoms is present throughout the year. Normalized (to 1.0 at the end of August) flow for four rivers in d'Iberville Fiord in 1974 show the differences between rates of runoff for the rivers of the types mentioned above (Figure 5-5).

To sum up, even crude calculations can give some information but care should be taken to ensure that representative precipitation values are used. Further investigation of evaporation rates in Arctic conditions are needed I feel. Very little river gauging had been carried out in the archipelago until recently. However particularly in the eastern Arctic runoff data are becoming available (McCann and Cogley, 1972), (Anon, 1976) now which may be used to firm up crude estimates such as those given above, or perhaps in time make them unnecessary. As an example runoff gauging stations operating in the Northwest Territories and Yukon in 1975 are shown in Figure 5-6 (Anon, 1976). Runoff histograms for the four records in the archipelago show typical rapid summer runoff discussed in an extreme form for rivers in d'Iberville Fiord, (Figure 5-7).

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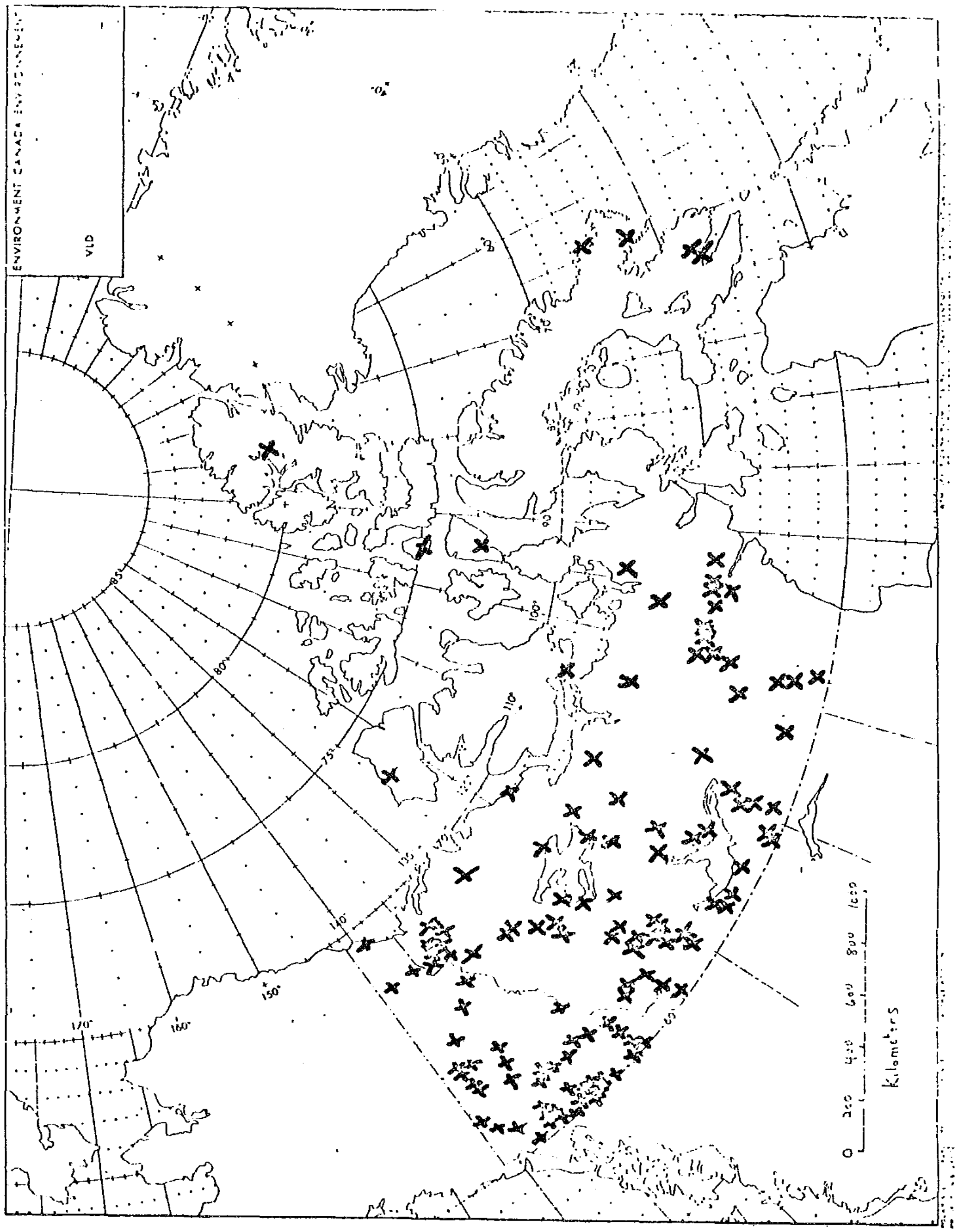
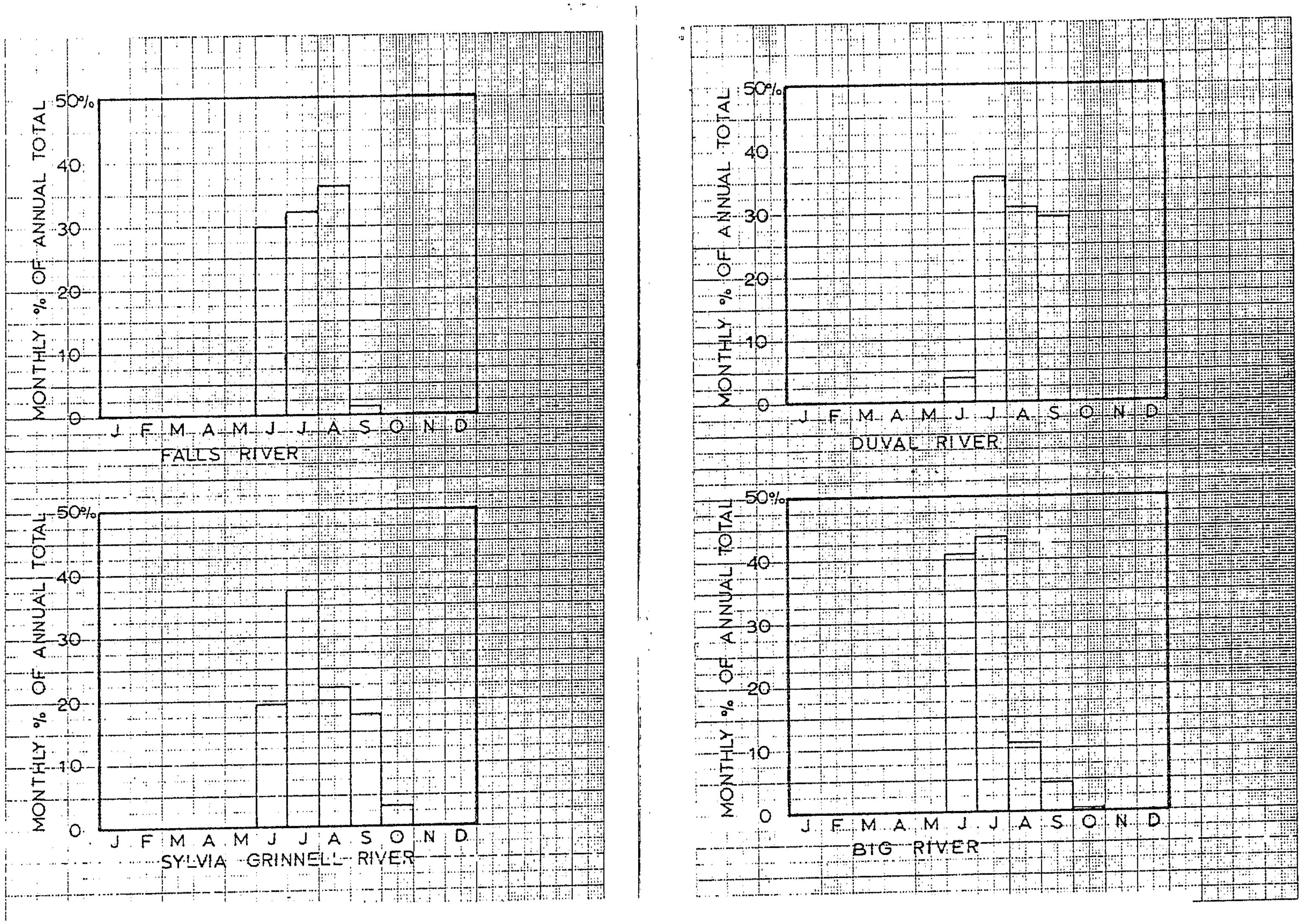


5-6 River runoff gauging stations in the Arctic, 1975 (Anon, 1975).

5-6 River runoff gauging stations in the Arctic, 1975 (Anon, 1976).



5-7 Monthly runoff percentages for Falls R (81N,80W), Sylvia Grinnell R (64N,69W), Duval R (66N,65W), and Big R (72N,123W) in 1975 (after Anon, 1976).

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TIDES IN THE ARCHIPELAGO

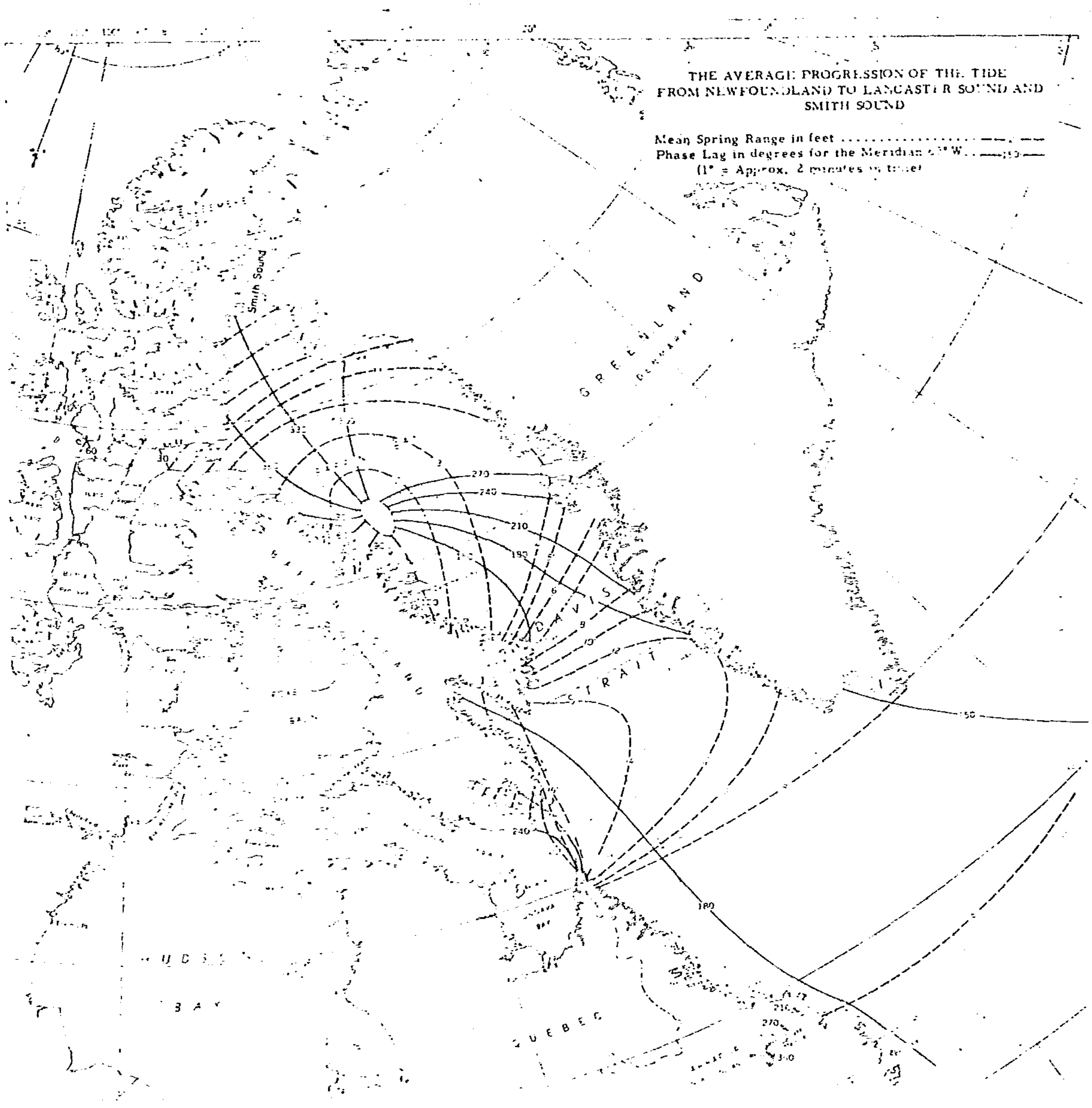
Tidal records in the archipelago, particularly in the northwest are scanty, and short. The Atlantic dominates the daily tides of the eastern portion of the Arctic Islands while in the northwest water levels are governed by the smaller tides of the Arctic Ocean. Exceptionally large mean tidal ranges are found on the southeast coast of Baffin Island south of Cape Dyer. Clearwater Fiord, Frobisher and Lake Harbour have mean tidal ranges of 5.9 m, 7.5 m, and 7.8 m respectively. The remainder of the east coast of Baffin Island appears to have mean values ranging between 0.6 and 2.0 m. In Figure 6-1 is shown the progression of the tide from Newfoundland to Lancaster Sound and Smith Sound (after Dohler, 1964). As one travels from east to west through Parry Channel the mean tidal range at points more or less evenly separated are Dundas Harbour, 1.9 m; Resolute, 1.3 m; Winter Harbour, 1.0 m; and Sachs Harbour 0.4 m. A more marked drop off in mean tidal range occurs as one travels north from Smith Sound to the Arctic Ocean. Pim Island on Smith Sound recorded a mean range of 3.1, Discovery Harbour 1.5 m, Wrangel Bay 1.2 m, Lincoln Bay 0.5 m.

Within the Queen Elizabeth Islands the tides are small. The mean tidal ranges in the Nansen Sound - Greely Fiord system of Ellesmere Island are less than 0.3 m with Eureka recording a mean range of only 0.1 m. Tides on the Arctic Ocean coast of the Islands are similarly small. From north to south values are 0.5 m, 0.3 m, 0.1 m, and 0.2 m for Alert, Cape Aldrich, Disraeli Fiord and Kleybolt Peninsula respectively on Ellesmere Island. Isachsen recorded a mean tidal range of 0.2 m. The majority of the mean tide ranges cited above are based on single 29 day records.

The character of the tide in northern Canadian waters is shown in Figure 6-2 after Dohler (1964). As the Pilot of Arctic Canada, Vl, p 104 remarks: "In all parts of the Arctic where observations have been made, the tides are semi-diurnal in character with two high waters and two low waters each day. In the central part of the Archipelago there is considerable inequality in the heights of successive high and low waters".

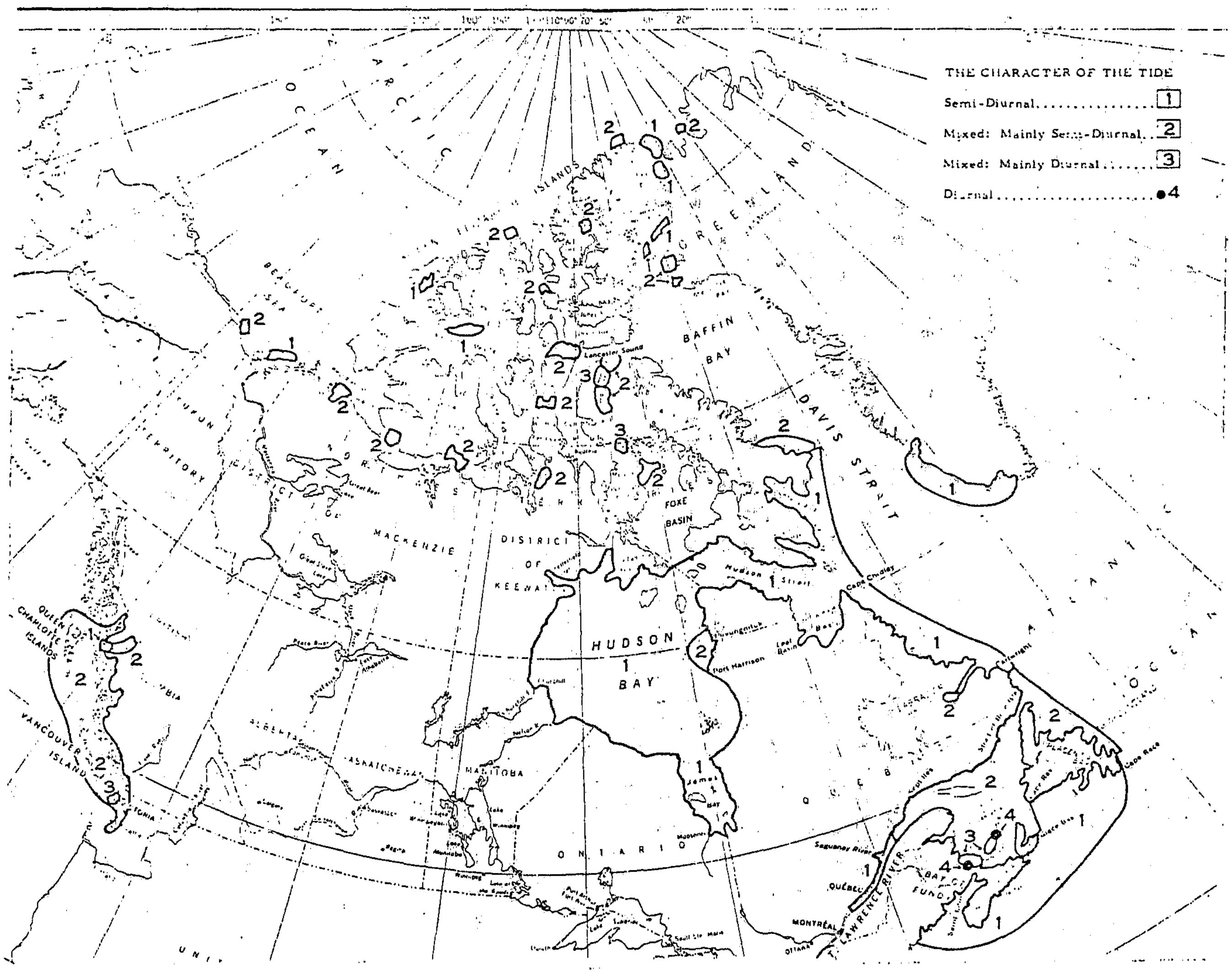
Despite the small size of daily tides in many parts of the archipelago, in narrow passages appreciable tidal currents can occur. Apart from areas kept open by these currents, tides and currents occur under continuous ice cover for most of the year. Such currents are noted in Bellot Strait (72°N, 95°W), Fury and Hecla Strait (70°N, 84°W) and the Hell Gate area (76°N, 90°W).

That part of the tide and water level spectrum between the daily tides and monthly mean values of water levels has been little investigated. The Frozen Sea Research Group obtained tide records from April 1973 to July 1974 in d'Iberville Fiord. When the daily tides were removed the residuals correlated very well, in an inverse sense, with the atmospheric pressure records from the AES weather station at Eureka N.W.T., (the sea surface was ice covered except for August, 1973). The predominant period in both records was 4 to 10 days. The difference in water levels peak to peak associated with these effects was about 20 cm. This type of sea surface level variation caused by weather systems of scale 10^2km and 10^5sec must have effects on movements of



6-1 The progression of the tide from Newfoundland to Lancaster Sound and Smith Sound (after Dohler, 1964).

6-2
The character of the tide in northern Canadian waters (after Dohler, 1964).



6-2 The character of the tide in northern Canadian waters (after Dohler, 1964).

waters in the channels of the Archipelago. Indeed in some of the (unpublished) current meter records we have seen disturbances of this time scale have appeared. These effects should be further investigated.

Some time ago the author (Walker, 1971) made a brief review of monthly values of mean and extreme water levels in the Canadian Arctic over the period 1963-1969. To bring the data base up to date recourse was made to the annual publications "Tidal" of the Tides and Water levels, Marine Science Branch then in the Canadian Department of Energy, Mines and Resources, but now in Ocean and Aquatic Sciences, Canada Department of Fisheries and Environment. A list of tide-recording stations, with months for which records were available is shown in Table 6-1. Of stations in the northwest archipelago only Cambridge Bay and Resolute records are in any sense complete over the period 1963-1969. Churchill, Frobisher and Nain are much further south and east.

When the monthly mean water levels were plotted as a function of time, similarities between curves are quite evident, a low level about the time of the spring equinox and a flat maximum during late summer and autumn. Despite the large difference in amplitudes of the daily tides between northwest and some of the southeast stations, the curves of the twelve monthly water level values for each station have very nearly the same amplitude of about 20 to 30 cm. A group of stations, Cambridge Bay, Resolute, Nain, and Churchill have curves which were similar for each year. At other stations the water levels differed more widely from month to month and year to year. Over the period of record there is no marked change in water level at any one station from year to year except at Alert. Plots of the monthly mean levels from 1963-1974 from selected stations are shown in Figure 6-3.

Of the several causes of the low frequency variations in sea level the ones of importance in northerly latitudes (Pattullo, 1963) are annual variations in atmospheric pressure and annual changes in water density (the tide gauges herein measuring water levels rather than pressure). Atmospheric pressure is measured by the Meteorological Service of Canada. Through their kindness values of mean monthly sea level atmospheric pressure were obtained for four stations in the Archipelago. Linear correlation coefficients between monthly values of sea level and atmospheric pressure were over 1963-1969, Alert, +0.04; Resolute, -0.70; Tuktoyaktuk, -0.49; Cambridge Bay, -0.83, markedly lower than the correlations at synoptic periods mentioned earlier.

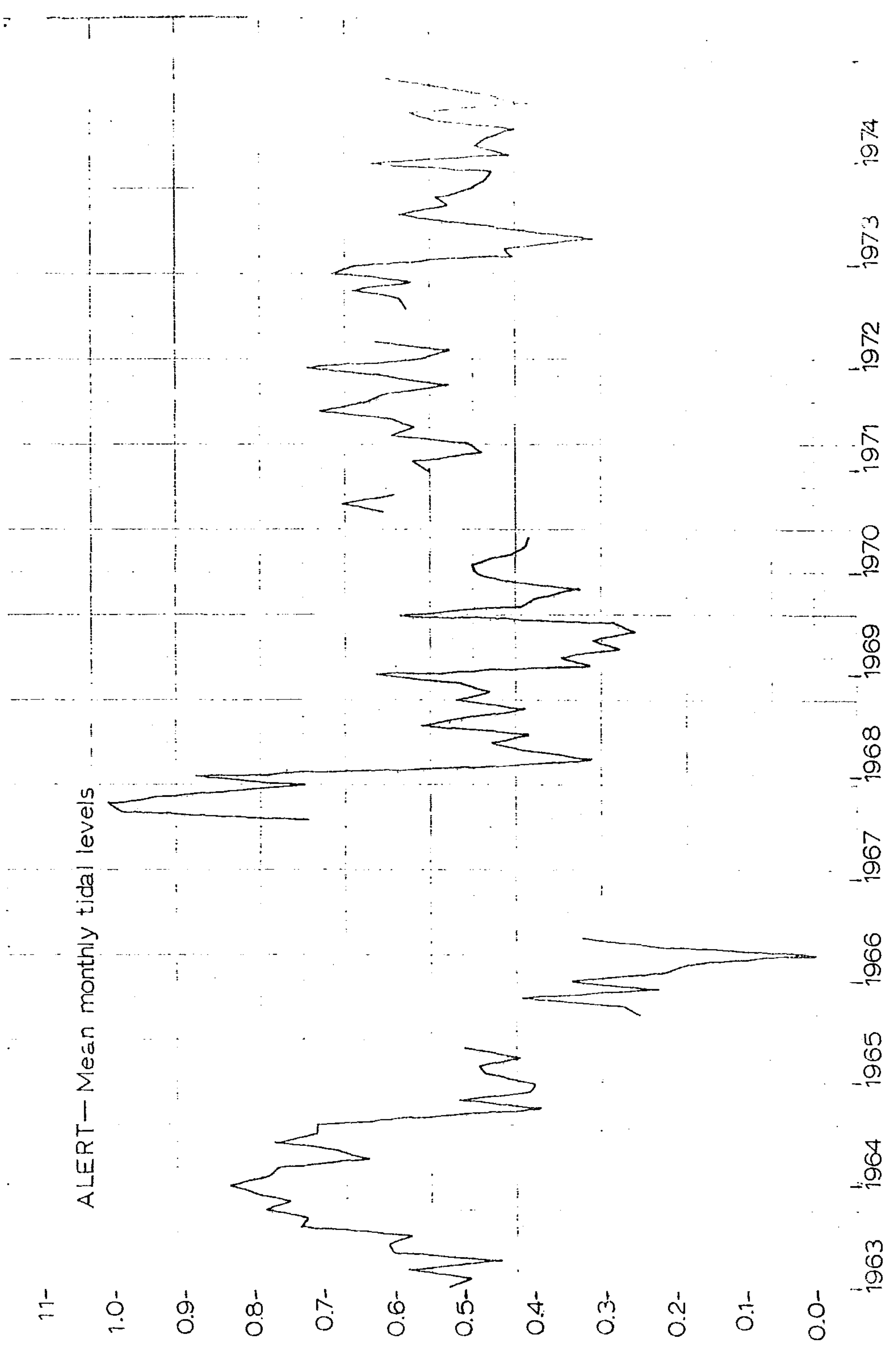
The changes in water density through the year in the Canadian Archipelago have nowhere been measured although as shown below they may be appreciable in surface water layers. The period of runoff of freshwater from land extends from May to September in the Archipelago. The actual amounts of freshwater in any area is critically dependent upon local topography, and rapidity of mixing over larger areas of sea. Unreliable estimates suggest a runoff freshwater depth of less than 0.3 m in general, with greater depths occurring (probably briefly) in restricted waters. In addition there may be an effect from the summer melting of sea ice. Two meters of sea ice, density 0.91 gcm^{-3} floating in seawater density of 1.02 gcm^{-3} would, if completely fresh raise the water level by 0.04 m. This compares with the rise in monthly water levels at Cambridge Bay of 0.2 m and at Resolute of 0.2 m from April to August. So quite probably some of the annual variation unexplained by

TABLE 1

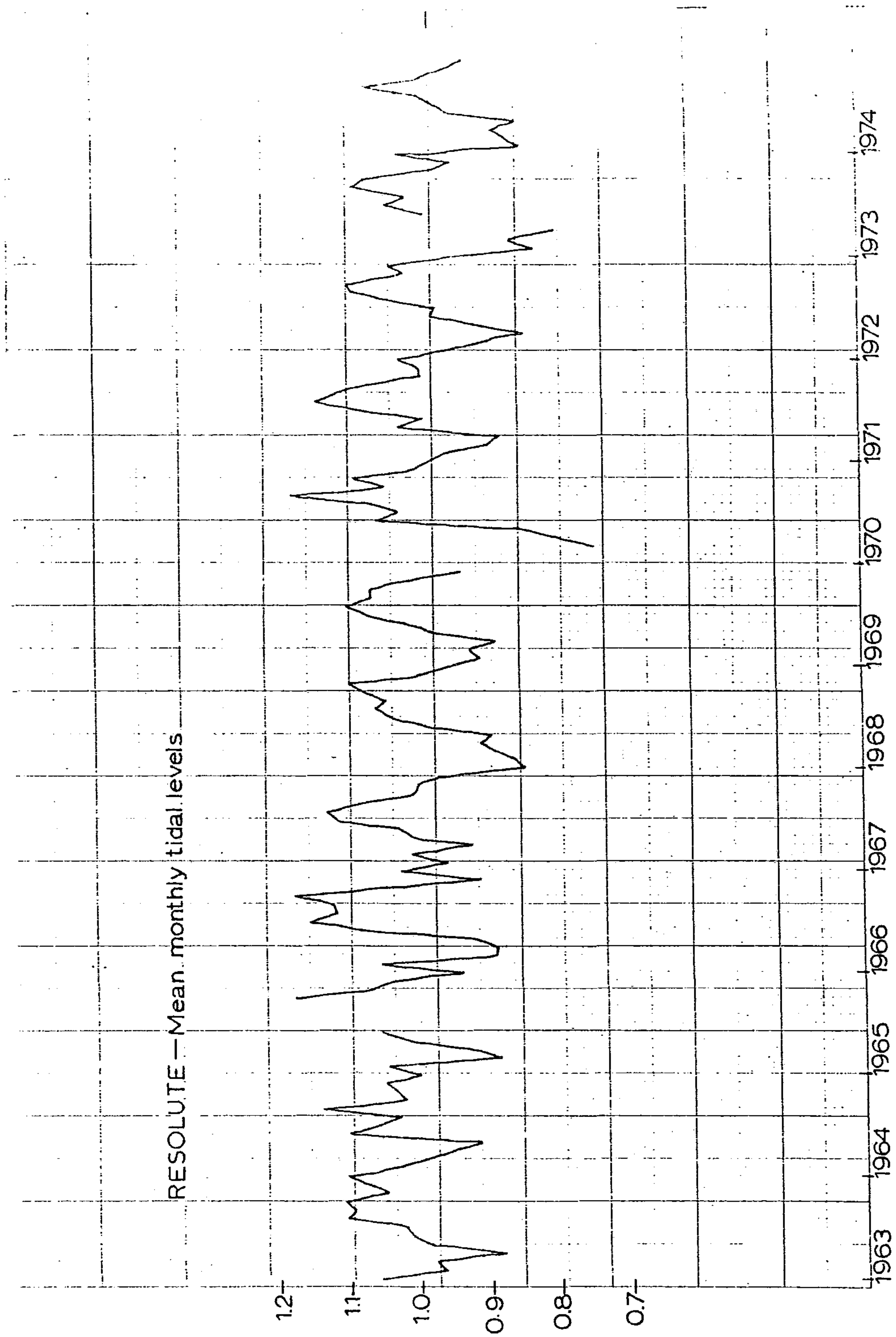
Arctic Water Levels Available 1963 - 1974

Station	Years of Record to 1974	1963	1964	1965	1966	1967	1968	1969	1970	1971	1972	1973	1974
Alert (82N,62W)	14	C	I	I	I	I	C	C	I	C	I	C	C
Cambridge Bay (69N,105W)	14	I	C	C	C	C	C	C	C	C	C	C	I
Cape Parry (70N,125W)	10	M	M	M	I	C	C	C	C	C	C	I	I
Churchill (59N,94W)	34	C	C	C	C	I	C	C	C	C	C	C	I
Frobisher Bay (64N,69W)	9	I	I	I	M	M	I	I	I	I	I	I	M
Resolute (75N,95W)	17	C	C	I	C	C	C	C	I	C	C	C	C
Tuktoyaktuk (69N,133W)	15	C	C	I	I	C	I	I	I	I	I	C	C
Nain (52N,62W)	11	I	C	C	C	C	C	C	C	C	C	I	M

6-1 Water level data available for Canadian Arctic Waters 1963-1974. Records are characterized as complete (C) if 11 or more monthly values were available, incomplete (I) if 1 to 10 monthly values were available, and missing (M) otherwise. (From Tide and Water Level publications, Canadian Hydrographic Service).



6-3 Monthly mean water levels (m about an arbitrary level) for Alert, Resolute Bay, Cambridge Bay and Nain. (data from Canadian Hydrographic Service).



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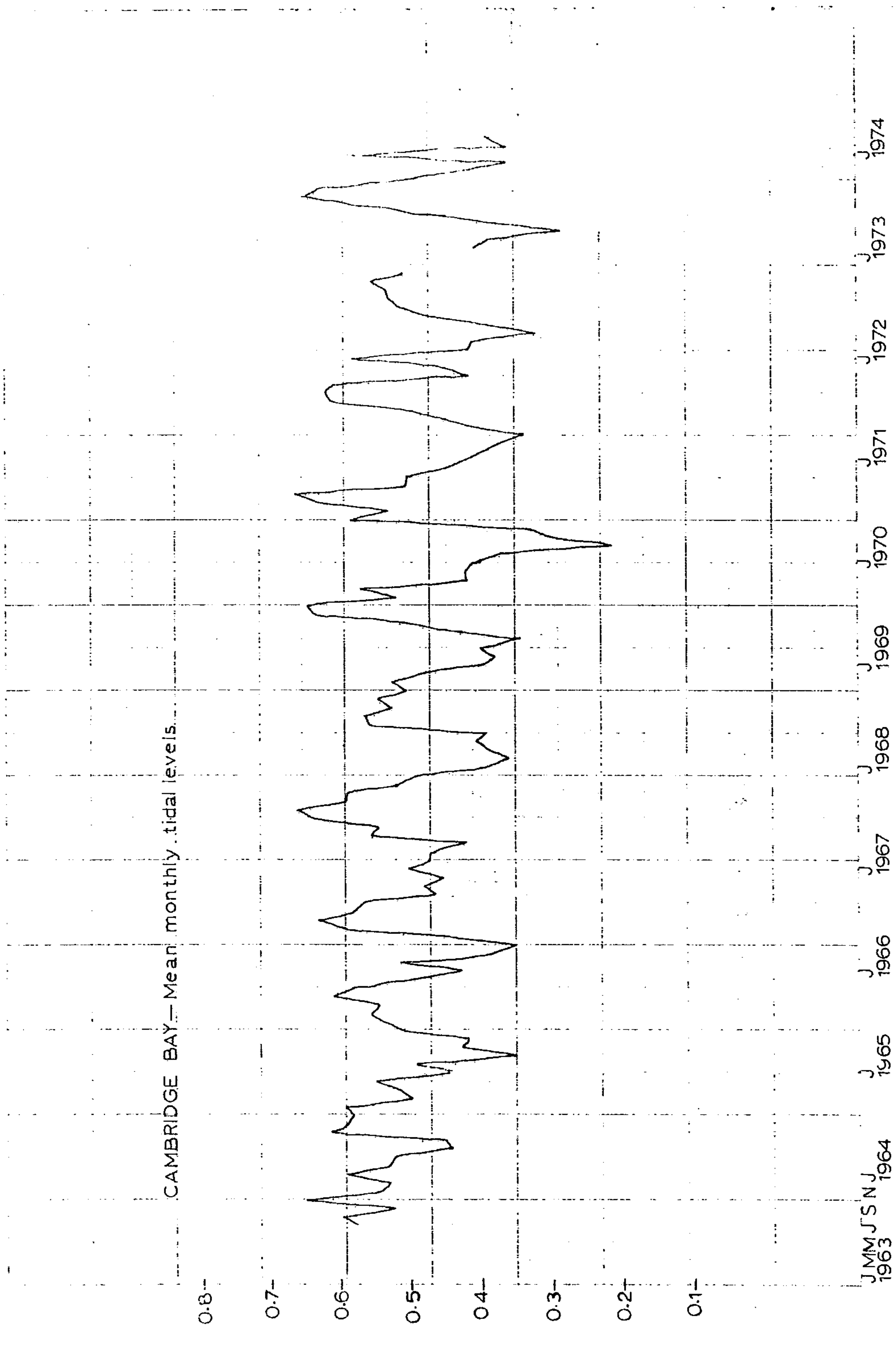
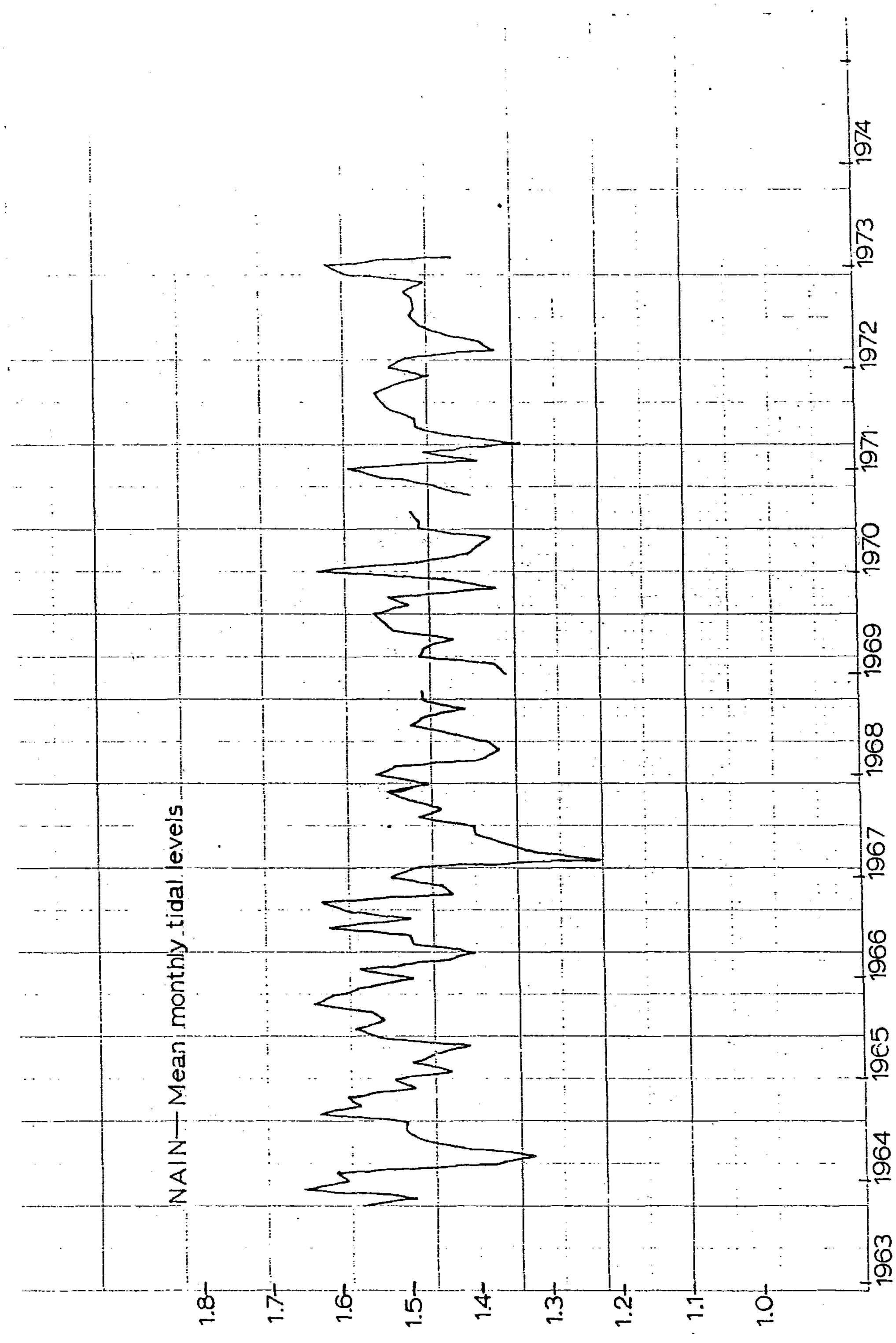


6-3 Monthly mean water levels (m about an arbitrary level) for Alert, Resolute Bay, Cambridge Bay and Nain. (data from Canadian Hydrographic Service).



6-3 Monthly mean water levels (m about an arbitrary level) for Alert, Resolute Bay, Cambridge Bay and Nain. (data from Canadian Hydrographic Service).

atmospheric pressure correlations may be due to water density changes. The analysis of variance in any quantitative fashion will have to await more complete oceanographic measurements.

Averages, over 1963-1974, of monthly values of mean sea level, high and low values of daily mean, and high and low values of the instantaneous water levels are plotted in Figure 6-4 for stations in the high Arctic and for Nain in Labrador. The overall range of sea level values during the period was about 1.5 m in the Cambridge Bay, Cape Parry records, 3.0 m at Tuktoyaktuk, and 2 m at Resolute and Alert, and 3.5 m at Nain. The outstanding features of the Tuktoyaktuk records are the occasional very high or low values of instantaneous water level, usually accompanied by corresponding values of the daily mean level. These are undoubtedly storm surge type phenomena lasting long enough to affect the daily high level. No extensive investigation was undertaken but at the more prominent peaks in July, October 1963, September 1966, daily mean water levels were high for periods of one to four days and the instantaneous high occurred on the day with the highest mean.

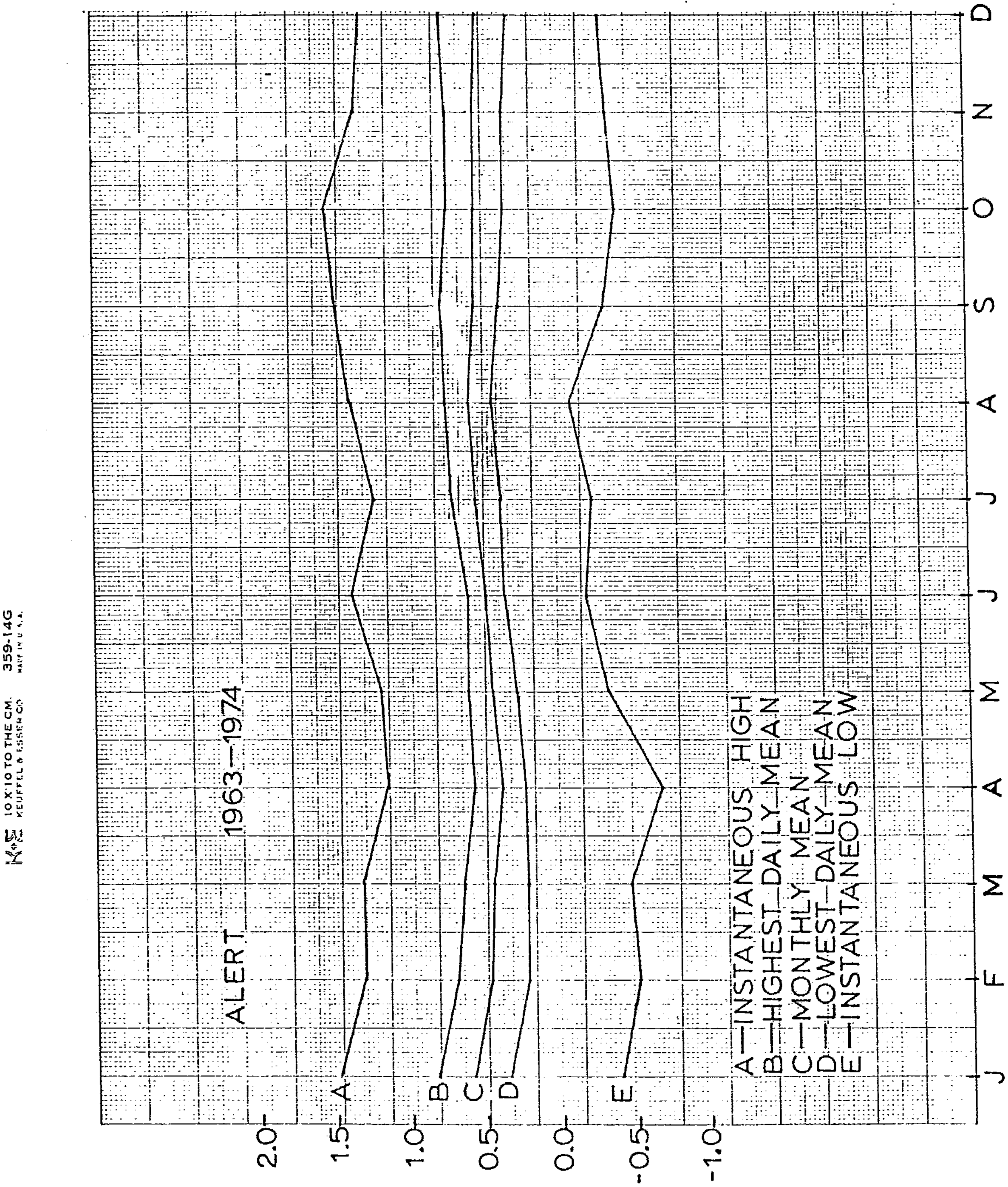
In Figure 6-5 the monthly values of mean water levels were plotted in pairs from 1968-1974 with the mean levels for the period aligned. It can be seen that differences in water levels do exist, do persist, and presumably have relevance to water movements. However obviously a more complete analysis is necessary. The water levels from the archipelago should also be related to the water levels in the Arctic Ocean. The only Arctic Ocean data presently available to the author are those referenced in section two and they are hardly adequate even for a qualitative comparison.

To sum up and generalize rather broadly, we may say that the overall water level ranges (over the years of record) are small in the upper Canadian Arctic archipelago of the order of 1.5 - 3.5 meters. This range is increased to over 3 m at Tuktoyaktuk probably by storm surges. It would appear likely that Mackenzie River - Beaufort Sea zone is the only area in the northern and western arctic subject to pronounced storm surges, ice-free fetches being limited elsewhere. Alert, the northernmost station, had in its record a longer period change, as evidenced in the segment of its record from 1963-1966, although this may be due to instrumental difficulties.

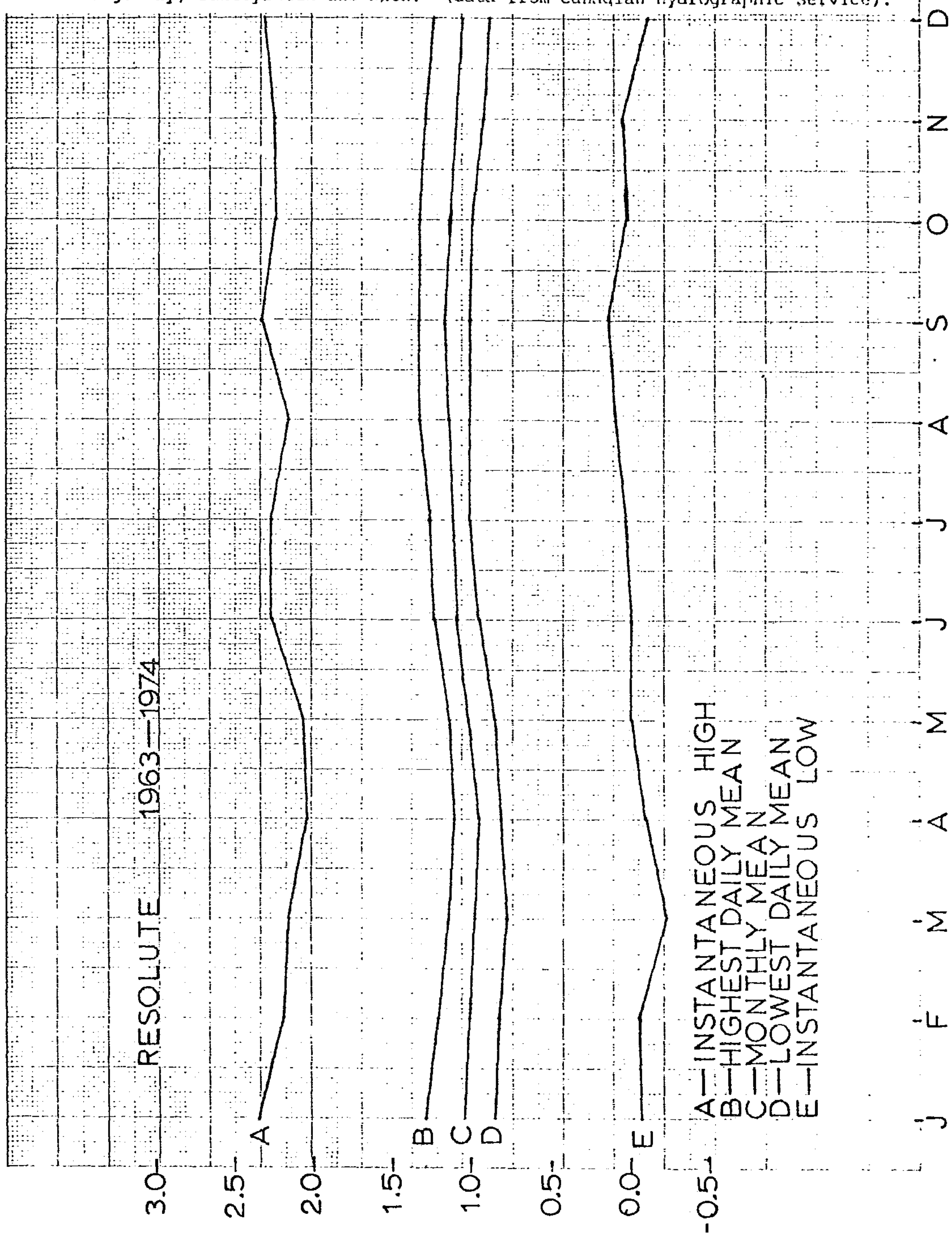
The size of very long term changes in water levels in the Canadian Arctic is not known, except for isolated areas in which water levels have been related to isostatic rebound from the glacier loading.

The importance of tides and water level differences in causing water movement through the archipelago is still in question, for the longer period phenomena. Daily tides certainly cause appreciable water movements. The loading and relaxation from the moving atmospheric cyclones should cause movement. Presumably to date any such effects have been obscurred by wind-caused currents. Shifting values of the monthly or longer term spatial values of water levels should be instrumental in causing flow through the channels of the Archipelago with annual variation (which latter has not been observed in the archipelago though it has in the inflow currents to the Arctic Ocean from the North Atlantic as noted in section 2). However, it seems that current data to establish these effects are still lacking.

6-4 **Average** over 1963-1974 of monthly values of mean water level, high and **low values** of the daily mean, and extremes of high and low values of **instantaneous** water levels (m about an arbitrary level) for Alert, Resolute, **Cambridge** Bay, Tuktoyaktuk and Nain. (data from Canadian Hydrographic Service).



6-4 Average over 1963-1974 of monthly values of mean water level, high and low values of the daily mean, and extremes of high and low values of instantaneous water levels (m about an arbitrary level) for Alert, Resolute, Cambridge Bay, Tuktoyaktuk and Nain. (data from Canadian Hydrographic Service).



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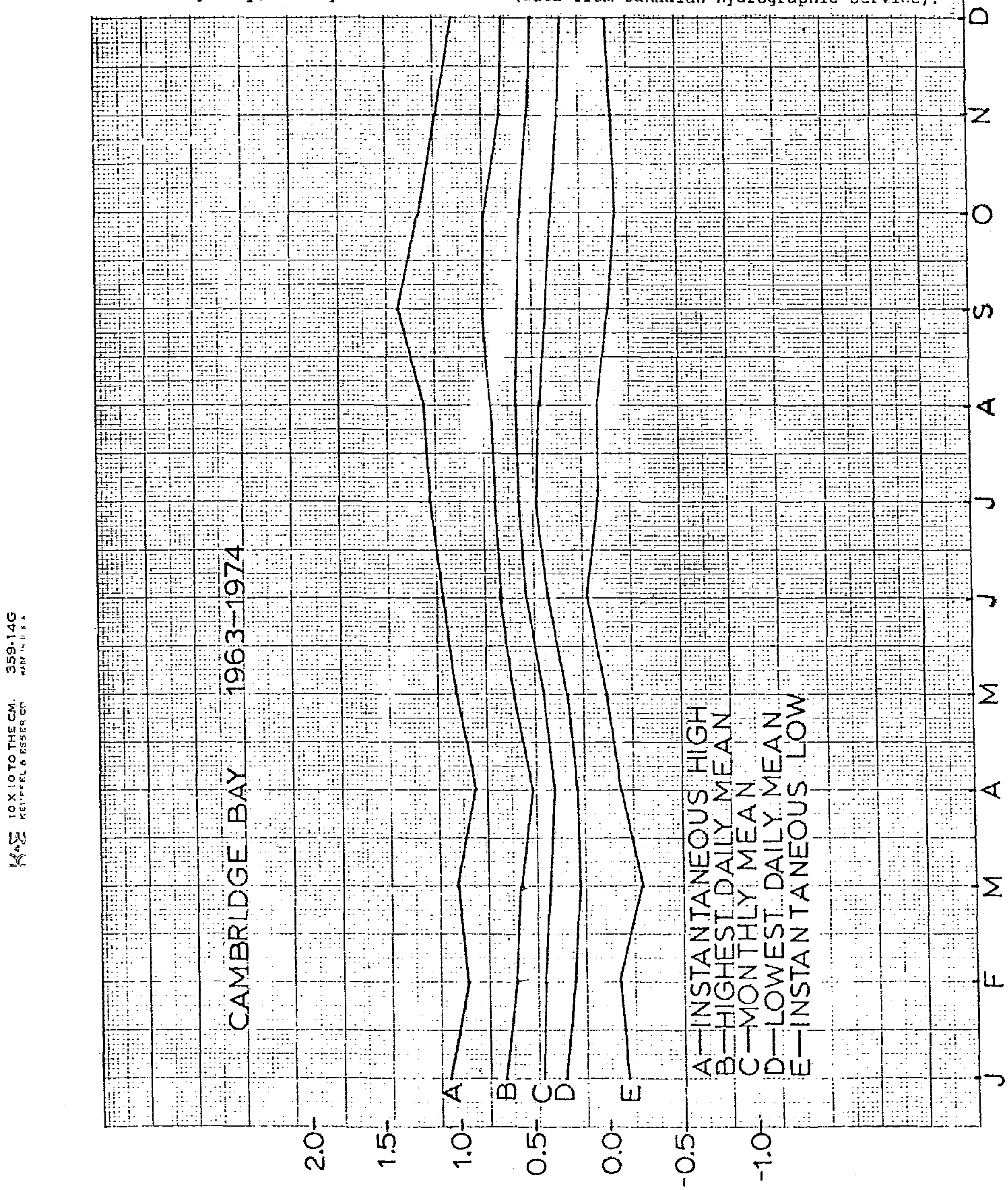

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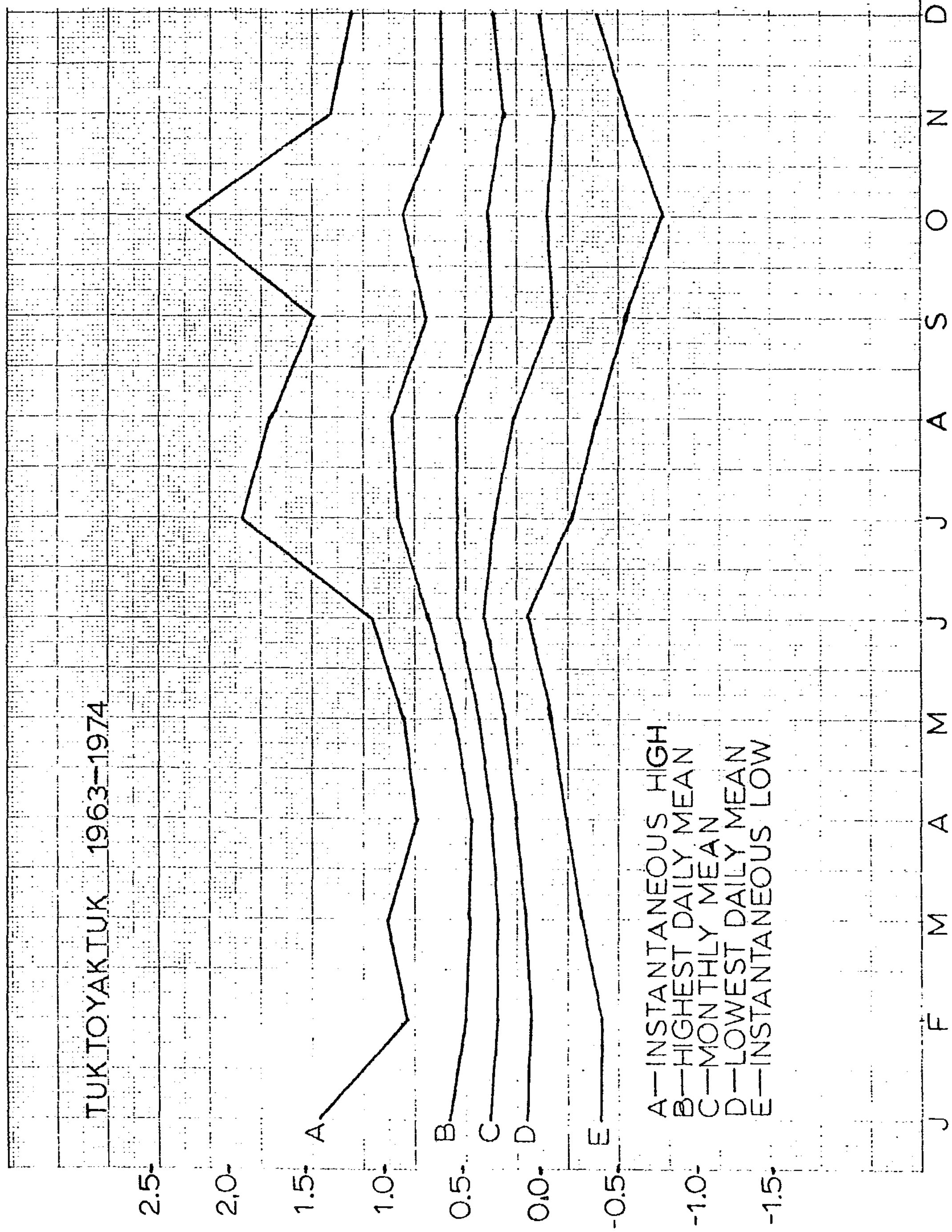

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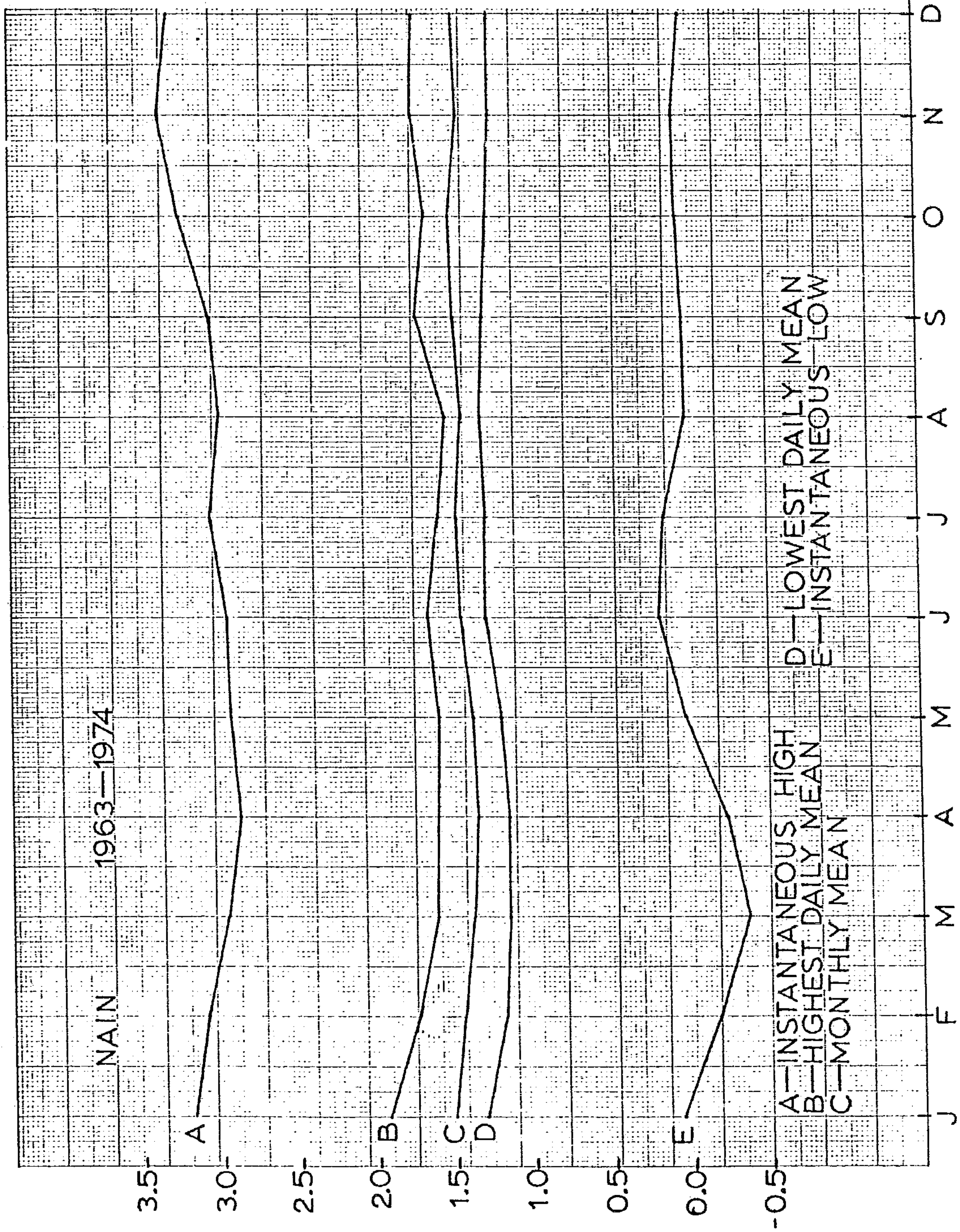
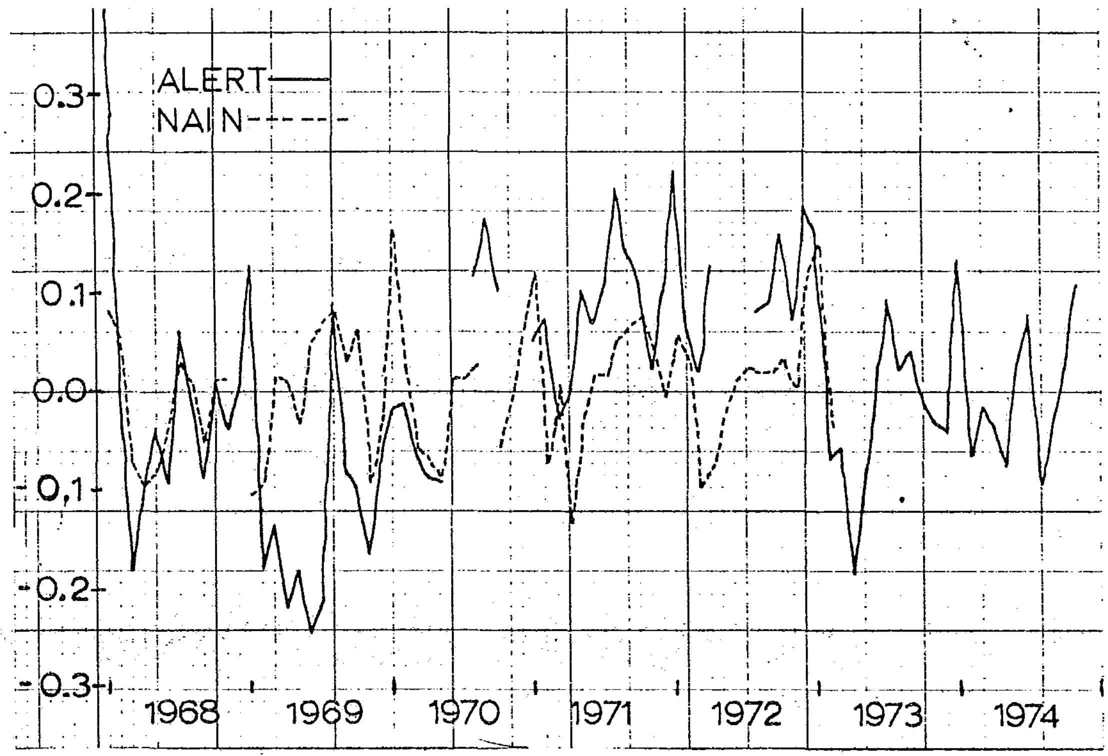
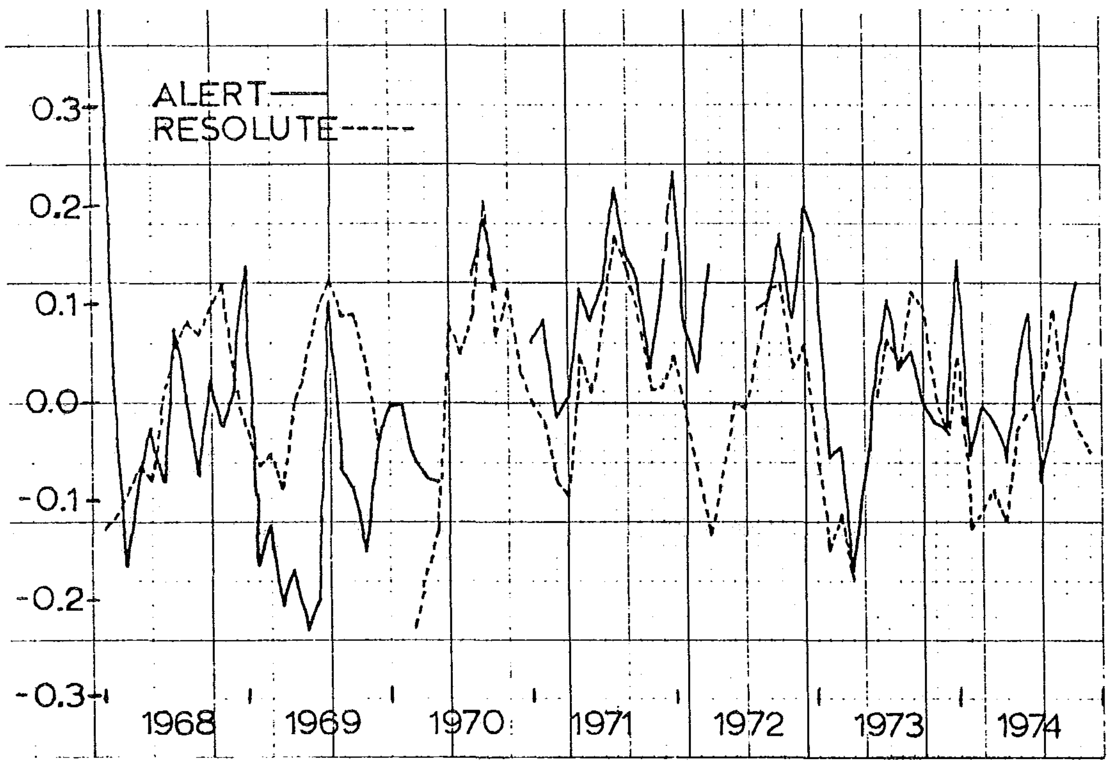
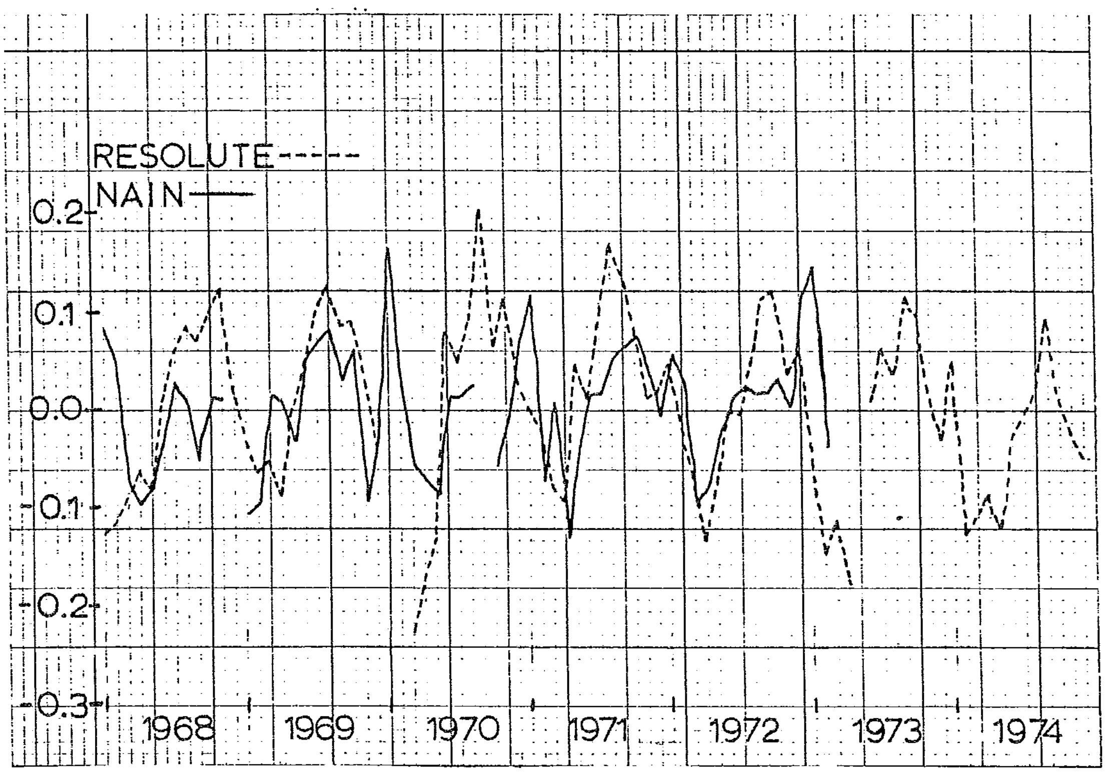




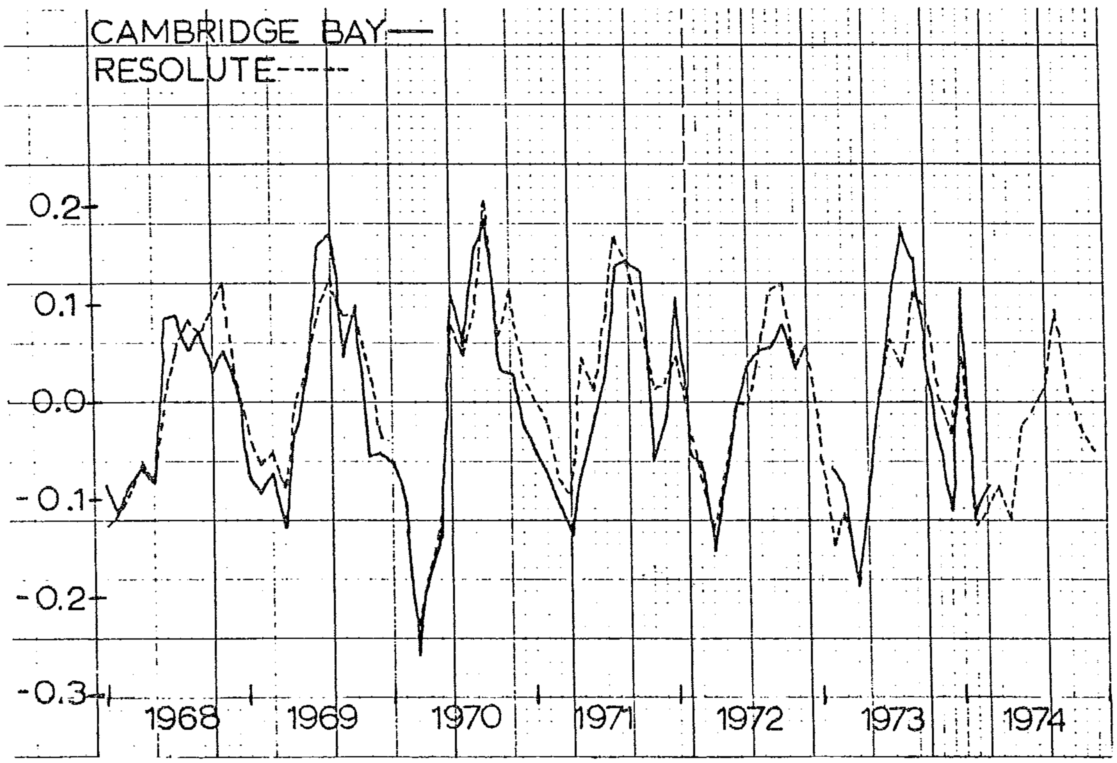
6-5 Values over 1968-1974 of monthly mean water levels for pairs of stations (a) Alert-Nain, (b) Alert-Resolute, (c) Resolute-Nain, (d) Cambridge Bay-Resolute, (m about an arbitrary level), data from Canadian Hydrographic Service.



6-5 Values over 1968-1974 of monthly mean water levels for pairs of stations (a) Alert-Nain, (b) Alert-Resolute, (c) Resolute-Nain, (d) Cambridge Bay-Resolute, (m about an arbitrary level), data from Canadian Hydrographic Service.



6-5 Values over 1968-1974 of monthly mean water levels for pairs of stations (a) Alert-Nain, (b) Alert-Resolute, (c) Resolute-Nain, (d) Cambridge Bay-Resolute, (m about an arbitrary level), data from Canadian Hydrographic Service.



6-5 Values over 1968-1974 of monthly mean water levels for pairs of stations (a) Alert-Nain, (b) Alert-Resolute, (c) Resolute-Nain, (d) Cambridge Bay-Resolute, (m about an arbitrary level), data from Canadian Hydrographic Service.

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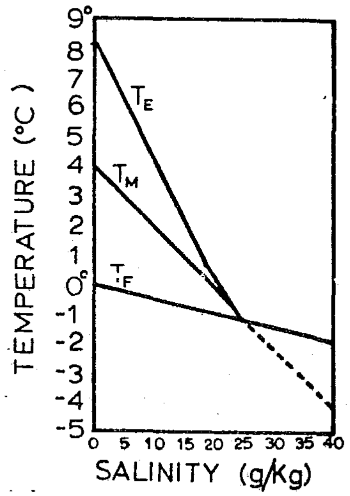
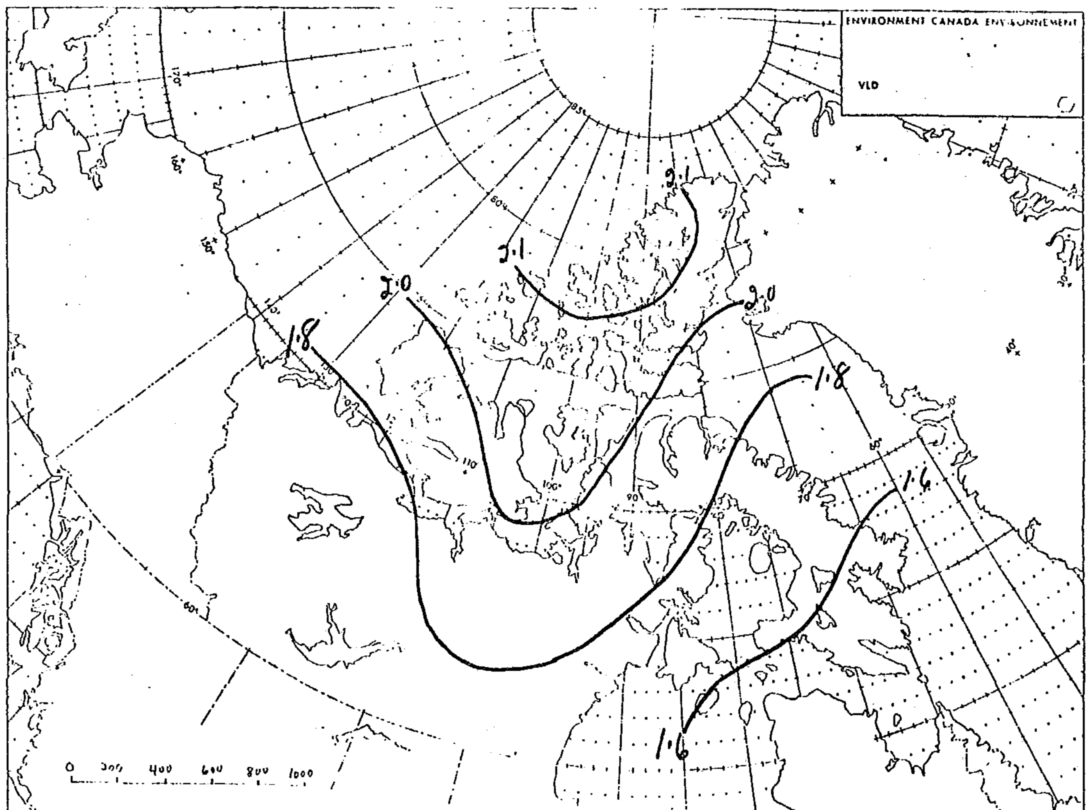
Because of the large scale surface heat balance in high latitudes, water surfaces in the Canadian Arctic commonly freeze in autumn. The freezing point of water and the temperature of maximum density is shown in Figure 7-1, (after Zubov, 1943). The maximum density of freshwater occurs at a temperature of about +4°C, while for sea water of salinity greater than 24.7 g kg^{-1} the temperature of maximum density coincides with the freezing temperature. The freezing temperature T_{FR} (°C) of seawater is given by $T_{FR} \approx -0.054\ S$ where S is salinity in g kg^{-1}. In autumn fresh water cools at the surface to +4°C with gravitational mixing occurring but further cooling leads to lighter surface water (T<+4°C) which can cool to freezing with ice formation in a gravitationally stable situation. The same process occurs in sea water whose salinity is less than 24.7 g kg^{-1}. For water of salinity greater than 24.7 g kg^{-1} the cooling water at the surface is at maximum density for a specific salinity and hence will mix downward at freezing temperature. At freezing temperatures, density (ρ) is largely a function of salinity with $\partial\rho/\partial S \approx 0.8\ kg\,m^{-3}(g\,kg^{-1})^{-1}$ while $\partial\rho/\partial T \approx 0.04\ kg\,m^{-3}(°C)^{-1}$ at $T \sim 0°C$, $S \approx 30\ g\,kg^{-1}$. As indicated below in the sections on under-ice convection, and water column structure, what usually happens when sea water freezes is that gravitational convection occurs from the surface downward to those depths where pre-existing higher values of salinity cause a water density too great for the mixing from the surface layers to penetrate.

The time of onset of freezing and the rate and amount of winter ice growth have been studied for many years. For empirical purposes, since knowledge of terms in the surface heat budget was not commonly available, ice growth was related more or less simply to accumulated degree days below freezing. Bilello (1960) has summarized formulas of type $I = a(\Sigma\phi)^b$ giving for snow free ice $I^2 = 12.6\ (\Sigma\phi)$ where I = ice thickness in cm, $\Sigma\phi$ = degree days below -1.8°C, a, b are constants. Zubov (1943) gave an equation of form $I^2 + 50I = 8\Sigma\Theta$ where Θ = degree days below 0°C. Leahey (1966) has provided an equation to deal with growth of snow covered sea ice of form $I^2 = 8.6\,\Sigma(\theta_s + 1.8 - \frac{Q\delta}{85})$ where Q = surface heat balance (cal $cm^{-2}day^{-1}$), θ_s = snow surface temperature (°C), and δ = depth in cm of snow on ice. Evaluation of the latter equation needs knowledge of Q, a quantity not generally available. Other more complicated formulations are available but a simple formula like that of Zubov is adequate for many purposes. The average maximum of ice thickness over the archipelago using Zubov's equation on the isopleths in Figure 4-8 is shown in Figure 7-2.

More complex numerical models, incorporating more of the physics of the process have recently been developed, among them the one dimensional thermodynamic models of Maykut and Untersteiner (1971) and Semtner (1976). The formation and growth of sea ice is also affected by water structure as noted below and several models incorporating these effects are included and discussed in Zubov (1943), Solomon (1973) and others. For large open areas where movement of ice and the internal stresses in the ice sheets are important mixed thermodynamic and dynamic models in two dimensions are being developed, (Thorndike et al, 1975; Bugden, 1976). The ultimate model, not yet attained, will couple water, air and ice. Descriptions of the decay of sea ice in the archipelago are to be found in a paper of Langleben (1968), Jacobs et al (1975)



7-1 The freezing point of seawater T_F, the temperature of greatest density T_M, and the temperature T_E of the density equal to density at the freezing point. (after Zubov, 1943).



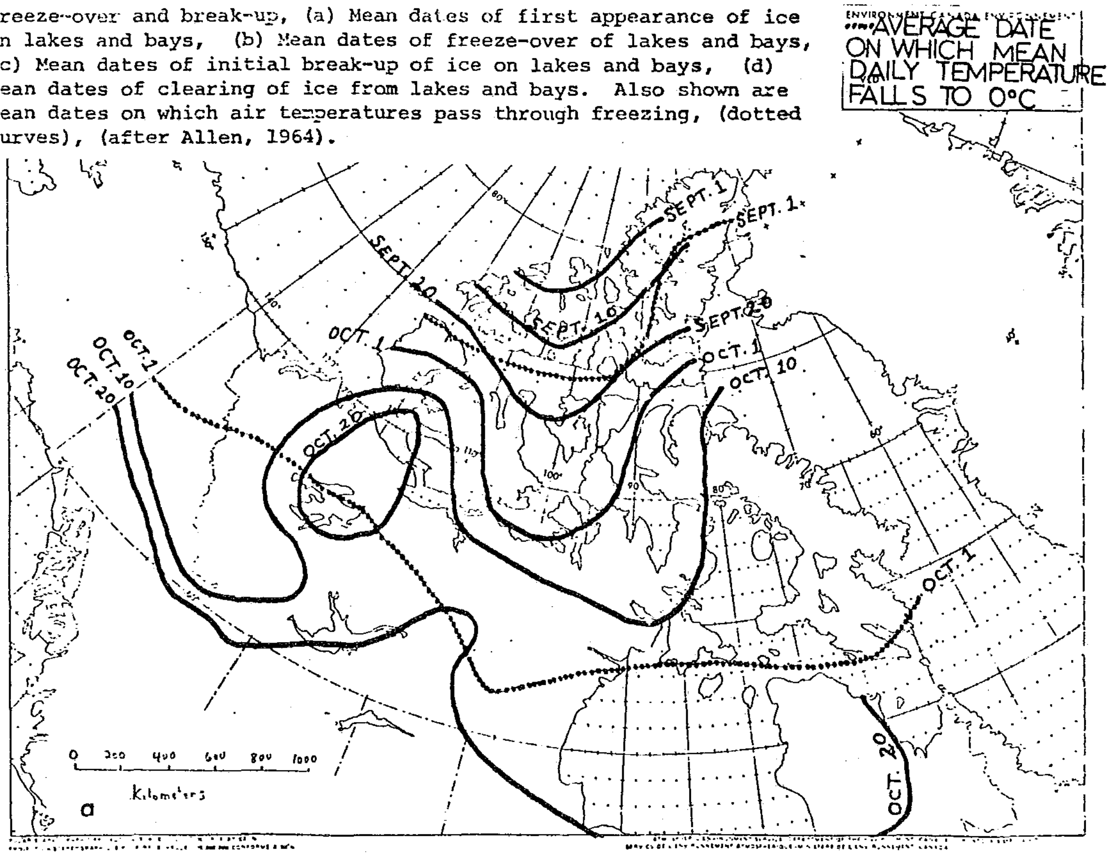
7-2 Average maximum of ice thickness expected over the Canadian Arctic using Zubov's (1943) equation on the average degree days from Figure 4-8 (m).

and in the ice coverage atlases noted below. The processes involved in ice sheet decay are not so easily modelled by the one-dimensional thermodynamic models as is ice formation. Bilello (1960) has provided an empirical formula $h = 0.55\ \Sigma\phi$ where h is ice melt (cm) and ϕ are degree days above -1.8°C. Langleben (1972) has revised an ice decay model of Zubov to take into account radiation absorbed by the ice which increases the rate of disintegration to realistic values. However, it does seem that precise forecasting of ice break-up will have to await the ultimate model mentioned above.

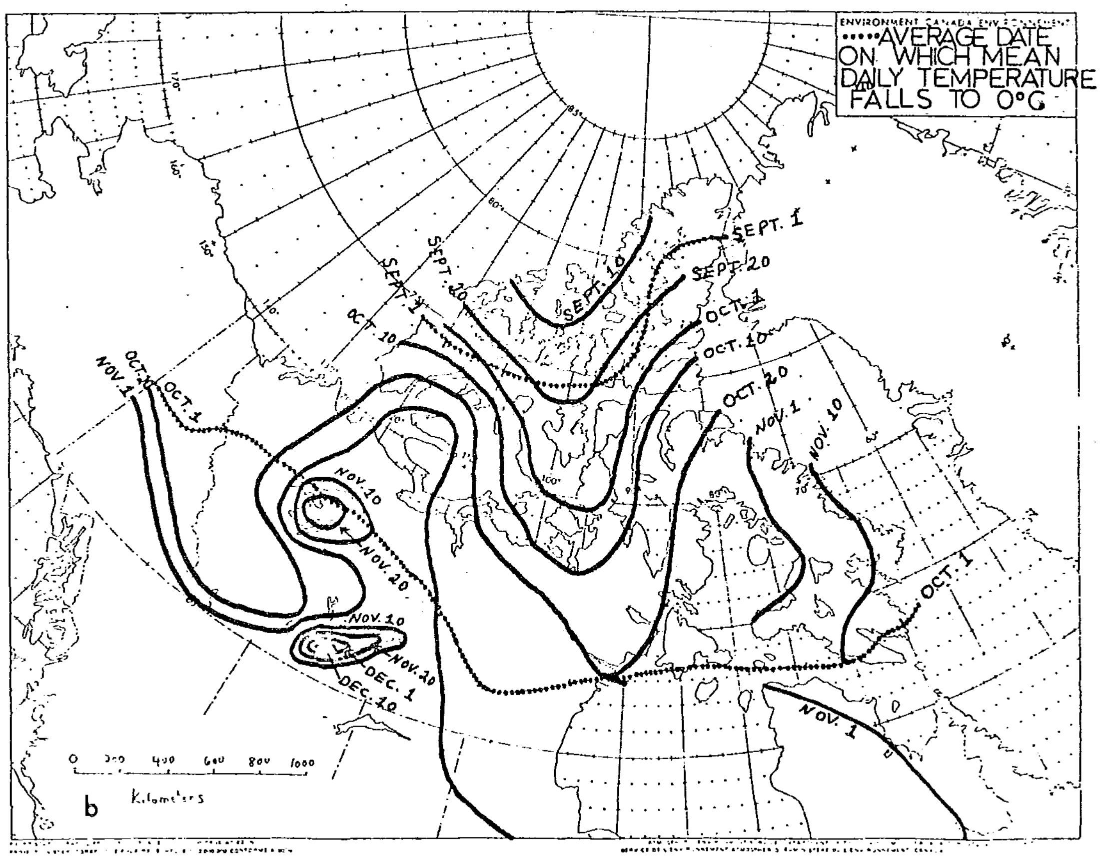
Observations of ice in the Canadian Arctic have been of three kinds. The first kind consists of aerial inspections of most channels in the archipelago each summer from which a good idea of ice coverage is gained by, but only qualitative information on ice thickness. Another type of data is timing of freeze up on lakes, rivers and bays by the Atmospheric Environment Service, Canada, which material is published annually. In Figure 7-3, average values of these quantities after Allen (1964), are shown. Also shown in the figure (dotted curves) are the dates on which air temperatures pass through 0°C.

The Atmospheric Environment Service has since 1958 also observed weekly values of ice thickness on bodies of water adjacent to their Arctic and sub-Arctic meteorological stations. This material has recently been summarized by Richardson and Burns (1975). Their publication allows easy preparation of such diagrams as average ice growth at representative arctic weather stations, Figure 7-4. The average maximum sea ice thickness, with the snow cover at maximum ice thickness, is shown for the years 1964-1974, in Figure 7-5. Although the material for Figure 7-5 was prepared independently it agrees very closely with material in Richardson and Burns (1975). This figure might be compared to Figure 7-2 illustrating Zubov's formula. The small scale discrepancies between Figures 7-2 and 7-5, and indeed small patterns in the isopleths of Figure 7-5 are probably because of local conditions. Inspection of the annual reports on which Figure 7-5 is based would suggest that even at stations in the northern archipelago ice measuring sites were usually chosen near shore in areas which were open water in summer. This would mean that the ice thicknesses in Figure 7-5 are pre-dominantly representative of first year ice near shore. In certain restricted areas, such as persistant polynyas, in which open water persists for some time in winter it may be possible for greater thicknesses of ice to form in a season, (Sadler, 1976) although in this situation this rapidly forming ice would have to be removed continuously or at frequent intervals. Although it is not evident from the data source the very high ice thicknesses measured at Indoucdjouac, P.Q. may be due to piling up of this sort. On the other hand, where areas of sea ice persist through the summer sea ice growth in autumn begins not from open water but may be from quasi-continuous ice cover. The only station where this was observed over 1964-1974 was at Isachsen where in three of the years initial and maximum thicknesses were 112 and 254 cm, 251 and 340 cm, and 152 and 267 cm. This is the situation typical of growth in channels where ice cover has persisted through a summer. In areas where ice cover persists annual growth is less than the open water values in Figure 7-5.

Despite exploration in Canadian arctic waters since the 16th century, quantitative data on ice cover and thicknesses in Canadian Arctic waters in a convenient form are rather hard to come by. A list of references is given in Bradford and Smirle (1970). Compilations by Swithenbank (1960), 7-3 Freeze-over and break-up, (a) Mean dates of first appearance of ice in lakes and bays, (b) Mean dates of freeze-over of lakes and bays, (c) Mean dates of initial break-up of ice on lakes and bays, (d) Mean dates of clearing of ice from lakes and bays. Also shown are mean dates on which air temperatures pass through freezing, (dotted curves), (after Allen, 1964).

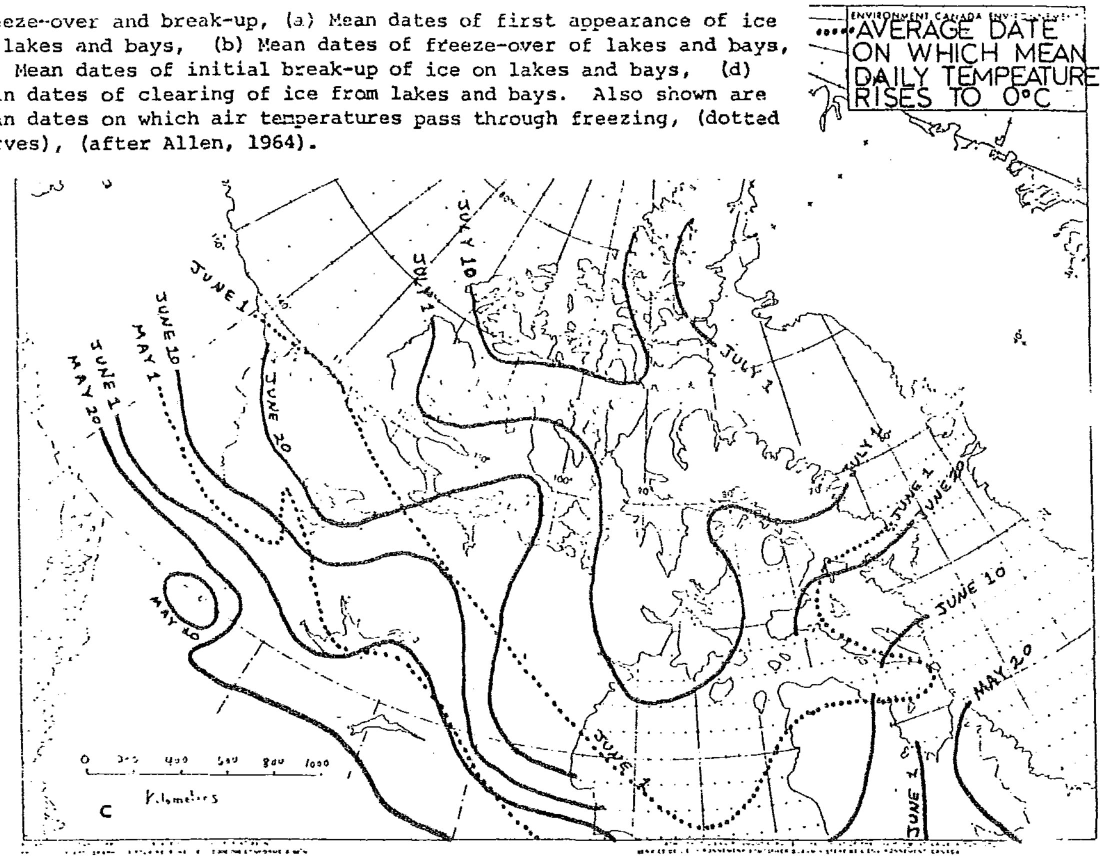


MEAN DATES OF THE FIRST APPEARANCE OF ICE IN LAKES, BAYS, ETC.

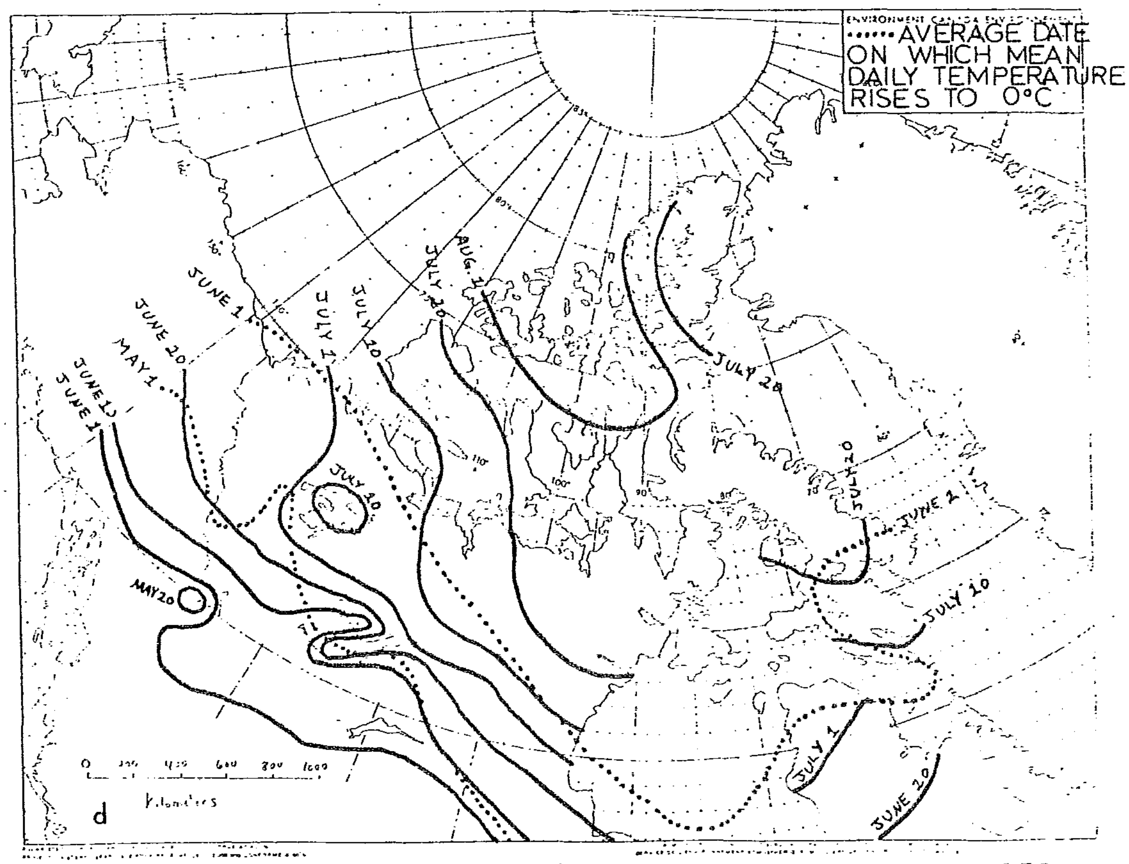


MEAN DATES OF THE FREEZE-OVER OF LAKES, BAYS, ETC.

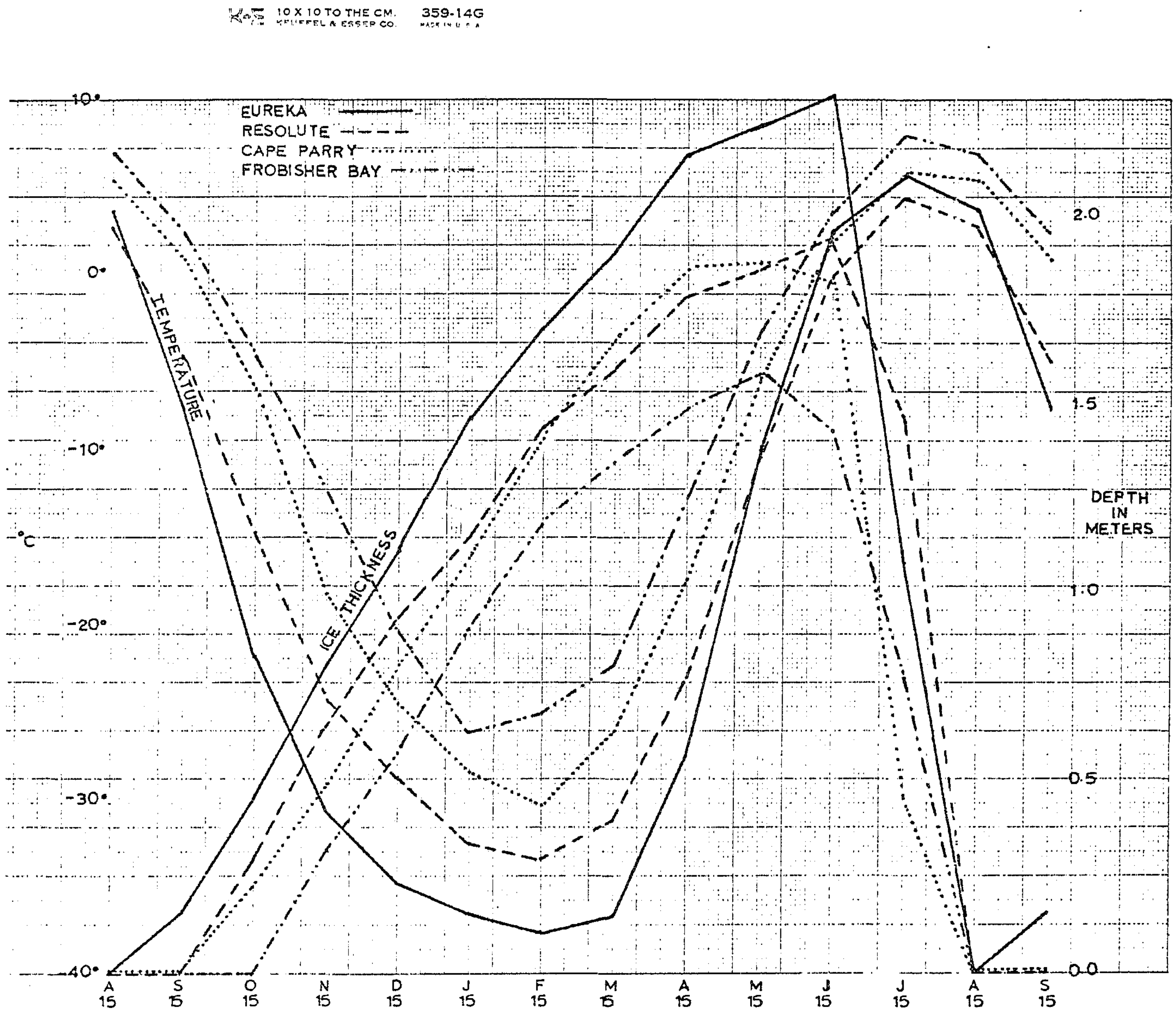
7-3 Freeze-over and break-up, (a) Mean dates of first appearance of ice in lakes and bays, (b) Mean dates of freeze-over of lakes and bays, (c) Mean dates of initial break-up of ice on lakes and bays, (d) Mean dates of clearing of ice from lakes and bays. Also shown are mean dates on which air temperatures pass through freezing, (dotted curves), (after Allen, 1964).



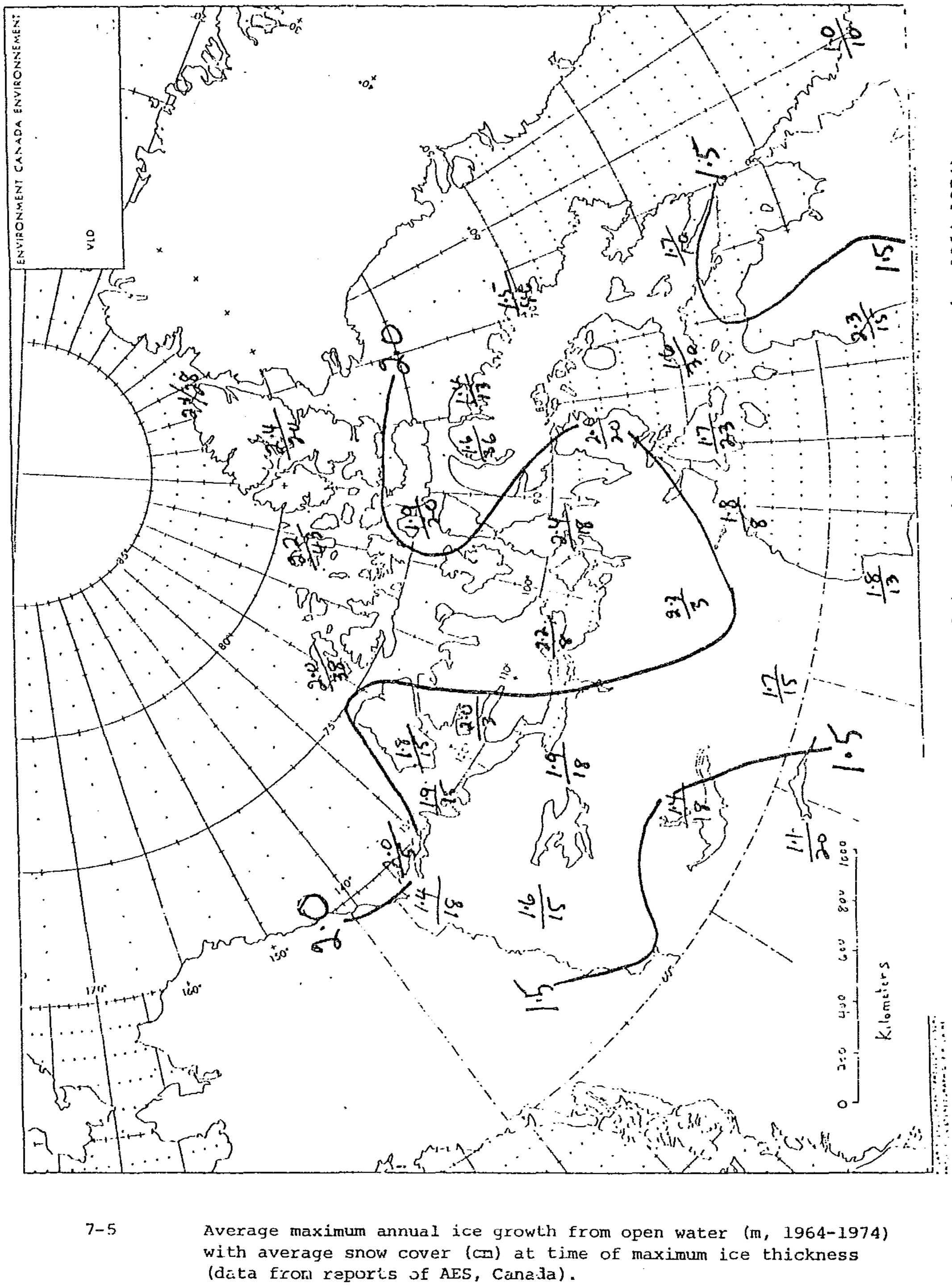
MEAN DATES OF THE INITIAL BREAKING OF ICE ON LAKES, BAYS, ETC.



MEAN DATES OF THE CLEARING OF ICE FROM LAKES, BAYS, ETC.



7-4 Annual course of air temperature (°C, 1941-1970) and ice thickness (m, after Richardson and Burns, 1975).



7-5 Average maximum annual ice growth from open water (m, 1964-1974) with average snow cover (cm) at time of maximum ice thickness (data from reports of AES, Canada).

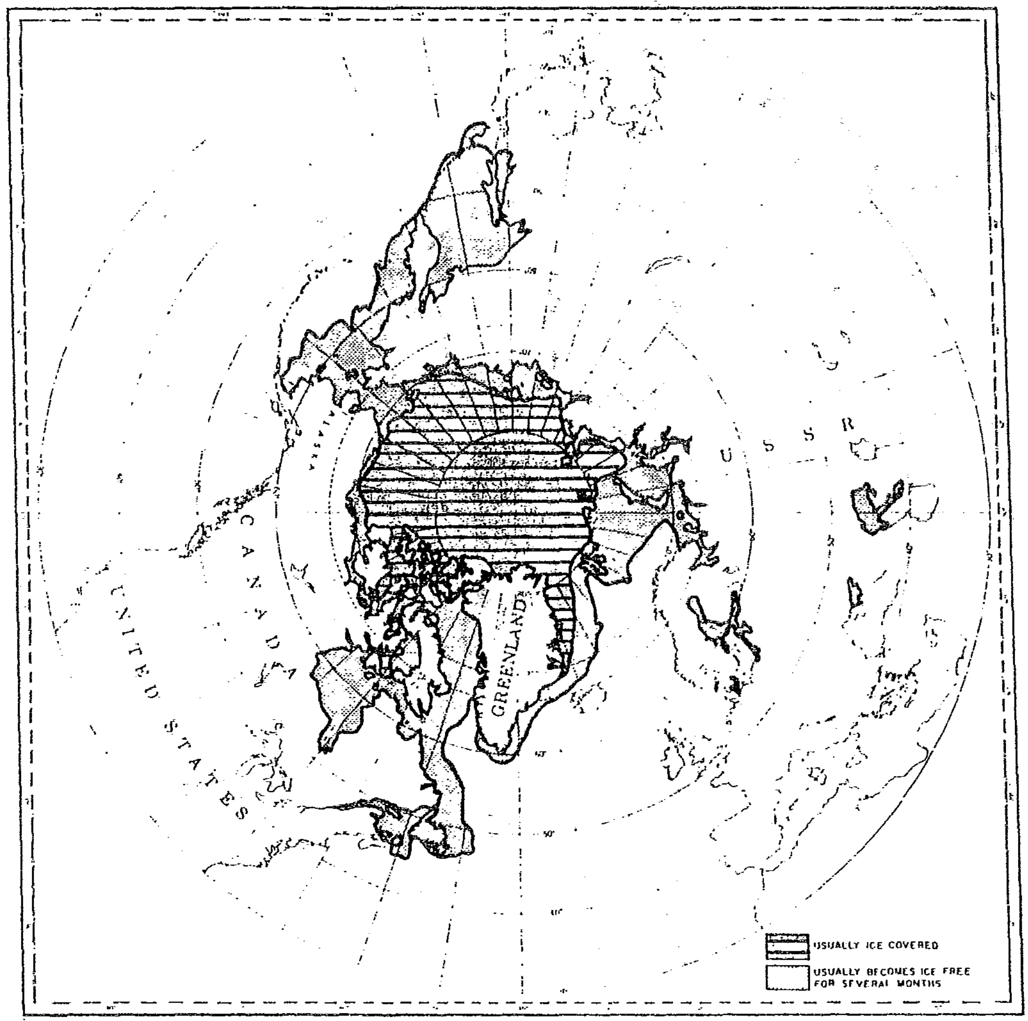
7-5 Average maximum annual ice growth from open water (m, 1964-1974) with average snow cover (cm) at time of maximum ice thickness (data from reports of AES, Canada).

Anon (1958) are fairly good on sea ice coverage as it affects ship movement. Sea ice surveys by the Canadian Polar Continental Shelf Project and surveys by the Canadian Ice Central of the Atmospheric Environmental Service are perhaps of broader coverage than Swithenbank's and the U.S. Navy's atlases.

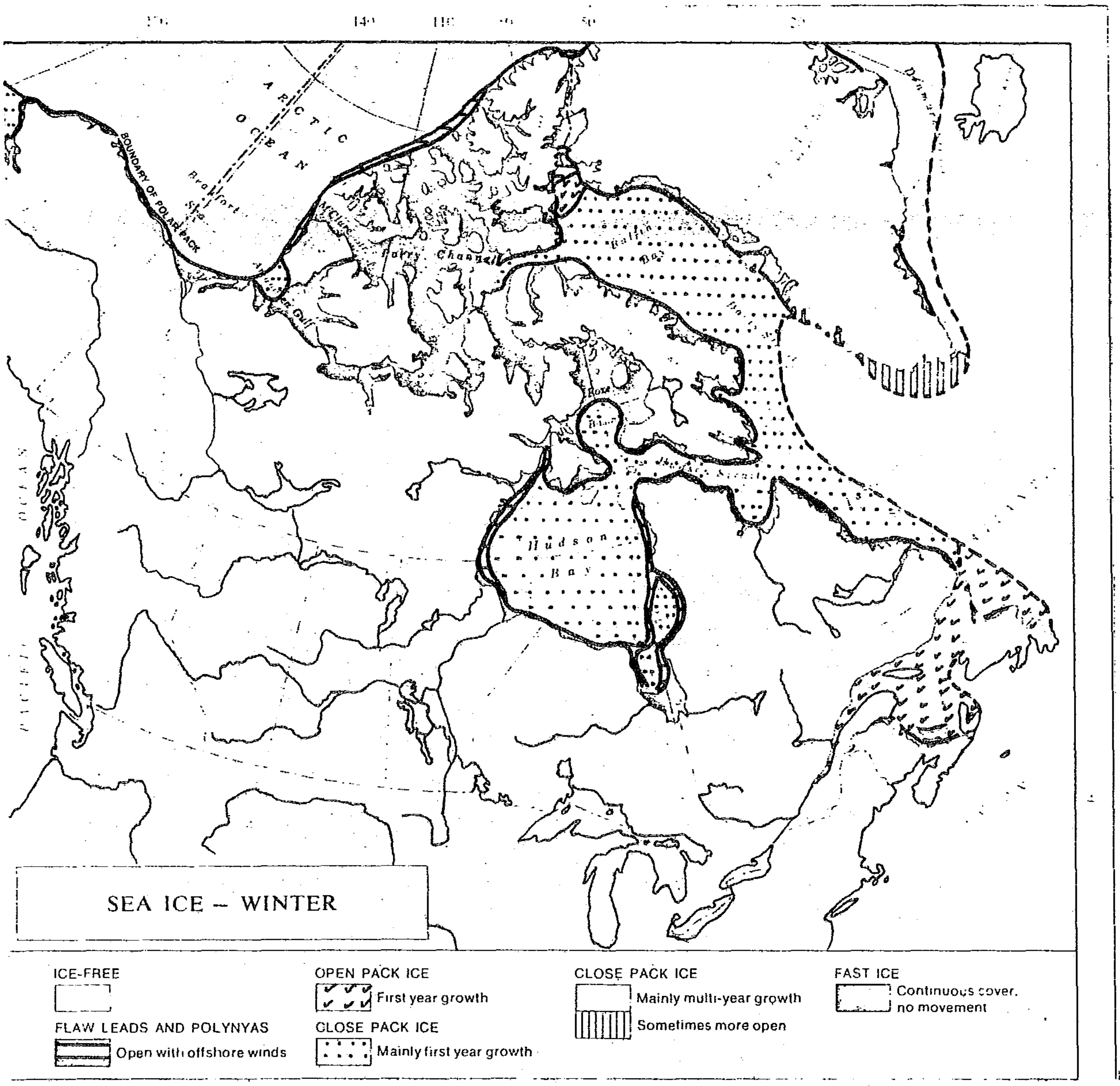
Although not strictly pertinent the extent of sea ice in the northern hemisphere is shown in Figure 7-6 (after Sater et al, 1971). In winter only Canada's Pacific coast and the southern Atlantic coast are free of sea ice. The details of clearing of this sea ice and reformation in autumn is perhaps best seen for the arctic archipelago in the handsome atlas produced by the Polar Continental Shelf Project, Canada (Lindsay 1976) and in the annual reports by the Canadian Ice Central of the Atmospheric Environment Service. The detail in these atlases is far too great to reproduce here. In Figure 7-7 are schematic drawings of winter and summer ice conditions in the Canadian Arctic Waters after the Atlas of Canada (Anon, 1973). Summer conditions vary particularly in the northwestern archipelago from one year to another. Shown crudely is the summer sea ice coverage in a 'good' ice year and a 'poor' ice year. In Parry Channel and the Queen Elizabeth Island channels Lindsay (1968) estimated minimum coverage of 32% for 1962, 67% for 1966, and 72% for 1964. Broad estimates of percentage ice coverage for the archipelago, and surrounding waters is shown in Table 7-1, for each month. Lindsay (1968) notes the importance of recurring polynyas as centres for early breakup. In the northern part of the archipelago these polynyas occur at 78°N, 75°W, North Water, at 76°N, 90°W, Cardigan-Hell Gate, at 77°N, 96°W, Penny Strait, at 76°N, 95°W, Queen's Channel, in Lancaster Sound, and along the west coast of the archipelago when contact is made between land-fast ice of the archipelago and the polar pack.

From the atlases and other references listed above some indication can be gained of the types of ice to be found in the archipelago. Ice forming from open water will reach a thickness of about 2 m in a winters growth. As noted earlier undisturbed ice in the Arctic Ocean will reach winter thicknesses of about 3 m. Ridging can of course greatly increase the thickness of undisturbed sea ice. A study of such ridging for the Beaufort Sea has been reported by Wadhams (1975). Polar ice would be expected to enter the archipelago from the north and west. Ice islands, pieces of ice shelf up to 30 m thick and believed to originate in northern Ellesmere Island might be expected to follow the same path. Icebergs from glaciers in Greenland and in the eastern archipelago are found in quantity in the Baffin Bay area, although drifting into passages leading from Baffin Bay.

Quantative budgets of mass of ice formed, melted in the archipelago or exported have been attempted by Lindsay (1968), Barber and Huyer (1971), Barber et al (1975), Huyer and Barber (1970), Collin (1963). Errors in the analyses resulted partly from lack of firm data on ice coverage, but probably more seriously from lack of knowledge of the thickness of that ice cover. Hopefully this lack may be improved with increased use of satellite data, plus more surface surveys. Extensive use of ice thickness measurements from a vehicle or aircraft have not yet been possible. When these measurements are available then a graduate student, in oceanography or in geography, could combine the Atmospheric Environment Service measurements of annual ice growth with areas of old ice and model of sea ice growth under observed meteorology and oceanography to obtain a best estimate of sea ice mass formed annually. A check on the calculations would be possible against import to and export from the archipelago.

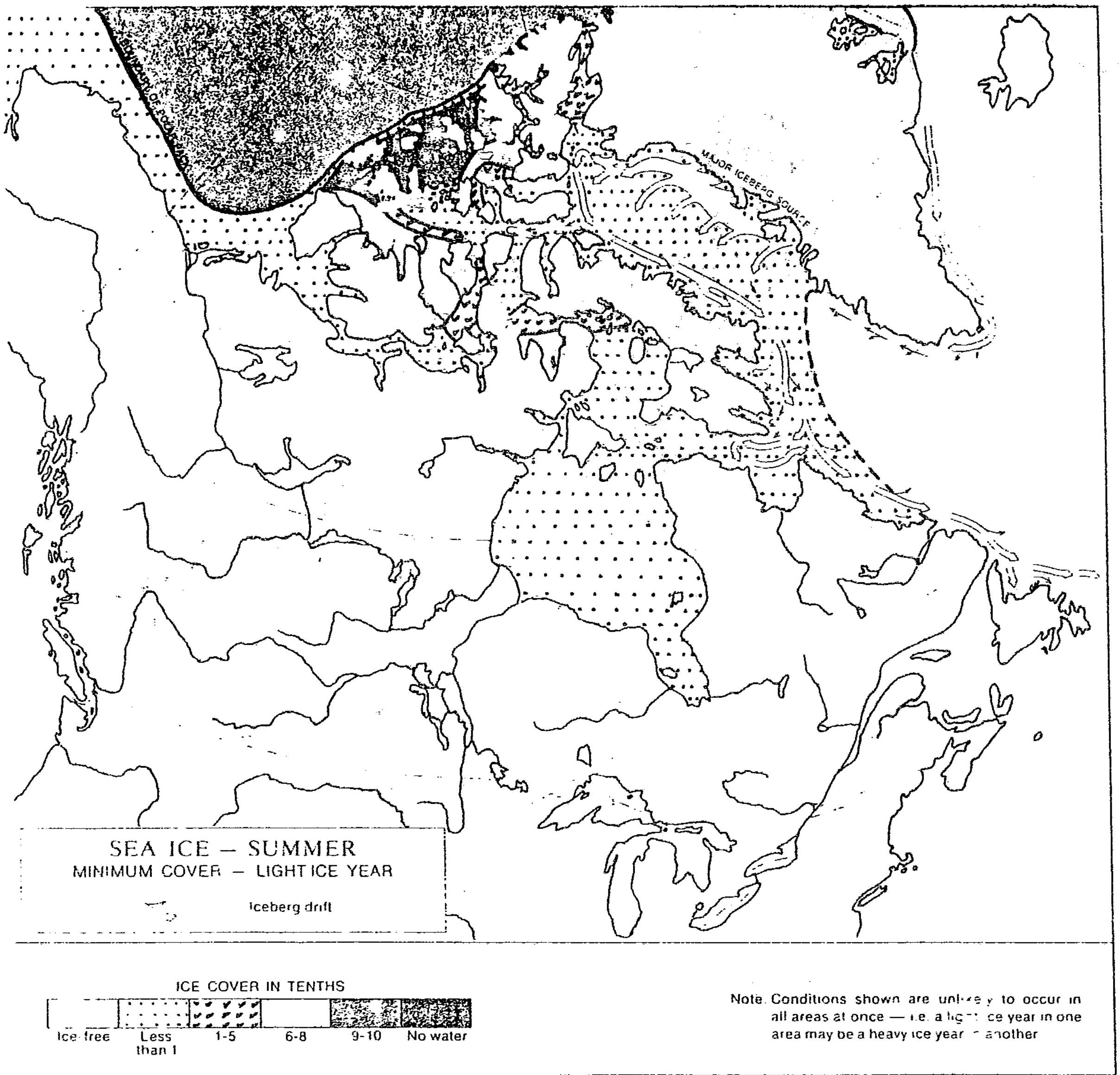


7-6 The maximum and minimum extent of sea ice in the northern hemisphere (after Sater et al, 1971).

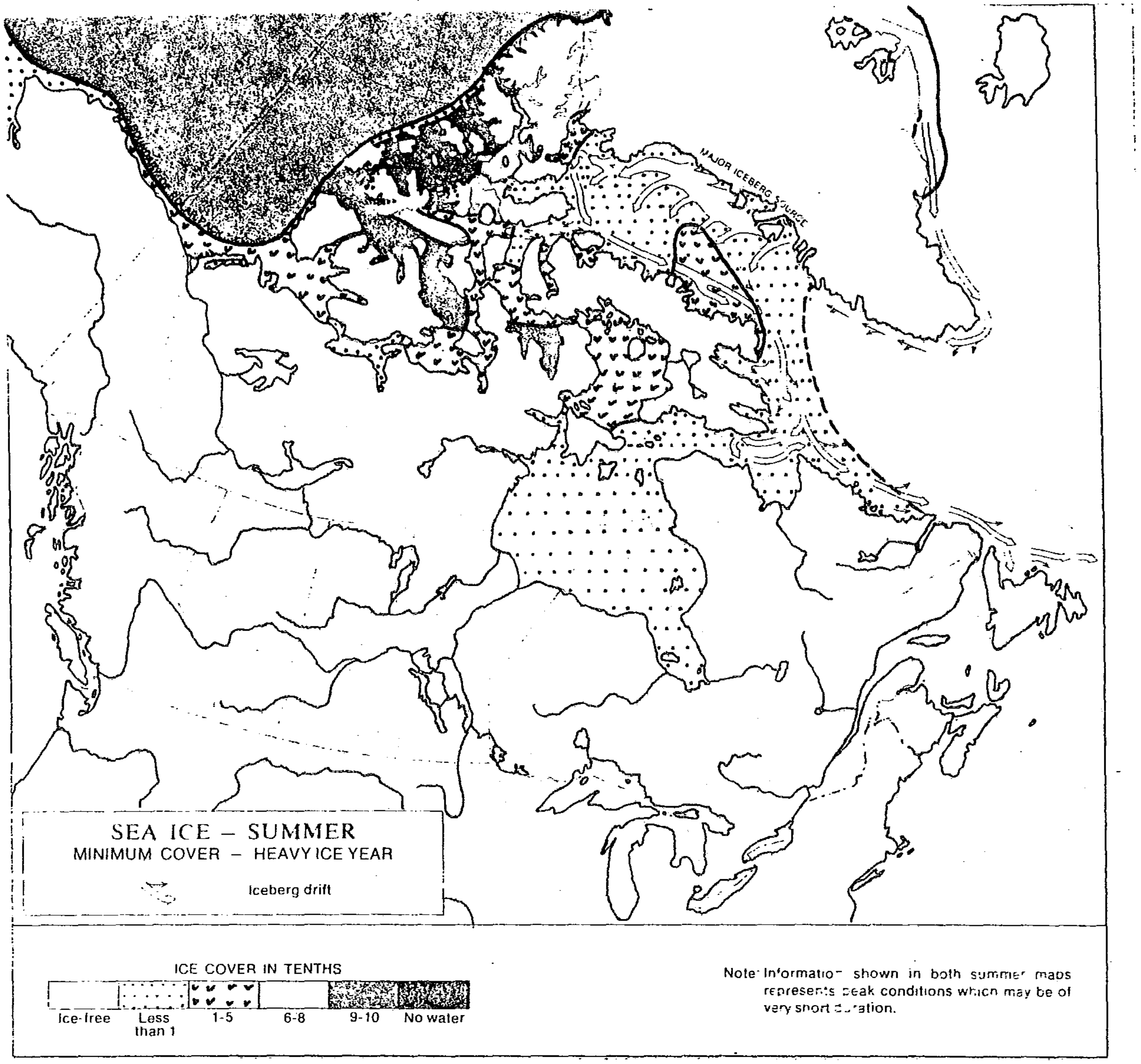


7-7 Schematic of ice conditions, Canadian waters

(a) winter ice conditions



7-7 (b) summer ice conditions in a light ice year,



7-7 (c) summer ice conditions in a heavy ice year, (after Anon, 1973).

TABLE 7-1

	J	F	M	A	M	J	J	A	S	O	N	D
Beaufort Sea	97	97	97	97	97	91	94	63	63	89	97	97
Canadian Archipelago	97	97	97	97	96	95	93	79	84	93	97	97
Davis Strait-Baffin Bay	97	97	97	96	93	87	68	34	16	75	87	94
Central Polar Ocean	99	99	99	99	99	99	97	97	96	97	99	99

Percentage ice cover in the Canadian Archipelago and nearby waters by months (after Sater et al, 1971).

The forces acting upon ice to cause movement as noted by Campbell (1968) are shown in Table 7-2. In the central Arctic Ocean Basin where movement is not subject to such restrictive boundary conditions as are imposed by the narrow channels of the archipelago ice movement has been the subject of many studies as noted in Table 7-2. Both gross ice movements and surface currents in the Arctic Ocean are largely a result of wind patterns (Hunkins 1966). Complications enter from the internal stresses within the ice cover (Campbell 1965), correct formulations of which are still under investigation. The wind effects on ice movement are important in the archipelago without doubt, and the internal stress of ice effect is probably more effective in the narrow channels of the archipelago then in the Arctic Ocean. It may be speculated that in the channels of the archipelago the component of ice movement due to stress imposed on the ice bottom by currents caused by the large scale oceanic circulations may be relatively more important than in the Central Polar Basin. Such features as the West Greenland Current, the Canadian Current and the Labrador Current, are observed to have appreciable transport of sea ice (and icebergs) semi-independent of current wind fields. The same effect is observed in tidal currents in straits (Hell Gate) and in the estuarial currents in an isolated fiord (Lake and Walker, 1976), and in Parry Channel (Marko, personal communication).

Observations of ice movements, mostly in the summer season by mariners, have been made on a local basis for many years, but have not, so far as I am aware, been systematically collected. Apart from these many scattered observations the most extensive ice movement observations are probably those made by the Polar Continental Shelf Project in the Queen Elizabeth Islands since 1961 and summarized in the atlas issued by that organization (Lindsay, 1976). To summarize further, sea ice in summer shows a movement towards the southwest off the northwest coast of the archipelago a southward flow through Nansen Sound - Eureka Sound - Norwegian Bay, a southern flow through Sverdrup Channel - Peary Channel - Massey Sound - Norwegian Bay - Hell Gate, through Peary Channel-Hassel Sound-Penny Strait-Wellington Channel/

TABLE 7-2

	Wind stress	Water stress	Coriolis force	Internal ice stress	Gradient currents	Boundary layer
Nansen	X	X	X			
Sverdrup	X		X	X		
Rossby and Montgomery	X	X				X
	X	X	X	X		X
	X		X	X		X
Shuleikin	X	X	X			X
Felzenbaum	X	X	X		X	
Ruzin	X	X	X	X		
Reed and Campbell	X	X	X			X
Campbell	X	X	X	X	X	X

Forces acting on ice sheets and considered in the various models of ice movement (after Campbell, 1968).

McDougall Sound, and through Prince Gustaf-Adolf Sea - Maclean Strait - Byam Martin Channel. These ice movements are accompanied by melting and deformation of the ice sheets of course but they give some indication of surface water movements in summer. A typical summary for the northwestern archipelago, for the year 1962 is shown in Figure 7-8 (after Black, 1965). The observations of Sadler (1976) indicate a prevailing movement to the south through Nares Strait. The southward flowing current and ice movement off the east coast of Baffin Island has been known for many years (Muench, 1971). In winter movement of sea ice is probably restricted to tens or hundreds of meters over most of the archipelago.

The growth and decay of sea ice have an effect on the underlying water structure. When sea water of appreciable salinity freezes, the sea salts and gases dissolved in the water are largely rejected by the ice although some saline water is trapped as shown in the brine pockets between growing ice crystals. In the laboratory Foster (1968) indicated that rejected brine is released in a rather uniform manner and the convection thus induced is amenable to treatment based on Bénard - Ellsworth theory. In unpublished laboratory work J.A. Elliott of the Frozen Sea Research Group found that brine was released from sea ice in irregular streamers from holes in the

7-8 Ice circulation, June to September, 1962 in the northwestern archipelago (Black, 1965).

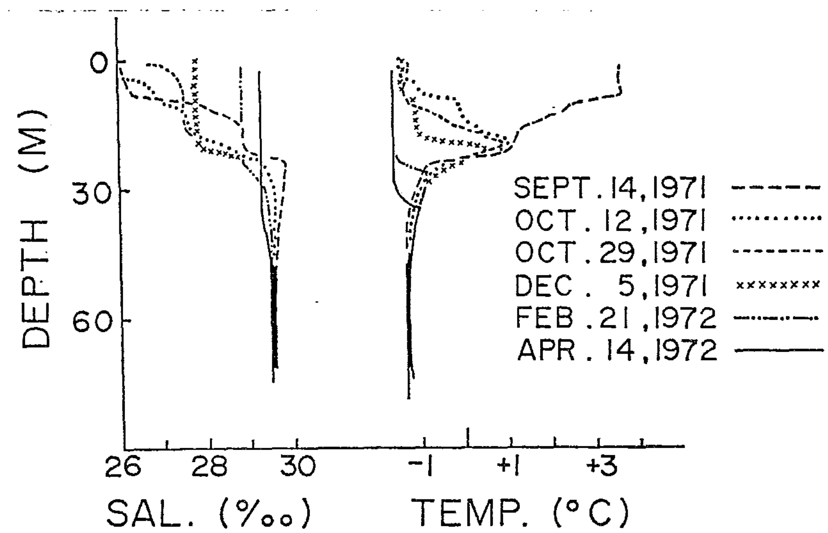
ice bottom. The difference between Foster's and Elliott's work remains unresolved. Eide and Martin (1975) under laboratory conditions have found pumping of sea water in and out of tubes extending high into the ice.

Investigating release of brine in the growth of natural sea ice, Lake and Lewis (1970) described the structure of a block of sea ice taken in March 1968 from Cambridge Bay. In addition to large tubes (filled with ice crystals?) extending well up into the ice (about 2 m thick) the bottom of the ice was covered with small holes or tubes. The release of salt from the ice bottom was considered by Lake and Lewis (op cit) to recur in random fashion from these small tubes, This would imply that most of the salt rejected by the growing sea ice would, on any but the very smallest scale, occur uniformly over the ice bottom. Larger scale features found in natural sea ice would seem superficially to correspond to the tubes of Martin's laboratory work. Stalactites have been observed in the Arctic (H. Welch, personal communication) mainly in areas of thin ice indicating a greater than usual temperature gradient in the ice.

Although all details of the processes of salt release by growing sea ice are not known, the result is usually convective overturn in the water layer below the growing sea ice. The mechanics of this convective overturn have never been fully observed in the Arctic, but the gross features have been observed, understood and forecast in terms of simple one dimensional models which balance potential energy of the released salt by changes in the potential energy of the underlying water structure, (Zubov, 1943; Solomon, 1973). The depth of mixing is dependent upon the pre-existing density structure. In the central Arctic Ocean the convection under growing sea ice can reach 50 m or more in some circumstances. In the Canadian Arctic Archipelago where more runoff water can collect the convective depths vary but tend on the whole to be less than in the Central Arctic Ocean. In many of the channels the convective process under growing sea ice can not really be regarded as one dimensional.

In Cambridge Bay, N.W.T. the winter time convective mixing depth has been reported by Gade et al (1974) to reach almost to the bottom (82) m of the bay: (Figure 7-9). In d'Iberville Fiord, on the other hand, the under ice layer in which we believe the convective processes dominate have been observed to be from 20 cm to 15 m deep. As may be noted from sea water structures profiled, the water in the under-ice layer in which the convective processes dominate is characteristically almost isothermal and very close to the freezing point. The layer is almost isohaline although increased salinities in the lower levels are sometimes observed to be slightly higher than near to the ice bottom. The convective layer usually has a rather sharp increase in salinity and temperature at its bottom. Matter under the ice will tend to be spread through this convective layer if it is not too buoyant. Observations in d'Iberville Fiord (Lake and Walker, 1973) have suggested interleaving of layers of saltier water in the convective layer just above its bottom, hypothesized as resulting from disturbances on the interface with denser water beneath.

The decay of sea ice adds very low salinity water to the surface layers, to add to any fresh layer previously laid down by run off from snow the ice surface and nearby land. The depth of these fresh layers has



7-9 Increasing depth of the convective layer under growing sea ice in Cambridge Bay, N.W.T. from September 14, 1971 to April 14, 1972 (after Gade et al, 1974).

been remarked upon earlier but is of the order of a meter or two in thickness. In conditions of low wind these fresh layers can persist on the surface for some time (or under any persisting sea ice). In windy weather the fresh layer can be mixed down so that when autumn comes the process repeats, matter which is in the freshwater tending to be distributed to some depth. Estimates of the energetics of these processes are given below in section 10.

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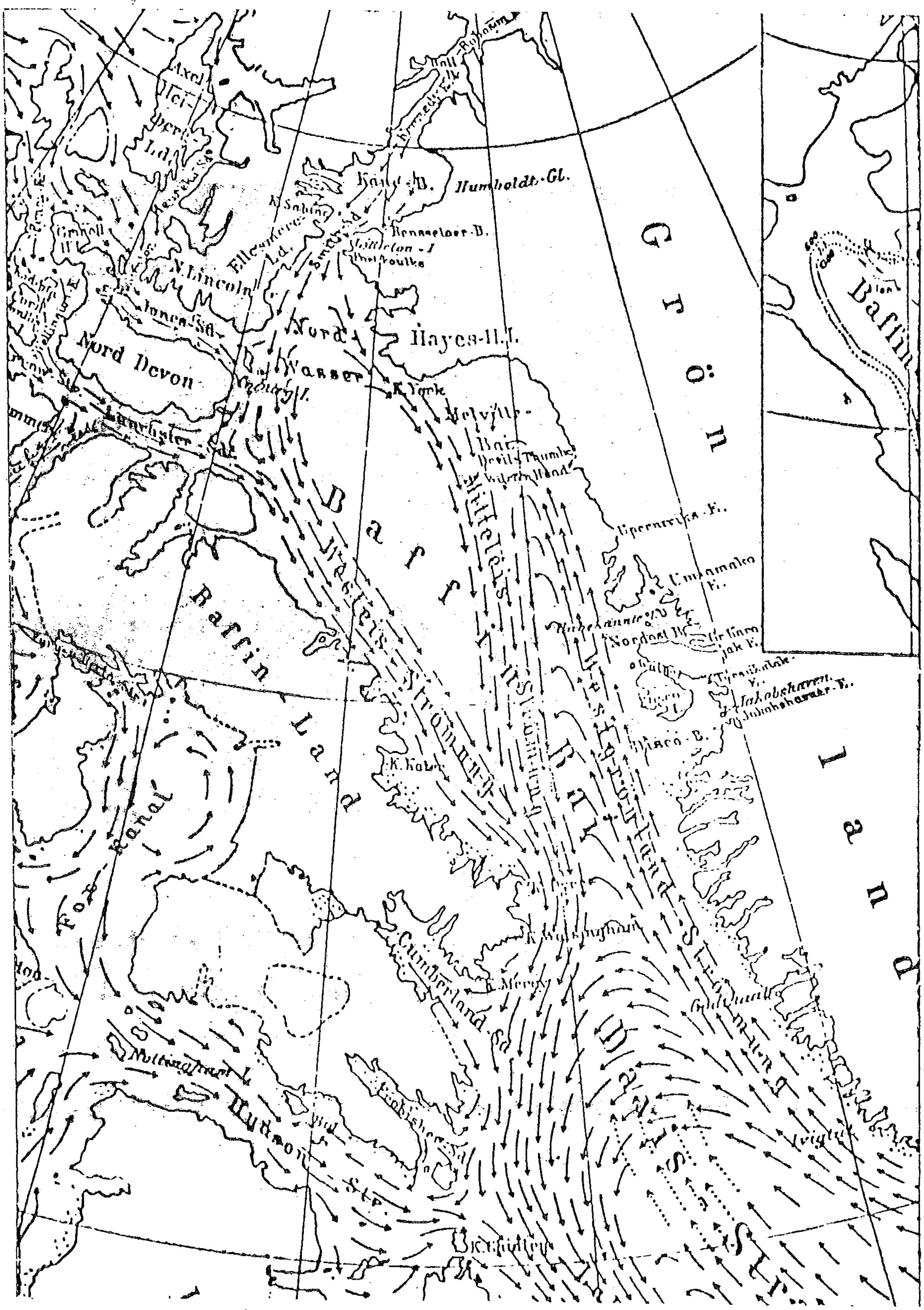
CURRENTS

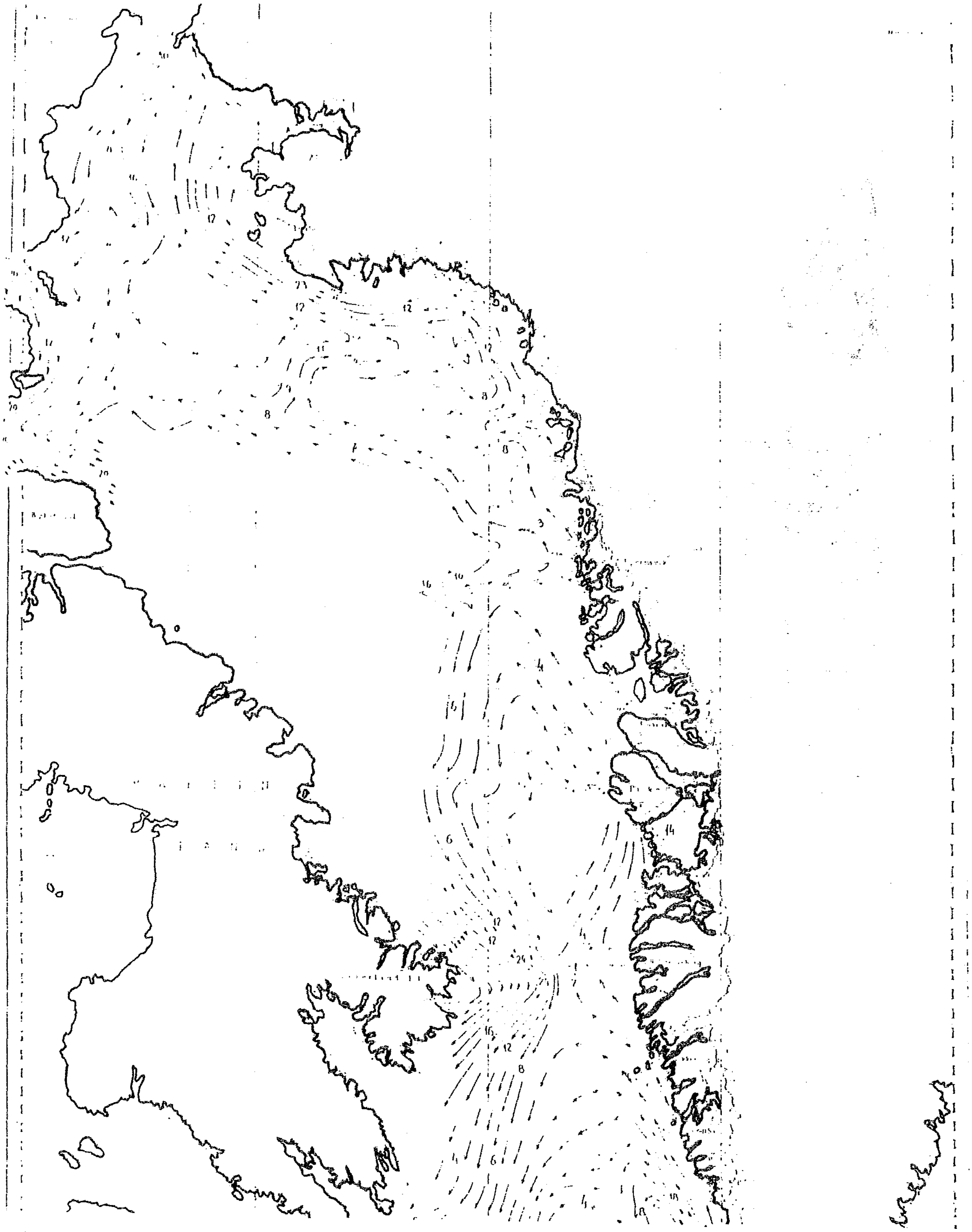
The conservation equations in Appendix A reveal that water parcels may be accelerated by various forces. Many horizontal motions can be considered approximately geostrophic, the pressure gradient forces balanced by the Coriolis forces. The scale in time and space of current systems vary widely. Tidal currents are tied to tide periods and many to fairly small topographic features. Those caused by wind stress, and the pressure-caused sea surface variations are mainly tied to weather system periods of a few days although shorter and longer term effects may be discernable in a sufficiently precise analysis. The magnitude or even reality of currents which may be caused by long term variation of sea level (section 5) or the variation of inflow to the Arctic Ocean (section 2) are really unknown, and current or sea level data within the archipelago seem inadequate to investigate this question. As to spatial scales the synoptic systems span radii of 10^3km, the longer term variations perhaps occur on a somewhat larger scale.

Currents may be measured in a variety of ways. Most straight forward is use of the usual current meter measuring a point value of velocity. Other measurements (probably less reliable) can measure integrated currents across a channel by changes of electric potential in a wire cable across the channel caused by currents in water of the channel (Robinson, 1976). Geostrophic current shear normal to a cross section of a channel is estimated by measuring spatial differences in water densities. The procedure gives current patterns normal to the section chosen and relative to velocity values measured by current meters or relative to a level in which the velocity might be insignificant. There are difficulties with all of these methods in the archipelago. Ice cover makes year-round data collection difficult. This explains why current data in the archipelago consists mainly of ship derived data in summer, or spring data obtained through sea ice. Other data on surface currents may be gained from drift in summer usually of ice (Lindsay, 1969) or other debris.

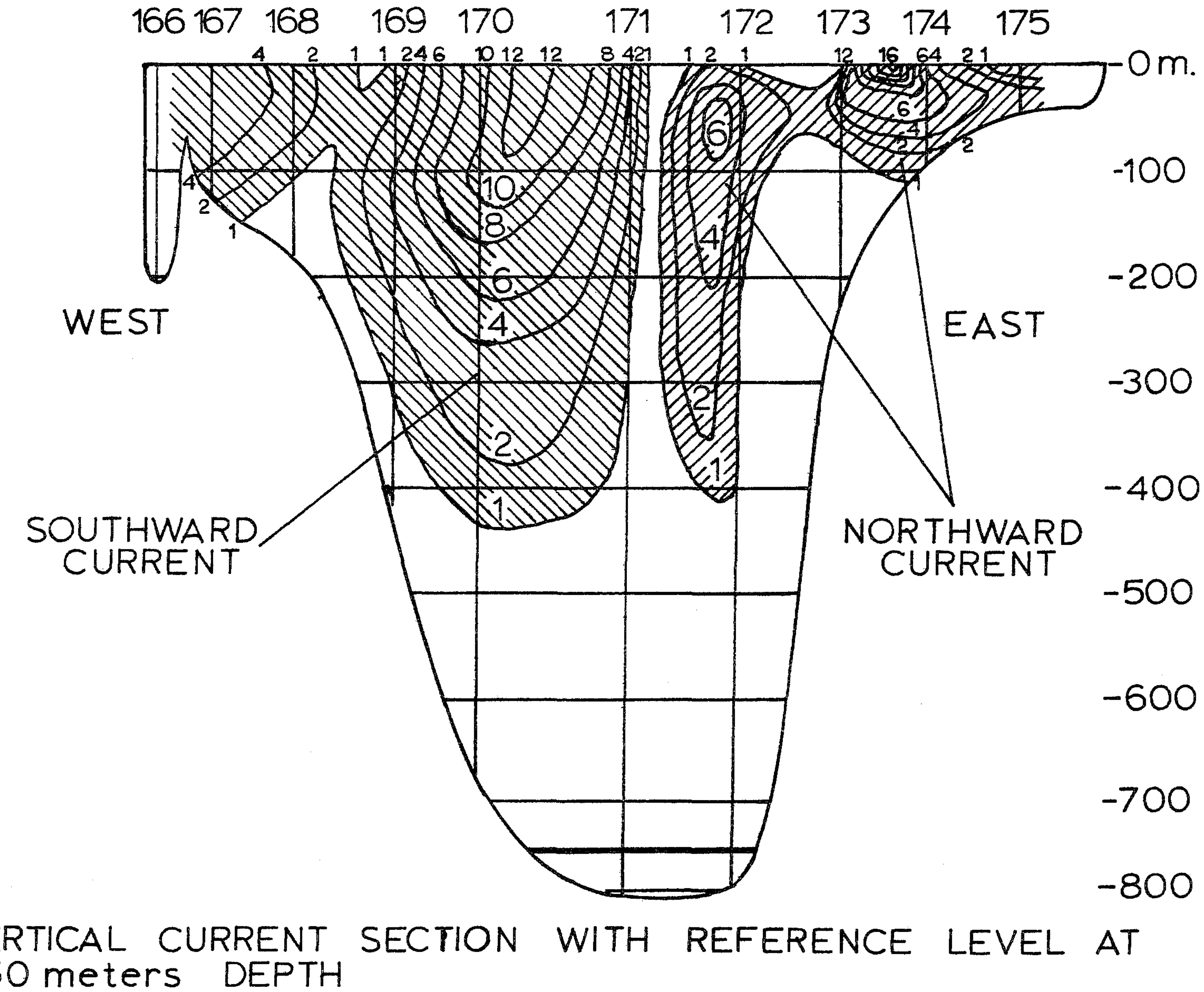
The circulation in the Canadian Arctic Archipelago is generally considered to be from the Arctic Ocean towards the south and east. The basis consists of inferences from observations of surface water flow by mariners and explorers, Taylor (1955), Collin (1958), Muench (1971). The current pattern in Figure 8-1 (after Mecking (1906) quoted by Dunbar et al,(1967) is in error in Baffin Bay but elsewhere in the archipelago similar to present ideas. The current structure in Baffin Bay in Figure 8-2 is from Kiilerich (1939) as are those in the cross section in Figure 8-3 showing geostrophic currents relative to 750 m through the Davis Strait cross section whose position is shown in Figure 8-4. Data from the extended cruise of HMCS Labrador in 1954 from Baffin Bay through Parry Channel, Prince of Wales Strait to the Beaufort Sea allowed Bailey (1957) to obtain currents through numerous cross sections. Bailey summarized his analysis of currents from sections in the diagram shown in Figure 8-5. Herlinveaux (1974) summarized his current measurements as shown in Figure 8-6. Muench (1971) investigated currents in northern Baffin Bay, Smith Sound, Lancaster Sound and Jones Sound, using recent oceanographic material. An example of his results is given in Figure 8-7 showing the rather complicated and variable current picture in the upper layers of Lancaster Sound.



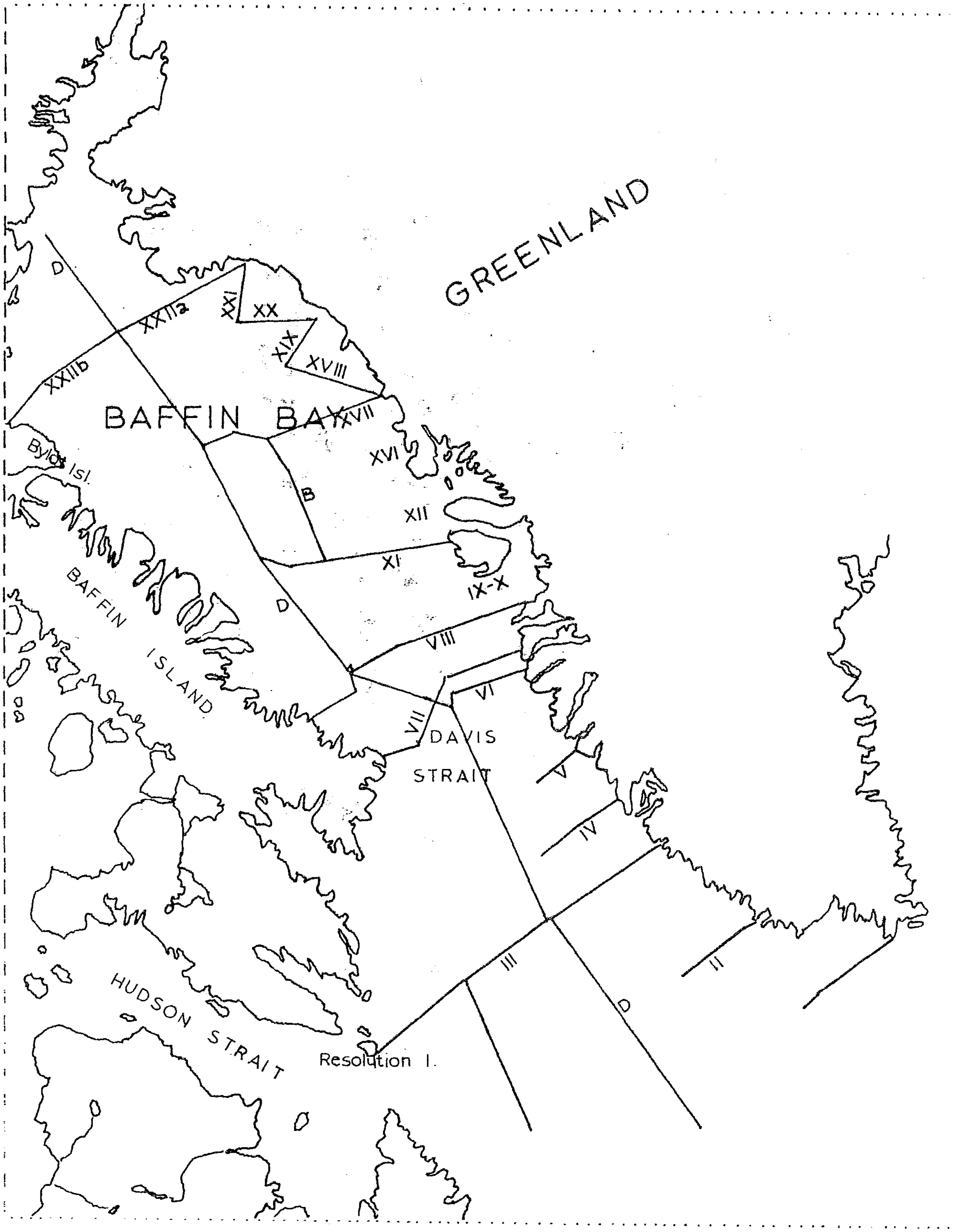
8-1 Currents of Baffin Bay and the eastern archipelago after Mecking (1906) quoted by Dunbar et al (1967).



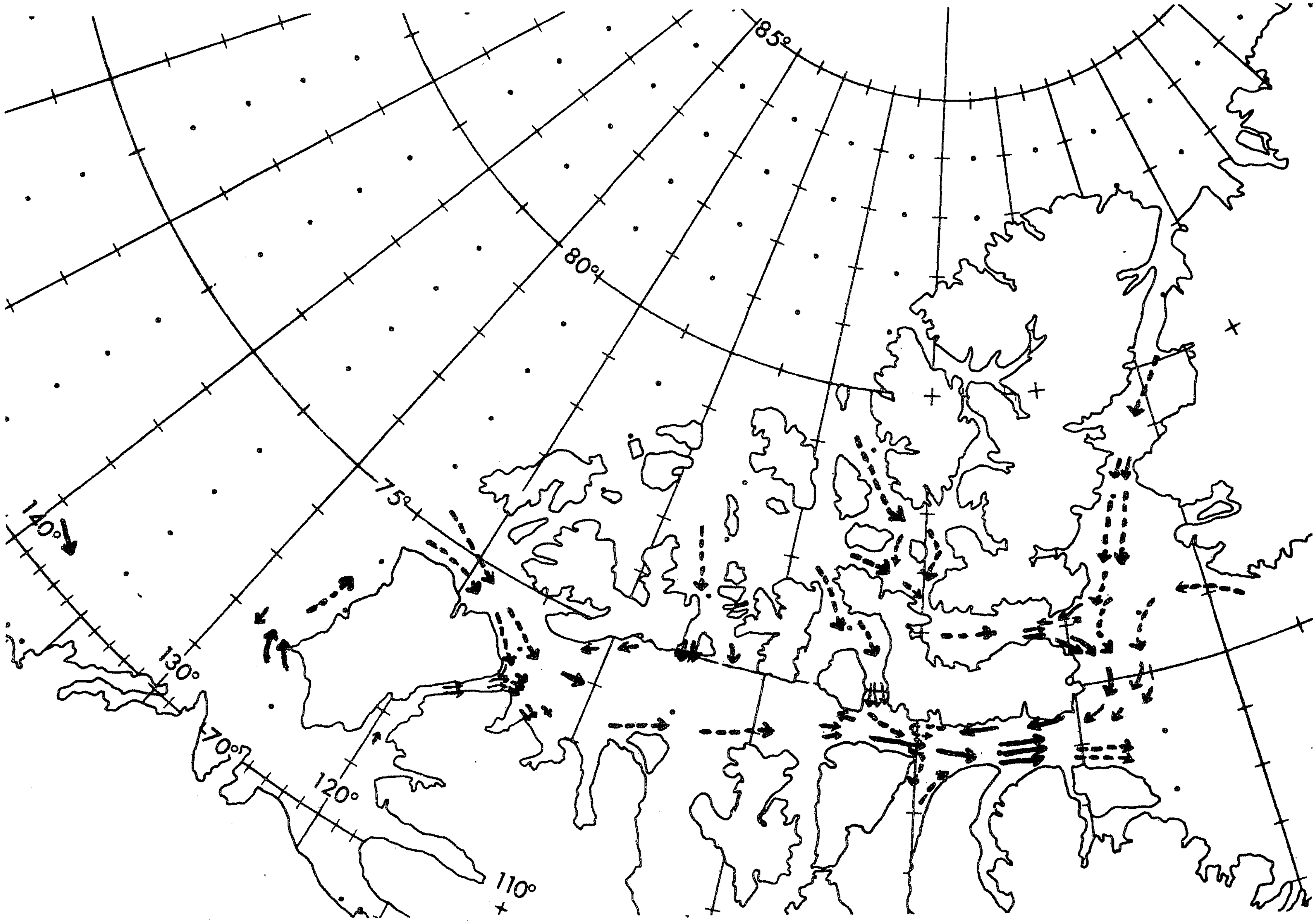
8-2 Surface water currents in Baffin Bay in May - October, 1928, obtained by computations from the cross sections shown in Figure 8-4 and using a reference level of 1500 m (after Kiilerich, 1939 [Plate 5]).



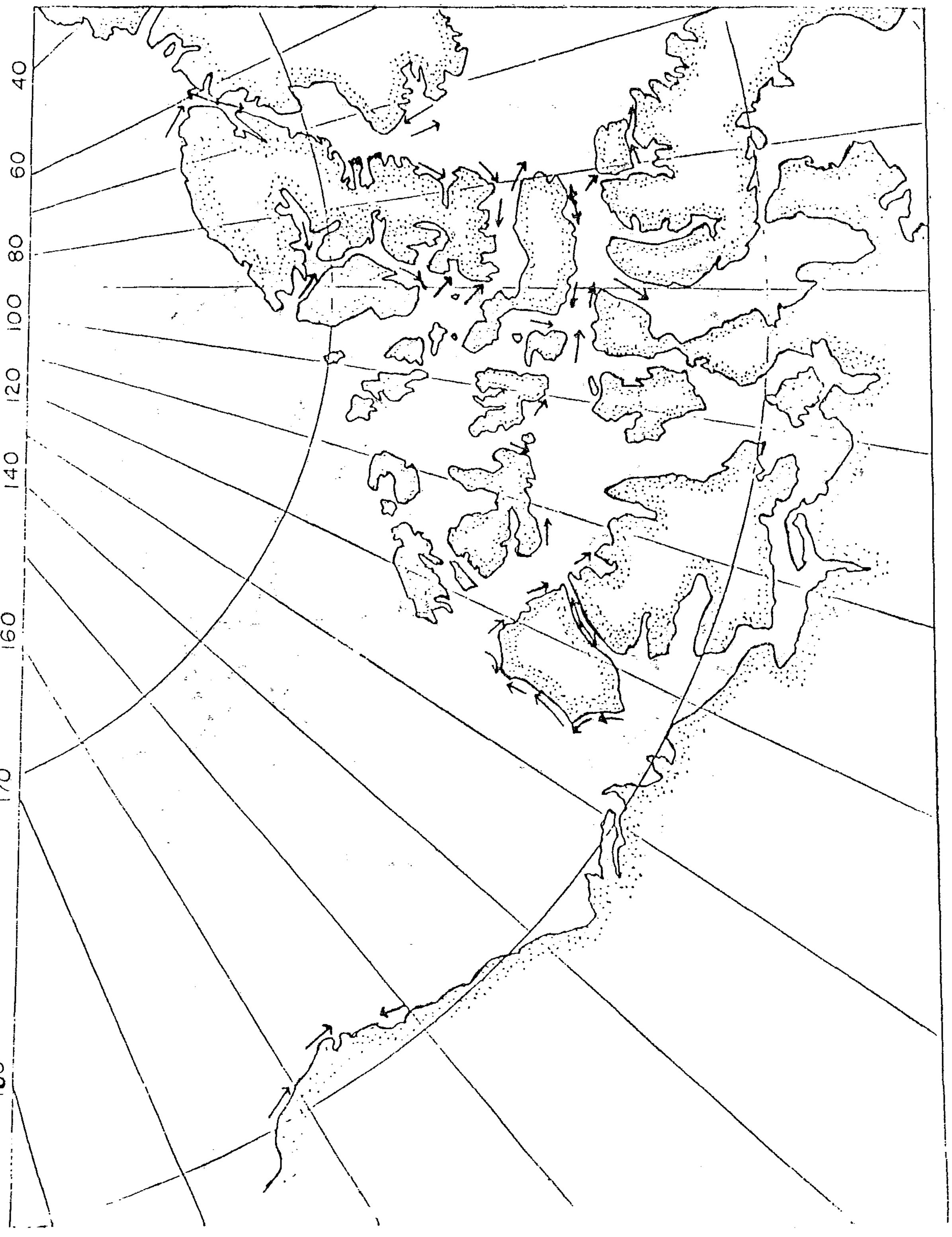
Current relative to 750 m computed for cross section VIII (in Figure 8-4) on the Godthaab expedition in September, 1928.



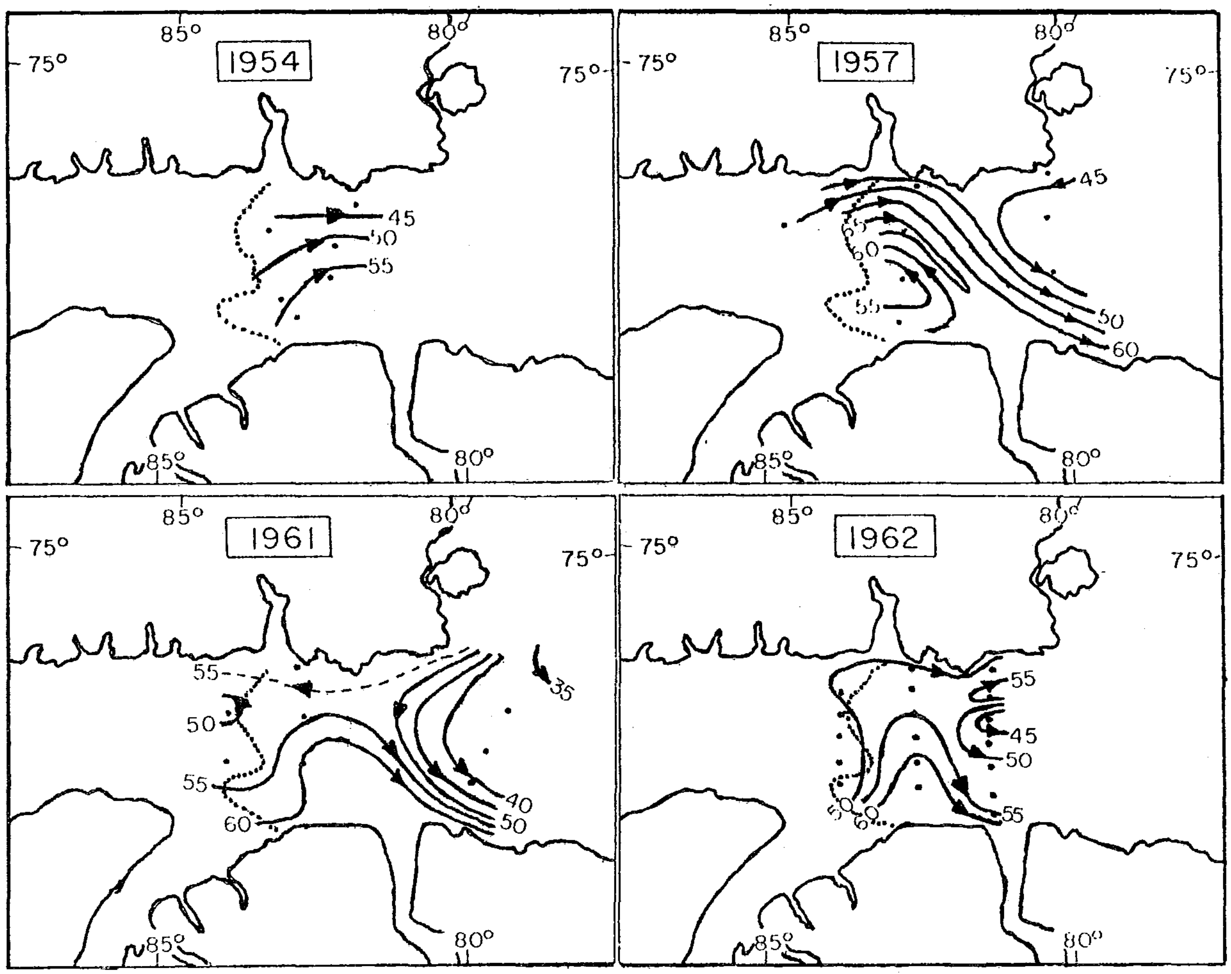
8-4 Oceanographic stations and sections in Baffin Bay on the Godthaab expedition May - October, 1928 (after Kiilerich, 1939).



8-5 Water movements in the Canadian Archipelago in summer 1954 as derived from the density distribution (·→) and as inferred from these movements (--→) after Bailey (1957).



8-6 Composite of all surface water movements observed during summer operations 1954-1973, (after Herlinveaux, 1974).



8-7 Dynamic topography of the water surface in Lancaster Sound relative to 500 db with a contour interval of 5 dynamic cm and inferred flow. Dotted line is the 650 m isobath, (after Muench, 1971).

Muench also measured currents in northern Baffin Bay in 1968. Muench's (1971) analysis corroborates the circulation pattern described by Kiilerich (Figure 8-2).

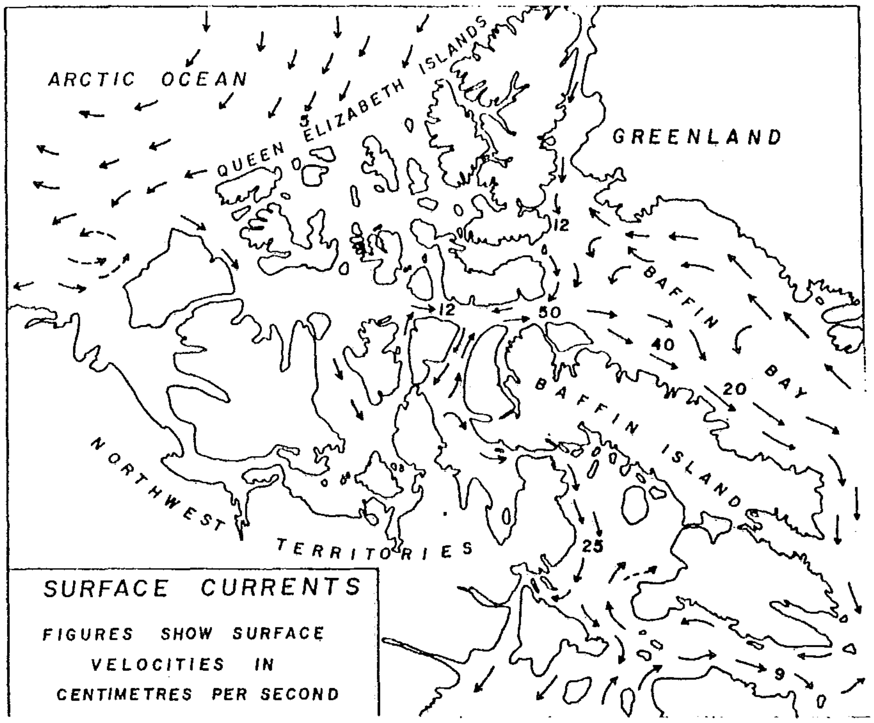
Sadler (1976) has investigated currents in Nares Strait both from cross-sections and by actual current measurements in the spring of 1972. Sadler used his measurements to estimate water transport through Nares Strait. Some estimates of transport through channels in the eastern archipelago are given in Table 8-1.

TABLE 8-1

Author	Channel	Average annual flow in $10^4 km^3$
Kiilerich (1939)	Davis Strait	4.6
Collin (1960)	Archipelago	4.0
Mosby (1963)	Archipelago	3.5
Muench (1971)	Northern Baffin Bay	6.5
Sadler (1976)	Davis Strait	6.3
Kiilerich (1939)	Smith Sound	1.4
Kiilerich (1939)	Jones, Lancaster Sound	3.0
Bailey (1957)	Lancaster Sound	4.7
Bailey (1957)	Smith Sound	-1.3 (northward)
Bailey (1957)	Jones Sound	-1.2 (westward)
Collin (1963)	Lancaster, 1957	3.1
Collin (1963)	Jones, 1957	.8
Sadler (1976)	Smith Sound	2.1

Estimate of average annual transport southeastward through various channels in the eastern archipelago.

The estimates of current patterns up to 1962 were summarized by Collin (1963) as shown in Figure 8-8. This may be compared with the current estimates in other figures in this section and such indications of ice movement as are shown in Figure 7-8. It is I think obvious that current data in the archipelago needs to be extended in coverage, both in space and time. Instrumental innovations such as year-round moorings of current meters with directional capability in low intensity magnetic fields, combined with year round moorings of temperature and salinity chains are needed for adequate coverage at reasonable cost.



8-8 Surface water currents in the archipelago summarized by Collin (1963) from Kiilerich 1939, Dunbar 1951, Campbell and Collin 1956, Campbell 1958, and Timofeyev, 1960.

WATER MASS STRUCTURE IN THE ARCHIPELAGO

Many oceanographic data are being gathered in the archipelago but rather few analyses have been made in recent years. Apart from the Beaufort Sea Studies the recent studies of Sadler (1976) in Nares Strait and Muench (1971) in Baffin Bay, and work of Barber (1967, 1972), Barber and Huyer (1971), Huyer and Barber (1970), little has been published since the era of Bailey (1957), Collin and Dunbar (1964).

Delineation of water mass characteristics may reveal systematic differences across the archipelago which give some indication of motion and processes within the archipelago. Characteristics of use in this regard include water temperature and salinity as a function of pressure, which are by far the observations most commonly made. Dissolved oxygen observations are available in small quantity. Chemical observations on a broad scale in the archipelago are almost non-existent (Canplan 1976). Biological observations are somewhat more abundant than chemical observations. Derived parameters may be chosen to represent water structure, heat or potential energy content or anomalies, or the freshwater content above some arbitrary standard (Huyer and Barber, 1970). It was decided to use profiles of temperature and salinity plotted as a function of water depth.

Unfortunately, data have been assembled over many years. The question of temporal variability will be dealt with at a later time, since adequate data are not now at hand. Instead a brief survey of temporal variation of such material as was available gave some indication of the size of spatial variations which have to be disregarded because they may be due to temporal variations. For the present, this section upon water mass characteristics will consist of general remarks upon temperature and salinity structure. The data is taken from a variety of sources, mostly the Canadian Oceanographic Data Centre, Environment Canada.

Changes in water column temperature and salinity are governed by the conservation equations given in Appendix A. Apart from changes caused by horizontal and vertical advection the upper layers under an open water surface may be cooled (generally) by losses at the surface caused by the long wave radiation balance, or caused by losses of sensible and latent heat, both largest at open water surfaces in the winter season. Heat will continue to be lost through an ice sheet in winter. Water column heating is mainly by solar radiation absorption and by mixing in of warm fresh runoff in summer. Salt will be added to surface layers by evaporation from open water or by sea ice formation while surface layers will be 'freshened' by mixing in of fresh water from melting ice, land runoff and to some extent from direct precipitation. The depth to which mixing proceeds is dependent upon the static stability of the water column as well as the strength of the mixing process. Wind mixing is far stronger than other surface mixing processes. Over most of the archipelago internal mixing by tides and other energy sources should be small. Flow through restricted channels (Hell Gate) greatly enhances such mixing. Flow over shallow sills will impede passage of waters from much below sill depths. As noted above, a series of sills at depths given in Table 3-1 run across the central archipelago from Hell Gate, Penny Strait, and through Barrow Strait.

Contours on the Canada bathymetric chart 800A show connections to the Arctic Ocean below 500 m exist into Amundsen Gulf, M'Clure Strait and Viscount Melville Sound, the Sverdrup Basin, and the Nansen Sound - Greely Fiord areas. Therefore of course, unless there is extensive slopping of water over the sill system, the bottom waters in east and west will be different. The surface waters should be similar except for areas in which the processes affecting surface waters differ. These differences are dominated by the access to the Atlantic through Davis Strait, and the large volumes of runoff in the southwestern part of the archipelago. On a smaller scale local differences in surface layers might be expected, as well as large annual changes. The main characteristic of these important annual changes are shown for Cambridge Bay in Figure 7-9 and below in section 10 for the waters in the Greely Fiord system. These annual changes should be kept in mind because most of the water column profiles below were taken either through ice cover in spring when the under-ice convective layer was near maximum depth, or in summer from ships, where the warming, freshening processes of summer may be marked.

Selected temperature and salinity profiles chosen as typical are listed in Table 9-1. The geographical position is shown in Figure 9-1 (a). In Figure 9-1 (b) are shown characteristics of water along the Arctic Ocean coasts of the archipelago from the Lincoln Sea (1) to the Beaufort Sea. They are of course similar to water masses of the Arctic Ocean described above. They are mainly isohaline from the surface to about 50 m, and near isothermal down to about 100 m. In waters off the Queen Elizabeth Islands (2,3) the small temperature maximum at 50-120 m is a remnant of the Bering Sea water caught in the circulation of the Beaufort (Pacific) gyre. The water column in the Lincoln Sea (1) has a slight salinity maximum at about 50 m, the explanation for which is lacking at the moment. The profile from the southeastern Beaufort Sea (4) is quite different from those along the coast to the northeast. A fresh water layer extends from the surface to more than 10 m. The freshwater excess over 31‰ represents roughly the equivalent of about 2 m of freshwater which might be derived from melting of about 2.5 m of sea ice or from freshwater outflow from the Mackenzie River. The temperature is correspondingly warm representing a combination of relatively warm fresh runoff plus some increased solar input in ice free conditions.

Figure 9-1 (c) shows that water characteristics to the south and east of Victoria Island are very similar to the profile from the southeastern Beaufort Sea (4), all showing summer time warm fresh surface layers due to the enhanced melting of sea ice but more importantly to the larger amounts of warm runoff in these areas. The Cambridge Bay profile in Figure 7-9 comes from this area. As the water depths throughout this area are not great, no deeper water exists.

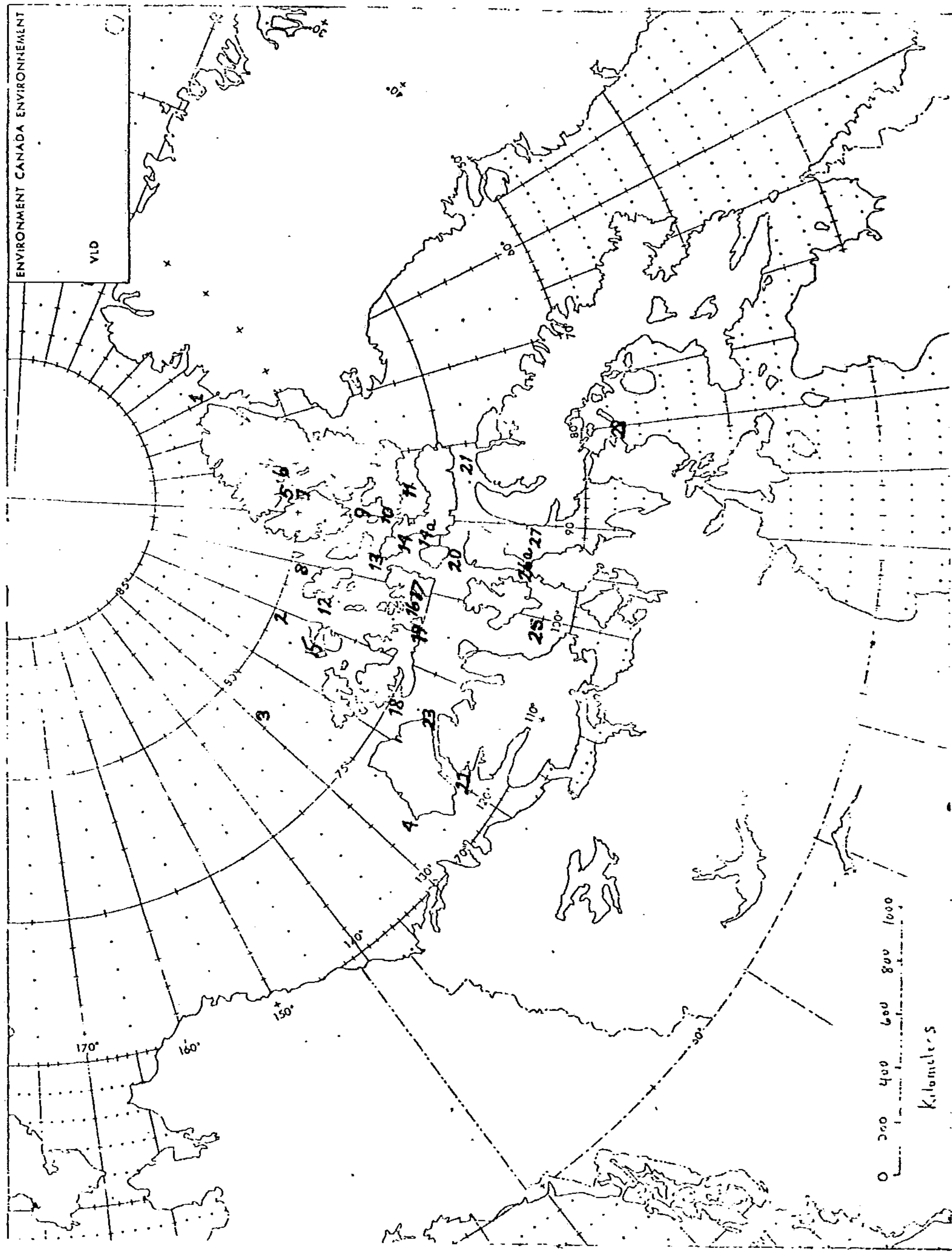
In Figure 9-1 (d) profiles through Parry Channel, from the Beaufort gyre to Lancaster Sound are shown. In M'Clure Strait (18) there is warm fresh water in the surface layers. Considering the limited ice melt in the area, this indicates intrusion of warm fresh runoff into the area probably from the south around Banks Island. Rather less heat and freshwater are found in surface layers in Viscount Melville Sound (19) and Barrow Strait (20). The profile in Lancaster Sound (21) is different from those of Barrow Strait west having temperature maxima at 50 - 100 m. The temperature maximum at

TABLE 9-1

Location - Oceanographic Stations

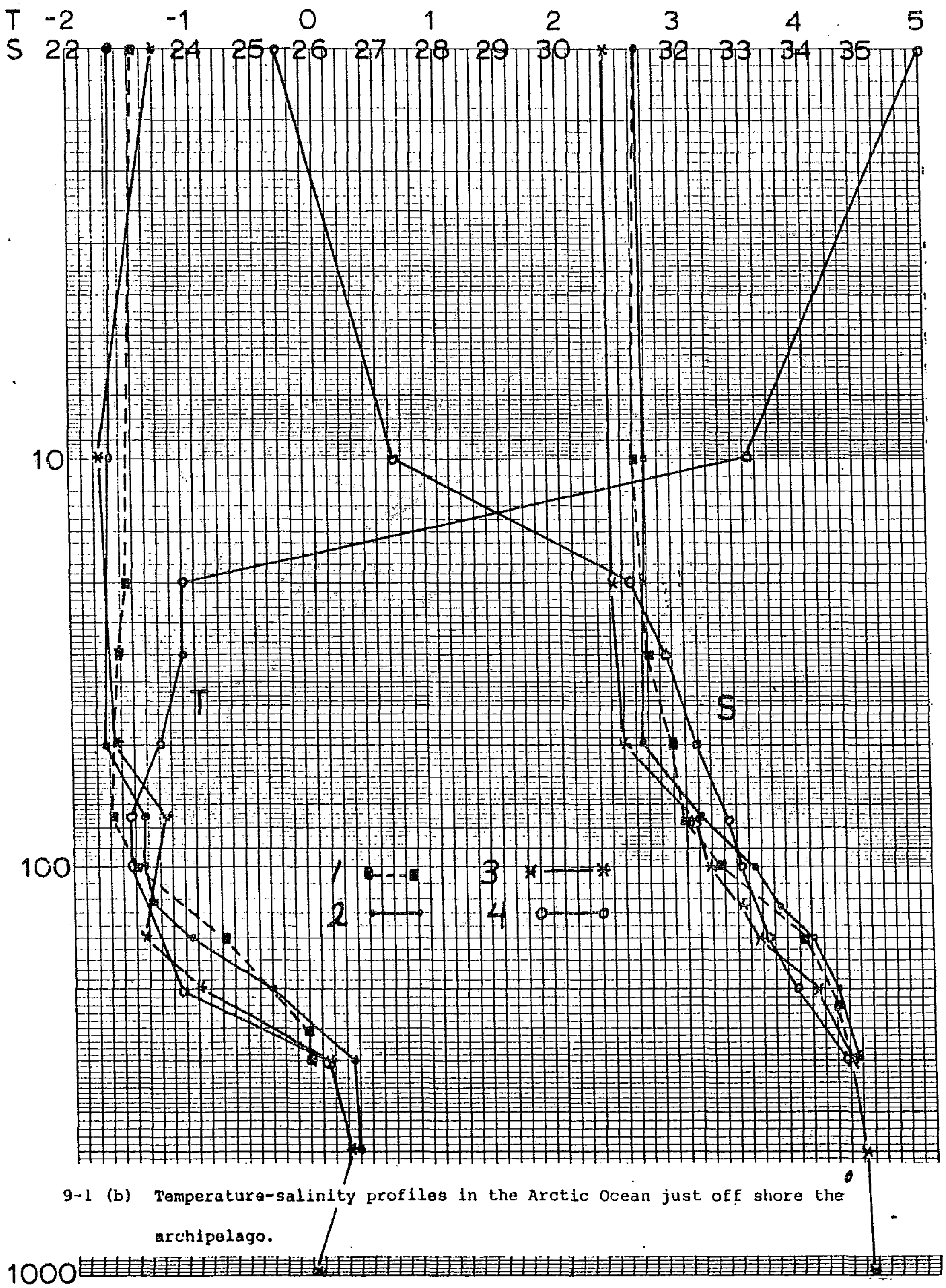
I.D.	Location	Lat.	Long.	Date
1	Lincoln Sea	82°18'N	59°05'W	22/08/71
2	Off Queen Eliz. Is.	79°41'N	111°09'W	07/05/62
3	Gyre	78°12.5'N	129°40'W	27/04/62
4	SE Beaufort Sea	71°58'N	127°21'W	12/09/54
5	Nansen Sound	80°29'N	86°42'W	27/03/76
6	Greely Fiord	80°30'N	81°45'W	23/03/76
7	Eureka Sound	80°00'20"N	86°49'W	30/03/76
8	Peary Channel	79°46.5'N	101°14'W	29/04/60
9	Norwegian Bay	77°05'N	89°38'W	15/08/62
10	Hell Gate	77°00'N	89°46'W	22/08/67
11	Jones Sound	76°50'N	87°00'W	15/08/62
12	Pr. Gustav Adolf Sea	78°28'N	105°16'W	13/04/61
13	Penny Str.	76°40'N	97°50'W	16/09/57
14	Wellington Channel	75°56'N	93°47'W	17/09/57
14a	Wellington Channel	74°47'N	92°43'W	23/04/73
15	Wilkins Strait	78°19'N	114°25'W	23/05/60
16	Byam Channel	75°26'45"N	104°46'32"W	29/04/76
17	Austin Channel	75°30'16"N	103°01'26"W	04/05/76
18	M'Clure Str.	74°28'N	115°58'W	31/08/62
19	V. Melville S.	74°59.5'N	105°26.5'W	04/09/64
20	Barrow Str.	74°15'N	95°30'W	23/08/54
21	Lancaster Sound	74°15'N	82°42'W	03/10/62
22	Amundsen Gulf	71°34'N	119°27'W	06/09/54
23	Pr. of Wales Strait	73°08'N	116°14'W	31/08/54
25	McClintock Channel	71°04'N	100°52'W	05/09/62
26a	Peel Sound	71°56'N	95°32'W	03/09/62
27	Pr. Regent Inlet	71°51'N	91°49'W	08/09/62
28	Foxe Basin	68°53'N	80°05'W	12/09/62

List of temperature-salinity profiles used in Figure 9-1.

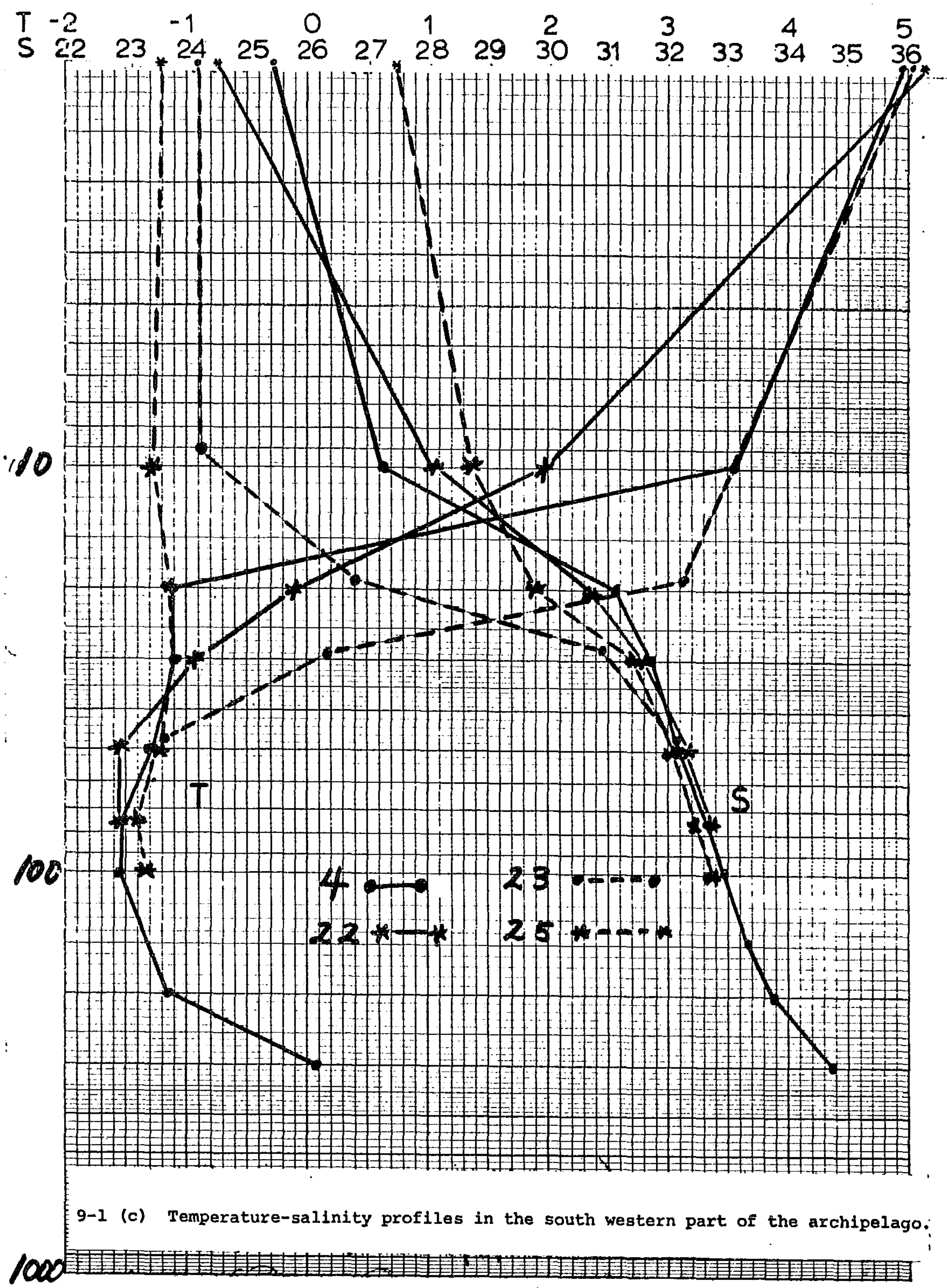


9-1 (a) Geographical location of temperature-salinity profiles in Table 9-1.

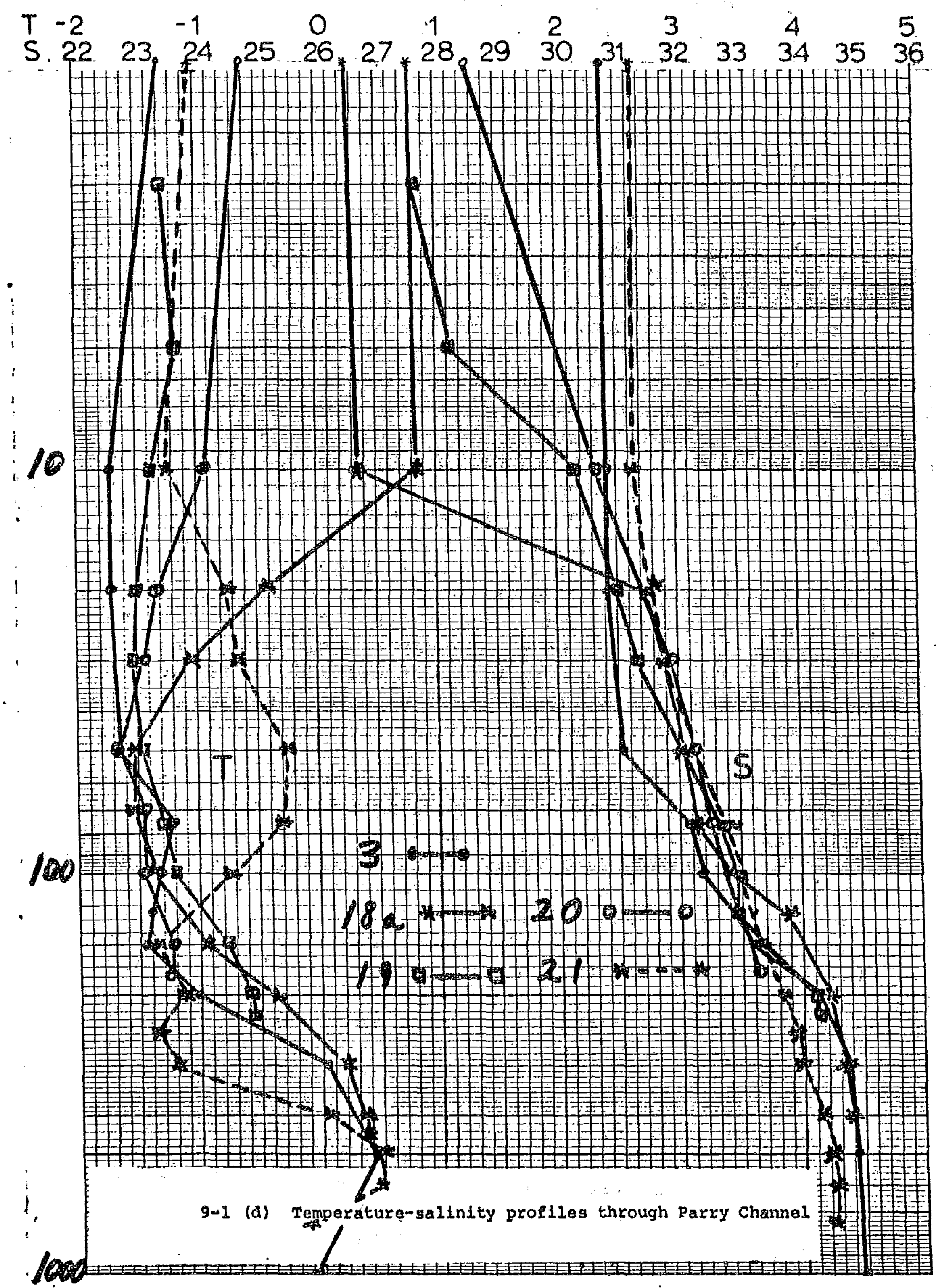
9-1 (a) Geographical location of temperature-salinity profiles in Table 9-1.



9-1 (b) Temperature-salinity profiles in the Arctic Ocean just off shore the archipelago.



9-1 (c) Temperature-salinity profiles in the south western part of the archipelago.



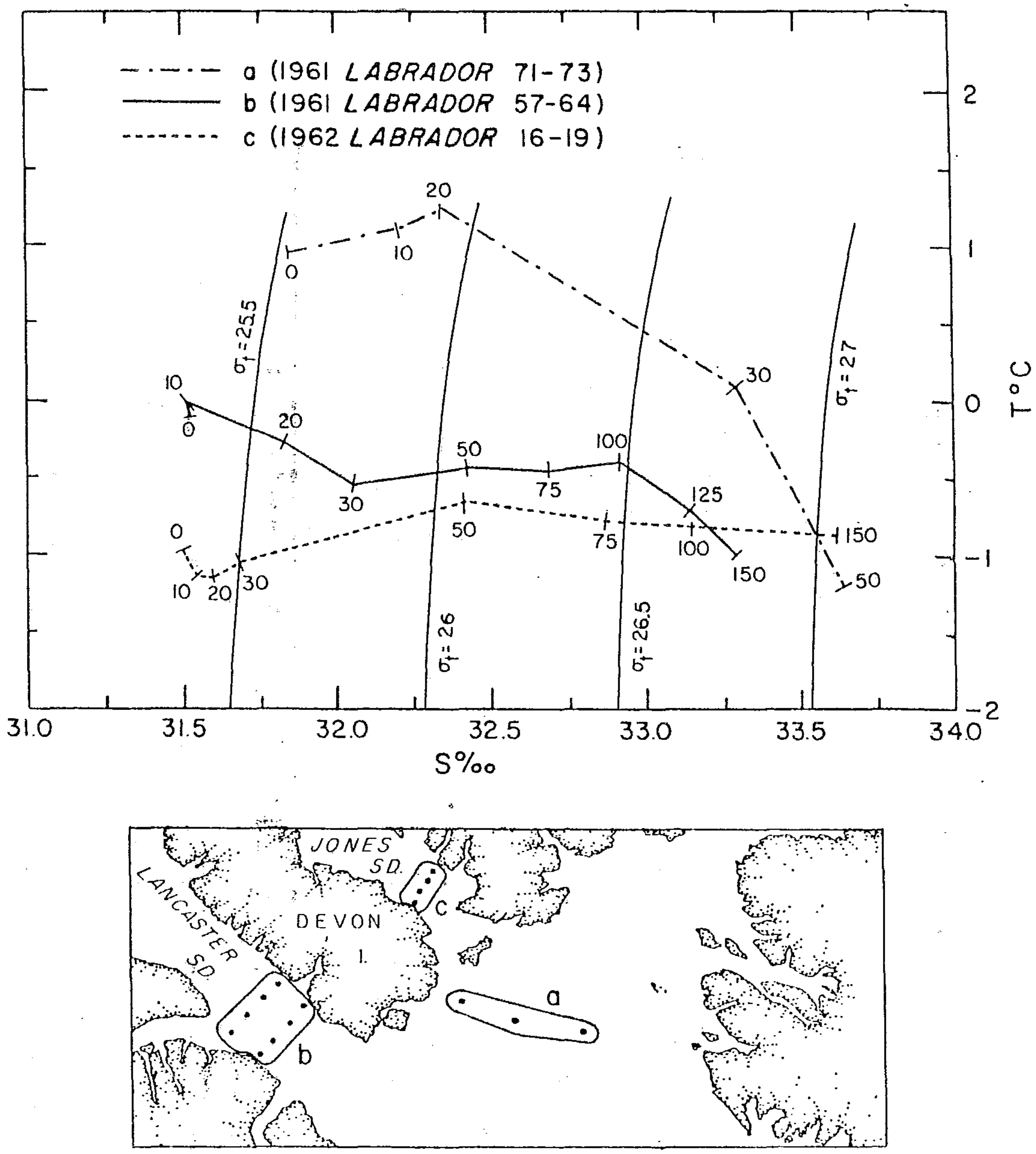
9-1 (d) Temperature-salinity profiles through Parry Channel

50-100 m is believed to come from an area to the northeast of Lancaster Sound in which surface warming frequently occurs in summer. The location of this area, with temperature-salinity curves for the areas are shown in Figure 9-2 (after Muench, 1971). It seems obvious that water in Lancaster Sound below 300 m is different than deeper water west of Barrow Strait and neither can be derived directly from the other profile by vertical motion. To sum up, water characteristics from M'Clure Strait to Barrow Strait are closely related to waters to the west, while in Lancaster Sound are found evidences of mixing from the east.

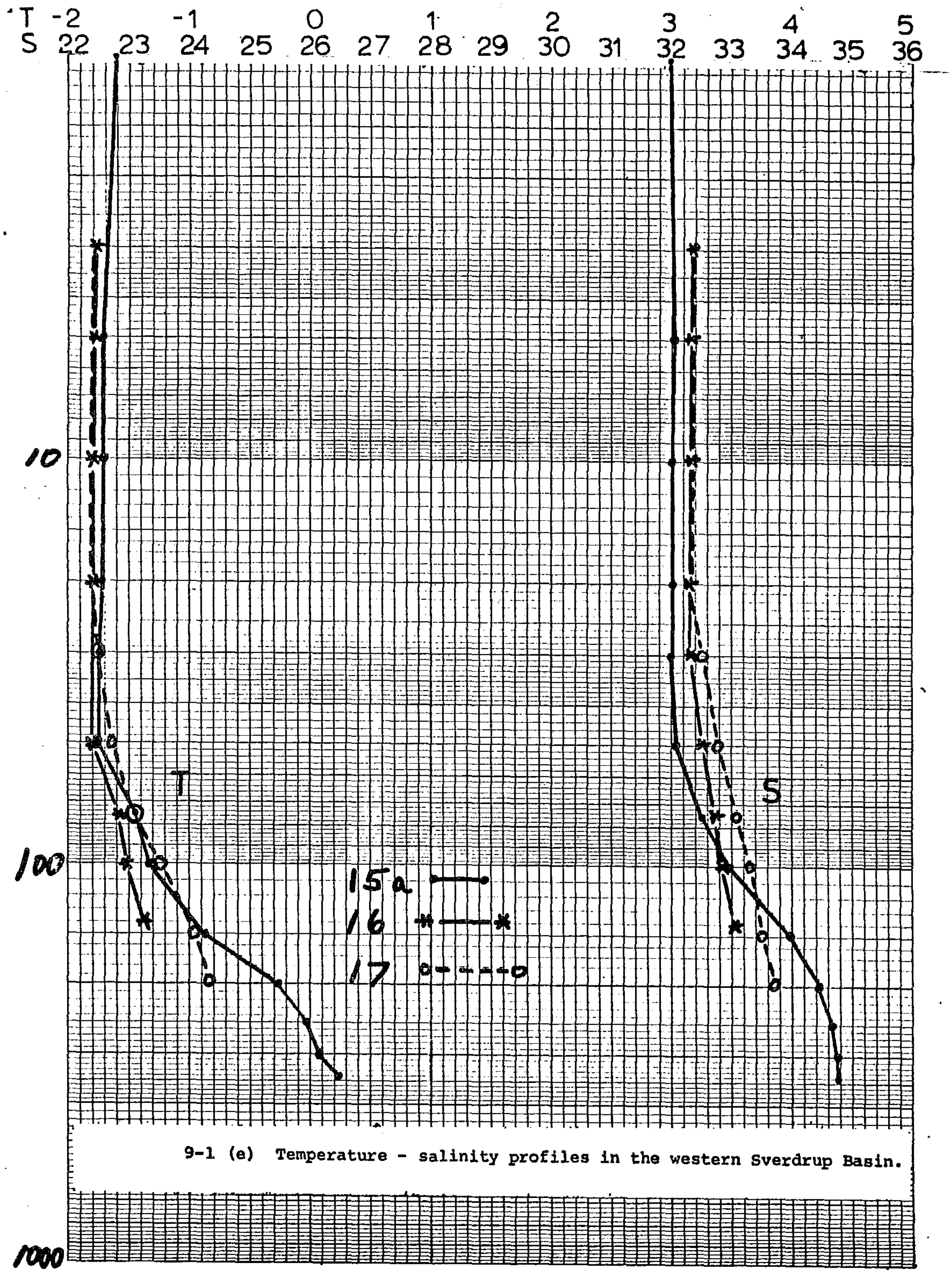
In Figure 9-1 (e) are shown spring time profiles from Wilkins Strait (15), Byam Martin Channel (16) and Austin Channel (17). They are similar to each other, and apart from salinity difference in surface layers, to the Arctic Ocean water offshore the Queen Elizabeth Islands. The three profiles are characteristic of water in the western Sverdrup Basin. It appears that the deeper water in Wilkins Strait is similar to that in Viscount Melville Sound (so far as the data indicates) but the bottom of the profiles in Byam and Austin Channels suggest flow over a sill of depth 100-150 m is occurring either north (or possibly south) of these profiles. There is nothing in this section to say whether circulation is north or south through the Sverdrup Basin although deeper water in Wilkins Strait (and Viscount Melville Sound as far as is shown) are the same as water at the same depths in the Arctic Ocean. These spring soundings lack the surface warm fresh layer shown in the Melville Sound profile in Figure 9-1 (d). Considering the scanty precipitation in the Sverdrup Basin, its greater distance from the large freshwater input of the Mackenzie River and from the mainland coast, the lesser amount of ice melt in the Sverdrup Basin with less melt water input and less wind mixing than in Viscount Melville Sound we may expect summer water profiles in the Sverdrup Basin to have a shallower surface layer than the profile in Viscount Melville Sound.

In Figure 9-1 (f) is shown water structure profiled in the Gustaf Adolf Sea (12), Penny Strait (13) and Wellington Channel (14, 14a). The Gustaf Adolf Sea profile is very similar to those in Figure 9-1 (e) and presumably is representative of water structure in the western Sverdrup Basin. The profiles in Penny Strait and Wellington Channel are saltier from 10-100 m than in the Gustaf Adolf Sea. This suggests strong mixing probably over sills in the Penny Strait, and is strongly suggestive of a flow from the north. The water below 100 m in Wellington Channel is colder and fresher than water at similar depths in the Prince Gustaf Adolf Sea. This water at depths of 200 m in Wellington Channel is fresher than water at the same levels in Lancaster Sound (Figure 9-1 (d)) or the Sverdrup Basin. This suggests that the basin in Wellington Channel is isolated by sills to north and wouth. This matter should be cleared up. The surface layers in Penny Strait and Wellington Channel are similar to those in the Gustaf Adolf Sea in salinity but somewhat warmer in one of the profiles in Wellington Channel, probably because of seasonal variations in surface processes. So while this section strongly suggests northwest to southeast flow through this area, more observations are needed in Wellington Channel to delineate details of deeper water.

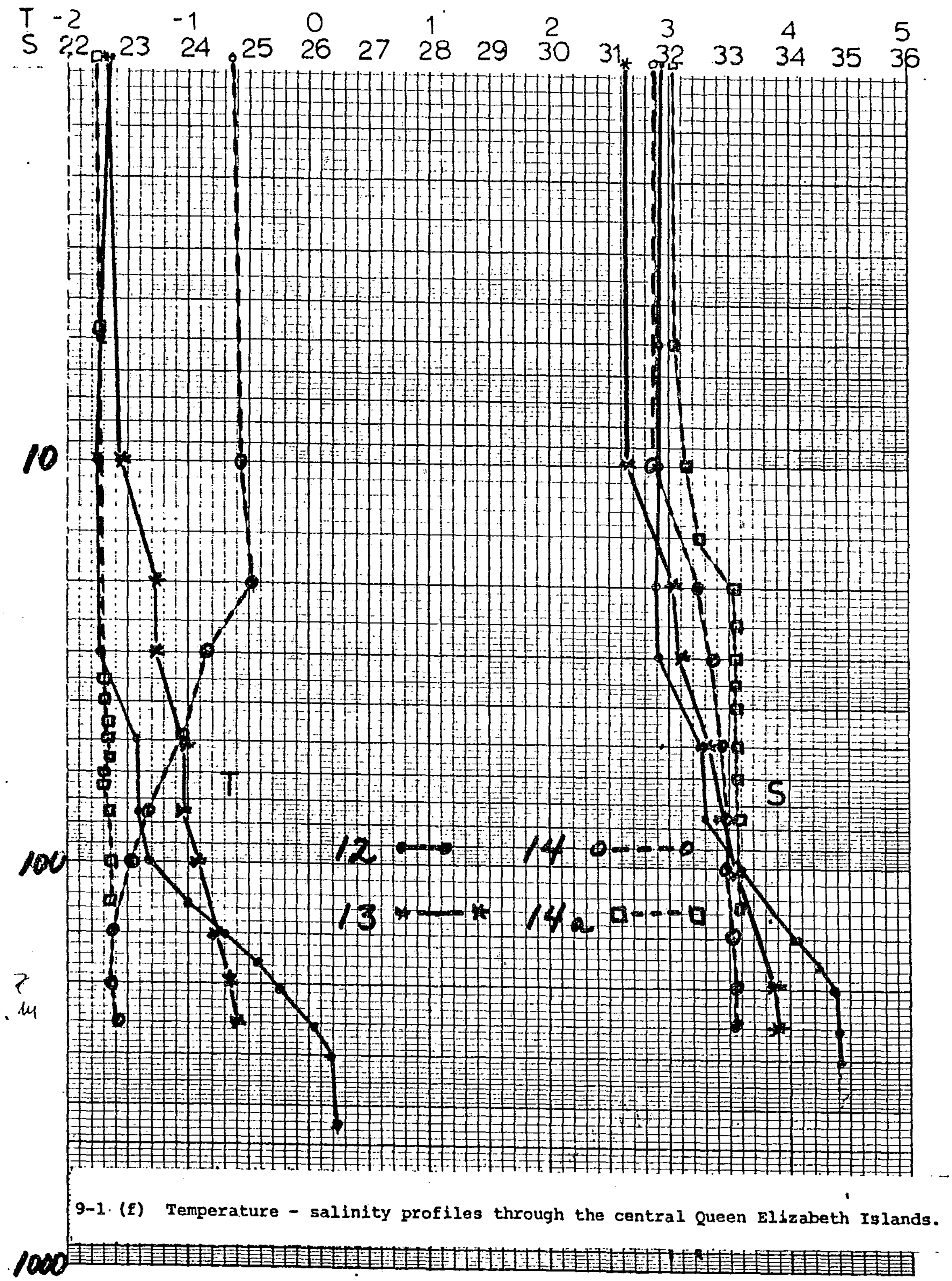
Another passage through the archipelago is from Peary Channel across Norwegian Bay to Hell Gate and Jones Sound. Figure 9-1 (g) shows water profiles from these areas. Water structure in Peary Channel (8) is similar to



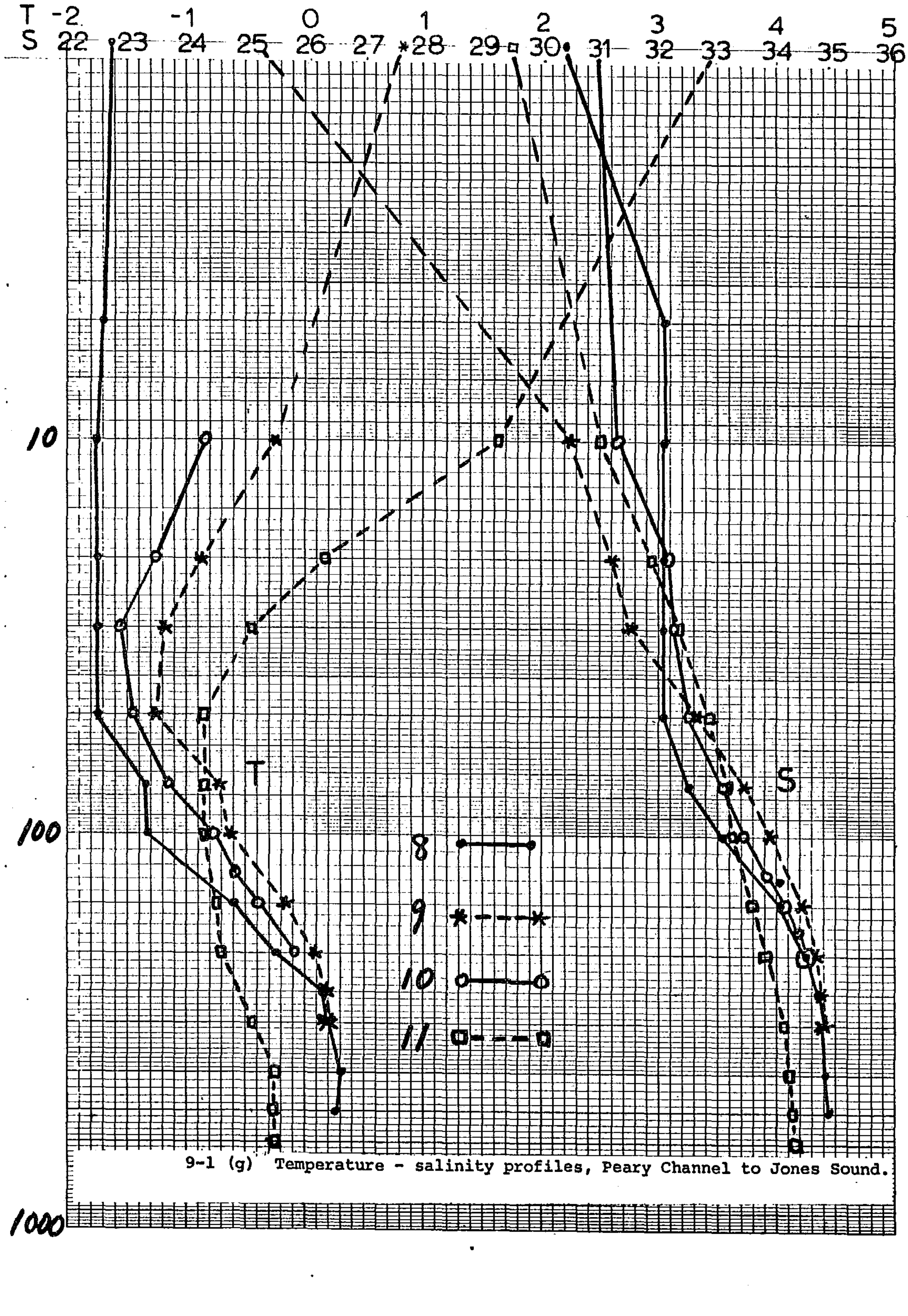
9-2 Temperature-salinity diagram and location key illustrating the origin of the 50-100 m depth warm core observed in Lancaster Sound and less regularly in Jones Sound (after Muench, 1971).



9-1 (e) Temperature - salinity profiles in the western Sverdrup Basin.



9-1 (f) Temperature - salinity profiles through the central Queen Elizabeth Islands.



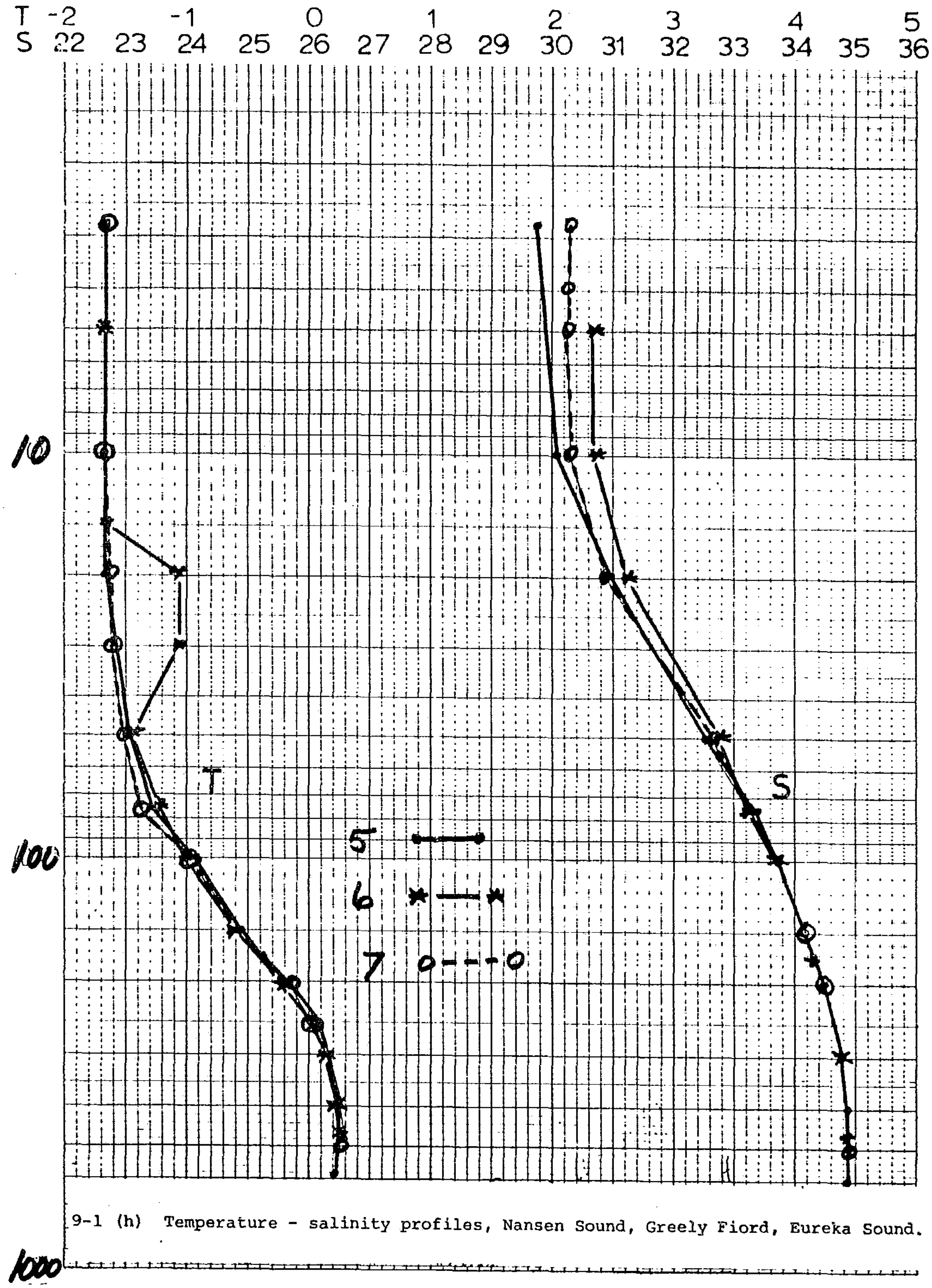
9-1 (g) Temperature - salinity profiles, Peary Channel to Jones Sound.

that in the Sverdrup Basin. The profile in Norwegian Bay (9) is slightly warmer than the profile in Peary Channel, a little fresher above 40 m and a little saltier below. These differences may be due to long term variation, annual variation in structure or processes, possibly combined with some internal mixing. The water structure in Jones Sound (11) is somewhat different. The warmer and slightly fresher surface layers may perhaps be due to annual variation in surface processes. However, below 110 m the water in Jones Sound in colder and fresher than in the other profiles. From Figure 9-1 (g) it might be possible to argue that the deep water in Jones Sound came from the 100 m level of water to the northwest, but the temperature is perhaps too warm particularly if non-adiabatic cooling should occur in the polynya near Hell Gate. The salinity structure at depth in Jones Sound is very similar to that in Lancaster Sound but the temperature in Jones Sound is cooler. If additional cooling could be attributed to something like open North Water than it is probably more likely the water at depth in Jones Sound came from the east rather than over the sill at Hell Gate.

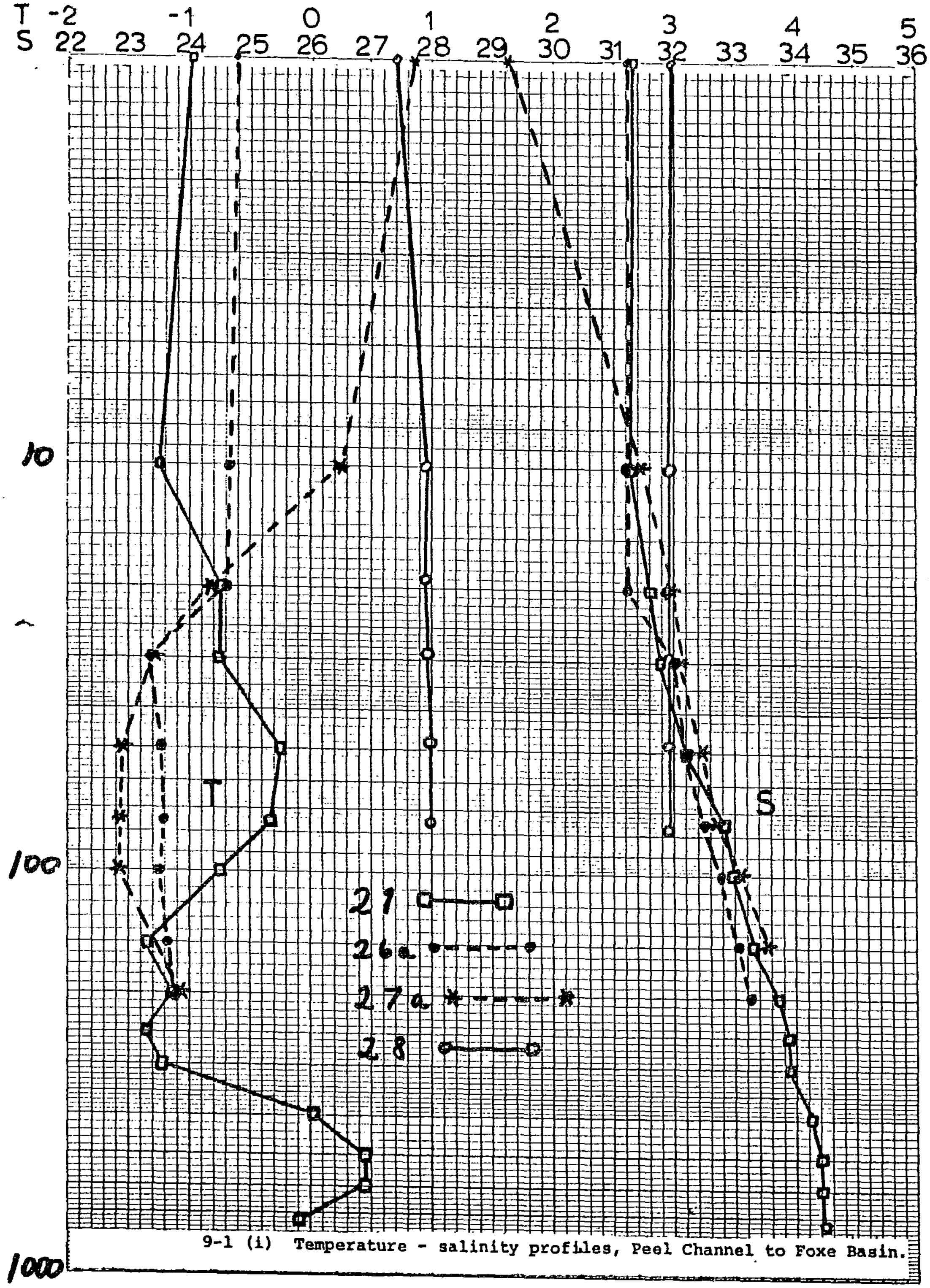
Another passage is that between Axel Heiberg and Ellesmere Island through Nansen Sound, Eureka Sound to Hell Gate and Jones Sound. In Figure 9-1 (h) are shown water structure profiles in the northern part of this channel. With the exception of one feature they are similar to water structure profiles in the northwest area of the archipelago. The exception is the temperature maximum in Greely Fiord (6) between 15 and 50 m. This feature which is dealt with at length below suggests that upper reaches of Greely Fiord are in a sense a stagnant backwater peripheral to the motion through Nansen Sound and Eureka Sound.

In Figure 9-1 (i) are shown water structure profiles from Peel Sound (26a), Prince Regent Inlet (27), Lancaster Sound (21) and Foxe Basin (28). The Peel Sound profile is similar to that from McClintock Channel in Figure 9-1 (c) but is saltier in surface layers. Peel Sound water is similar to water in Prince Regent Inlet although in the latter the surface layers are somewhat warmer. At depth the Prince Regent Inlet profile is similar to that from Lancaster Sound (21) but the latter's temperature maximum from 10-100 m is entirely lacking in Prince Regent Inlet. The water in Foxe Basin is thoroughly mixed but might be derived from water in the upper layers of Prince Regent Inlet or the Gulf of Boothia.

The water structure in the channels east of Ellesmere and Baffin Islands will not be dealt with here. The water sturcture in Nares Strait has been recently discussed at length by Sadler (1976). To summarize, he finds the water column in Nares Strait appears to originate in the Lincoln Sea. A layer of Arctic surface water ($T < 0°C$, $S < 34°/_{oo}$), about 200 m thick overlies water from the Atlantic layer of the Arctic Ocean, ($T > 0°C$, $S > 34.6°/_{oo}$). To a depth of 50 m the water is affected by seasonal factors of insolation, runoff and wind caused turbulence. Sadler finds occasional incursions of Baffin Bay water northward into Smith Sound, to depths of 250 m. He believes the bottom water in central Baffin Bay might be formed by mixing in the (anomalous) North Water area with its marked surface influences, of water from Nares Strait with Atlantic water from Baffin Bay.



9-1 (h) Temperature - salinity profiles, Nansen Sound, Greely Fiord, Eureka Sound.

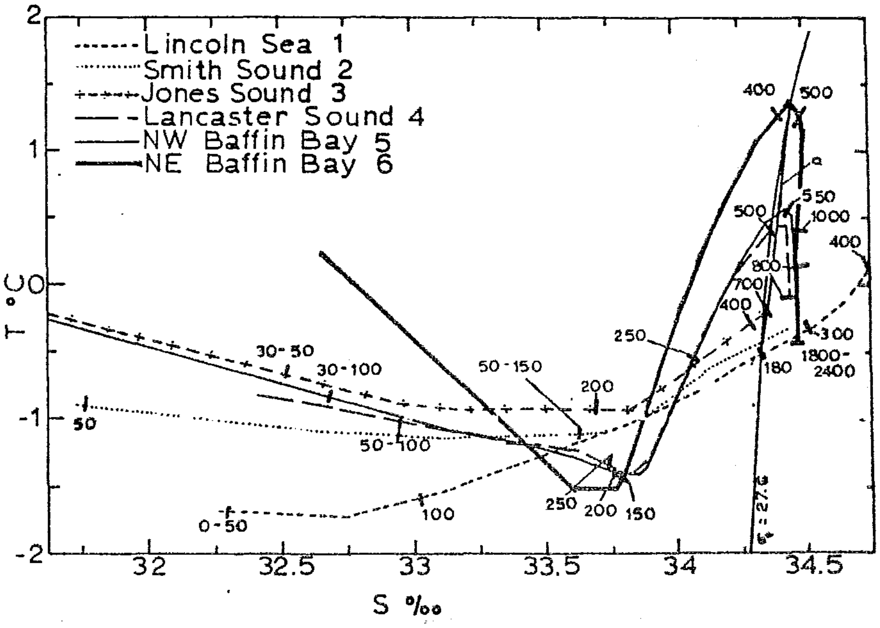


9-1 (i) Temperature - salinity profiles, Peel Channel to Foxe Basin.

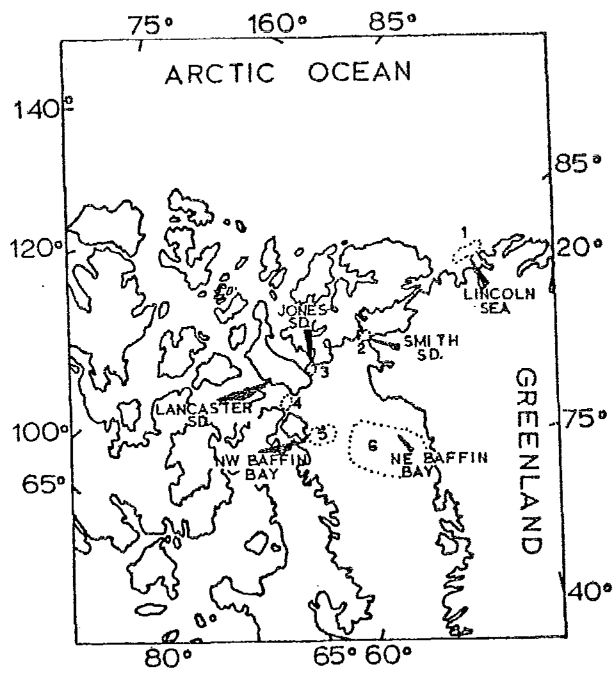
Baffin Bay has enjoyed oceanographic activity that is rather extensive compared to the rest of the Archipelago. The most recent summary is that of Muench (1971). Muench found three layers in the water structure in northern Baffin Bay. On top is a layer of relatively cold ($0°C < T < -1.8°C$; $31°/_{oo} < S < 35.5°/_{oo}$) 'Arctic' water, about 100-150 m deep in E Baffin Bay, and 200-300 m deep in NW Baffin Bay and Lancaster Sound. An intermediate layer of Atlantic water ($0°C < T < 2°C$; $34.2°/_{oo} < S < 34.5°/_{oo}$) enters via eastern Davis Strait, is advected northward by the West Greenland current while mixing with Arctic Water. This water and other water masses are shown in Figure 9-3, (after Muench, 1971). After further mixing presumably this Atlantic water moves southward along eastern Baffin Island. The cross section across Davis Strait shown in Figure 9-4 whose position is shown in Figure 8-4, shows the contrast between northward flow in eastern Davis Strait and the southern flow in western Davis Strait. The circulation and water masses in Baffin Bay may be clearly seen in Barber and Huyer's (1971) atlas.

The picture shown in the preceding salinity and temperature profiles is that of Arctic Ocean water at all depths west of the sill system shown in Figure 3-2. East of this sill system the water column structures are more complicated with mixing, particularly below the surface layers, of water that has come from the Atlantic via Davis Strait with upper level Arctic Ocean water. The mixing is perhaps complicated by the greater proportion (in space-time) of open water, with greater water mass modification than occurs in the northwestern archipelago.

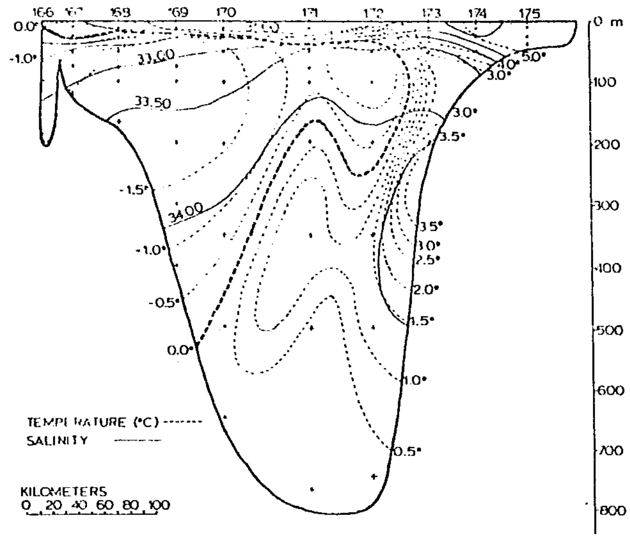
Additional confirmation of this is provided by Grainger who in a series of papers (Grainger 1961, 1962, 1963) classified biological parameters found in surface water layers as being characteristic of Arctic or Atlantic water. His analysis resulted in the estimates of characteristic surface water layers shown in Figure 9-5 (after Grainger, 1961). Analysis of the reactive silicate budget for the Arctic Ocean (Codispoti and Lowman, 1973) suggests that much of the Bering Sea water in the arctic may tend to pass southeastward through the Canadian Arctic Archipelago but supporting data are scanty.



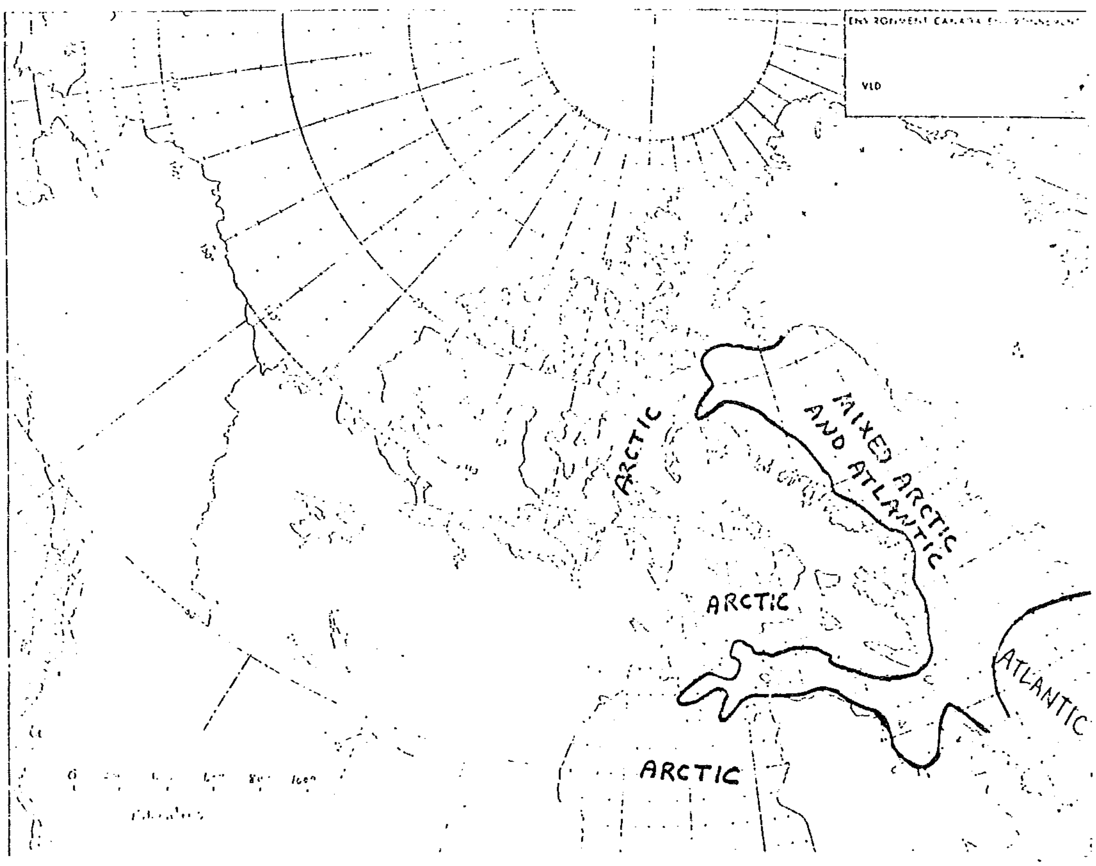
9-3 (a) Temperature-salinity diagram contrasting water profiles in northern Baffin Bay with those found at locations in Figure 9-3 (b). Numbers on curves indicate depth in meters.



9-3 (b) Geographical location of water masses illustrated in Figure 9-3 (a).



9-4 A temperature-salinity cross section across Davis Strait, illustrating warmer northward flow and colder southward flow on the west (after Riis-Carstesen, 1936).



9-5 Type of water in surface layers as deduced from biological evidence by Grainger (1961, 1962, 1963).

FIORD OCEANOGRAPHY

At d'Iberville Fiord, at the eastern end of the Nansen Sound - Greely Fiord system, the Frozen Sea Research Group, Institute of Ocean Sciences, Patricia Bay, B.C. has maintained a small base since 1968. At Strathcona Sound, Baffin Island, operation of a lead-zinc mine beginning in 1976 was preceded by a brief oceanographic program, carried out principally by B.C. Research, Vancouver, on behalf of the mining company. The Frozen Sea Research Group has worked in Cambridge Bay. The results of the most extensive experiments there in 1971-1972 are summarized below.

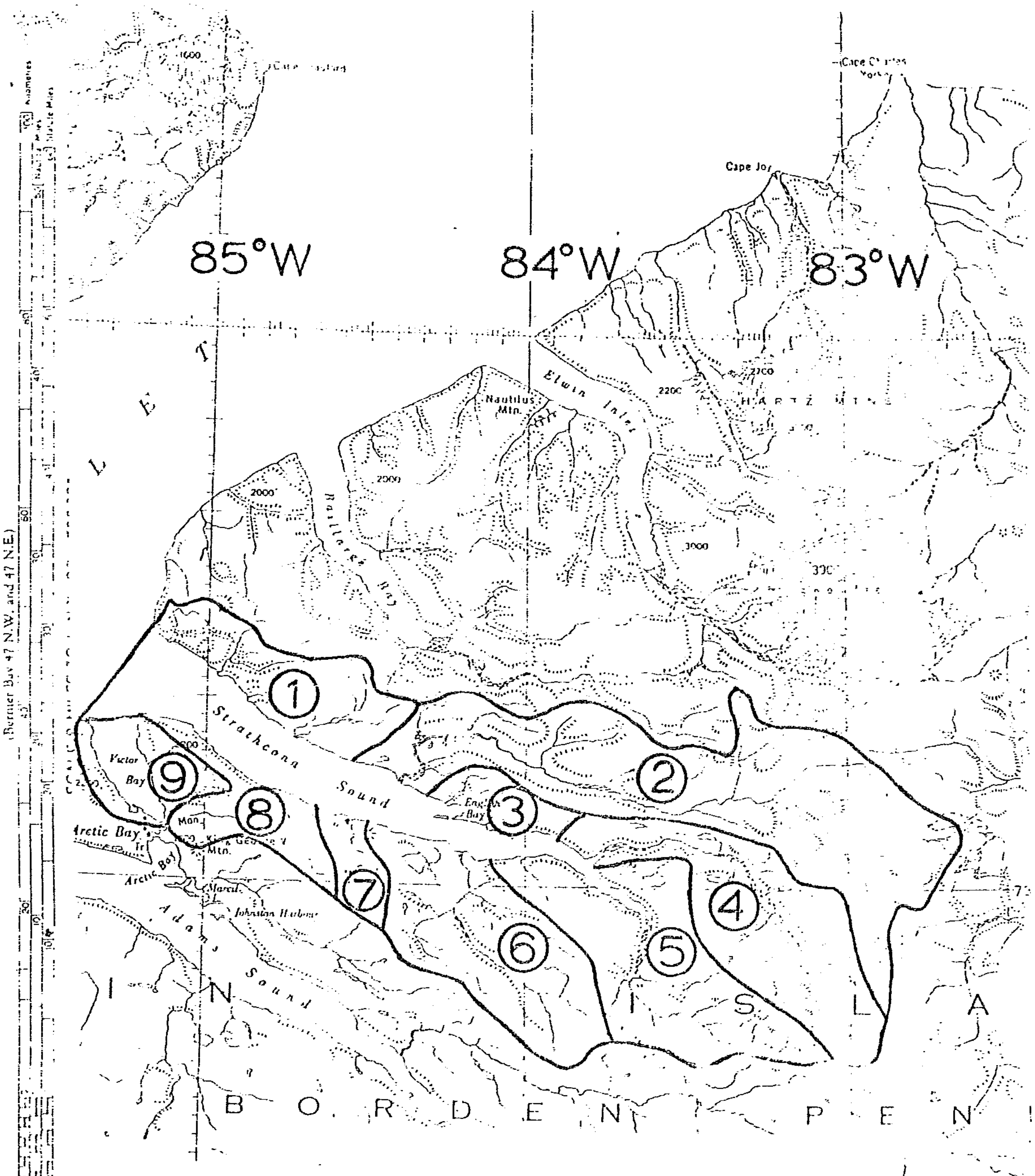
Strathcona Sound

Oceanographic knowledge of Strathcona Sound is based mainly on measurements made by B.C. Research in 1974 and 1975. Strathcona Sound, which extends NNW by ESE for about 30 miles at 73°N, opens into Admiralty Inlet, Figure 10-1. Maximum depths in Strathcona Sound are about 300 m. A sill with depths about 200 m lies across the Sound about seven miles in from the mouth. Surface area of Strathcona Sound is $3 \times 10^8 m^2$ while its drainage area is $21 \times 10^8 m^2$. Admiralty Inlet extends roughly N-S at about 86°W from nearly 74°N to 71°N and is thus about 200 miles long. A sill with depths about 300 m lies across the mouth of Admiralty Inlet. Depths to 600-700 m apparently lie down the centre of Admiralty Inlet off Strathcona Sound, although Hydrographic Chart 7503 indicates a shallower shelf off the mouth of Strathcona Sound. Admiralty Inlet in its turn opens into Lancaster Sound with depths of 600-800 m. Lancaster Sound opens on the east into Baffin Bay.

The physiography of the local area is shown in Figure 10-2, a portion of the National Topographic Series, Sheet 72/88. Drainage areas are sketched in. Preliminary estimates of meteorological parameters, ice growth and so on for the Strathcona sound area are shown in Table 10-1. Estimates of runoff for the various drainage basins are given in Table 10-2. As discussed above, the precipitation amounts from the Arctic Bay weather station records are serious underestimates of amounts over surrounding water sheds. Corrections of the size in sections 4 and 5 are necessary. In Figure 10-3 is shown the bathymetry of Strathcona Sound. Depth contours at 50 m intervals indicate a sill structure with depths about 220 m seven miles inside the mouth of the sound. A graph of water depths, along the centre of the sound on the section shown in Figure 10-3 is shown in Figure 10-4. The cross sections at X2 at the sill and cross section X3 are also shown in Figure 10-4.

B.C. Research undertook oceanographic sampling August 8 - 11th, 1974 and August 17 - 24th, 1974. Current measurements were made over the period August 18 - 24th, 1974. Strong winds on August 16 - 17th, 1974 disrupted tidal and wind records and changed the water structure in the upper layers. These data are available in a 1975 unpublished B.C. Research report. In March 1975, at three stations, observations were made of conductivity, salinity, temperature and pH. Measurements of currents were made by lowering a current meter at the same locations. Three current meters were moored just below the ice for removal in June. In May - June of 1975 the ice-bottom current meters were recovered. Dissolved oxygen and pH measurements were

10-1 Location of Strathcona Sound, Baffin Island, with surrounding areas. After Canadian Hydrographic Chart No. 7000 (fathoms).



10-2 Physiography of Strathcona Sound area, with drainage basins, on National Topographic Series, Sheet 72/68. Heights in feet.

TABLE 10-1

Monthly Values of Various Parameters in the Strathcona Sound Area

With Units, Sources and Other Details

	JAN.	FEB.	MAR.	APRIL	MAY	JUNE	JULY	AUG.	SEPT.	OCT.	NOV.	DEC.	YEAR
1	-21.7	-23.9	-17.9	-3.7	18.4	35.7	42.5	40.7	29.2	12.1	-5.9	-16.3	7.4
2	0	0	0	0	Tr.	.16	.75	.91	.33	Tr.	0	0	2.15
3	2.3	1.7	2.2	1.6	2.5	1.4	0.1	0.3	5.3	6.0	2.7	1.8	28.0
4	.23	.17	.22	.16	.26	.30	.76	.95	.85	.60	.27	.18	4.95
5							15th			15th			
6	36	42	50	55	55	47	34	-	-	11	20	30	
7	10	10	12	11	9	-	-	-	1	6	7	8	
8	7.6	7.6	9.1	8.4	6.9	-	-	-	.8	4.6	5.3	6.1	
9	0	20	140	360	560	600	460	270	130	40	10	0	
10	80	80	80	80	60	50	40 / 6	9	15	34 / 70	80	80	
11	0	∿0	∿0	∿0	∿.2	50	123 / 447	246	110	∿0	∿0	∿0	
12	80	-90	-80	-40	+50	+125	+150	+110	-20	-80	-110	-90	
13						∿0	.3	1.5	.2	0			

1. Mean daily air temperature (°F) 1941-70 Arctic Bay (AES).
2. Mean monthly rainfall (inches) 1941-70 Arctic Bay (AES).
3. Mean monthly snowfall (inches) 1941-70 Arctic Bay (AES).
4. Mean monthly total precipitation (inches of water) 1941-70 Arctic Bay (AES).
5. Mean dates of ice clearance and formation Arctic Bay (AES).
6. Average month-end ice thickness (inches) Arctic Bay (AES).
7. Median snow depth at month's end (inches) Arctic Bay (Potter, 1965).
8. Snow water content from (7) with ρ_S = .3 g cm^{-3}. (cm of water)
9. Mean daily global solar radiation (cal $cm^{-2}day^{-1}$) (Titus and Truhlar, 1969).
10. Sea surface albedo 70°N (Larsson and Orvig, 1961).
11. Estimated shortwave radiation at water surface using median snow depths with transmissivity factor 0.2 cm^{-1} (O'Neill and Gray, 1972), ice thicknesses with transmission factor .015 cm^{-1} and (9). (cal $cm^{-2}day^{-1}$).
12. Estimated net radiation (Vowinckel and Orvig, 1964). (cal $cm^{-2}day^{-1}$).
13. Evaporation from lakes (inches of water) (Ferguson, O'Neill, Cork, 1970).

TABLE 10-2

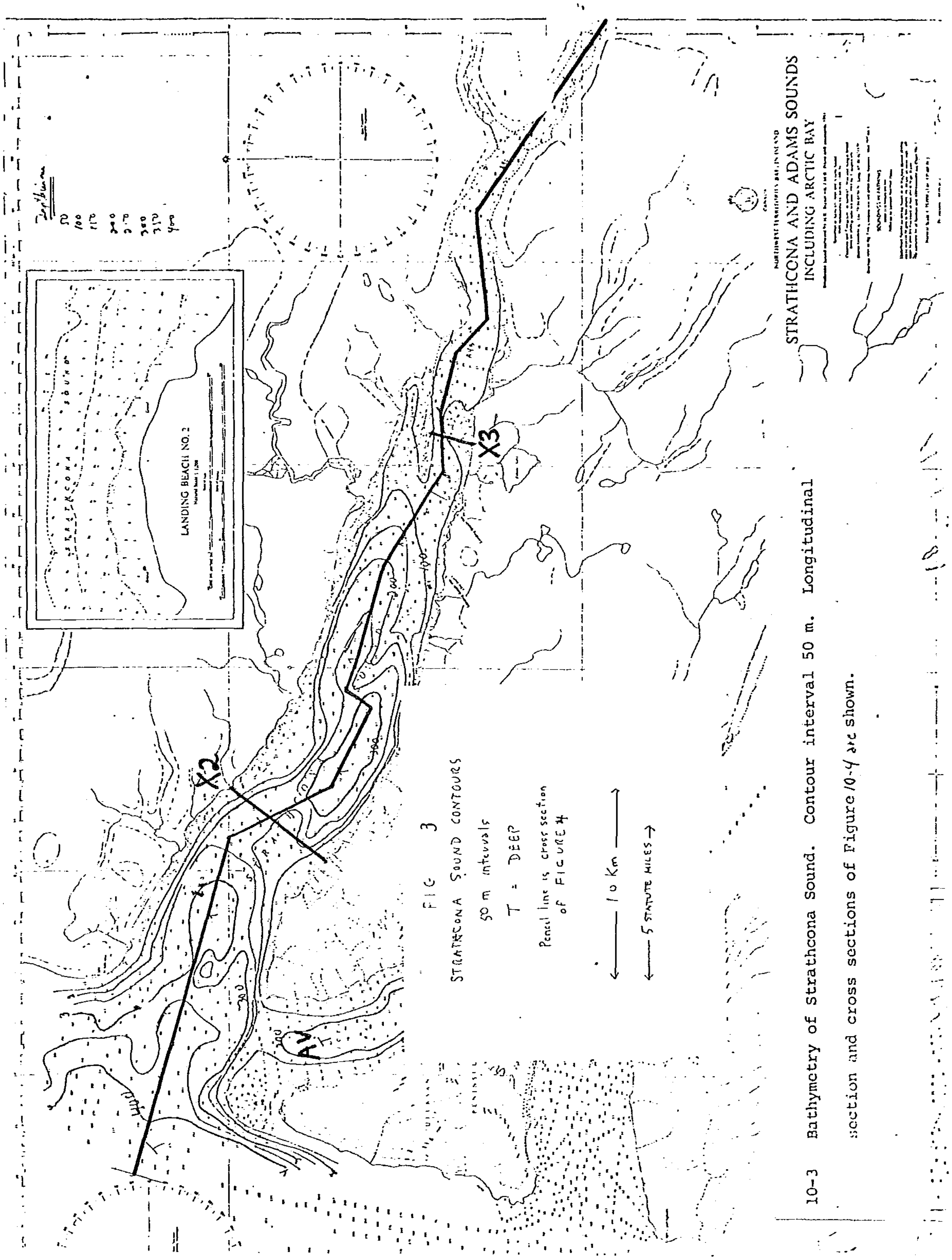
Estimated Run-Off Into Strathcona Sound From Areas in Figure 2

BASIN	AREA (10^8 m^2)	ARCTIC BAY PRECIPITATION				
		JUNE (10^6 m^3)	JULY (10^6 m^3)	AUG (10^6 m^3)	SEPT (10^6 m^3)	ANNUAL (10^6 m^3)
1	1.9	8	7	4	4	23
2	6.3	27	23	15	14	79
3	.6	3	2	2	1	8
4	3.3	14	12	8	7	41
5	3.3	14	12	8	7	41
6	3.0	13	11	7	6	38
7	.6	2	2	1	1	7
8	1.0	4	3	2	2	12
9	.7	3	3	2	1	9
Victor Bay	.5	2	2	1	1	6
Strathcona Sound	2.9	12	11	7	6	36

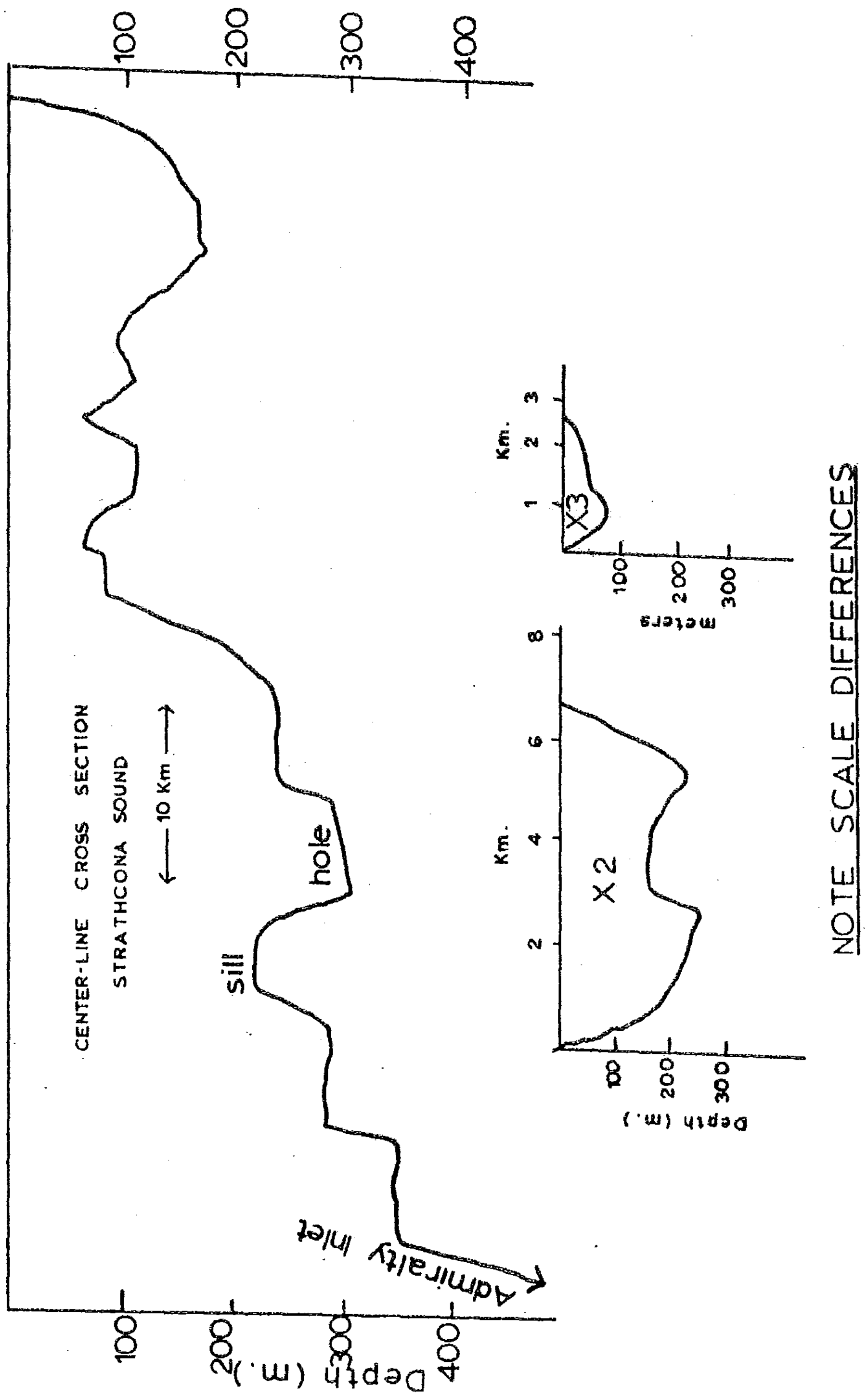
Assumes:

1. Snowmelt (Oct. - May) split 2/3 to June, 1/3 to July
2. No allowance for evaporation
3. Area of Victor Bay plus Strathcona Sound is 3.4×10^8 m^2.
   Depth of run-off $\sim$ 0.9 m over whole area.

Estimated runoff into the Strathcona Sound from areas marked in Figure 2, using Arctic Bay precipitation from Table 10-1,



10-3 Bathymetry of Strathcona Sound. Contour interval 50 m. Longitudinal section and cross sections of Figure 10-4 are shown. Areas



10-4 Section along the centre of Strathcona Sound, and the cross sections X2 and X3.

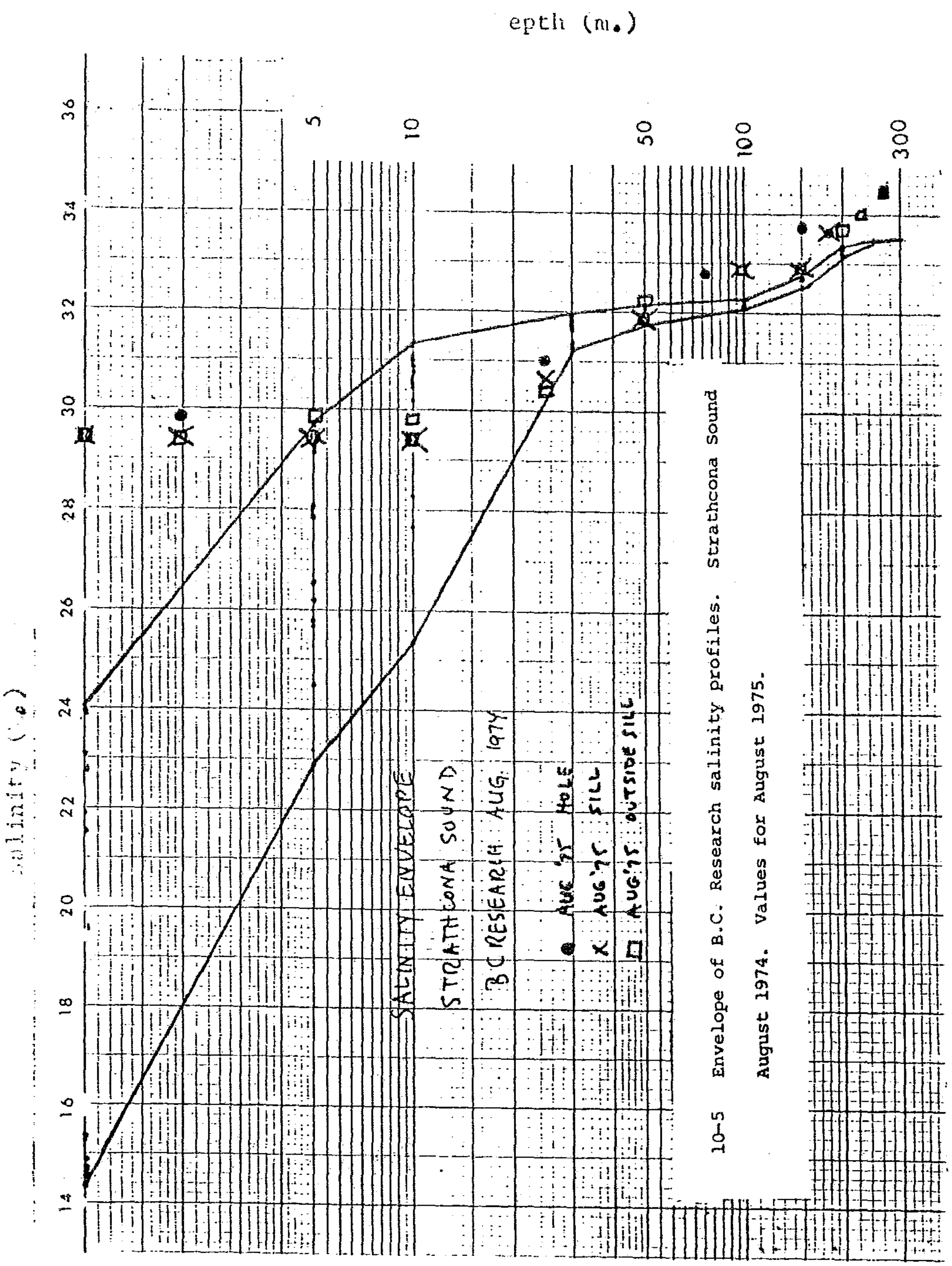
made in the hole. In August 1975, dissolved oxygen, pH, S and dissolved solids were made at three locations. Three current meters were suspended for two weeks in a chain.

In August, 1974 the range of salinities and densities was large above 30 m depth but small below. The upper levels can be grouped into samples taken before the storm of August 17, 1974, with 1 m salinities about 15 °/₀₀ and those after the storm with salinities at 1 m being 22 - 24 °/₀₀. It would appear that the storm of August 17, 1974 caused changes in salinity profiles to at least 30 m depth (Figure 10-5). Temperature samples were taken with the salinity samples in August 1974. The envelope of observed values indicates little change below 50 m depth but the surface layers, which were warm before the storm of August 17, 1974 were then subjected to mixing extending downward to 50 m, one would estimate, from the temperature records in Figure 10-6. In March 1975 three salinity-temperature samplings were made at water columns just outside the sill of Strathcona Sound, on the sill, and over the deepest part of Strathcona Sound. These three salinity profiles indicate mixing to about 30 - 40 m. Differences in values at lower depths in Strathcona Sound and at levels above 80 m in the mouth of Strathcona Sound cast some doubt upon the accuracies of these data (Figure 10-7). Temperature profiles at the same three stations in March 1975 indicate mixing to depths of about 50 m. The temperature structures are similar at lower depths (Figure 10-8). August 1975 salinity samplings made at the same locations as in March 1975 are shown superimposed on Figure 10-5.

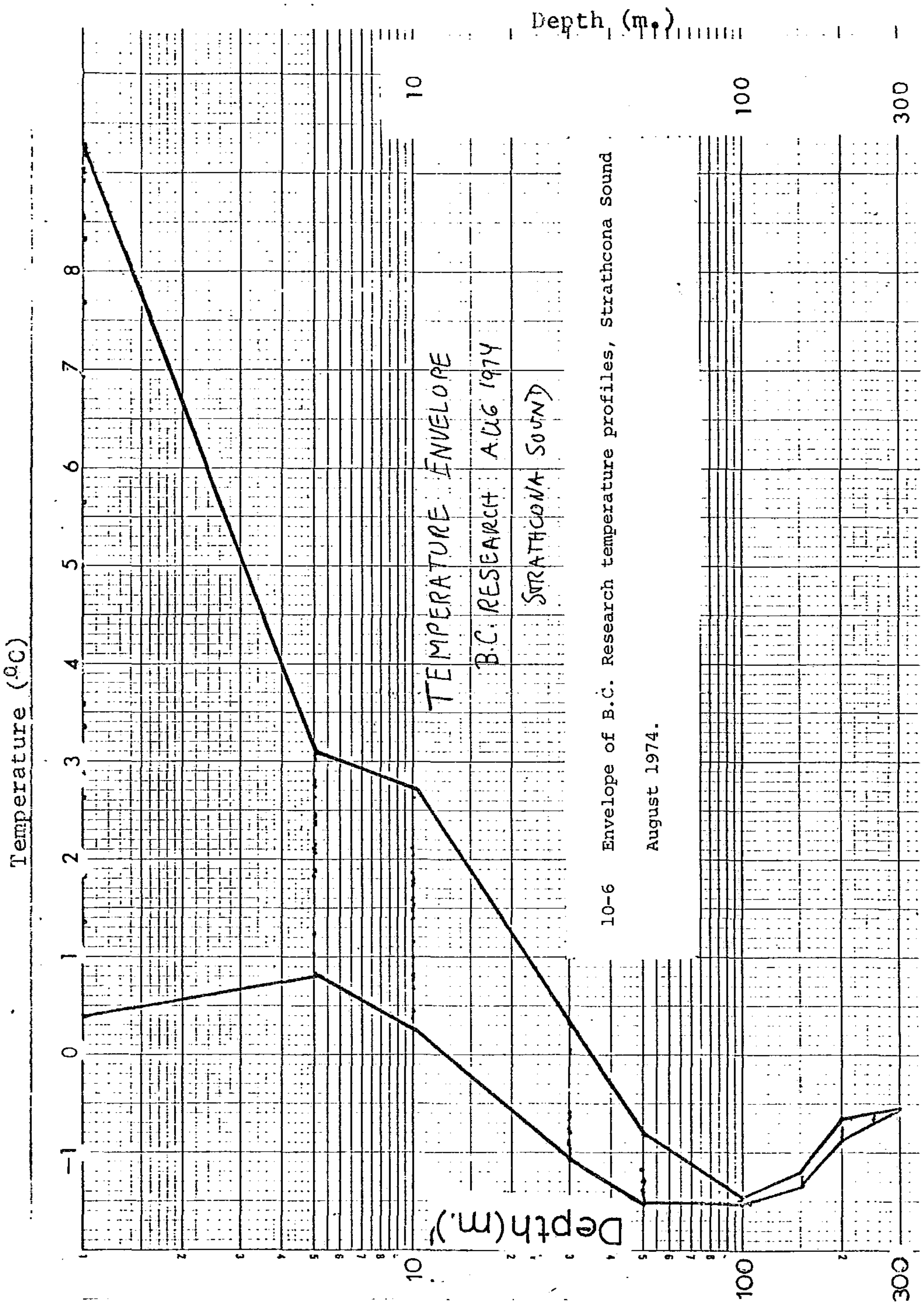
In August 1974 dissolved oxygen measurements were taken with an in situ instrument on the same schedule as the temperature and salinity measurements. The envelope of values is shown in Figure 10-9. Most values were on the low side of the range contained within the envelope, particularly below 100 m. The B.C. Research report also contains dissolved oxygen levels in fresh water systems flowing into the sound. These values were about 13 ppm. In June 1975 a sample was taken in the hole by standard methods. In August 1975 samples were taken in the hole, on the sill, and outside the sill by the same method as in June. These June 1975 and August 1975 values are plotted on Figure 10-9.

Current measurements were also made from a launch by B.C. Research in Strathcona Sound in August 1974. Surface currents in the upper layers ranged up to about 0.5 kt (25 cm sec^{-1}). Currents above the current meter start speed were not measured at depths except near the bottom on the sill where currents of up to 0.2 kt (11 cm sec^{-1}) were measured. In March, 1975 current profiles were taken through the sea ice at the three stations mentioned above. No currents above the meter start speed were recorded at any depth. The start speed of their current meters is stated in B.C. Research Report on August 1974 activities as 3.0 cm sec^{-1}. Tidal measurements were made by B.C. Research from August 5th to October 16th, 1974. Tides are semi-diurnal. The largest range measured by B.C. Research in the period was 2.74 m (9 feet).

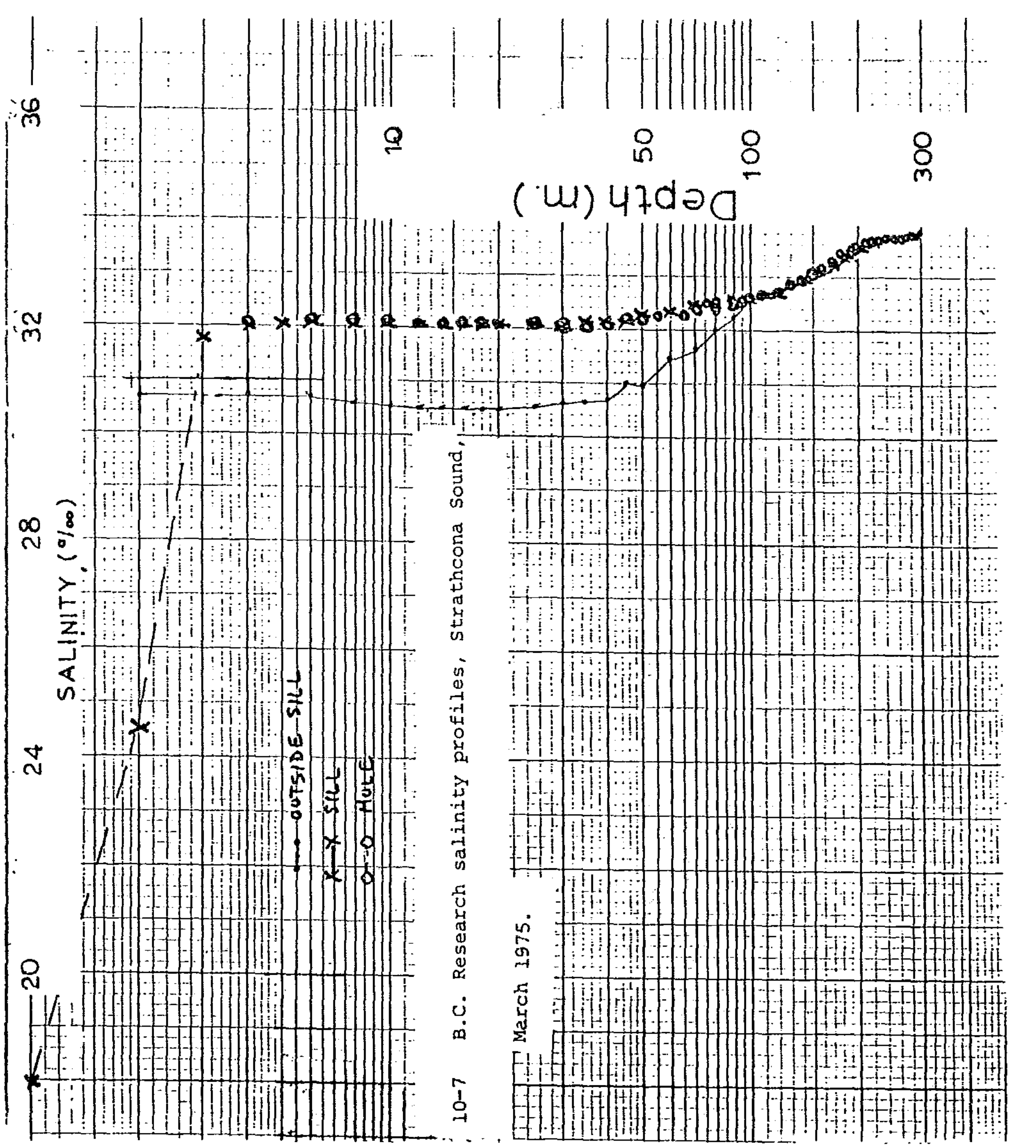
Chemical oceanography samplings were made in August 1974 by B.C. Research and by Environment Canada. Samplings were also made of stream and lake water. Further studies were made in August 1975. The high levels of heavy metals in streams were found near an orebody, roughly in drainage area #7 in Figure 2. Heavy metal levels in sea water were relatively uniform over different water depths at different sampling stations. High heavy metal levels in bottom sediments were more easily related to stream outflow from the vicinity of the orebody.



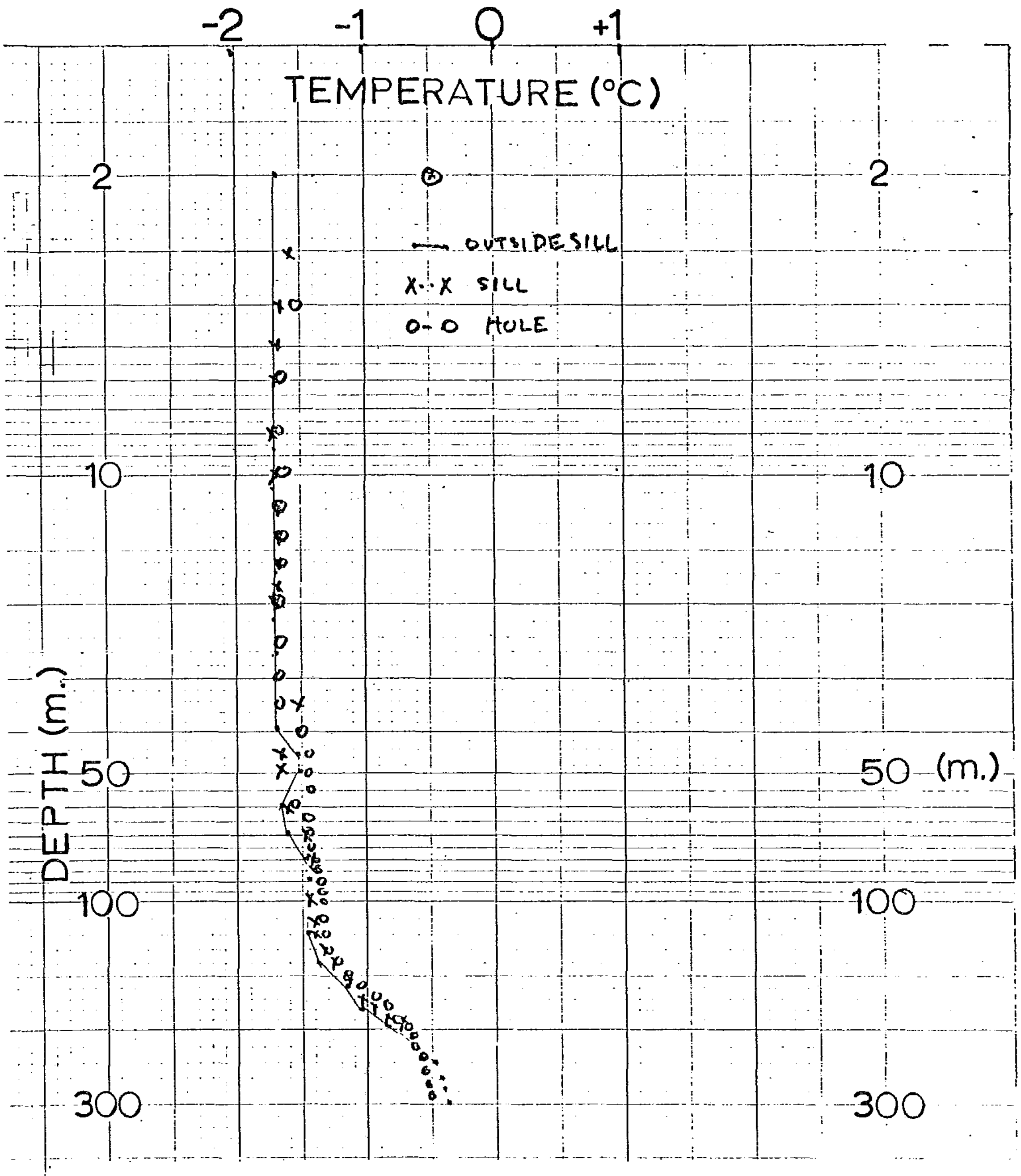
10-5 Envelope of B.C. Research salinity profiles. Strathcona Sound August, 1974. Values for August 1975.



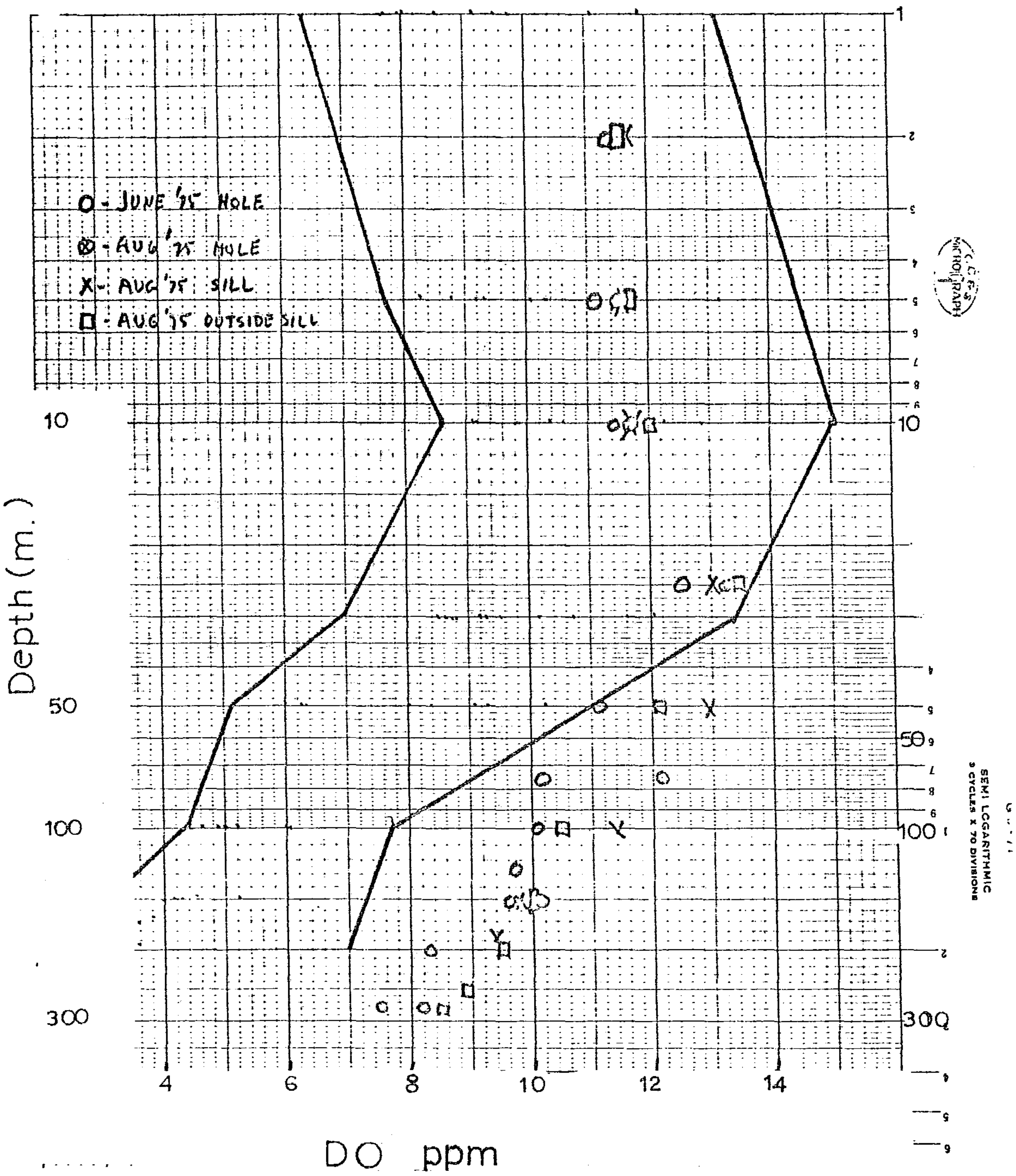
10-6 Envelope of B.C. Research temperature profiles, Strathcona Sound August 1974.



10-7 B.C. Research salinity profiles, Strathcona Sound, March 1975.



10-8 B.C. Research temperature profiles, Strathcona Sound, March 1975.



10-9 B.C. Research dissolved oxygen profiles, Strathcona Sound, August 1974, June 1975, August 1975.

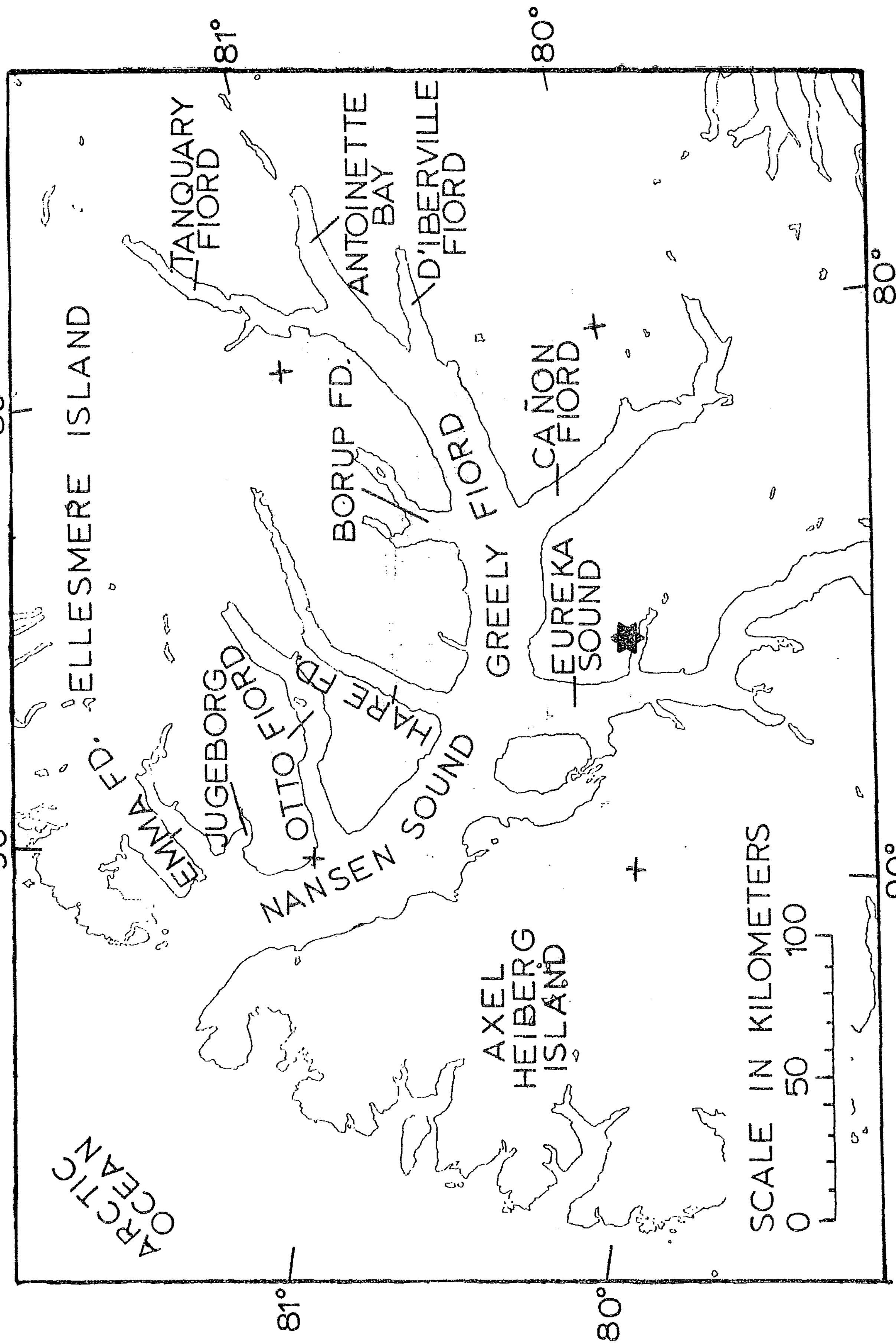
Levels of pH were recorded in August 1974, ranging from 7.6 to 8.2. Measurements of pH were made in March 1975, in the hole, on the sill and outside the sill. Values then varied from 5.8 to 7.8 but their accuracy under difficult winter sampling conditions is not certain. In August 1975, pH values ranged from 7.7 to 8.1. In August 1974, turbidity, measured with a Secchi disk, tended to be higher close to shore. In August 1975 total suspended solids ranged from 6 to 8 mg/l at the surface, 8 to 10 mg/l near the bottom.

To summarize, water structure from Figures 10-5 and 10-7 shows stable stratification except for near neutral-stability in wind mixed layers after storms in summer and under ice convection layers in late winter. Mixing in both cases can extend to 40 or 50 m. This same depth of mixing is evident in the temperature records of Figure 10-6 and Figure 10-8. The dissolved oxygen sampling shown in Figure 10-9 suggested values as low as 2 to 6 ppm at depth in August 1974. However, samplings in June and August 1974 determined by standard chemical methods indicated values of about 8 ppm at depth. Probable instrumental trouble was encountered in August 1974. No comparisons with standards were reported for August 1974 for dissolved oxygen, temperature or salinity. Surface layer water currents were reported in August 1974 as well as near bottom currents on the sill. No currents above the threshold of the meter were reported from March 1975. Time lapse movies made in 1975 showed considerable surface water movement as did surface silt plume movements in August 1974. Crude estimates of tidal currents from tide heights and cross section areas suggest area-mean currents of at least 3 cm sec^{-1} over the sill and about 9 cm sec^{-1} at the cross section X3 in Figure 10-3. The chemical analyses of the water of Strathcona Sound imply mixing in the summer surface layers. The stream runoff estimated in Table 2 is rather small and distributed along the sound in a fashion rather different than in a typical estuary with a river at its head.

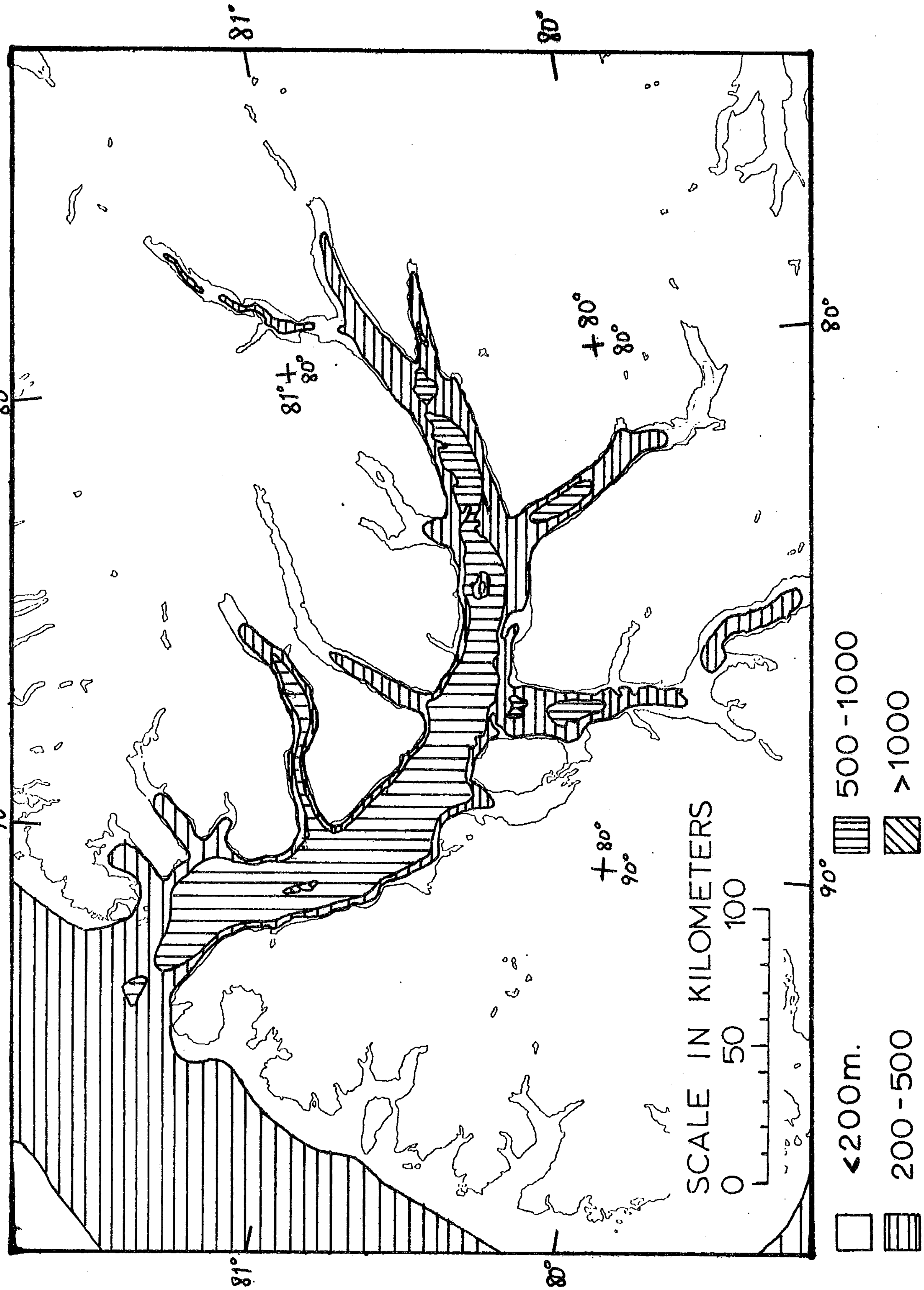
Greely Fiord

The Nansen Sound - Greely Fiord system was first seen by white men in 1883. Oceanographic studies did not begin until after 1950, and have intensified since 1962 when the Canadian Defence Research Board first operated a base at Tanquary Fiord. In 1968 the Frozen Sea Research Group occupied a base on d'Iberville Fiord. A fair amount of oceanographic data is available for the system and has been published by the Defence Research Board, Canadian Oceanographic Data Centre (now M.E.D.S.) and the Institute of Ocean Science, Patricia Bay, B.C. A modest number of papers have discussed oceanography of the system [Hattersley-Smith and Serson,(1966); Ford and Hattersley-Smith (1965); Lake and Walker (1973); Lake and Walker (1976)]. This brief section in many ways is a summary of these previous studies.

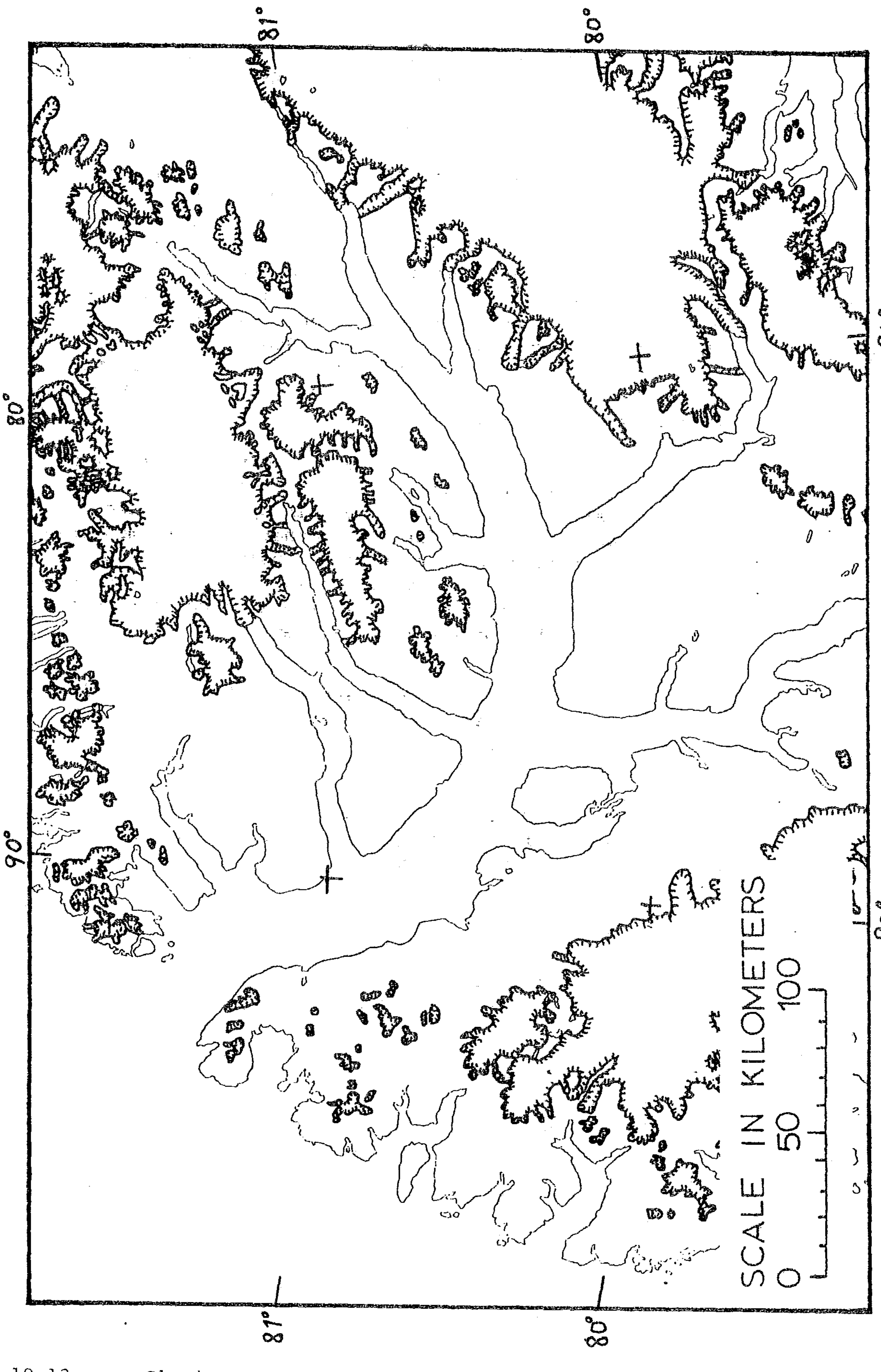
The names of the various water bodies making up the system are shown in Figure 10-10. The bathymetry, on a larger scale than that in Figure 3-2, is shown for the Nansen Sound - Greely Fiord system in Figure 10-11. The main stem of the system is between 500 and 1000 m deep, which is deeper than the continental shelf in the area. The upper reaches of the subsidary fiords tend to be quite shallow. Tanquary Fiord and d'Iberville Fiords, at least, have sills near their entrances, at depths of about 200 m. The upper reaches of the fiords extend into mountainous terrain. Much of this mountainous terrain is glaciated and many of the glaciers reach sea level (Figure 10-12).



10-10 The various water bodies making up the Nansen Sound - Greely Fiord system. The location of Eureka weather (80 00N, 85 56W) is marked with a star.



10-11 Bathymetry of the Nansen Sound - Greely Fiord system.



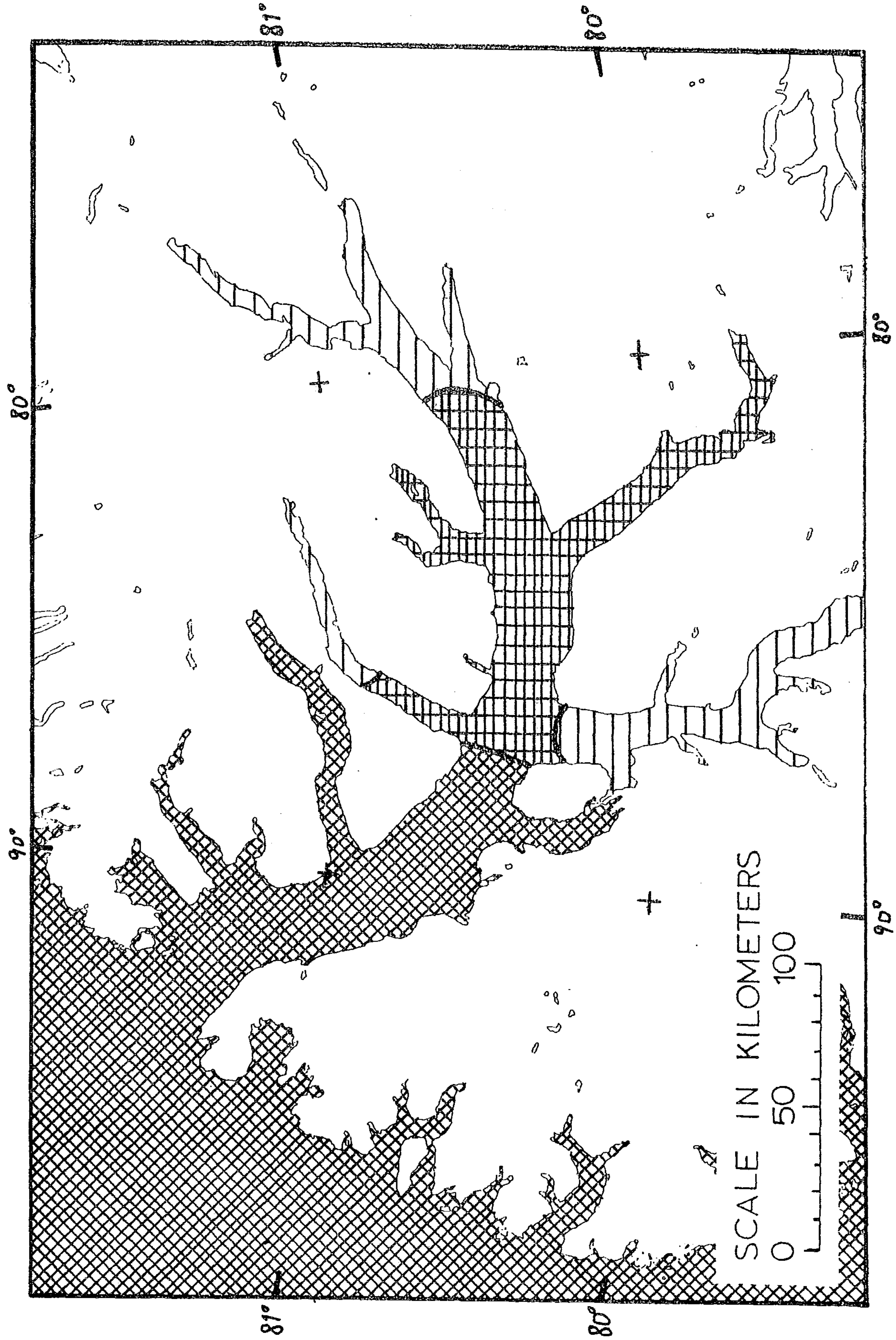
10-12 Glaciers in the Nansen Sound - Greely Fiord area.

Published material on details of sea ice cover of the system is scanty, although much data must be hidden in logs of the icebreakers which have been venturing into the upper part of the system for the past 15 years, and have come to the Eureka weather station (80 00N, 85 56W), for the past thirty years. The authors personal assessment of relative summer ice cover is shown in Figure 10-13. The upper reaches of the Greely Fiord system, of Hare Fiord and much of Eureka Sound have some open water each summer and the sea ice breaks up rather more completely in perhaps one half the summers. In the body of Greely Fiord, Canon Fiord and Borup Fiord, while patches of open water may be present in most or every summer, any extensive clearing of sea ice is less frequent. In Nansen Sound the ice cover is generally heavier and more persistant. The exception to this can occur when a massive floe or ice island jams across the northwestern reaches of Nansen Sound interrupting the movement of ice from the northwest. Over most of the area, as far as sea ice cover is concerned, summer extends from the end of July to early September as might have been guessed from Figure 7-3.

As noted in section 6, daily tides in the area are very small, the mean tidal ranges being much less than one meter. Tidal currents should then be rather small. The weather station at Eureka has been in operation since 1947. A selection of the average values of different weather elements measured there are given in Table 10-3. Estimates of precipitation volumes over the fiord system have been discussed in section 5. There it was shown that because of differing ratios of water surface to drainage areas, the laydown of runoff tended to be higher on the upper reaches of the fiord system.

It seems that in summer at least the main surface water current is that flowing to the southeast through Nansen Sound and southward through Eureka Sound. The variation in this current through the year is unknown but appreciable variation may be present. The upper part of the Greely Fiord system should contain an estuarine circulation which will presumably be stronger in summer when runoff occurs. Each individual fiord will presumably have its own estuarine sub-circulation. Apart from these time-mean currents, weak tidal currents have been observed, as have wind caused surface water movements in open water.

As noted in the discussions of water structure in section 9, the water masses in the Nansen Sound - Greely Fiord system came from the Arctic Ocean. Small amounts of runoff are added annually as is freshwater from ice melt. An interesting feature of the temperature structure has already been shown in Figure 9-1 (h). It is the warm layer, at depths of about 20 to 50 meters in Greely Fiord, and its upper reaches which is not present in the water profiles from Nansen Sound and upper Eureka Sound. It has been noted in practically all profiles of the area. Figure 10-14 is a reproduction of Figure 3 in Ford and Hattersley-Smith (1965). In Figure 10-14 the layer in question is, in the upper reaches of Greely Fiord and Tanquary Fiord, separated from the warm fresh surface layers characteristic of warm summers with open water as 1962 was. Probably the most detailed profiles so far as temperature and salinity profiles are concerned down Greely Fiord came from a traverse from upper d'Iberville Fiord to eastern Nansen Sound by a team from the Frozen Sea Research Group in March-April 1976. Cross sections of temperature and salinity from these profiles are shown in Figure 10-15. The scale in depth is logarithmic. The features shown are very similar to those in the temperature cross section of Figure 10-14, from data taken fourteen years earlier.

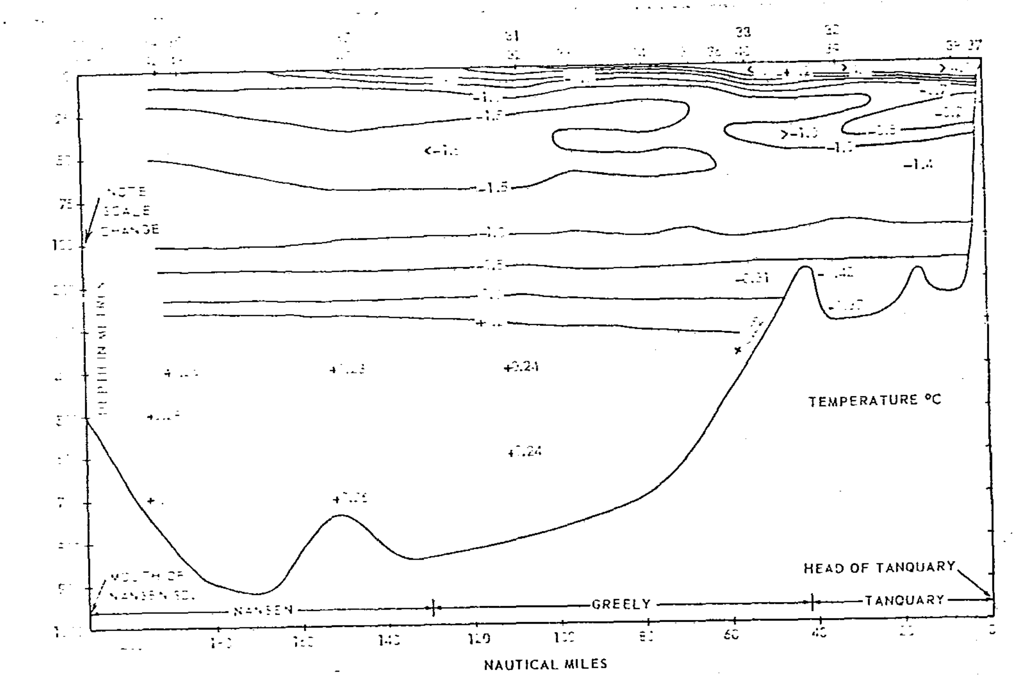


10-13 An assessment of relative amount of summer ice cover in the Nansen Sound - Greely Fiord system. See text discussion of rationale.

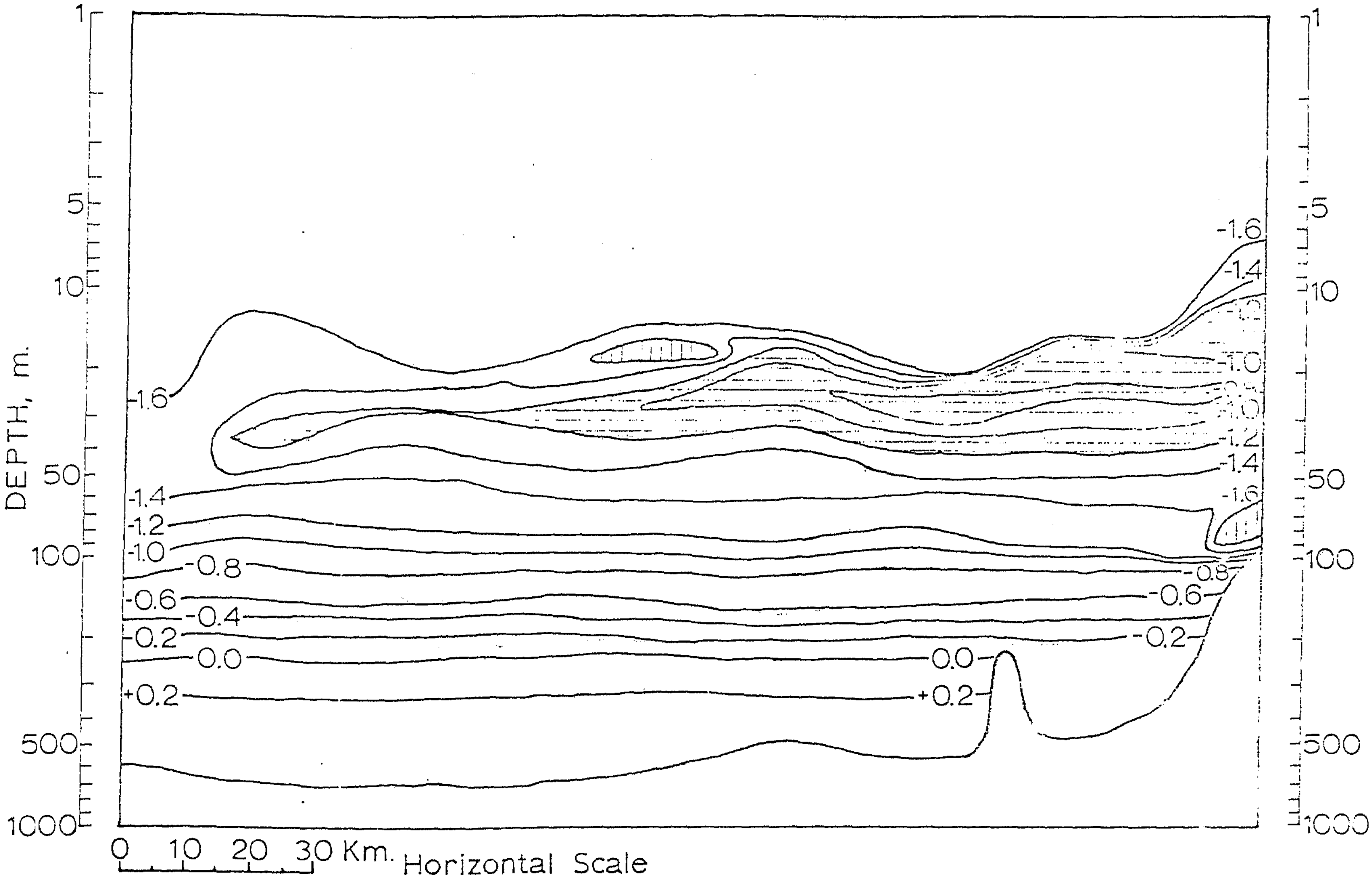
TABLE 10-3

Monthly mean values of selected meterological parameters at Eureka, N.W.T., mostly after Atmospheric Environment Service publications. A, Mean air temperature 1947–70 (C). B, Average total precipitation (centimeters of water 1947–70). C, Median depth of snow cover at month's end (centimeters). D, Average of all wave net radiation (calories per square centimeter per day) 1969–72. F, Average absorbed solar radiation over sea ice (open water during August) (calories per square centimeter per day). G, Average wind speed (miles per hour) 1951–60.

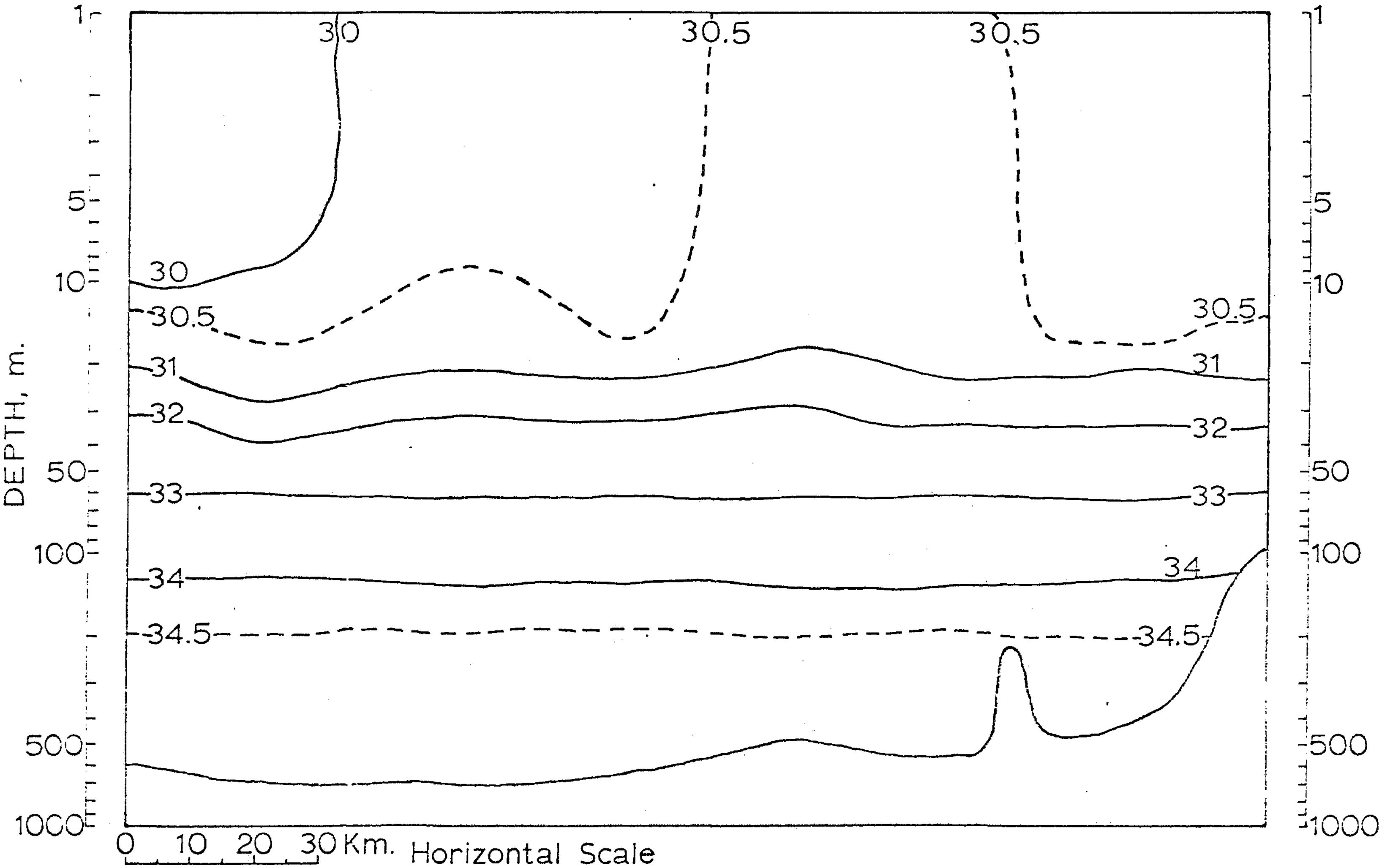
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
A	−37	−38	−37	−28	−10	2	6	3	−8	−22	−31	−35	−19
B	.3	.2	.2	.2	.3	.4	1.3	.9	1.0	.6	.3	.2	5.8
C	15	18	18	18	15	—	—	—	5	10	10	13	—
D	5.6	5.6	6.1	7.4	5.1	—	—	—	1.3	3.3	4.6	4.3	—
E	−35	−30	−30	0	80	310	240	140	10	−20	−40	−30	—
F	0	0	14	70	138	210	345	225	35	2	0	0	—
G	7.2	6.6	5.4	5.8	8.4	10.9	11.3	9.6	7.8	6.6	6.2	5.4	7.6



10-14 Profile on water temperature distribution from the head of Tanquary Fiord to the mouth of Nansen Sound in August 1962 (after Figure 3 of Ford and Hattersley-Smith, 1965).



10-15 Cross section of water structure from upper d'Iberville Fiord to eastern Nansen Sound, March 1976, (a) Temperature (°C),

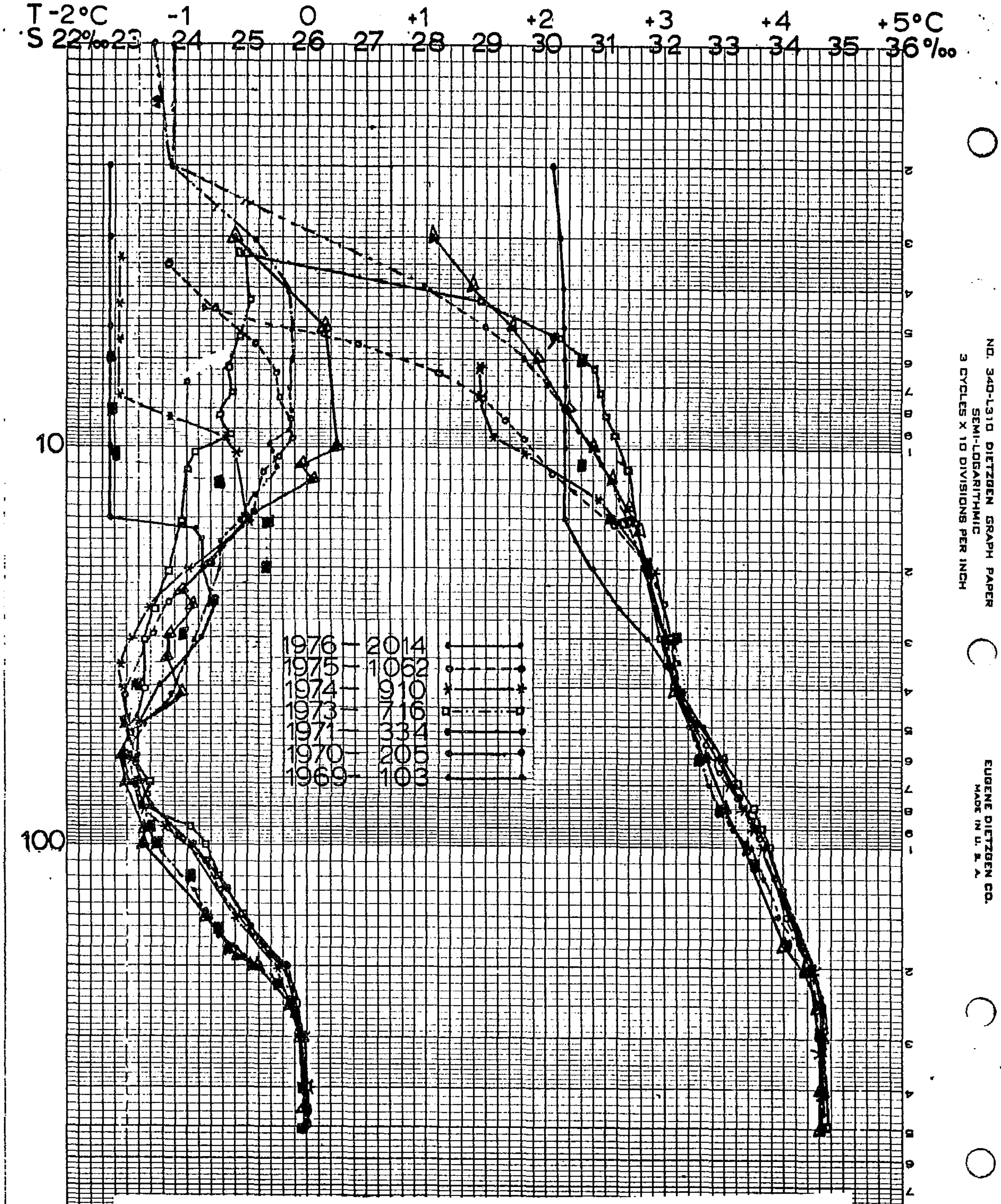


10-15 Cross section of water structure from upper d'Iberville Fiord to eastern Nansen Sound, March 1976, (b) Salinity (°/°°), In the upper reaches of d'Iberville Fiord an additional small feature, a cool layer at a depth of about 80 m, occurs. It has been discussed by Lake and Walker (1976) and was attributed to cooling under the floating ice tongue in upper d'Iberville Fiord. Probably one of the longest time series, profiles taken annually in the same spot, is that from central d'Iberville Fiord. In Figure 10-16 are shown temperature and salinity profiles for this location from 1969 to 1976.

A pair of simple one-dimensional models were coded. The first was the Maykut-Untersteiner (1971) one-dimensional model for growth of sea ice. The results, when best estimates of meteorological parameters were input, indicated that, in years when the surface heat budget was somewhat cooler than normal, the sea ice in the Greely Fiord system would not melt completely. In years warmer than usual the sea ice would just melt. This is observed as noted above.

The other model was a simple one-dimensional model of the vertical water column. Advection was neglected. Potential energy of the column was conserved as were heat and salt. In each cycle (of arbitrary length but usually one day) the water profile of density, salinity and temperatures were mixed downward to the energy balance necessitated by wind stirring. Then heat increments were added, over the mixed layer for eddy heat quantities, and over the water column for solar heat. After adjustment of water temperatures any salt addition was made and the profile adjusted again for energy balance. Cycling ended after an arbitrary time had elapsed, or if the mixing had reached the sea bottom. The density (ρ_W) in the model was a function of salinity (S) only since $\frac{\partial \rho_W}{\partial T} \approx 0.04$ Kgm^{-3}(°C)$^{-1}$ while $\frac{\partial \rho_W}{\partial S} \approx 0.80$ Kgm^{-3}(°/$_{\circ\circ}$)$^{-1}$ and variations in temperature (T) were small. Wind stirring caused energy changes of 0.2 (kgm^{2}sec^{-2}m^{-2}day^{-1}) for a wind speed of 1 m sec^{-1} and 200 for a wind speed of 10 m sec^{-1}. Salt releast from sea ice growing at 1 cm day^{-1} in water of salinity 25 gkg^{-1} caused energy changes of 10-20 (kgm^2 sec^{-2}m^{-2}day^{-1}) while over brief periods in autumn evaporation would cause energy changes of about 5-10 kgm^{2}sec^{-2}m^{-2}day^{-1}. The latter were neglected.

The models were run on initial profiles taken in Greely Fiord and d'Iberville Fiord by the Frozen Sea Research Group in 1974-1976. Inputs simulated seasons with open water and seasons in which ice cover persisted. Meteorological inputs simulated open water seasons in which winds were strong and in which winds were weak. No startling results were obtained. The water structure was modelled to an accuracy which reflected the expected profiles of temperature and salinity which we had observed. The effects of open water in d'Iberville Fiord is shown schematically in Figure 10-17. Upon nominal summer profiles of temperature and salinity are imposed the very shallow freeze-up and late winter changes appropriate to a calm open-water summer. The deeper changes following a windy open-water summer are also shown. Note the difference in height and temperature of the warm "nose" between these two cases. In real life the winter temperature nose, particularly in the shallow case, is not as sharp as depicted, some heat from it passing upward to balance the surface heat budget. However the large differences we have observed in surface water structure in d'Iberville Fiord is to be contrasted to that in Cambridge Bay (in Figure 7-9) which is much the same each year. The ice-in case is similar to the calm open-water situation, although the heating is not



10-1b Temperature and salinity profiles from central d'Iberville Fiord from 1969 - 1976 inclusive. Data from Frozen Sea Research Group.

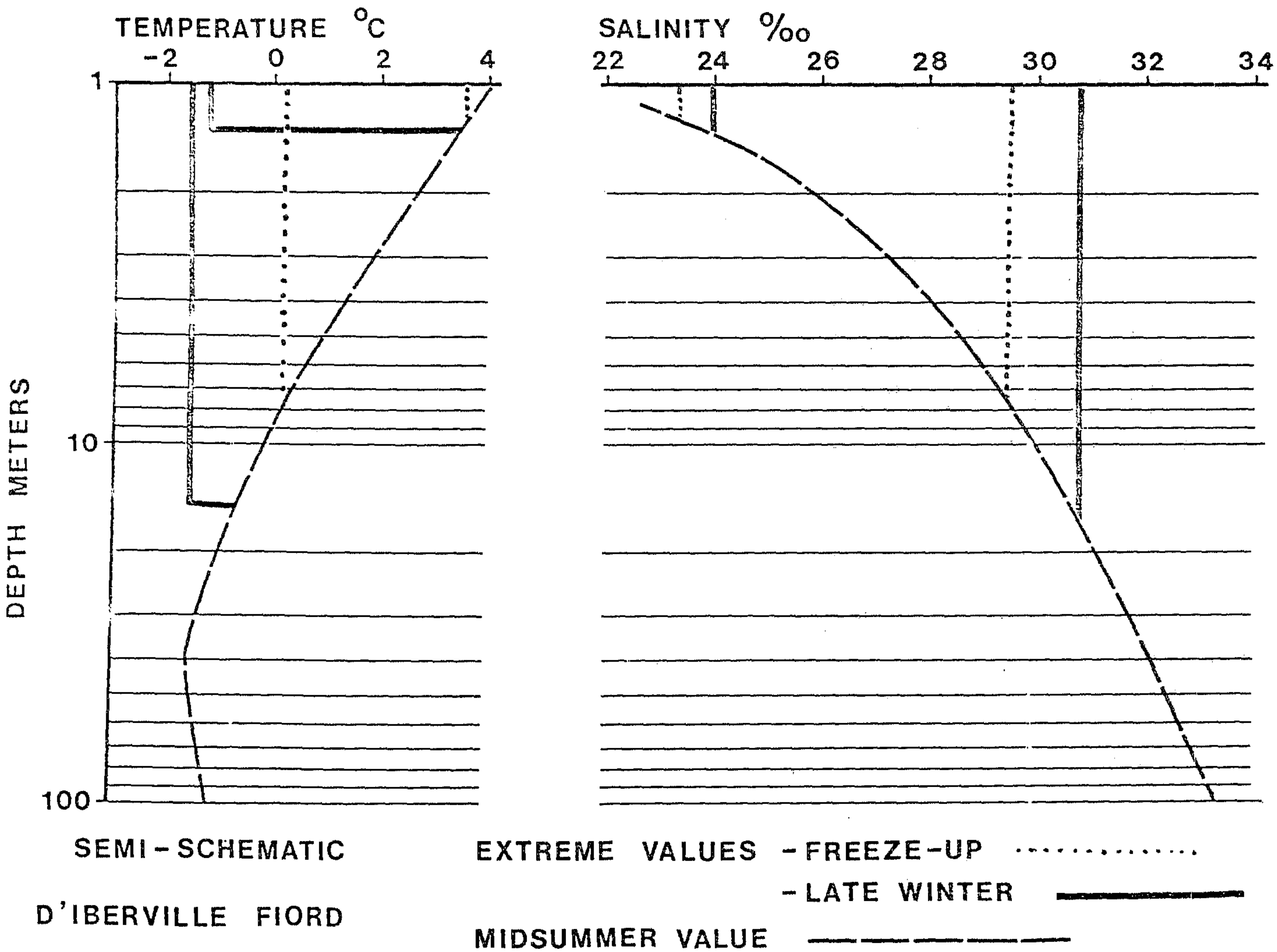
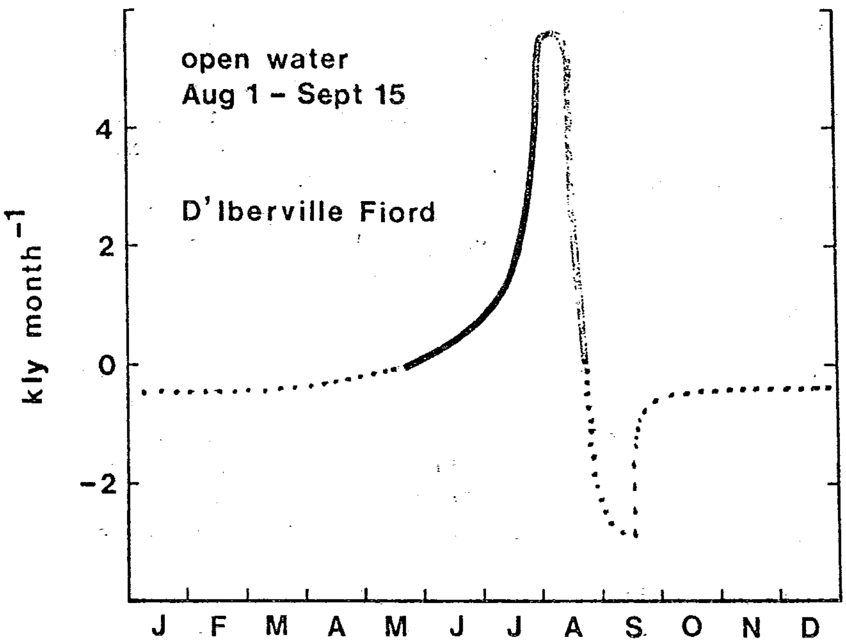


Fig. 10-17. Semi-schematic of water structure in d'Iberville Fiord. Dashed line shows typical mid-summer values. Short dotted lines show upper structure after ice-free windless early autumn while long dashed lines show changes by freeze-up in a windy summer. Solid lines show upper levels after winter convection.

as great. In Greely Fiord, where the runoff and ice-melt fresh water input tend to be less the water column is less stable than the d'Iberville Fiord situation depicted in Figure 10-17 and the winter time mixing occurs to greater depths.

Probably the main benefits from the model results were in helping elucidate the heat budgets of water column under different conditions. As has been mentioned earlier, the accuracy of and confidence in budgets for most quantities in the archipelago tends to be low. The model results pointed out the importance of the heat inputs of the upper reaches of the system. If for example, the less ice-prone upper reaches in Figure 10-13 are considered to be ice free in August the heat input to the water column there is enough to ensure an annual heat increment of about 4 kcal cm^{-2}. If however, a windy autumn occurs, eddy heat losses can bring the annual heat increment in the water column down to about 1 kcal cm^{-2}, roughly what might be found under ice-in conditions. This situation can be seen, semi-schematically in Figure 10-18. Most of the year the water column is losing heat. In spring heat gain occurs and dramatically increases if the ice clears, as shown, at the beginning of August. However, by late August the eddy heat losses are increasing, and if open water continued into September, losses become appreciable. If areas are considered for a year in ice conditions as shown in Figure 10-13, about half the heat input to the Greely Fiord system east of Nansen Sound occurs in the upper reaches. The heat input by runoff appears to be an order of magnitude less than solar input. Inter-annual variability is appreciable in the conditions noted just above, particularly in ice coverage, but the same processes occur to greater or lesser extent each year and lead to the same features.

The water structure features shown above are of course the result of one dimensional processes upon which are superimposed other processes. Horizontal mixing caused by the weak tidal currents present. There must be an estuarine circulation, particularly in summer since average depth of runoff laid down in the upper reaches of the system is double that in the main body of Greely Fiord. Local currents suggesting such a summer circulation have been measured by Herlinveaux (1974). However in d'Iberville Fiord we have been unsuccessful in measuring an appreciable estuarine circulation. It must on the whole be very weak, as is also suggested by the level character of the fields in Figure 10-15. Unfortunately the cross sections, profiles and model outputs do not define (to the author at least) any useful details of the estuarine circulation in the Greely Fiord system, although chemical studies of d'Iberville Fiord presently underway may provide some clues. However a one-dimensional in the vertical Fickian diffusion model was coded. For use of eddy diffusion coefficients of size 5 x 10^{-6} to 5 x 10^{-5} $m^2 sec^{-1}$ suggest that the temperature minimum at 40-60 m could not persist from year to year without a constant inflow of similar water at that depth. Similarly the temperature maximum above must be reinforced every year or it too would not persist.



Heat change in the water column

Fig. 10-18. Semi-schematic of modelled heat loss in water column in d'Iberville Fiord for open water August 1st to September 15th. The dotted part of the curve denotes heat loss.

Cambridge Bay

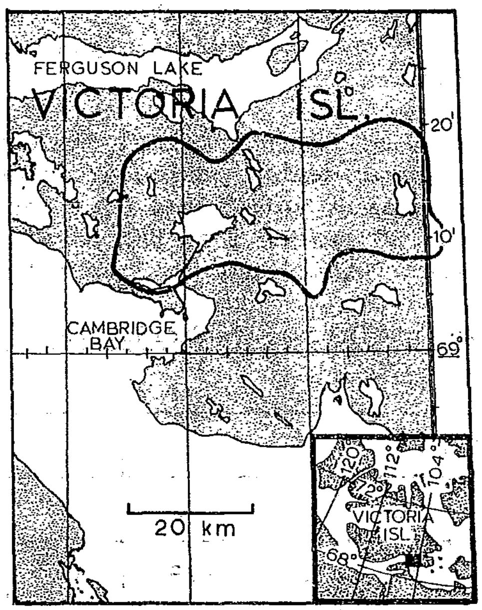
Although oceanographic studies have been sporadically carried out in Cambridge Bay for many years the most intensive program was that of the Frozen Sea Research Group, in the season 1971-1972. The results have been reported by Gade et al (1974) and the following is a brief summary of that paper.

Cambridge Bay is on the southern shore of Victoria Island, itself in the southern region of the Canadian Arctic Archipelago. The source water of Cambridge Bay flows eastward from the Beaufort Sea through the Gulf of Amundsen and Coronation Gulf while being modified by outflow of northward flowing continental rivers, particularly the Mackenzie and Coppermine. As may be seen from Figure 10-19 the southern part of Victoria Island is of low relief, consisting of Palaeozoic lowland. The bathymetry of Cambridge Bay and its entrance is indicated in Figure 10-20. The bay itself trends east-west and is separated into two sections by a sill of minimum depth 48 m. The greatest depth in the eastern part is 86 m and the greatest depth in the western part 57m. The wide entranceway is extremely shallow with two sills located at AA and BB on the figure of maximum depths 20 and 11m, respectively. Between the two sills, water depths are as great as 31 m.

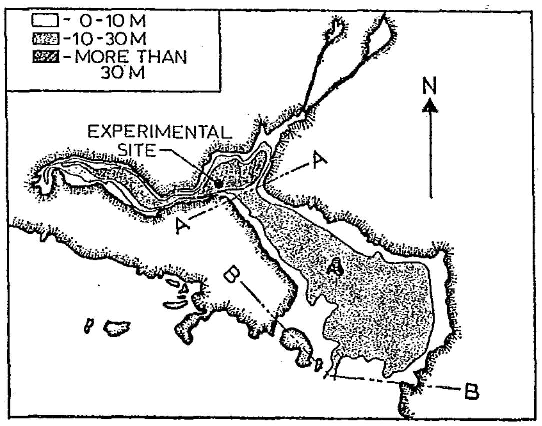
Tides at Cambridge Bay are mainly semi-diurnal. Daily range of heights at highest tides is 0.6 m while the average daily range is 0.4 m. In Cambridge Bay sea ice is formed annually in open water. Over the years 1958-1971 the mean date of complete ice cover in the bay as reported by the Atmospheric Environment Service (A.E.S.) of the Department of the Environment, Canada was 6 October, the earliest date for complete ice cover 25 September in 1965, the latest, 19 October in 1962. The sea ice thickness increases until May, with the average maximum thickness measured by the A.E.S. being about 2.0 m. Melting commences in June and by the second half of July the sea ice has usually completely cleared from Cambridge Bay. During much of the winter the sea ice is covered by snow to variable depths, usually less than 0.3 m.

In summer the water column has a pronounced two-layer structure, freshened in the upper 20-30 m from run-off and ice melt. Salinities in the upper layer can approach 5°/₀₀ and surface temperatures can reach 5°C. Wind mixing is of major importance in determining the vertical details of this upper layer. Below depths of 20-30 m the water column salinities have been about 29.5 °/₀₀ znd temperatures about -1.2°C. After freeze-up, salt released by the growth of sea ice causes an increase in the salinity of the upper layer, which becomes isohaline and isothermal early in the winter. The salt-driven convection continues to cause downward mixing throughout the winter. By spring this process may produce a condition approaching neutral static stability throughout the water column.

Oceanographic measurements were made in the periods 8 September - 29 October, 1971; 29 November - 5 December, 1971; 15 - 26 February, 1972; and 8 - 15 April, 1972. Temperature and salinity profiles of the water structure were taken daily and, on occasion, at rapid intervals; time series of temperature structure in the water column, including a very detailed high speed scan of the upper 5 m, were made using thermistor chains installed before freeze-up.



10-19. Cambridge Bay. The area of the large scale map is indicated on the inset small-scale map. The region bounded by the heavy black line on the large-scale map is the drainage basin of Cambridge Bay.



10-20. Bathymetry of Cambridge Bay showing sills AA and BB at entrance.

Some meteorological data were taken to supplement those available from the A.E.S. station at Cambridge Bay settlement using conventional meteorological instrumentation. The thickness of the sea ice and its snow cover were observed by direct measurement and were a supplement to those available from the local weather station.

Mean monthly values of meteorological elements measured by A.E.S. at Cambridge Bay during the experimental period are shown in Table 10-4. The late summer and autumn were warmer, wetter and cloudier than usual. However, from the end of October until after our last field trip in April the weather was colder, drier and somewhat less cloudy than normal.

Table 10-4. Meteorological conditions, Cambridge Bay, 1971-72 (after A.E.S.).

	1971						1972			
Month	J	A	S	O	N	D	J	F	M	A
A	13	9	3	-4	-22	-29	-32	-35	-28	-27
B	0	0	16	134	663	899	985	1026	870	802
C	407	277	49	0	0	0	0	0	0	0
D	3.3	7.4	1.8	1.6	0.8	0.9	0.6	0.3	0.8	T
E	0	1.5	10.4	25.4	11.9	14.0	9.9	3.0	13.0	T
F	6.9	8.4	8.6	8.3	4.6	5.2	5.5	3.8	4.8	1.8
G	E13.6	E6.5	E3.8	1.7	0.3	0	0	1.4	5.8	13.4

A. Mean air temperature (°C)
B. Freezing degree days (°C)
C. Thawing degree days (°C)
D. Total precipitation (cm water equivalent)
E. Snowfall (cm)
F. Mean cloud cover (tenths)
G. Global solar radiation (kcal cm^{-2})
T = Trace
E = Estimated

The temperature and salinity structure in the water of Cambridge Bay during the experimental season has been shown in Figure 7-9. From Figure 7-9 one can see that surface water temperatures were above +3°C in early September. By mid-September water column cooling commenced. About 5 kcal cm^{-2} were lost by freeze-up. Surface temperatures had dropped below -1°C by freeze-up but temperatures of +1°C persisted near 20 m.

After freeze-up, the water column was cut off from the atmosphere and salt-induced convection commenced. Until mid-December, the depth of the convective layer at the sampling site was moving downward through the layer mixed earlier by the wind. In December, 1971, the convective layer passed below the wind-mixed depth (20 m) and during the February and April sampling periods the convective layer, cooling and with increasing salt content, was subject to less variation from cast to cast in both temperature and salinity than had been the case earlier in the winter.

The water column in early September was stable from the surface to a depth of 25 m. From 25 m to the bottom the water was nearly neutral, possibly due to flow into the bay. By the end of October, wind mixing had reduced stability from the surface to about 15 m. From 30 m to the bottom of the bay the water had near neutral stability. During the winter and spring experiments, both the mixed layer and the bottom layers in the bay were very slightly stable.

To supplement the curves in Figure 7-9 the seasonal isotherm pattern is shown in Figure 10-21 to illustrate some of the points mentioned above.

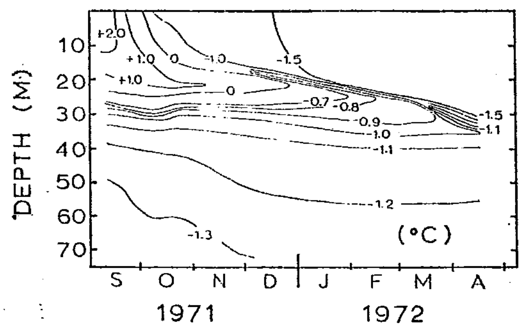


Fig. 10-21. Seasonal isotherms in Cambridge Bay, winter 1971 and 1972.

Oceanographic observations and measurements of ice-snow thicknesses and temperatures were used to estimate the heat flow upward to the surface of the water or ice. During the autumn, the water column heat loss was the only component (Q_W). After freeze-up, computation of the heat furnished by formation of ice (Q_{if}) used values of ice density $\rho_i = 0.9$ g cm^{-3} and heat of fusion $L_i = 68$ cal g^{-1}.

Components of the meteorological heat furnished to the surface from above were computed on a daily basis. From the beginning of September until freeze-up on 12 October daily mean values of fractional coverage of low and mid-cloudiness, wind speed, air temperature and air vapour pressure were taken from A.E.S. reports. The sea surface 'bulk' temperature was taken from our

Table 10-5. Heat budget components, Cambridge Bay, 8 September, 1971 to 12 October, 1971 (kcal cm^{-2}).

$Q\uparrow$	$Q\downarrow$	Q_S	Q_L	Q_H	Calc. heat loss	Q_W
24.3	17.7	3.9	2.7	2.5	7.9	5.1

For explanations of symbols see caption of Table 10-6.

readings. Conditions were assumed to be homogeneous over the bay and no allowance was made for the bay's finite size. The values of clear sky infra-red radiation were calculated from the twice daily radio-sonde ascents at a site about 3 km north of the raft sampling site. The sum of the five components of the meteorological heat budget listed in Table 10-5 for 8 September, 1971, to 12 October, 1971 inclusive indicate about 8 kcal cm^{-2} must be furnished from the water to attain balance. From the oceanographic heat budget the water column heat loss during this same period was about 5 kcal cm^{-2} difference. An important reason for this discrepancy is undoubtedly the skim of ice present on the water surface at intervals before freeze-up. This would tend to reduce water column heat loss. The water column heat loss is accurate to 1 kcal cm^{-2}, considering instrumental errors, runoff and some sewage inflow.

Terms in surface heat budgets, cumulative from freeze-up, 12 October, 1972, to dates of our field observations are shown in Table 10-6. The calculated solar radiation agreed to within 10% of that measured by the A.E.S. at Cambridge Bay from freeze-up to April. It was not possible to compare in a meaningful way the net radiation calculated here with that measured by A.E.S., for their network was too sparse. The accuracy of our meteorological budgets may be estimated by comparing them with the heat lost from below the surface.

Table 10-6 Cumulative heat budget from freeze-up, 12 October, 1971, Cambridge Bay (kcal cm^{-2}).

	$Q\uparrow$	$Q\downarrow$	Q_s	$(Q_L + Q_H)$	*Calc. heat loss*	*Measured heat from below surface* Q_w	Q_{if}	Q_{ic}	*Sum*
Oct. 29/71	10	9	0·3	0·2	2	1·4 (47%)	1·4 (47%)	0·2 (6%)	3
Dec. 2/71	26	21	0·5	0·2	4	2·2 (38%)	3·1 (53%)	0·5 (9%)	6
Feb. 24/71	57	46	1·0	0·2	10	3·6 (26%)	8·6 (62%)	1·7 (12%)	14
Apr. 14/72	77	60	5·1	0·9	13	4·1 (26%)	9·7 (62%)	1·9 (12%)	16

$Q\uparrow$ Long wave radiation flux upward from surface.
$Q\downarrow$ Long wave radiation flux downward to surface.
Q_s Absorbed solar radiation.
Q_L Latent heat flux from evaporation upward.
Q_H Sensible heat flux upward.
Q_w Water column heat loss.
Q_{if} Heat of ice formation.
Q_{ic} Heat released by ice cooling.

If we estimate the heat fluxes in Cambridge Bay for late spring and summer and combine them with our measurements for the winter season, then on an annual basis the heat budget appears to be in balance to within the accuracy of our estimates. The ice forms and largely melts in situ. The mixing from runoff appears to be very small insofar as heat is concerned. The main components of water column heat balance are solar gain in summer versus long wave losses all year and latent and sensible heat losses in autumn.

Computations of total salt content for the period September and October showed large day-to-day variations. More detailed computations were carried out using data obtained in Cambridge Bay in December, February, and April. The bay was considered to extend seaward as far as the outer sill wherethe maximum depth is 11 m. The calculation shows that the total salt mass increases by 1.4% from December to February and 0.9% from February to April. An analysis of possible errors in these estimates due to scatter of data, uneven ice thickness and inaccuracies in water volumes and salinity values indicate that these changes are not significant.

Thus from December to April Cambridge Bay may be considered to be a closed basin with no detectable net exchange of heat and salt with water outside the bay. This condition cannot extend to the summer period when the bay is partially or wholly ice-free. Surface currents of the order of 10cm s^{-1} were noted during the summer observation period. Thermistors measuring the temperature of the water column recorded the downward deflection of isotherms in response to water being set up in the bay by winds with a strong southerly component. The isotherms returned to their previous positions when the winds abated. In such periods exchange of water with the outside can be expected. Figure 10-22 is a TS diagram showing the characteristic water mass below 30 m for Cambridge Bay and outside seas. Water above 30 m was not included, due to the large scatter in values caused by fresh water from runoff and melting sea ice. Compared to the outside, Cambridge Bay water is characteristically of

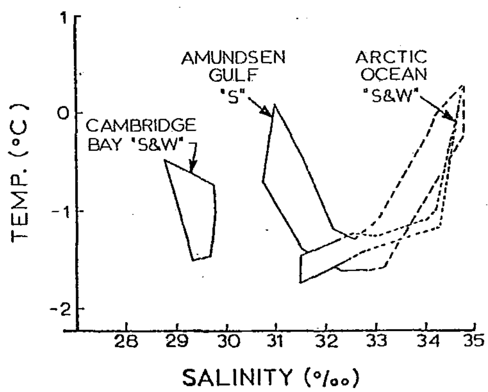


Figure 10-22. The water masses of Cambridge Bay, Amundsen Gulf (Cameron, 1953) and the Arctic Ocean (Worthington, 1959). Solid lines enclose TS values in the depth interval 30-70 m and dotted lines 70-300 m. S and W indicate summer and winter values. No winter values are available for Amundsen Gulf.

the same temperature but 2 or 3‰ fresher than water at similar levels. On this basis, it is expected that the exchange of water between the bay and outside seas would result in the importation of salt to the bay without causing a significant exchange of heat. The 14 September salinity profile of Figure 7-9 shows a salinity maximum near the 25 m level indicating an incursion of saltier water from outside the bay, the deep inflow of an estuarine circulation system. Sparse observations outside the bay just before freeze-up showed the density structure in the surface layers to be the same as that inside the bay, so we consider that freeze-up marks the cessation of the exchange of water with the outside with the exception of a possible small exchange depending on tidal motion.

Superimposed upon the large-scale seasonal temperature changes (development) exhibited in Figure 21 both non-periodic and periodic disturbances are present. These disturbances are of a wide range of periodicities, from a few minutes to several days.

In order to examine the nature of these disturbances it was convenient to compute the levels of selected isotherms by a linear interpolation technique. Figure 10-23 shows an example seven days before freeze-up, where the local wind registrations are also plotted. Attention is directed toward the rapidly varying depths of the isotherms of the upper levels, reflecting surface convergence and divergence by the sustained wind. The record of Figure 10-24 taken six days after freeze-up, shows a reduced level of high frequency disturbances, the variations in isotherm depths now being clearly related to the tidal cycle which is also shown.

The isolated temperature maxima seen in Figure 10-24 do not indicate a change of phase above this level, but is likely to be caused by uplift of the underlying temperature maximum seen in Figure 7-9.

Vertical eddy exchange must take place at all levels throughout the year to account for the changes seen in Figure 7-9. In the upper levels, energy for this exchange may be furnished by wind disturbances in summer and salt-induced convection in winter. Eddy exchange also takes place in the lower levels of the basin, although not so energetically as in the upper levels. Energy to drive this process could originate in convection reaching the bottom of the bay or from tidally caused disturbances.

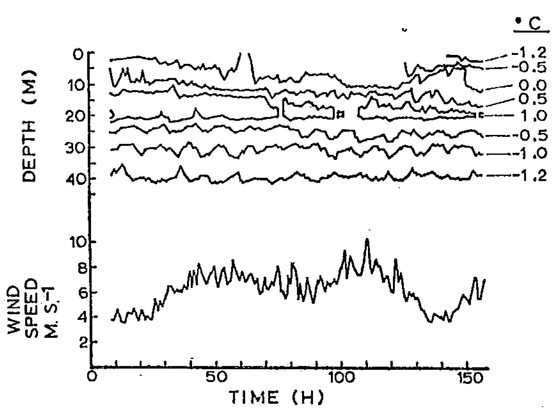


Figure 10-23. Isotherm depths in Cambridge Bay before freeze-up compared with wind speeds. Record commences October 5, 1971.

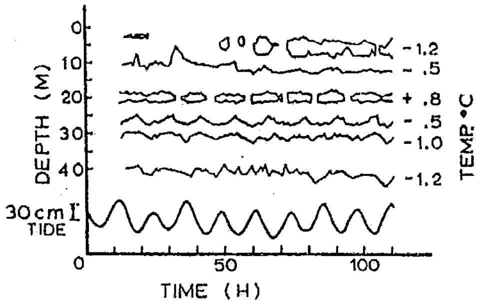


Figure 10-24. Isotherm depths in Cambridge Bay after freeze-up compared with tide levels. Record commences October 18, 1971.

Consideration of the budgets and processes within the water column led, as described by Gade et al (1974) to a winter circulation picture as shown in Figure 10-25, in which water from shallow parts of the bay penetrated into the lower reaches, below the mixed under-ice layer in which convection was occurring.

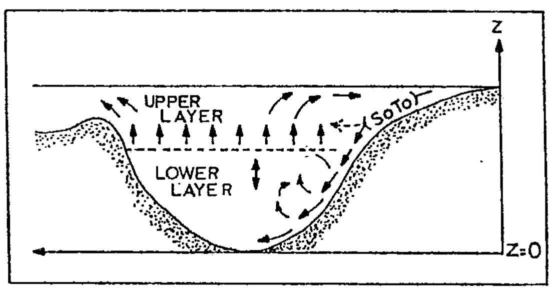


Figure 10-25. Schematic diagram of circulation in Cambridge Bay during fall and winter caused by a flow of dense water from the shallows penetrating the main pycnocline.

The volume flux of the deep-reaching convection was used to estimate the mass of salt transported from the shallows. Assume that the salinity of water entering the shallows is that of the under-ice mixed layer and the salinity of water entering the shallows is that of the under-ice mixed layer and the salinity of the water leaving the shallows to enter deep-reaching convection is $29.65^{\circ}/_{\circ\circ}$. Then the removal of excess salt from the shallows in deep convection currents would be about 2×10^{10} g over December-February, and 2×10^{9} g over February-April. For the earlier period this is about 30% of excess salt from the shallows in the inner bay, and 7% of salt formed in shallows of the whole bay to the outer sill. In the period February-April the percentages are 10 and 2% respectively. This implies that most of the excess salt in the entraceway does not enter the inner bay. It probably remains, through the winter, ponded between the two sills. The percentages also support the suggestion above that a larger fraction of salt from the shallows moves horizontally in the under-ice layer as the winter progresses. This circulation is shown in the dashed arrow in Figure 10-25.

Our study of Cambridge Bay has indicated that the outward flow of fresh and brackish water in the surface layers associated with summer runoff is balanced in part by an influx of salty water at a depth of about 25 m in the late summer. The actual salinity of Cambridge Bay during the winter months has varied from year to year, and this reflects differing salt circulations in response to variations in local oceanographic and meteorological conditions. The great influence of the wind is apparent. One gale immediately before freeze-up could influence water structure for the following nine months.

Although the external exchange of salt is small or nil through the ice-in season there is a lively movement of salt within the bay. Short-term variability of the temperature field superimposed on its seasonal changes indicates considerable water movement after the sea ice cover is complete. Probably much of the energy for this is supplied by convective processes, but there is also a spectrum component directly associated with the tides. The under-ice convection is most noticeable causing a very well-mixed layer of increasing depth during the winter season. However, heavy, salt-rich water from the extensive shallows moves throughout the bay both in the upper mixed layers and also to the bottom of the bay. The lowest levels of Cambridge Bay are stirred annually. The budgets of this lower circulation were modelled successfully using physically reasonable values of the eddy diffusivities of heat and salt.

Our observations indicate that after freeze-up a 'double layer' structure persists just under the ice beyond the time when any such layering due to wind 'set up' processes would have disappeared. It must reflect the nature of the convective processes in early winter. Half the total annual heat loss from the waters of the bay occurred in the month before freeze-up, and it is probably that this proportion would also apply to most Arctic areas where the ice is totally removed during the summer period. As might be expected, the bay behaved, as far as heat loss is concerned, as a combination of an open lead during the summer and the Arctic Ocean during the rest of the year.

BUDGETS

In making a budget one is evaluating a conservation equation of form

$$\frac{D\phi}{Dt} = \frac{\partial \phi}{\partial t} + \mathbb{V}\cdot\nabla\phi = J_\phi$$

where $\frac{D}{Dt}$ is the rate of change in a parcel, $\frac{\partial}{\partial t}$ is time change at a fixed point, $\mathbb{V}\cdot\nabla\phi$ is the change at a point due to advection and J_ϕ includes turbulent exchanges, and sources and sinks of the scalar ϕ. In what follows we may imagine the equation evaluated over the area of the archipelago and over a period of time for which $\frac{\partial \phi}{\partial t}$ may be small enough to be neglected. Turbulent exchange will also be neglected so that advection may be balanced against sinks or sources. Usually of course it is impossible to measure or estimate all terms sufficiently accurately, but frequently the exercise of trying to obtain a balance is illuminating.

One of the common exercises is estimating the flow of sea water through the archipelago. As will be shown below the sinks and sources may be neglected and only the advection measured or estimated. This has been done by those referred to in Table 8-1 of Chapter 8. As may be seen the latest estimates average about 6 x 10^4 km^3 outflow southward through the southeastern archipelago. The latest estimates are based in part on actual current measurements, earlier ones on temperature salinity measurements across channels, or perhaps residual terms in Arctic Ocean budgets such as shown in Table 2-1, although in that table the archipelago fluxes were estimated independently of the overall budget. The latest estimates (Sadler 1976) seem to be somewhat higher than earlier estimates. The freshwater budget has been considered from an oceanographic viewpoint by Huyer and Barber (1970) who show that most of the freshwater in the water column has been advected from the Arctic Ocean surface layers. The freshwater input within the archipelago has been briefly considered in Chapters 2 and 5. To extend the material there, in a crude way, we have in the archipelago northwest of Baffin Island a water area of 8 x 10^5 km^2. From Figure 4-9 we might estimate precipitation as 0.14 m, from Figure 4-14 evaporation of 0.05 m, and from Figure 5-2 runoff of 0.07 m. The freshwater volume input over the northwestern archipelago is then 130 $km^3 year^{-1}$, much smaller than the sea water fluxes through the archipelago. Using roughly the same techniques over Baffin Bay with Walmsley's (1966) estimate of 0.27 m evaporation gives a freshwater input of 30 $km^3 year^{-1}$ there.

The conservation of salt in sea water has been considered along the lines indicated in Table 201. Sadler (1976) has recently estimated the flux of salt southward through Nares Strait as 6.7 x 10^{11} tonnes $year^{-1}$ which compares with the estimate in Table 2-1 of 23 x 10^{11} tonnes $year^{-1}$ for the whole archipelago. Sea ice net transport southward through Davis Strait has been estimated by Vowinckel and Orvig (1962) as 491 km^3 $year^{-1}$. Lindsay (1968, 1969) has attempted to obtain a sea ice budget over the Queen Elizabeth Islands for the average ice year 1966. His estimates, I believe, indicated that the northwestern archipelago is a weak sink of sea ice. Barber et al (1975) estimated that about 350 km^3 of multiyear sea ice melted in the southeastern Beaufort Sea in each of 1969, 1970. Lindsay's figures for average summer ice cover (50%) can be combined with the winter ice growth figures discussed earlier (2.0 m from open water, 1.2 m from old ice areas) to give an estimate of about 1300 km^3 of sea ice formed annually in the northwestern archipelago, within the area shown in Figure 5-1. Similarly, using Walmsley's (1966) figures for average ice coverage in Baffin Bay it can be crudely estimated that about 1000 km^3 of sea ice forms in Baffin Bay anually. Sadler (1976) estimates about 110 km^3 of sea ice passes south through Nares Strait annually. These figures taken together indicate that advection of sea ice through the archipelago is smaller than the volumes forming and melting there each year. It is to be noted that ice export southward through Davis Strait implies a fresh water transport southward, and a transport of 'cold' southward, or of heat northward.

Heat budget studies have been of two types. The first is of the sort producing the results shown in Table 2-1, with transport of a volume of water at a certain temperature resulting in a certain heat flux. The other type of study is a localized effort to evaluate the terms in the heat budget. These, exemplified by studies of Huyer and Barber (1970) for the heat budget of Barrow Strait and Walmsley (1966) for Baffin Bay, are somewhat similar to the type described in chapter 10. As indicated in Table 2-1 the heat transport by water movement, shown in Table 2-1 is 4×10^{16} kcal $year^{-1}$. Vowinckel and Orvig (1962) have estimated the (northward) heat flux caused by ice transport through Davis Strait as 4×10^{16} kcal $year^{-1}$ too. Sadler (1976) has estimated the northward flow of heat through Nares Strait as 2.1×10^{16} kcal $year^{-1}$ which is probably higher than the earlier estimates when other passages through the archipelago are considered. The results of Barber and Huyer (1970) indicated about 3 kcal cm^{-2} was put into the water column in Barrow Strait in 1962. The results in chapter 10 also seem to indicate that in most years the water column over much of the archipelago has a slight surplus of the order of 1 kcal cm^{-2}. Walmsley's (1966) results for Baffin Bay indicate a higher heat surplus but his confidence in this seems low. If for the northwestern archipelago 1 kcal cm^{-2} is put in over the water area of Figure 5-1 of $8 \times 10^5 km^2$ the total heat input is about 1×10^{16} kcal almost comparable with the transport values mentioned just above, and Baffin Bay was not considered.

In all of the budget studies mentioned above, and in others not mentioned, while the results are useful, they should be firmed up by measurements possible with modern equipment. By this I mean chains measuring water flow, temperature, salinity and pressure in adequate quantities in strategic locations, supplemented by measurements of ice volumes and other quantities, and an increase in localized studies of terms in the heat balance.

DISPERSION OF POLLUTANTS

This section reviews briefly some aspects of pollutant dispersion in waters within the archipelago. Pollutants of interest may include materials which go into solution, freshwater in salt water, gases, solid material from fine particles to large lumps. The pollutant may be put into the sea on the top, bottom or at mid depth, instantaneously, continuously, at one point or over an appreciable area. As was noted in Chapter 8, water currents in the archipelago are caused by mechanisms which act over widely differing periods. Some operate over cycles a year or more in length. Periods of tidally caused currents are mainly semi-diurnal or diurnal. Atmospheric influences cause currents over a wide range of periods from a few seconds upward, although diurnal, multidiurnal or annual periods dominate. The net result is that water movements occur on a wide range of scales in space and time. We may postulate that, after due consideration of the scale of the polluted water parcel, that this parcel is carried bodily by currents of scale much larger than that of the polluted volume, while being mixed or diffused by water currents of scale comparable to or smaller than the polluted volume. The practical problem is usually to decide where the polluted volume is going to be moved, and how rapid will be the mixing with consequent decrease in concentration of pollutant.

Some instances, such as a sea bed gas release, a sea bed fresh water release, or a surface release of heavy material may effect the water circulation over an appreciable area, although it can almost always be disregarded on a geophysical scale. Apart from these cases we may state that the water currents will be caused by the forces embodied in the equations of Appendix A. The pollutant will then be governed by a conservation equation. If we disregard for the moment buoyancy effects the polluted volume will be subject to conservation in form:

$$\frac{DP}{Dt} = \frac{\partial P}{\partial t} + \mathbb{V} \cdot \nabla P = J_P$$

where as in Chapter 11, the parcel change rate may be broken into rate of change at a fixed point and the advective change. The right hand member again includes diffusive and sink/source quantities. The methods of solution of the above equations have been discussed by Csanady (1973). Practical methods usually include separation of flow into a mean flow and a 'diffusing' flow as noted above. Fairly drastic assumptions are needed to evaluate the latter. By analogy with molecular diffusion, we may write a diffusion equation upon which most practical attempts to deal with diffusion are made. It is:

$$\frac{\partial P}{\partial t} + \mathbb{V} \cdot \nabla P = K_x \frac{\partial^2 P}{\partial x^2} + K_y \frac{\partial^2 P}{\partial y^2} + K_z \frac{\partial^2 P}{\partial z^2}$$

where Kx, Ky and Kz are (eddy) diffusion coefficients, with the x direction frequently chosen in the direction of mean flow, y direction normal to the mean flow and z upward in the vertical. If necessary a sink/source term, involving, usually, decay by physical, chemical, biological or radioactive processes may be included.

The problem of estimating the magnitudes of the eddy diffusion coefficients is frequently attacked empirically by relating some parameter such as observed variance (σ^2) of an experimental distribution to the factor K. Frequently Kx is neglected in crude treatments of continuous sources. Experimentally derived values of Ky vary as $K_y \approx \sigma_y^2/2t$. The values of Ky vary widely around 10^2 - $10^5 cm^2 sec^{-1}$ in coastal waters. Ky does not increase with time as quickly as does Kx on the average. The vertical eddy diffusivity coefficient is a strong function of water column static stability. Experimental values vary from 10^{-2} to over $10^2 cm^2 sec^{-1}$. There are other possible mechanisms which may aid vertical diffusion such as the double diffusion or salt finger mechanism. In the case of appreciable shear of the horizontal current with depth, vertical diffusion, although slow, can aid horizontal diffusion. This is because pollutant advected with the stronger horizontal currents is diffused vertically into layers of slower horizontal current, effectively increasing horizontal diffusion there.

Mathematical models have been devised to predict outcomes of pollution episodes. Models tend to be based upon statistical theory in which lower moments of the distribution of pollutants are forecast either for relative diffusion, as of a spreading cloud of pollutant moving in a larger scale current, or for diffusion of pollution as from a continuous source. The quantities of interest are usually the concentration of pollutant or its integral over time, called dosage. The models relate the 'idealized' spread and movement of pollutant to atmospheric or ocean parameters such as mean currents and turbulence levels. Consideration of the possible maximum value of the concentration of pollutant in any one episode as compared to modelled concentrations is discussed by Csanady (1973). Boundaries, absorbent, or reflecting, may be taken into account also possible decay, deposition, or other loss of pollutant. Buoyancy of the pollutant is also included if necessary. If buoyancy effects connected with the pollutant are so marked as to affect the circulation, as with chimney plumes in the atmosphere, heavy material dumping in the sea, the tendency is to model this part of the process separately, and match this model to another diffusion model when the density effects become unimportant insofar as ambient conditions are concerned.

In the Canadian Arctic Archipelago sea ice covers the water surface much of the year. While some pollutants (ie, crude oil) may be incorporated en masse in growing sea ice, most solutes (including sea salt) are rejected by growing sea ice. This rejection of salt causes convective mixing to a depth of a few meters or tens of meters (Figures 7-9, 10-16), with consequent pollutant mixing. In summer shallow surface layers of fresher water will inhibit mixing, but if the water becomes free of ice, winds can cause horizontal and vertical mixing in surface layers. Lower layers of water tend to be saltier and denser than upper layers, and mixing between upper and lower layers, and mixing within the lower layers may be slow. Frictional influences on side and bottom boundaries may cause appreciable mixing in appropriate circumstances. Coriolis effects may cause enhanced currents against lateral boundaries. Otherwise lateral boundaries probably inhibit mixing in the horizontal. The mixing in most arctic fiords, bays and estuaries may be less pronounced than in similar localities elsewhere, since the meagre runoff is not favorable for a strong surface outflow or strong returning circulations. Tides tend to be small over the northwest archipelago although appreciable in the southeast, while ice cover restricts wind influences on mixing. Many of the fiords have appreciable sills which result in deep areas inside them being very stagnant (Lake and Walker, 1976).

Before analyzing and attempting to predict pollutant dispersion for any specific case characteristics of both pollutant and the receiving environment should be known. For the pollutant, parameters might include type of source, quantity of emissions, density of pollutants, rate of decay or deposition and other relevant characteristics. For the oceanographic environment parameters generally necessary will include geometry of boundaries, sea ice cover, relevant meteorological conditions, water density structure and movement over periods long enough to permit determination of characteristic oscillations and variations, ie, the spectral content. The vertical resolution must be such that layers can be resolved and the density structure clearly defined. Other data, when relevant, such as water levels should also be observed. Since seasonal variation is so great in Arctic regions the foregoing implies adequate observations over at least one annual cycle before use of predictive models other than on a very short time scale. Above all the analysis and prediction of pollutant problems should be entrusted to a person well versed in both dispersion theory and the oceanography of the locality.

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APPENDIX A

GOVERNING EQUATIONS

The processes and motions described earlier are governed (although they do not know it) by equations of geophysical hydrodynamics and thermodynamics. A very brief outline of the set of equations is given below as they refer to earlier parts of this note. For a fuller description, Phillips (1966) is a good source.

The equations may be described as conservation statements, relating changes in quantities to factors which may cause changes in these quantities. The conservation of momentum is a statement of Newtons second law

$$\mathbb{F} = \frac{D}{Dt}(m\mathbb{V})$$

which may be written for a parcel upon a rotating earth as

$$\frac{D\mathbb{V}}{Dt} = \mathbb{P} + \mathbb{C} + \mathbb{G} + \mathbb{F} \qquad (A1)$$

where $\frac{D\mathbb{V}}{Dt}$ is the acceleration of a fluid parcel of unit mass, which can, as noted in Chapter 11, be divided into acceleration at a point $\frac{\partial\mathbb{V}}{\partial t}$, and the 'advective' acceleration $\mathbb{V}\cdot\nabla\mathbb{V}$; $\mathbb{P}$ is the pressure force, $\mathbb{C}$ the Coriolis force due to the earth's rotation, $\mathbb{G}$ is the gravity force, and $\mathbb{F}$ represents other (frictional) forces. The equation is three-dimensional, double-lineal characters being vectors. Subject to considerations of scale, imposition of proper boundary conditions and other assumptions as appropriate, equation (A1) forms part of a set of equations by which flow (in air or water) may be described. Some specific formulations and nomenclature of currents include (a) Buoyancy currents (in the vertical) acting through $\mathbb{G}$ and $\mathbb{P}$, with $\mathbb{F}$, $\mathbb{C}$ neglected usually, (b) Gradient currents with $\mathbb{F}$ neglected and $\frac{D\mathbb{V}}{Dt}$ frequently very small (or zero, geostrophic), (c) inertial currents with $\mathbb{P}$, $\mathbb{F}$ neglected but $\frac{D\mathbb{V}}{Dt}$ sizeable, (d) Frictional currents with $\mathbb{F}$ important, (e) Periodic currents as from tides and seiches.

Particularly used, as noted in section 8, is the horizontal geostrophic current with velocity frequently denoted $\mathbb{V}_g$ where the horizontal pressure force is balanced by the Coriolis force. Where the water mass is homogeneous and the pressure force is due to the slope of the water surface the current which is the same through the water column is called 'barotropic'. When the water mass is not homogeneous (but baroclinic), that is the density surfaces do not coincide with the pressure surfaces the geostrophic equation can be differentiated in the vertical and shear found. If for example on a section across one of the channels in the archipelago temperature and salinity measurements are available so that density variations along isobaric surfaces can be calculated then the shear of the geostrophic current perpendicular to the cross section can be calculated. If the absolute value of the perpendicular current can be found at any point on the cross section then the absolute value of the perpendicular current can be found over the whole section. This is illustrated in Figure 8-3 where, if the absolute current at a depth of 750 m can be assumed to be zero, then the 'baroclinic' current is the absolute current normal to the cross section.

By manipulations of equation A1 various other conservation equations can be obtained which are indispensable in certain problems. They will not be discussed here but include conservation equations for mechanical energy, vorticity and other quantities.

Other equations in the set include a thermodynamic equation of state giving, for water, density as a function of temperature salinity and pressure, and for a perfect gas, air density as a function of pressure and temperature. Another equation is that expressing the conservation of mass. Conservation equations, of the form given in Chapter 11 may be written for scalar quantities of interest for example, sea salt. Such an equation, for the conservation of entropy, after drastic simplification, was used in the form

$$\frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2}$$

where is the eddy thermometric conductivity, to estimate temperature changes in the water column in d'Iberville Fiord, as noted in Chapter 10. The equation for the conservation of entropy, drastically pruned in a different way forms the basis for the evaluation of heat budgets at the earth/ice/water/air interface. Such an interface heat budget may be written

$$R = I(1-A) + R{\downarrow} - \varepsilon\sigma T_s^4 = H + LE + S$$

where R = radiation balance

I = incoming solar radiation

A = albedo

$R{\downarrow}$ = incoming long wave radiation

ε = surface long wave emissivity

σ = Stefan-Boltzmann constant

T_s = surface temperature (°K)

H = sensible heat exchange

L = latent heat of evaporation

E = evaporation

S = storage of interface element of soil/water/ice heat flux (if any).

The uses of this relationship are described in chapters upon meteorology and upon the heat budget of the water column in fiords. An example given in Chapter 4 above relates the annual fresh water balance and the annual surface energy budget, to ensure compatibility and (usually) to confirm estimates of quantities difficult to measure directly. The surface heat budget $R = H + LE$ and a conservation equation for fresh water, $P = N + E$ where P is annual precipitation, N is annual runoff and E is annual evaporation, may be combined in forms such as

$$(1 + B)(1 - C) = D$$

where the ratios $B = H/LE$ (Bowen ratio), $C = N/P$ (runoff ratio) $D = R/LP$ (dryness ratio), may be used to characterize different regimes.

APPENDIX B

Because different systems of units are used in the text a brief conversion table follows.

Length

1 meter (m)	= 3.28 ft
	= 39.37 in
	= 0.55 fathoms
1 kilometer (km)	= 0.62 miles

Area

1 kilometer squared (km^2)	= 0.386 miles squared (mi^2)

Volume

1 cubic meter (m^3)	= 35.3 ft^3
1 cubic kilometer (km^3)	= 0.24 mi^3

Velocity

1 meter per second ($m\ sec^{-1}$)	
	= 3.28 $ft\ sec^{-1}$
	= 2.24 $mi\ hr^{-1}$
	= 3.60 $km\ hr^{-1}$
	= 1.94 knots

Mass

1 kilogram (kg)	= 2.2 lb
	= 10^{-3} tonne

Pressure

1 atmosphere	= 101.3 Pascal (pa)
	= 1013.3 mb
	= 10.3 m of water

Heat Flow Rate

1 $cal\ cm^{-2}min^{-1}$	= 1.0 langley (ly) min^{-1}
	= 698 $watt\ m^{-2}$

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APPENDIX C

GEOGRAPHICAL LOCATION OF PLACE NAMES IN TEXT

Admiralty Inlet	72°45'N	86°10'W
Aldrich, Cape	83°09'N	70°22'W
Alert	82°30'N	62°20'W
Amundsen Gulf	70°30'N	123°00'W
Antoinette Bay	80°53'N	77°30'W
Arctic Bay	73°01'N	85°08'W
Arctic Ocean	78°00'N	160°00'W
Austin Channel	75°28'N	103°15'W
Axel Heiberg Island	79°30'N	90°00'W
Baffin Bay	72°00'N	65°00'W
Baffin Island	68°00'N	72°00'W
Baker Lake	64°18'N	96°00'W
Ballentyne Strait	77°25'N	114°05'W
Banks Island	73°30'N	120°00'W
Barrow : Alaska	71°20'N	156°46'W
Barrow Strait	74°25'N	95°00'W
Bear Island	74°30'N	19°00'E
Beaufort Sea	73°00'N	140°00'W
Bellot Strait	72°00'N	94°45'W
Bering Sea	58°00'N	167°00'W
Bering Strait	66°00'N	170°00'W
Boothia, Gulf of	70°00'N	89°00'W
Borup Fiord	80°34'N	83°47'W
Byam Channel	75°15'N	105°15'W
Byam Martin Channel	75°55'N	104°40'W
Cambridge Bay	69°06'N	105°07'W
Canada Basin	78°00'N	105°00'W
Canon Fiord	80°00'N	82°40'W
Chukchi Sea	72°00'N	168°00'W
Churchill	58°45'N	94°05'W
Clearwater Fiord	66°33'N	67°23'W
Coral Harbour	64°12'N	83°22'W

Cornwallis Island	75°10'N	95°00'W
Davis Strait	65°00'N	57°30'W
Denmark Strait	66°00'N	30°00'W
Devon Island	75°00'N	86°00'W
d'Iberville Fiord	80°30'N	79°00'W
Discovery Harbour	81°44'N	65°15'W
Disraeli Fiord	83°00'N	74°30'W
Dundas Harbour	74°32'N	82°27'W
Dyer, Cape	74°00'N	102°00'W
East Siberian Sea	73°00'N	160°00'E
Edmonton	53°30'N	113°30'W
Ellef Ringnes Island	78°30'N	102°02'W
Ellesmere Island	79°30'N	81°00'W
Emma Fiord	81°27'N	89°00'W
Eurasian Basin	87°00'N	40°00'E
Eureka	80°00'N	85°56'W
Eureka Sound	79°10'N	85°00'W
Fitzwilliam Strait	76°25'N	116°15'W
Fort Nelson	58°50'N	122°30'W
Fort Smith	60°01'N	111°58'W
Foxe Basin	67°30'N	78°00'W
Franz Josef Land	76°00'N	62°00'E
Frobisher Bay	63°45'N	68°33'W
Fury and Hecla Strait	69°53'N	84°00'W
Greely Fiord	80°23'N	84°00'W
Greenland	66°00'N	45°00'E
Greenland Sea	73°00'N	10°00'W
Hall Beach	68°47'N	81°15'W
Hare Fiord	80°55'N	85°45'W
Hassel Sound	78°20'N	98°45'W
Hell Gate	76°40'N	89°40'W
Hudson Bay	60°00'N	86°00'W
Inuvik	68°18'N	133°29'W
Isachsen	78°47'N	103°32'W

Jones Sound	76°05'N	85°00'W
Jugeborg Fiord	81°15'N	89°30'W
Kara Sea	75°00'N	70°00'E
Kleybolt Peninsula	81°45'N	90°45'W
Lake Harbour	62°15'N	69°52'W
Lancaster Sound	74°10'N	85°00'W
Laptev Sea	76°00'N	125°00'E
Lincoln Bay	82°08'N	61°50'W
Lincoln Sea	83°00'N	58°00'W
Lomonsov Ridge	90°00'N	
MacLean Strait	77°40'N	103°40'W
Massey Sound	78°00'N	94°00'W
McDougall Sound	75°10'N	97°05'W
Mackenzie River Basin	67°00'N	130°30'W
M'Clintock Channel	71°45'N	102°30'W
M'Clure Strait	74°45'N	118°00'W
Melville Island	75°15'N	110°30'W
Mould Bay	76°14'N	119°20'W
Nain	56°34'N	61°40'W
Nansen Sound	81°00'N	90°30'W
Nares Strait	80°00'N	70°00'W
Norman Wells	65°17'N	126°48'W
North Water	76°30'N	74°00'W
Norway	67°00'N	11°00'E
Norwegian Bay	77°45'N	91°30'W
Norwegian Sea	66°00'N	01°00'E
Novaya Zemlya Islands	75°00'N	56°00'E
Otto Fiord	81°02'N	86°40'W
Parry, Cape	70°10'N	124°41'W
Parry Channel	74°30'N	105°00'W
Peary Channel	79°20'N	100°00'W
Peel Sound	73°00'N	96°10'W
Penny Strait	76°37'N	97°15'W

Pim Island	78°44'N	74°25'W
Prince Gustaf Adolf Sea	78°30'N	107°00'W
Prince of Wales Strait	72°40'N	118°30'W
Prince Patrick Island	76°40'N	119°30'W
Prince Regent Inlet	73°00'N	90°30'W
Queen Elizabeth Islands	75°00'N	95°00'W
Queen's Channel	76°10'N	96°20'W
Resolute	74°43'N	94°59'W
Robeson Channel	82°00'N	61°15'W
Rousskaya Gavane : USSR	76°14'N	62°32'E
Sachs Harbour	71°59'N	125°17'W
Smith Sound	78°25'N	73°50'W
Spitzbergen	78°00'N	17°00'E
Strathcona Sound	73°07'N	84°44'W
Sverdrup Basin	78°00'N	102°00'W
Sverdrup Island	78°30'N	95°00'W
Tanquary Fiord	81°05'N	79°20'W
Tuktoyaktuk	69°27'N	133°00'W
Victor Bay	73°06'N	85°15'W
Victoria Island	70°30'N	110°00'W
Viscount Melville Sound	74°15'N	105°30'W
Wellington Channel	75°15'N	93°00'W
Whitehorse	60°43'N	135°04'W
Wilkins Strait	78°10'N	111°00'W
Winter Harbour	74°47'N	110°39'W
Wrangel Bay	82°00'N	62°27'W