LATE WINTER HYDROGRAPHY OF THE NORTHWEST PASSAGE: 1982, 1983 AND 1984

by

B.R. de Lange Boom, H. Melling and R.A. Lake

Seakem Oceanography Ltd.
Sidney, B.C.

and

Institute of Ocean Sciences
Department of Fisheries and Oceans
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PREFACE

This report is one of eleven pertaining to an investigation of the physical oceanography of the Canadian Arctic Archipelago. Three reports appear as volumes of the Canadian Data Report of Hydrography and Ocean Sciences No. 39, and document and present CTD measurements acquired during the three years of the project. A further three reports appear as volumes of Report No. 51 in the same series, and document and present current and tide measurements. Two contractor reports document and discuss wide-ranging CTD surveys within the Archipelago and adjacent waters in the late winters of 1982 and 1983 (Canadian Technical Report of Hydrography and Ocean Sciences Nos. 15 and 16). The present report deals with water masses and baroclinic structure and flow and is one of three attempting to synthesize present and past measurements within this vast and complex system of channels into an up-to-date description of its physical oceanography. The two companion reports discuss tides and tidal currents, and non-tidal residual flows, respectively.

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ABSTRACT

de Lange Boom, B.R., H. Melling and R.A. Lake. 1987. Late Winter Hydrography of the Northwest Passage: 1982, 1983 and 1984. Can. Tech. Rep. Hydrogr. Ocean Sci. 79: 165 p.

Observations acquired during extensive surveys of seawater temperature and salinity within the waterways of the Canadian Arctic Archipelago between 1982 and 1984 are combined with existing data to produce an up-to-date description of the physical oceanography of this part of Canada's Polar Continental Shelf. The data are presented in terms of contoured channel sections, maps of selected hydrographic properties and composite temperature-versus-salinity diagrams. These graphical presentations are discussed in terms of baroclinic flows derived using the dynamical method, and in terms of water-mass transformations of both seasonal and advective origins. Where possible, observed features have been related in a heuristic manner to physical processes in the atmosphere and ocean. The central Archipelago in the vicinity of Barrow Strait is seen to be a meeting ground for waters from as widely separated regions as the southern Beaufort Sea, the northeastern Canada Basin of the Arctic Ocean and Baffin Bay. Mixing, and non-conservative processes in this area ensure that water outflowing to Baffin Bay is quite distinct from the water inflowing from the Arctic Ocean.

Keywords: Northwest Passage, Canadian Arctic Archipelago, physical oceanography, baroclinic flows, water masses, sea water temperature, salinity.

RÉSUMÉ

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Les résultats d'observations réalisées de 1982 à 1984 das le cadre de relevés détaillés de la température et de la salinité de l'eau de voies navigables de l'archipel arctique canadien ont été combinés aux données actuelles afin d'obtenir une représentation à jour de l'océanographie physique de cette partie du plateau continental polaire du Canada. Les données sont présentées sous forme de coupes, en courbes de niveau, des chenaux, de cartes de certaines propriétés hydrographiques et de diagrammes composites température-salinité. Les auteurs traitent de ces répresentations graphiques dans le contexte des écoulements baroclines déterminés par la méthode dynamique, et dans celui des transformations eau-masse d'origine saisonnière ou dues à l'advection. Les caractéristiques observées ont, dans la mesure du possible, été corrélées de façon heuristique aux processus physiques de l'atmosphere et de l'océan. Le centre de l'archipel, au voisinage du détroit Barrow, apparaît comme un point de rencontre de masses d'eau de provenances aussi éloignées que le sud de la mer de Beaufort, le nord-est du bassin Canada dans l'océan arctique et la baie Baffin. Le mélange et les processus non conservateurs caractérisant cette région font que l'eau s'écoulant vers la baie Baffin est passablement différente de celle en provenance de l'océan Arctique.

Mots-clés: Passage du Nord-Ouest, Archipel arctique canadien, océanographie physique, écoulements baroclines, masses d'eau, température de l'eau de mer, salinité.

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The effort expended on the Northwest Passage Oceanography Programme since its inception 5 years ago is considerable, and the number of people and organizations who have contributed to its success is correspondingly large. Without detailing their specific contributions, our appreciation is extended to the following: Syd Moorhouse, Denny Richards, John MacNeill, Gary Moonie, Bob Sudar, Ron Cooke, Paul Johnson, Al Koppel, Andrew Wharton, Bill Green, Jeff Richards, Lyn Lewis and Ron Perkin, all of the Arctic Group within the Physics Division at the Institute of Ocean Sciences; David Fissel, David Lemon, Rick Birch and Doug Knight of Arctic Sciences Ltd.; Simon Prinsenberg and Bert Bennett, formerly of the Bayfield Laboratory for Marine Science and Surveys, Burlington, Ontario; George Hobson and the staff of the Polar Continental Shelf Project, Energy Mines and Resources, Canada; Dick MacDougall of the Canadian Hydrographic Service; Pan Arctic Oils Ltd.; Bradley Air Services Ltd.; Kenn Borek Air Ltd.; Quasar Helicopters Ltd.

In respect of particular contributions to this report the authors wish to thank Bill Buckingham of the Institute of Ocean Sciences for his painstaking and careful work in processing and documenting the CTD data and to acknowledge the efforts of the following Seakem Oceanography Ltd. personnel: Peter Brouwer for the processing and plotting of data, Brenda Fraser and Julia Tudway-Cains for preparing the numerous figures and Gerrie Hosick for word processing the report. We thank Larry Giovando for reviewing the manuscript and L.S.C. Thomson for technical editing.

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1. INTRODUCTION

1.1 Observational Programme

Over the years 1982 to 1984, a programme of physical oceanographic studies was undertaken in the Canadian Arctic Archipelago. The studies, by the Ocean Physics Division of the Institute of Ocean Sciences (IOS), Department of Fisheries and Oceans (DFO), Canada, concentrated on the series of channels and waterbodies generally referred to as the Northwest Passage. The study area extended from Amundsen Gulf and M'Clure Strait in the west through Parry Channel to Lancaster Sound in the east and included adjacent waters from Prince Gustaf Adolf Sea in the north to M'Clintock Channel, Peel Sound and Prince Regent Inlet in the south (Figure 1.1). The bathymetry of the region is shown in Figure 1.2.

The basic objective of the project was to gain a better description and understanding of the physical oceanography of the Arctic Archipelago. In recent years, the Archipelago has become a focus of environmental interest as the result of the search for and extraction of natural resources in the region. This interest has provided the impetus for the current oceanographic studies whose results are discussed in part below.

Specific objectives of the studies included:

a) the identification of the magnitude and direction of tidal and non-tidal currents;

b) the description of tidal propagation,

c) the identification of sources and destinations of various water masses,

d) the identification of the relative importance of physical forces which determine currents and water mass distribution,

e) the identification of temporal and spatial variability in currents and water mass distributions,

f) the estimation of the volume of water transport into and out of the study area and the partition of that transport.

The present report deals specifically with the results of the CTD surveys and hence deals with water structure and water movement as deduced from density distributions. The results of concurrent measurements of water levels and currents will be the subject of other reports.

The CTD data were collected in the years 1982, 1983 and 1984 during early spring, a time when the solid ice cover makes a suitable working platform for obtaining the measurements. The measurements are reported in data reports in the DFO publication series (Buckingham et al., 1985, 1987a, b). During each of the years, the observational emphasis was on a different region of the study area.

In 1982 data were collected in the western part of the study area, consisting of Prince of Wales Strait, eastern M'Clure Strait and western Viscount Melville Sound (Figures 1.3, 1.6, 1.7, 1.8, 1.9). The central region, (eastern Viscount Melville Sound, Byam Martin Channel and M'Clintock Channel) was studied in 1983 (Figures 1.4, 1.9). In 1984 the eastern portion consisting of Barrow Strait, Peel Sound, Penny Strait, Queens and Wellington Channels was covered (Figures 1.5, 1.10, 1.11).

Concurrent with these studies, other related studies were undertaken. In 1982 and 1983, Arctic Sciences Ltd. conducted CTD measurements in adjacent areas, under contract to the Ocean Physics Division (IOS) (Fissel et al., 1984a, b). The 1982 work covered Amundsen Gulf and other areas shown in Figure 1.3. In 1983 the areas visited were similar with the omission of Amundsen Gulf (see Figure 1.4). In 1984, measurements in adjacent areas were combined with the Arctic Shelf Programme of the Ocean Physics Division. Stations are shown in Figure 1.5. Relevant data from this programme are discussed here. The Bayfield Laboratory for Marine Science and Surveys of Burlington, Ontario was active during the springs of 1981, 1982 and 1983 in the Barrow Strait region (Prinsenberg and Sosnoski, 1983a, b, c).

The instrument employed in these studies was a Guildline model 8706 digital CTD probe with a Guildline model 87102 control unit (see Table 1.1 for typical calibration accuracies). The sampling rate of 25 times per second gave a vertical sampling interval of 0.06 m at the controlled lowering rate of 1.5 m/s. The lowering rate was chosen to match the responses of the temperature and conductivity sensors (Perkin and Lewis, 1982). The data were recorded digitally on magnetic tape. In addition water samples were obtained for calibration purposes.

Transportation between stations was by Bell 206L1 helicopter using a GNS 500A Omega/VLF navigation system. This system was capable of an accuracy of better than 1 km under careful use. Further details on instrumentation can be found in Buckingham et al., (1985, 1987a, b).

The station positions were chosen so that the spacing of the stations allowed a large-scale definition of water properties. Station locations were more closely spaced near the shore lines with the intent that narrow coastal currents could be resolved (LeBlond, 1980). Further details on procedures can be found in Buckingham et al., (1985, 1987a, b), as can summary data listings and plots.

1.2 Background

1.2.1 Bathymetry

The bathymetry of the Arctic Archipelago is shown in Figure 1.2. Local bathymetry is found on the detailed station maps (Figures 1.6, 1.7, 1.8, 1.9, 1.10 and 1.11). In general, the channels and basins of the Archipelago are substantially shallower than the adjacent ocean basins. The continental shelf effectively limits free passage of water to maximum depths of about 450 metres into Sverdrup Basin, to about 380 metres into M'Clure Strait, and to about 360 metres into Amundsen Gulf. To the east Baffin Bay has depths up to 2300 metres, but the sill in Davis Strait limits exchange with the North Atlantic to waters at depths less than 700 metres. The other entrance to the Archipelago from the North Atlantic, through Hudson Strait and Foxe Basin, has a maximum passage depth of less than 80 metres. The deepest continuous passage through the Archipelago is through Parry Channel with a limiting sill depth of 125 metres in Barrow Strait. The waterways connecting Sverdrup Basin to Parry Channel (M'Clure Strait, Viscount Melville Sound, Barrow Strait, etc.) are in general constricted by shallow sills on the order of 100 metres in depth. South of Parry Channel, the waters are typically shallow (less than 200 metres) with the exception of some deeper areas having soundings greater than 350 metres (Prince Regent Inlet, Peel Sound/Franklin Strait and M'Clintock Channel).

1.2.2 Regional Oceanography

Due to its remoteness and the difficulty of ship operations during the short and often ice-inhibited navigation season, the study area has not been extensively investigated. Before the 1950's little was known of the oceanography of the Archipelago. Starting in the late 1940's and continuing into the 1960's summer cruises by icebreakers provided the opportunity to collect oceanographic data using water bottles and reversing thermometers. By the early 1960's these data allowed description of the regional water-mass characteristics (Bailey, 1957; Barber and Huyer, 1971; Collin, 1963). During the 1960's the first significant winter and spring oceanographic measurements were made from the ice (Collin, 1961; Herlinveaux, 1961; Hattersley-Smith and Serson, 1966). Gradually the use of tracked vehicles and aircraft for winter and spring ice-based observations became more common. As new and better instruments and techniques developed in the 1970's, a better quality and greater quantity of data became available. During this period the studies became more focussed but tended to concentrate on smaller geographic areas. In the 1980's a series of wide-ranging measurement programs were undertaken, from which some results have been reported (Prinsenberg, 1978; Peck, 1978, 1980; Prinsenberg and Sosnoski, 1983a, b, c; Fissel et al., 1984a, b; Melling et al., 1984). Inventories of physical oceanographic measurements in the Canadian Arctic Archipelago can be found in Birch et al. (1983) and Fissel et al. (1983).

Early data indicated that below about 150 metres, waters differ significantly from east to west across the Archipelago, presumably due to the sills which limit free passage of the stratified waters along the channels. Average flow, in summer, was inferred to be to the east and south from the Arctic Ocean through the Arctic Islands to Baffin Bay. Estimated volume transport (via dynamic calculations) was 2.1 x 10^6 m^3/s through Lancaster, Jones and Smith Sounds with half of that transport passing through Lancaster Sound (Muench, 1971). The volume transport through the Islands is thus about 20% of the total outflow from the Arctic Ocean to the Atlantic, but about 50% of the outflow of Arctic Waters (Aagaard and Greisman, 1975).

The water in the study area consists of Arctic Water in the upper 250 to 300 metres and Arctic Intermediate Water (usually termed Atlantic Water) at greater depths. The Arctic Water is characterized by a surface mixed layer, and within the halocline by a temperature maximum layer, a temperature minimum layer and the main thermocline. The surface mixed layer in winter is usually near the freezing temperature for its salinity. The temperature maximum layer, found intermittently throughout the study area in winter, is not as strong a feature as in the western Arctic Ocean. Temperatures are within about 0.1 to 0.5 C degrees of freezing in the cold halocline layer. The main thermocline is a layer of marked gradients in both temperature and salinity.

Atlantic Water is arbitrarily defined to have a temperature greater than 0 degrees C. West and north of the limiting sills in the central part of the Archipelago, Atlantic Water is similar to that found in the Arctic Basin (T = 0.0 to 0.5 degrees C, S = 34.55 to 34.85). Baffin Bay Atlantic Water is found south and east of the sills and it is warmer (up to 2.0 degrees C) and less saline (about 34.08 to 34.50).

In the upper 200 metres there is a trend towards increasing temperature and salinity at constant depth between the Arctic Ocean and the central sills. Horizontal gradients are largest over the continental slope and over the slopes of the central sills. In the shallow waters around Cornwallis Island (Barrow Strait, Penny Strait, Wellington Channel) the water column is well mixed. To the east of this shallow area, near freezing temperatures are found at depths up to 200 metres and waters are substantially cooler than to the west of the sills.

Near the surface the overall mean flow is generally south and east towards Baffin Bay. Mean flows are weak (less than 2 cm/s) except in Penny Strait, where persistent flows greater than 15 cm/s have been observed (Buckingham et al., 1987e). In Lancaster Sound flows can reach 20 cm/s.

Below the mixed layer, temperature maxima and minima have been observed sporadically in the western part of the Archipelago but these layers, particularly the former, are much less evident in the eastern region and non-existent in most central passages. The strong mixing in the central passages, resulting in the upward diffusive transport of heat, apparently smooths away these features. The slowly increasing temperature to the east along isohaline surfaces in the western Archipelago is also a consequence of vertical diffusion (Melling et al., 1984). Any intrusion of near-freezing water into the halocline is insignificant in the Archipelago.

The temperature of Atlantic Water decreases between the Arctic Ocean continental slope and northwestern entrances of M'Clure Strait and Prince Gustaf Adolf Sea, but within the channels the horizontal gradients are much reduced in magnitude. It is the heat diffused upward from the Atlantic Water which warms the Arctic Water. Melling et al., (1984) found no evidence of annual variability in the properties of the Atlantic Water in the Archipelago.

1.3 Environmental Factors

1.3.1 Ice conditions

During the summer of 1981, ice conditions were generally light (see Figure 1.12). Ice cover consolidated by early January 1982 in most of the study area except M'Clure Strait and Amundsen Gulf which consolidated by the end of January. The western part of Lancaster Sound and Barrow Strait consolidated in the last weeks of February whereas both Lancaster Sound east of the Brodeur Peninsula and Prince Regent Inlet south of 73 N remained unconsolidated with a complex pattern of leads.

Ice conditions were moderate in the summer of 1982, the degree of clearing of the previous year not being reached (see Figure 1.13). Maximum simultaneous open water occurred at the beginning of September. As is typical, the eastern part of the study area showed a greater degree of clearing than did the western part. During the winter of 1982/83, the ice in the Queen Elizabeth Islands was already completely fast by mid-October except in the vicinity of Penny Strait and by the end of the month McDougall Sound was consolidated. Prince of Wales Strait, M'Clure Strait, Viscount Melville Sound, M'Clintock Channel, Queens Channel, Wellington Channel and Peel Sound were covered by fast ice by the end of November. Barrow Strait was consolidated by the end of December but Lancaster Sound and Prince Regent Inlet remained unconsolidated until late February when the ice in western Lancaster Sound and northern Prince Regent Inlet became fast.

The summer of 1983 was a better ice year than the previous year (see Figure 1.14) but not as good as in 1981. Again, maximum simultaneous open water occurred in early September, with more open water in the east than the west. By the third week in September, Viscount Melville Sound reached its maximum extent of open water, while Prince of Wales Strait was completely open by the end of the month. During the winter of 1983/84, the eastern part of the Queen Elizabeth Islands and McDougall Sound were in fast ice by the end of October. By about mid-November the ice in the Queen Elizabeth Islands was completely fast, as was the ice in Byam and Austin Channels and Prince of Wales Strait. Peel Sound was covered by fast ice by the end of November and M'Clintock Channel by the end of December. At the end of January 1984 M'Clure Strait, Viscount Melville Sound, Wellington Channel and Barrow Strait were covered with fast ice and Lancaster Sound as far east as Prince Leopold Island. However, northern Prince Regent Inlet and Lancaster Sound as far east as the northern tip of Borden Peninsula were not consolidated until the end of March. No significant changes occurred from March until after April.

The results of ice thickness measurements (generally for level first year ice) taken during the field programs are indicated in Figures 1.15, 1.16 and 1.17 for the years 1982, 1983 and 1984 respectively. The general pattern is that the thickest ice ($\sim$2 m) is found in the Queen Elizabeth Islands, Viscount Melville Sound, M'Clure Strait and west of the entrance to M'Clure Strait whereas the thinnest ice (1 to 1.5 m) is found in Penny Strait, Queens Channel and Lancaster Sound. Elsewhere ice of intermediate thickness is found. These measurements are somewhat biased by operational constraints for making CTD casts; i.e. very thick (rough) or very thin ice is avoided.

1.3.2 Weather Conditions

The following description of meteorological conditions was abstracted from information in the Monthly Record of Meteorological Observations in Canada published by the Canadian Climate Centre of Environment Canada. Monthly temperature, snow fall and wind data were examined.

A comparison of the temperature and therefore the potential for sea-ice growth during the three winters can be made by means of the cumulative freezing degree-days for each winter. The freezing degree-day plots for Resolute, a representative station, are shown in Figure 1.18. The winter of 1981/82 was generally slightly warmer than normal with a slightly cooler March and April whereas the winter of 1982/83 was colder than normal with a colder March and April. In 1983/84 the winter was slightly warmer than normal except in the eastern Archipelago where the winter was somewhat cooler than normal. March and April 1984 were slightly cooler than normal except in the west. The difference between April 1982 and April 1983 amounted to about 1000 freezing degree-days. Figure 1.18 also gives similar plots comparing four different stations in each winter.

Snowfall accumulation on sea ice can insulate underlying seawater from low atmospheric temperatures. During the winter of 1981/82 (up to the end of April) snowfall was near normal (from 22.8 cm at Rea Point to 64.3 cm at Resolute Bay) except in the most easterly part of the study area which had about 120% of normal. The winter of 1982/83 had a significantly lower than normal snowfall (48% to 79% of normal). The following winter (1983/84) showed a highly variable snowfall pattern with Resolute and Mould Bay above normal and Rea Point and Pond Inlet below normal. Snowfall ranged from 35% of normal to 160% of normal. Table 1.2 lists some snow and ice parameters prevailing during the study.

During the period between cessation of ice melting and ice cover consolidation (generally September to December), winds can have a significant effect on the upper water column through the induced mixing of the water. Winds at Pond Inlet generally show a similar pattern to those at Resolute Bay, although they are not as strong. Mean winds at both Resolute Bay and Mould Bay were below normal in the fall of 1982 (Figure 1.19) whereas they were above normal in 1981. In 1983, the pattern was less distinct with Resolute somewhat above normal and Mould Bay below normal.

2. CHANNEL SECTIONS AND FLOWS

2.1 Discussion of Errors

The accuracy of the temperature and salinity measurements following data calibration were about ± 0.005 C degrees and ± 0.008, respectively. In relation to the temperature and salinity sections discussed below these errors are not significant. However, they do have a significant effect on the accuracy of geostrophic shear (baroclinic current) profiles calculated from the temperature-salinity data.

Baroclinic currents were calculated for each of the cross-channel sections measured in the project. The reference level chosen in these calculations was 0 dbar (the surface) in order that sections with different water depths could be easily compared. As will be seen, in some cases this was not a bad assumption while in others the level of no motion was obviously not at the surface.

In the calculation of baroclinic currents, the horizontal positioning errors are generally negligible in comparison with errors in computed seawater density. Even where stations are closely spaced, such as in Prince of Wales Strait (spacing about 2 km), the positioning errors (about 500 m, associated with GNS and dead-reckoning positions) are not dominant. Thus, the error in the calculated baroclinic current is approximatly

(1) $\Delta v \approx \sqrt{2}\, gh\, \Delta\sigma \;/ (\rho f \;\; \Delta x)$

where Δv is the speed error, g is the gravitational acceleration, h is the vertical distance of the calculated current from the reference level, $\Delta\sigma$ is the error in sigma-t, ρ is the mean water density, f is the Coriolis parameter and Δx is the station separation. As the density is strongly dependent on salinity and only weakly on temperature in the waters under consideration, a sigma-t error of 0.006 based on a salinity error of 0.008 has been used to determine the size of the speed error in the calculations. Table 2.1 gives some representative values for the errors. The combination of possible values of station separation and depth separation relative to the reference level gives a range of estimated errors of two orders of magnitude. However, typical speed errors would lie in the range of 0.5 to 1.0 cm/s. Only at relatively close station spacing in deep water would errors exceed the above values.

The confidence limits of the transport calculations are affected by a number of factors of which the most important are: the accuracy of the velocity profiles, the accuracy of the velocity profile extrapolation, the accuracy of the channel cross-section area, the correct choice of the reference level and the degree to which the flow is in geostrophic balance. The largest range of confidence limits is that due to the accuracy of the velocity profiles. Error calculations were done for one cross-section in Prince of Wales Strait (1982 stations RCM01 to RCM08) and for one section across Lancaster Sound (1984 stations 84191 to 84196), based on measured velocities and estimated uncertainties. The uncertainty in the transport values associated with the calculated velocities gave 1.2×10^4 m^3/s for Prince of Wales Strait and 16×10^4 m^3s for Lancaster Sound. These values were 12 times and 15% of the respective calculated transports.

Estimates of the uncertainty in the calculated baroclinic velocity transport were also made as follows. The volume transport, Q, is given by:

$$(2) \qquad Q = \int_0^L \int_{h(y)}^0 u\,(y, z)\,dz\,dy$$

where L is the channel width, h(y) is the channel depth at position y ($0 \leq y \leq L$) and u (y,z) is the along channel current.

The uncertainty in the volume transport, ΔQ, is given by:

$$(3) \qquad \Delta Q = \int_0^L \int_{h(y)}^0 \delta u(y, z)\,dz\,dy$$

where $\delta u(y, z)$ is the uncertainty in the velocity at position (y, z). Typically the station layout was to use close (∽ 2 - 10 km) spacing for stations in shallow (50 - 200 m) water and wider (∽ 10 - 30 km) spacing for stations in deeper (200 - 500 m) water. Thus a typical uncertainty in baroclinic flow at the seafloor is 1 cm/s and zero at the surface. We can use the approximation:

$$(4) \qquad \delta u(y, z) \backsim 0.01Z/h(y)$$

which gives:

$$(5) \qquad \Delta Q \backsim 0.01 \int_0^L \frac{-\,h(y)}{2}\,dy$$

or:

$$(6) \qquad \Delta Q \backsim 0.005\,DL$$

where D is the average depth of the cross section

$$(7) \qquad D = \frac{1}{L} \int_0^L -\,h(y)\,dy$$

and D and L are in metres.

Note added in proof: This calculation of uncertainty assumes that transport errors between adjacent station pairs are independent. This is incorrect, since an erroneously large geopotential at one station will cause overestimate of transport in combination with one of the adjacent stations, and underestimate in combination with the other. To correct approximately for this oversight, transport error estimates in this re port should be divided by (n-1), where n is the number of stations in the section.

The method of extrapolating the velocity profiles to the bottom of the channel sections introduces an uncertainty on the order of 12%. The calculated channel cross-sectional areas have an uncertainty estimated to be 10% to 20% and possibly higher in channels that are poorly surveyed. The uncertainty about the correct reference level, or alternatively the fact that the magnitude of the barotropic current component is not known, introduces further uncertainty which can easily be of the same magnitude or larger than the calculated transports. For example on the above noted Prince of Wales Strait section, the calculated transport corresponds to an average velocity for the section of 0.08 cm/s while the corresponding average velocity for the above Lancaster Sound section is 4.1 cm/s. Thus relatively small barotropic currents can have a large affect on the net transports. In view of the above uncertainties, any errors due to lack of geostrophic balance are likely to be relatively unimportant, since the channels are mostly wide and deep (thus friction effects will be small), and except in channels where strong tidal currents are present, the flows are approximately steady.

2.2 Longitudinal Sections

The longitudinal sections show the variation of temperature and salinity with depth along a channel. Over the three years of these studies, a total of five sections are available. From 1982 there are sections along Prince of Wales Strait and along Parry Channel. No useful longitudinal sections were measured in this study in 1983 (see Fissel et al., 1984b, for 1983 sections from composited data) but there are three from 1984: a Parry Channel section and two sections through Penny Strait to Lancaster Sound, one through McDougall Sound and one through Wellington Channel. The stations used to make up the individual sections were generally not occupied sequentially nor within a day or two. In some cases as much as a month separates occupation of the first and last stations.

The first 1982 section extends from Amundsen Gulf in the south to Viscount Melville Sound in the north through Prince of Wales Strait (Figure 2.1; also Figures 1.6, 1.7). In this section temperature maximum and minimum layers are evident in the upper 100 m near Amundsen Gulf but these features disappear, and there is a decrease in stratification as distance from Amundsen Gulf increases. The mixed-layer salinity and thickness increase to the northeast along the channel. These observations suggest a modification by mixing of a water mass originating in Amundsen Gulf and moving north. A likely source of energy to drive mixing is tidal flow through the relatively shallow and constricted channel. Spring tides reach 35 cm/s in the northern half of the channel (IOS, unpublished data). Note the very rapid loss of stratification near the central sill at Princess Royal Islands, and the approximate two-layer structure found in the northern part of the strait. A two-layer structure was observed also in 1983 (Fissel et al., 1984b: station Q04), but data from 1977 near CTD 20 (Peck, 1978) show a more continuous stratification below the surface layer in this area (Figure 2.14). Bailey's (1957) sections taken in the summer of 1954 also show a two-layer structure with salinities generally lower throughout the water column but particularly near the surface on the eastern side, where salinities as low as 24.18 were found (Figure 2.15). A warm, strongly northward moving surface flow was found by Bailey and a resultant net northward transport was inferred.

Figures 2.2 and 2.3 are sections along Parry Channel taken in 1982 and 1984 respectively. Figure 2.2 extends from eastern M'Clure Strait to eastern Barrow Strait while Figure 2.3 extends from eastern Viscount Melville Sound to western Lancaster Sound (refer to Figures 1.8, 1.9 and 1.10 for station locations). Arctic Water within the halocline east of the central sills is colder and less saline at depths below the sill depth

(125 m) than in western Parry Channel. Also evident are a slight temperature maximum layer in western Parry Channel (near the end of Prince of Wales Strait) and a lens of low salinity surface water in Viscount Melville Sound (Figure 2.2). These observations are similar to those in 1983 (Fissel et al., 1984b; Figure 2.4). Melling et al. (1984) have discussed the reasons for the higher temperature within the halocline west of the Barrow Strait sill. It is clear from these observations that the mixing and advection producing this characteric are persistent from 1982 to 1984. The sub-surface temperature maximum seen at stations 27 and 39 in 1982 was also observed in 1983 and 1984 (Fissel et al., 1984b; Buckingham et al., 1986a, b). This feature is most pronounced east of Prince of Wales Strait in western Viscount Melville Sound, where it is associated with a cyclonic gyre.

This gyre is revealed by a pinching of isohalines near 100 m depth in both longitudinal and transverse sections (Figure 2.20), and is therefore cyclonic and most intense beneath the surface. A pool of low salinity surface water in central Viscount Melville Sound (Station CTD39 in 1982) is associated with the gyre.

Of particular interest in Figure 2.3 is the salinity of the densest waters which pass over Barrow sill and thus contribute to the properties of Arctic water in Lancaster Sound. Figures 2.2 and 2.3 (1982 and 1984) indicate that this salinity is approximately 33.4.

Sections extending from the Prince Gustaf Adolf Sea through Penny Strait to Lancaster Sound in 1984 are found in Figures 2.5 and 2.6. In Figure 2.5 the section passes through Queens Channel and McDougall Sound and then Barrow Strait to Lancaster Sound, while in Figure 2.6 the section passes through Wellington Channel. Both sections show that the waters between Penny Strait and eastern Barrow Strait have higher surface salinities and lower stratification than the waters either to the northwest or to the east. In Queens Channel, McDougall Sound, Barrow Strait and Wellington Channel waters are extremely well mixed. The overall pattern here is similar to that found by Fissel et al. (1984b) in the spring of 1983 (Figure 2.7). The highest salinity passing over the sills southeast of Penny Strait is about 33.1. The higher surface salinity and reduced stratification in the waters around Cornwallis Island is almost certainly a result of tidally induced mixing. Tidal currents in this area are strong; spring tidal flows of 50 cm/s are typical and values of 120 cm/s are attained at some locations. Since the length of the sill region is substantially greater than a tidal excursion, there is ample opportunity for homogenization of water parcels during their slow transit through the region.

An interesting consequence of this mixing is the maintenance of a longitudinal internal pressure gradient which drives flow away from the sill in both directions with a maximum at the depth where the horizontal density gradient is zero. Flow continuity requires a barotropically driven inflow towards the sill. The result is a surface inflow and a subsurface outflow. The Coriolis effect will, of course, cause the resultant flows to be separated horizontally rather than vertically in these wide channels (internal Rossby Radius ≪ Channel width). See, for example, Figures 2.32 and 2.40 in Barrow Strait and Wellington Channel, respectively. Given that net transport is southward and eastward through the Archipelago, these longitudinal sections illustrate that waters from the Arctic Ocean are mixed, far within the Archipelago, with waters from Baffin Bay, and modified prior to their eventual efflux to the east.

2.3 Prince of Wales Strait

The 1982 data provide the series of cross sections and baroclinic flow profiles from south to north in Prince of Wales Strait shown in Figures 2.8 to 2.13. Station locations and bathymetry can be found in Figures 1.6 and 1.7.

All sections illustrate a tendency for convex upward isolines below about 50 m depth and convex downward isolines at shallower depths. This implies a maximum in baroclinic current at about 50 m, and a cyclonic horizontal shear in this current. Flow is southwesterly on the west side of the channel and opposing on the east side. An upward slope in the isolines to the east, in addition to the curvature, reveals a tendency for the northeasterly flow to occupy the deeper and more easterly part of the section, and the southwesterly flow to occupy the shallower more westerly part.

In any particular section, the observed tendency for counter-flowing boundary currents may be obscured by other features. In the southernmost section (Figure 2.8), there are indications of an anticyclonic eddy located near the channel centre, in association with a lens of warm water at 40 m depth. In the section near the Princess Royal Islands (Figure 2.9), the southwesterly flow occurs near the channel axis, between stations 8 and 9, rather than against the west shore. This is presumably a consequence of flow obstruction by the extensive shoal to the southwest of station 7 and 8. In the deeper, narrower Armstrong Point section (Figure 2.10), the counterflows appear to be separated vertically, not horizontally, and the deeper waters move to the northeast. In the section near 116 degrees west (Figure 2.11), as with the Princess Royal Islands section, topography again appears to modify the baroclinic flow; steep slopes associated with the canyon on the channel axis and an upstream shoal on the Banks Island side of the channel force the southwesterly flow to follow the channel axis, rather than the western shore.

The general tendency for northwesterly flow at depth is consistent with the northward modification of water properties in the strait discussed in connection with Figure 2.1. The interpretation of this figure implies a persistent northward flow of water. The second occupation of the section at 116 degrees west (Figure 2.12), three months after the first, revealed a flow structure very similar to, though more intense than, that revealed by the first. The flow observed is thus representative of the late winter period, at least in 1982.

Observations in the late winter of 1977 (Peck, 1978) show a similar type of water column structure near the north end of Prince of Wales Strait (Figure 2.14). The deep salinities are similar and isohaline lines slope up to Victoria Island (eastern) shore. However surface layer salinities were much lower in 1977 (< 31) than in 1982 (> 32.2). Bailey's (1957) summer measurements indicate warm, low salinity water flowing northeastward along the east side of the Strait (Figure 2.15). Although the direction of flow is consistent from summer to winter, the type of water associated with the currents changes from summer to winter. In summer the northward flow is warm and of low salinity while the counter flow on the west side is cold and of high salinity; in winter the northward flow is cold and of high salinity.

The observations reveal that despite the narrowness of the strait (10-20 km), there are significant cross-channel variations in baroclinic flow. Since the relevant length scale for baroclinic motions is the Internal Rossby Radius, which ranges between 4 km in the south and 2 km in the north, such cross-channel variability in internal motions is not unexpected.

2.4 Western Parry Channel

Hydrographic sections were occupied in western Parry Channel in all three years of the programme (see Figures 1.3, 1.4 and 1.5).

By appearances, there is less spatial variability in the sections of this wide, deep channel than in Prince of Wales Strait. While this appearance is probably genuine away from the boundaries of the channel (wide station spacings make a definitive statement difficult), near the boundaries the appearance is somewhat deceiving due to the dramatic difference in plotting scales between Prince of Wales Strait and Parry Channel (a four-fold change in depth, and a five-fold change in distance). Baroclinic flows up to 15 cm/s are calculated near the boundaries of Parry Channel (Figures 2.16 to 2.24), while flows more than about 10 km from boundaries are very small. The implied weakness of residual flows in the main body of the channel has been corroborated by direct current measurement during this programme (Buckingham et al., 1987c).

Surface salinities are lower on the south side of Viscount Melville Sound (this is evident from the sections but is best shown on Figures 3.4, 3.5, 3.6 and 3.7). These values in the range 31.0-31.5 are characteristic of basin waters of the Beaufort Sea in winter. Their presence in Viscount Melville Sound may be evidence for a summertime surface inflow from the Beaufort Sea via M'Clure and/or Prince of Wales Straits. Winter inflow seems unlikely since both straits in winter contain surface waters of higher salinity. The observations of a sub-surface temperature maximum at stations north of Peel Point and Natkusiak Point in 1982 and north of Stefansson Island in 1984, in association with low surface salinities, are further evidence for an origin for these waters in an area subject to considerable radiative warming in summer, namely Amundsen Gulf or the southeastern Beaufort Sea. Ice cover in southern Viscount Melville Sound was heavy in 1981 (Figure 1.12), thereby permitting little radiative heating locally. This deduction regarding surface-water origin is supported by the observations of Bailey (1957), who derived a northward transport of water through the strait, with surface temperatures greater than 5.0 degrees C and salinities below 25 in the summer of 1954 (Figure 2.15).

Differences in surface salinity can be a strong reflection of differences in the thickness of the mixed layer, which grows by entrainment of deeper more saline waters from the halocline. For example, surface salinity at the western end of Parry Channel (see Figures 2.16 and 2.17) was lower in 1984 than in 1982, whereas the surface layer was about twice as thick in 1982 as in 1984. Since the average salinity of the uppermost 50 m is about the same in the two years (32.7 vs. 32.4), the differences reflect a more energetic mixing in 1981/82. Reference to weather statistics cited earlier reveals that in the winter of 1981/82 at Mould Bay, mean and maximum hourly winds were greater in October than in 1983/84 during the period prior to the consolidation of the ice cover. The fall of 1983 brought heavier ice cover than the fall of 1981, and therefore less potential for rapid ice movement and total ice growth, both of which cause mixing in the upper ocean.

Below about 50 m, cross-channel isohaline surfaces are nearly level. Only near the boundaries are there significant slopes (and significant baroclinic flows). Beneath the surface layer, local maxima and minima in temperature are generally absent. The occurrence north of Victoria Island of a maximum immediately beneath the surface layer, which is probably an advected remnant of summertime heating, has been discussed above. Another exception to this generality is the broad temperature minimum in the halocline (a characteristic feature of the Beaufort Sea) which was observed in

1984 in a layer centered at about 120 m depth in the southern part of western M'Clure Strait (Figure 2.16). The existence of this feature suggests that inflow from the Arctic Ocean occurs in this area, a conclusion reached on the basis of 1982 data by Melling et al. (1984). This interpretation is, however, not clearly consistent with the computed shear profiles (Figure 2.16). Some smaller-scale, short-lived flow variations may be influencing the latter.

Eastward flow in the halocline along the southern shore of M'Clure Strait appears to feed into the cyclonic sub-surface gyre in Viscount Melville Sound (velocity maximum at about 80 m depth; see Natkusiak Peninsula section: Figure 2.20). This gyre is revealed by the 'pinching' together of the isohalines, concave upwards above 80 m and concave downward below this depth. Near the northern end of this section, there is a spreading of the isohalines between 50 and 110 m depth so that the westward sub-surface flow associated with the gyre here has somewhat more uniform salinity than the surrounding waters. Further to the east, in the Stefansson Island Section (Figure 2.21), the pattern on the north side is similar, and since the T-S characteristics of the water in this depth range match those in Byam Martin Channel (see section 4.2), a continuous flow south and west around Melville Island, then joining the gyre in Viscount Melville Sound, is implied. This pattern of counterflowing right-bounded currents, noted in Prince of Wales Strait and also in western Parry Channel, will be seen to be a very characteristic feature of flow within the Archipelago. The existence of this type of flow has been discussed by LeBlond (1980) in terms of geostrophic dynamics and channel geometry.

The counterflowing currents noted in Prince of Wales Strait must interact with the southeastward flow across its northern end. Since the T-S characteristics (section 4.2) of the upper 50 m in the northern end of the strait lie between those on the southern side of M'Clure Strait (CTD22-CTD23) and those in southwestern Viscount Melville Sound (CTD32-CTD33), the southwestward flow of surface waters in western Prince of Wales Strait clearly originates north of Banks Island. The northeastward flow of slightly more saline deeper water in Prince of Wales Strait, as revealed by its T-S characteristics, moves eastward into Viscount Melville Sound, underflowing less saline, warmer waters (which may have moved this way out of the strait during the preceding summer). Eastward baroclinic flows were computed from station data acquired north of Peel Point in 1982, although a band of apparently westward flow about 40 km offshore does confuse the picture slightly.

Observations to the north and east of Stefansson Island, where M'Clintock Channel joins Viscount Melville Sound, were acquired in 1982 and in 1983 (Figures 2.21, 2.23 and 2.24; see also Figure 25 in Fissel et al., 1984b). Progressive changes in temperature-salinity characteristics suggest an eastward, then southward, flow around the island from Viscount Melville Sound into M'Clintock Channel, at least within the halocline. However, baroclinic flow calculations (assuming no motion at the surface) suggest a northwesterly flow in the halocline on the south side of the sound (Figures 2.21 and 2.23). To resolve this apparent contradiction, it is necessary that the relatively fresh surface water also be moving south and eastward into M'Clintock Channel. This surface water may originate in the pool observed in 1982 north of Victoria Island. Unfortunately in 1983 there are no CTD measurements in that area to confirm the continued existence of this pool. In northern M'Clintock Channel, (Figure 2.25) there are significant isohaline slopes only on the east side (similar deductions were made by Fissel et al., 1984b). Over most of the channel baroclinic flows are small, a result consistent with the data from a moored current meter in the middle of the channel (Buckingham et al., 1987d). The T-S characteristics of the stations on the eastern side of this section do not give any conclusive indication as to the origin of the water found here, since they resemble both those of Viscount Melville Sound to the north and those of Larsen Sound to the south. However, near the northwest coast of Prince of Wales Island (Figure 2.24) the T-S plots of the stations resemble those of M'Clintock Channel more closely than those of Viscount Melville Sound, particularly in the upper 100 m. Although the baroclinic flows are southwesterly relative to the surface on this section, a current meter at this location indicated a northeasterly surface flow (Buckingham et al., 1987d). The surface salinity distribution (Figure 3.5) also suggests a northeasterly flow. In summary, this junction of channels again shows a cyclonic excursion into the side channel (M'Clintock) of an eastward surface flow across its mouth. Right-bounded counterflows appear in M'Clintock Channel at least as far south as 73 degrees N.

To the north, where Austin Channel joins Viscount Melville Sound at the southwest corner of Bathurst Island, the channel sections (Figure 2.23) reveal a thick layer of water of salinity 33.0 ± 0.2. Baroclinic flow is to the northwest relative to the surface. The T-S characteristics of this layer closely resemble those of north central Barrow Strait and McDougall Sound but differ from those of Viscount Melville Sound (Figures 4.11a, b, c; section 4.2 and 4.4). It thus appears that a westward subsurface flow south of Bathurst Island here turns north into Austin Channel. This flow counters the eastward flow on the opposite side of Viscount Melville Sound at this longitude. Current measurements (this project) in this area support in the main this deduced circulation.

The 100 degree W section between Bathurst and Prince of Wales Islands (Figure 2.28) shows more variation in the isohaline depths and spacing and correspondingly larger baroclinic flows than other sections in western Parry Channel. The surface salinity distribution (Figure 3.6) suggests an eastward surface flow on the south side of the channel. At the northern end of the section sub-surface flow is deduced to be westward on the basis of both the baroclinic flow calculation and the progressive T-S changes discussed in the preceding paragraph.

The baroclinic velocity profiles on the sections across M'Clure Strait and Viscount Melville Sound show common features. Generally the vertical shears are small (1 or 2 cm/s over 300 or 400 m) in the middle of the channel but up to an order of magnitude larger at the channel walls or near steep topography. On the south side of the channel the flow below the surface layer is eastward while on the north side it is to the west. Significant flows are confined to a region within about 15 km of the coast (approximately the internal Rossby radius), although the westward flow near the south shore of Melville Island appears somewhat broader than this in the more easterly of the three sections occupied in 1982.

At junctions between Parry Channel and side channels, flows appear to follow the tendencies of the shoreline on their right. Thus on the northern side of Parry Channel a westward flow makes an excursion into Austin Channel, and a southward flow from Byam Channel moves westward along the southern shore of Melville Island. On the southern side of Parry Channel, an eastward flow curves into Prince of Wales Strait and M'Clintock Channel, while outflows from these side channels join the eastward flow in the main channel. The principal circulation features deduced in this section are depicted schematically in Figure 2.52.

2.5 Queen Elizabeth Islands

Channel cross-sections were occupied between Ellef Ringnes and Melville Islands in 1982, 1983 and 1984, and in Byam Martin Channel in 1982 and 1983 (see Figures 2.26, 2.27 and Figures 14 and 16 in Fissel et al., 1984a). The 1984 section from Melville Island to Ellef Ringnes Island is shown in Figure 2.26. A relatively low salinity surface layer with temperatures at freezing found near Melville Island was observed here also in 1982 and 1983. The value of salinity (31.6) and the pattern of isopleths on the maps of surface salinity strongly suggests that this water originates in the Arctic Ocean northwest of the Archipelago. The only other significant source of fresh water is summer ice melt, and there is no reason to assume that melt at this location should be any greater than that to the northeast on this section. The surface mixed layer associated with the low salinities is, however, quite thin, and the low salinity anomaly averaged over the upper 50 m is not as dramatic as the surface anomaly depicted (see section 3, Figure 3.13). Bottom water in the southwestern part of the section is less saline than that in Maclean Strait presumably because of the isolation between the two parts of the section due to the sill (at 350 m) between Lougheed Island and Borden Island. Dynamic calculations suggest flow in the same direction as surface movement, only weaker. Flow measurements by Peck (1980b) support this interpretation, with generally south and easterly flow, both at the surface and 50 m depths. The exception to this was the currents at 50 m depth in Maclean Strait which showed a north and west mean flow over the record period of about two weeks. Within the halocline, a progressive warming (see section 3) with distance into the Archipelago is consistent with a generally southerly flow of Arctic water.

The section across Byam Martin Channel in 1983 is shown in Figure 2.27. A section was occupied here in 1982 (Fissel et al., 1984a) and in 1979 (Peck, 1980). In these sections the lower salinity surface water is found on the west side of the channel implying, on the basis of surface salinity maps a southward moving surface layer.

Vertical shears are large, particularly near the sides of the channel. Although a northwest flow relative to the surface is indicated, the existence of a significant southerly barotropic component is suggested by the character of the T-S plots on either end of the Channel and by the type of water found in Viscount Melville Sound (see section 4.2). The existence of a deep counterflow to the northwest near the eastern shore is a possibility, however, Greisman and Lake (1978) report net southerly flow through both Byam and Austin Channels. Current measurements taken concurrently with the CTD transect in 1983 (Buckingham et al., 1987d) also indicate a southerly mean flow at the surface.

2.6 Central Sills Region

2.6.1 Barrow Strait

The 'Central Sills Region' is an informal designation for the waters surrounding Cornwallis Island, namely Barrow Strait (Figures 2.29 to 2.32), Penny Strait, Crozier and Pullen Straits, McDougall Sound (Figures 2.35 to 2.37), Wellington Channel (Figures 2.38 to 2.40) and Peel Sound (Figures 2.43). In general this region has irregular bottom topography.

The sections across Barrow Strait generally show much similarity: on the south side there are lower surface salinities and greater vertical temperature and salinity gradients than on the north side, where salinity is close to 33.0, particularly below 50 m depth, and temperature is mostly within 0.2 C degrees of freezing. This pattern is consistent with coastal flows which are westward on the north side of the channel and eastward on the south side: the cold, relatively high salinity water observed on the north side has moved south from Penny Strait, McDougall Sound and Wellington Channel whereas the water on the south side has its principal origins in southern Viscount Melville Sound although some of this water may move northward from M'Clintock Channel and Peel Sound. Near the seafloor, Figures 2.3 (a longitudinal section through Barrow Strait) and 2.29 both suggest an eastward flow over the Barrow sill located south of Lowther Island.

The southern part of the section in Figure 2.29 actually crosses the sill diagonally (see Figure 1.12) and the spilling of 33.4 salinity water down the southeast slope of the sill is evident.

The section from Lowther Island to Cornwallis Island (Figure 2.30) shows even weaker stratification over the 300 m channel depth than does the northern part of the section to the west. This can be attributed to the joint influences of a well mixed water mass coming south out of McDougall Sound and of the cold homogeneous water in the westward coastal current, along the south side of Cornwallis Island, which originates in Wellington Channel. These waters apparently exclude the lower salinity waters of southern Barrow Strait from this section. The general pattern of this salinity distribution is reflected in the surface salinity maps (Figure 3.5, 3.6, 3.7, 3.8 and 3.9) every winter between 1981 and 1984. Figures 3.7 and 3.8 show the surface salinities in 1981 and 1982 (Prinsenberg and Sosnoski, 1983a,b). The 33-salinity water from the north undercuts the lower salinity water on the south side of Barrow Strait, producing some temperature layering such as is evident north of Somerset Island in Figure 2.32. This layering appears to be a winter feature resulting from the strong modifying effects of wintertime mixing and freezing on water inflowing from the north. The summer results of Bailey (1957) in Barrow Strait (his section IX, August 1954, see Figure 2.33) do not reveal large north-south gradients of salinity across the strait even near the surface, presumably because ice-meltwater from all source regions, in shallow mixed layers, obscures the source signatures. In addition, although mixing is still active, the formation of freezing brine is stopped in summer. The summer section from Jones and Coote (1980) shows a similar pattern (Figure 2.34), particularly in the temperature, with the warmest water occurring on the north side of Barrow Strait, suggesting it is coming from Wellington Channel.

The baroclinic flow profiles on sections across Barrow Strait (Figures 2.29 to 2.32) show moderate vertical shear west of the sill and large shear east of the sill. Again the smallest shears are generally found near mid-channel and the largest near the shores or other steep topography. Relative to the surface, flows are generally westward. However, as discussed in connection with surface salinity patterns above, surface flows along the southern shore are certainly eastward, and the shears observed therefore imply decreasing eastward flow with depth, and on the north side the possibility of westward flow at depth, consistent with progressive T-S changes along the channel.

In the section by Griffith Island an easterly flow is found north of the island and a westerly flow to the south (Figure 2.31). These combine to suggest an anticyclonic circulation about the island. Such a circulation was very evident in drift-buoy trajectories (Fissel and Marko, 1978) and in ice-floe tracks (MacNeill et al., 1978) in the summer of 1977. The strong eastward flow north of Somerset Island evident in Figure 2.32 was also clearly seen in the drift-buoy studies. Vertical shears are large north of

Somerset Island (Figure 2.31, 2.32 and 2.44), as much as 15 cm/s over a 50 m depth change, and the baroclinic flow is strongly westward relative to the surface, except close to Somerset Island where flow is strongly eastward at about 30 m depth on all three sections. Since the surface salinity distribution (section 3) and the above-noted surface current and ice drift results establish the existence of a barotropic eastward component to the flow in this area, the shear profiles indicate most intense eastward flow at about 30 m depth, and decreasing eastward flow to the seafloor.

2.6.2 Penny Strait

The 1984 section across Penny Strait is illustrated in Figure 2.35. This section was occupied also in 1982 and 1983 (Fissel et al., 1984a, b). Surface salinities are higher than those to the northwest, with values similar to those at 70 to 80 m in the Prince Gustaf Adolf Sea. Temperature and salinity are less in the deep part of the strait than at similar depth in the area northwest of the Strait, and temperature on a given isohaline is somewhat higher. The former is a consequence of the blockage by the sill at about 150 m depth to the northwest of the section in Figure 2.35 while the latter fact reflects the greater degree of diapycnal mixing in the approaches to Penny Strait. Examination of the longitudinal sections through Penny Strait (Figures 2.5 and 2.6) indicates that diapycnal mixing occurs in and downstream of Penny Strait also and that a nearly homogeneous water type is produced after passage over the sills in southern Queens Channel and northern Wellington Channel.

Figure 2.35 contains the baroclinic velocity profiles in Penny Strait. The shears are large and show one or more flow maxima and minima with depth. Generally the flow pattern is complex but a northwesterly flow relative to the surface appears to dominate. The complex bottom topography of the strait and the existence of large amplitude internal waves (Fissel et al., 1984b) introduce significant uncertainty into the dynamic calculations. Uncertainties of ± 3 to 5 cm/s with a station spacing of 5 km are possible. Fissel et al. (1984b) inferred a southerly flow relative to a 150 db reference level, and current meter data acquired during this project confirm this inference (Buckingham et al., 1987e). The temperature and salinity longitudinal sections (Figures 2.5 and 2.6) as already discussed are only consistent with a southerly flow.

2.6.3 Queens Channel and McDougall Sound

The effect of continued mixing through Queens Channel is shown in Figure 2.36 on the section between Bathurst Island and Cornwallis Island through Crozier and Pullen Straits. Over a depth of more than 300 m the salinity increases by less than 0.3 and the temperature by less than 0.2 C degrees. In the next section to the south between Bathurst and Cornwallis Islands across McDougall Sound (Figure 2.37) waters are also nearly homogeneous. Water-mass variations increase again in Barrow Strait, as this southerly flow meets the eastward flow with different characteristics moving in from Viscount Melville Sound.

The presence of cold (within 0.05 C degrees of freezing temperature) water near the bottom on the eastern side of the main channels in Figures 2.36 and 2.37 suggests a conditioning by brine rejection during ice formation in the relatively shallow area north of Cornwallis Island (Figure 2.38). The well mixed, nearly isopycnal water column would allow sinking of water to the seafloor with only a small salinity increase (i.e., relatively little ice formation is required). The -1.6 degree C core in these figures is the relatively warm southward moving water from Queens Channel.

The two sections show similar baroclinic flow patterns (Figures 2.36 and 2.37). Shears are generally moderate and show a southerly flow on the west side of the channel and a northerly flow on the east side of the channel, relative to the surface. Once again surface flow must be southerly in order to generate the observed progressive changes in temperature and salinity. Current measurements by Greisman and Lake (1978) revealed southerly flows in Crozier Strait except for a weak northward current at depth on the eastern side.

2.6.4 Wellington Channel

Figures 2.38 to 2.40 show a sequence of sections across Wellington Channel from north to south. Most of the water in the channel has temperatures below -1.5 degrees C and salinities between 32.8 and 33.2, but a markedly different water is found in the upper 75 m on the eastern side of the channel. The observed distribution is indicative of a southward flow of well mixed, cooled water that has come through Penny Strait and Queens Channel and a northward surface flow on the east side of the strait of a lower salinity water originating in Lancaster Sound.

Figure 2.38 shows water properties to be nearly homogeneous just east of the sill barring the channel in the north near Baillie-Hamilton Island. The warmest water is found at station 84141, presumably because it is closest to the deeper channel leading north to Penny Strait, where warmer waters of this salinity are found.

The Helpman Head section (Figure 2.39) near the middle of Wellington Channel, does show lower salinity water near the surface on the eastern side. This layer is broader and thicker in the Innes Point section to the south (Figure 2.40). The lowest surface salinities are found at the southeastern corner of Wellington Channel.

The vertical profiles of baroclinic flow indicate little or no vertical shear in the section at the north end of Wellington Channel. Presumably flow here must be barotropic, due to the almost non-existent stratification. This is not the case for the central and southern sections across the channel, Figures 2.39 and 2.40. In both these sections the vertical shear increases from west to east across the channel with a southerly flow relative to the surface over most of the width of the sections. Near the eastern shore a narrow band of northerly flow is found in the upper 50 m. The absence of lower salinity water in the section through Baillie-Hamilton Island suggests that the northerly coastal current does not extend as far as the north end of Wellington Channel. The signature of this inflow must be lost during mixing with adjacent southward flowing waters. The unpublished measurements in April 1978 of Lake (Figure 2.41) along a section at the same location as Figure 2.39, also show the existence of a northward coastal flow on the Devon Island side.

Bailey's (1957) results indicate a similar flow pattern for the summer (Figure 2.42). In his data the lowest salinities (as low as 30.55) are found in the southward flowing water on the west side of the channel. The explanation for the reversed cross-channel surface salinity gradient is not clear from the data although it appears that the high salinity (32.36) on the east side is anomalous compared with surface salinities in Barrow Strait. The high salinity water may originate in the westward coastal flow occasionally found on the north side of Lancaster Sound (Fissel et al., 1982; see Jones and Coote, 1980 for summer sections across Lancaster Sound). Again the waters below about 50 m have salinities within 0.3 of 33.0 although temperatures are about 0.5 to 1.0 C degrees warmer than in winter.

Although only one surface drifter passed through Wellington Channel, the results of Fissel and Marko (1978) indicate the north flowing coastal current on the east side extends as far north as Prince Alfred Bay in the summer season.

2.6.5 Peel Sound

The section across northern Peel Sound (Figure 2.43) crosses part of the basin extending north into Barrow Strait (Figure 1.10). Below about 70 m the water characteristics are the same as those in eastern Viscount Melville Sound (Figure 4.23, section 4), indicating that region is the likely deep-water source. It has already been noted, in the discussion of the section across Barrow Strait in the vicinity of the sill (Figure 2.29), that there is an indication of water of salinity as high as 33.4 spilling eastward over the sill. Such water is free to flood the basin in Peel Sound (see Figure 1.10).

Since surface salinities are lower than in most of Barrow Strait, the surface water in Peel Sound could originate in the southern part of Viscount Melville Sound and move eastward close to the north shore of Prince of Wales Island and/or from further south in Larsen Sound and move north along the west side of Somerset Island. Low surface salinities were measured in 1982 and 1983 by Fissel et al. (1984a, b) in Larsen Sound. These values (near 31) are suitably close to the relatively low surface salinities (about 31.4 ± 0.3) found in the northern part of Peel Sound.

The baroclinic velocity profiles for this section (Figure 2.43) show varying amounts of shear. Near the shore on both sides a southerly flow relative to the surface is evident above about 150 m, whereas in the centre the flow is northerly at these depths with southerly flow beneath. The southerly flow near the bottom fits with the idea that the deeper water found here originates west of the sill in Barrow Strait. Surface current measurements by CCIW in April 1981 (B. Bennett, private communication) revealed a southerly flow (about 3 cm/s) on the west side of the Sound and a stronger northerly flow (about 7 cm/s) on the east side.

2.7 Eastern Parry Channel

The pattern of temperature and salinity in eastern Parry Channel continues the transition from the water properties of the central sills to the water properties of Baffin Bay. The relatively cold (-1.5 ± 0.2 degree C) water of salinity 33.0 ± 0.2 that is found in Barrow Strait can be traced in the sections across Lancaster Sound and Prince Regent Inlet. All the sections have sizeable baroclinic flows, particularly near the shores. The sections indicate cyclonic geostrophic shears indicative of right-bounded counterflows and perhaps of mid-channel eddies such as observed by Fissel et al. (1982).

The Garnier Bay section between Somerset and Devon Islands (Figure 2.44) is located east of Wellington Channel where Barrow Strait deepens into Lancaster Sound. The continuation eastward of the cold core at 100 m depth seen on the south side of the channel in the Cape Dungeness section (Figure 2.32) is consistent with an eastward flow of water over most of the water column at the south shore.

There is a slight (< 0.5) increase in surface salinity towards the east evident from the sections across Lancaster Sound from Hobhouse Inlet and across northern Prince Regent Inlet at Batty Bay (Figures 2.45 and 2.48; see also Figure 3.6) although this seems too small to merit speculation. The upper 100 m is somewhat warmer and less saline than in 1983 (see Fissel et al., 1984b) in both Lancaster Sound and in Prince Regent Inlet. In both sections the deeper water shows similar structure in 1983 and 1984. Isohalines slope down to the south across Lancaster Sound in both years, suggesting a westward flow at depth relative to the surface. A similar situation exists in summer (Figure 2.46) although salinities are lower, and temperatures are higher, near the surface. This figure suggests that in summer the eastward surface flow continues on the south side of the channel with a cool (< 0 degrees C) low salinity (s < 31) core.

The velocity profiles in Figure 2.44 are consistent with those in southern Wellington Channel and further west in Barrow Strait. Mid-channel shears are small and in the upper 70 m an easterly flow is indicated near the south shore while a westerly flow is found near the north shore. Further east in Lancaster Sound (Figure 2.45), the velocity shears are somewhat larger. Relative to the surface, the baroclinic flow is mainly westward so that eastward flow decreases from a maximum near the surface (except between about 50 and 200 m in the north central part of the section). The reality of westward flow on the north side of Lancaster Sound is supported by the similar T-S structure in the upper 50 m found at stations along the Devon Island shore. Surface drift paths in summer (Fissel and Marko, 1978) revealed the same pattern of surface counterflows. These results and those of Fissel et al. (1982) indicate that the flow in central and western Lancaster Sound is complex and irregular, at least in summer, a characteristic attributable in part to energetic eddies in the flow. Variable wind directions may also contribute to the variability in current. Much of the character of surface flow deduced from the hydrography in this study is evident in the drift patterns of sea ice, particularly in the eastern part of the sound (Marko, 1977), and of icebergs (see Figure 2.47 and de Lange Boom et al., 1982).

In Prince Regent Inlet (Figure 2.48) the pattern of vertical shear is complex with southerly flow on both sides of the channel and northerly flow in the middle of the channel below the surface mixed layer. The drifter paths reported by Fissel and Marko (1978) revealed a southward flow on the west side of the inlet and a northward flow on the east side. Since the T-S characteristics in the upper 100 m at stations in Prince Regent Inlet are similar to those in Lancaster Sound near the Brodeur Peninsula shore but 0.02 C degrees warmer at a given salinity, the hydrography supports the existence of a northeasterly flow along the Brodeur Peninsula.

2.8 Dynamic Height Anomalies

Maps of the dynamic height anomaly referenced to various pressure levels are sometimes useful for delineating patterns of horizontal flow. Figures 2.49, 2.50, and 2.51 depict 0/150 db dynamic topography for the years 1982 to 1984. Since the data permit similar patterns of contours to be drawn for these years, the principal circulation features must remain the same from year to year. The maps do not, however, clearly illustrate the general eastward flow through Parry Channel and the southeastward flow through the Queen Elizabeth Islands which have been inferred from progressive changes in water properties.

Interpretation of these anomaly maps is greatly complicated by the absence of any quasi-plausible reference surface in this complicated network of channels. It has been noted above that in some instances the surface is apparently a suitable "level of no motion", but more typically the entire water column is in motion, and that motion varies with cross-channel position. The only realistic solution is to tie together the dynamical calculations with simultaneous direct current measurements where these are available. (See the companion volume: Residual Flows in the Northwest Passage.)

2.9 Regional Circulation and Baroclinic Volume Transports

Referring to Figure 2.52 which summarizes the regional circulation in winter, the overall flow pattern can be described as follows. There is a general west to east flow from the Arctic Ocean into M'Clure Strait, through Parry Channel and into Baffin Bay via Lancaster Sound. In a similar way there is a general northwest to southeast flow from the Arctic Ocean through the channels of the Queen Elizabeth Islands and Lancaster Sound to Baffin Bay. Superimposed on this are more detailed patterns that show a high degree of consistency throughout the region. The right-bounded counterflowing currents in the channels and the cyclonic flows at channel junctions are the most notable. The patterns is consistent with flows derived from ice motion (MacNeill et al., 1978; Black, 1965; Marko, 1977; de Lange Boom et al., 1982), surface drifter studies (Fissel and Marko, 1978) and results from other investigators (Bailey, 1957; Collin, 1963; Herlinveaux, 1974; Fissel et al., 1982).

Although counterflowing currents are found in Prince of Wales Strait, a net northeastward transport is inferred for winter conditions. This transport direction is consistent with that found by Bailey (1957) in summer. The ice drift data of Black (1965) and Marko (1977) and the direct current measurements of Greisman and Lake (1978) are consistent with the southward flow through Byam Martin Channel and Crozier and Pullen Straits.

In Barrow Strait, Lancaster Sound and adjacent channels, the summer surface drift data (Fissel and Marko, 1978; Fissel et al., 1982) confirm such features as the clockwise circulation around Griffith Island, the northward flow on the east side of Wellington Channel, counterflowing currents in Prince Regent Inlet, southward flow on the west side of southern Peel Sound and the more complex and confused motion in Lancaster Sound. Furthermore, the ice drift results of Marko (1977) also show a net eastward motion in Barrow Strait and Lancaster Sound. The counterflowing currents in Lancaster Sound are also supported by the distribution of nutrients, with the westward flowing water on the north side having the lower nutrient levels typical of Baffin Bay water as sugggested by the results of Jones and Coote (1980).

For the channel cross-sections measured, volume transports based on the calculated baroclinic flows were determined. The transports were estimated for each pair of adjacent stations separately and were then combined to give the resulting transport for the section as a whole. First the flow cross-section was calculated for a station pair using the station separation and the mean water depth determined from the cross-sectional plots (ie. the depth at the mid-point between stations since the bottom was approximated by straight line segments). Next the baroclinic velocity profile for the station pair was extrapolated to the surface and to the mean depth for the section: since the velocities were determined assuming the surface was the level of no motion, a straight line was used to extend the upper part of the profile to a zero value at the surface; at the lower end of the velocity profile a constant value equal to the velocity at the bottom of the profile was used to extrapolate to the depth equal to the mean depth of the section. A planimeter was then used to measure the area under the velocity-vs-depth curve and the depth averaged velocity was calculated. At channel boundaries it was assumed that the mean velocity for the closest station pair applied right in to the shore. The method of extrapolation at the bottom of the velocity profile was a compromise although it was consistent for all the sections. Some velocities tended to zero with depth while others tended to increase with depth at the bottom of the velocity profile. To check the effect of the method of extrapolation, a few sections had the transports calculated using the assumption that the velocity at the bottom went to zero. The difference between the two methods was about 10% in volume transport for the section as a whole.

Uncertainties in calculated transports due to observational error have been estimated and tabulated in Section 2.1. These estimates do not include uncertainty introduced by the possible inability of the station array to resolve the baroclinic flows.

The calculated volume transports for the channel cross sections and estimated observational uncertainties are displayed in Figures 2.53 to 2.58. As can be seen, based on continuity considerations alone there is a non-zero barotropic component in a number of the cross sections. The assumption that the level of no motion is at the surface is not true in all cases.

Using the Lancaster Sound section as an example, a mean easterly surface flow of 6.7 cm/s (Fissel et al., 1984) over this section (cross-sectional area 26.8 x $10^6 m^2$) changes the volume transport from a westerly flow of 1.11 x 10^6 m^3/s to an easterly flow of 0.69 x 10^6 m^3/s. The latter value is in general agreement with values reported by Collin (1963) and Muench (1971) that range from 0.6 x 10^6 to 1.5 x 10^6 m^3/s towards the east.

In addition, at a number of channel cross sections the uncertainty in the flow estimates is of the same magnitude as the calculated flows, e.g. in the 1982 data (Figures 2.53, 2.54, 2.55). Further inconsistencies can be attributed to the different dates on which the measurements were made. Anywhere from one week to over two weeks elapsed between the first and last section measurements in a given region. The best results (in terms of signal-to-noise ratio and consistency) are found in the central sills and Lancester Sound areas in 1984 (Figures 2.57 and 2.58). As can be seen in Figure 2.58, the results are quite consistent with the exception of the Baillie-Hamilton Island section.

Note that in Section 2.1 it was assumed that transport errors between adjacent station pairs were independent. This is incorrect. Transport uncertainties quoted herein should be corrected (crudely) by division by (n-1), where n is the number of stations in the section.

3. HORIZONTAL DISTRIBUTION OF HYDROGRAPHIC PROPERTIES

3.1 Surface Layer Thickness

The thickness of the surface layer in winter in Arctic waters can be defined either as the depth at which sigma-t differs by some threshold (here 0.02 kg/m^3) from a nominal "surface" value (this is the conventional "mixed" layer) or as the depth at which the temperature differs by some threshold (here 0.01 C degrees) from the "surface" value. Since the surface temperature is invariably close to freezing, the second definition identifies waters which have had recent contact with the surface via an oblique path, even though they appear stably stratified in vertical profile. It thus refers to a "ventilated" layer depth; generally temperature increases monotonically with depth in the Archipelago but in some instances temperature is observed to decrease before increasing. The mixed-layer thickness is an indication of how deep mixing has penetrated during the winter. In most areas the mixing is convectively driven by brine rejection during ice formation and augmented by shear-forced mixing due to ice movement and wind in autumn, but near the shallow central sills the dominant mechanism may be shear-forced mixing induced by tidal flows in benthic and lateral boundary layers.

The thickness of the surface layer whether defined by density difference or by temperature difference is similar in the overall pattern. Figures 3.1 to 3.3 show the surface layer thickness in the years 1982 to 1984 respectively. Each figure shows both the density-based and the temperature-based layer thickness. In 1982 the surface layer in M'Clure Strait was much thicker than in the following years (30 to 50 m as opposed to 15 to 30 m) whereas in western Barrow Strait the layer was thicker in 1984 than in 1983.

A number of factors influence the thickness of the surface mixed layer: the amount of old ice cover before and during freeze-up, the strength of the winds, the degree of stratification, the rate of heat loss (which depends on temperature, ice thickness and snow cover) and the degree of current induced mixing. The noted inter-annual differences in M'Clure and Barrow Straits appear to be due to interannual difference in ice and wind conditions in the fall. For example, in 1981/82 M'Clure Strait had the greatest amount of open water and the strongest winds compared to the following two years (see Figure 1.19), creating the best conditions for surface layer deepening by wind events. However, because the mean 5-to-50 dbar salinities were also higher in 1981/82 (see Figures 3.11 to 3.13), a slightly weaker overall stratification in this winter may also have played a role. Annual differences in air temperatures and snow cover did not show a pattern that correlated with the surface layer thickness.

In western Barrow Strait, 1983/84 brought higher winds and more open water than did 1982/83. Again low air temperatures did not correlate with a thicker mixed layer as 1982/83 had below normal temperatures (see Figure 1.18). Because both the 5-dbar salinities and the mean 5-to-50 dbar salinities showed a similar horizontal distribution in the two years (see Figures 3.5, 3.6, 3.12 and 3.13), changes in overall stratification were probably not a factor in accounting for the differences.

The overall pattern of surface layer thickness is similar to that reported by Fissel et al. (1984b), with the thickest surface layers found on the western approaches into the channels leading into the Archipelago and the thinnest layer in the western approaches to the central sills. The central sills region shows a highly variable and sometimes ill defined surface layer due to the mixing occurring in this region. Highly variable surface layer thickness is also found in the Prince Regent Inlet and Lancaster Sound region.

Generally, because the surface layer thickness as deduced on the basis of temperature is close to that based on density, the archipelago can be judged as an area where advective halocline ventilation in winter does not occur. The regions excluded from this generality are those close to open ice edges: Amundsen Gulf (1982) and eastern Barrow Strait, Lancaster Sound and Prince Regent Inlet (1982 - 4).

3.2 Surface Layer Salinity

The salinity at 5 decibars (usually within the mixed layer) reflects the opposing influences of summertime additions of freshwater (runoff, ice melt, etc.), of wintertime brine rejection from growing ice and of the entrainment of saline waters through turbulent erosion of the halocline. At locations where a 5-decibar value was not available, the value at the nearest available depth (within about 2 decibars) has been chosen. No significant errors are introduced by this compromise since the mixed-layer depth was generally 10 meters or more.

The mean salinity of the water column between 5 and 50 decibars is useful when examining changes in water-column freshwater content between years, since it avoids the misleading correlation between surface salinity and mixed-layer depth, (i.e. low surface salinities generally coincide with shallow mixed layer depths).

The salinity at 5 decibars is shown in Figures 3.4, 3.5 and 3.6 for the years 1982, 1983 and 1984 respectively. In general the lowest salinities (around 31) are found in southern Viscount Melville Sound. However in 1984 the lowest salinity values in the study area (as low as 30.0) were found in Hazen Strait; these values were found in a layer only 2 to 5 m in thickness. Relatively low salinities may be a feature of the Hazen Strait area; in 1983 the 5-decibar salinities were also low (31.4) and Peck (1980) reported low values (31.6) at his 1979 southern station in Hazen Strait (Figure 3.9). A combination of summer-time ice melt with a persistent ice cover throughout the summer (allowing relatively little new ice formation and reducing wind induced mixing) could produce the shallow surface layer of low salinity observed. Ice cover data in Section 1.3 illustrate the persistence of ice throughout the summer in this area. Low surface salinity in Viscount Melville Sound is hypothesized to be a consequence of a summertime transport of water from Amundsen Gulf, where surface salinities are generally lower because of significant run-off and ice melt. MacDonald et al. (1978) measured surface salinities in the range 23-30 in August 1977 in Amundsen Gulf, and in the summer of 1954 Bailey (1957) observed the northward movement of low salinity (in the range 24 to 30) surface water in Prince of Wales Strait. In the summer of 1962, salinities below 27 were reported by Barber and Huyer (1971); see Figure 3.10. Once within the Sound, these waters pool within the gyre, and are subject to negligible salinity increase since new ice growth is severely limited by a heavy persistent ice cover. In 1982 the low surface salinities in the Sound coincided with low mean salinities in the upper 50 m (Figure 3.11) and with a region of relative thin surface mixed layer (Figure 3.1). This area also shows a subsurface temperature maximum in the temperature profile, similar to that found in Amundsen Gulf and the southern Beaufort Sea.

Higher 0 - 50 dbar average salinities in the region of the central sills were observed in all three years (see Prinsenberg and Sosnoski, 1983b for 1982 data). This observation is consistent with an upward flux of salt over the sills, due to strong vertical mixing, to isohaline upwelling and to a possible greater injection of salt resulting from greater wintertime ice growth in this seasonal ice zone. Surface salinities in the Lancaster Sound and Prince Regent Inlet area are of intermediate value while those of M'Clintock Channel and Peel Sound are relatively low. As illustrated by Figures 3.7, 3.8 and 3.9 (for 1981, 1982 and 1978 respectively), the north to south salinity gradient in Barrow Strait is a consistent feature in winter, with the higher salinities on the north side.

The mean salinity between 5 and 50 dbar (Figures 3.11 to 3.13) shows a similar pattern to the surface salinities. The low mean salinity in Viscount Melville Sound in 1982 (Figure 3.11) coincides with the gyre that is evident in the dynamic height anomaly plots (Section 2.8). As previously suggested, the low salinity water likely moves north in pulses from Amundsen Gulf in summer. Figure 3.10 illustrates the near surface (10 m) surface salinity in the summer of 1962 with low values in Viscount Melville Sound. The general pattern is one of salinity increasing from the north and west towards the central sills region (reaching about 33.0) and then decreasing slightly (to about 32.3) towards Lancaster Sound in the east.

3.3 Temperatures

An indicator of the importance of diapycnal sensible heat flux is the freezing-temperature-departure at 5 decibars pressure. The freezing-temperature-departure is the difference between the actual temperature and the freezing temperature at atmospheric pressure for the salinity at the depth in question. In winter the freezing of the surface in the absence of significant oceanic fluxes of heat produces a mixed layer at or near its freezing temperature. However shear-induced mixing in some areas results in a transfer of heat and salt to the surface layer from the warmer and more saline water below, sufficient to supply the sensible heat demand (by the atmosphere) at the surface and to raise the surface-layer temperature.

The mean freezing temperature departure between the surface and the pressure at which salinity is equal to 33.0 is a measure of the heat content in the water column between the surface and the depth of the 33-isohaline. This layer within the western Archipelago is, roughly speaking, that subject to seasonal and interannual variations in heat content due to solar radiation and surface cooling locally, although advection will also play a role in determining its properties.

An indicator of wintertime halocline ventilation is water of small freezing-temperature departure below the surface mixed layer. Water whose density is increased by brine rejection during ice formation sinks to a depth where its density matches that of the surrounding water, and then may flow along the isopycnal surface (or the seafloor) to a location where it lies below a warmer, less saline water mass.

The temperature on an isohaline surface can be used as a tracer to indicate the evolution or movement of a water mass. It is a peculiarity of low temperature seawater at shallow depths that isohalines and isopycnals are coincident and temperature is a dynamically passive tracer. The temperature on the 33.5 isohaline has been used to track warming of the halocline (Melling et al., 1984), while the temperature on the 34.83 isohaline has been used as a measure of the Atlantic water temperature in the Arctic Archipelago.

Over most of the study area in all three years, small freezing temperature departures at 5 decibars indicate that waters are at or close to freezing (generally within 0.01 C degrees). There are some slightly higher departures (up to .021 C degrees in central Prince of Wales Strait in 1982, 0.026 C degrees off Bathurst Island in 1983 and .046 C degrees in western M'Clure Strait in 1984), which, in view of the calibration accuracy (± 0.005 C degrees), are probably real. The temperature profiles in the surface layer at the 1982 and 1984 stations show no unusual features and are generally close to isothermal with slightly warmer (by about 0.2 C degrees) water below the mixed layer. At the 1983 station located in a depth of 78 m, there is a well mixed water column (salinities ranging from 32.82 to 32.86 top to bottom). Only in the 1984 data are there significant systematic departures from freezing temperature (Figure 3.14), all of which occur in the vicinity of the central sills. The largest values are found in Wellington Channel (0.172 and 0.162 C degrees) with southern Barrow Strait giving one value as high as 0.134 C degrees. Penny Strait showed a range of values (0.039 to 0.103 C degrees) with no cross-channel pattern. This contrasts with the 1983 results reported by Fissel et al. (1984b) where the largest freezing temperature departures were found in Penny Strait (0.086 to 0.242 C degrees). As do high salinities, positive freezing-temperature-departures in winter imply significant vertical oceanic fluxes (of heat) which are driven, presumably, by strong tidal flow in the shallow constricted channels of the central Archipelago. The largest freezing-temperature-departures generally coincide with the waters that have passed through Penny Strait, that is those on the west side of McDougall Sound and the west side of Wellington Channel.

The distribution of the heat content in upper part of the water column, as given by the mean freezing-temperature-departure between 5 decibars and the pressure at which salinity is 33.0, can be found in Figures 3.15 to 3.17. The reason for the choice of this isohaline is that it marks the approximate boundary within the halocline between progressively cooling and progressively warming waters flowing south and east through the Archipelago (Melling et al., 1984; see also Figure 4.5). The general pattern is similar to the 5 decibar salinity and the mixed layer thickness plots, with higher heat content corresponding to lower surface salinity and a thinner mixed layer. Some caution in the interpretation of these figures is required as in some cases the upper salinities exceed 33.0 and heat content is computed to be zero when in fact some heat may be present (eg. in northern Wellington Channel in 1984, Figures 3.14 and 3.17). The heat content in late winter is related in part to the amount of entrainment of warmer halocline waters into the mixed layer, which in turn depends on the amount of ice formation during the winter (and thus on the amount of ice cover left at the end of the previous summer) and on the amount of wind-induced mixing occurring before ice consolidation (which also depends on remnant ice cover at summer's end). Other factors are the upward diffusion of heat from warmer underlying layers (Melling et al., 1984) and radiant heating of ice-free waters in summer. Radiantly heated waters may also be advected from other areas (for example from the Beaufort Sea and Amundsen Gulf into the Viscount Melville Sound). Figure 3.18 shows the distribution of 1962 summertime water temperature at 10 m depth.

The heat content is generally higher in Viscount Melville Sound and the Sverdrup Basin and drops off to the south and east of the central sills. The exceptions to this pattern are the high values found in Peel Sound and Prince Regent Inlet which may reflect local solar heating due to the clearing of the ice cover the previous summer. To the east and west of the central sills heat content is largely determined by heat in halocline water below the surface mixed layer, not by heat in the upper mixed layer (see Figures 3.14, 2.32, 2.43 and 2.48). The heat content of the halocline is related to the circulation of these waters. The water passing into Queens Channel becomes very well mixed with temperature around freezing and salinity about 33 (see section 2.6). On the other hand the waters of Peel Sound and Prince Regent Inlet are more stratified, deriving their heat from summer insolation and from an upward diffusion of heat from warmer underlying layers (Melling et al., 1984). Year to year variations in a given area are evident although some may be an artifact of the different re-sampling locations used. For example in Viscount Melville Sound the mean departure exceeded 0.2 C degrees in only one location in 1982, did not exceed 0.168 C degrees in 1983 and exceeded 0.2 C degrees over an apparently larger area in 1984. The pattern suggests that the variation in the amount of summer heating of the water is responsible: 1981 and 1983 had more open water than 1982 in the late summer.

The temperature on the 33.5 isohaline surface is a measure of progressive thermal change within the halocline in the Archipelago. This parameter is contoured in Figures 3.19 to 3.21 for the years 1982, 1983 and 1984. These plots display the warming of Arctic Water in the pycnocline as this water moves eastward through the Archipelago (Melling et al., 1984). Although there is some year-to-year variation, the overall pattern is maintained. The water entering through the Queen Elizabeth Islands shows greater or more rapid warming with distance from the Arctic Ocean than that entering Parry Channel. Note that water of 33.5 salinity does not appear to cross the central sills (see section 2, particularly Figures 2.2 to 2.6).

In a similar way the temperature on the 34.83 salinity surface is an indicator of the temperature change in the Atlantic water layer in the Arctic Archipelago. Because this water is found at greater depth, it was sampled at fewer locations within the Archipelago. Thus only the 1982 and 1983 data provide sufficient detail for contouring these temperature values (Figures 3.22 and 3.23). The broad features are similar in the two years. The warmest water is found on the north side of M'Clure Strait and Viscount Melville Sound. The significance of these apparent spatial variations may be low, since their magnitude is little greater than the estimated accuracy of temperature measurement. Melling et al. (1984) discuss the implications of these low Atlantic water temperatures found in the Archipelago. One feature of note is that in 1984 the temperature on the 34.83 salinity surface in western M'Clure Strait was significantly cooler than in 1983 (0.26 to 0.29 degrees C compared to 0.30 to 0.33 degrees C). However, in central Viscount Melville Sound temperatures were closer (0.33 and 0.31 degrees C) in the two years.

4. REGIONAL T-S CHARACTERISTICS AND WATER MASS EVOLUTION

4.1 Prince of Wales Strait

Prince of Wales Strait is the transition zone between surface waters of Amundsen Gulf in the south and those of Parry Channel in the north. Because the T-S characteristics of all but the southernmost line of stations are close to those of M'Clure Strait and Viscount Melville Sound (see Figure 4.1; Figures 1.7 to 1.8 for station locations), the modification of northward flowing waters is largely completed in the region south of the Princess Royal Islands. The southern stations in the strait have T-S characteristics similar to Amundsen Gulf (Figure 4.2). However, within most of the strait, the variation in temperature at a given salinity is less than 0.03 C degrees. The near linearity of the T-S plot indicates that the water properties in the strait are maintained by mixing in the presence of advective inflows from both ends.

At the northeastern end of the strait, some nearby stations have been occupied in several years, thus providing information on interannual variations of T-S properties (Figure 4.3). Station Q04 (1983) of Fissel et al. (1984b) is close to 1982 station CTD14 and station 43 (1977) of Peck (1978) is close to 1982 station CTD19. Comparison of these stations indicates that although all were occupied at about the same time of year, the highest surface layer salinities (32.4 to 32.7) were present in 1982; in 1983 salinities were lower (31.8 to 32.6), while the lowest values (as low as 30.3 at the surface) occurred in 1977. The 1962 summer station (58) indicates the warmer surface water present in summer. Below 50m the variation between years is not much greater than between nearby stations in a given year. Although the data are not extensive enough to indicate the persistence of these features in any one year, the low surface salinities in 1977 remained unchanged for a period of at least 10 days (Peck, 1978). On shorter time scales, repeated casts in 1982 at several stations indicated that the T-S characteristics did not change by more than 0.03 C degrees at a given salinity, and that salinity variations at a given depth were less than 0.05 units over periods of half an hour to 69 hours at a station. The T-S curves showed no significant changes in shape.

4.2 Western Parry Channel

Figure 4.4, showing 1983 data in Western Parry Channel (Figures 1.4 and 1.9 for station locations), illustrates that the principal features of the water masses and of their variation within the waterway were the same as in 1982 (Fissel et al., 1984a,b; Melling et al., 1984): the sub-surface temperature maximum decreases from west to east, being virtually absent in eastern Viscount Melville Sound, and the nearly isothermal cold halocline of the Arctic Ocean warms towards the east in Parry Channel. A composite plot of selected stations from the years 1982, 1983 and 1984 is shown in Figure 4.5. Here it is clear that the water with salinity near 33.0 shows very little temperature variation (less than 0.1 C degrees) throughout the deep portion of the channel, the exception being the 1961 summer data. Much greater spatial variation is found at salinities between 33.0 and 34.0. The range of temperature values in the deeper water is also very small. Variations at salinities less than 33 occur both spatially and interannually.

It is observed that the T-S characteristics of the near surface waters at the northern end of Prince of Wales Strait lie between those of nearby waters in southeastern M'Clure Strait (salinity of 32.5) and those in southwestern Viscount Melville Sound (salinity as low as 31.5; refer to Figures 4.6, 4.7, 4.8). Close to the Victoria Island shore in Prince of Wales Strait (at station CTD21), the T-S curve is nearly linear, indicating the strong influence of mixing on the northeastwardly moving water. The plots suggest that the mixing has been less effective on waters in mid-channel (station CTD19) than on those near the southern shore and that the mid-channel characteristics are closest to those of nearby waters of Viscount Melville Sound (Figures 4.7 and 4.8). A mixing in Prince of Wales Strait of water from southeast M'Clure Strait with that found in the middle of Prince of Wales Strait could produce the characteristics of the water near the Victoria Island shore, and a mixing of this water with that from further offshore in Viscount Melville Sound could produce the water of intermediate properties found at station CTD32. As already mentioned water seen at CTD33 is thought to originate in Amundsen Gulf in summer.

The core of the westward flow south of Ross Point in northern Viscount Melville Sound contains a relatively large volume of water with T = -1.8 to -1.6 degrees C and S = 32.6 to 32.8 (see Figure 2.21). Water of these characteristics (see Figure 4.9) is not evident elsewhere and may be a pulse of water formed by brine rejection due to ice formation in the shallower waters southeast and east of Melville Island (see Figure 1.9 for bathymetry). Below this water mass, in the Ross Point section and further west in the Winter Harbour section (Figure 2.20) and in other locations, water with characteristics T = -1.6 to -1.5 degrees C and S = 32.8 to 33.0 is common (Figure 4.9). Water with similar characteristics is found in Byam Martin Channel (e.g., stations C3 and C5 in 1982), in Hazen Strait (Station D2, 1982, see Figure 4.10), and also in northern Prince of Wales Strait (Figure 4.6). Thus this water may have several sources. In Western Parry Channel water of this salinity range has a narrow temperature range (see Figure 4.5). Further east off the southeast side of Bathurst Island, water at similar salinities is significantly cooler (Figure 4.11).

Station 48 near the centre of Viscount Melville Sound was occupied in all three years and the measurements there provide a measure of the internnual variability at a particular site (Figure 4.12). Data for this site from other studies are also plotted in this figure. Again very little temperature variation at salinities between 32.9 and 33.0 is evident, with the exception of the 1967 summer data. In 1984 at this station a subsurface temperature maximum and a nearly isothermal halocline, both reminiscent of Beaufort Sea profiles were more evident than in the other years. Such a variation, however, is better viewed as a reflection of small changes in the spatial distribution of recurrent hydrographic features, rather than as an indication of significant regional interannual changes. For example, in 1983 at station O40 (about 100 km west of station 48) a temperature maximum was observed (Fissel et al., 1984b). Low surface layer salinities (30.2) have been observed here in 1977 by Peck (1978). An interannual difference of this magnitude at the surface may be more important, reflecting changes in the amount of the transport through Prince of Wales Strait or changes in the surface-layer entrainment and ice growth in Viscount Melville Sound due to differences in summertime weather and ice cover. The years 1982 and 1983 show the presence of the T = -1.6 to -1.5 °C, S = 32.8 to 33.0 water at depths between about 70 and 110 m. The variation of the temperature at salinities around 33.4 shows that the amount of heat possessed by water at these salinities is variable (but this variability again is likely the result of shifts in the flow pattern). Some change is evident over about a week in the 1983 data. The deeper, more saline water shows much less variability, particularly at salinities above 34.6.

4.3 Queen Elizabeth Islands

The waters among the Queen Elizabeth Islands show a transition in properties similar to that seen in western Parry Channel. Figure 4.13 is a composite based on data from selected stations taken in the area in the years 1982 to 1984. Aside from the two stations 83A08 and 83V01 (north of Penny Strait and in Norwegian Bay) there is less temperature variation at a given salinity, for salinities between 33.0 and 33.8, than in western Parry Channel (Figure 4.5). This could be due in part to the greater length of Parry Channel which gives more time for heat transfer. However, the main reason appears to be that by the time the water reaches Prince Gustaf Adolf Sea a significant amount of heating has already occurred relative to the Arctic Ocean water temperatures at these salinities. This is not the case in western Parry Channel. Temperatures are higher (by up to 0.2 C degrees at a given salinity) than in western Parry Channel. However, the very small temperature variation between stations (< 0.1 C degrees) seen at salinities near 32.9 in Parry Channel is not found in the data from the Queen Elizabeth Islands. Neither a sub-surface temperature maximum nor an isothermal halocline are evident in most of the data. Only one station (83V01 in Norwegian Bay) shows a slight but significant sub-surface temperature maximum.

The relatively high temperatures at salinities between 32.8 and 34.0 for two of the stations can be attributed to an upward turbulent diffusion of heat in the water column, forced by energetic tidal flows in the narrower, shallower channels of the Queen Elizabeth Islands. Evidence lies in the linear T-S line between salinities of 32.8 and 33.4 at station 83A08 located just northwest of Penny Strait. The water in Norwegian Bay at these salinities is even warmer than in Penny Strait. The higher temperature may reflect a longer residence time and flow path in the Archipelago with greater heat transfer potential from the deeper water. The non-linear nature of the T-S line indicates that Norwegian Bay is not an area where diapycnal mixing processes are dominant and that there is interleaving of water with slightly different properties.

In Figure 4.14 the progressive warming of the halocline with distance from the Arctic Ocean is evident. A few stations south and east of the central sills are included on this plot to illustrate differences from the stations in the Queen Elizabeth Islands. At station S01 in Wellington Channel, the very narrow ranges of temperature (within 0.2 C degrees of freezing) and of salinity (33.1 ± .05) are evidence that intense mixing and cooling of waters occurs after their passage through Penny Strait. The same water shows up at BLMSS station 65 (Prinsenberg and Sosnoski, 1983c) in eastern Barrow Strait, where it is overlain by lower salinity water originating south and west of Barrow Strait. In Lancaster Sound (station N07) the upper 200 m are colder and more saline than the upper 100 m in the Queen Elizabeth Islands, in contrast to the water below 250 m which is less saline than the deep water in the islands. The deep water in Lancaster Sound originates in Baffin Bay (Muench, 1971).

4.4 Central Sills Region

Because of the mixing of inflowing waters due to energetic tidal action over the sills, because of the higher losses of sensible heat brought to the surface by mixing through thinner or non-existent ice cover and because of the confluence of waters with different properties, the waters over the central sills have variable temperature over a salinity range limited by shallow sill depths. Figures 4.15 and 4.16 illustrate T-S properties for stations located in Barrow Strait. The region is the 'meeting ground' for waters of western Parry Channel, of the Queen Elizabeth Islands and of Lancaster Sound.

As the waters are mixed on passing over the sills, the range of T-S values is decreased. Figure 4.17 illustrates this for the path between Maclean Strait and the northern end of Wellington Channel. The upper water becomes more saline and is cooled, while lower waters are stripped away by sills, and the lower remnants are cooled and reduced in salinity. The current of southward flowing water on the western side of Wellington Channel (section 2.6) retains its narrow salinity range (33.05 ± .05; see Figure 4.18) until it reaches Barrow Strait where the upper 70 m freshens slightly (0.1 units) due to admixture of lower salinity surface waters from Barrow Strait.

The westward coastal current on the south side of Devon Island, which turns north along the east side of Wellington Channel, is distinct by virtue of salinities lower by 0.4 to 0.8 compared to the west side (Figure 4.19). These can be traced half way up the channel but disappear (presumably due to mixing) before the north end of Wellington Channel is reached.

Eastern Barrow Strait contains the transition from the limited T-S range of the waters passing through Penny Strait to the wider range of Lancaster Sound (Figure 4.20). In the upper 50 m is a layer of lower salinity formed by water originating west and south of Barrow Strait. Below 50 m is found modified water from the Queen Elizabeth Islands, and below 200m is found Baffin Bay Atlantic water (Muench, 1971).

There is a southerly flow also from Queens Channel through McDougall Sound to Barrow Strait (section 2.6). This water is slightly less saline (by about 0.2) and warmer (by about 0.1 C degrees) than that at the north end of Wellington Channel (Figures 4.19 and 4.21). The water can be traced from the north end of McDougall Sound to Lowther Island in Barrow Strait and along the south coast of Bathurst Island. Westward flowing water along the south coast of Cornwallis Island and Griffith Island can be traced from Wellington Channel into the east side of McDougall Sound to station 84073. Figure 4.22 indicates how well this water retains its characteristics.

The available data have not provided the detail to trace the water moving eastward along the north coast of Prince of Wales Island. However, as Figure 4.23 shows, the bottom water (below 100 m) in Peel Sound has the same T-S signature as that of water at about 100 m depth west of the sill in Barrow Strait that separates Peel Sound from Viscount Melville Sound (station 84002), confirming the eastward passage of this water over the sill. The low salinity surface waters with slightly elevated temperatures (0.3 C degrees) in Peel Sound appear to come from further south in Larsen Sound where similar surface water exists (Figure 4.24). Although water of similar salinity is also found in Viscount Melville Sound, it has temperatures at freezing (Figure 4.25) and thus is not a possible source of surface water in Peel Sound.

The water on the east side of Peel Sound that moves first north then east along the shore of Somerset Island shows a progressive cooling at salinities below 33, likely because of an isopycnal interleaving and mixing with waters resembling those north of Prince of Wales Island, just to the west (compare 4.25 and 4.26). Below 50 to 100 m the T-S characteristics are identical (Figure 4.26). On the north side of Somerset Island the T-S structure is more complicated as different water masses interact and interleave (Figure 4.27). In this plot the water from Peel Sound and the near-freezing water from Penny Strait/Queens Channel of salinity near 33.0 can clearly be distinguished on the basis of temperature. Local formation of this cold saline water along the Somerset Island shore is not likely, since the coast is very steep and the relative volume of water modified by brine rejection would not be significant. It is the confluence of these differing waters that produces the principal hydrographic characteristics of Barrow Strait: the north-south salinity gradient at the surface and the cold deeper water with salinities near 33.0.

Figure 4.28 shows the range of T-S variation in the central sills region. One incidental feature of interest is the lower temperatures found over much of the water column in Penny Strait in 1982 (station A1) relative to 1983 and 1984.

4.5 Eastern Parry Channel

The water found partway down Prince Regent Inlet on the west shows some similarity to the interleaved waters of eastern Barrow Strait (Figure 4.29), but generally this figure suggests that this area of Prince Regent Inlet is little influenced by outflow from Barrow Strait. On the east side of the inlet the T-S structure is very similar to that on the south side of Lancaster Sound. The slightly lower temperatures in Lancaster Sound could result from a greater sensible heat loss through the generally mobile ice cover of this channel (Marko, 1977) or from a greater influence of near freezing waters advected eastward from Barrow Strait. In 1984 the edge of the landfast ice did not reach to the east of the sampling stations until sometime in March, only about a month before the measurements were made; prior to that the edge was west of Prince Regent Inlet; however, the north end of Prince Regent Inlet was fast by the end of February.

The T-S plots for Lancaster Sound and Prince Regent Inlet show much less location-dependent variation than do those for western Parry Channel (Figure 4.30). However, on a year-to-year basis there is variation, particularly at the higher salinities. At salinities above 33.3, the eastern Parry Channel waters have higher temperatures than the same salinity water north and west of the sills.

Another feature of Lancaster Sound water is the presence of temperature minima well below the surface mixed layer, at salinities up to 33.8 (Figure 4.31). Such are not observed at these salinities over or to the west of the sills. Melling et al. (1984) postulate, and the measurements of Lemon and Fissel (1982) provide some empirical support, that convective mixing in eastern Lancaster Sound and northwestern Baffin Bay can extend to depths greater than 200 m and involve freezing-temperature waters of correspondingly high salinity. If this cold water were advected along isopycnals beyond the region of formation, specifically into western Lancaster Sound, it would provide a likely means for the generation of the observed temperature minima in a T-S envelope provided by Baffin Bay Atlantic water. The water from Barrow Strait and Wellington Channel though suitably cold is of too low a salinity to produce the deeper layers. A cold halocline at salinities of around 33.4 has not been observed in wintertime Lancaster Sound data.

5. FINE STRUCTURE ON PROFILES

In the present context fine structure is taken to be small scale (approximately 1 to 10 m) vertical variation in temperature or salinity. These variations are generally much less than 1 C degree or 1 salinity unit in magnitude in the profiles measured here. The judgment of whether a profile has fine structure is subjective.

Two types of fine structure are common: gradient-reversal fine structure and step-like fine structure. The gradient-reversal fine structures appear in temperature as layers of distinctly different temperature imbedded in the larger scale structure (see Figure 5.1 for an example). Step-like fine structure occurs in temperature and/or salinity (see Figure 5.2).

The causes of the two types of fine structure differ. The gradient-reversal type is likely the result of the advection and interleaving along its isopycnal of a layer of water of different temperature than the surrounding water. The source of the interleaved water is not often obvious. It is likely that such structures require investigation using closely spaced CTD profiles. An interleaved layer with temperature lower than its surrounds may be related to cold water formed at the surface due to ice formation. A layer with temperature warmer than the surrounds could originate at a front separating waters of the same density flowing together from different areas.

The step-like fine structure looks similar to the double diffusive layering discussed by Melling et al. (1984) although it is larger in vertical scale (~ 10 m) and somewhat asymmetric. Since it is found in areas of relatively strong flow (eg. narrow channels, points of land), it is likely related to diapycnal mixing. Melling et al. (1984) postulate that the layers are tidal boundary layer remnants advected away from the channel walls.

The following treatment of fine-structure observations is qualitative, and affected by such factors as non-uniform station coverage, and a programme emphasis on other facets of hydrography. Fine-structure observations herein may be regarded as incidental.

Figures 5.3 and 5.4 illustrate the observed distribution of late winter fine structure in the years 1982, 1983 and 1984. Prince of Wales Strait showed the least fine structure of any of the regions with about 58 % of the profiles indicating no significant fine structure. Of those profiles showing any type of fine structure, about 2/3 showed steps.

For the Central Sills area, 55 % of the profiles showed no significant fine structure but about 2/3 of the profiles with fine structure showed gradient-reversals.

Western Parry Channel fell in the middle in terms of the occurence of fine structure. About 1/2 of the profiles showed gradient-reversals and 31% showed no significant fine structure.

The Queen Elizabeth Islands and eastern Parry Channel areas showed the greatest amount of fine structure with only 15 % and 17 % indicating negligible fine structure, respectively. In the Queen Elizabeth Islands 1/2 of the profiles showed steps while in eastern Parry Channel about 1/2 of the profiles showed gradient-reversals.

The relationship between the type of fine structure and where it is found gives some indication as to the cause of the fine structure. The gradient reversal fine structure is found throughout the region although not with the same frequency as the step-like structure, as noted above. It tends not to occur in regions of strong mixing or in iso-thermal and iso-haline structure. It is sometimes found in profiles that also show steps. In addition it does not appear to occur with greater frequency in areas such as near side channels, shelf edges, etc. On the other hand, the step-like fine structure is definitely associated with areas where strong flows are found or where flows might be accelerated such as in channel constrictions and near points of land. It is found in the narrowest section of Prince of Wales Strait, Byam Martin Channel, Penny Strait, near shore on either side of Maclean Strait, near Lowther and Griffith Islands in Barrow Strait and other similar locations. In only a few cases was there no direct connection with strong flow, although even here currents likely play a role. For example in the middle of Lancaster Sound there is a relatively large vertical shear in the baroclinic current.

We conclude that fine structure is likely to be found in most areas of the Archipelago with the exception of areas with nearly iso-thermal and iso-haline structure such as Queens Channel, northern Wellington Channel and McDougall Sound. The step-like fine structure is likely to be found in areas of strong flow such as Byam Martin Channel and Penny Strait.

6. SUMMARY

During the years 1982, 1983 and 1984, oceanographic surveys were conducted in the channels of the Canadian Arctic Archipelago generally known as the Northwest Passage. The primary data on which this report is based are the CTD measurements acquired by the Ocean Physics Division, IOS during March and April of the above years. Data from concurrent programs and previous studies were also utilized.

The main water masses have been known for many years. Within the Archipelago the waters consist of Arctic (Surface) Water to depths of about 275 m and Arctic Intermediate Water (Atlantic Water) between this depth and about 900 m or the bottom in the shallower channels. The winter Arctic Water mass can be broken down into several sub-layers: a surface mixed layer extending down as deep as 50 m with a temperature at or close to freezing; a subsurface temperature maximum layer (found at some but not all locations) beneath the mixed layer but not deeper than 75 m, with a temperature maximum of -1.3 to -1.0 degrees C; a cold halocline with temperature within a few tenths of a degree of freezing and/or a thermocline with both temperature and salinity increasing with depth into the Atlantic Water mass. The Atlantic Water is generally defined as water with a temperature above 0 degrees C and typically is located below 200 m in the Eurasian Basin and northeastern Baffin Bay and below 275 m in the Canada Basin and northwestern Baffin Bay. Maximum salinities in the Atlantic Water are just below 35 except in Baffin Bay where maximum values are near 34.5.

This report has focussed on the small differences in the properties of these waters within the Canadian Arctic Archipelago, and has used this information, in conjunction with the dynamical method, to infer water circulation, mixing and property evolution in this very complex system of wide interconnecting channels. Principal findings are briefly reviewed for each waterway in the following paragraphs.

The movement of the water through the Archipelago is reflected in the systematic spatial distribution of water properties. The Arctic Water of the halocline warms as it moves east and south through the islands from the Arctic Ocean. Conversely the underlying Atlantic Water is cooled. The increased heat content of the halocline becomes available to the surface layer by entrainment and diffusion in the turbulent waters of the central sills region. Active mixing in this region is indicated by the linear T-S characteristics, the relatively warm saline water near the surface and the progressively narrowing range of T-S properties. A subsurface temperature maximum near 50 m at many stations west of the central sills is evidence for an inflow path from the southern Archipelago of summertime heated water from Amundsen Gulf and the Beaufort Sea, via M'Clure Strait, Prince of Wales Strait and possibly via Dolphin and Union Strait, Coronation Gulf, Larson Sound and Franklin Strait. Waters also flow to the central Archipelago from the east via a meander of the Baffin Current into Lancaster Sound. Waters from the east can be traced as relatively low salinity flows as far west as eastern Barrow Strait, and the eastern side of Wellington Channel. Thereafter their distinct properties are rapidly lost through mixing with waters inflowing from the west.

Surface-layer salinities increase from 31.0 to 31.5 in the Arctic Ocean, to 32.0 to 32.5 at the approaches to the central sills, and to 33.0 in the vicinity of the sills. This progression reflects the upwelling of isohalines towards the central sills, and the energetic vertical diffusion of salt by tidally-induced mixing over the sills. A decrease in surface salinity to about 32.5 east of the sills is a consequence of the inflow of lower salinity surface water from Baffin Bay to the east. The mean salinity in the upper 50 m shows the same pattern.

Wintertime mixed-layer thickness shows both regional and inter-annual variations that are related to the amount of remnant ice cover prior to freeze-up, the amount of ice growth, the occurrence of strong winds during freeze-up and the degree of flow-induced mixing. The thickest mixed layers are found in the western approaches to the Archipelago and in Wellington Channel and McDougall Sound. The thinnest are found in the western and northern approaches to the central sills region. East of the sills the mixed layer is highly variable in thickness, poorly defined or non-existent. Surface salinity and mixed-layer depth tend to be correlated in any area from year to year.

The increased variability in water characteristics east of the central sills reflects the confluence of waters of different properties in eastern Barrow Strait. In Lancaster Sound, where ice cover is mobile and variable, there are indications of a variable circulation pattern with large eddies. The summer circulation pattern is maintained in winter although with a reduced strength.

On the basis of the T-S relationships, regionally distinct water masses can be identified and traced over limited distances within the Archipelago. The uniform water of Queens Channel and a homogeneous water found in Byam Martin Channel are examples. Such signatures have enabled the detection of clockwise flows around Cornwallis, Bathurst and Melville Islands which counter the general flow tendencies in Parry Channel and the detection of a cyclonic gyre in Viscount Melville Sound.

A cursory examination of the fine structure in temperature and salinity profiles reveals two common forms: gradient reversals in the temperature profiles indicative of interleaving, and step structures in both the temperature and salinity profiles indicative of efficient mixing over limited vertical extents. Fine structure is found in most areas of the Archipelago except where water is nearly homogeneous (e.g. Queens Channel). The gradient-reversal fine structure is found in areas of weak flow, or areas where different water masses are confluent, whereas the step-like fine structure is most likely to be found in areas of strong tidal flow such as Byam Martin Channel, Penny Strait, Prince of Wales Strait and near Lowther and Griffith Islands.

Prince of Wales Strait separates stratified water having temperature maximum and minimum layers in the halocline (Amundsen Gulf) and less stratified water having a more-or-less coincident thermocline and halocline to the north (Viscount Melville Sound). Surface-layer salinity increases with distance up the strait from the Gulf indicating a modification by mixing due to the moderate (30 cm/s) tidal flows within the strait. A net northward transport is inferred, although counterflowing currents apparently co-exist on opposite sides of the strait (with northeasterly flow on the east side). This pattern of cyclonically sheared boundary currents is common throughout the Archipelago. There are indications that significant volumes of radiatively warmed water from the ice-free Amundsen Gulf move northward into southern Viscount Melville Sound in summer without substantial property changes.

Along the length of Parry Channel there is a transition from the Arctic Ocean waters through the variable central sills area to the waters of Baffin Bay. Within the halocline and below sill depth the waters are colder and less saline on the eastern side of the sill than they are on the western side. West of the sills an increasing temperature in the halocline with distance from the Arctic Ocean is a persistent feature over the years 1982 to 1984. A similar transistion in halocline temperature takes place through the Queen Elizabeth Islands from the Prince Gustaf Adolf Sea to Lancaster Sound, although the mixing in Penny Strait and Queens Channel is much more effective than that in the wider, deeper Barrow Strait. The result in Queens Channel is nearly homogeneous water with salinity 33.0 ± 0.1 and temperature of -1.7 ± 0.1 degrees C. The greater effectiveness of mixing north of Cornwallis Island can be attributed to the stronger tidal currents found there, the rugged sill region greater than a tidal excursion in length and the weaker initial stratification of inflowing water due to its origin in the northeast Canada Basin, far from regions of substantial summertime melting or riverine inflow.

Viscount Melville Sound is characterized by relatively low surface salinity (31.0 to 31.5) and a subsurface temperature maximum layer, particularly in the southern and western parts. The warmer waters enter the region in the summer through Prince of Wales Strait and perhaps through M'Clure Strait. Yearly differences in surface salinity within the Sound are more a reflection of changes in the thickness of the surface mixed layer than of changes in freshwater inflow. Wind strength and the degree of ice cover in the autumn are the key factors affecting the depth of surface-layer entrainment and hence surface salinity. Right-bounded coastal currents peaking near 150 m depth (up to 15 cm/s) are present in Viscount Melville Sound, evident both in the baroclinic currents and in spatial variations in water-mass T-S characteristics. A cyclonic gyre was detected in Viscount Melville Sound during the detailed surveys in 1982, but its existence in 1983 and 1984 was not obvious, perhaps due to limited sampling of that area.

More than a few internal Rossby radii away from the boundaries, flows are generally weak. At channel junctions the right-bounded coastal currents show a tendency to follow the curving shoreline, forming cyclonic meanders into the mouth of the side channel. As a result, for example, the flows are south and west on the east and south sides of Melville, Bathurst and Cornwallis Islands and west and north on the south and west sides of Devon, Cornwallis and Bathurst Islands.

In the Queen Elizabeth Islands a general southward flow from the Arctic Ocean towards the centre of the Archipelago can be inferred both from the distribution of water temperature on isohalines and from baroclinic flow calculations. This inference is substantiated by direct current measurements in Byam and Austin Channels which indicate a net southerly flow. Vertical shears in the baroclinic currents are large in Byam Martin Channel.

A consistent feature over the three winters is the relatively low salinity (near 31) that is found in conjunction with a shallow surface layer in the vicinity of Hazen Strait.

The central sills region around Cornwallis Island is characterized by variable water structure: from nearly homogeneous in temperature and salinity to well stratified differentiated water masses. In this area considerable mixing takes place and water properties undergo a marked change to produce waters distinct from those of the Arctic Ocean or of Baffin Bay. Barrow Strait is the place where the waters of the Arctic Ocean entering through western Parry Channel and those from the southern part of the Archipelago meet and interact with the waters moving south from Penny Strait and Queens Channel and those moving east from Baffin Bay. The waters from Queens Channel, with a narrow range of salinity (33.0 ±0.1) and temperature (-1.7 ± 0.1 degrees C), have a distinctly different T-S signature than those from western Parry Channel, which have a much wider range of both temperature and salinity. As a result on the north side of Barrow Strait, particularly below 50 m, the seawater has a salinity near 33.0 and temperature within 0.2 C degrees of the freezing point, whereas on the south side it has lower surface salinity and greater vertical temperature and salinity gradients. This pattern is consistent with the origins of the right-bounded flows that bring waters into this area. The waters moving south through Wellington Channel and McDougall Sound turn and move westward in Barrow Strait. The eastward moving water on the south side of Barrow Strait swings south into Peel Sound. Bottom water (salinity 33.4) found in Peel Sound originates west of the sill near Lowther Island in Barrow Strait; warmer water of this salinity in Lancaster Sound does not appear to cross the sill north of Somerset Island into Peel Sound.

The major part of the transformation of Arctic Ocean water takes place through mixing induced by the energetic tidal flows and through the increased sensible heat loss to the winter atmosphere through the open water and thin ice of Penny Strait and Queens Channel. Large baroclinic flows with a complex structure are deduced in Penny Strait. The complexity probably reflects the effects of rough bottom topography but may be due in part to internal wave fluctuations aliased into the dynamical calculations. The spatial pattern of water temperature and salinity and direct current measurements confirm the existence of a southerly mean flow.

Wellington Channel shows a greater range of temperature and salinity than McDougall Sound although most of the water has a salinity between 32.8 and 33.2 and a temperature below -1.5 degrees C. A northward flow of lower salinity on the east side of Wellington Channel is a continuation of the eastward flow on the north side of Lancaster Sound. The northward flow loses its identity as it mixes with the waters flowing south through the channel, so that it is no longer detectable beyond the junction with Queens Channel. The baroclinic flow profiles show that the greatest vertical shear is found on the east side of the channel, where the northward flowing coastal water overflows the southward moving water found in the rest of the channel. A similar flow pattern is found in summer.

The underflow of water from Wellington Channel beneath waters from Peel Sound and from western Barrow Strait results in a variable pattern of temperature maxima and minima off the north coast of Somerset Island in Barrow Strait. This area (east of the main sill) has a large baroclinic flow in comparison with the more moderate shears to the west of the sill. Again, the largest vertical shears are generally found near the coastlines. An anti-cyclonic circulation around Griffith Island is apparent.

The low surface salinities (near 31.5) found in Peel Sound could have their origin in Larsen Sound or in Viscount Melville Sound, although in the latter instance the flow cannot be continuous in time. Counterflows are found in Peel sound as in the other channels. However the largest vertical shears were found in the centre of the sound, apparently associated with a northward surface flow above a deep southward flow of higher salinity water.

The water in McDougall Sound has very uniform properties since it all originates, although via different routes, in Queens Channel. The northward flow along the Cornwallis Island shore is very similar in properties to the southward flow at the other shore despite its flow around Cornwallis Island from Queens Channel through Wellington Channel and Barrow Strait and into McDougall Sound. Moderate (3 cm/s) baroclinic flows are observed in the sound.

In eastern Parry Channel waters from both the central sills region and Baffin Bay can be identified. The relatively cold (-1.5 ± 0.2 degree C) near-surface water of salinity 33.0 ± 0.2 can be traced from Barrow Strait to Lancaster Sound and Prince Regent Inlet. It is clear that the waters with these characteristics found in Lancaster Sound are formed in the central Archipelago, and not in the Arctic Ocean as suggested by Muench (1971). The water properties found below 200m indicate that these waters cannot have come from the Arctic Ocean through the Northwest Passage, as might be anticipated considering the shallow depth ($\sim$ 125 m) of the central sills, and hence must have come from Baffin Bay. The near surface waters have different sources reflecting the different flow patterns on opposite sides of the channels. On the north side of Lancaster Sound the water flows westward from Baffin Bay while the eastward flowing water on the south side of Lancaster Sound comes from Barrow Strait and Prince Regent Inlet. The eastward flowing water from Barrow Strait follows the coastline in a meander south into Prince Regent Inlet, with a northward counterflow on the east side of the inlet. The flow patterns are more complex in Lancaster Sound as a result of the existence of large eddies in the flow which produce instantaneous flows that can be significantly different from the mean flow. The baroclinic velocities are larger in Lancaster Sound than further west in Barrow Strait.

Baroclinic transport calculations indicate that M'Clure Strait/Viscount Melville Sound, Byam Martin Channel and Penny Strait contribute equally to the flow through the Archipelago. This implies that about two-thirds of the volume flow passes through the Queen Elizabeth Islands, that is, about 0.54×10^6 ($\pm 0.06 \times 10^6$) m^3/s out of 0.76×10^6 ($\pm 0.24 \times 10^6$) m^3/s. This transport compares well with the results of Collin (1963) and Muench (1971) for Lancaster Sound, which range from 0.6×10^6 to 1.5×10^6 m^3/s. Transports are largely baroclinic in the Queen Elizabeth Islands (Maclean Strait, Hazen Strait, Byam Martin Channel), the Central Sills Region (Barrow Strait, Penny Strait, McDougall Sound, Queens Channel, Wellington Channel, Peel Sound) and eastern Parry Channel (Prince Regent Inlet, Lancaster Sound).

With the exception of the southern section in the strait, flow continuity is consistent in Prince of Wales Strait as is the case for western Viscount Melville Sound and M'Clure Strait. In central Viscound Melville Sound (bounded by Melville Island, Bathurst Island, Prince of Wales Island and Stefansson Island) continuity cannot be determined on the basis of the 1983 data. In the region of the central sills, baroclinic transports are very consistent between sections. The results indicate that the transport is all barotropic in the Baillie-Hamilton Island section and primarily barotropic in the Beechey Head to Somerset Island section. The calculated baroclinic flow directions indicate that, with the exception of Wellington Channel, the surface is not a suitable reference level of no motion. Lancaster Sound and Prince Regent Inlet have large barotropic components of flow.

The current meter measurements obtained during the studies suggest that in the main channels the barotropic transport may be equal to the baroclinic transport. Including crude estimates for barotropic transport (velocities of 1 to 10 cm/s), the transports through the main channels are 0.6 (± 0.5) x 10^6 m^3/s for both M'Clure Strait and Viscount Melville Sound, 0.5 (±0.1) x 10^6 m^3/s for Byam Martin Channel, 0.6 (± 0.1) x 10^6 m^3/s for Penny Strait, 1.0 (± 0.1) x 10^6 m^3/s for Barrow Strait and 1.2 (± 0.2) x 10^6 m^3/s for Lancaster Sound.

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Table 1.1

Typical Calibration Performance for the Guidline Model 8706 CTD, Northwest Passage

Function	Conductivity	Temperature	Pressure
Units	ppt	deg C	dbar
Range Utilized	28 to 35	-2 to +1	0 to 500
Accuracy	± .005	± .005	± .5
Resolution	± .001	± .0005	± .15
Stability	± .005/30 days	± .005/30 days	
Response time	< 50 msec	< 50 msec	< 50 msec

Note: Equivalent salinities used for conductivity specifications.

Table 1.2

Ice Thickness and Snow Depth

Season (Location)	Date Ice Thickness = 50 cm	Snow Depth when Ice 50 cm thick (cm)	End of March Thickness of Ice (cm)	Snow Depth (cm)
1981/82 (M)	29 Oct.	5	158	45
1981/82 (R)	28 Oct.	7	201	7
1982/83 (M)	before 4 Oct.	$\leq$ 12	218	21
1982/83 (R)	5 Nov.	7	163	37
1983/84 (M)	5 Nov.	12	166	42
1983/84 (R)	6 Nov.	3	185	25

Note: for location M = Mould Bay; R = Resolute Bay

Table 2.1

Geostrophic Shear Computation Error Estimates (cm/s) for $\delta\sigma_t = .006\ kg/m^3$, $\rho = 1026.5\ kg/m^3$

Depth (m)	Station Spacing (km) 1	2	3	5	10	20	30	50
10	.58	.29	.19	.12	.06	.03	.02	.01
50	2.9	1.4	.96	.58	.29	.14	.10	.06
100	5.8	2.9	1.9	1.2	.58	.29	.19	.12
150	8.6	4.3	2.9	1.7	.86	.43	.29	.17

Table 2.2

Typical Uncertainties in Calculated Baroclinic Volume Transport

Section	D(m)	L(m)	ΔQ (m^3/s)
Prince of Wales Strait	70	20×10^3	0.7×10^4
M'Clure Strait	450	100×10^3	22.5×10^4
Penny Strait	200	35×10^3	3.5×10^4
Barrow Strait	130	55×10^3	3.6×10^4
Wellington Channel	150	30×10^3	2.3×10^4

D = depth, L = width of channel

Note: It is assumed that the spacing between stations is sufficiently close to resolve the flow details.

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FIGURES

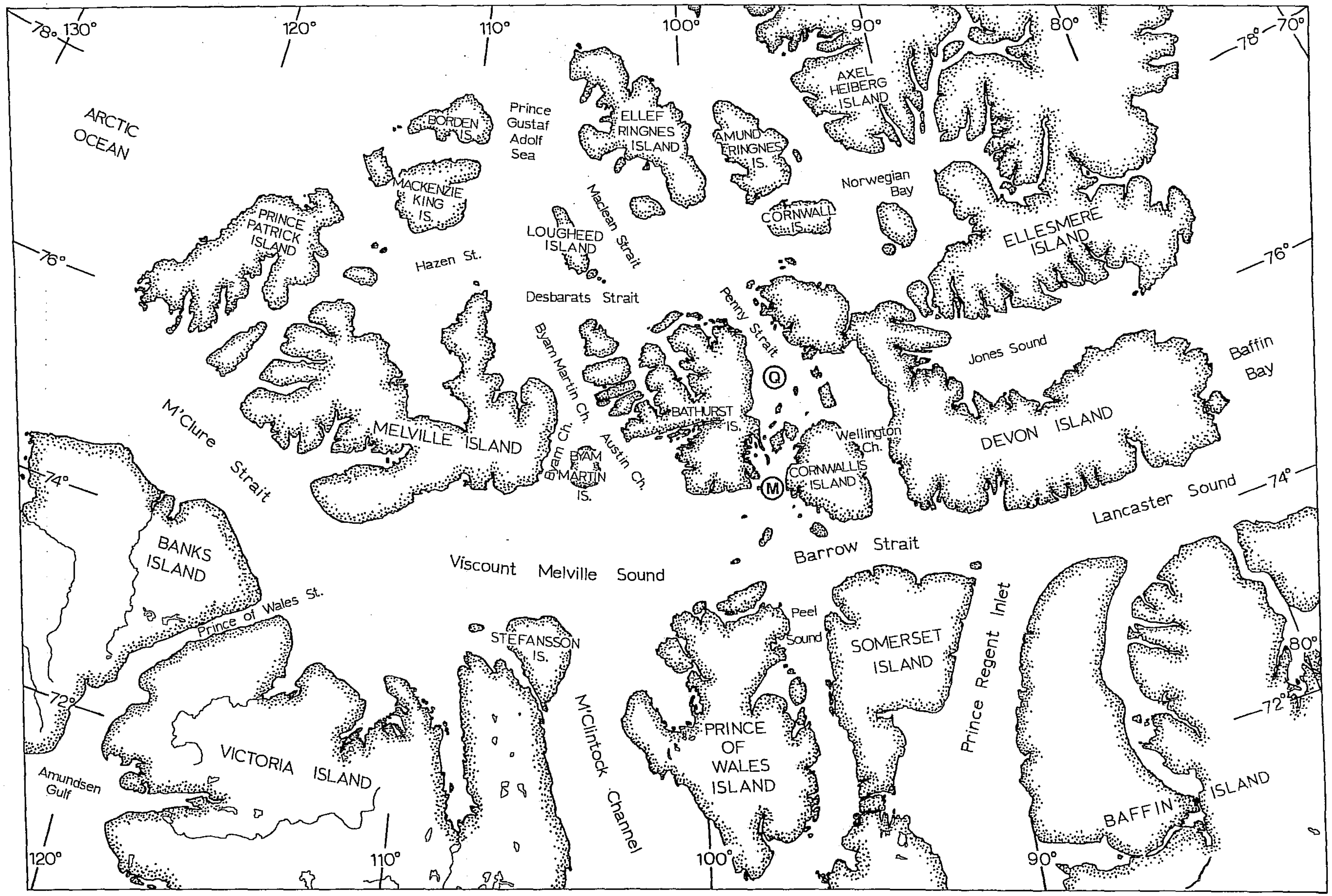

Figure 1.1 Map of the Canadian Arctic Archipelago. Here Q denotes Queens Channel and M denotes McDougall Sound.

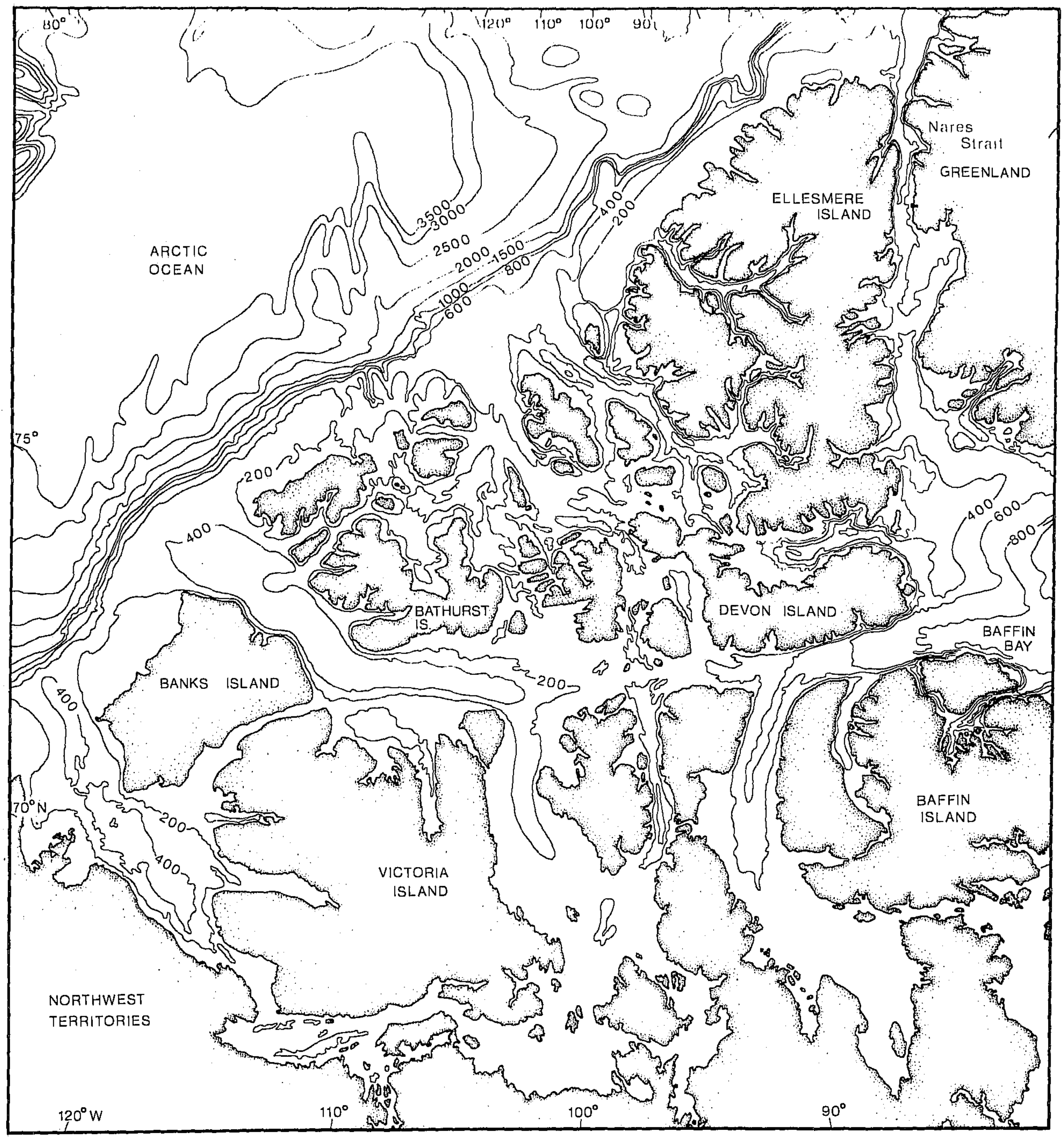

Figure 1.2 The bathymetry of the Canadian Arctic Archipelago. Depth contours are at 200 m intervals up to 1000 m. At greater depths, 500 m intervals are used.

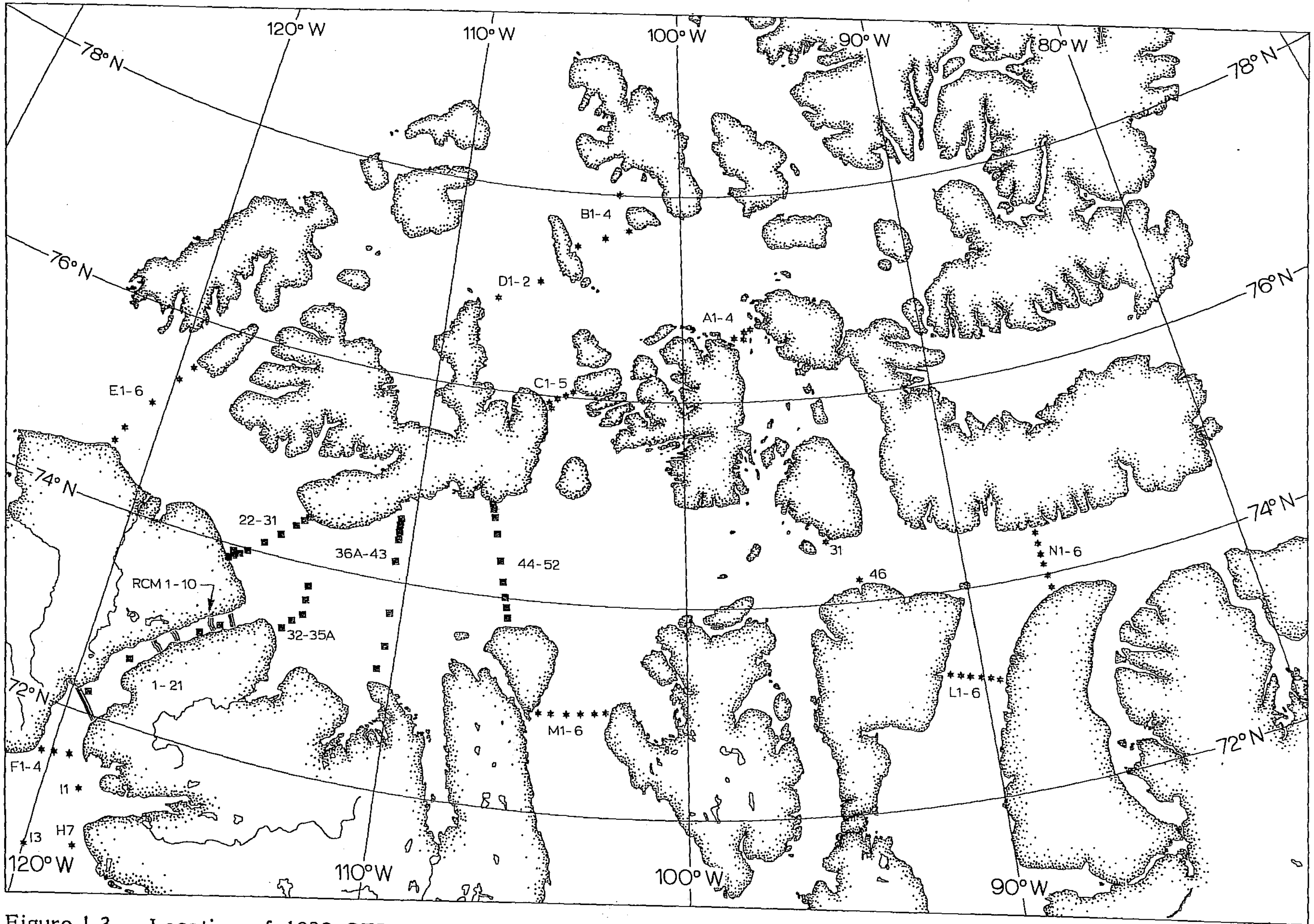

Figure 1.3 Location of 1982 CTD stations. Square symbols and triple lines indicate stations and transects occupied by IOS. Stars indicate stations occupied by Arcic Sciences Ltd. under contract.

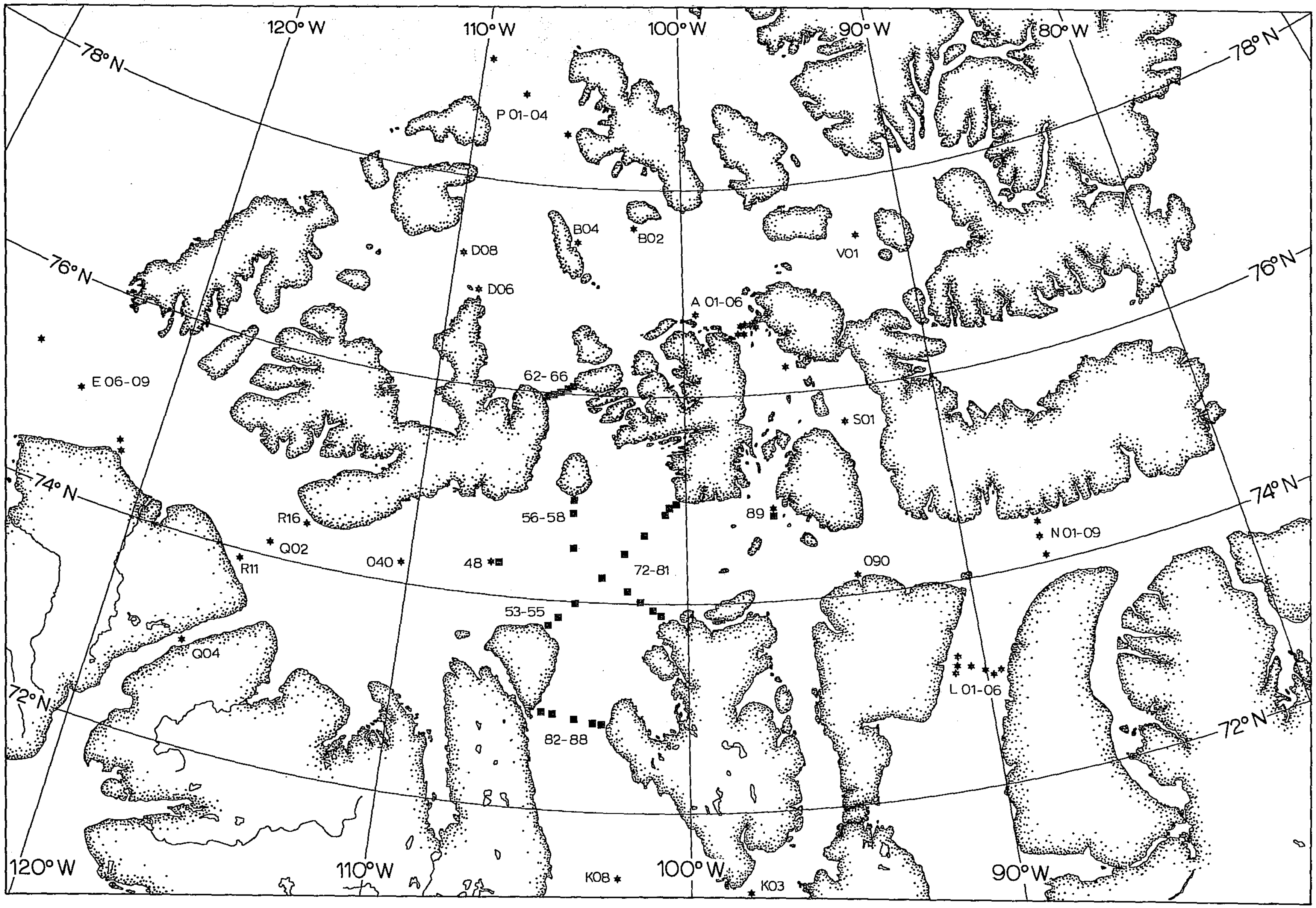

Figure 1.4 Location of 1983 CTD stations. Square symbols indicate IOS stations while stars indicate stations occupied by Arctic Sciences Ltd.

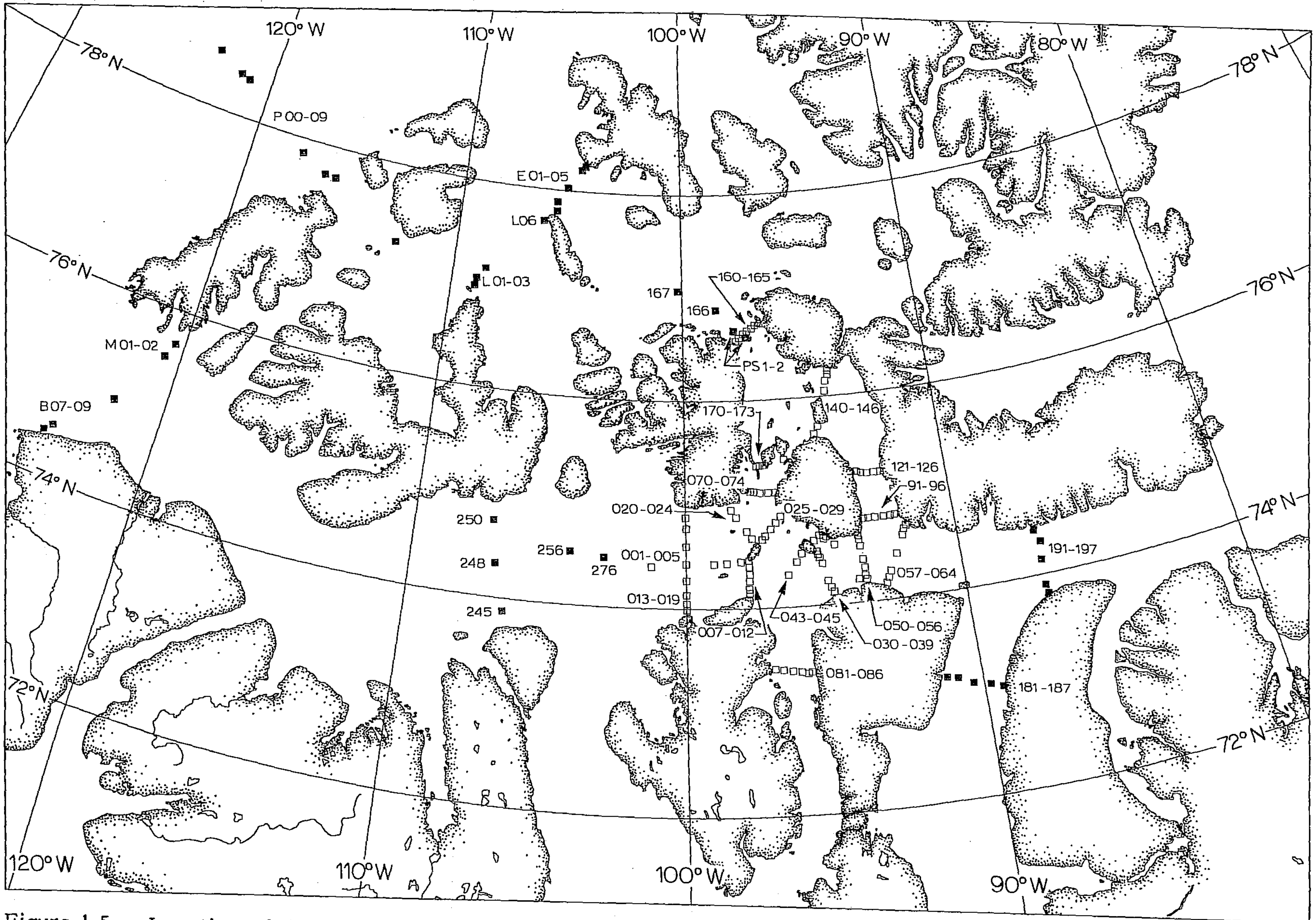

Figure 1.5 Location of 1984 CTD stations. Open squares denote stations in the main study area and solid squares indicate stations occupied as part of the Arctic Shelf Program.

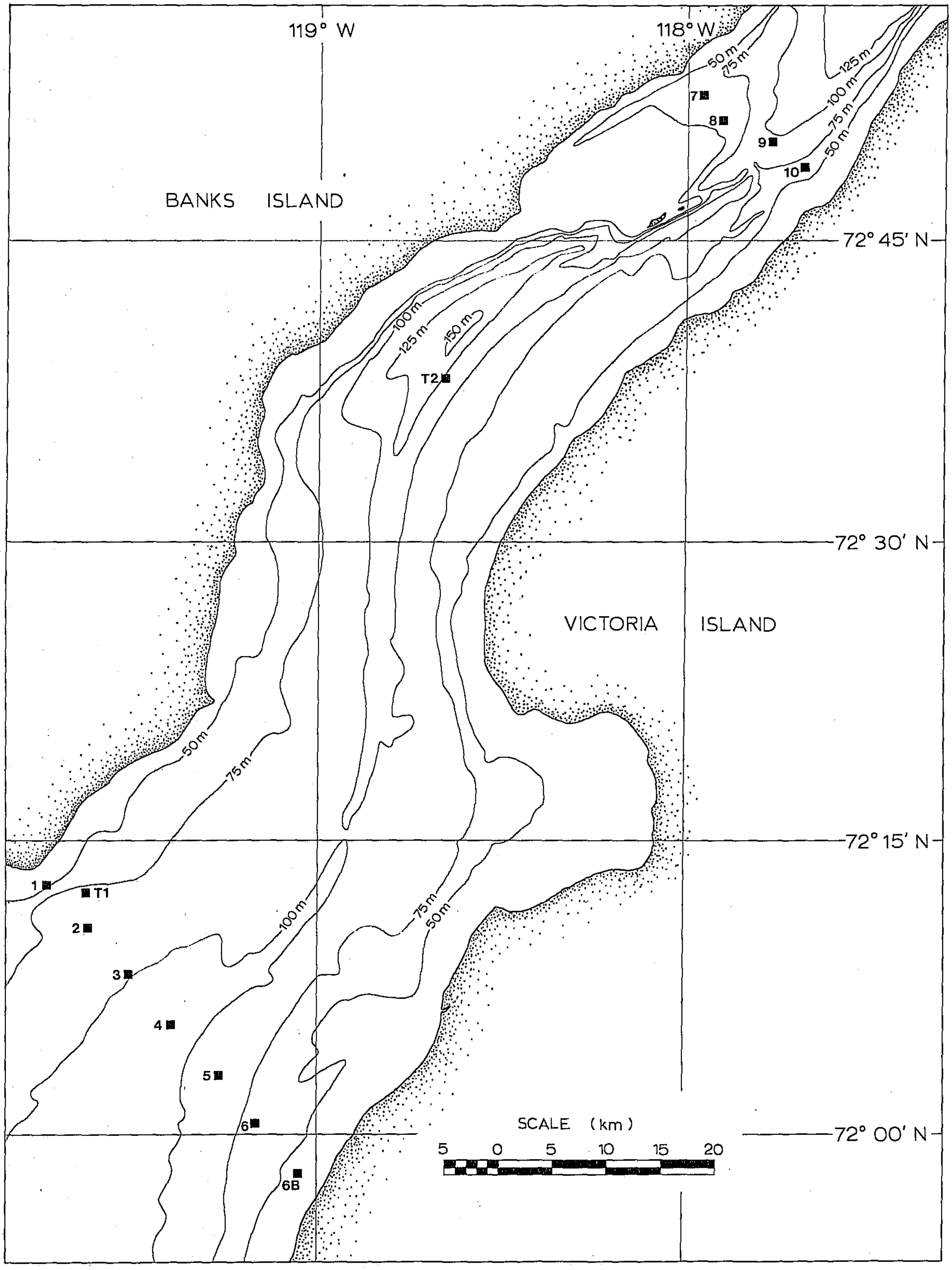

Figure 1.6 Detail of 1982 CTD stations in southwestern portion of Prince of Wales Strait including bathymetry.

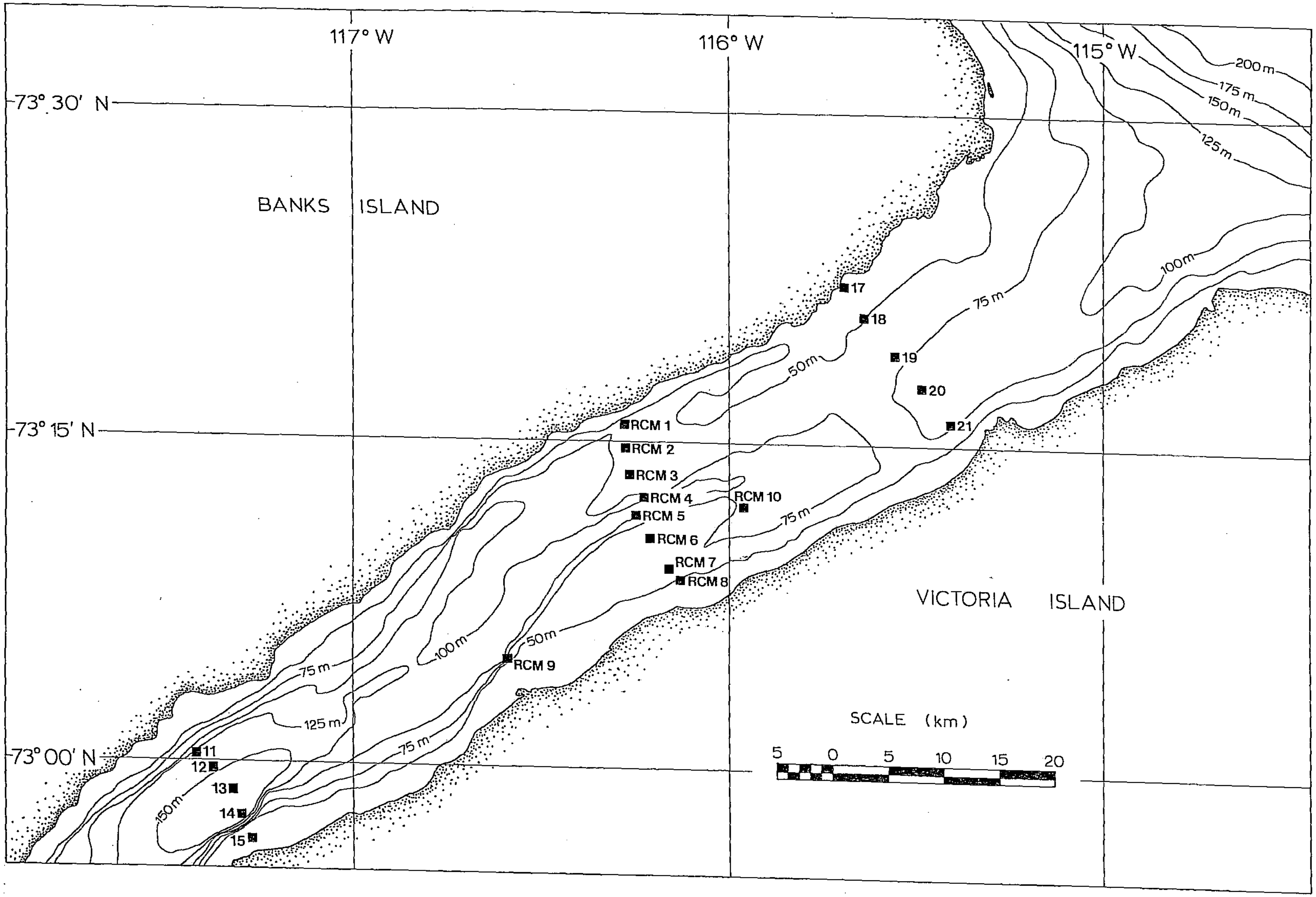

Figure 1.7 Detail of 1982 CTD stations in northeastern portion of Prince of Wales Strait including bathymetry.

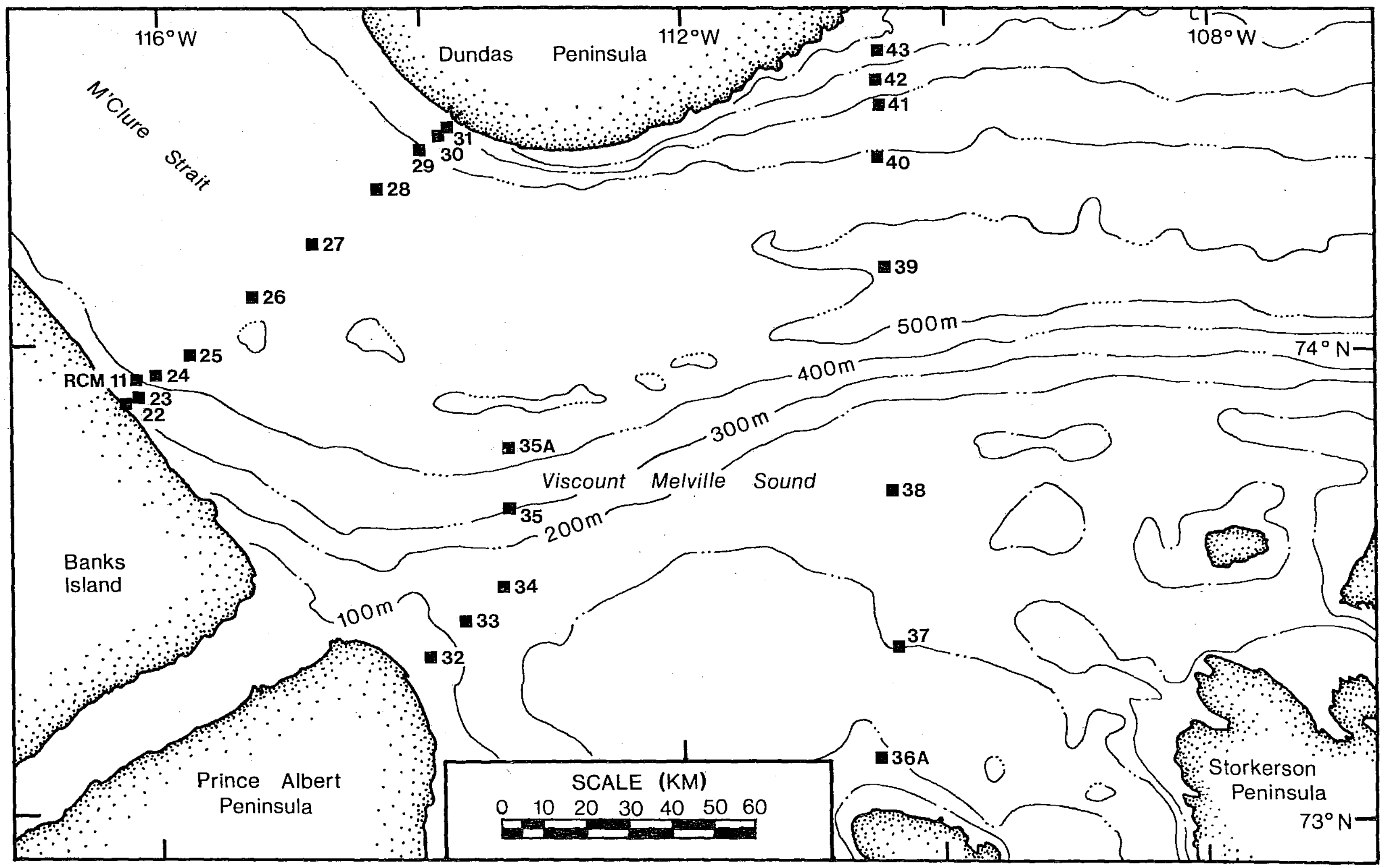

Figure 1.8 Detail of 1982 CTD stations and bathymetry in eastern M'Clure Strait and western Viscount Melville Sound.

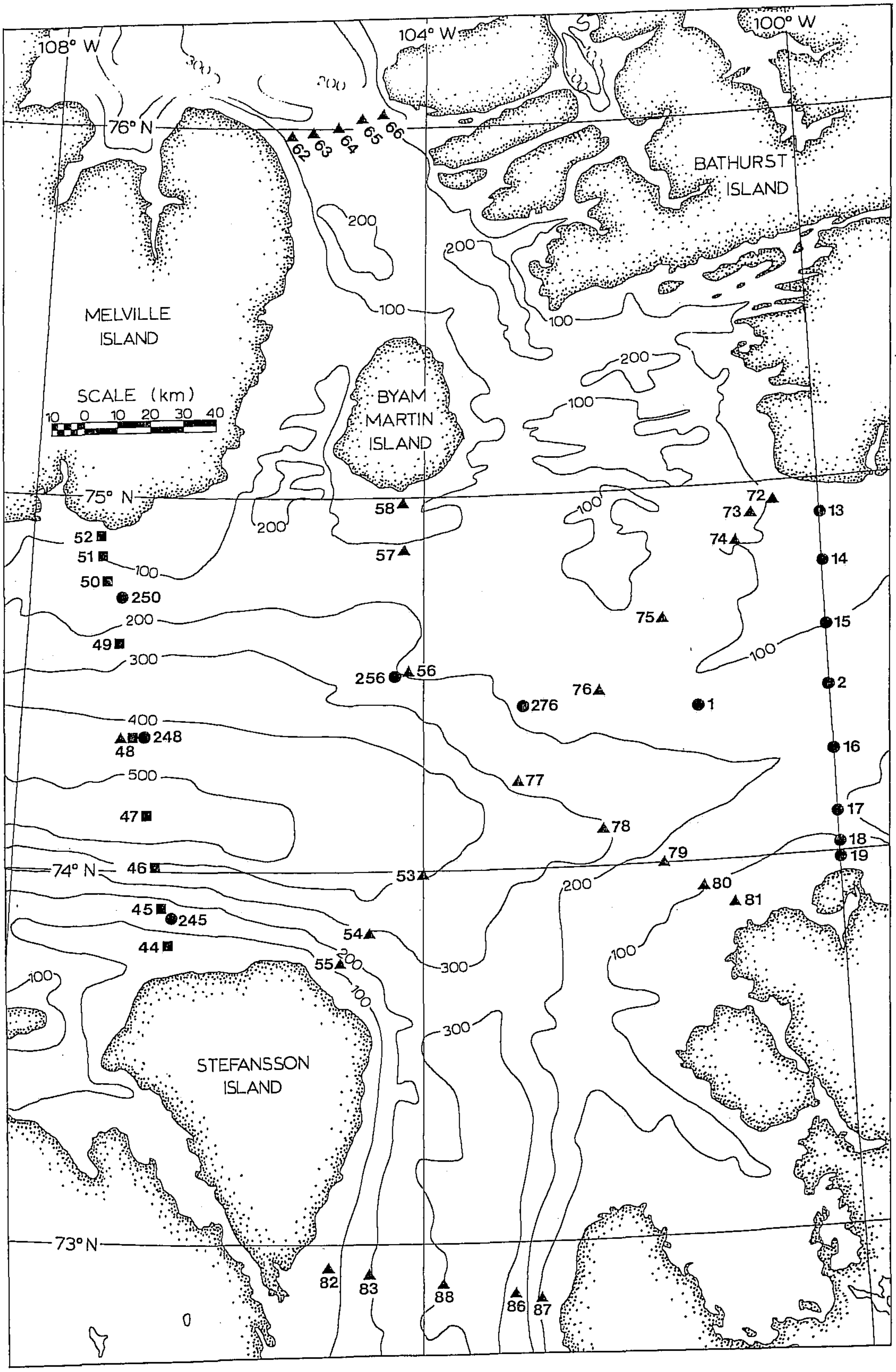

Figure 1.9 Detail of IOS CTD stations and bathymetry in eastern Viscount Melville Sound. Squares are 1982 stations, triangles are 1983 stations and circles are 1984 stations.

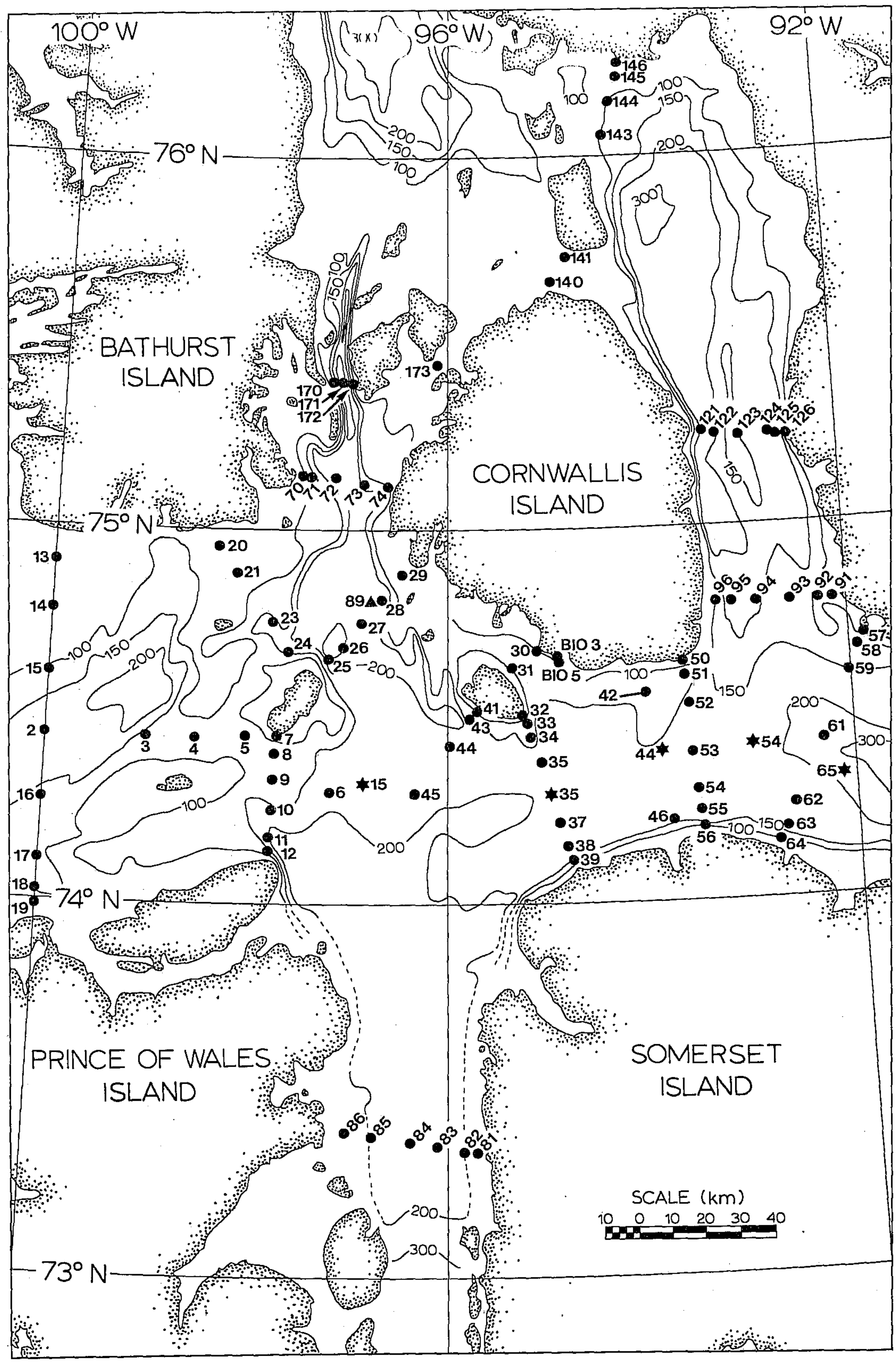

Figure 1.10 Detail of CTD stations and bathymetry in the Barrow Strait area. Circles are 1984 stations, stars are BLMSS stations and the triangle is a 1983 IOS station.

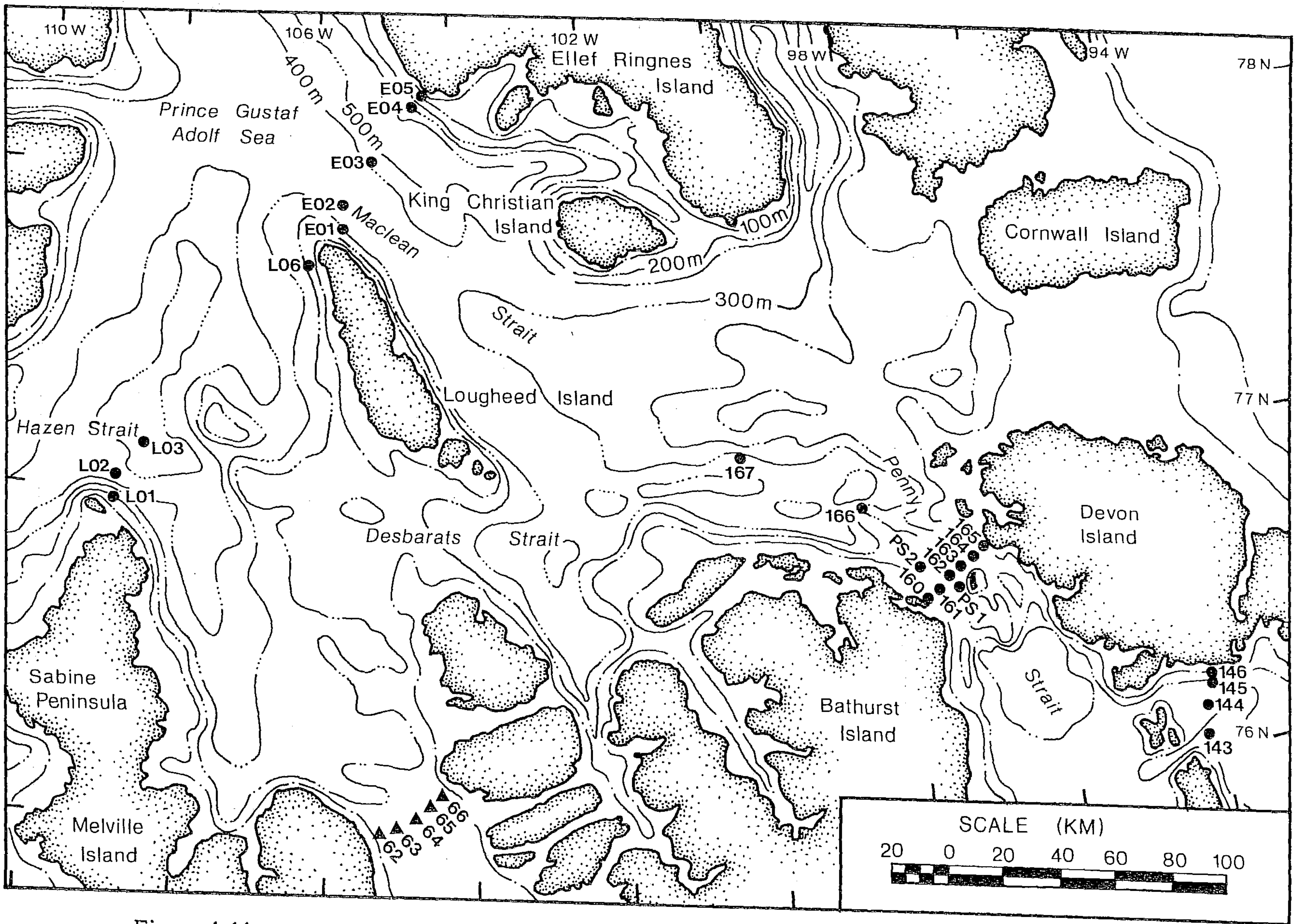

Figure 1.11 Detail of 1984 CTD stations and bathymetry in the Sverdrup Basin region. Triangles are 1983 stations and circles are 1984 stations.

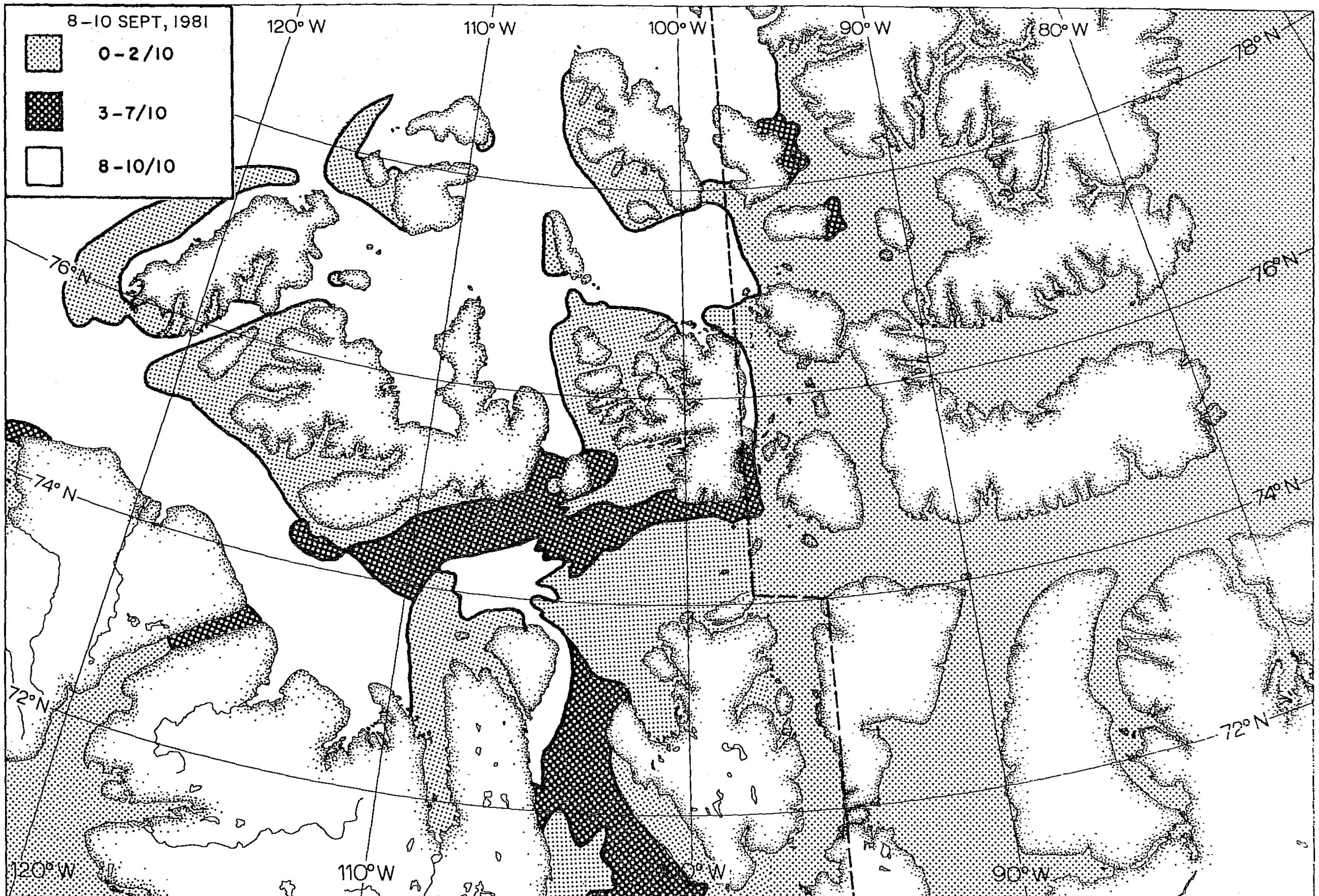

Figure 1.12 Ice cover in late summer 1981. Based on summary charts from Ice Central, Atmospheric Environment Service. The dashed line is boundary between charts prepared two days apart.

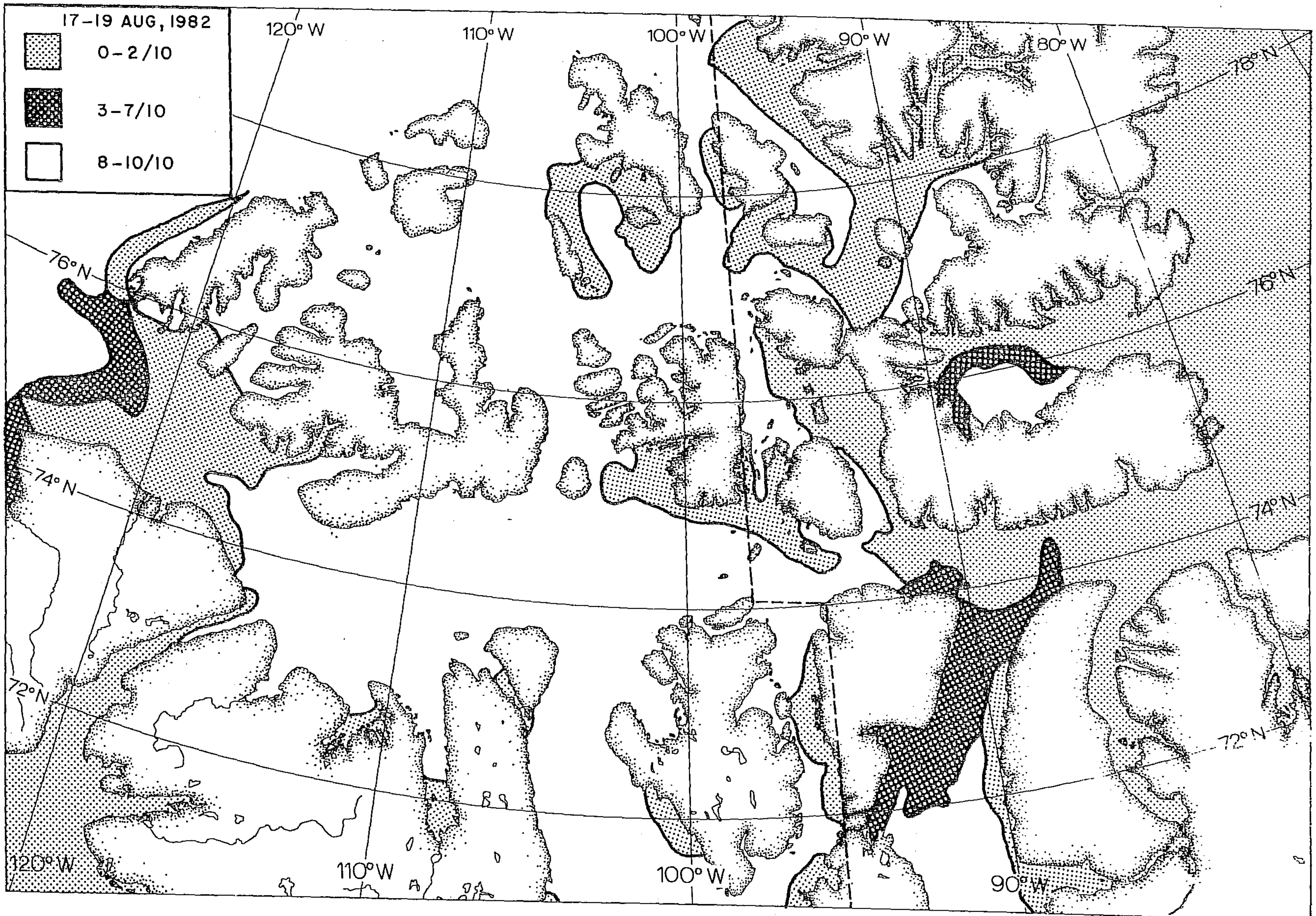

Figure 1.13 Ice cover in late summer 1982. Based on summary charts from Ice Central, Atmospheric Environment Service. The dashed line is boundary between charts prepared two days apart.

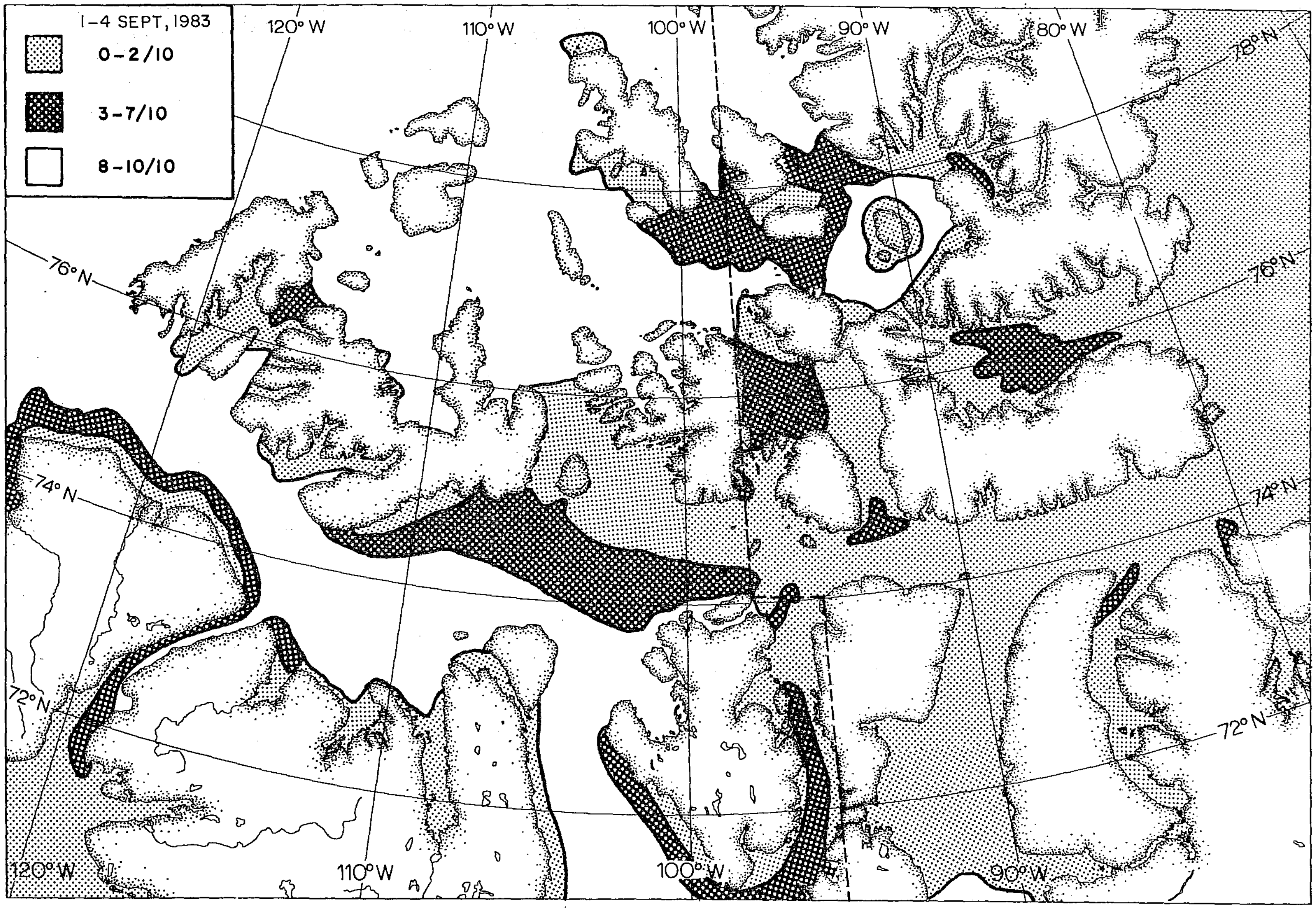

Figure 1.14 Ice cover in late summer 1983. Based on summary charts from Ice Central, Atmospheric Environment Service. The dashed line is boundary between charts prepared three days apart.

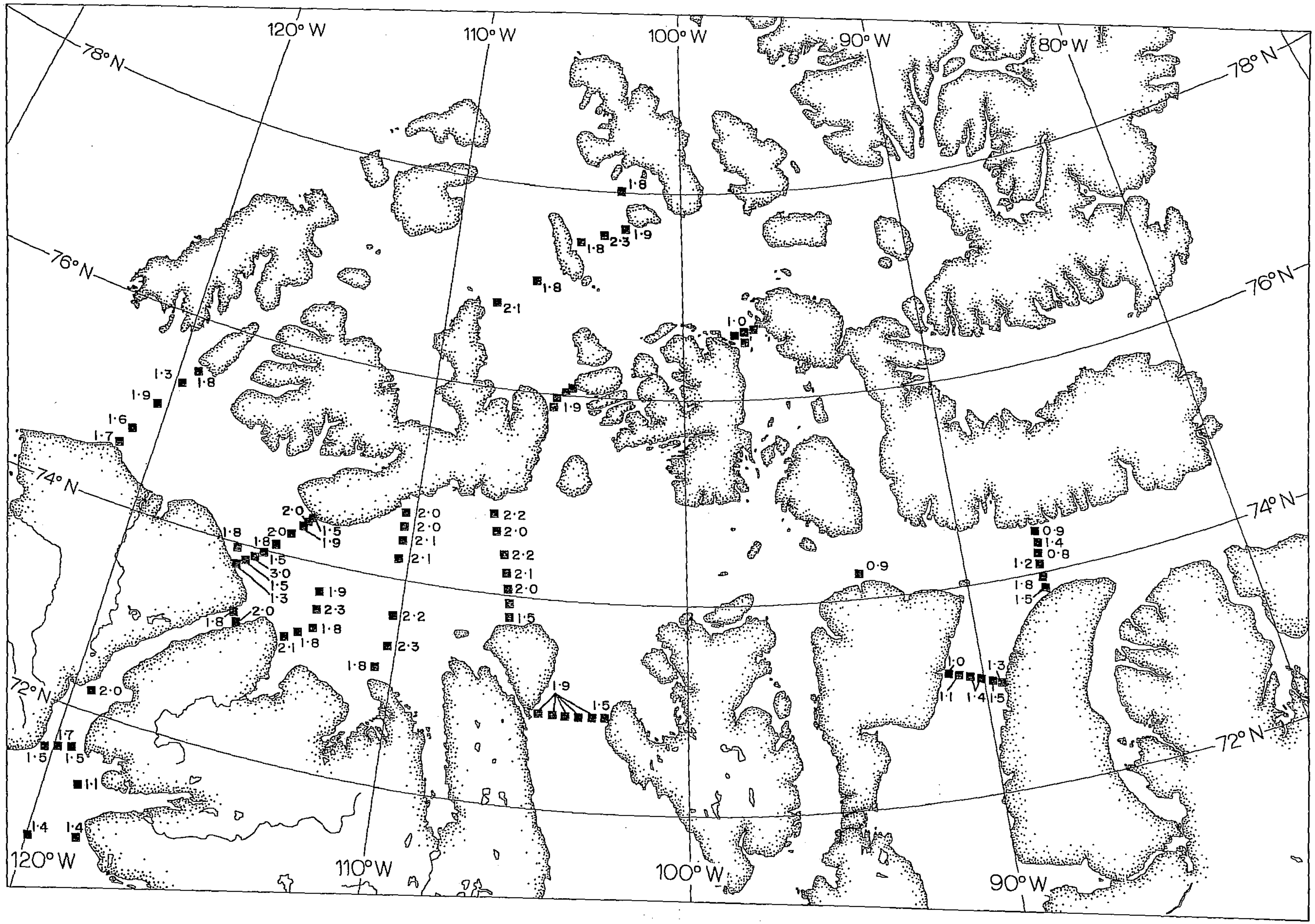

Figure 1.15 Measured ice thickness (m) during the spring 1982 field programs.

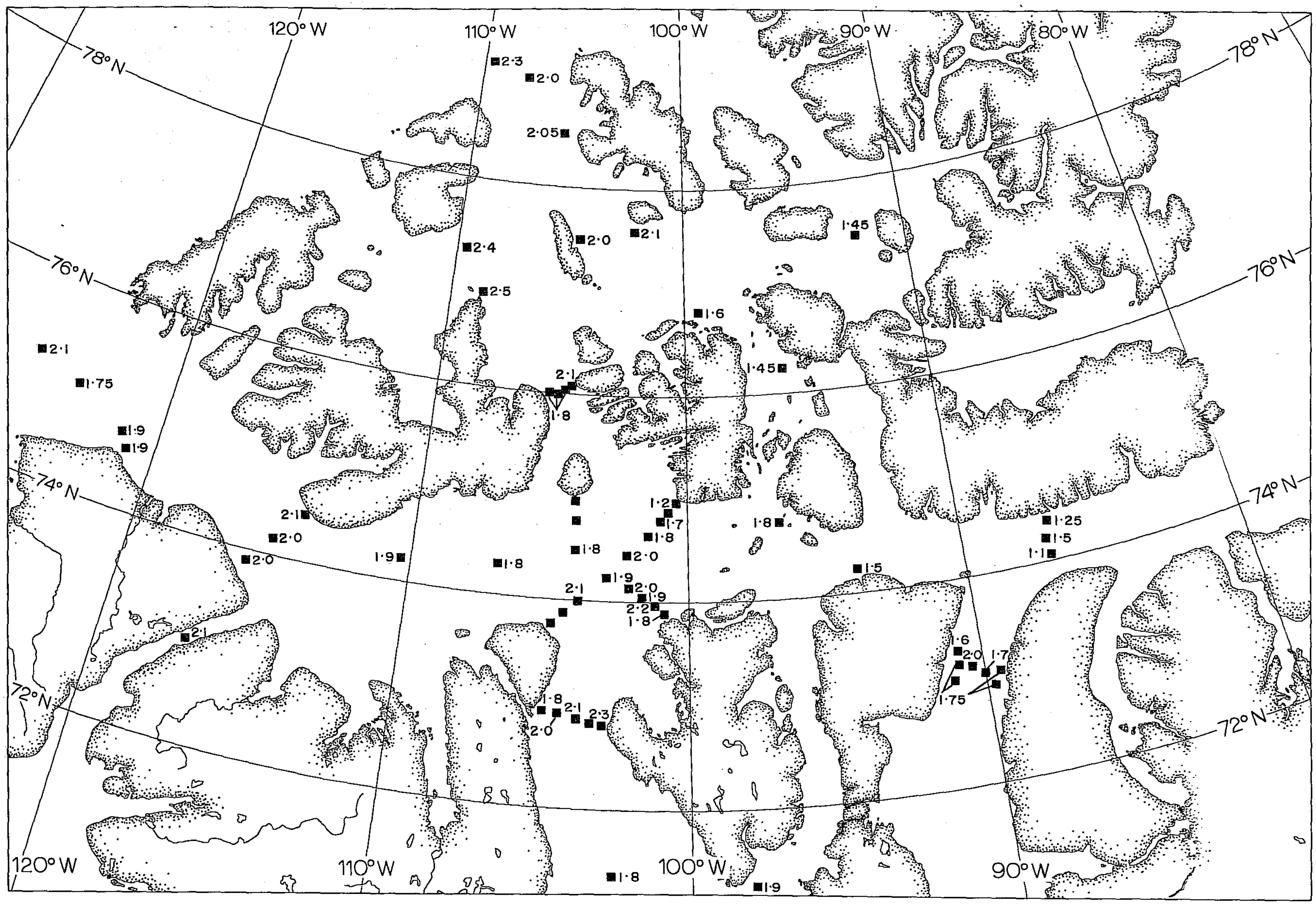

Figure 1.16 Measured ice thickness (m) during the spring 1983 field programs.

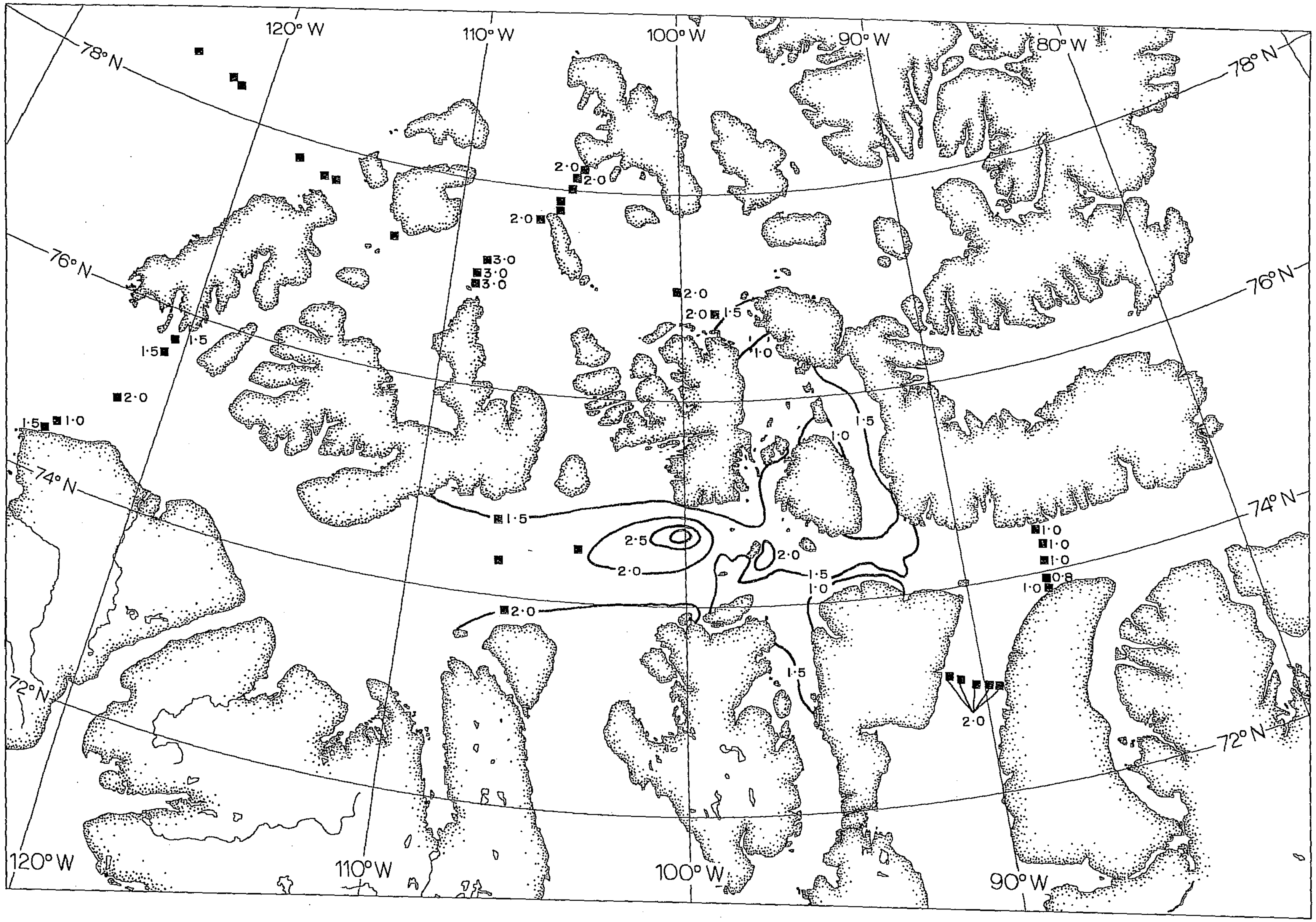

Figure 1.17 Measured ice thickness (m) during the spring 1984 field programs.

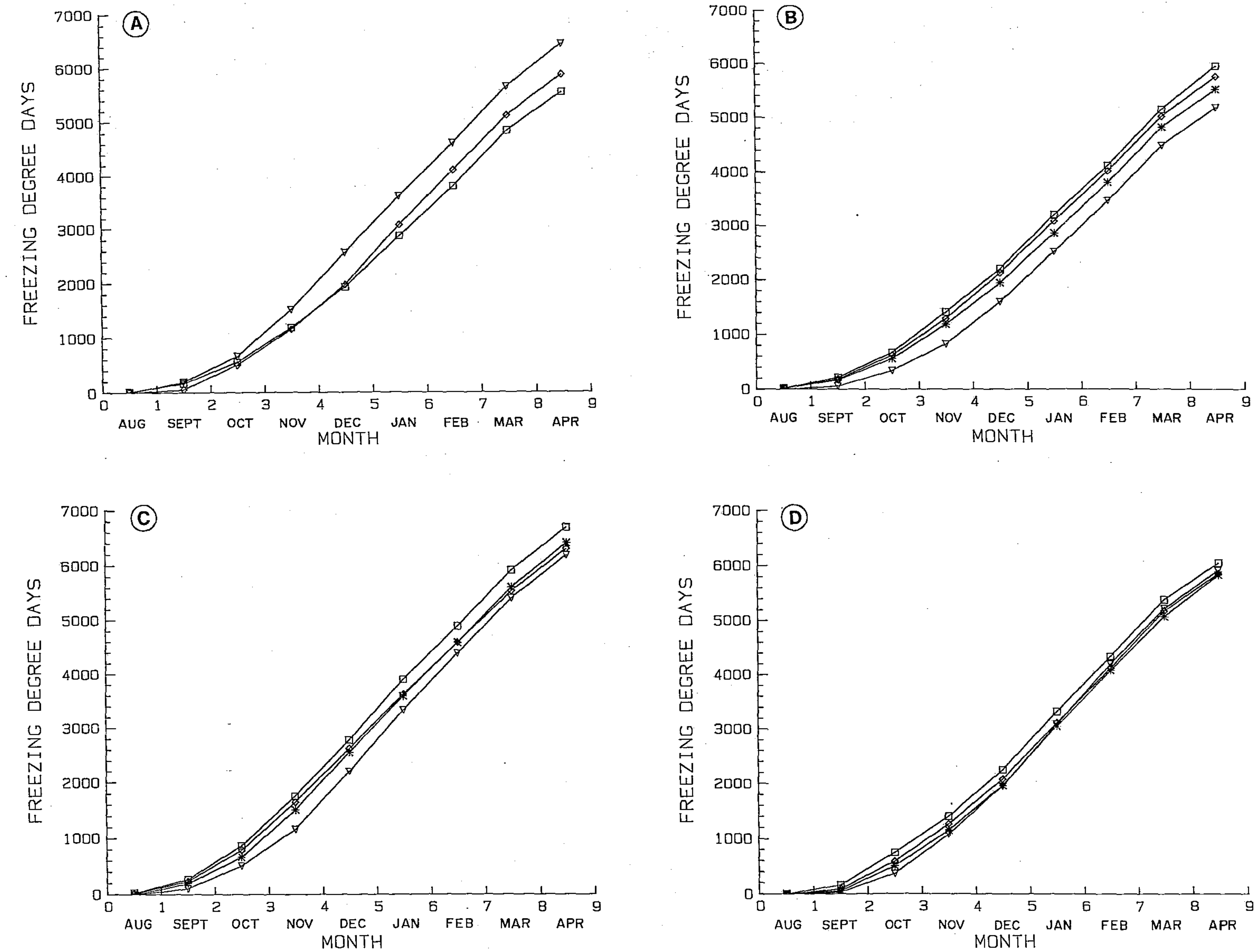

Figure 1.18 Plots of cumulative freezing degree days since August: (A) for Resolute Bay; squares denote 1981 - 82 winter, triangles denote 1982 - 83 winter and diamonds denote 1983 - 84 winter, (B) for 1981 - 82 winter; squares denote Mould Bay, triangles Pond Inlet, diamonds Rae Point and asterisks Resolute Bay, (C) for 1982 - 83 winter; symbols as for (B), (D) for 1983 - 84 winter; symbols as for (B).

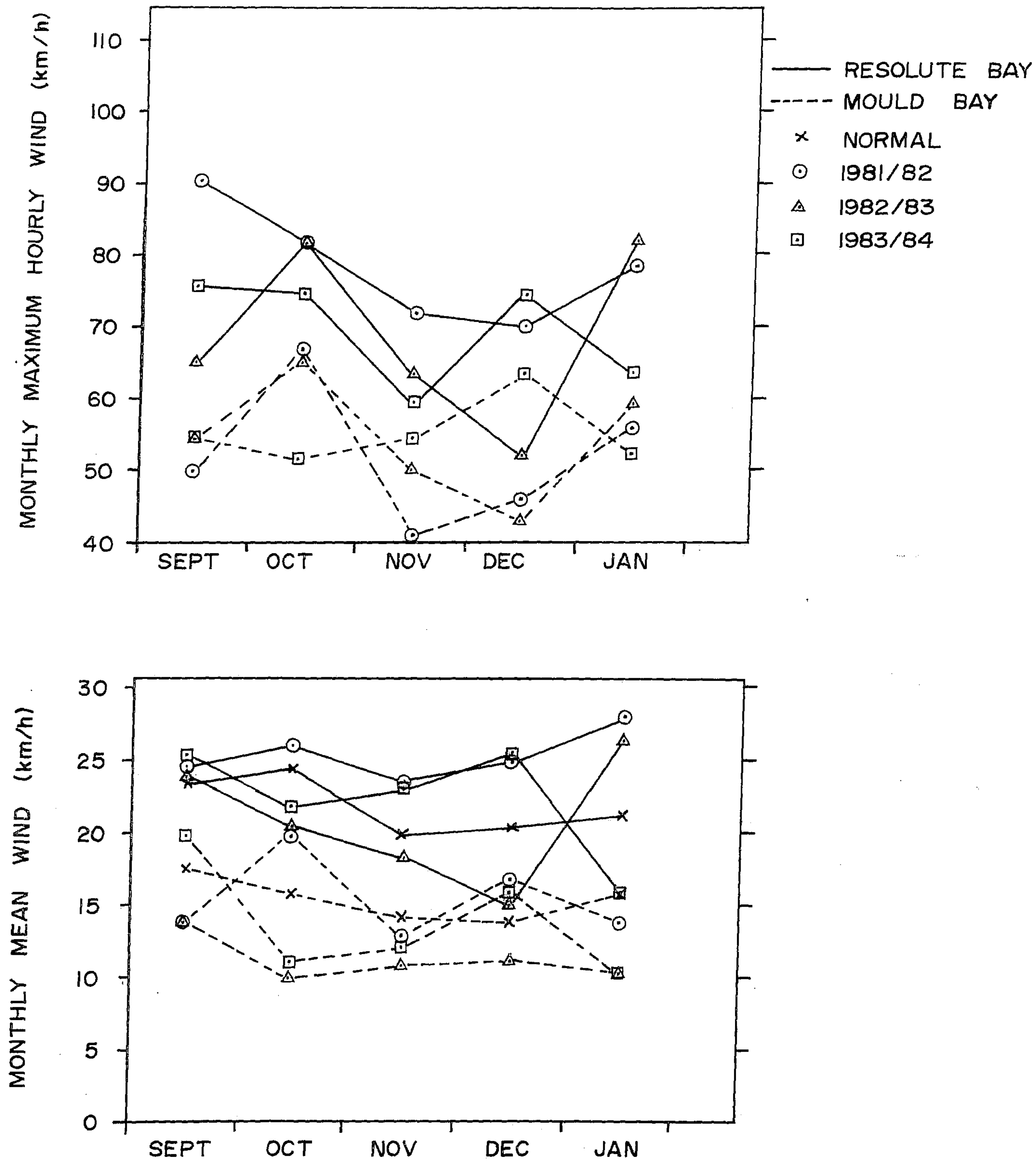

Figure 1.19 Monthly mean wind and maximum hourly wind for Resolute Bay and Mould Bay, September to January for the winters 1981/82 to 1983/84. 25 km/h equals about 7 m/s.

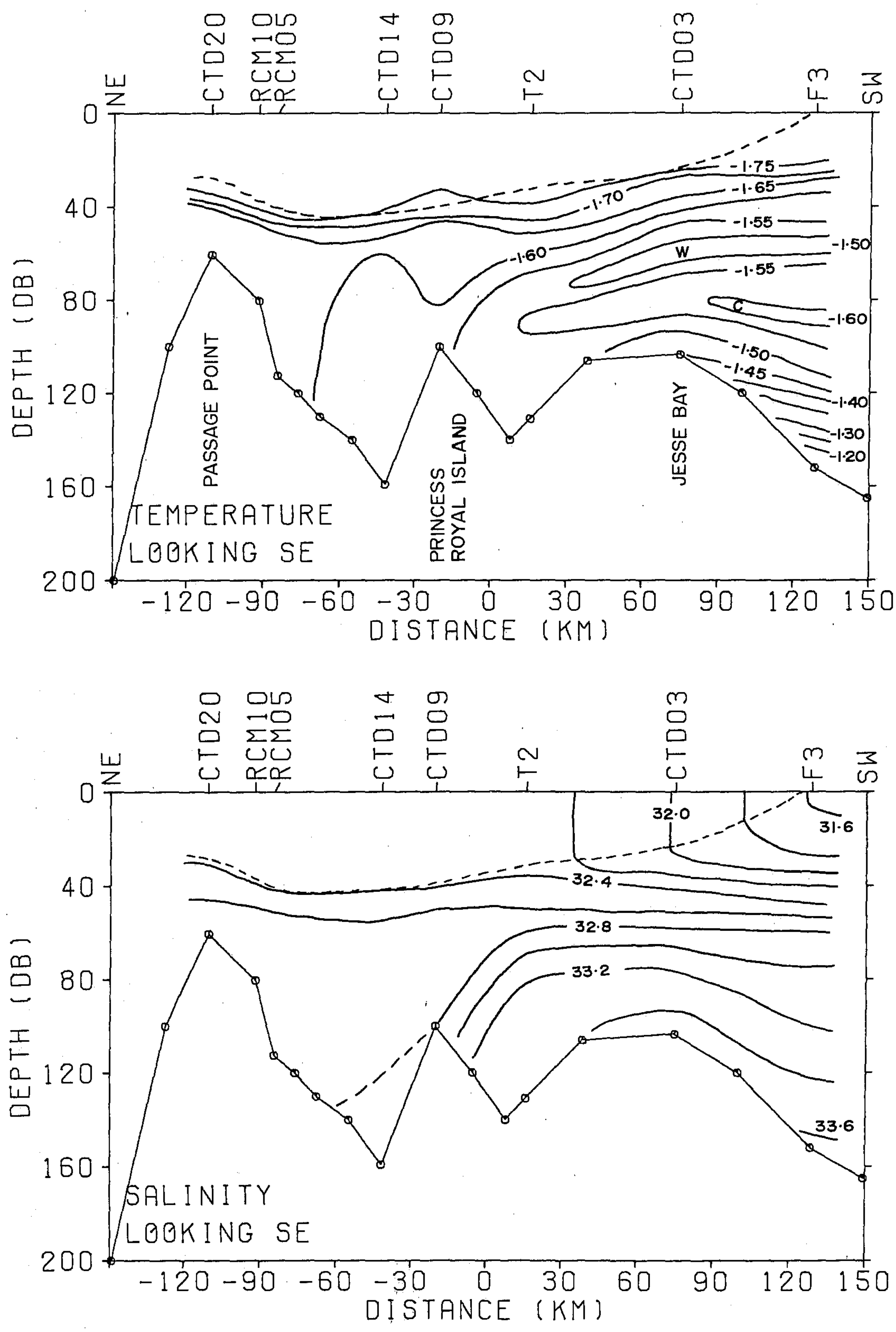

Figure 2.1 Longitudinal temperature and salinity section along Prince of Wales Strait in 1982. Note that station F3 salinities are low by 0.17 based on comparison with CTD03 measurements.

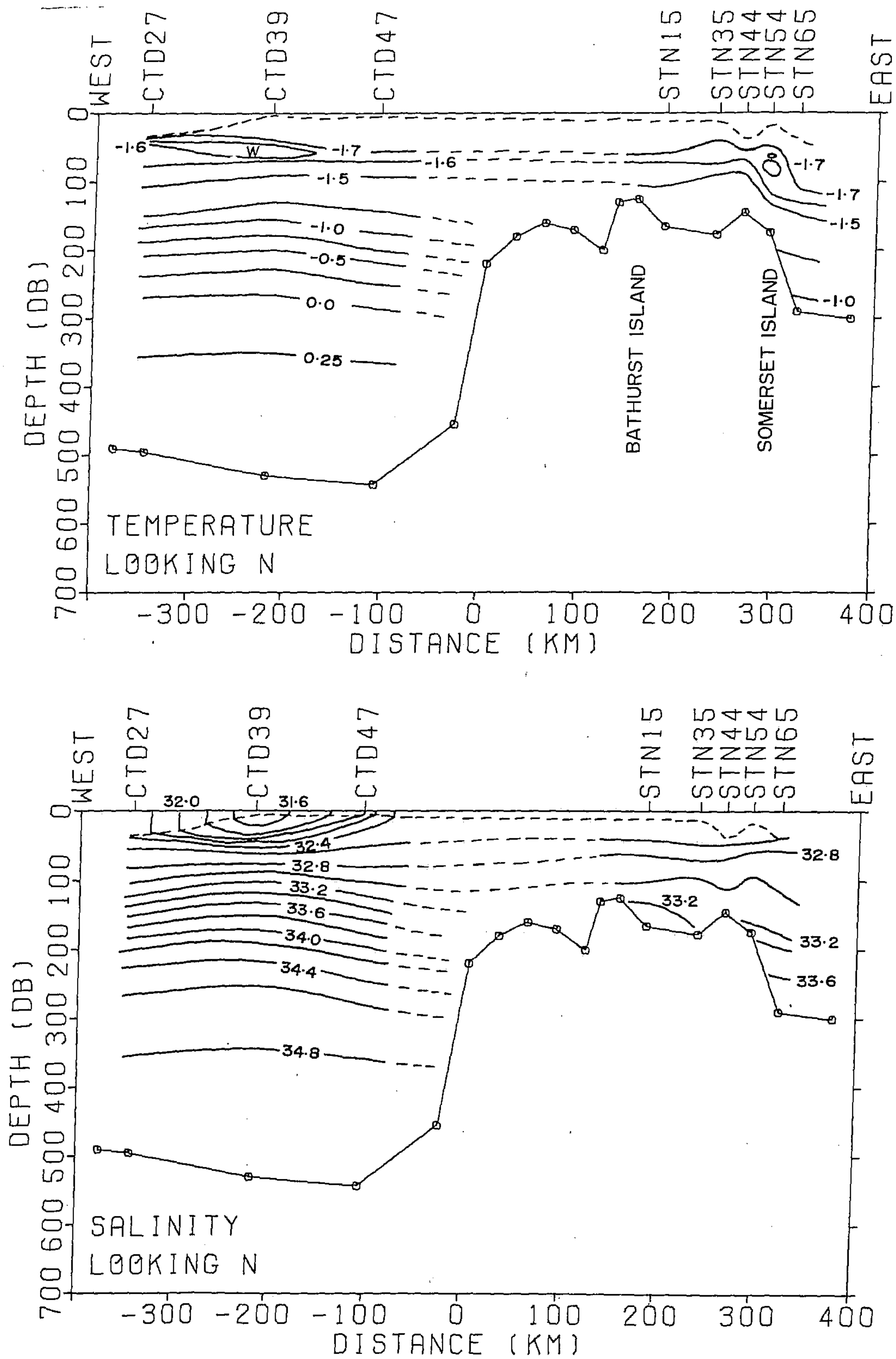

Figure 2.2 Longitudinal temperature and salinity section along western and central Parry Channel in 1982.

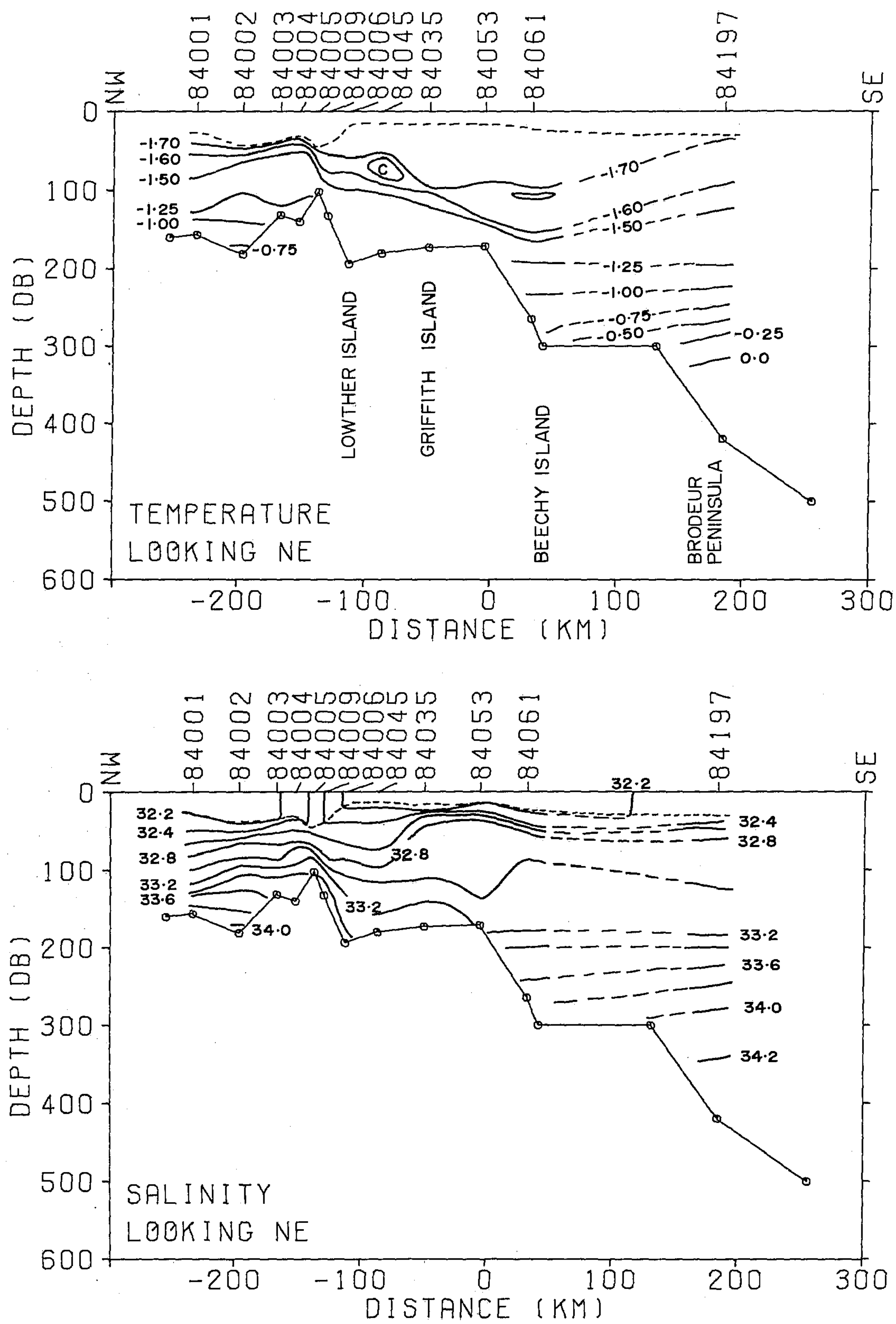

Figure 2.3 Longitudinal temperature and salinity section along central and eastern Parry Channel in 1984.

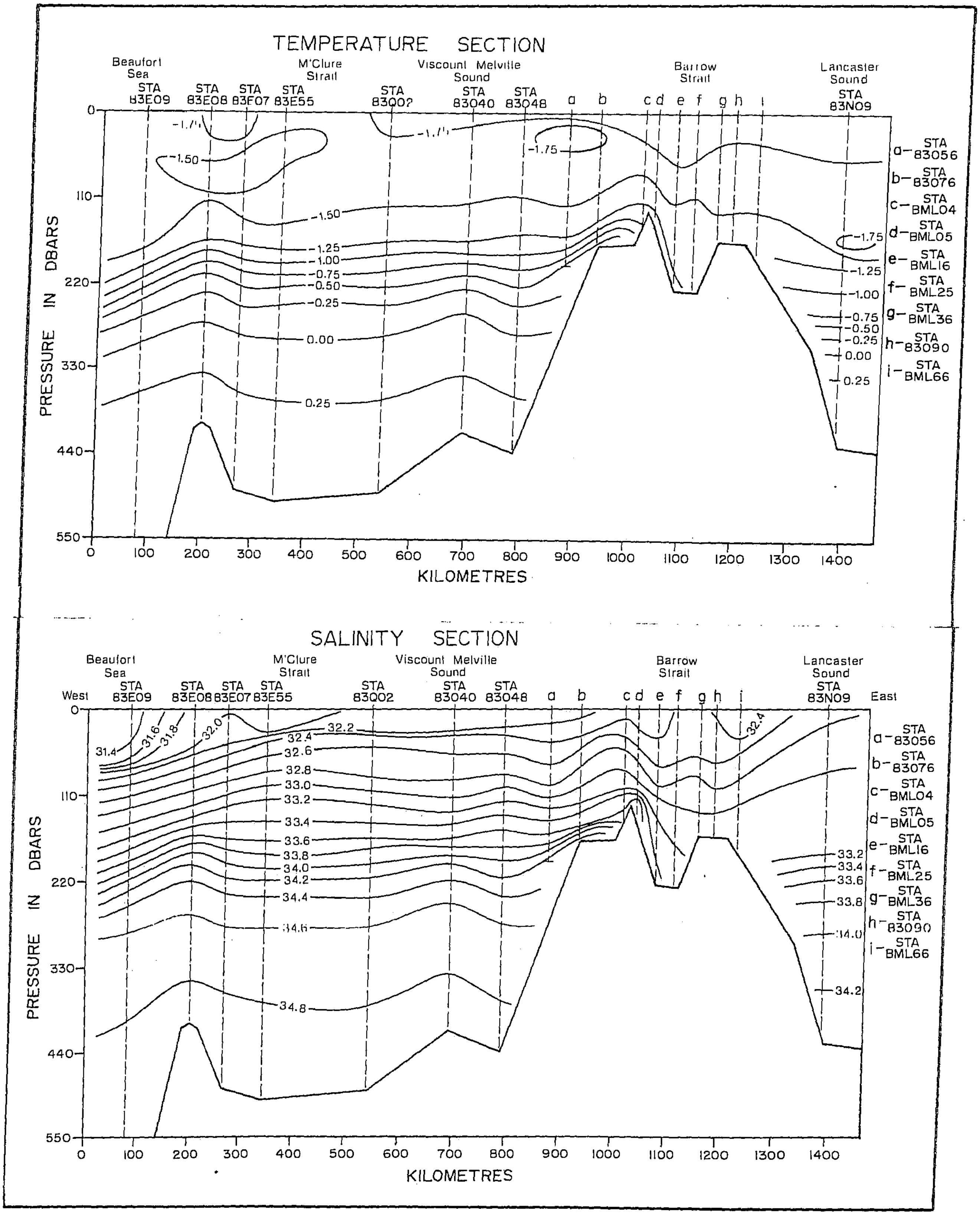

Figure 2.4 Longitudinal temperature and salinity section along Parry Channel in 1983. (from Fissel et al. 1984b.)

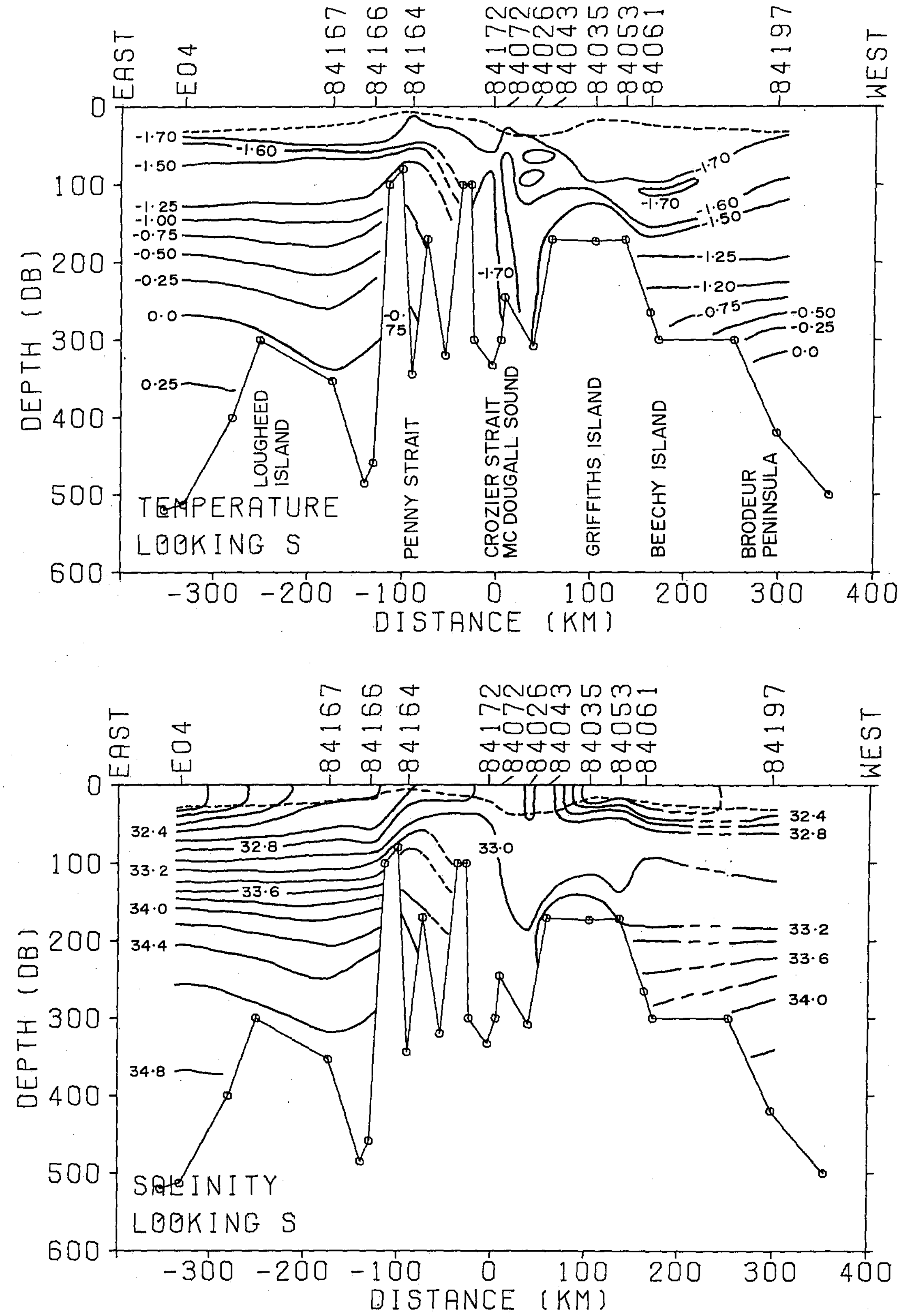

Figure 2.5 Longitudinal temperature and salinity section from Prince Gustaf Adolf Sea through Penny Strait and McDougall Sound in 1984.

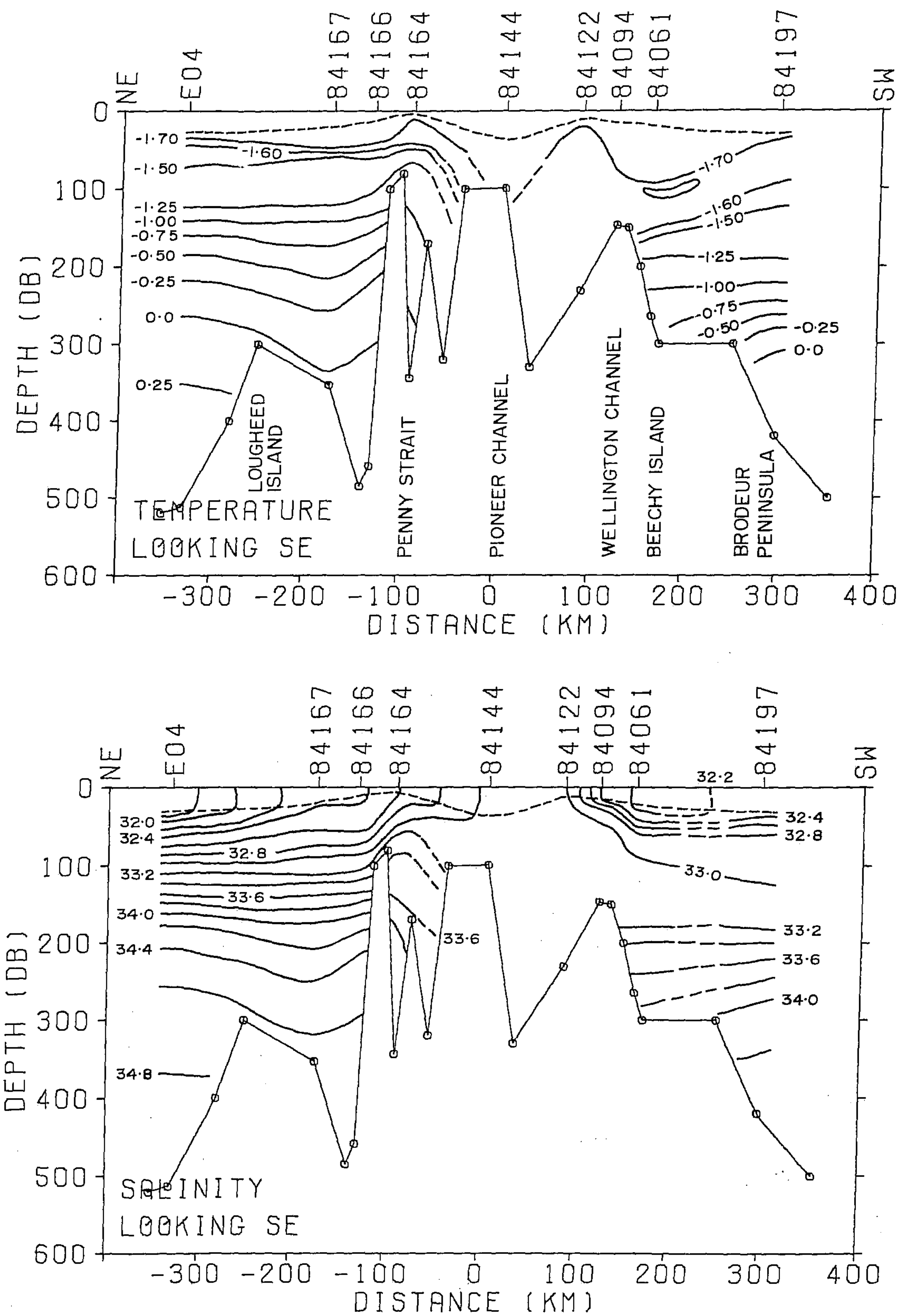

Figure 2.6 Longitudinal temperature and salinity section from Prince Gustaf Adolf Sea through Penny Strait and Wellington Channel in 1984.

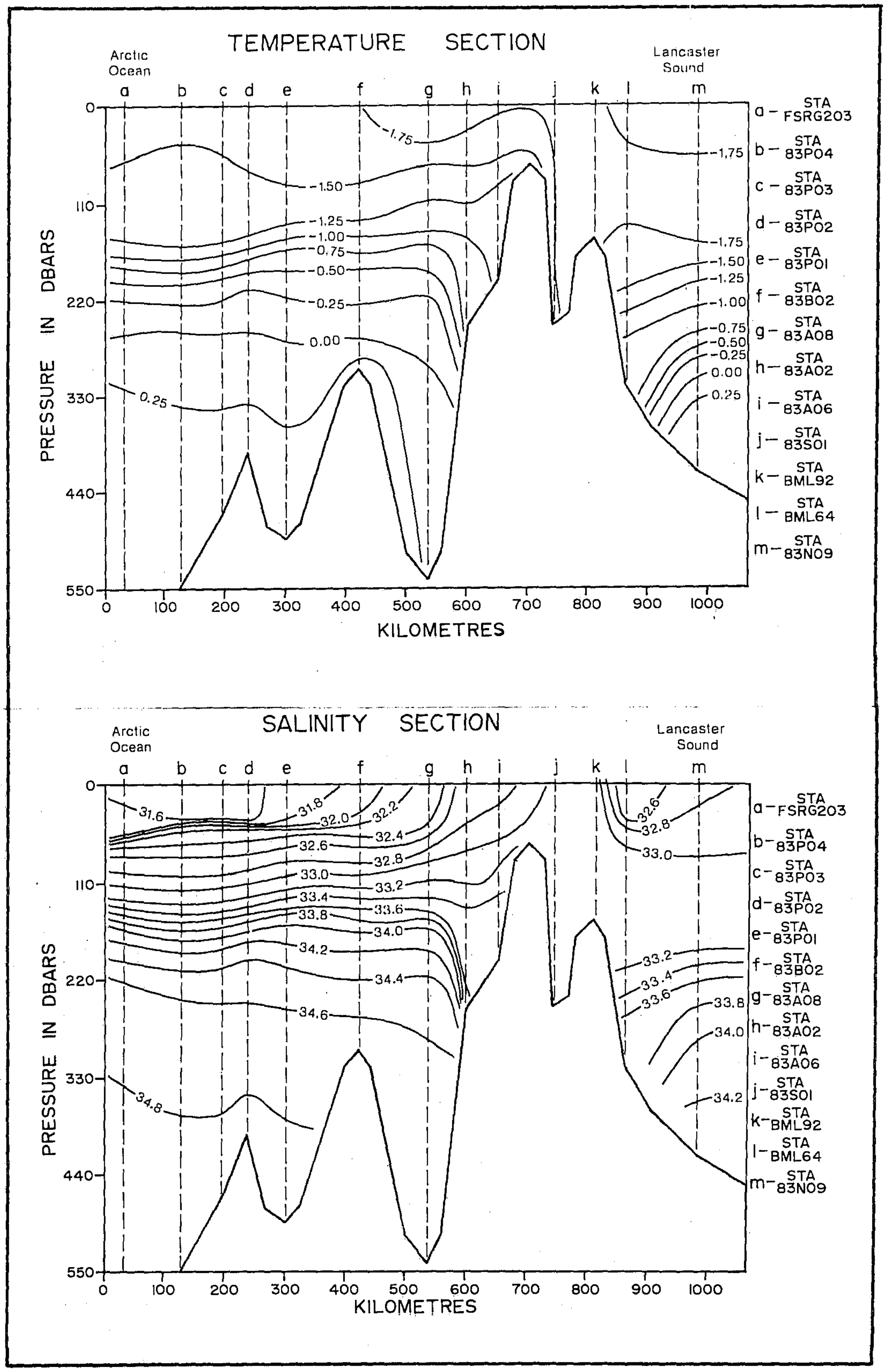

Figure 2.7 Longitudinal temperature and salinity section from Prince Gustaf Adolf Sea through Penny Strait and Wellington Channel in 1983. (from Fissel et al. 1984b.)

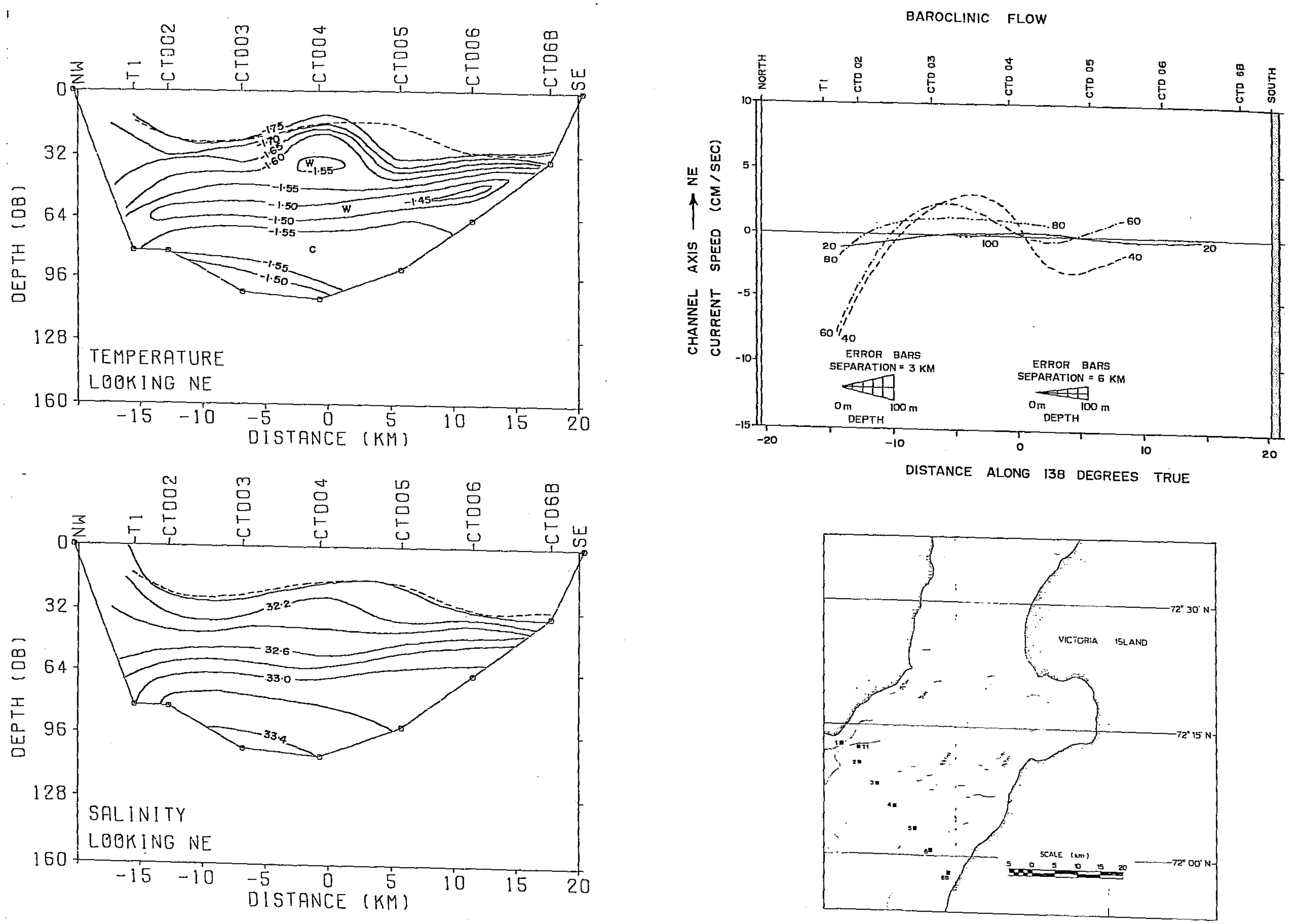

Figure 2.8 Temperature and salinity section and baroclinic velocity across southern Prince of Wales Strait near Jesse Bay, 1982.

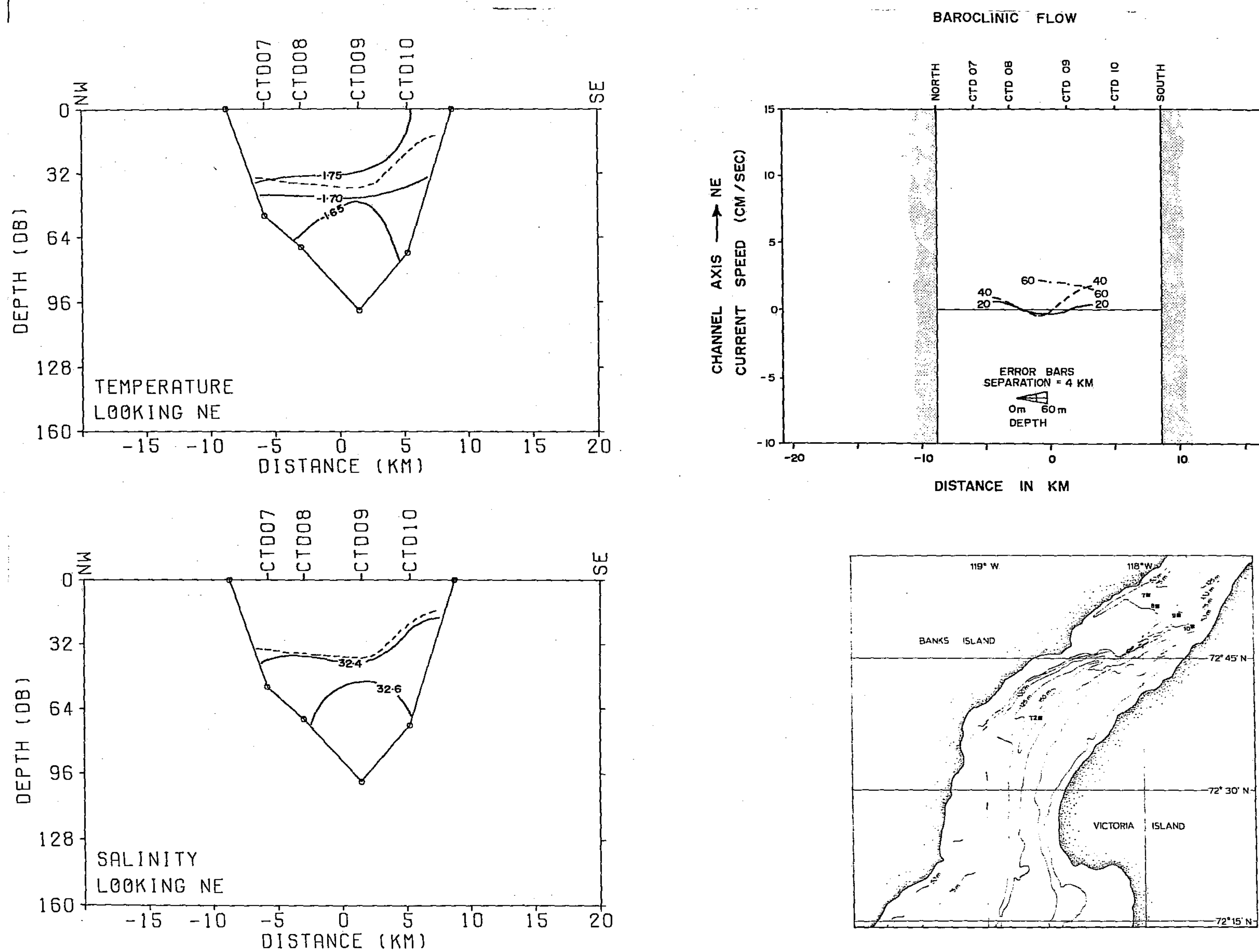

Figure 2.9 Temperature and salinity section and baroclinic velocity across central Prince of Wales Strait near Princess Royal Islands, 1982.

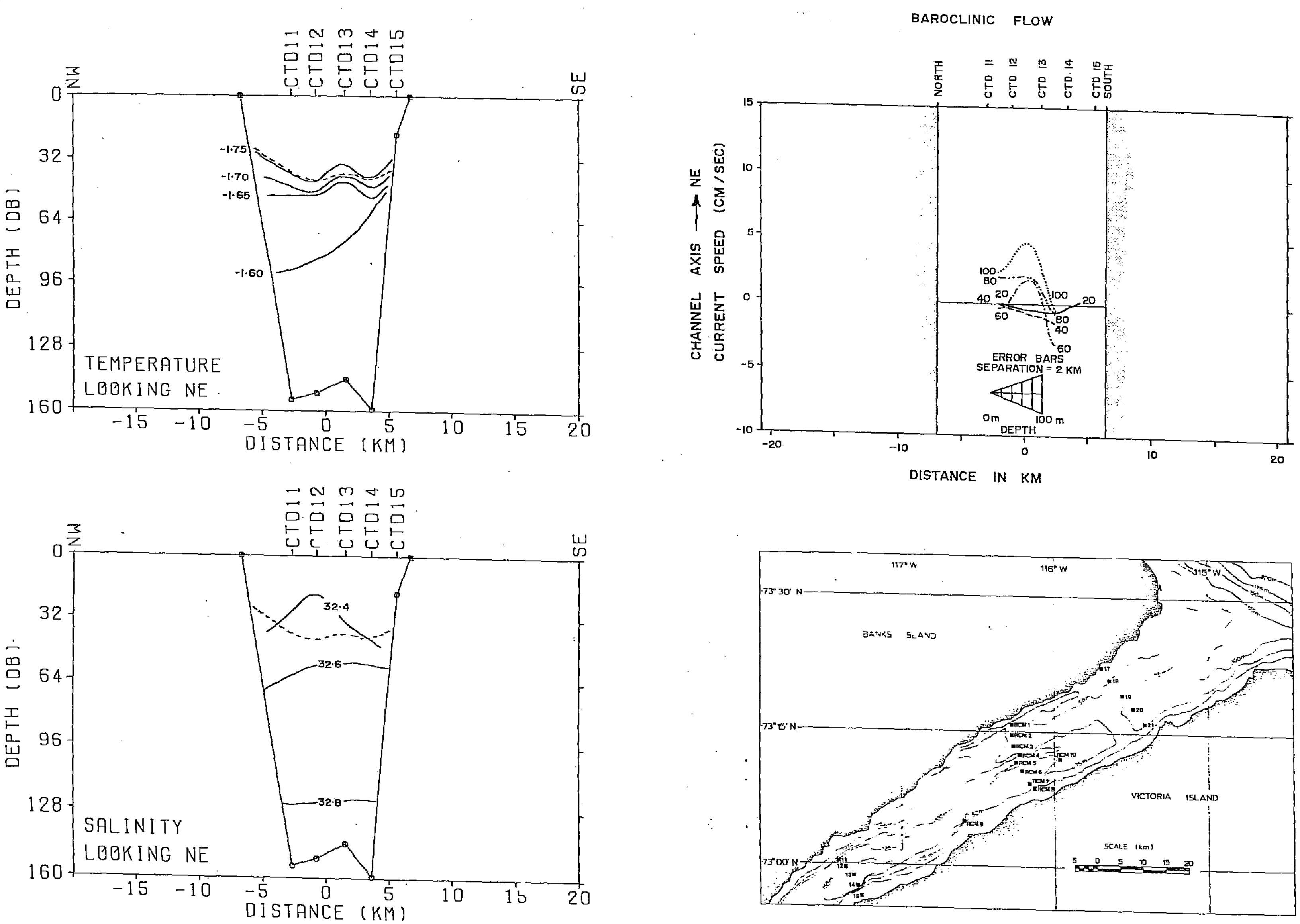

Figure 2.10 Temperature and salinity section and baroclinic velocity across central Prince of Wales Strait near Armstrong Point, 1982.

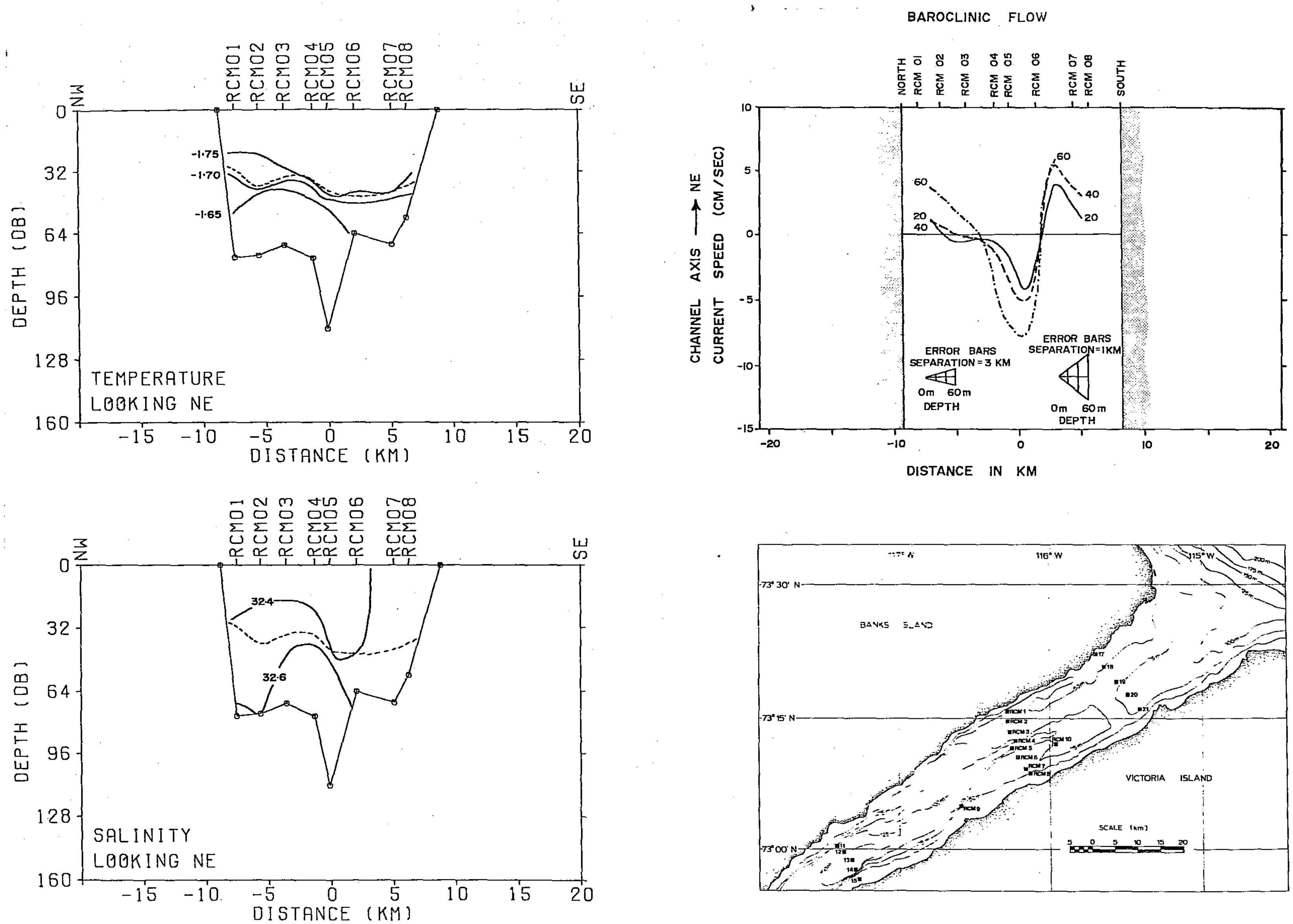

Figure 2.11 Temperature and salinity section and baroclinic velocity across northern Prince of Wales Strait, March 1982.

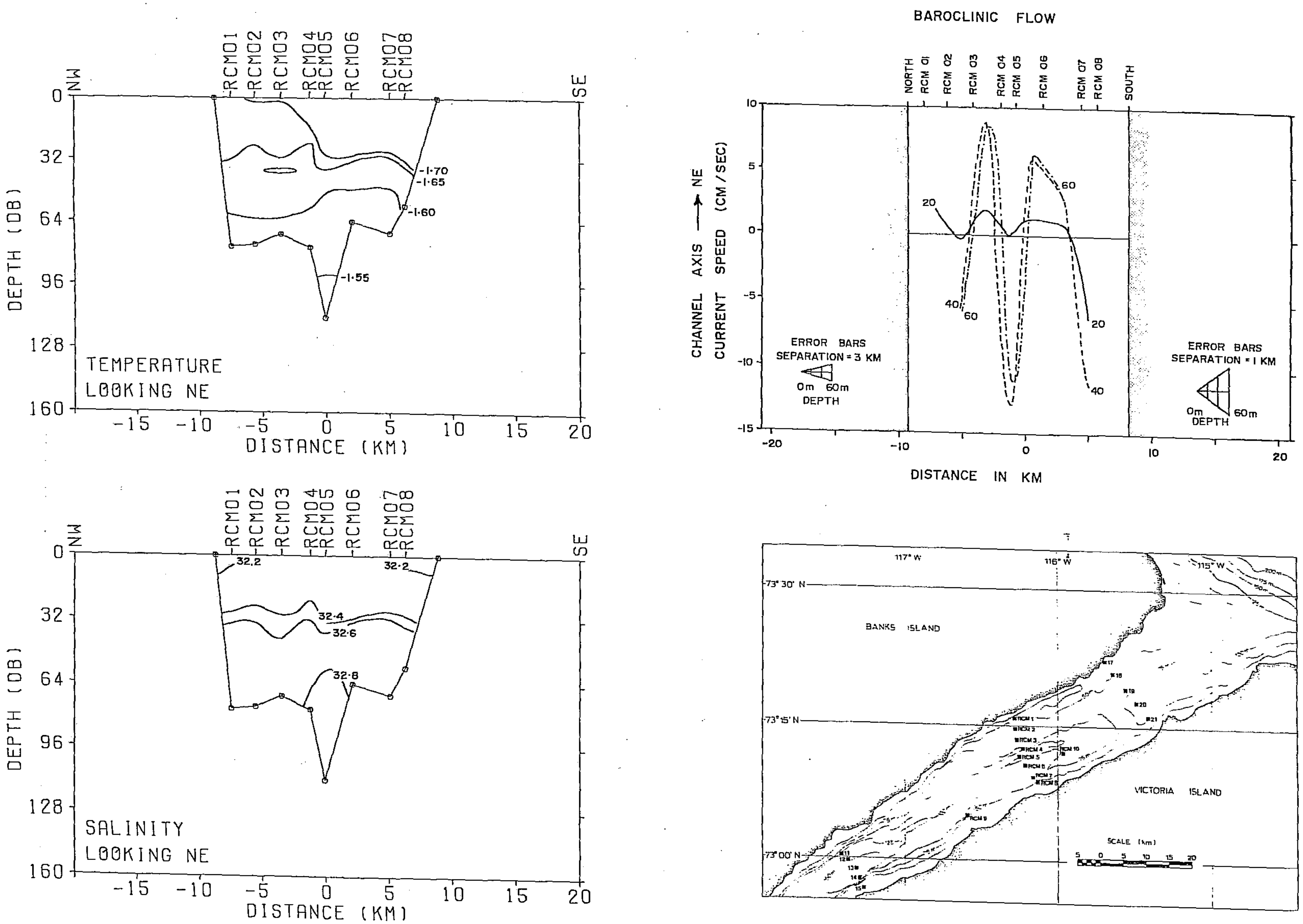

Figure 2.12 Temperature and salinity section and baroclinic velocity across northern Prince of Wales Strait, June 1982.

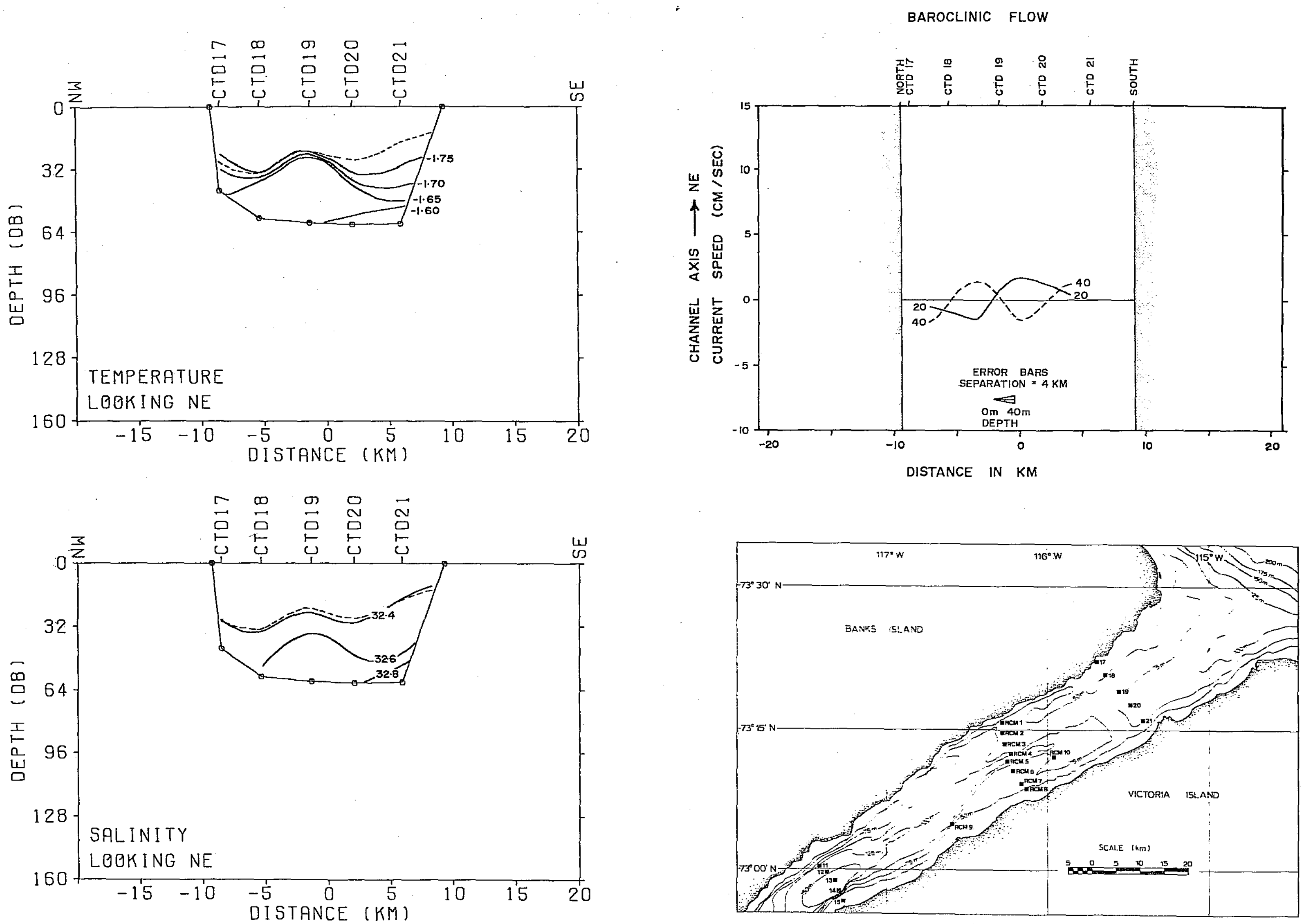

Figure 2.13 Temperature and salinity section and baroclinic velocity across northern Prince of Wales Strait near Passage Point, 1982.

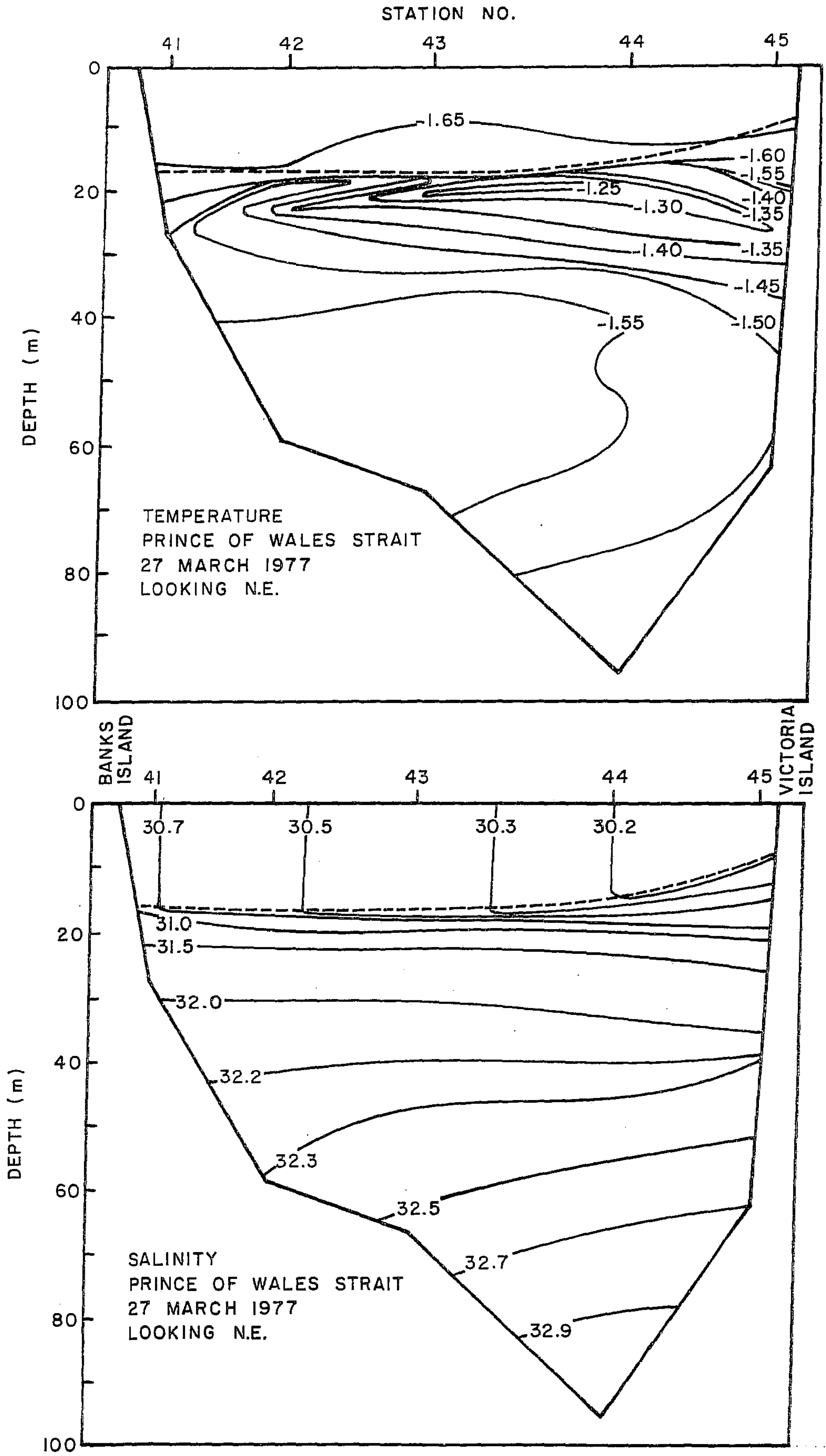

Figure 2.14 Temperature and salinity section across northern Prince of Wales Strait near Passage Point, late winter 1977 (data from Peck, 1978).

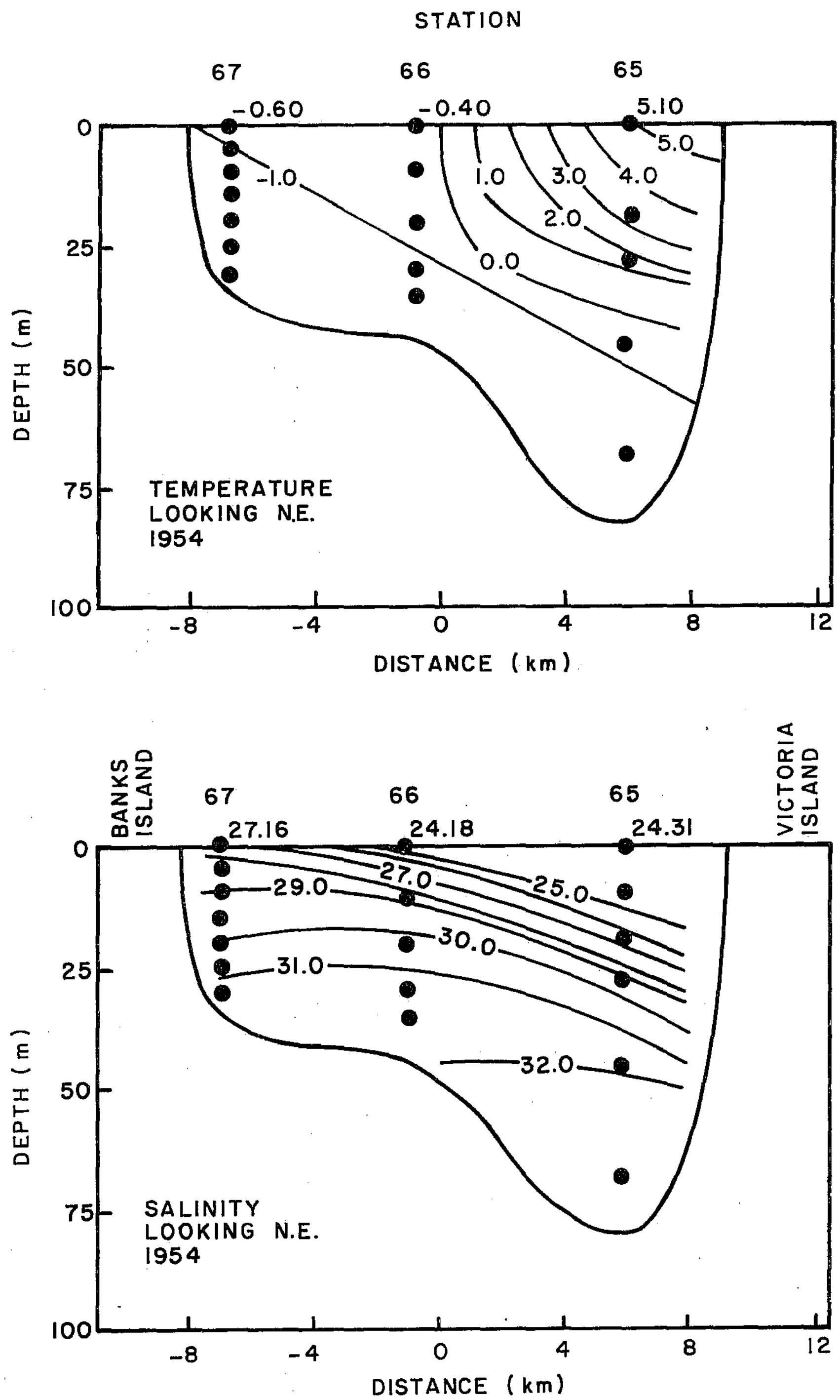

Figure 2.15 Temperature and salinity section across northern Prince of Wales Strait, summer 1954 (after Bailey, 1957).

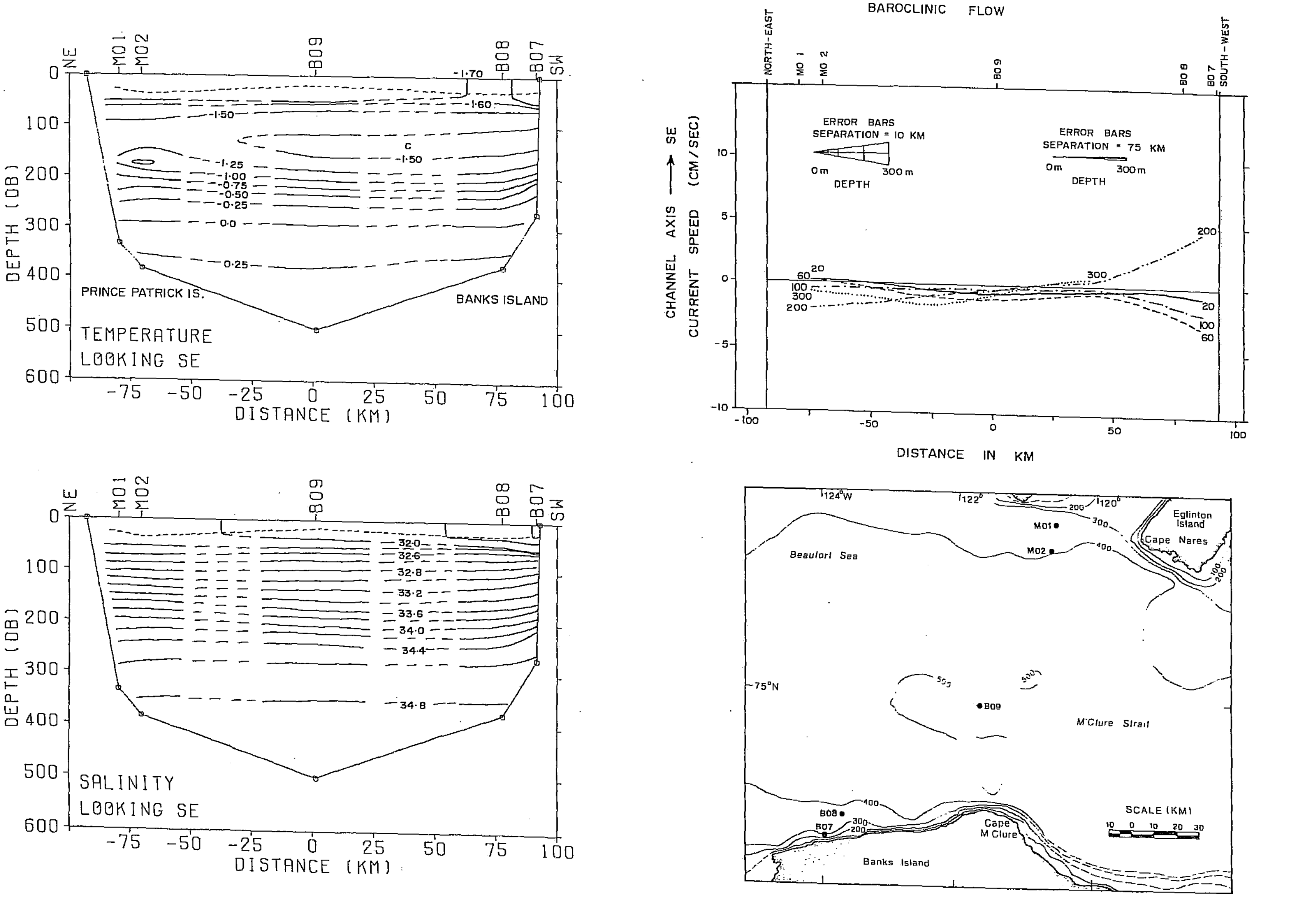

Figure 2.16 Temperature and salinity section and baroclinic velocity across western M'Clure Strait, 1984.

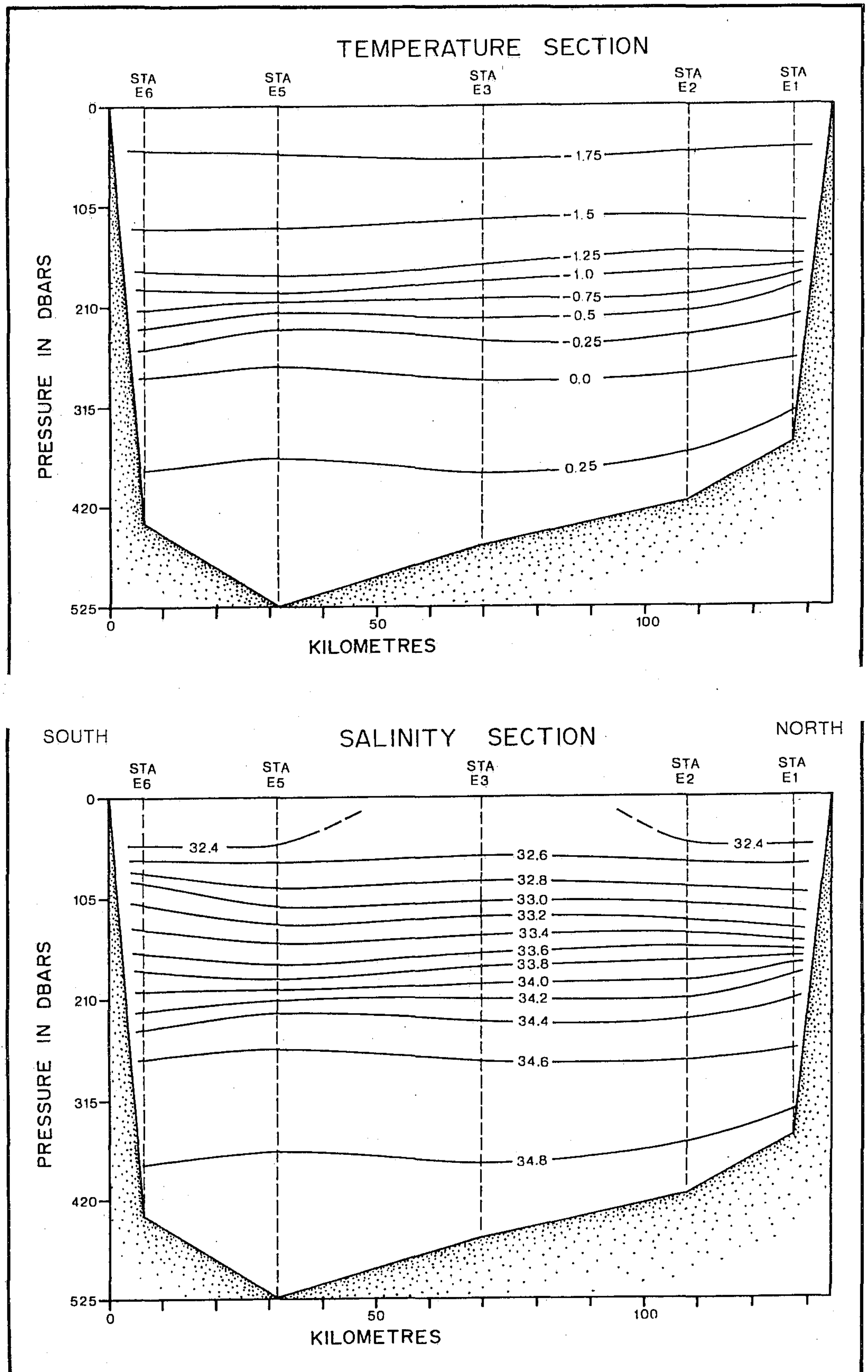

Figure 2.17 Temperature and salinity section across western M'Clure Strait, 1982. (from Fissel et al. 1984b). Note that the viewing direction is towards the west in this figure and towards the east in the previous figure.

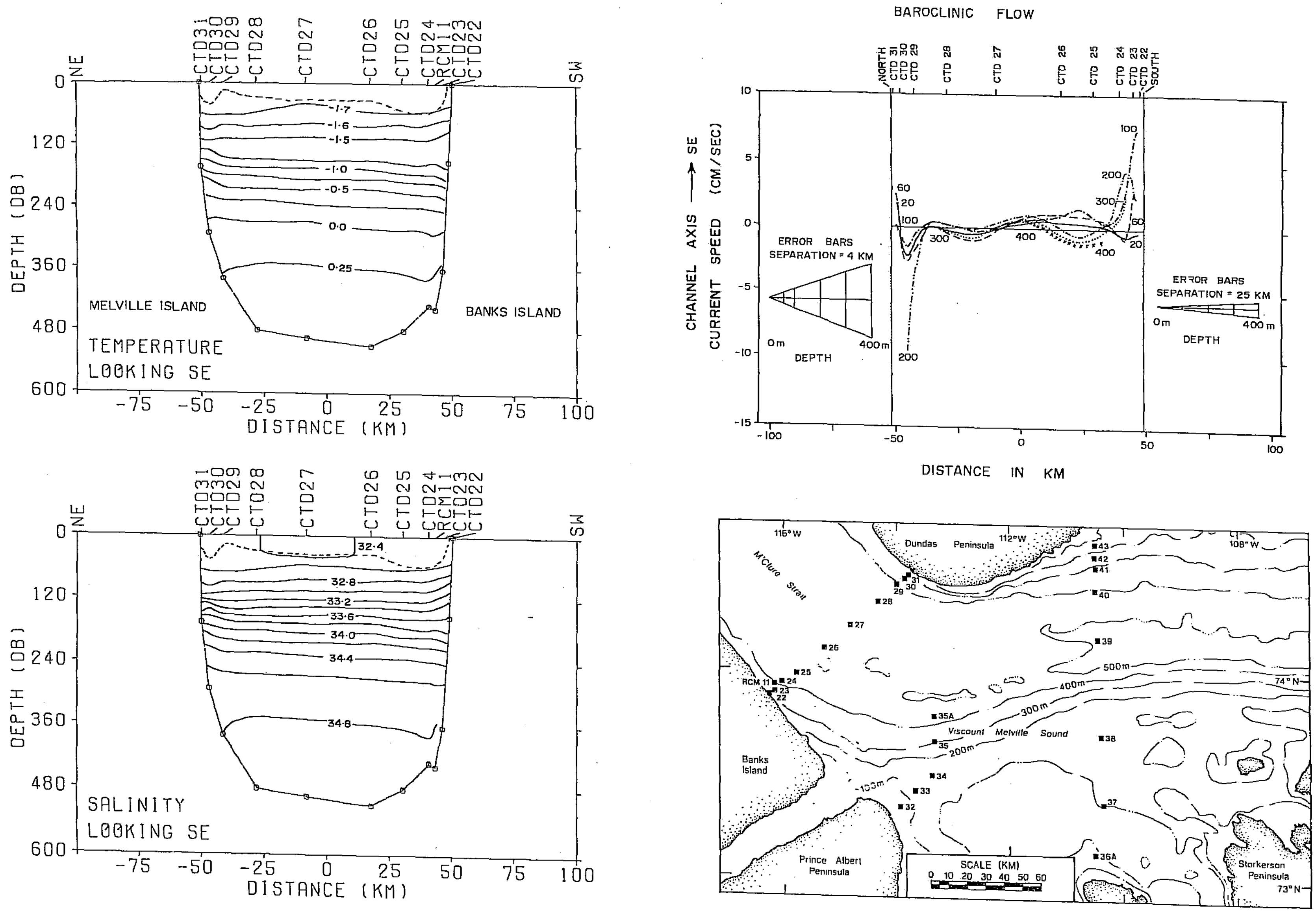

Figure 2.18 Temperature and salinity section and baroclinic velocity across eastern M'Clure Strait (Parker Pt. to C. Dundas), 1982.

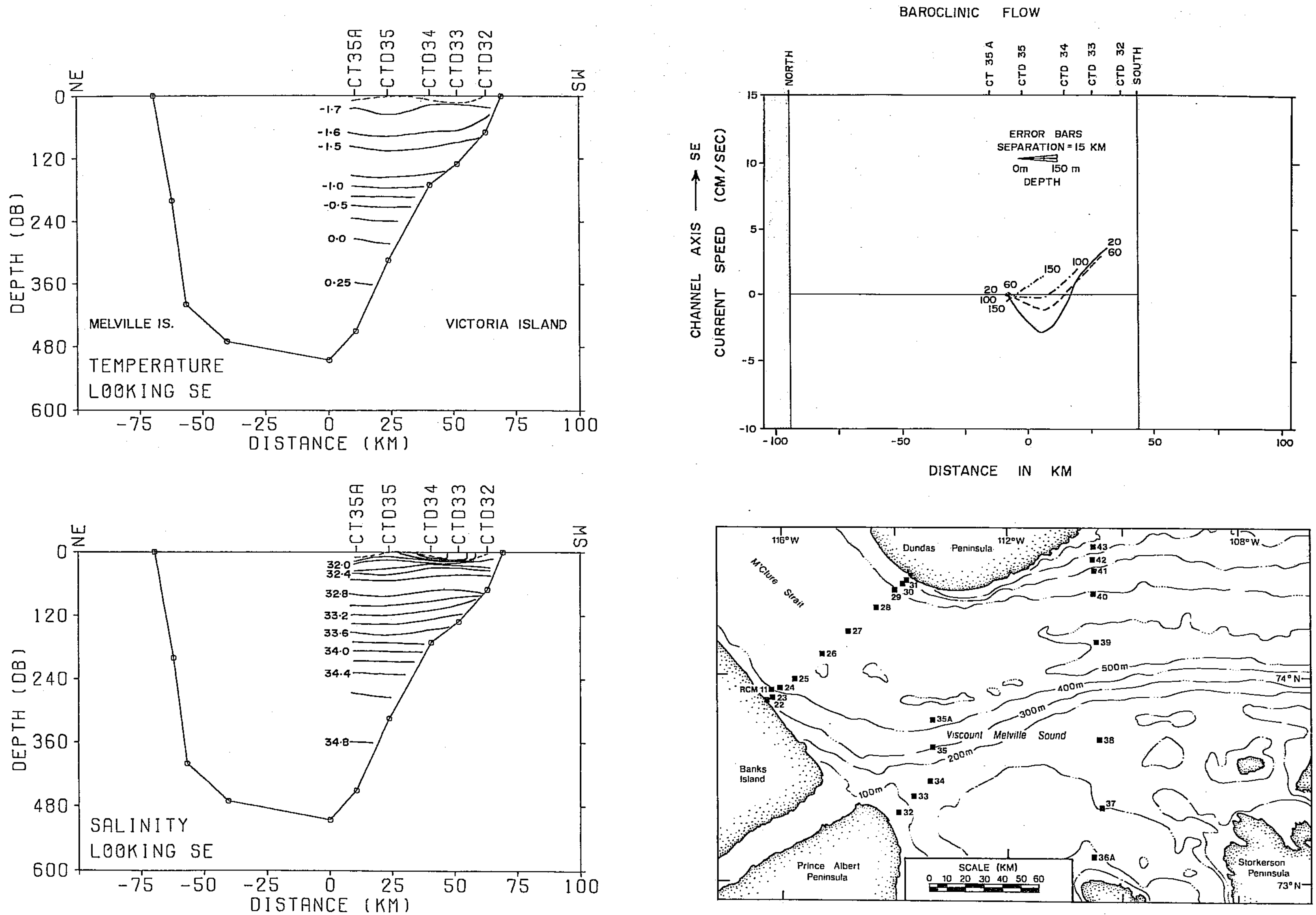

Figure 2.19 Partial temperature and salinity section and baroclinic velocity across Viscount Melville Sound (from Peel Pt.), 1982.

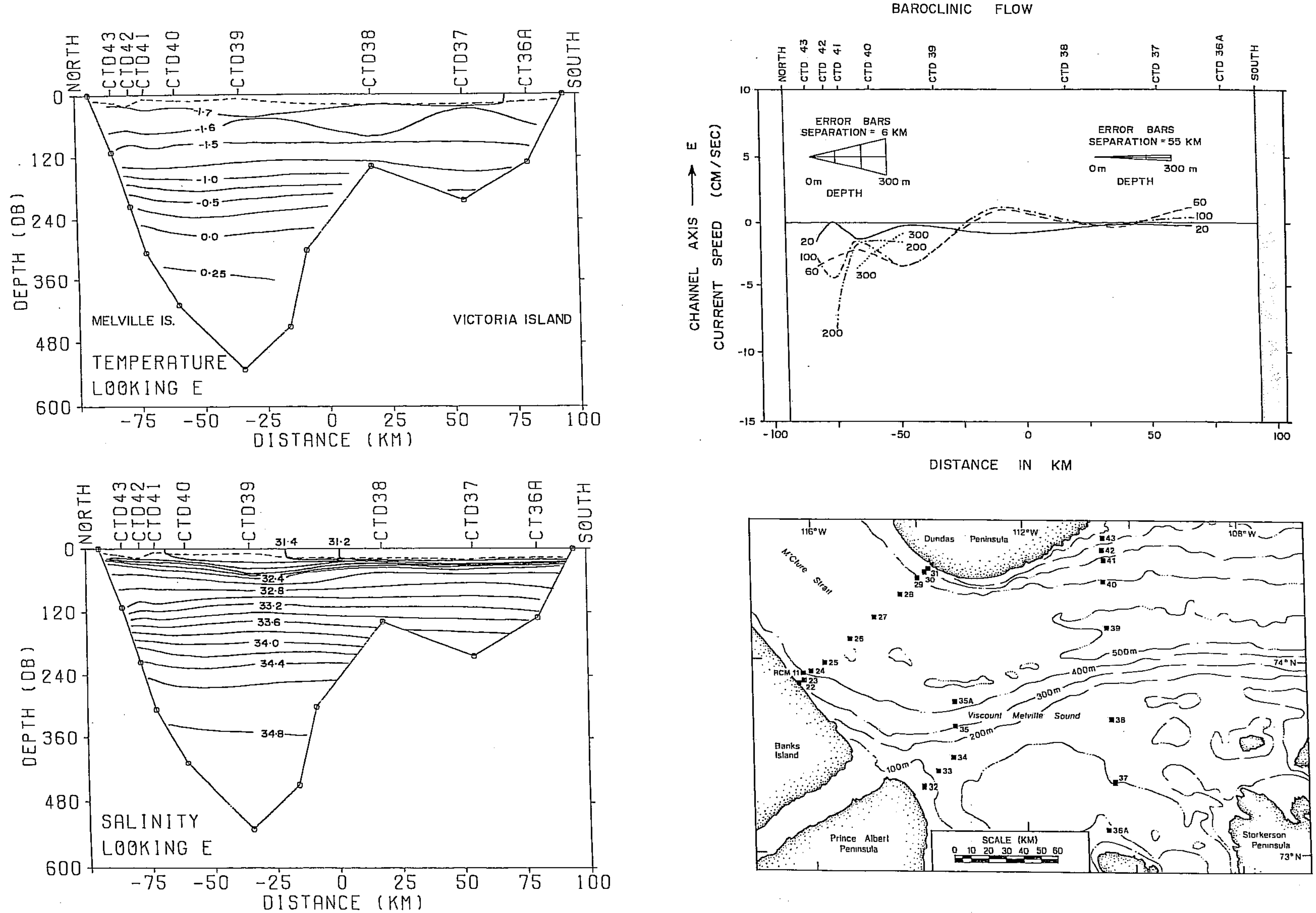

Figure 2.20 Temperature and salinity section and baroclinic velocity across central Viscount Melville Sound (south of Winter Harbour), 1982.

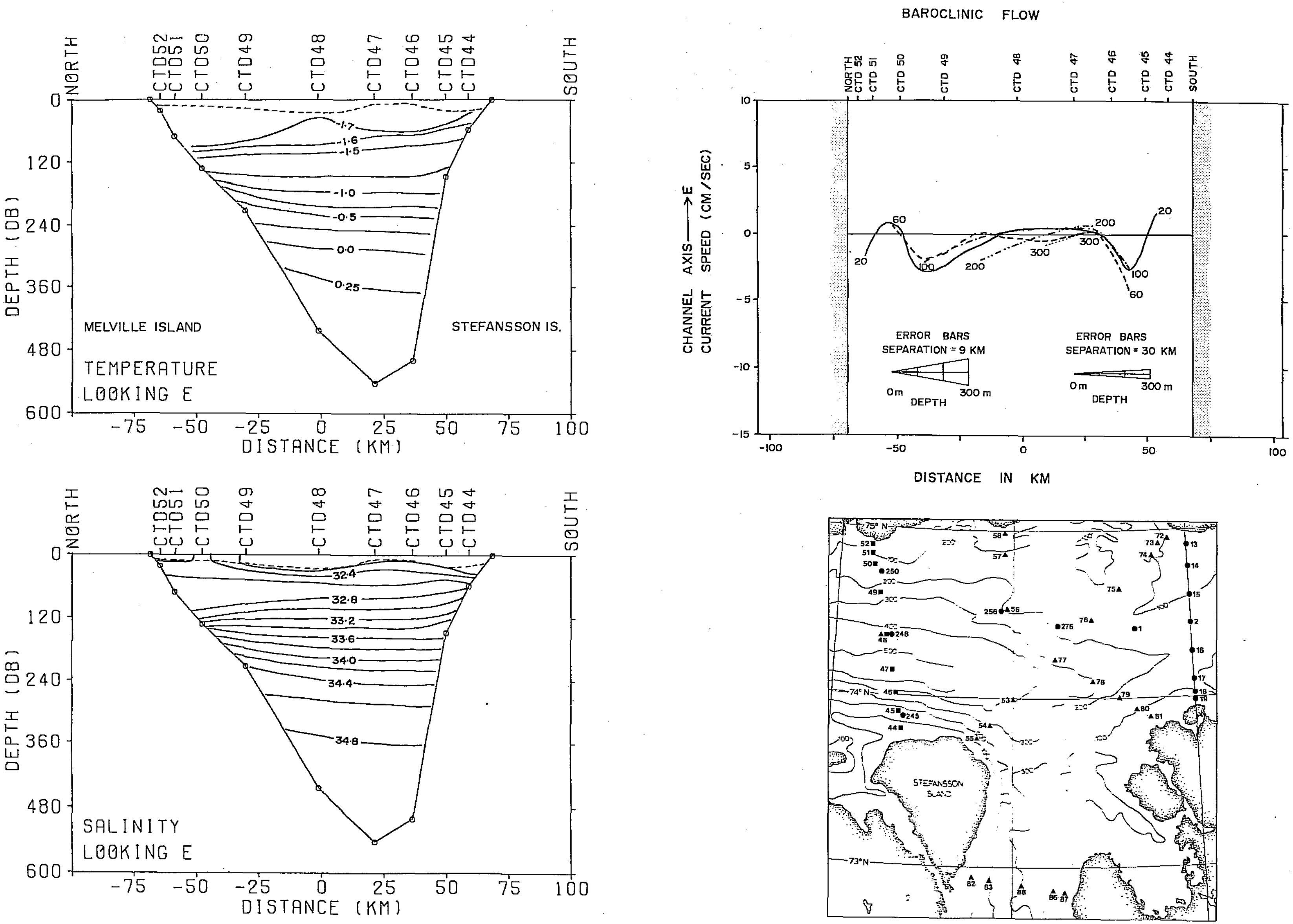

Figure 2.21 Temperature and salinity section and baroclinic velocity across central Viscount Melville Sound (south of Ross Point), 1982.

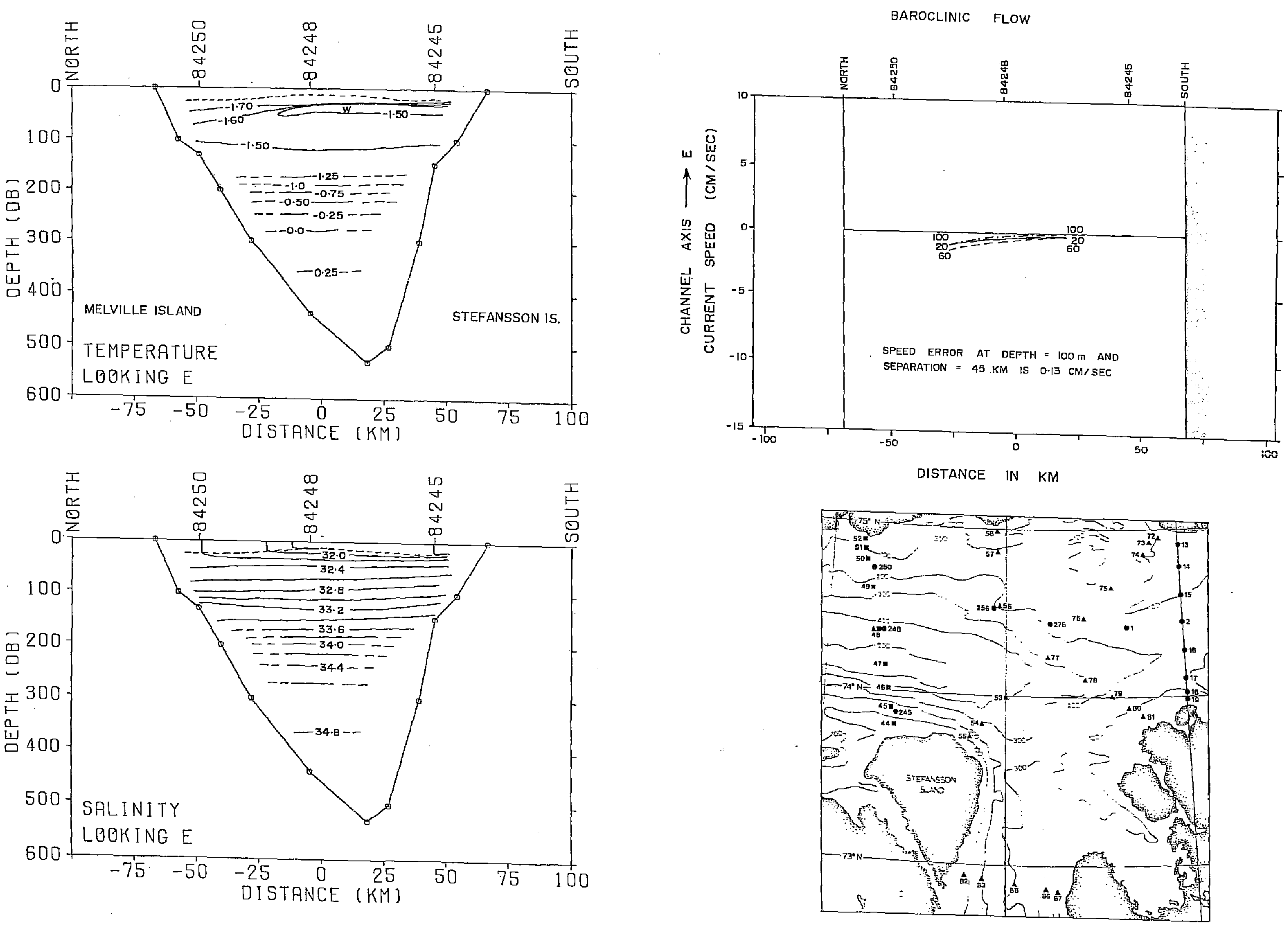

Figure 2.22 Temperature and salinity section and baroclinic velocity across central Viscount Melville Sound (south of Ross Point), 1984.

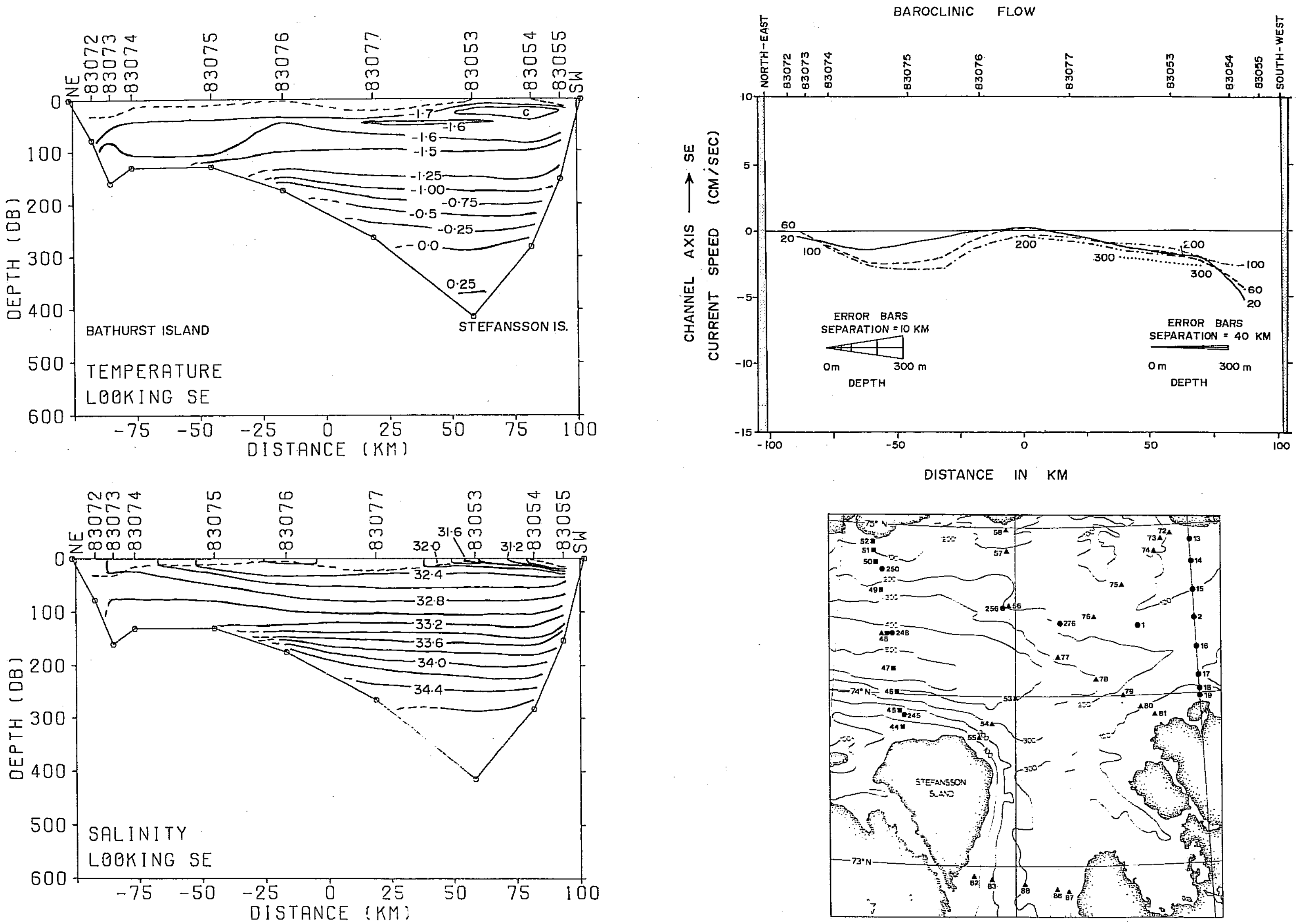

Figure 2.23 Temperature and salinity section and baroclinic velocity diagonally across eastern Viscount Melville Sound (Bathurst Is. to Stefansson Is.), 1983.

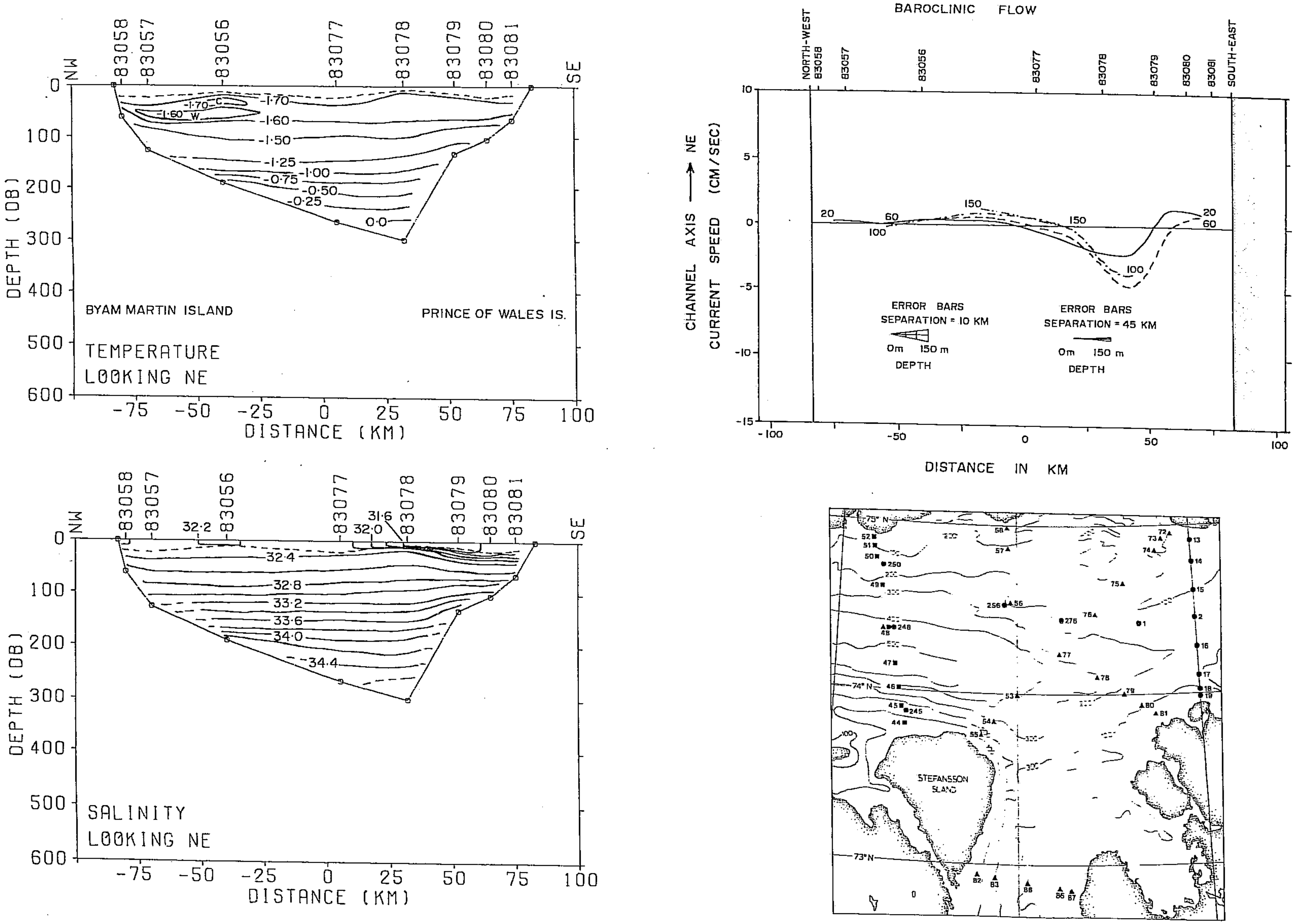

Figure 2.24 Temperature and salinity section and baroclinic velocity diagonally across eastern Viscount Melville Sound (Byam Martin Is. to Milne Pt., Prince of Wales Is.), 1983.

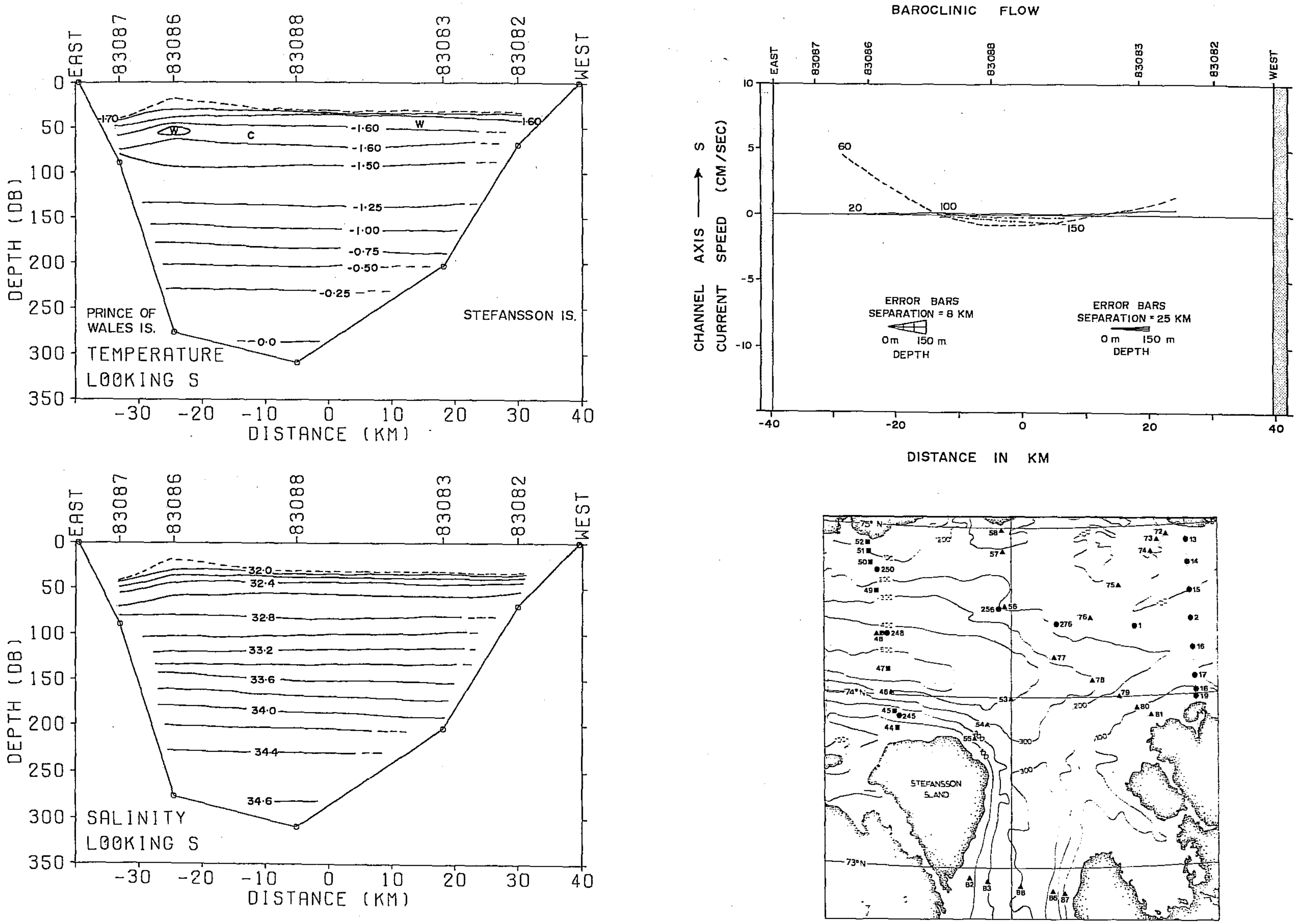

Figure 2.25 Temperature and salinity section and baroclinic velocity across northern M'Clintock Channel (C. Richard Collinson, Prince of Wales Is. to Stefansson Is.), 1983.

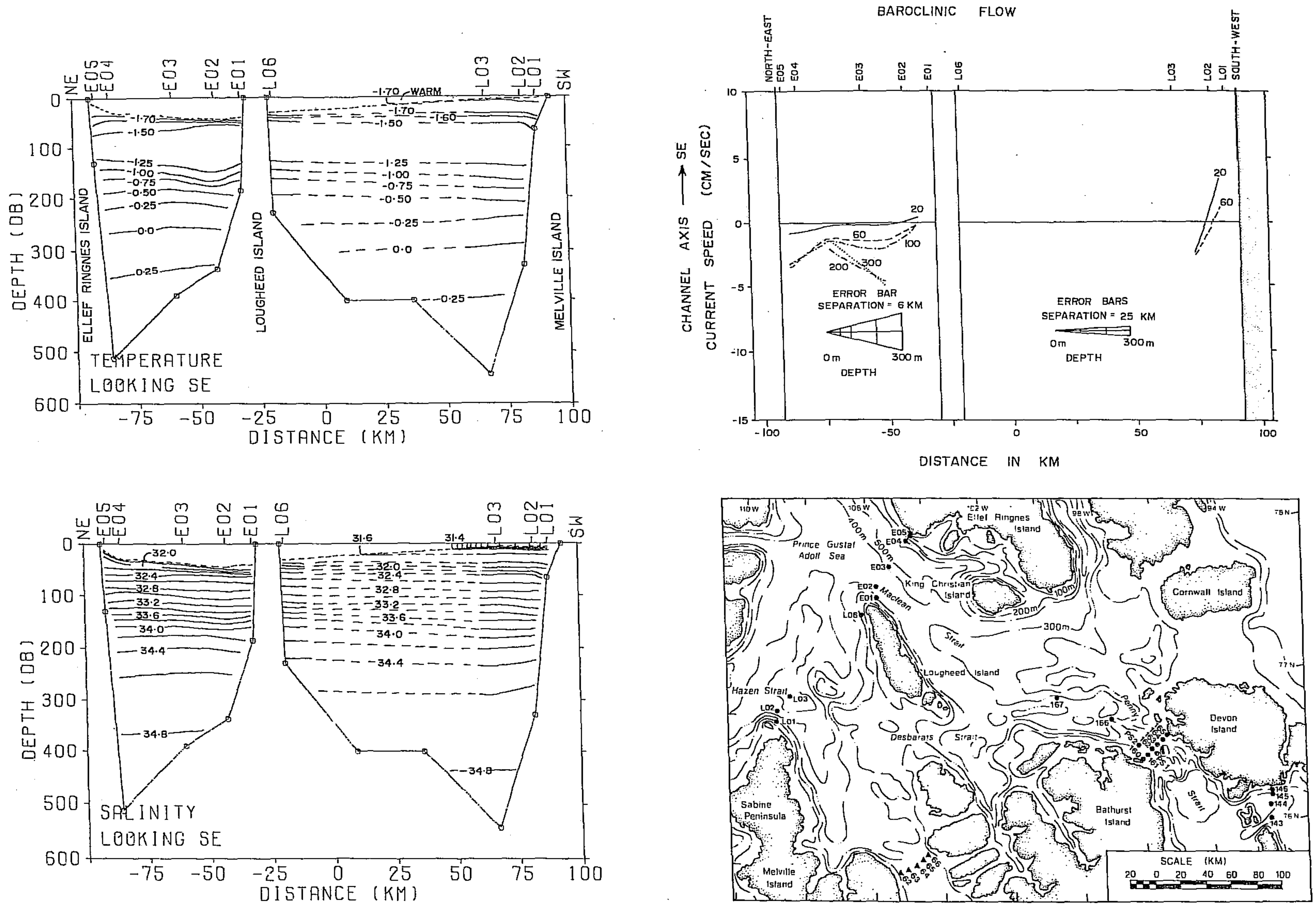

Figure 2.26 Temperature and salinity section and baroclinic velocity across the Sverdrup Basin (Ellef Ringnes to Melville Is.), 1984.

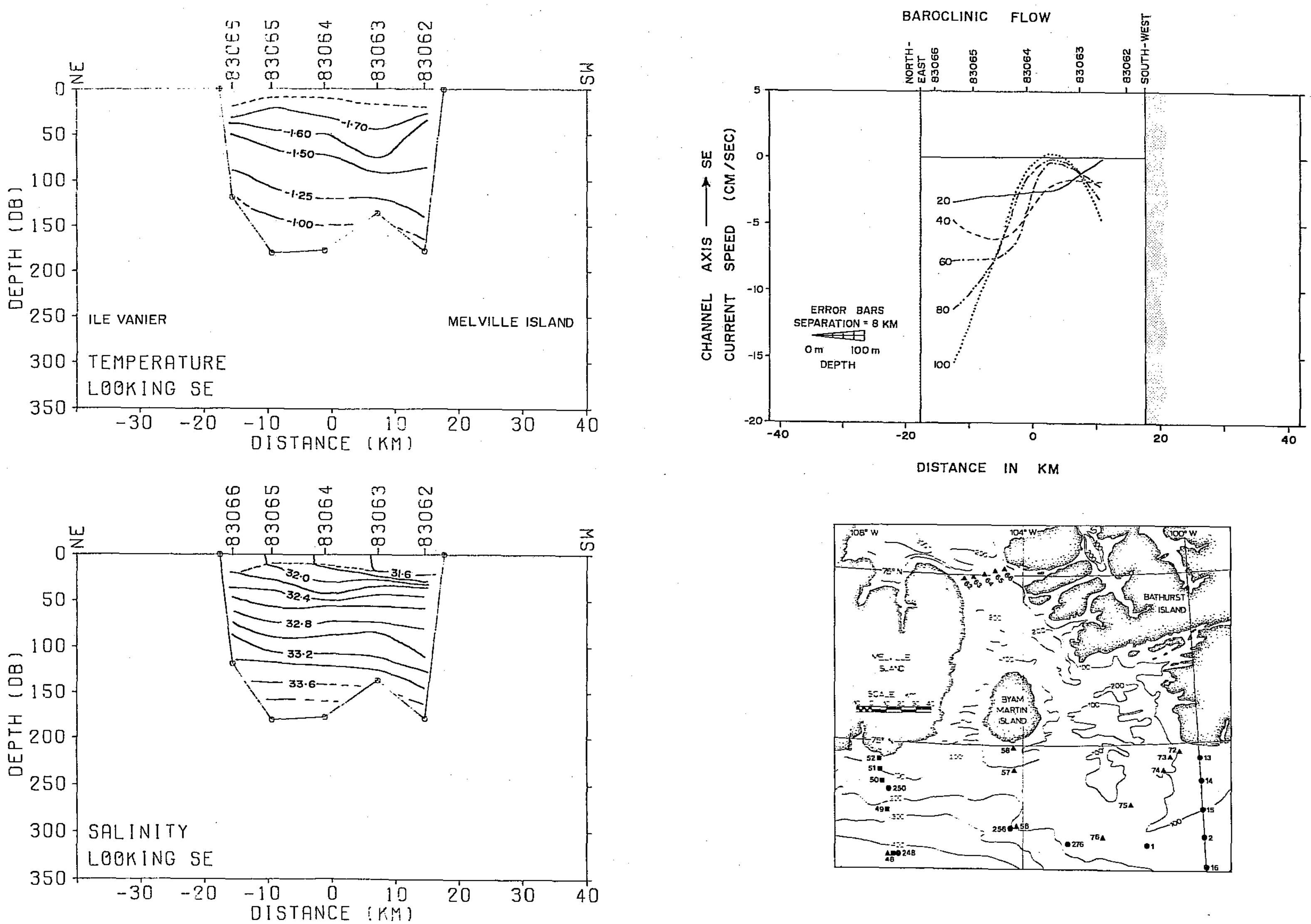

Figure 2.27 Temperature and salinity section and baroclinic velocity across Byam Martin Channel (Ile Vanier to Melville Is.), 1983.

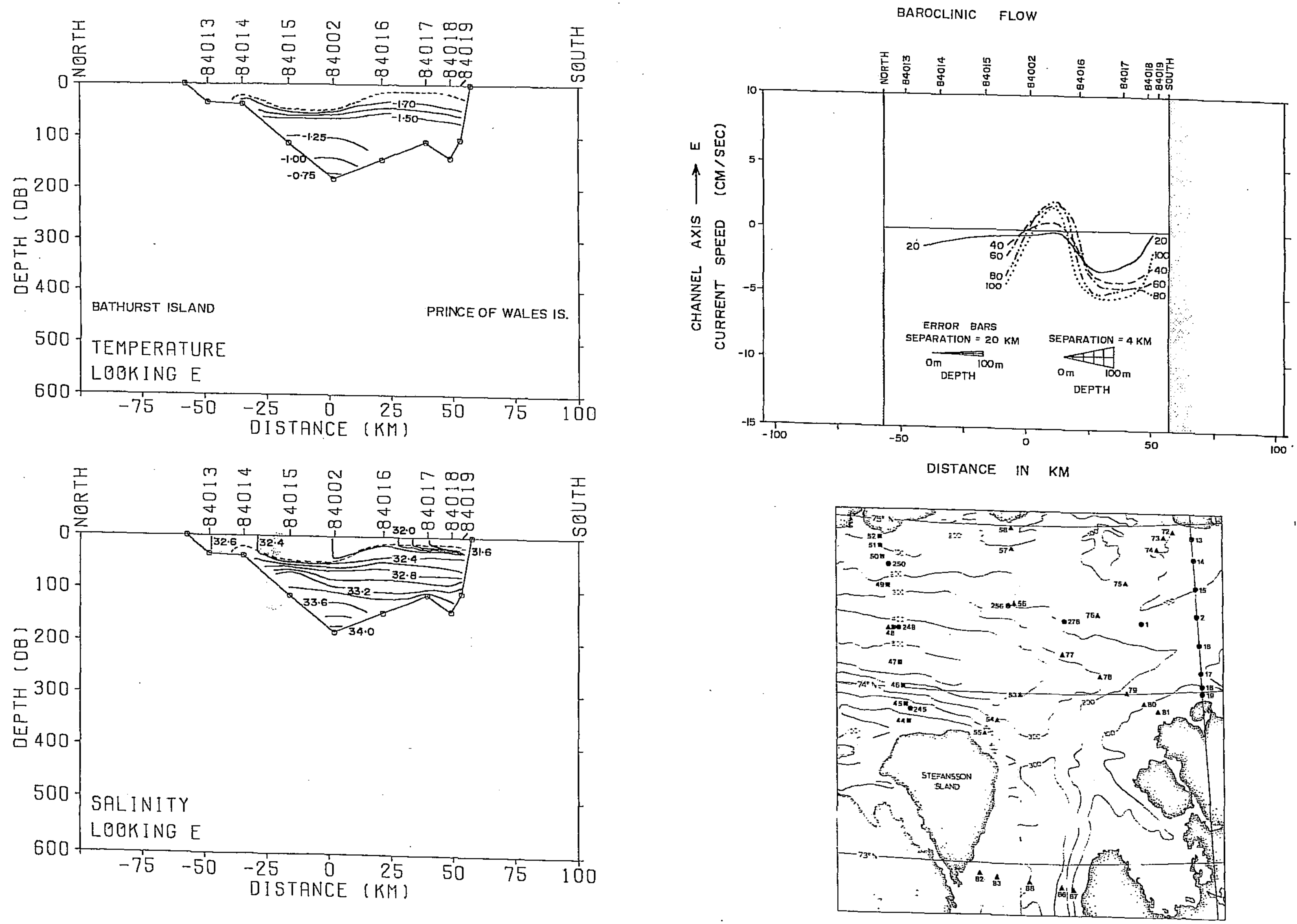

Figure 2.28 Temperature and salinity section and baroclinic velocity across western Barrow Strait (Bathurst Is. to Prince of Wales Is.), 1984.

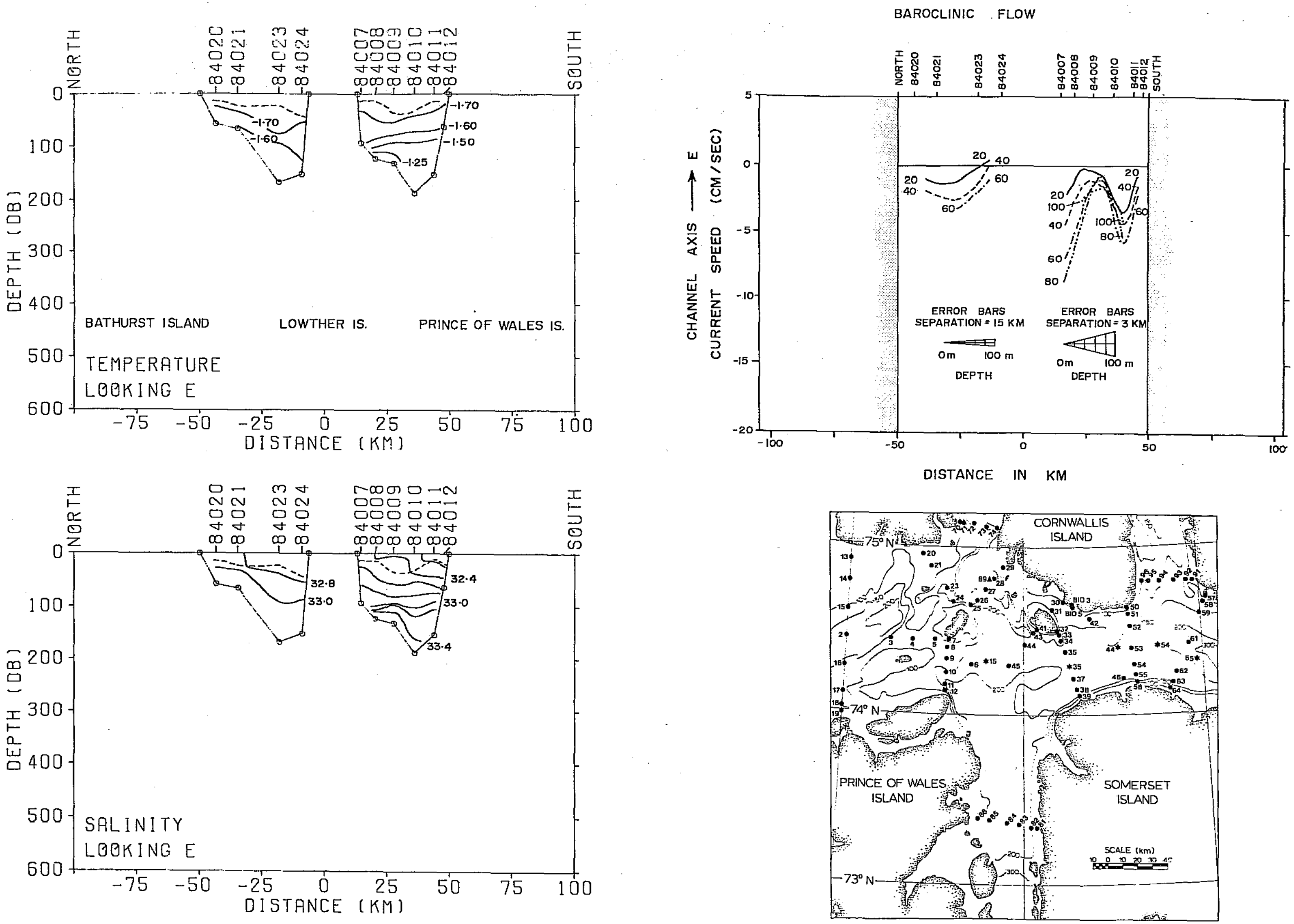

Figure 2.29 Temperature and salinity section and baroclinic velocity across western Barrow Strait at Lowther Is., 1984.

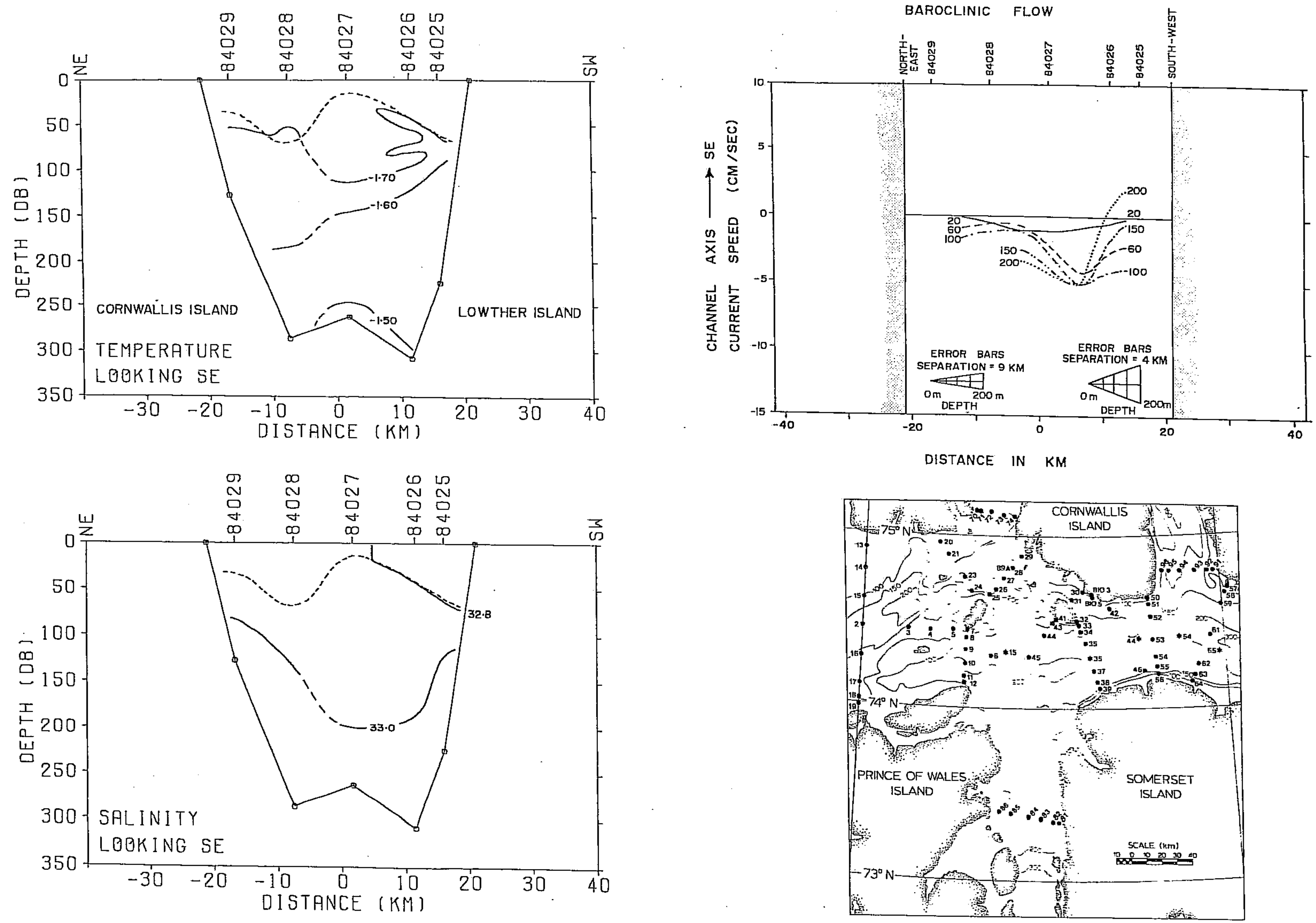

Figure 2.30 Temperature and salinity section and baroclinic velocity across north-central Barrow Strait (Cornwallis Is. to Lowther Is.), 1984.

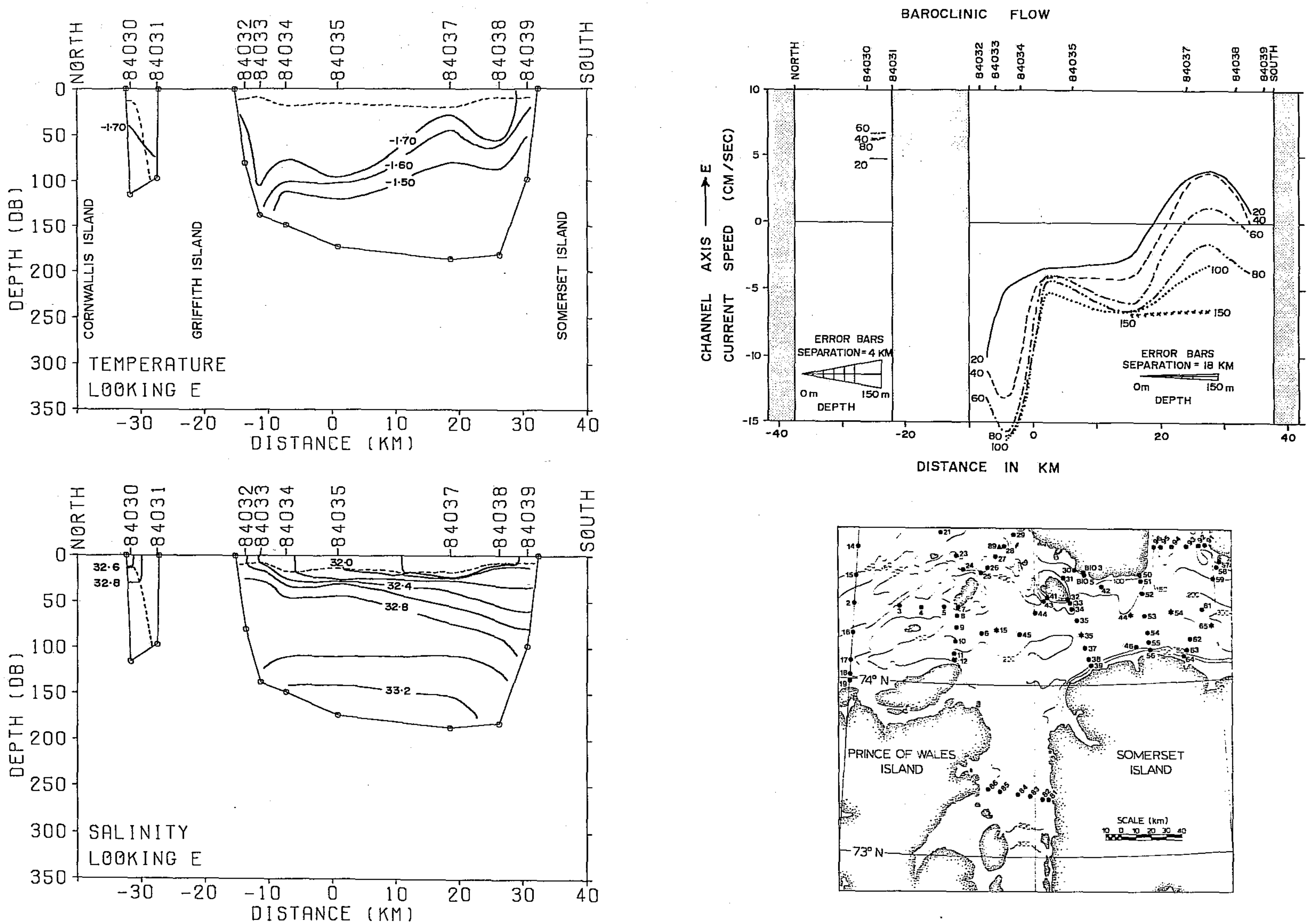

Figure 2.31 Temperature and salinity section and baroclinic velocity across central Barrow Strait at Griffith Is., 1984.

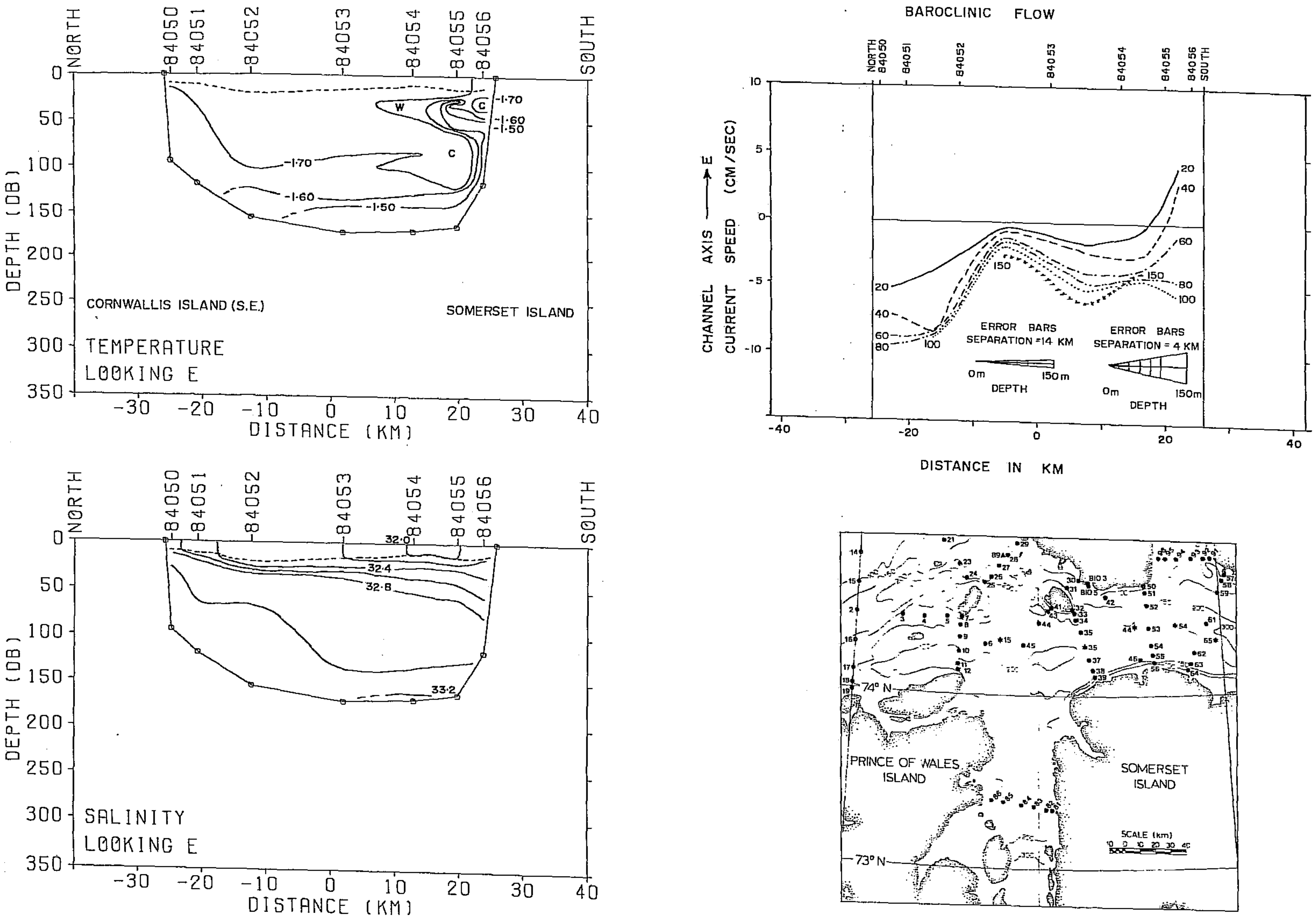

Figure 2.32 Temperature and salinity section and baroclinic velocity across central Barrow Strait (C. Dungeness, Cornwallis Is. to Somerset Is.), 1984.

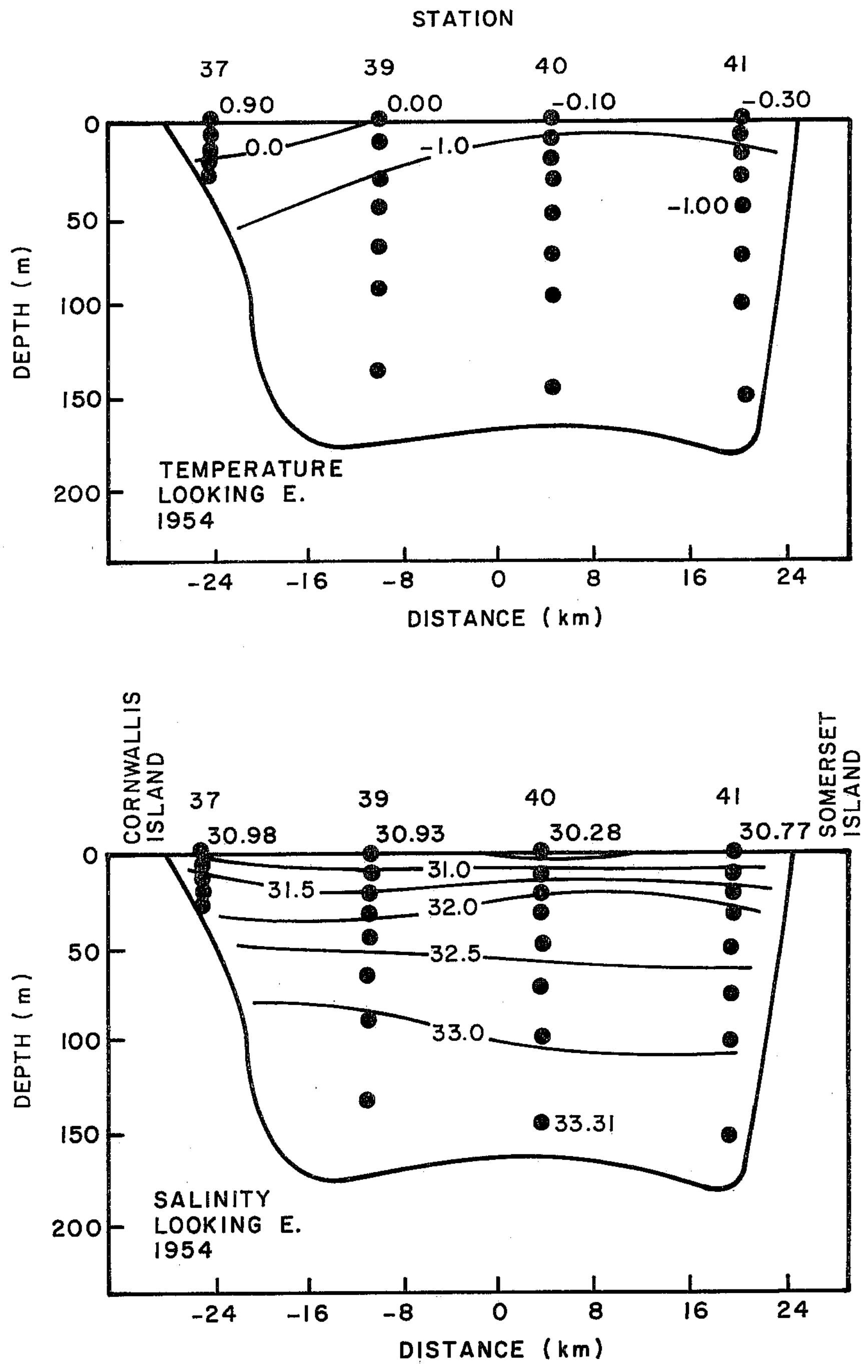

Figure 2.33 Temperature and salinity section across Barrow Strait (close east of the section in Figure 2.31), summer 1954 (after Bailey, 1957).

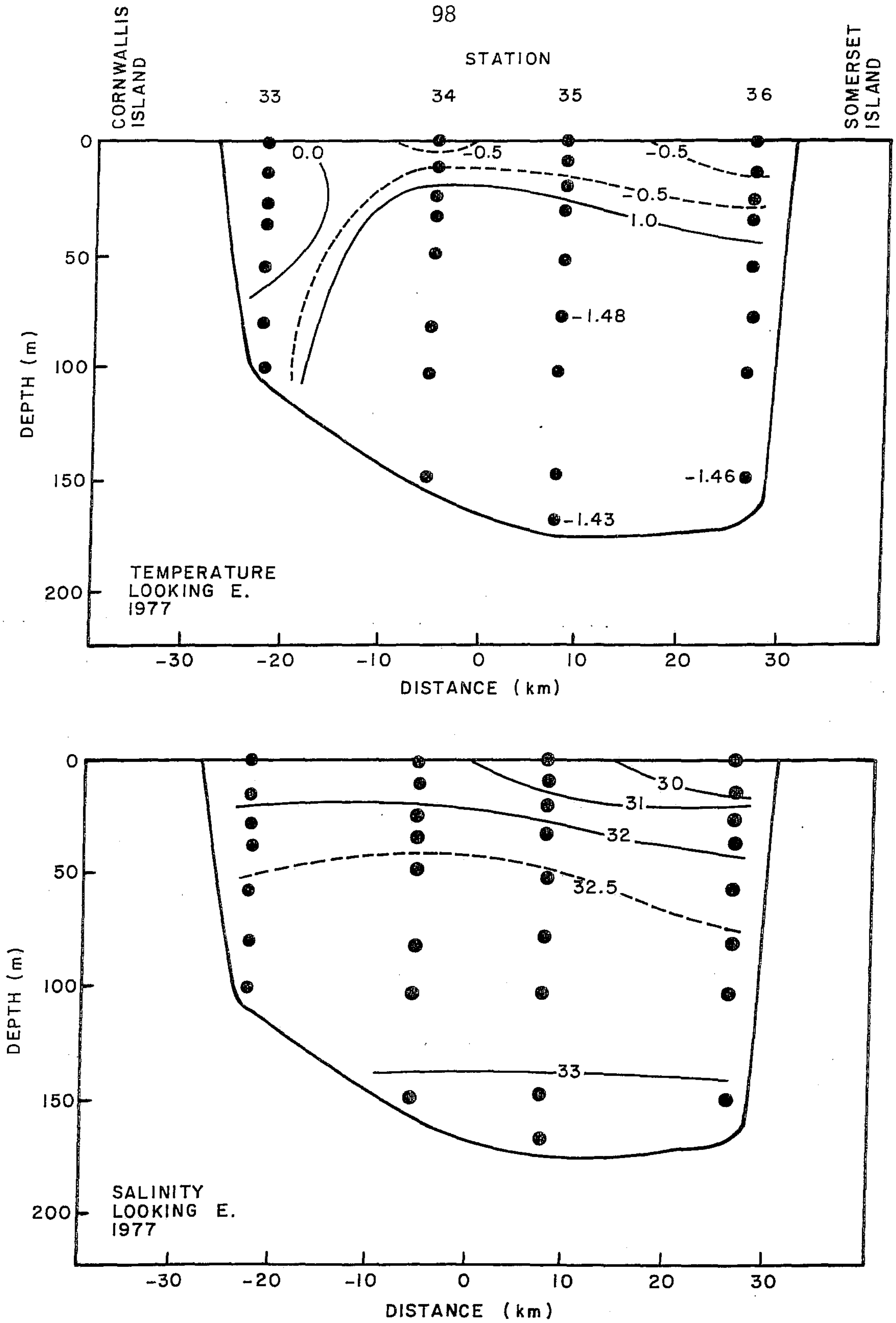

Figure 2.34 Temperature and salinity section across Barrow Strait (close to the section in Figure 2.32), summer 1977 (after Jones and Coote, 1980).

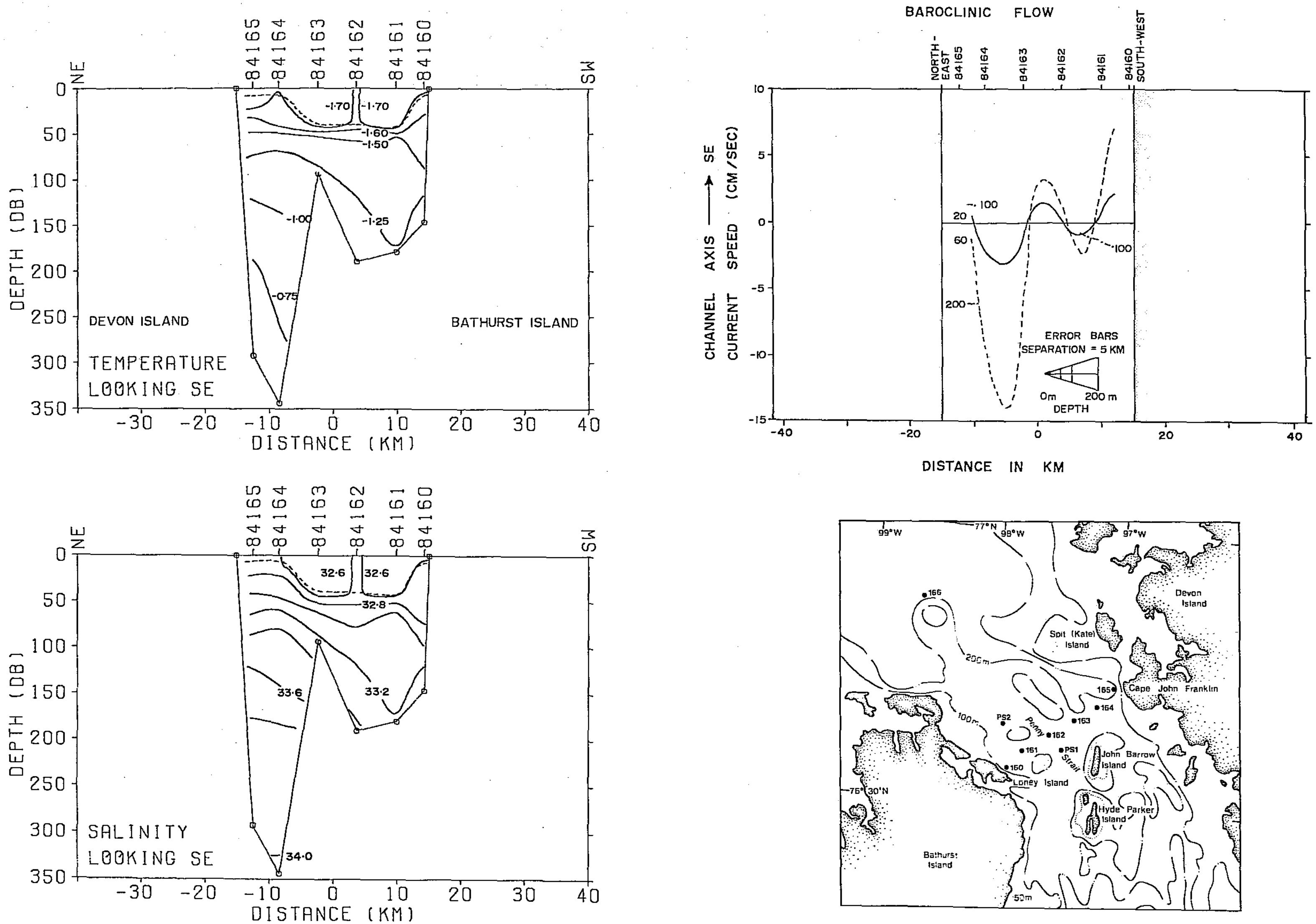

Figure 2.35 Temperature and salinity section and baroclinic velocity across Penny Strait at C. John Franklin, 1984.

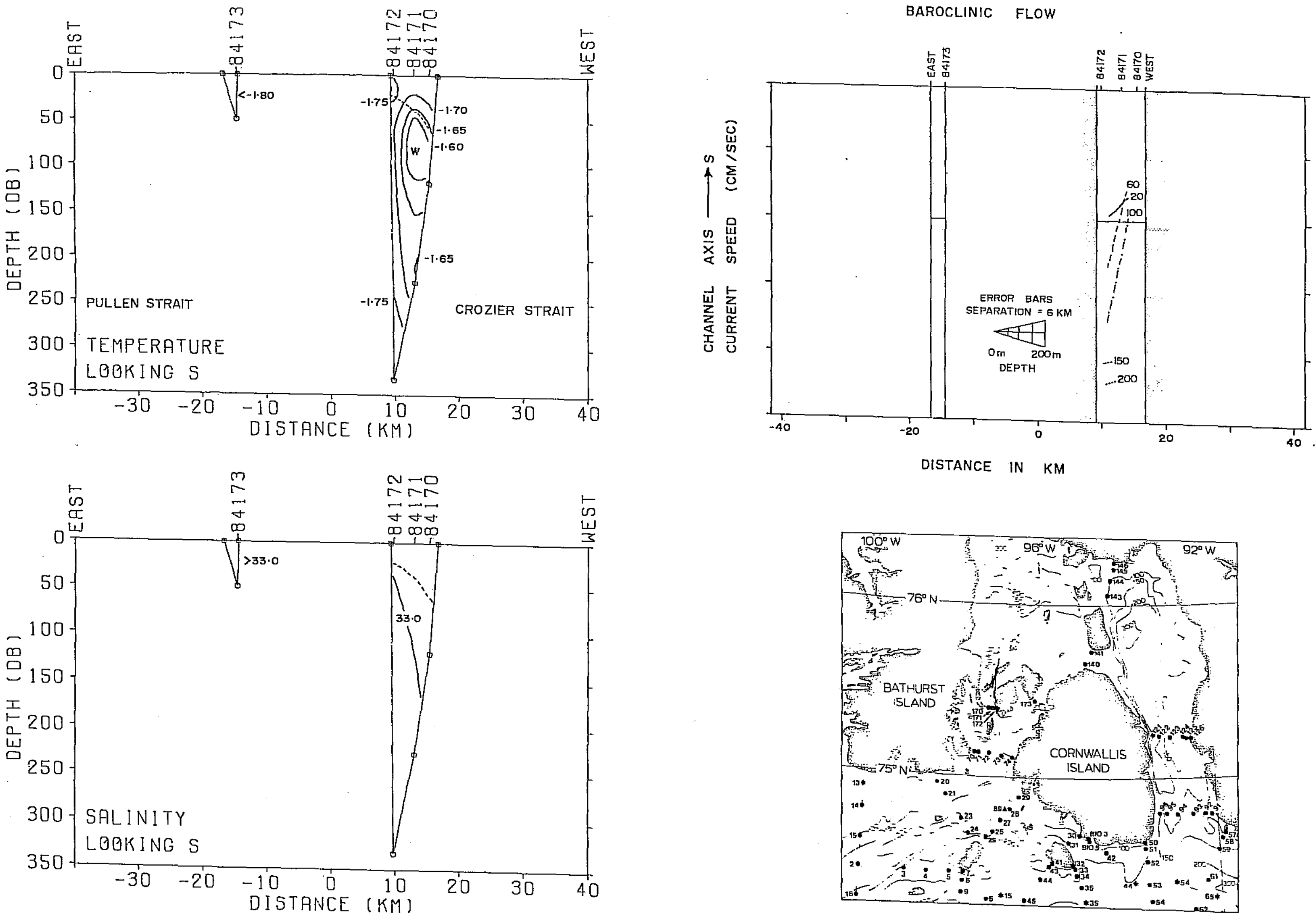

Figure 2.36 Temperature and salinity section and baroclinic velocity across Pullen and Crozier Straits, 1984.

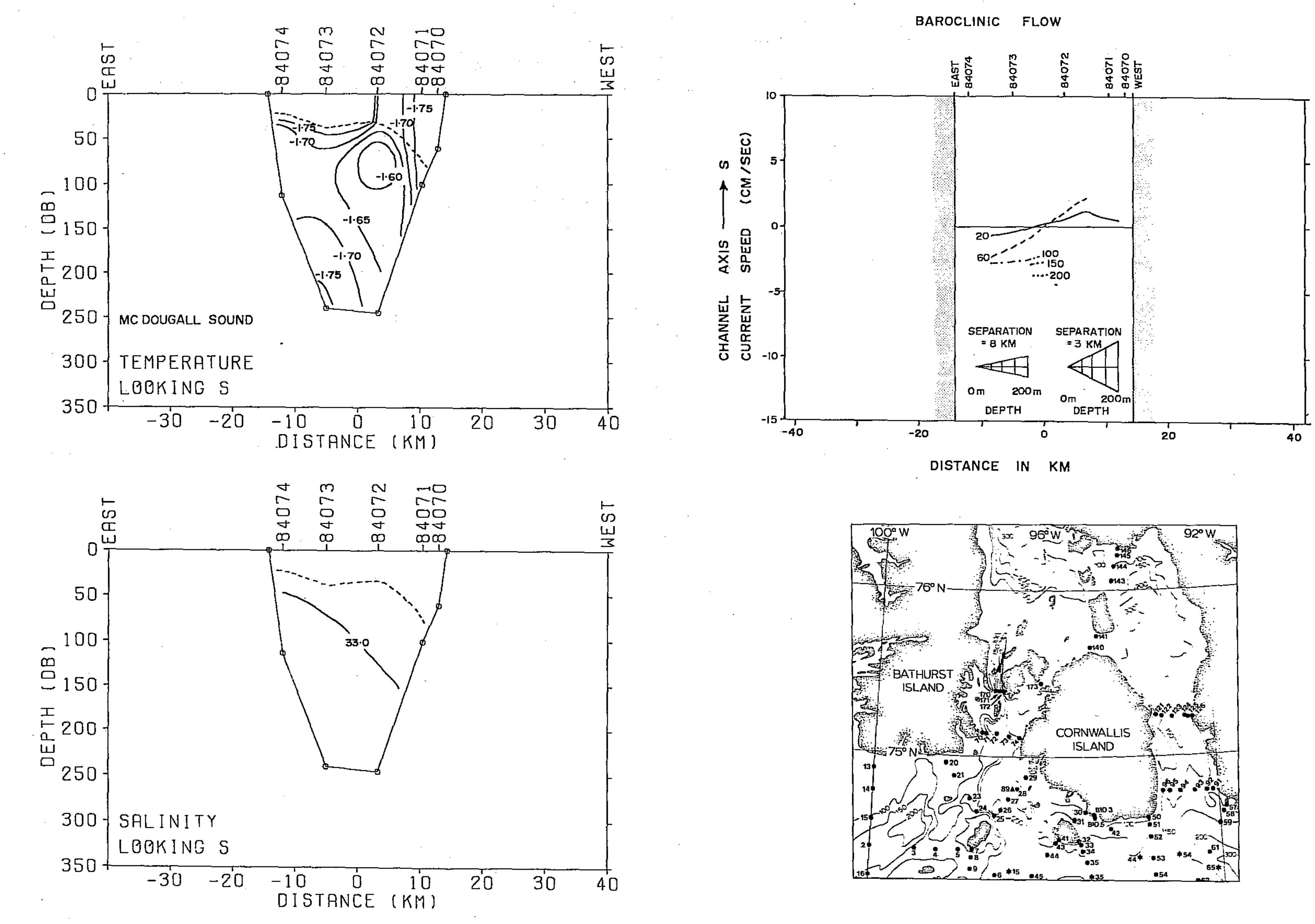

Figure 2.37 Temperature and salinity section and baroclinic velocity across central McDougall Sound, 1984.

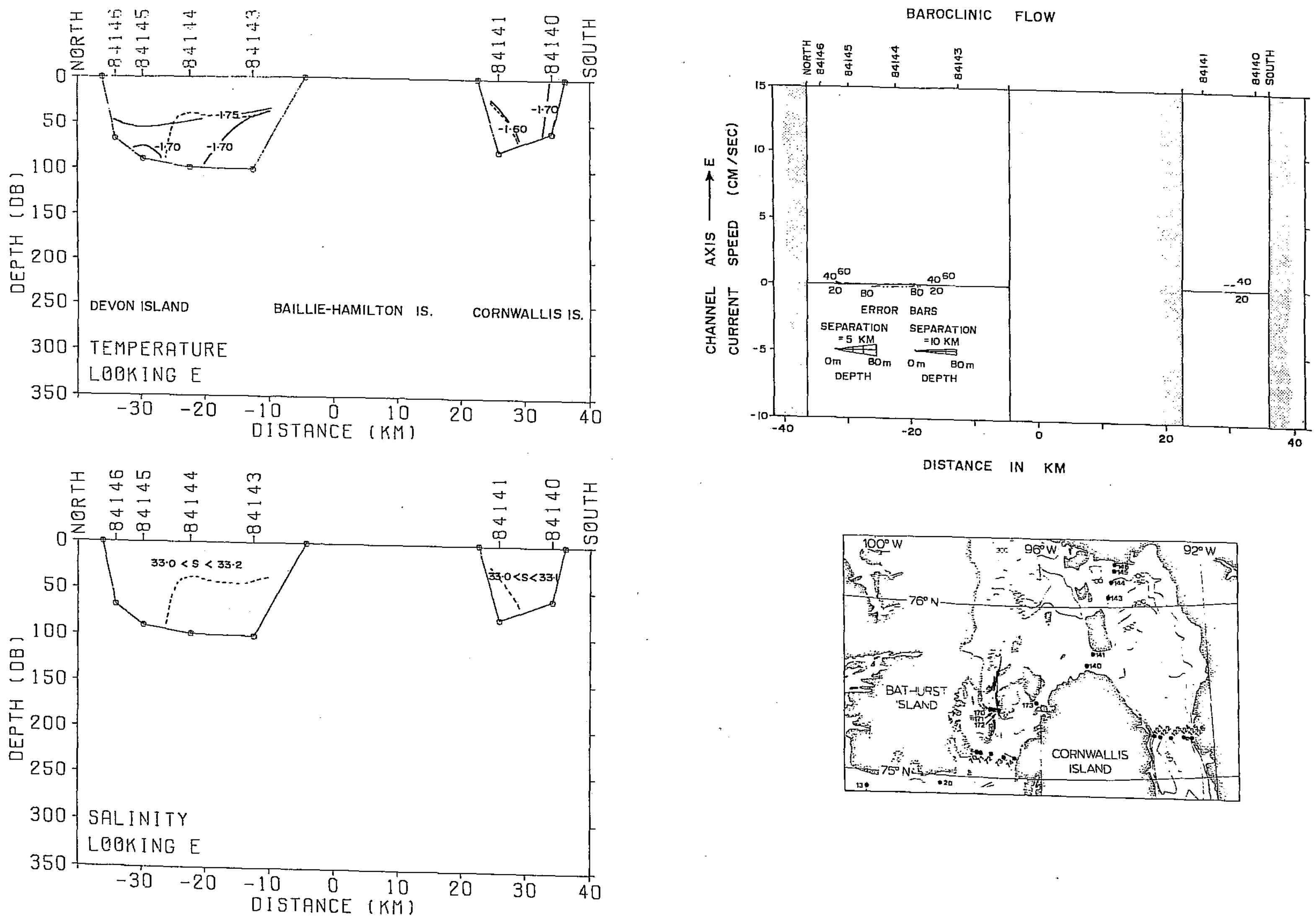

Figure 2.38 Temperature and salinity section and baroclinic velocity across northern Wellington Channel at Baillie-Hamilton Is., 1984.

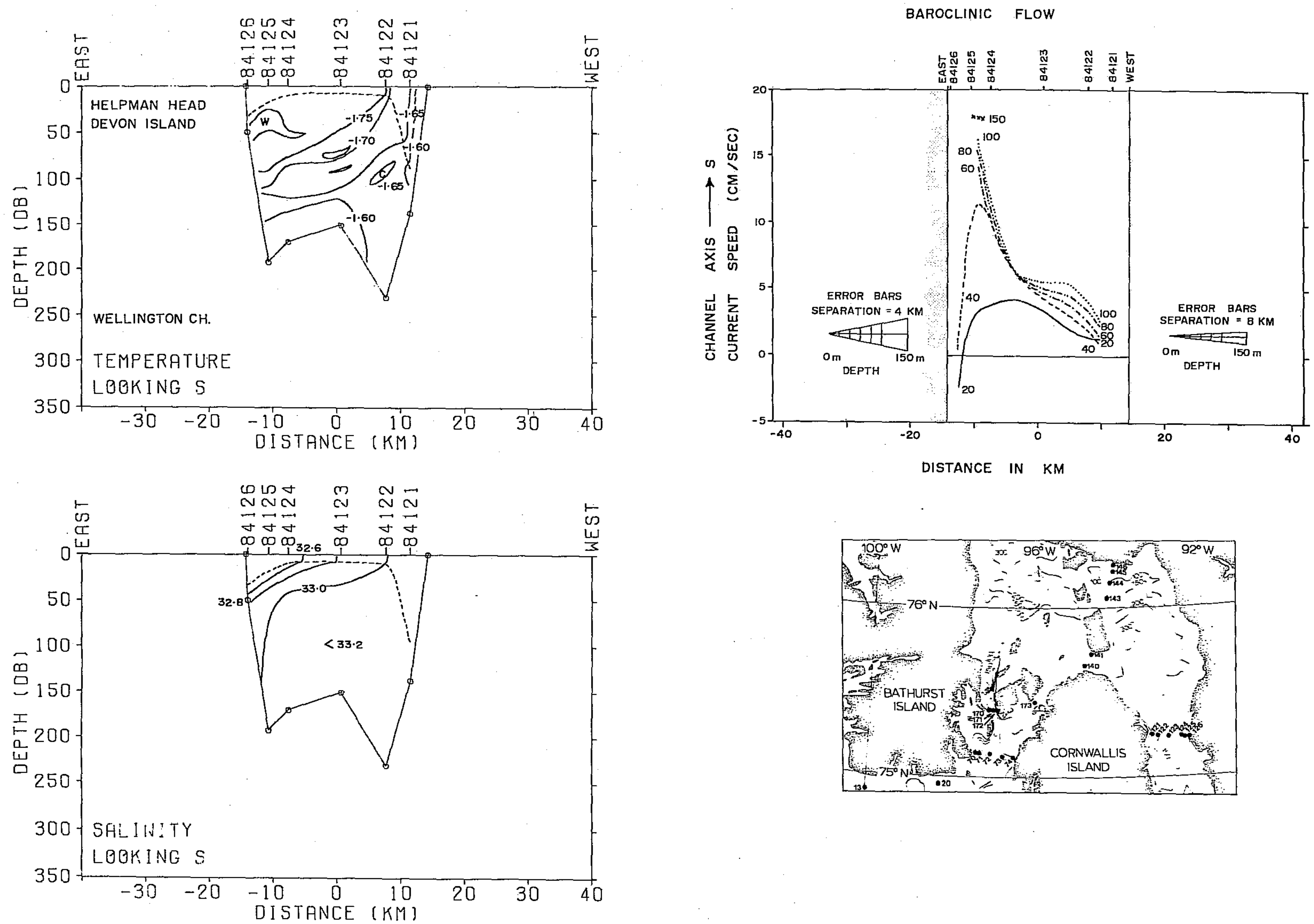

Figure 2.39 Temperature and salinity section and baroclinic velocity across central Wellington Channel at Helpman Head, 1984.

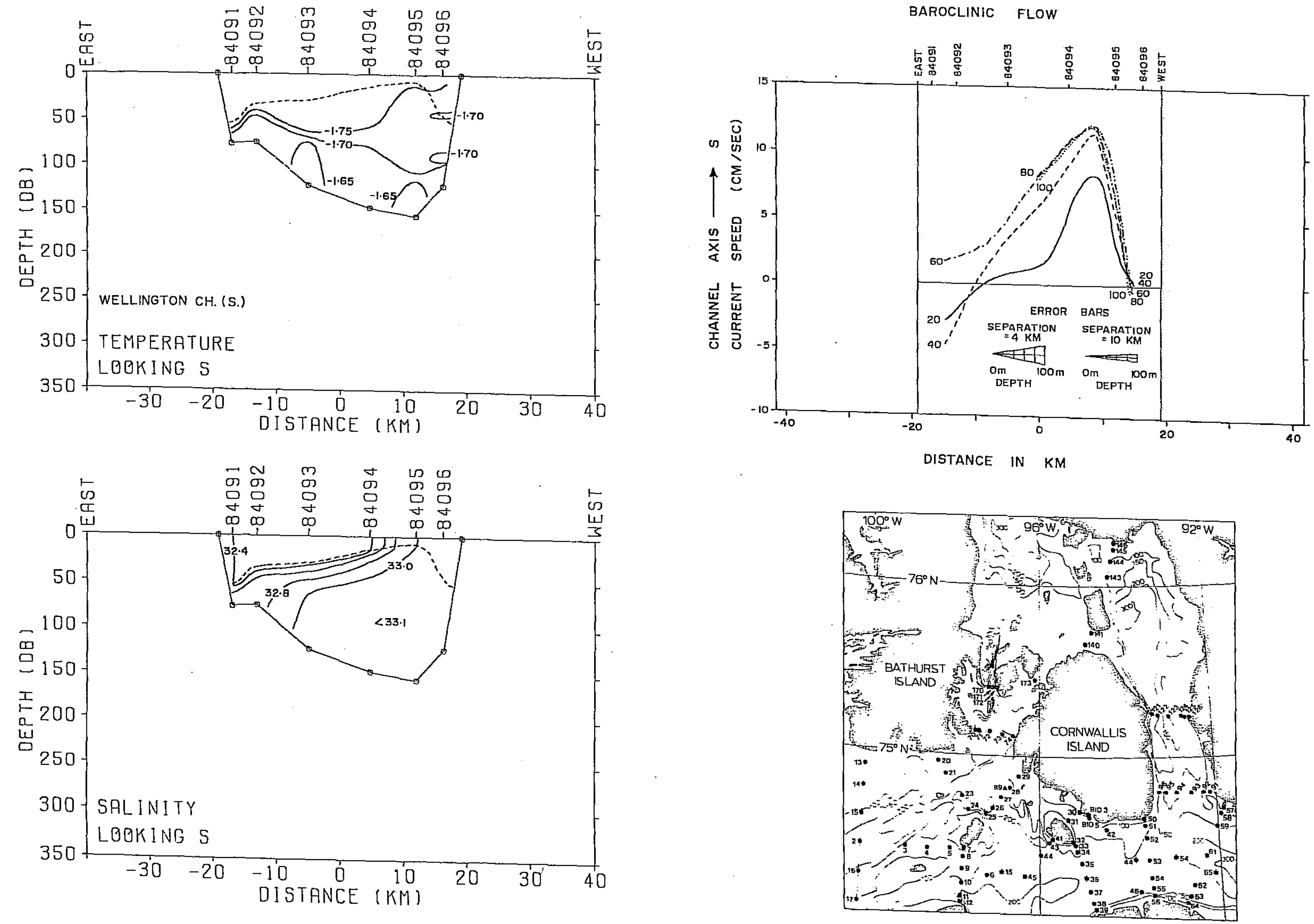

Figure 2.40 Temperature and salinity section and baroclinic velocity across southern Wellington Channel at Innes Point, 1984.

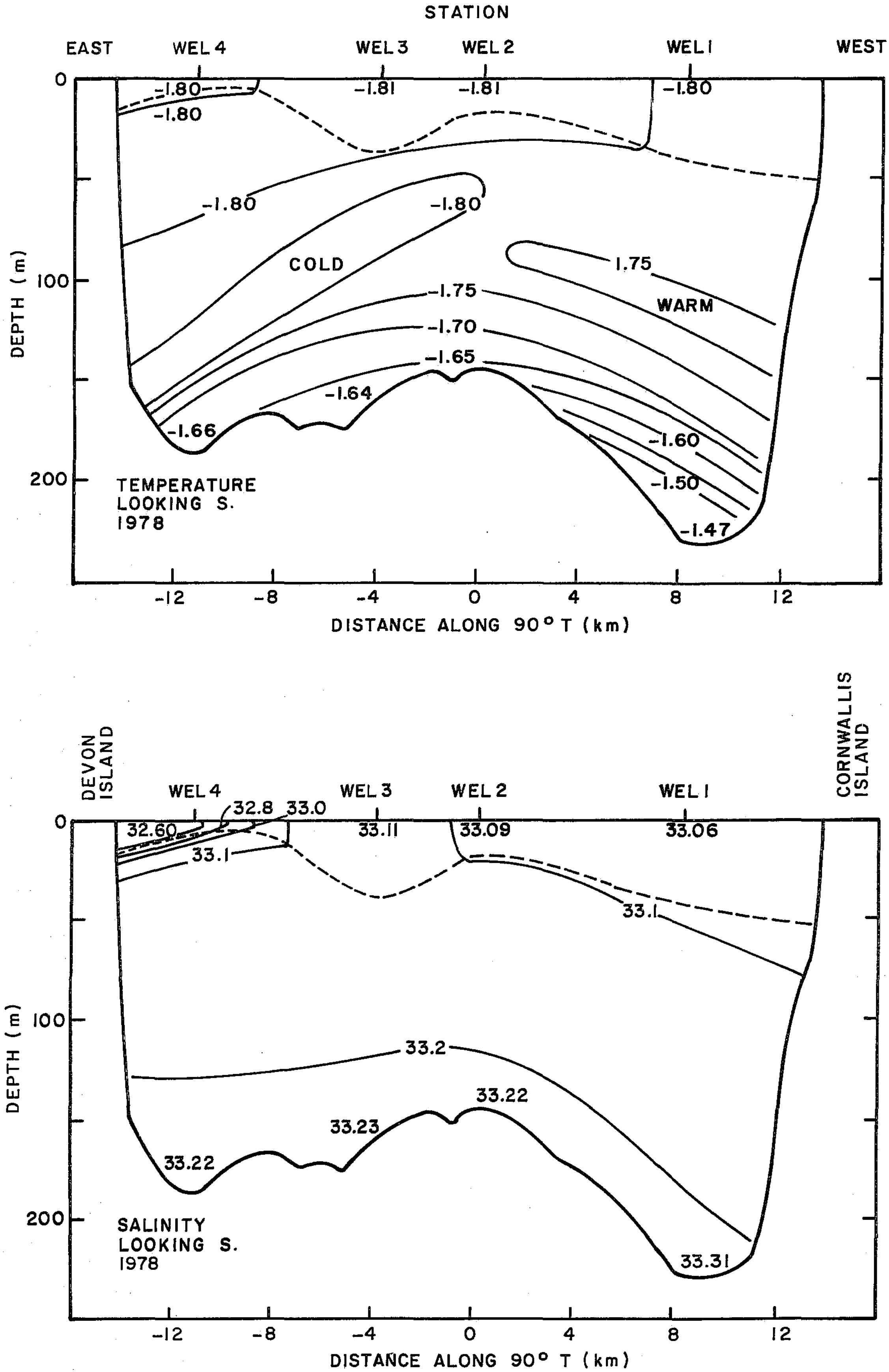

Figure 2.41 Temperature and salinity section across Wellington Channel (same location as Fig. 2.39) in late winter, 1978 (unpublished data, R.A. Lake).

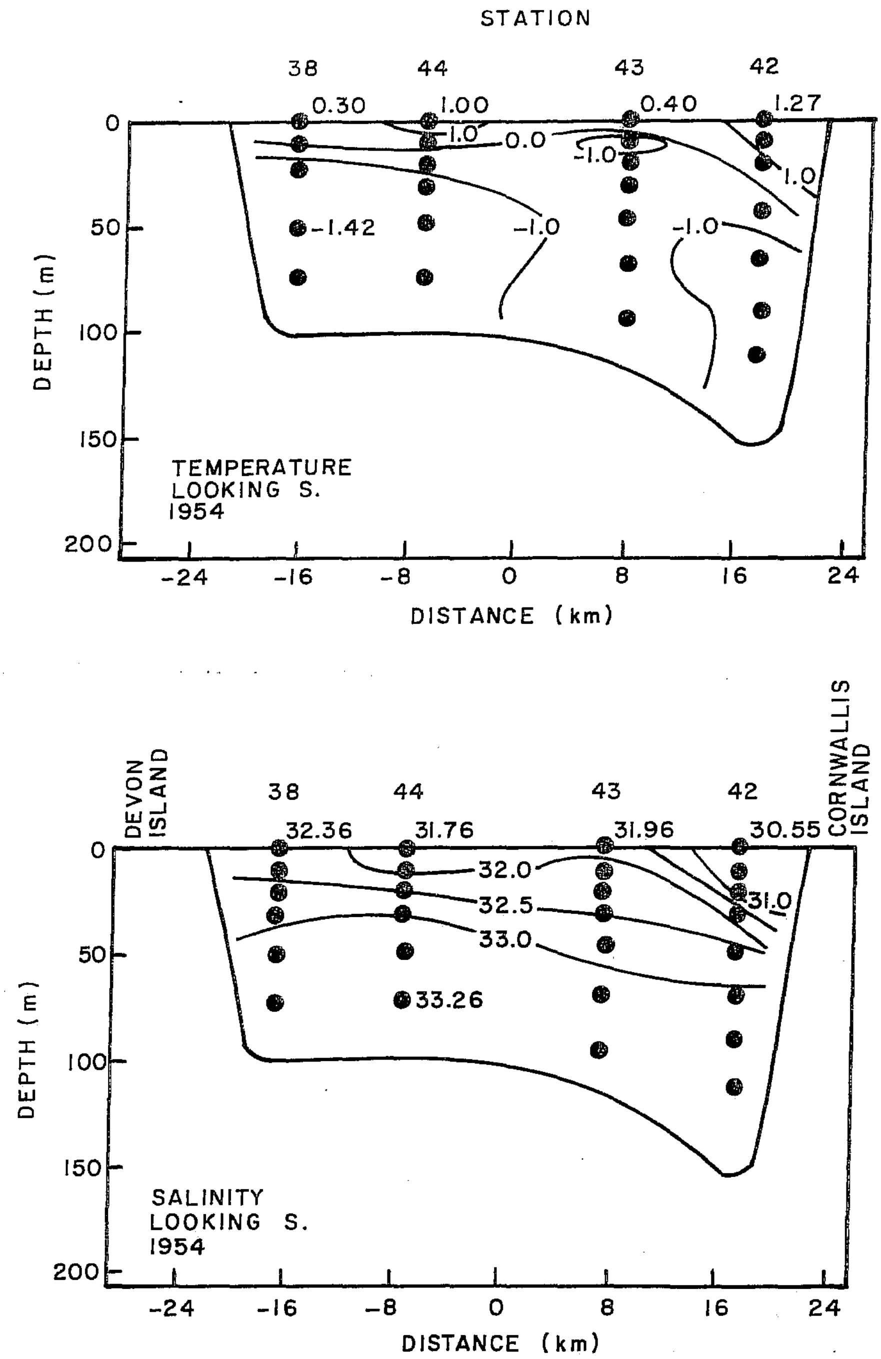

Figure 2.42 Temperature and salinity section across Wellington Channel (close south of the section in Fig. 2.40), summer 1954 (after Bailey, 1957).

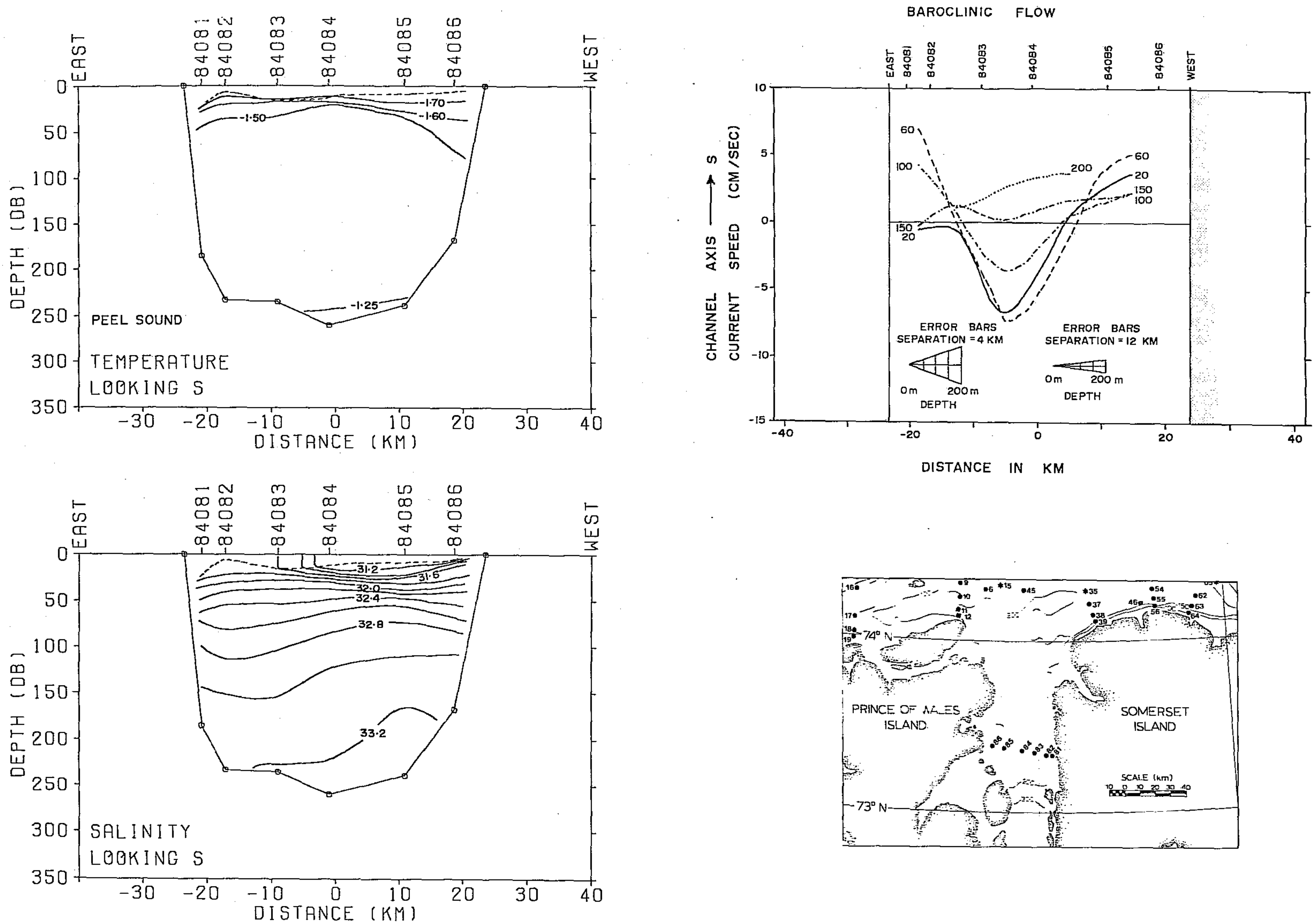

Figure 2.43 Temperature and salinity section and baroclinic velocity across northern Peel Sound (Sherard Head, Prince of Wales Is. to Somerset Is.), 1984.

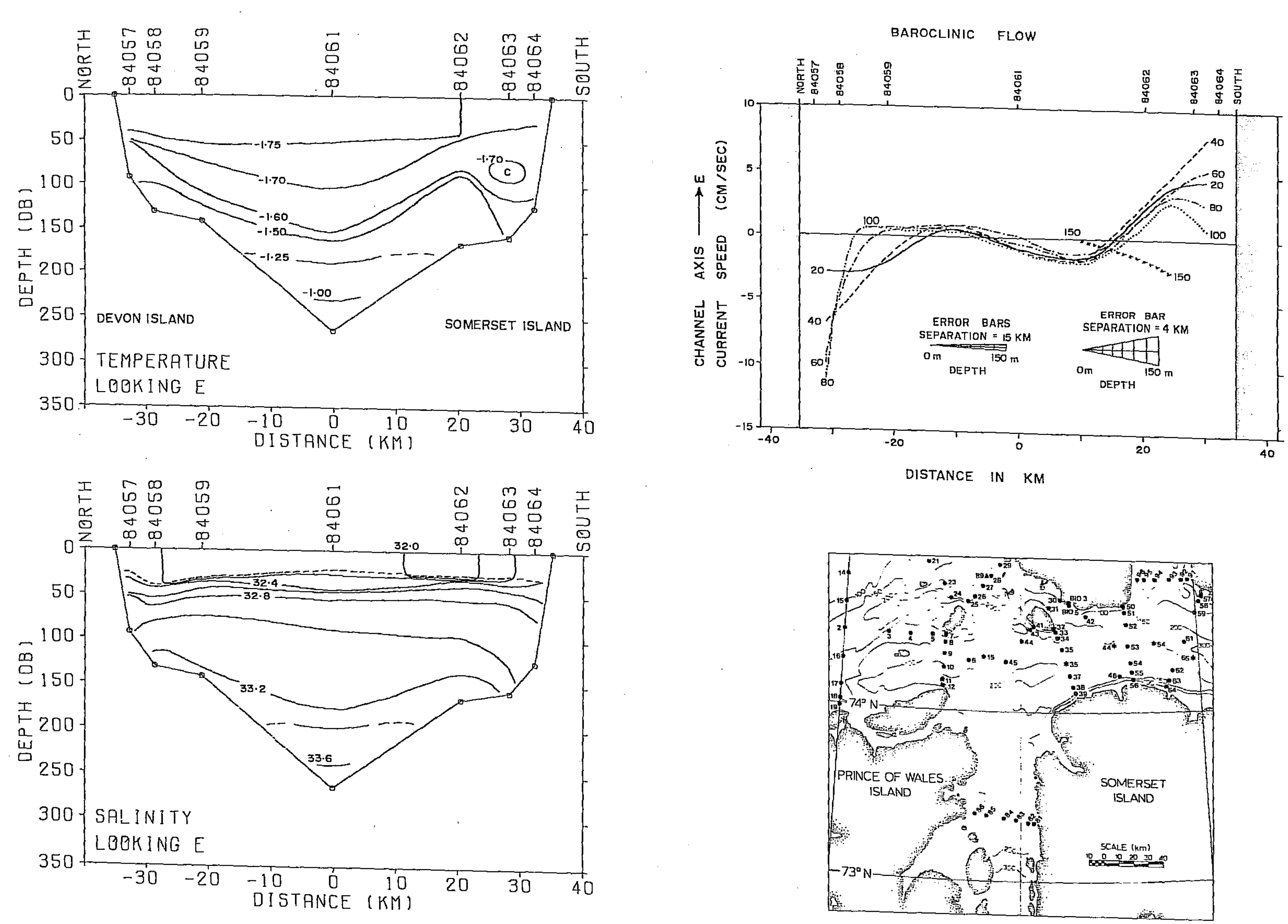

Figure 2.44 Temperature and salinity section and baroclinic velocity across eastern Barrow Strait west of Garnier Bay, 1984.

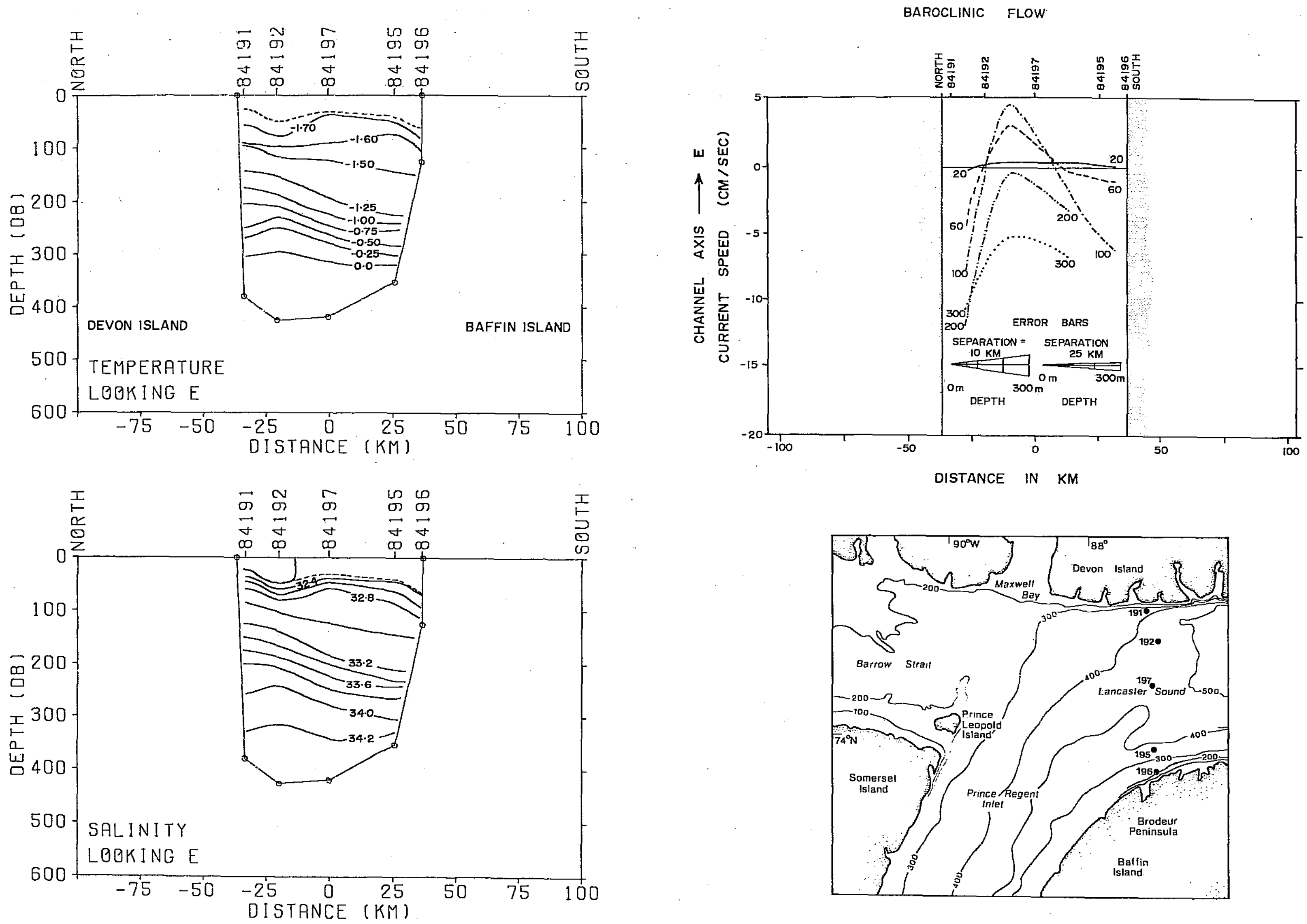

Figure 2.45 Temperature and salinity section and baroclinic velocity across western Lancaster Sound at Hobhouse Inlet, 1984.

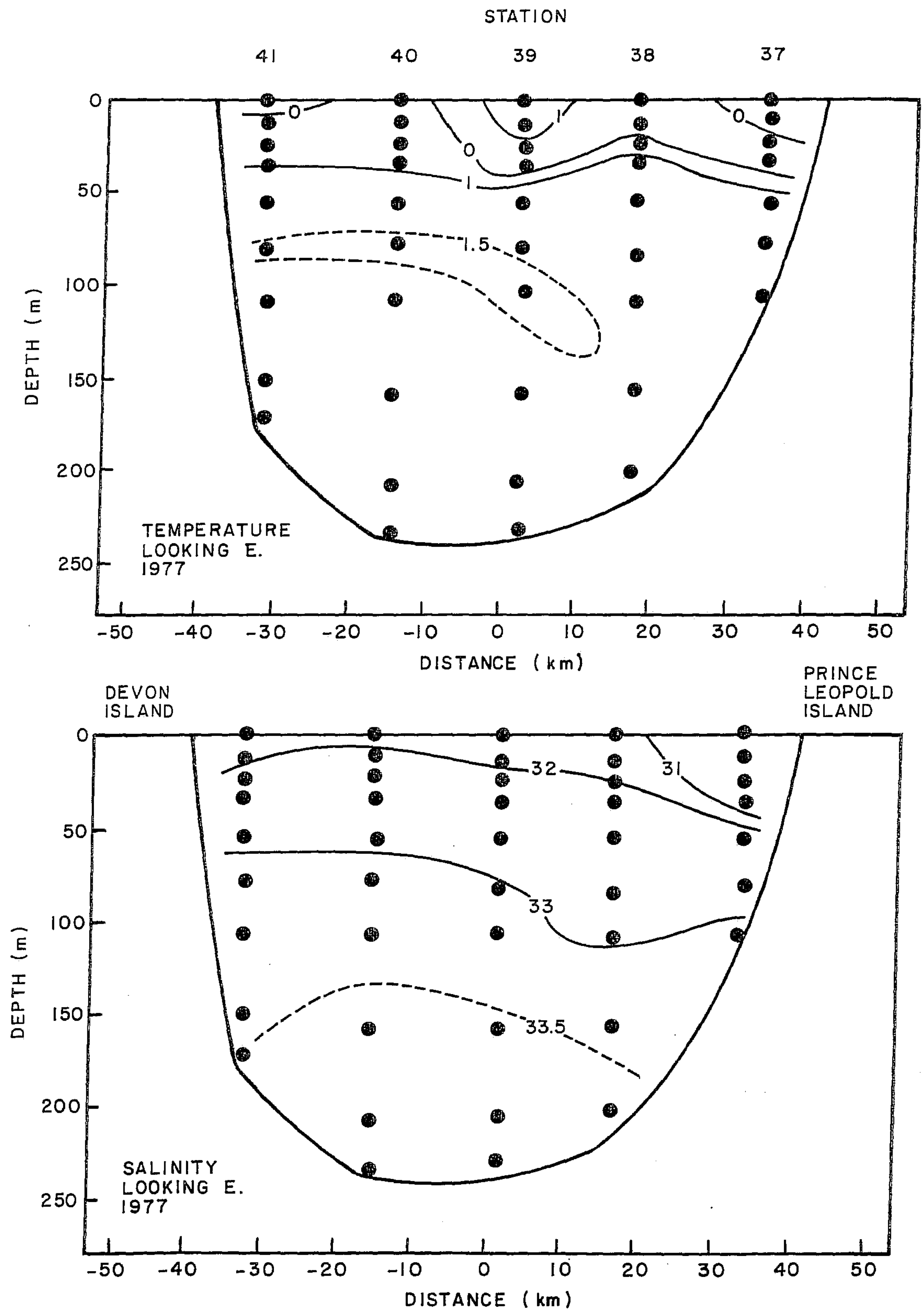

Figure 2.46 Temperature and salinity section across eastern Barrow Strait at Prince Leopold Island, summer 1977 (after Jones and Coote, 1980).

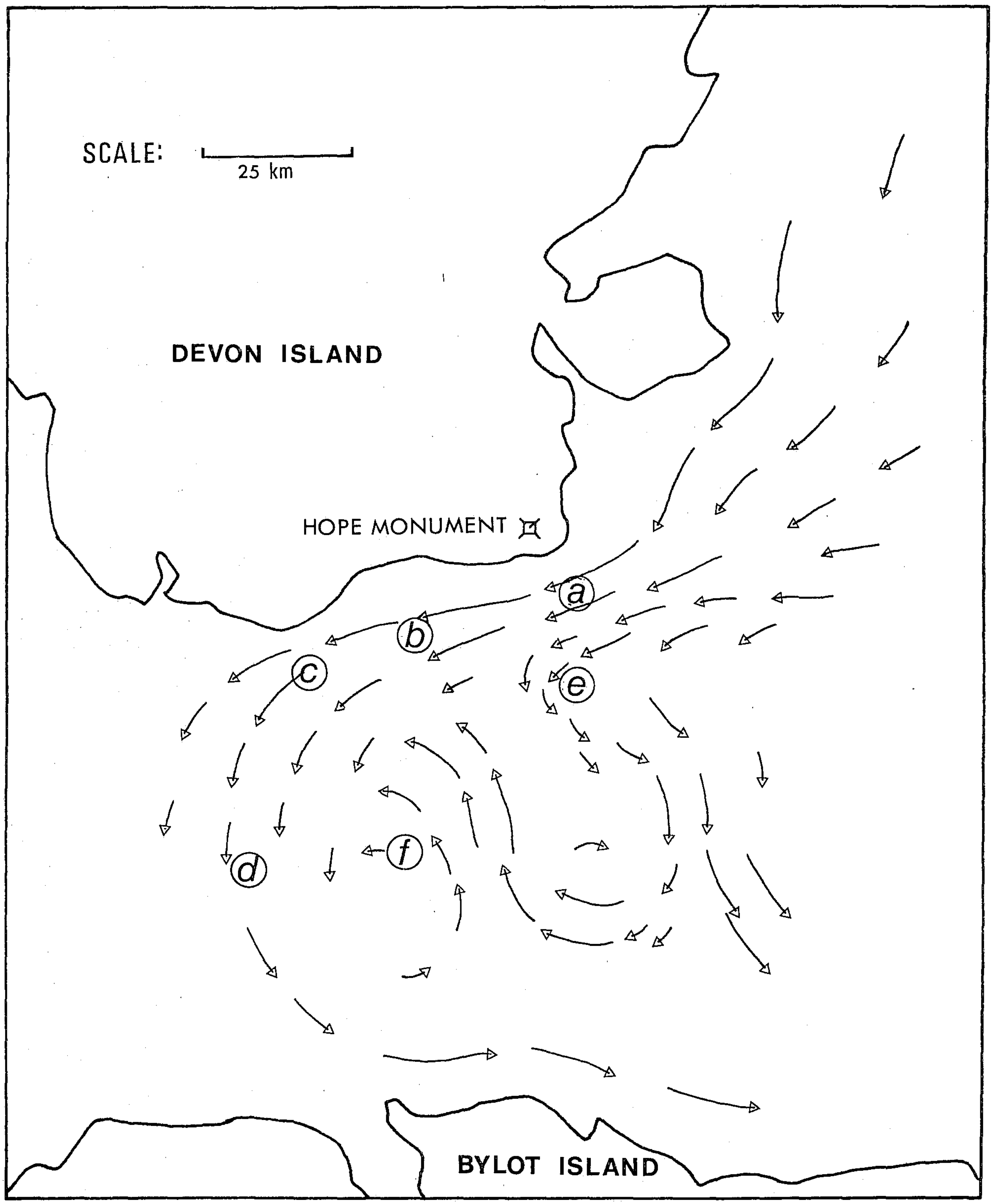

Figure 2.47 The general circulation pattern of icebergs as inferred from observations made during July to September, 1978 (from de Lange Boom et al., 1982).

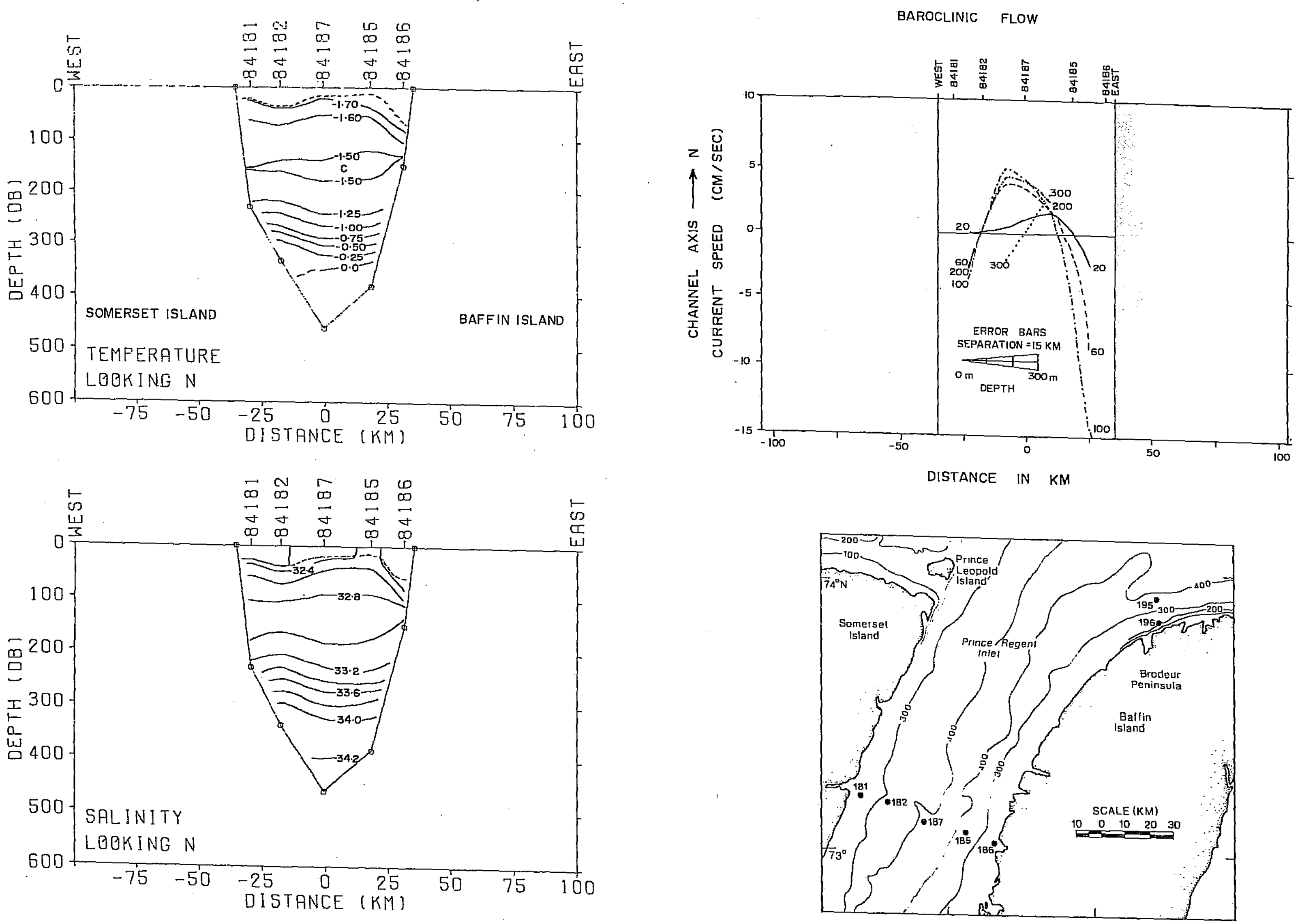

Figure 2.48 Temperature and salinity section and baroclinic velocity across northern Prince Regent Inlet at Batty Bay, 1984.

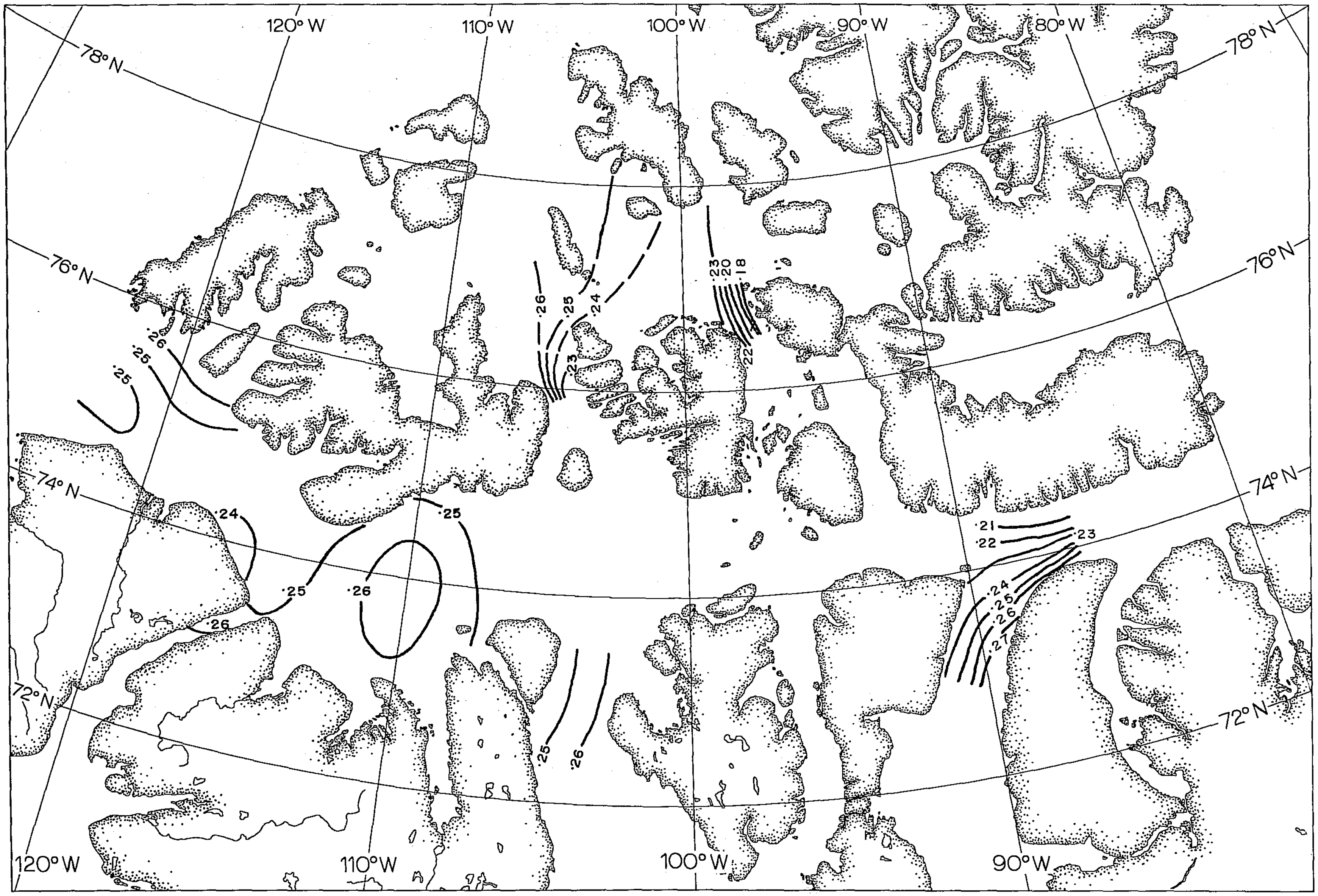

Figure 2.49 Dynamic Height Anomaly 0/150 dbar (dynamic m), 1982.

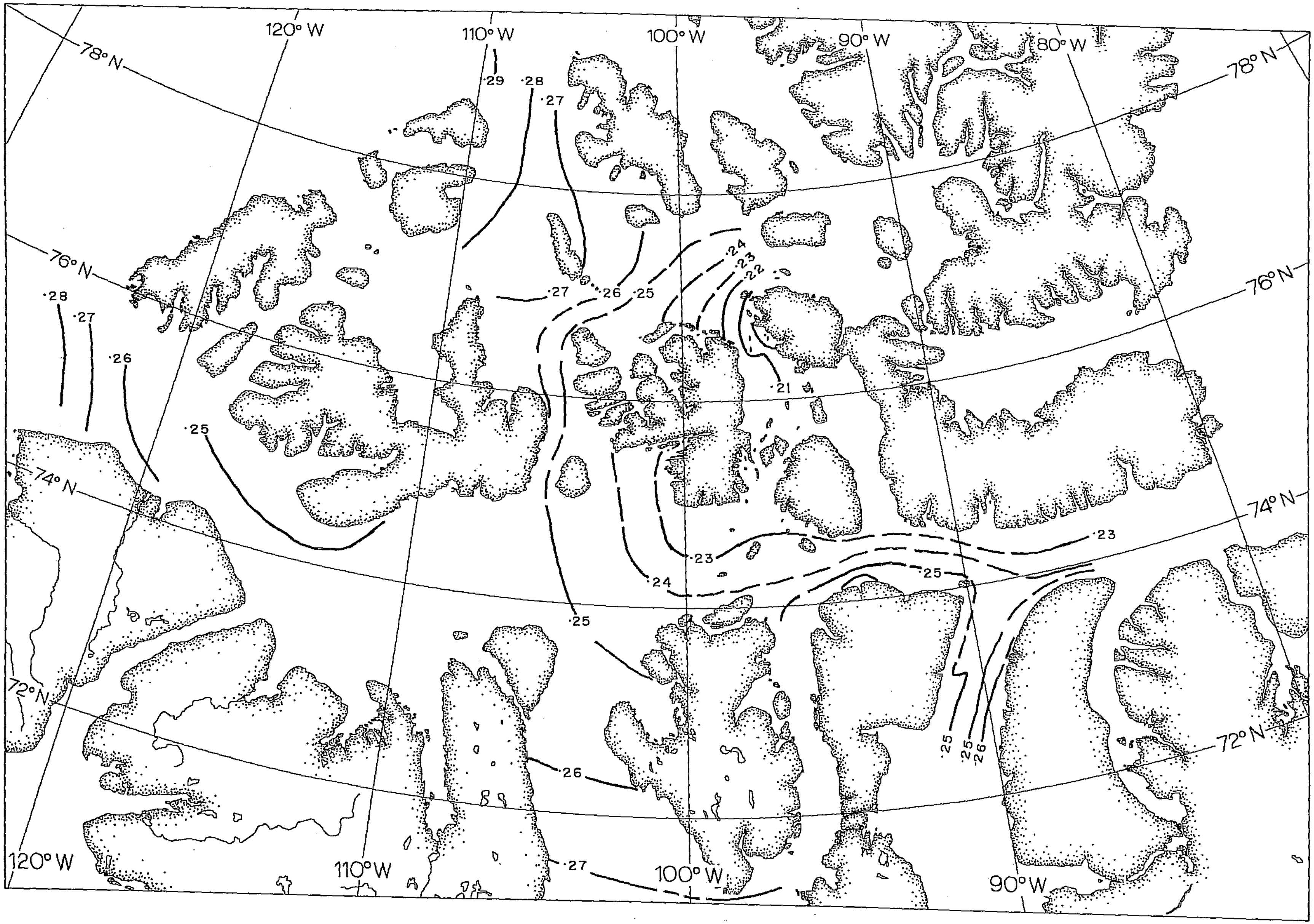

Figure 2.50 Dynamic Height Anomaly 0/150 dbar (dynamic m), 1983.

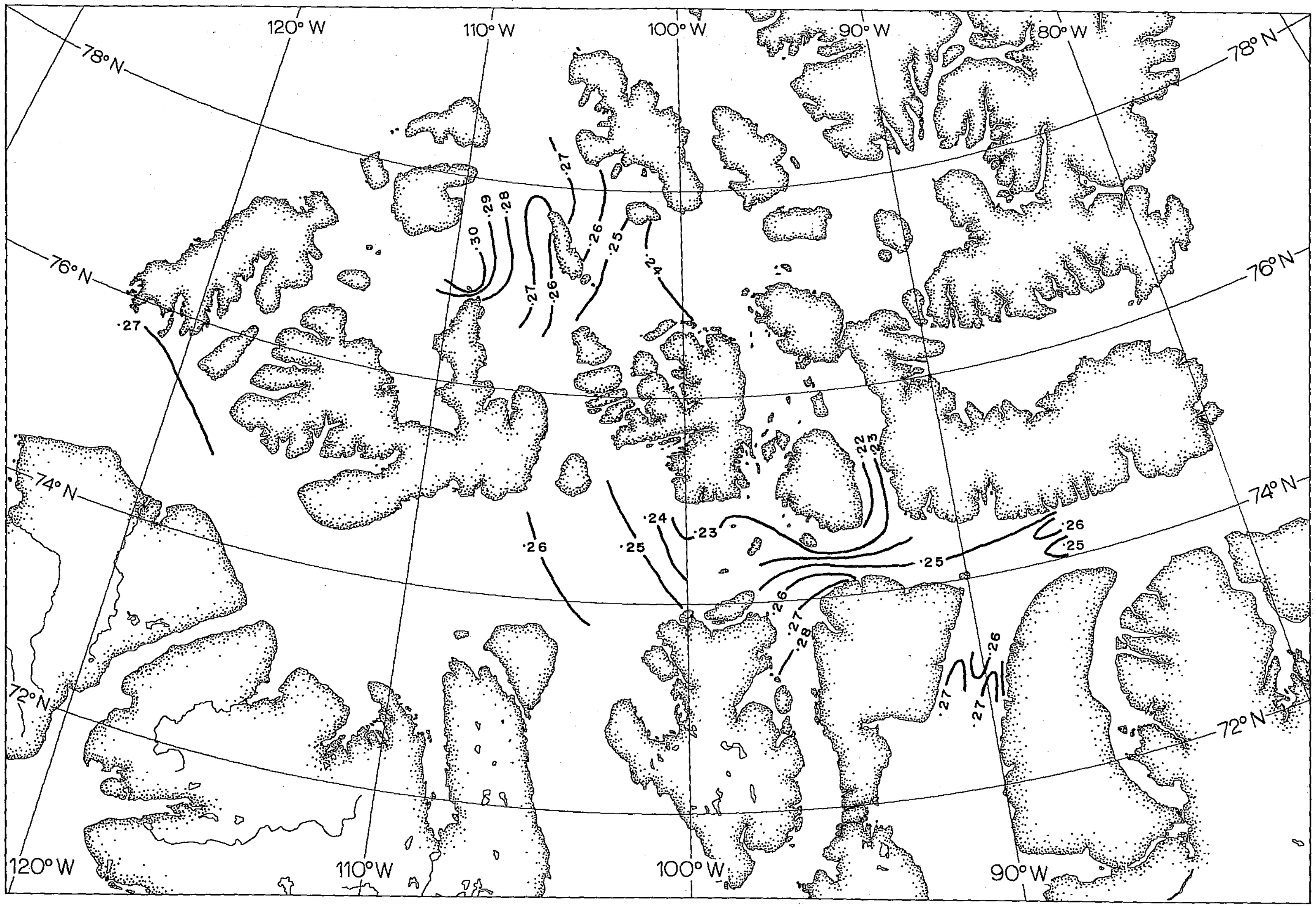

Figure 2.51 Dynamic Height Anomaly 0/150 dbar (dynamic m), 1984.

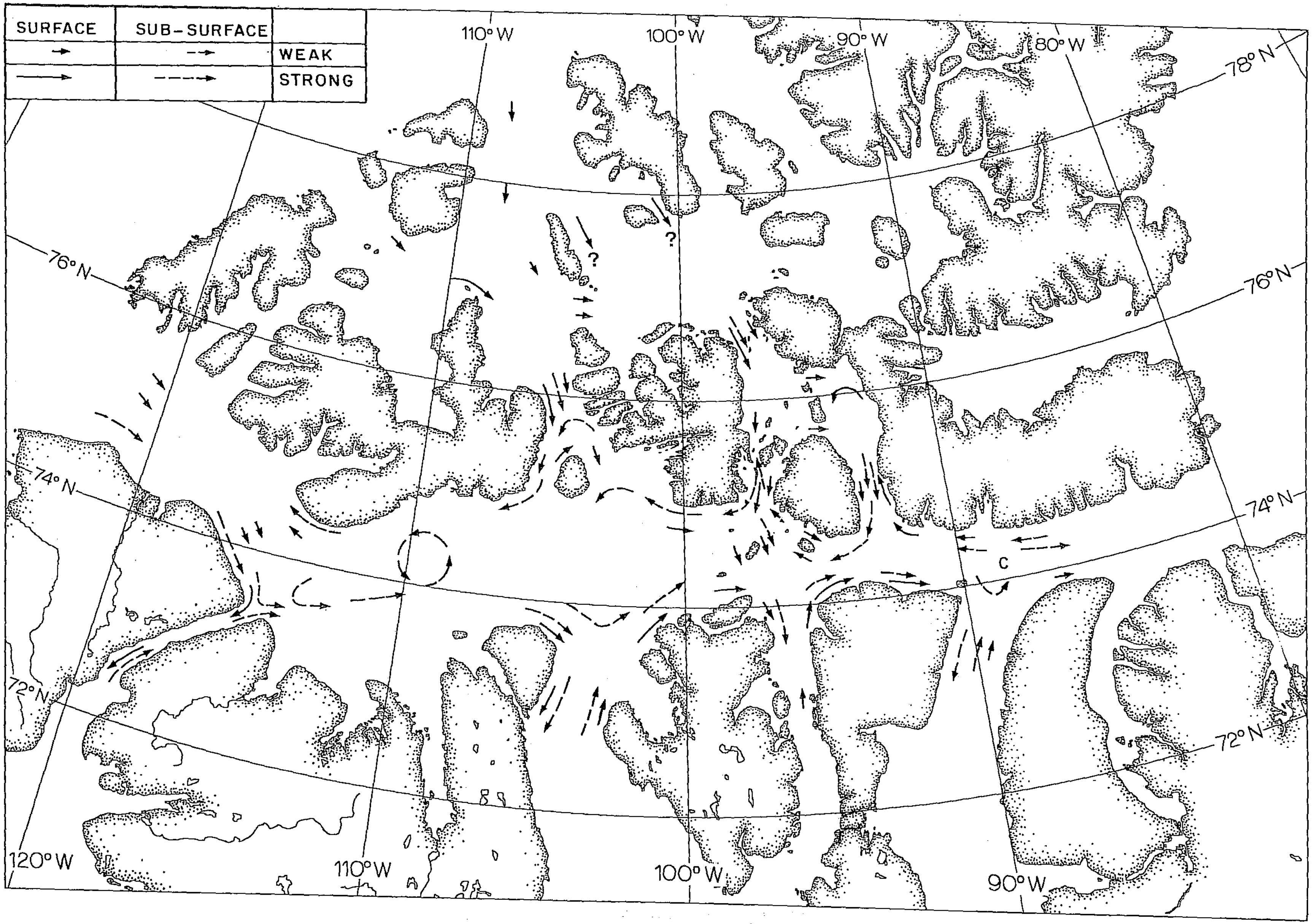

Figure 2.52 The general circulation as inferred for the study area. The 'C' denotes a region of variable, confused motion.

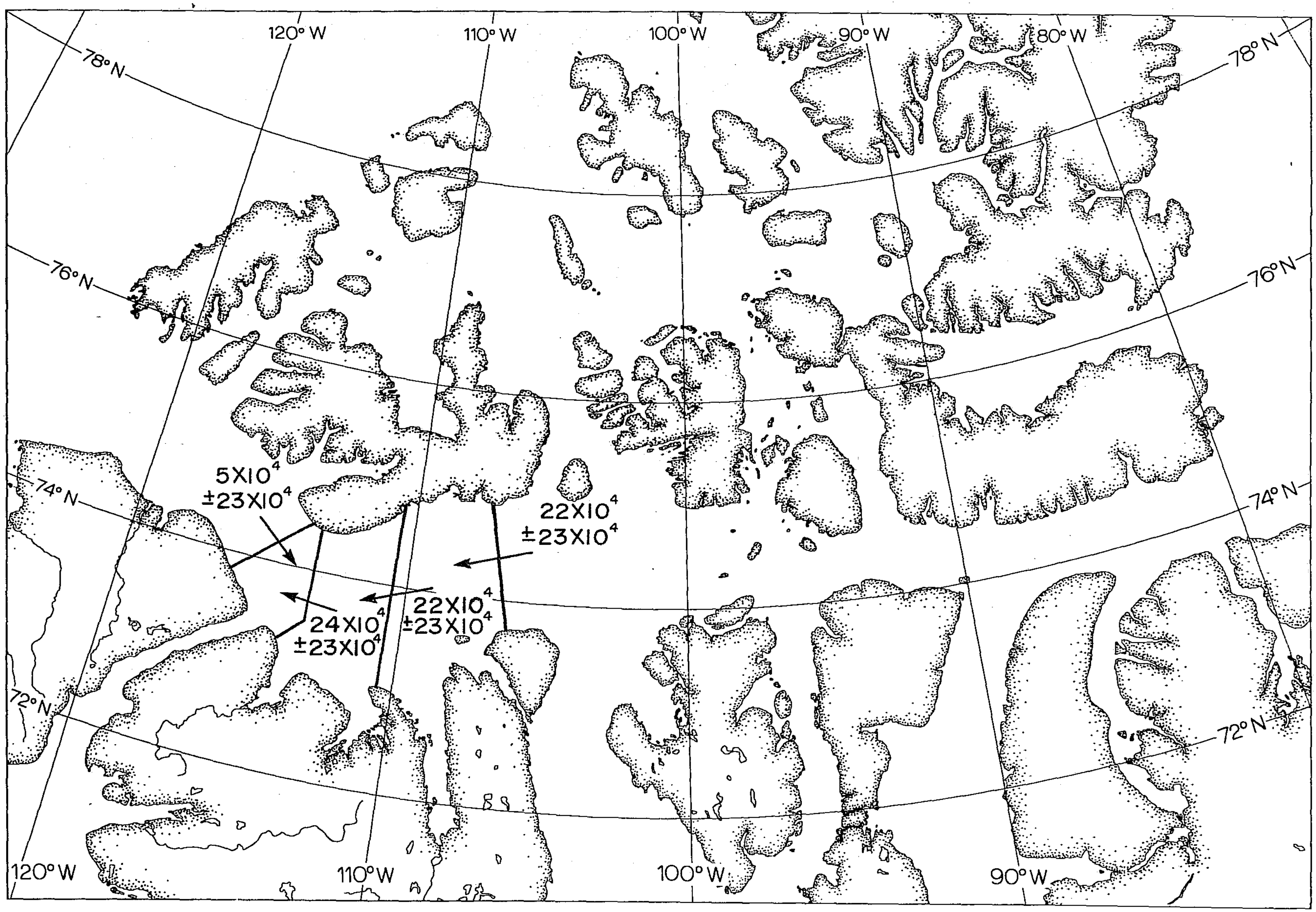

Figure 2.53 Volume transport (m^3/s) and uncertainty estimates for sections occupied in 1982.

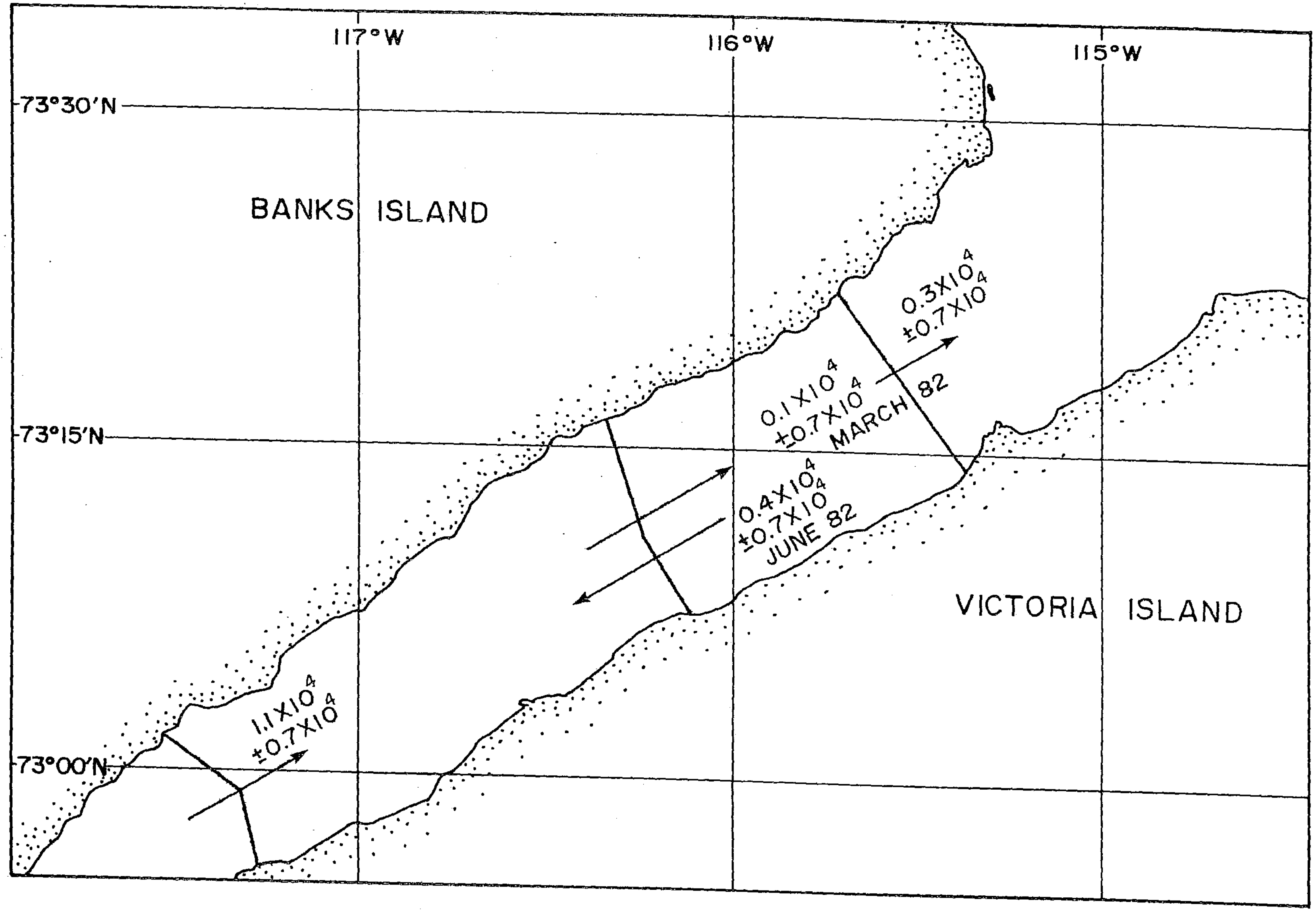

Figure 2.54 Volume transport (m^3/s) and uncertainty estimates for sections in northern Prince of Wales Strait, 1982.

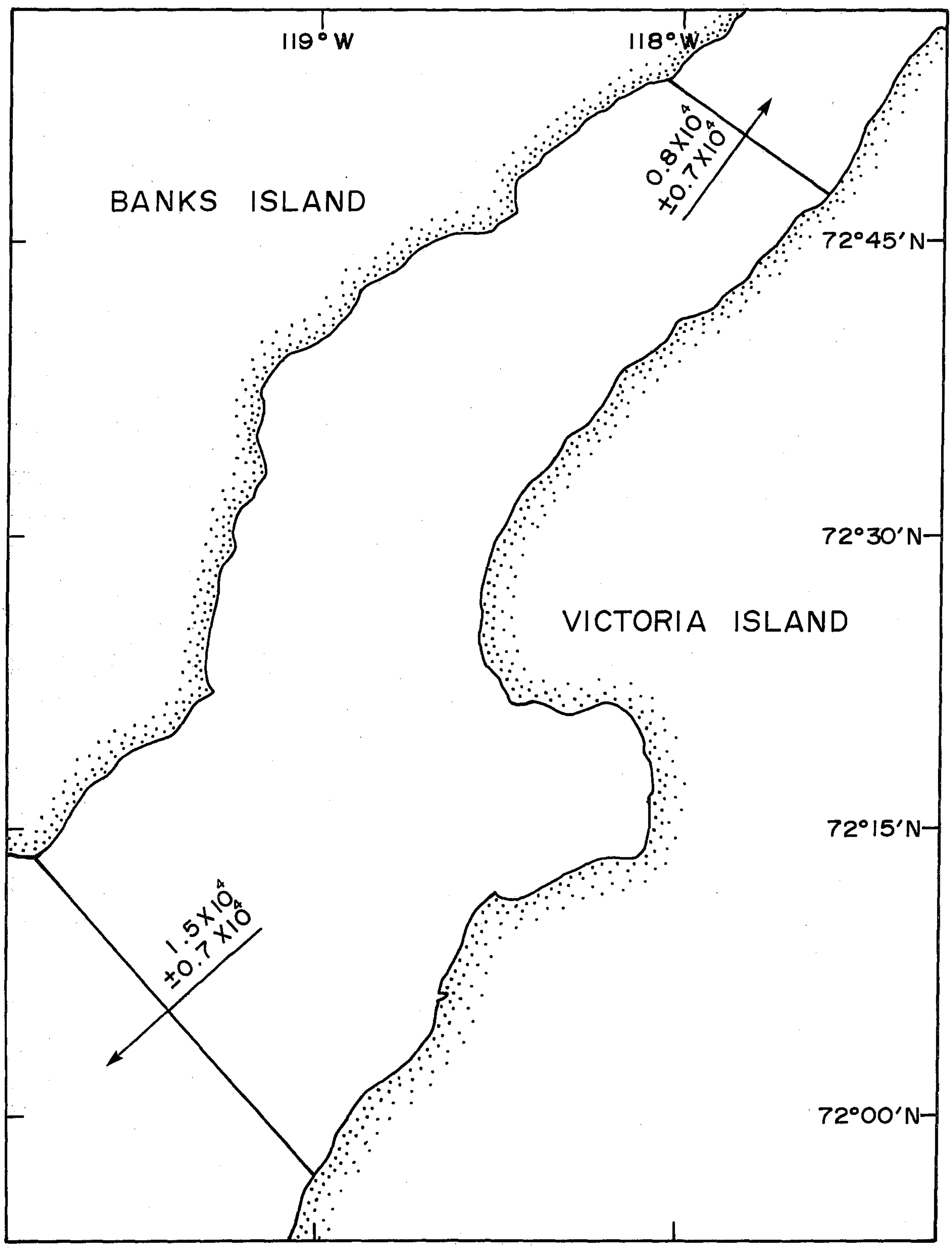

Figure 2.55 Volume transport (m^3/s) and uncertainty estimates for sections in southern Prince of Wales Strait, 1982.

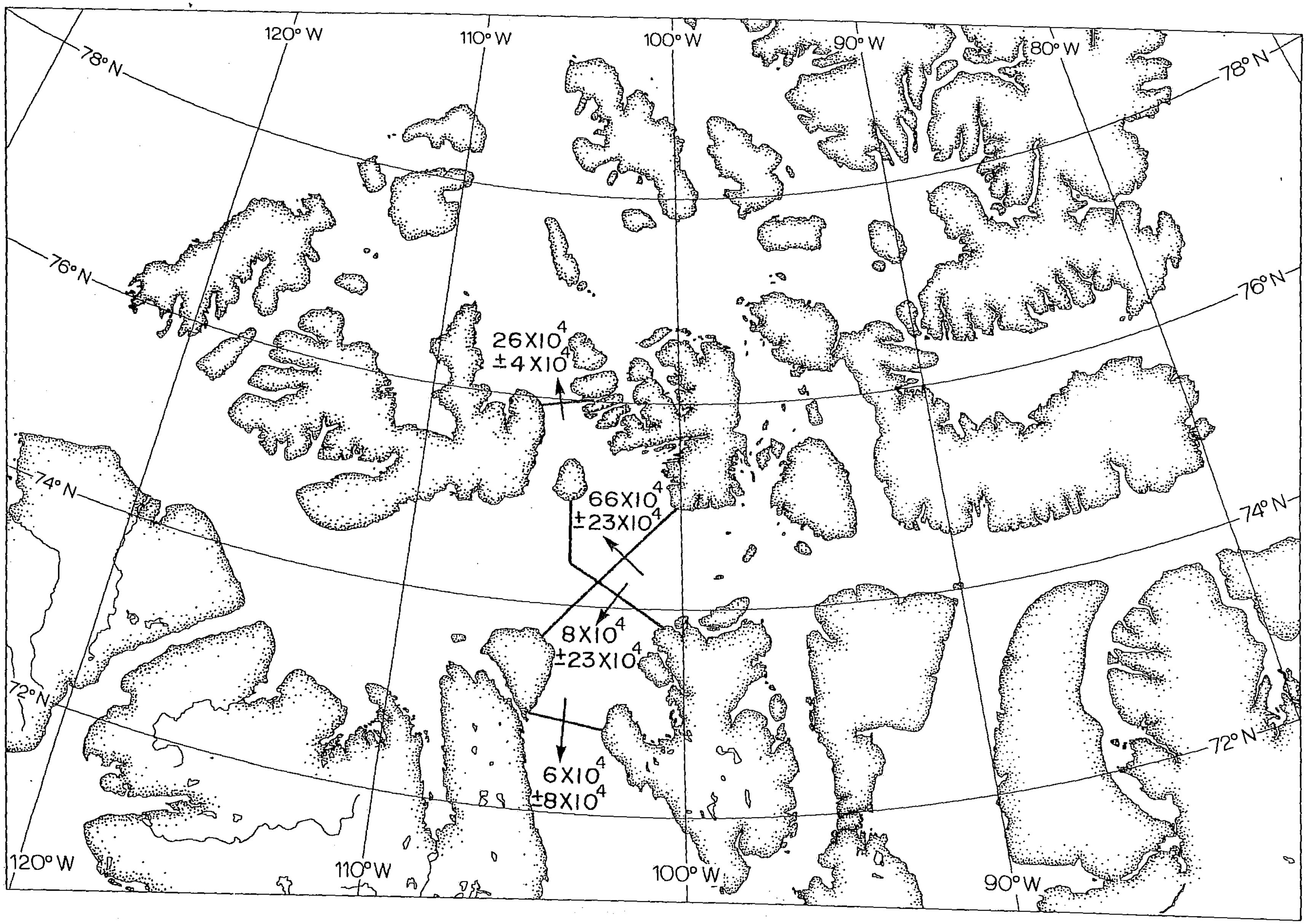

Figure 2.56 Volume transport (m^3/s) and uncertainty estimates for sections occupied in 1983.

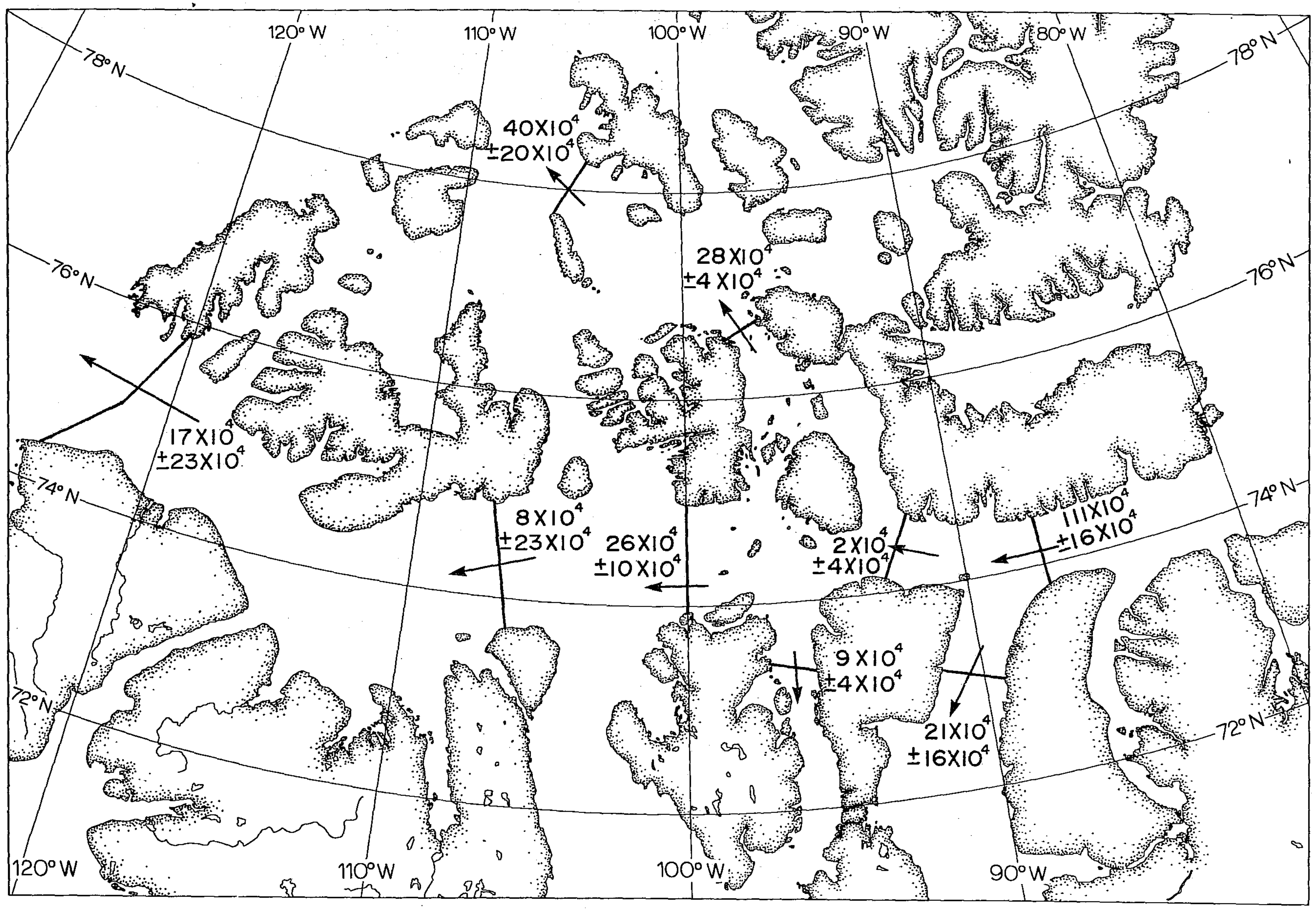

Figure 2.57 Volume transport (m^3/s) and uncertainty estimates for sections occupied in 1984.

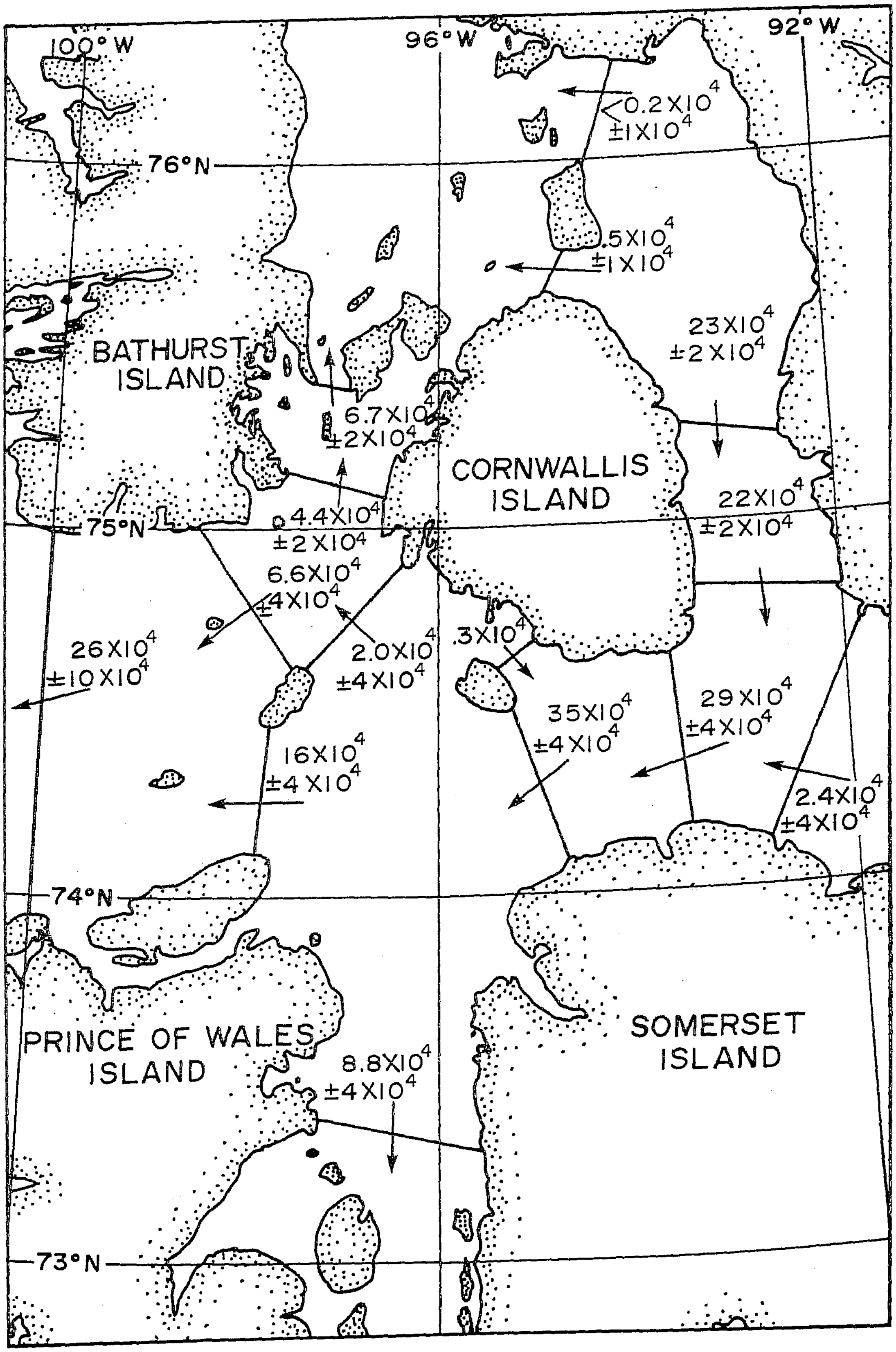

Figure 2.58 Volume transport (m^3/s) and uncertainty estimates for sections in the central sills area, 1984 (m^3/s).

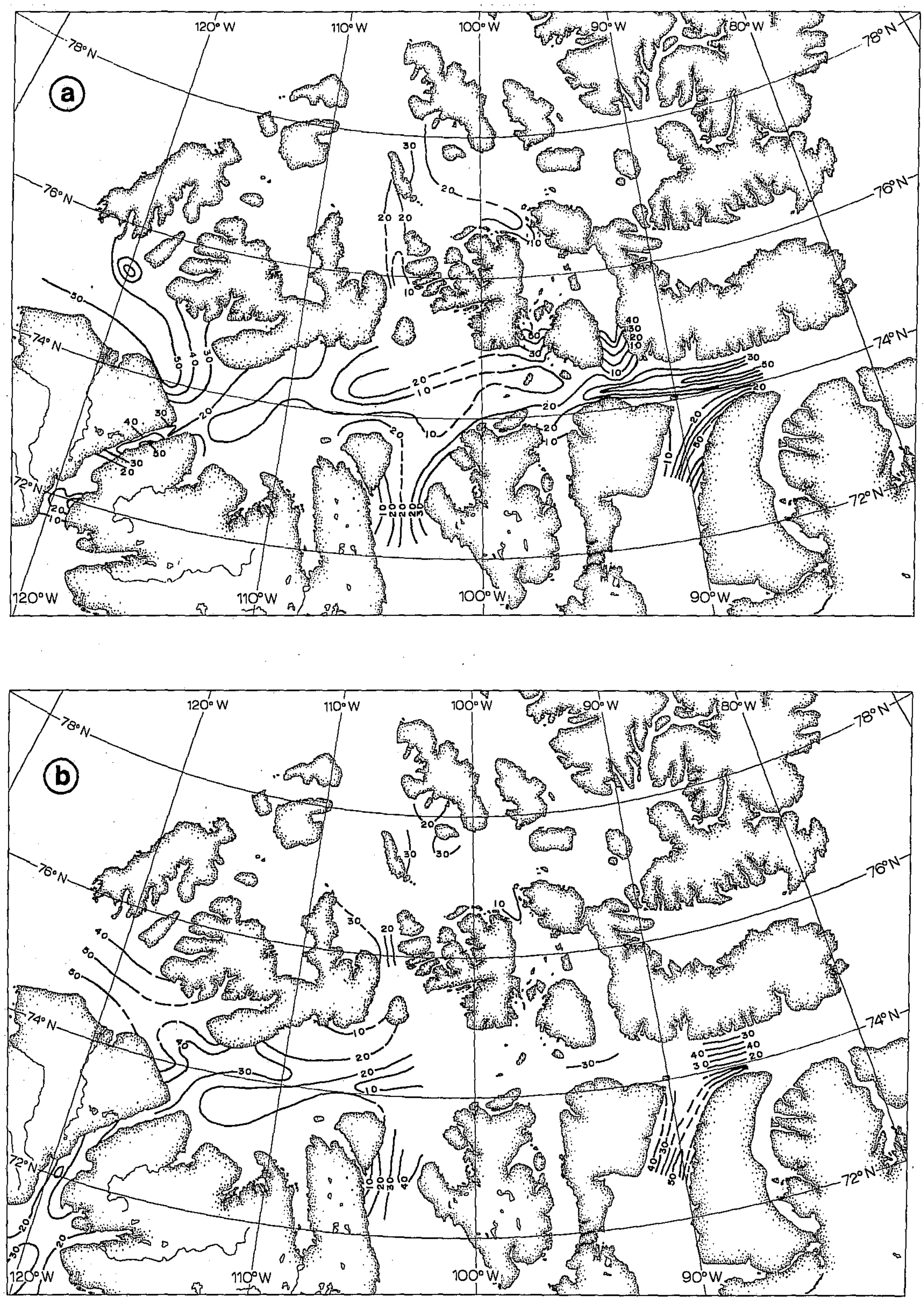

Figure 3.1 Thickness of the surface layer (dbar), 1982: a. depth at which sigma-t value exceeds the 5 dbar value by 0.02; b. depth at which the temperature differs from the 5 dbar value by 0.01 C degrees.

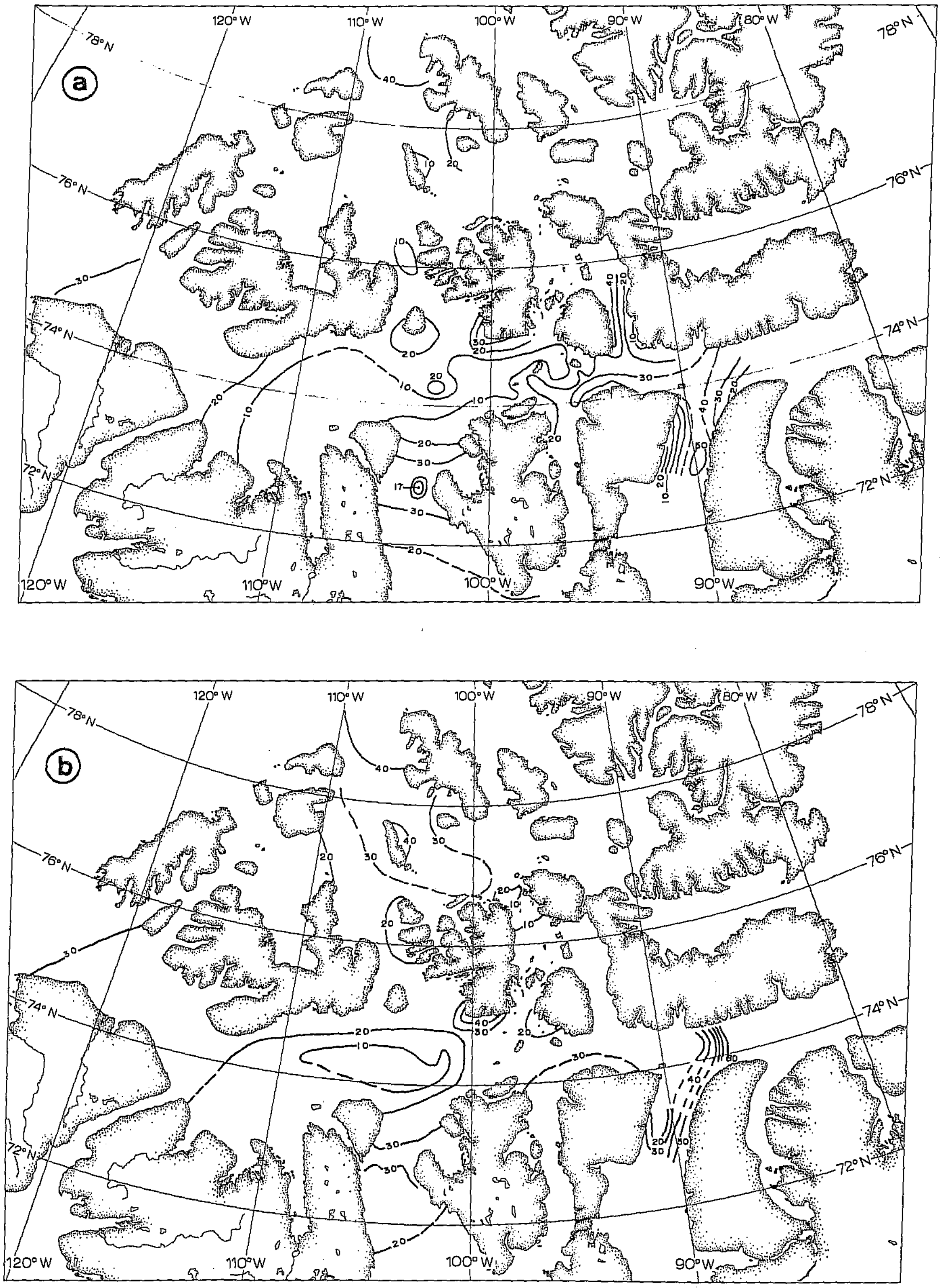

Figure 3.2 Thickness of the surface layer (dbar), 1983: a. density based, b. temperature based; see Figure 3.1 for details.

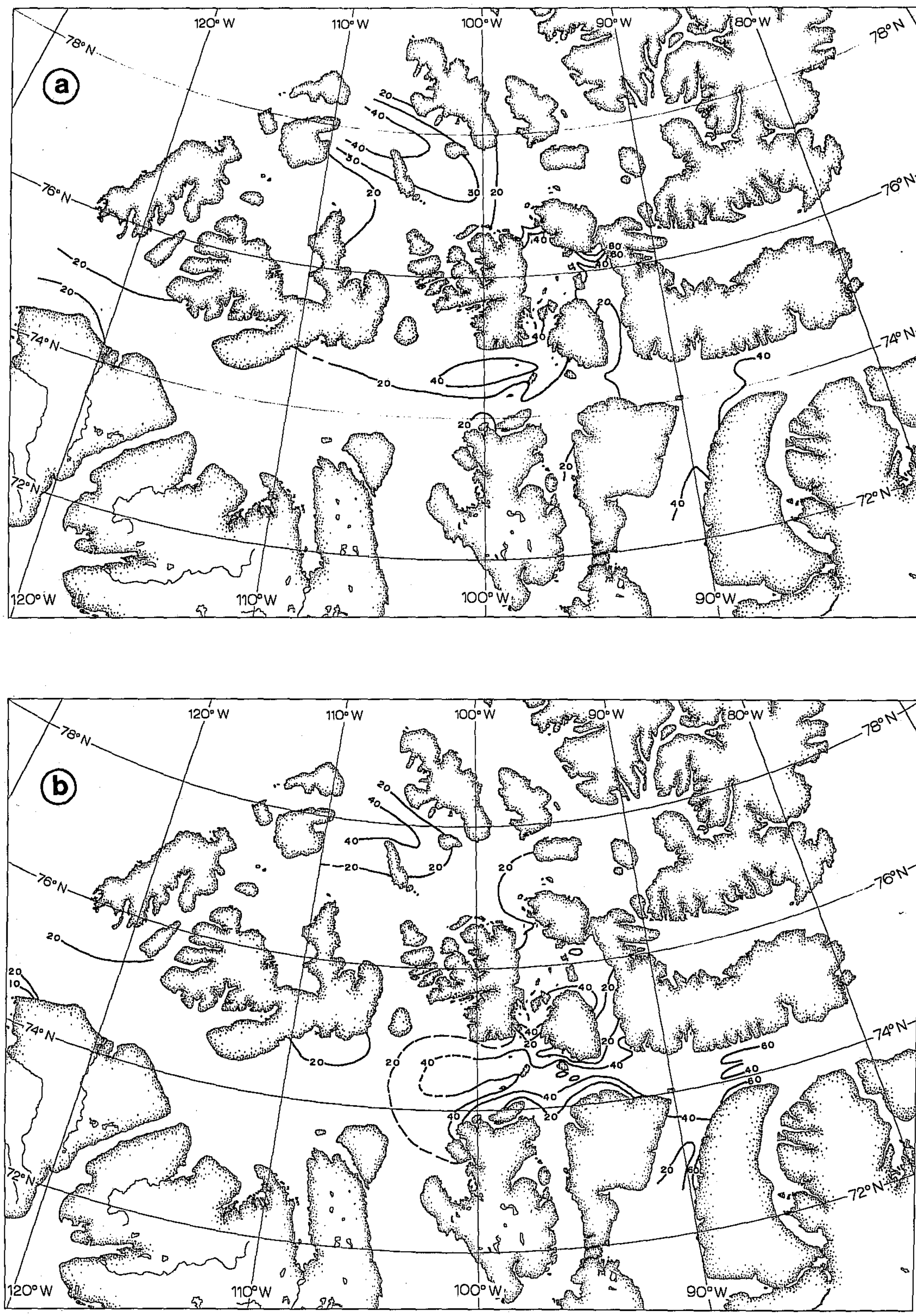

Figure 3.3 Thickness of the surface layer (dbar), 1984: a. density based, b. temperature based; see Figure 3.1 for details.

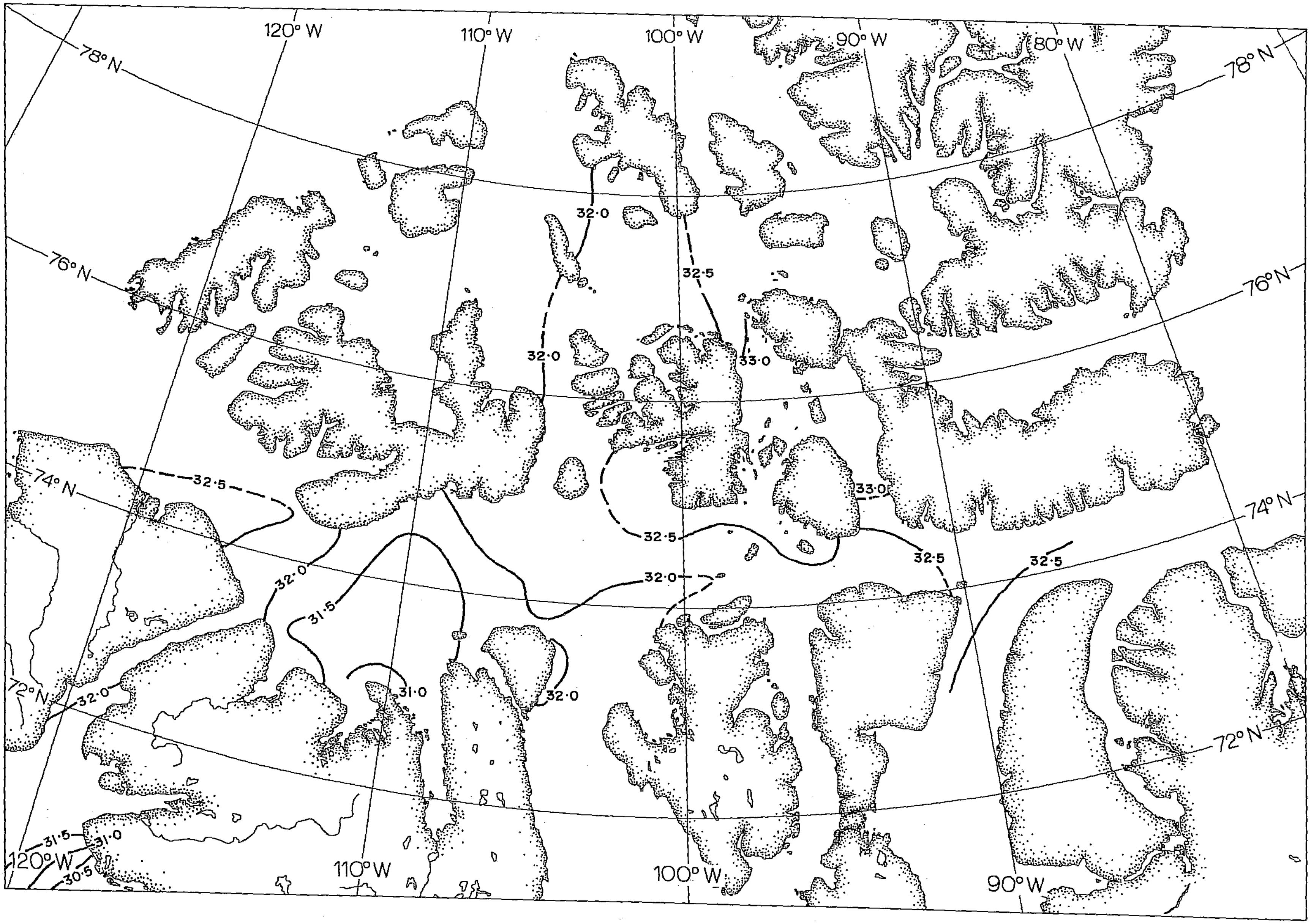

Figure 3.4 Salinity at 5 dbar pressure, 1982.

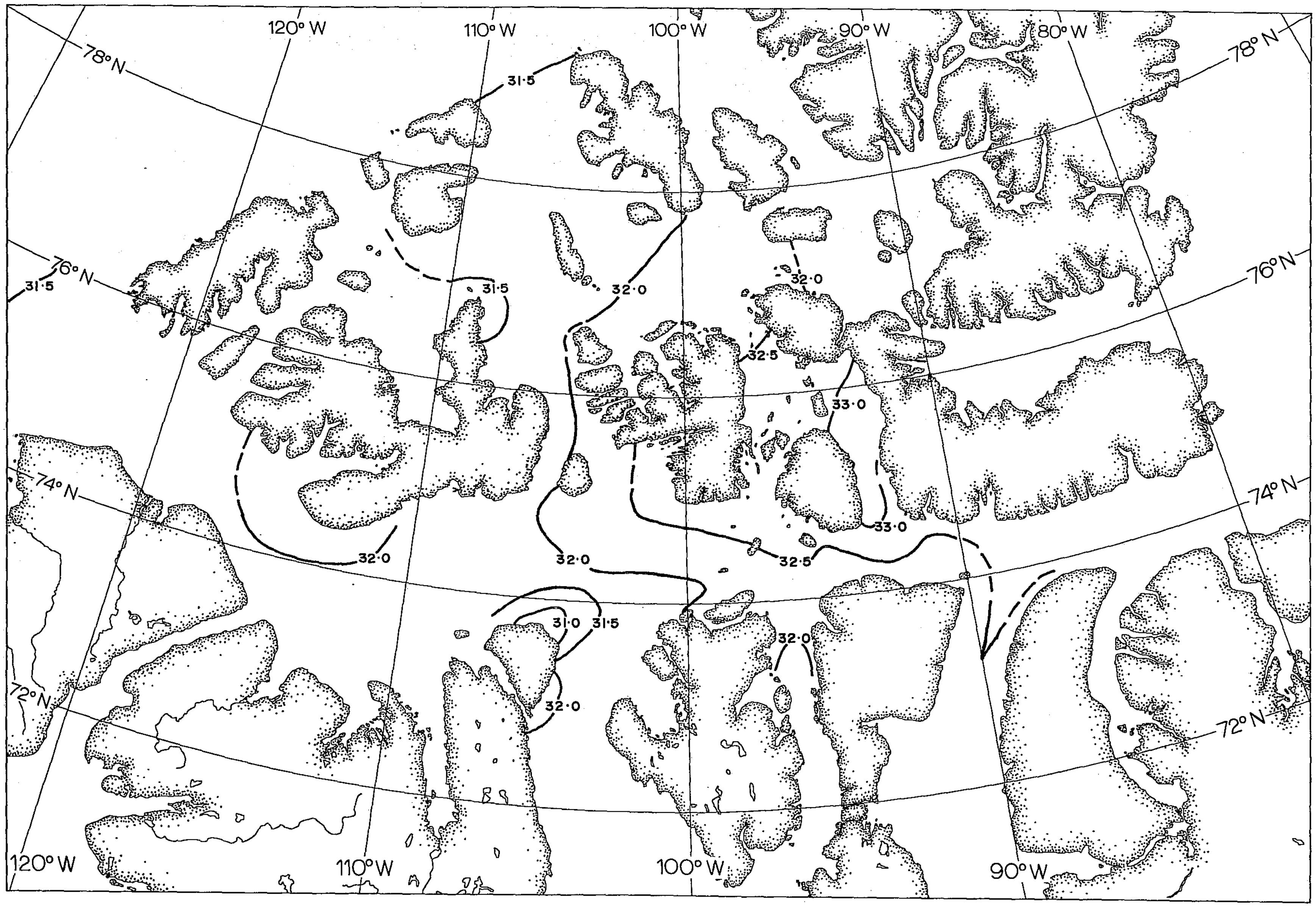

Figure 3.5 Salinity at 5 dbar pressure, 1983.

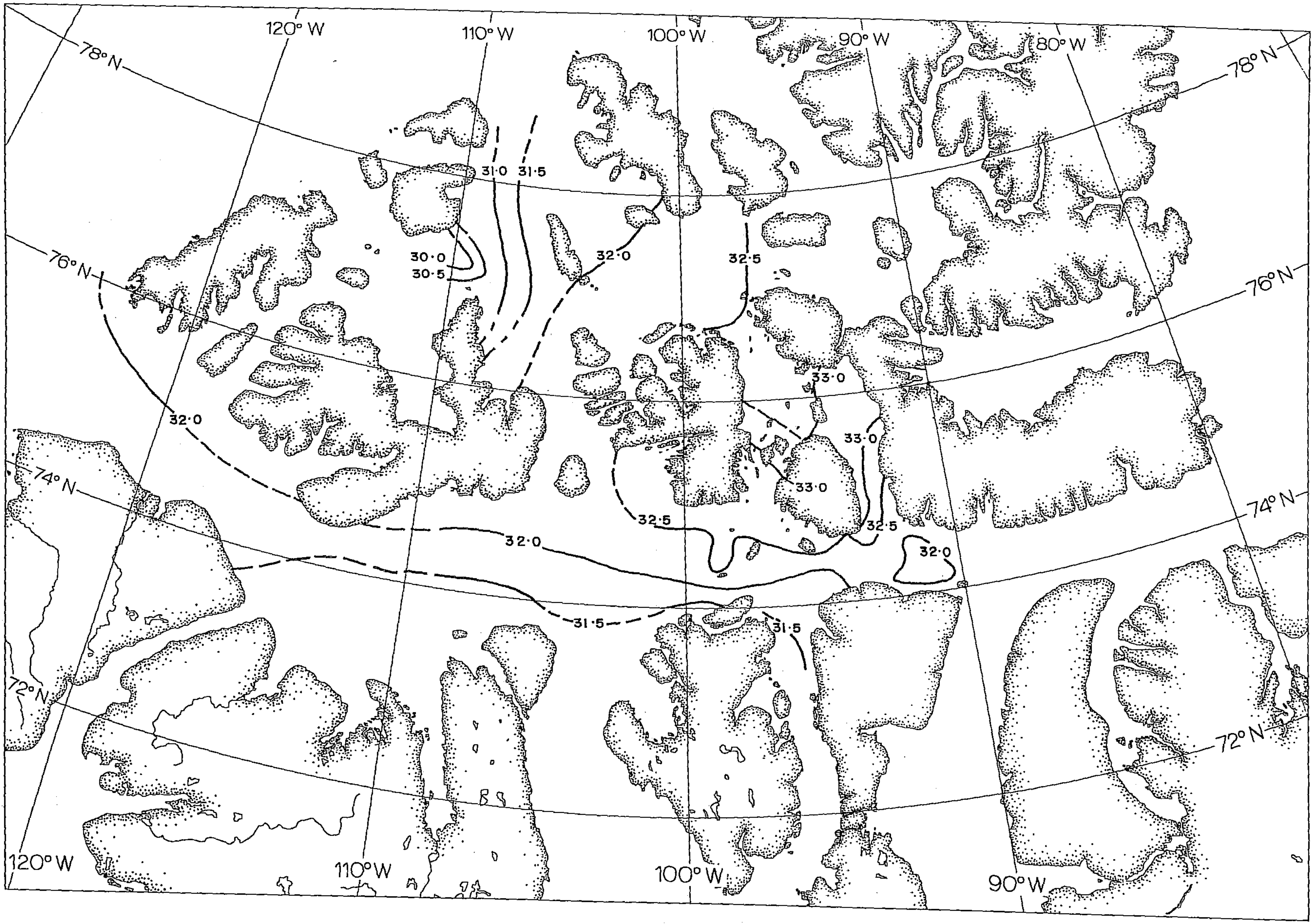

Figure 3.6 Salinity at 5 dbar pressure, 1984.

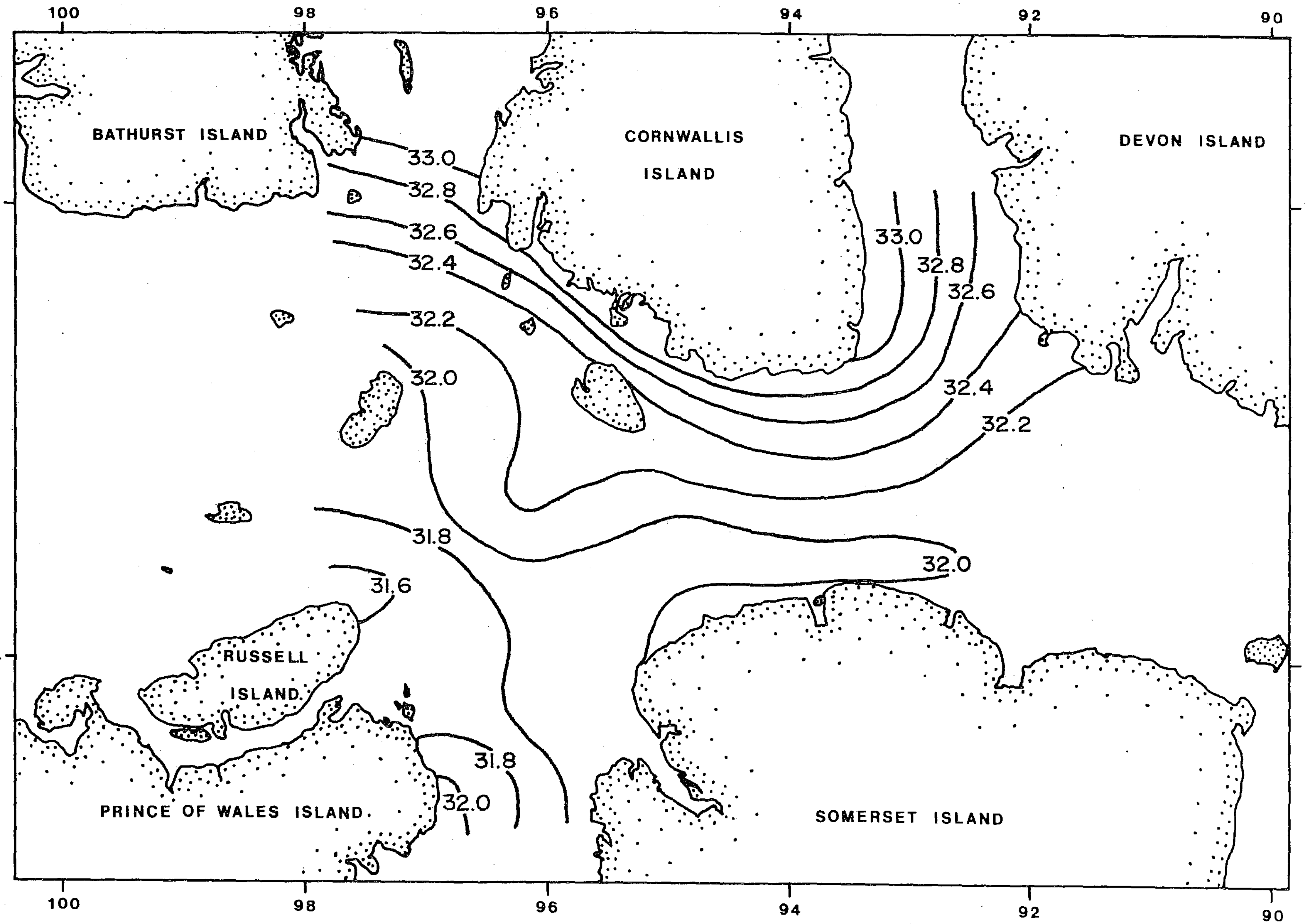

Figure 3.7 Salinity at 10 m, late winter 1981 (after Prinsenberg and Sosnoski, 1983a).

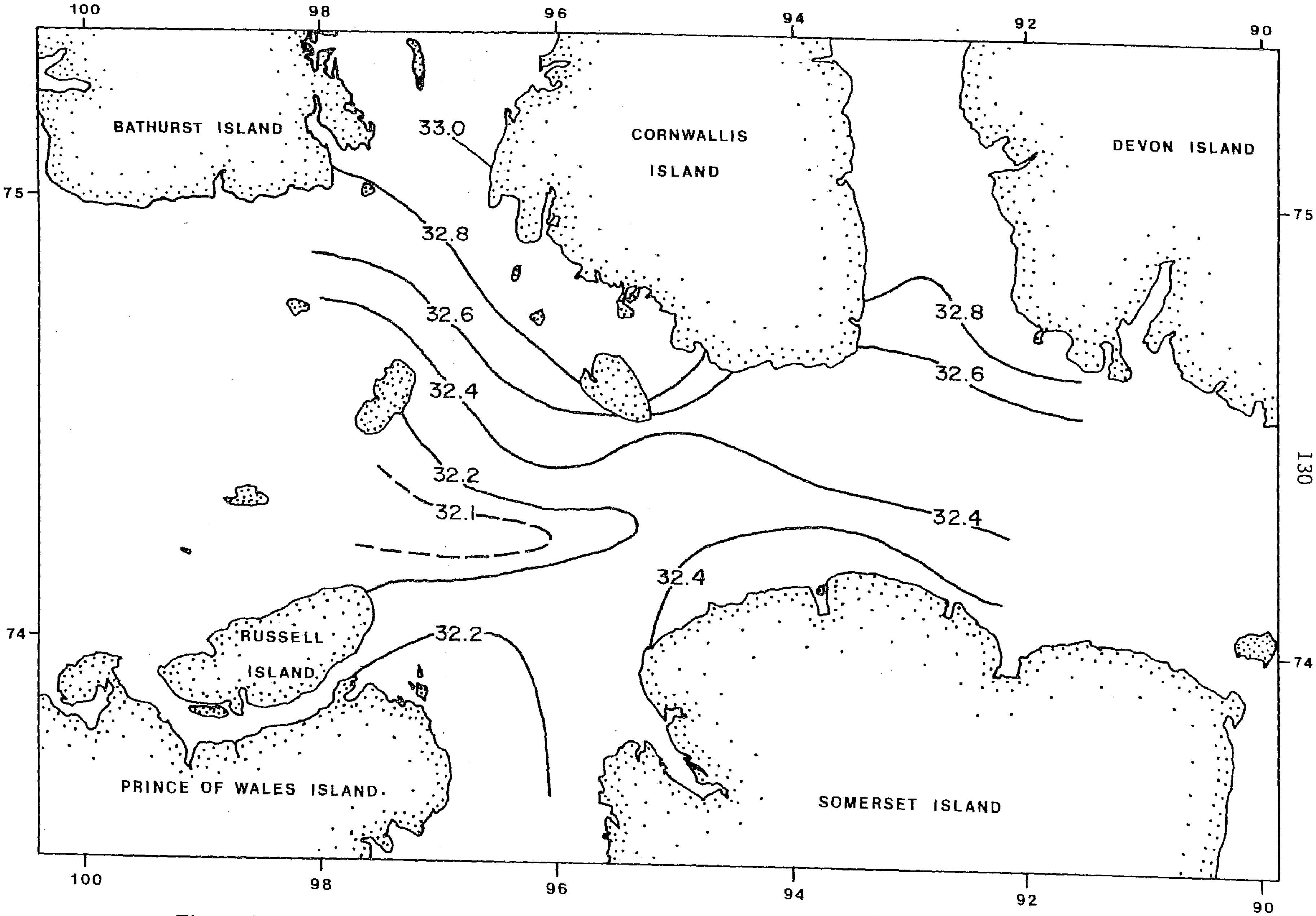

Figure 3.8 Salinity at 10 m, late winter 1982 (after Prinsenberg and Sosnoski, 1983b).

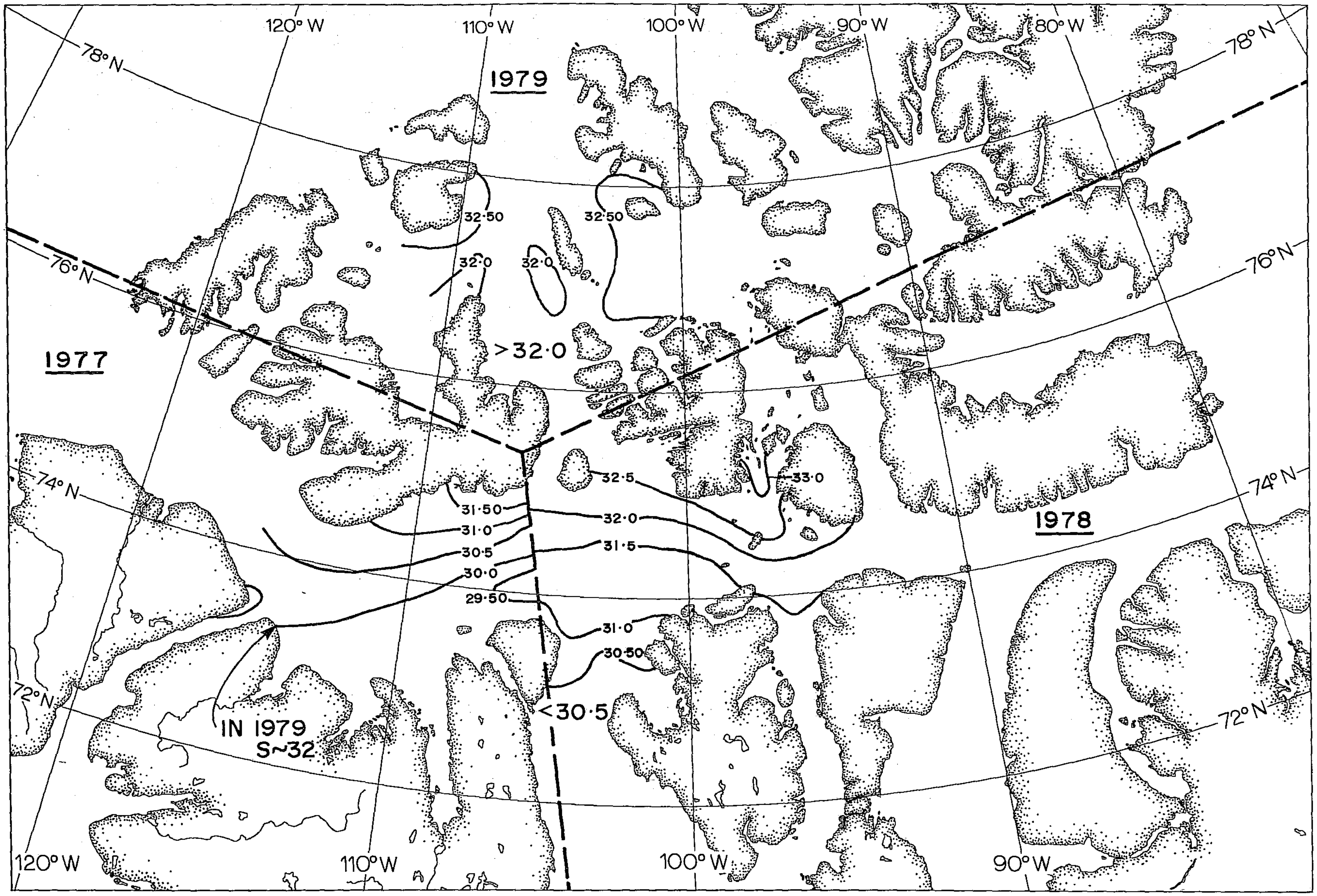

Figure 3.9 Salinity at 5 dbar pressure, late winter 1977 to 1979 (data from Peck, 1978, 1980a; Prinsenberg, 1978).

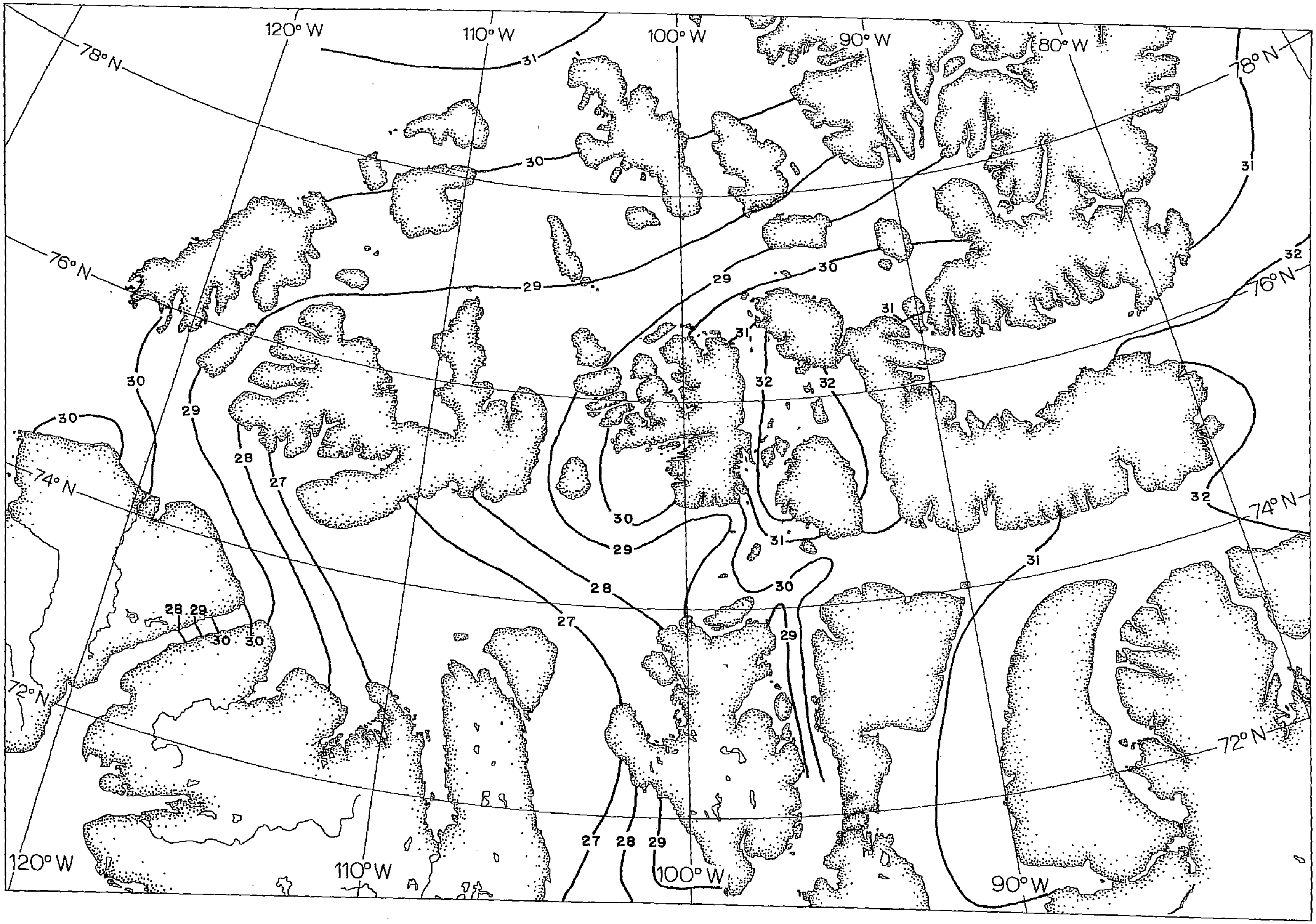

Figure 3.10 Salinity at 10 m, summer 1962 (redrawn from Barber and Huyer, 1971).

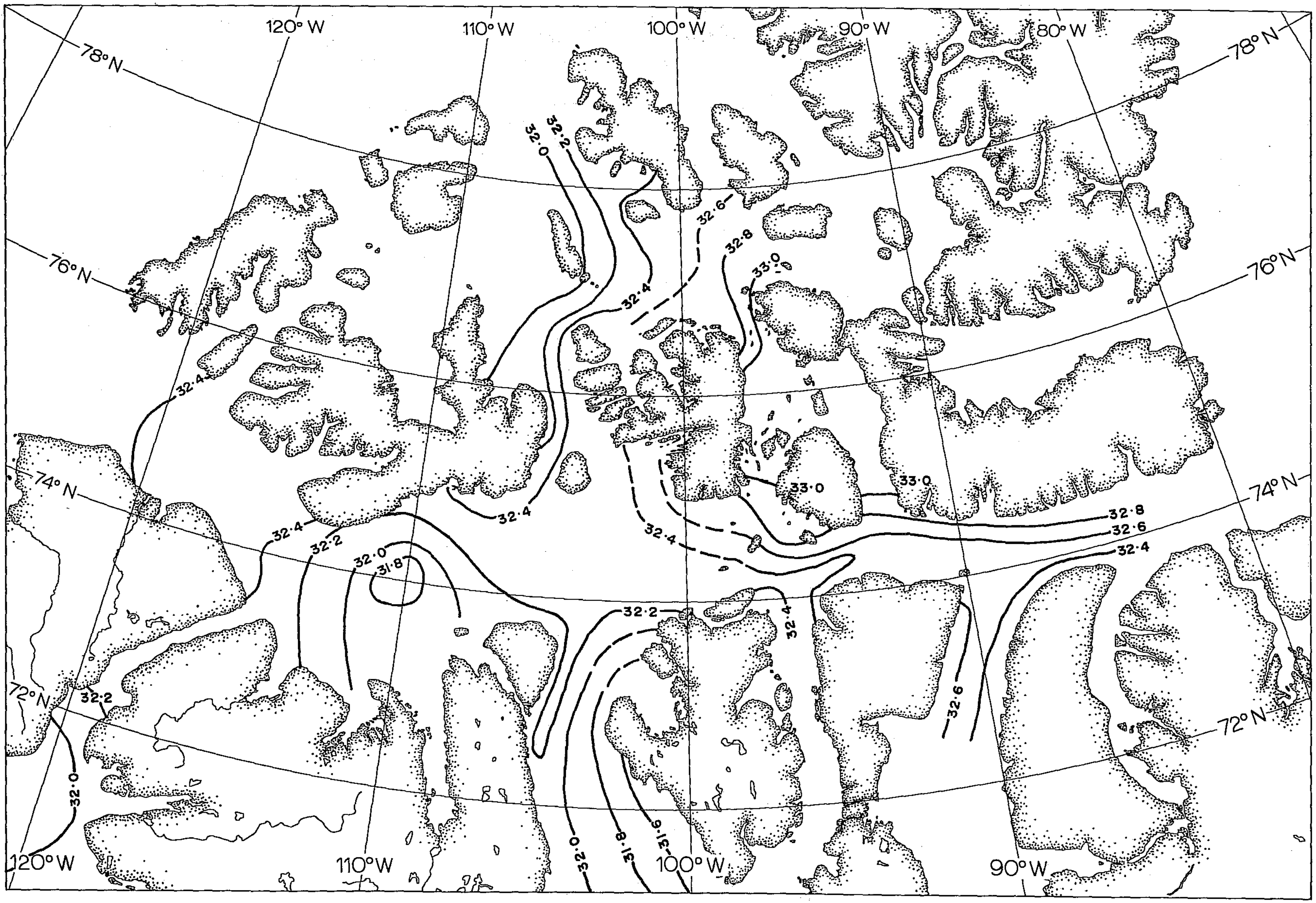

Figure 3.11 Mean salinity between 5 and 50 dbar pressure levels, 1982.

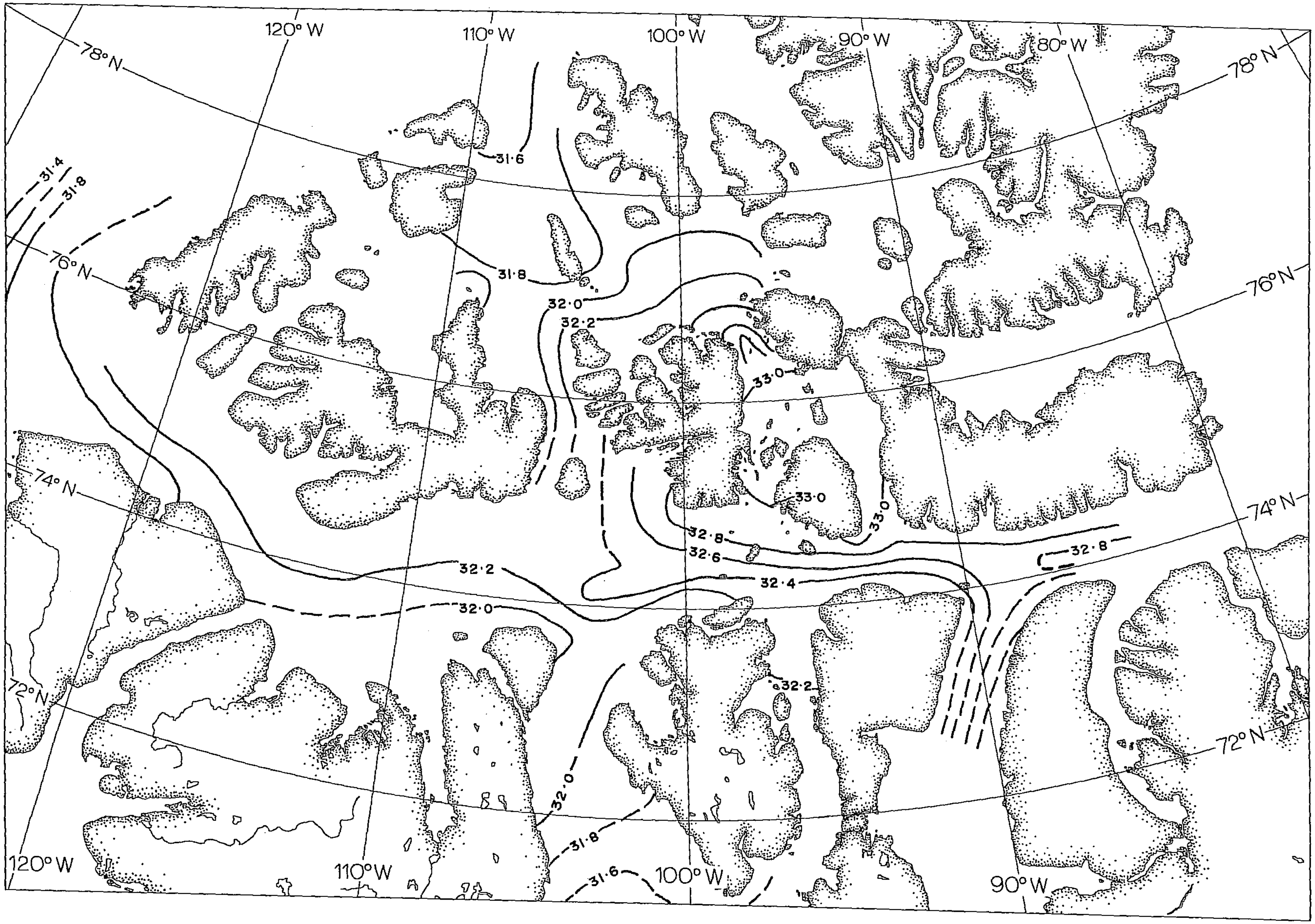

Figure 3.12 Mean salinity between 5 and 50 dbar pressure levels, 1983.

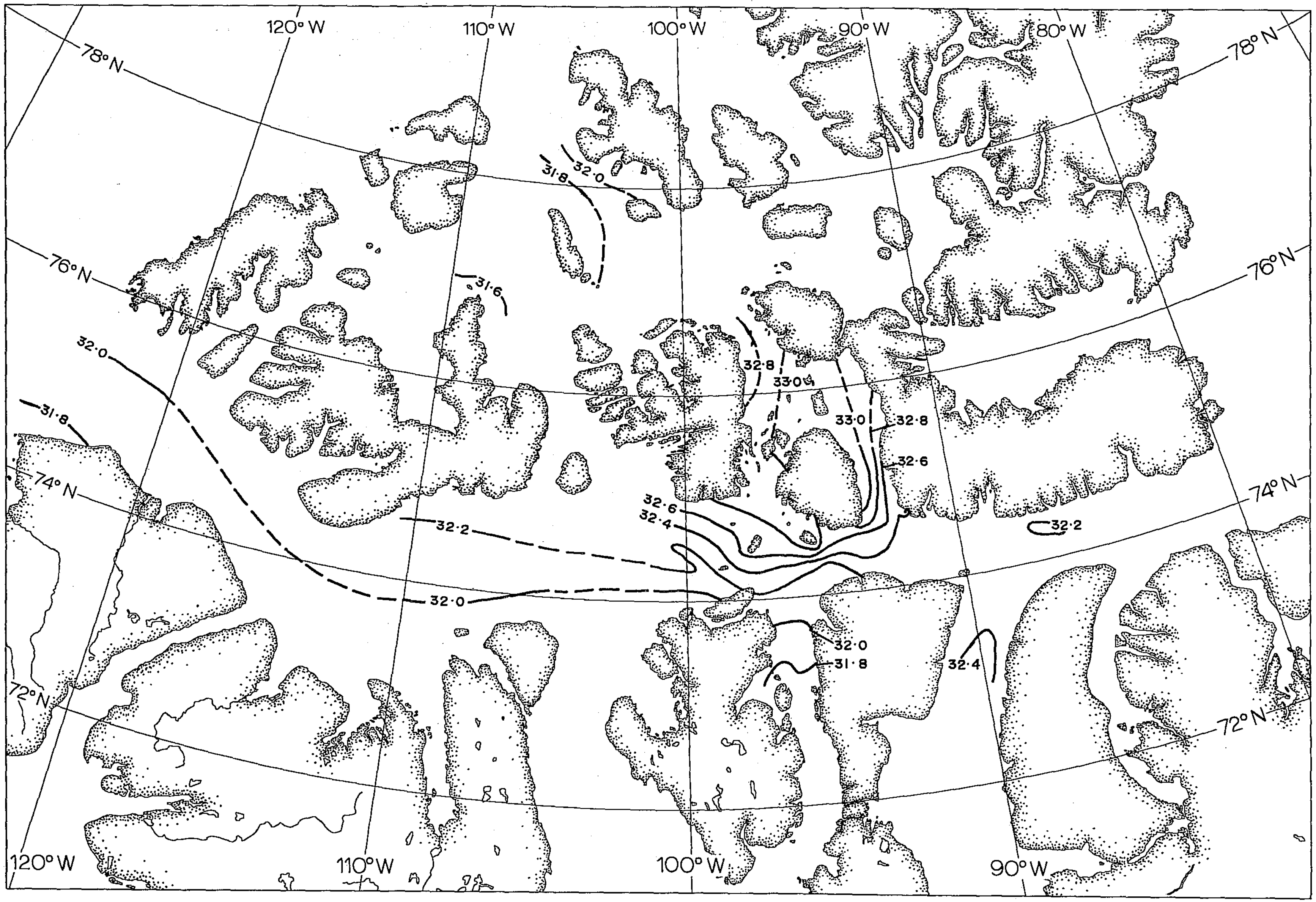

Figure 3.13 Mean salinity between 5 and 50 dbar pressure levels, 1984.

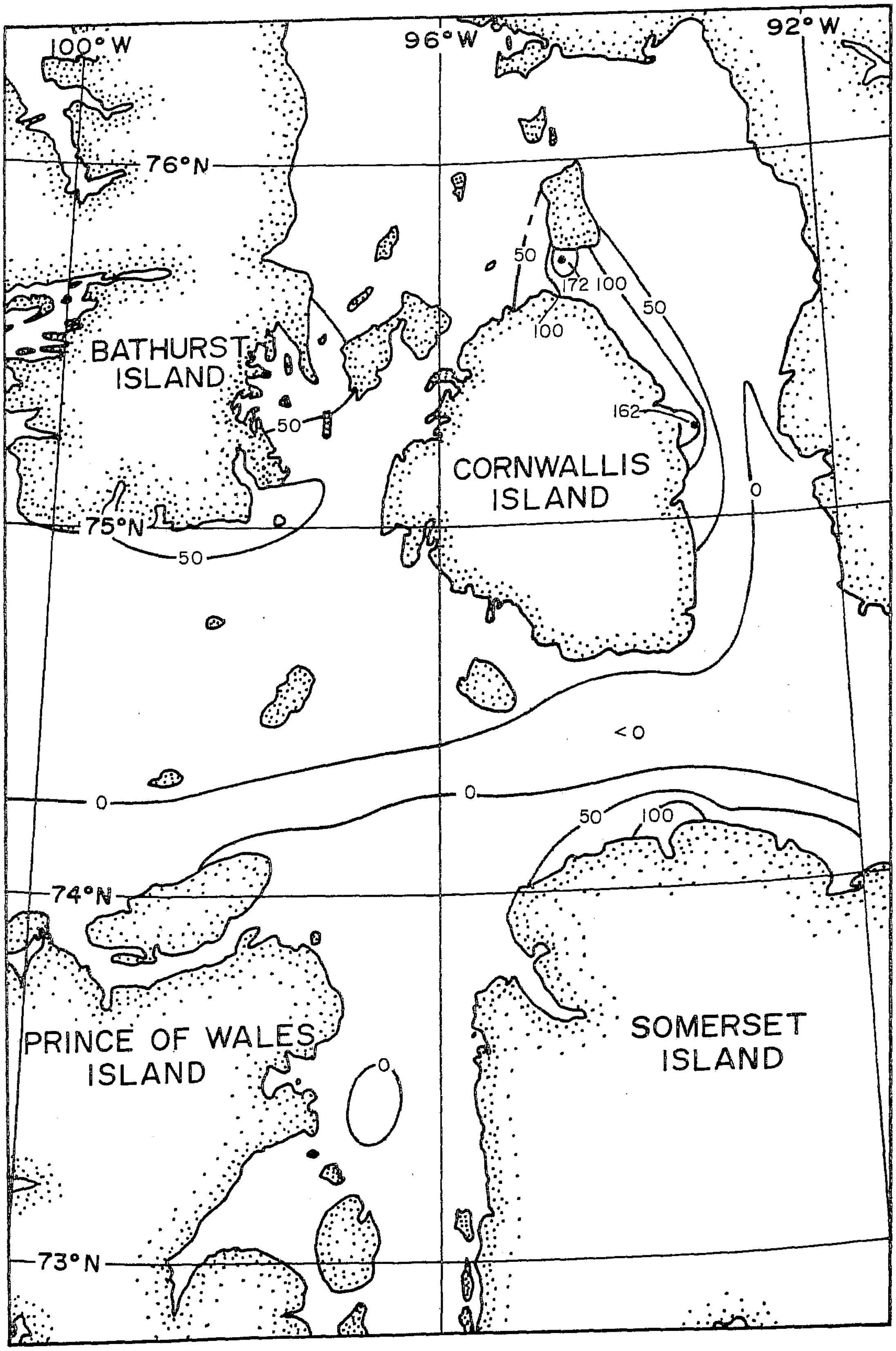

Figure 3.14 Freezing temperature departure at 5 dbar (C millidegrees) in the Barrow Strait region, 1984.

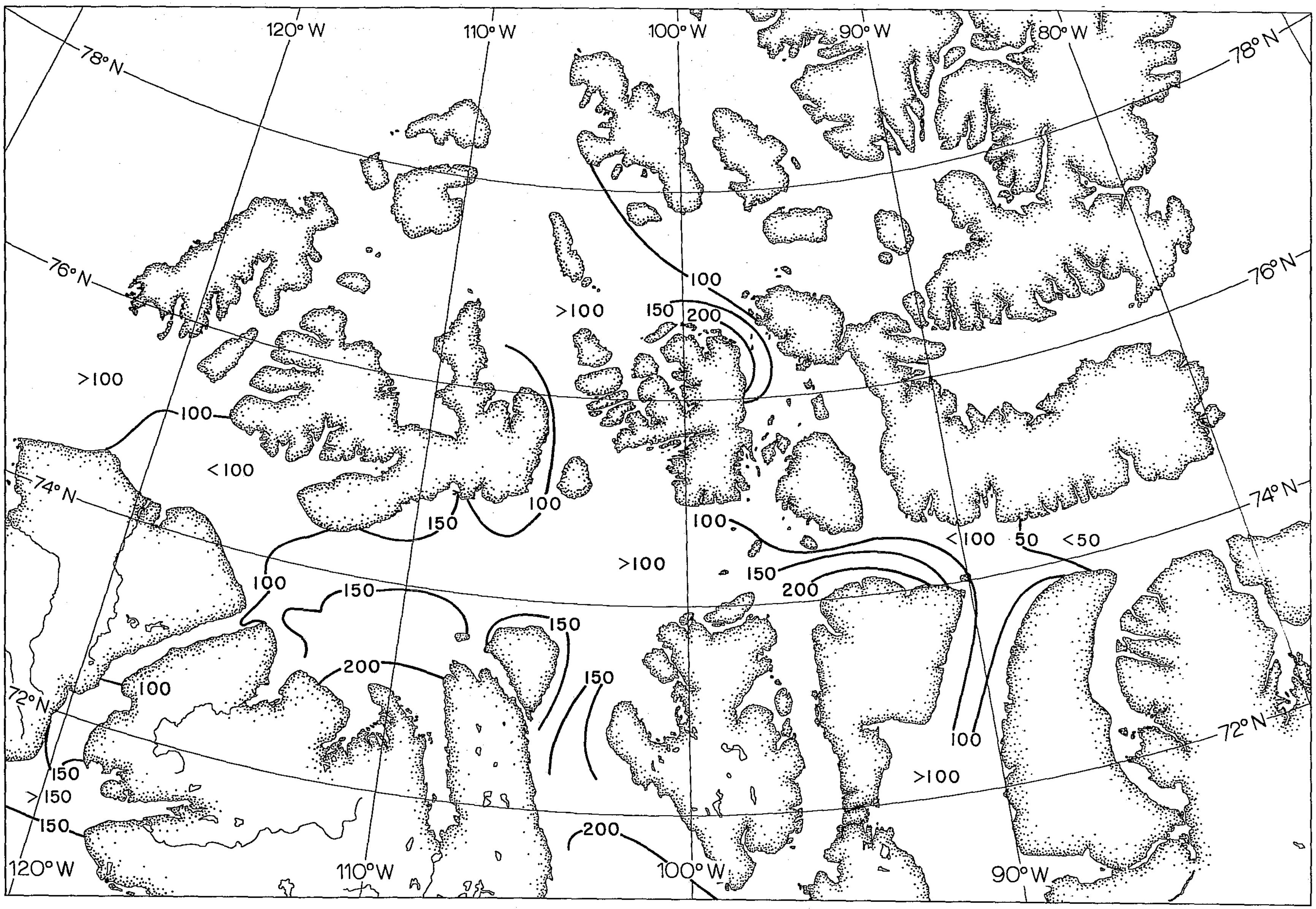

Figure 3.15 Mean freezing temperature departure between 5 dbar and the pressure where S = 33.0 (C millidegrees), 1982.

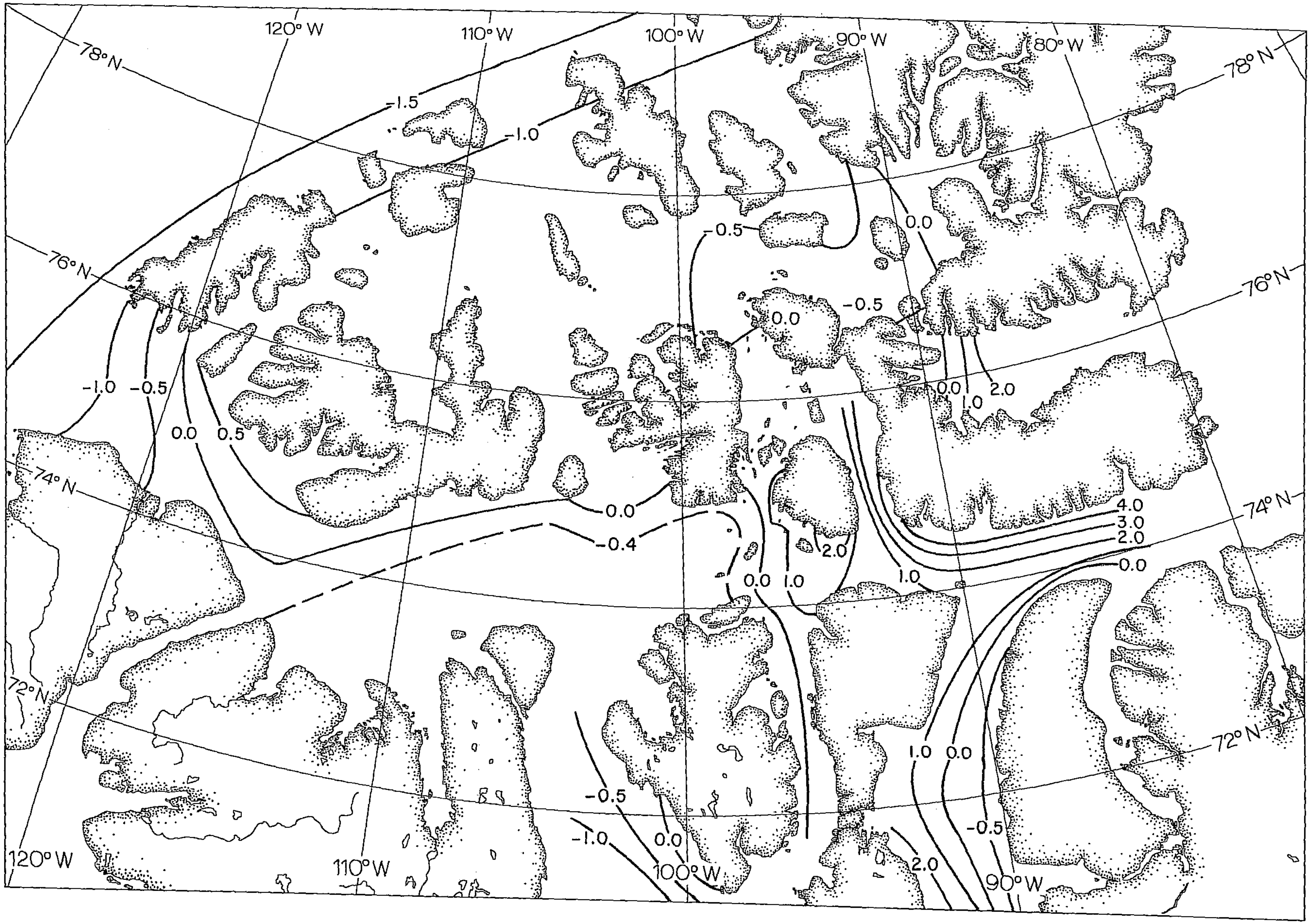

Figure 3.18 Temperature (degrees C) at 10 m, summer 1962 (redrawn from Barber and Huyer, 1971).

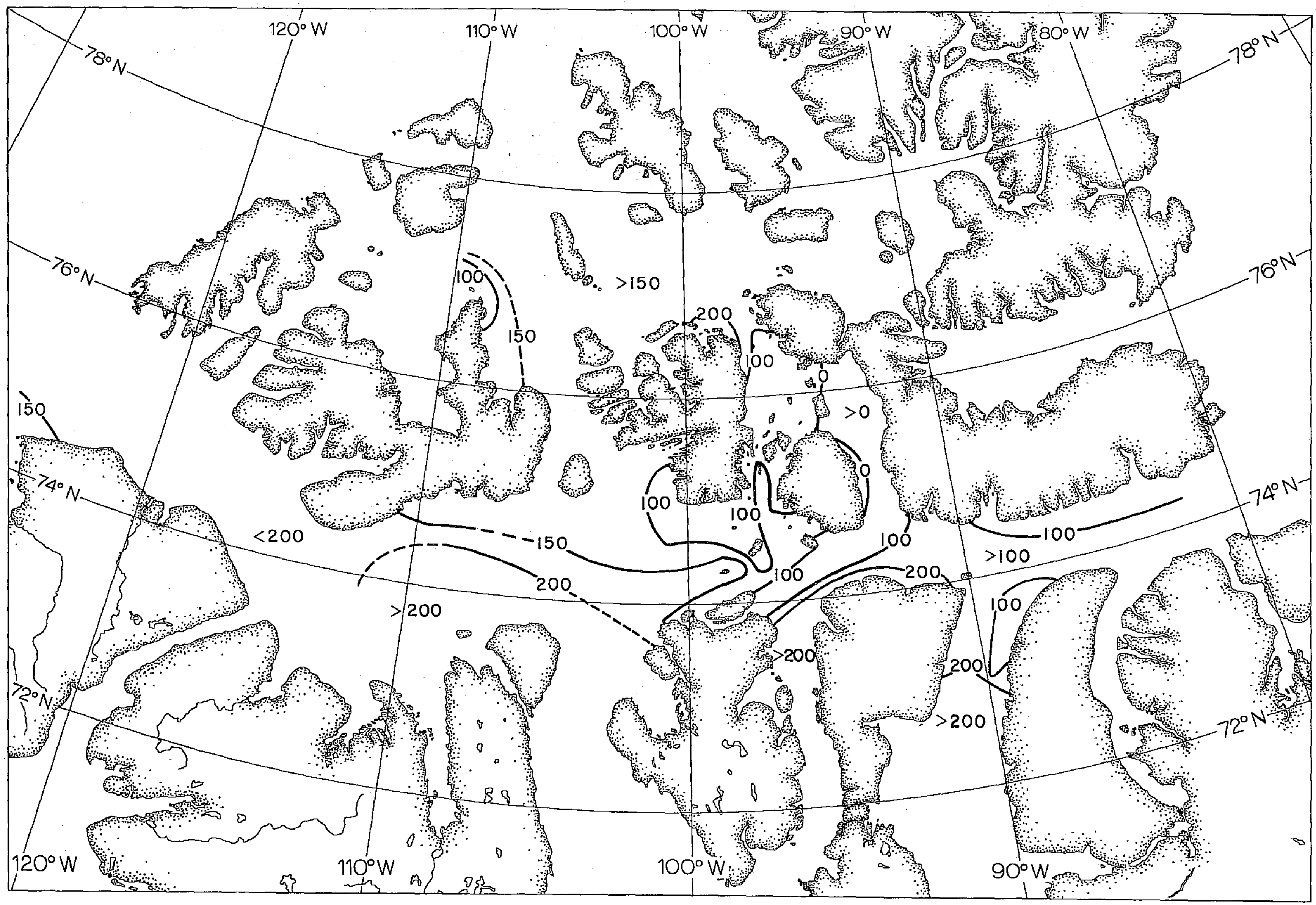

Figure 3.17 Mean freezing temperature departure between 5 dbar and the pressure where S = 33.0 (C millidegrees), 1984.

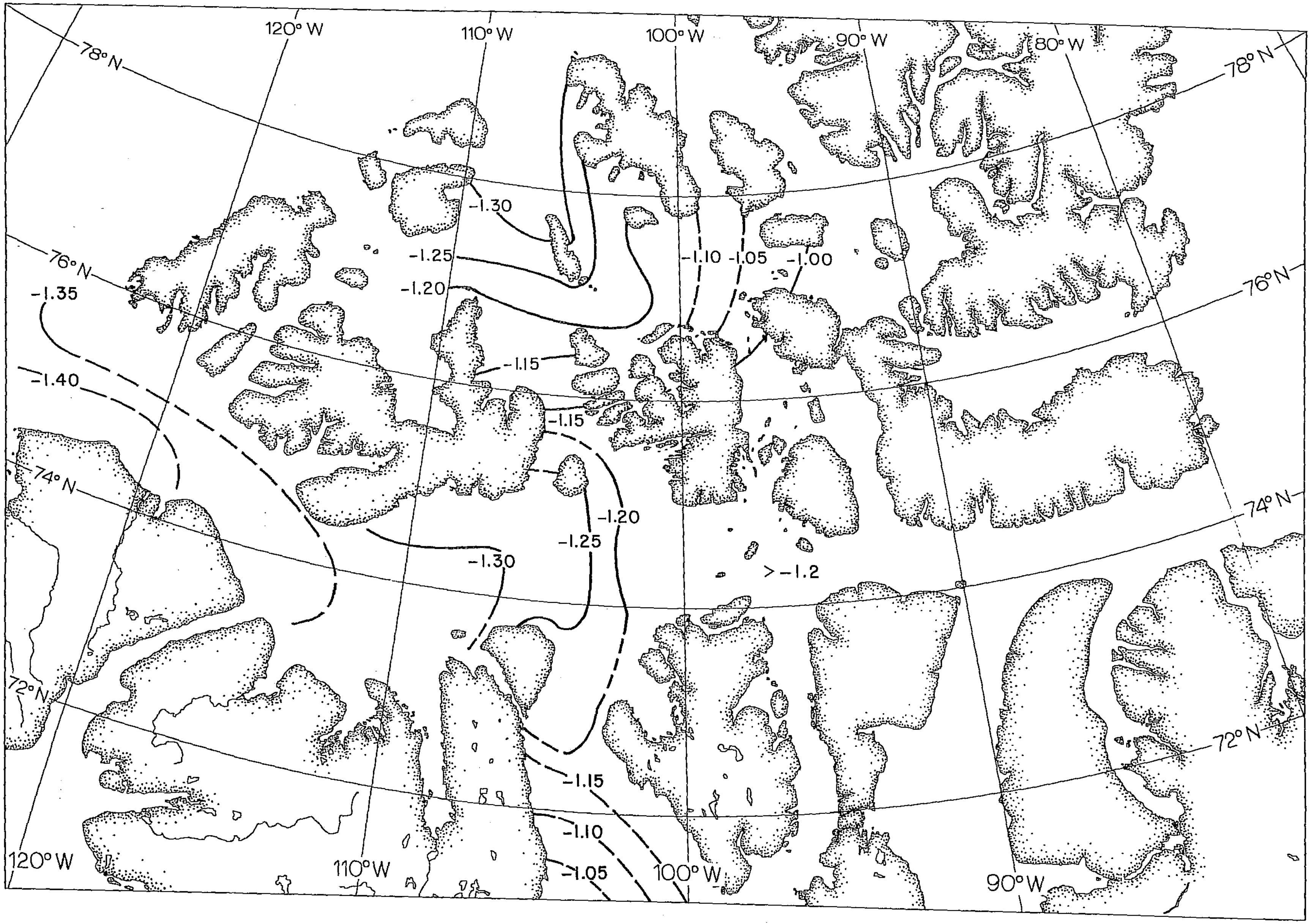

Figure 3.20 Temperature (degrees C) on the 33.5 isohaline surface in 1983. An indication of the halocline temperature.

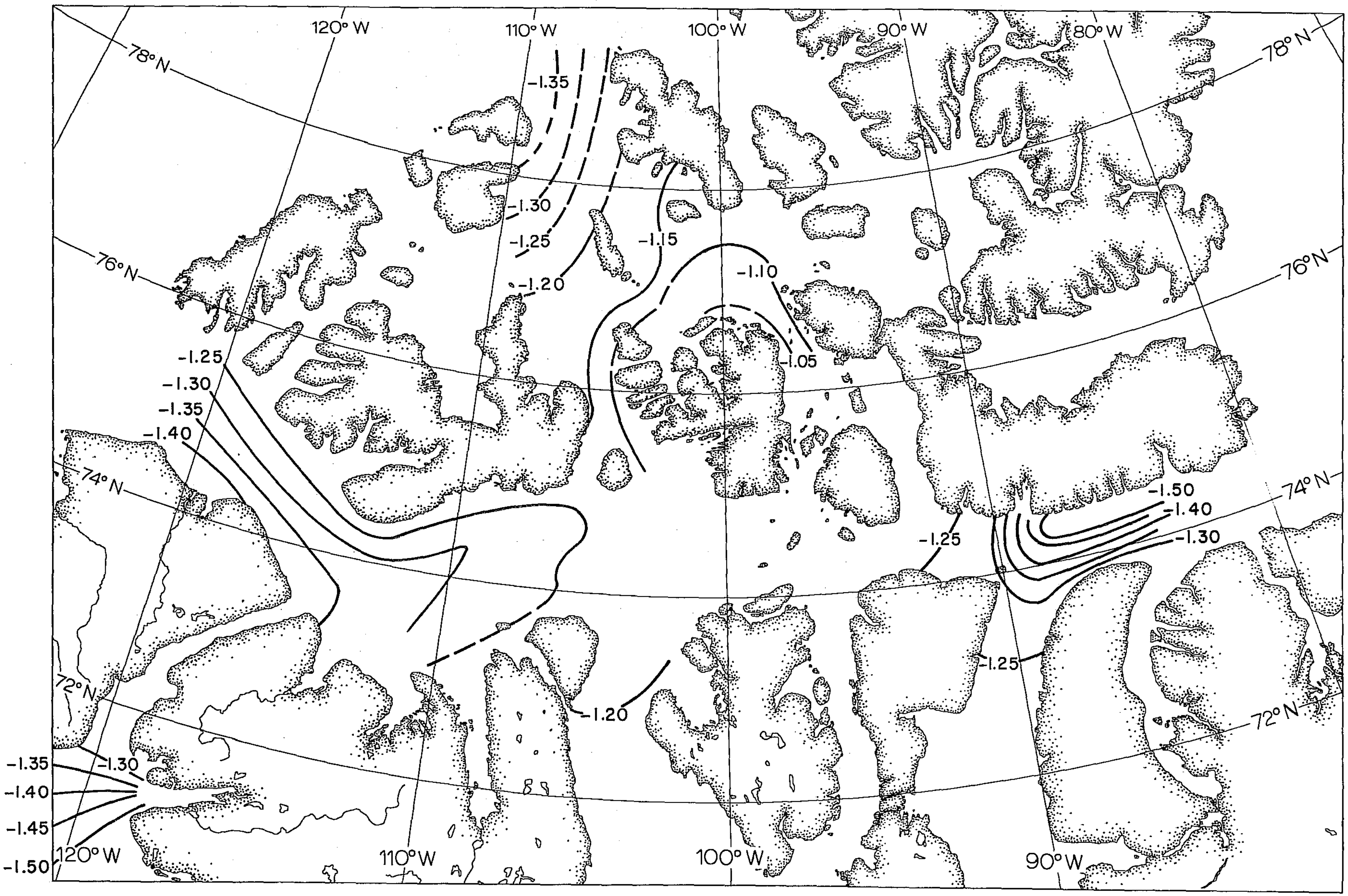

Figure 3.19 Temperature (degrees C) on the 33.5 isohaline surface in 1982. An indication of the halocline temperature.

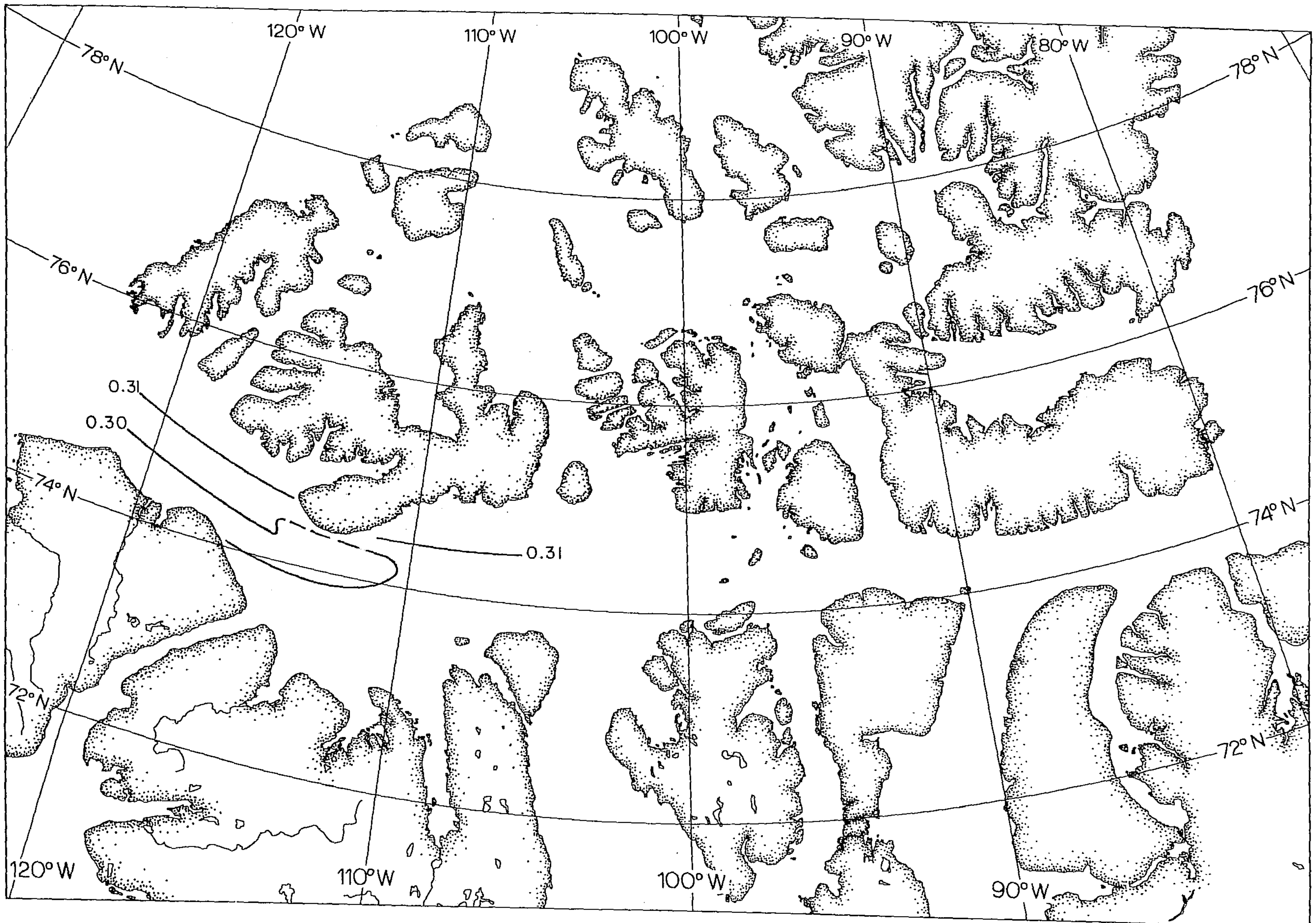

Figure 3.22 Temperature (degrees C) on the 34.83 isohaline surface in 1982, an indication of the Atlantic water temperature.

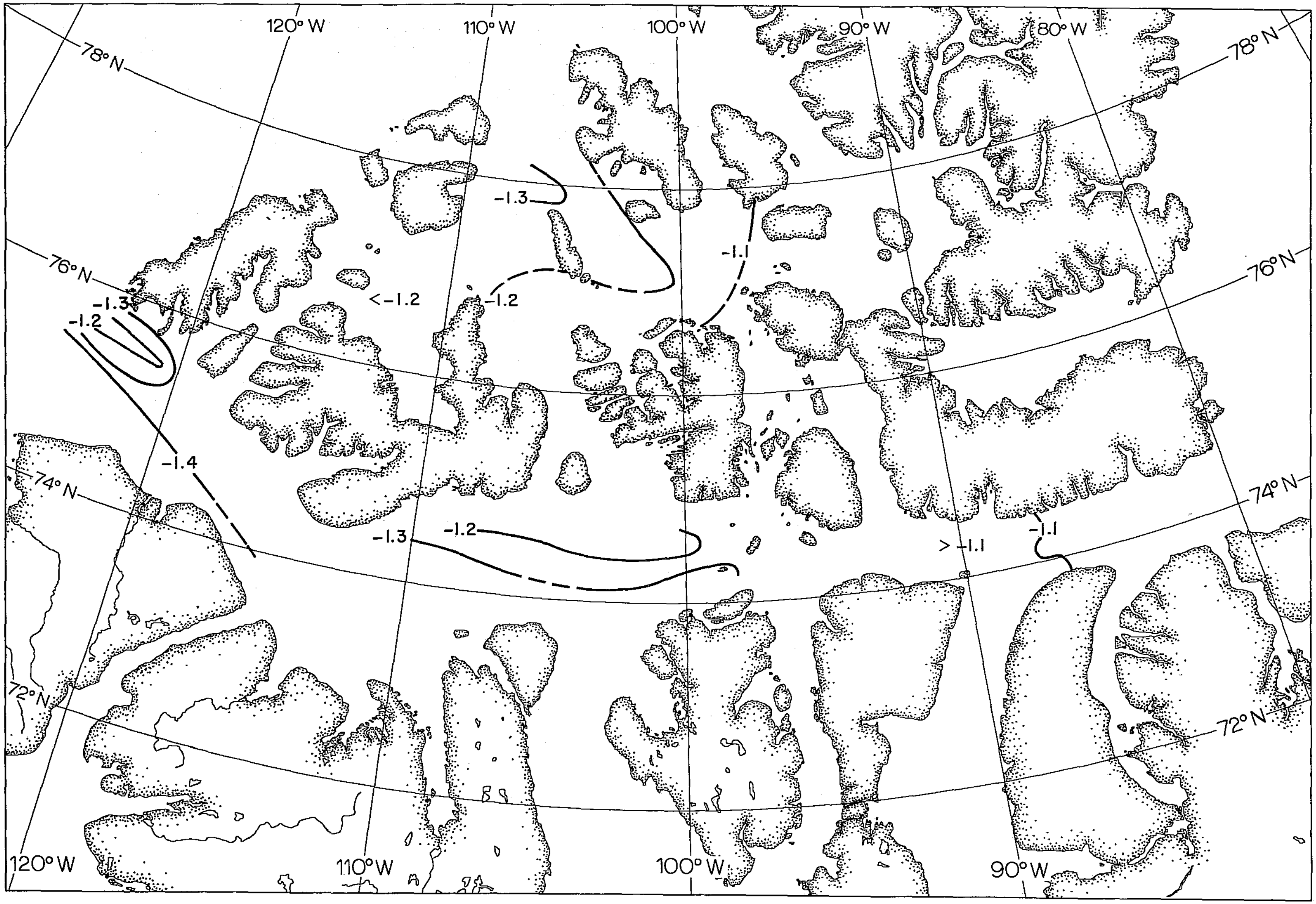

Figure 3.21 Temperature (degrees C) on the 33.5 isohaline surface in 1984. An indication of the halocline temperature.

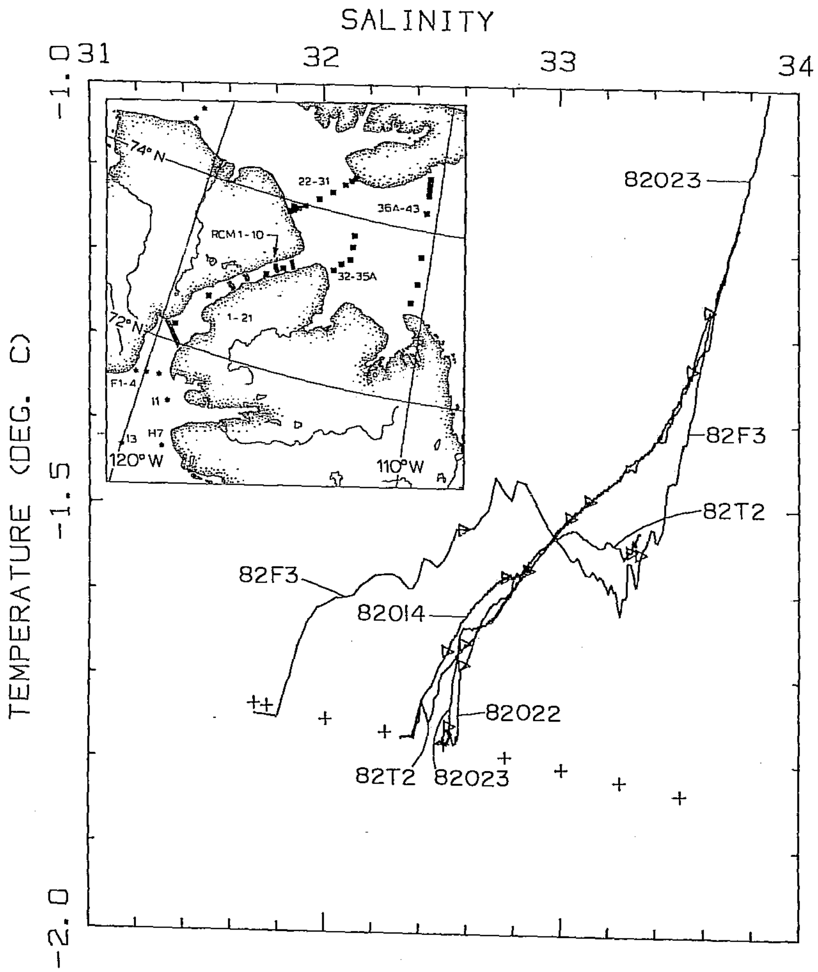

Figure 4.1 Temperature-salinity diagram for stations from Amundsen Gulf through Prince of Wales Strait to M'Clure Strait, 1982.

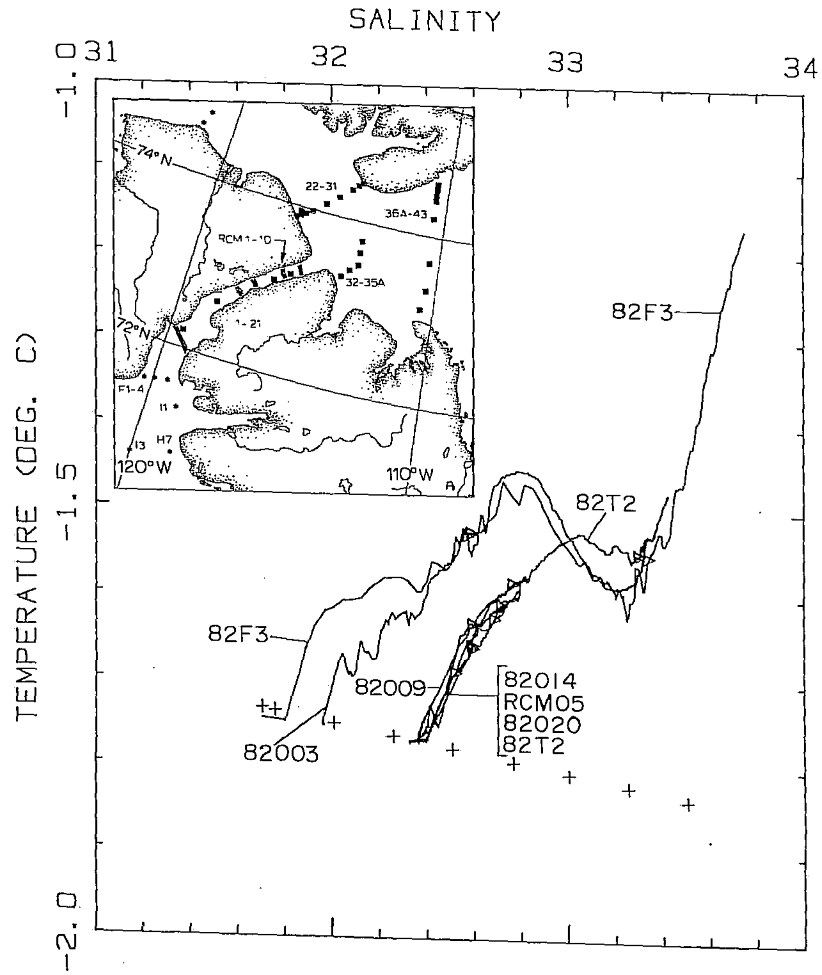

Figure 4.2 Temperature-salinity diagram for stations along the axis of Prince of Wales Strait from Amundsen Gulf, 1982.

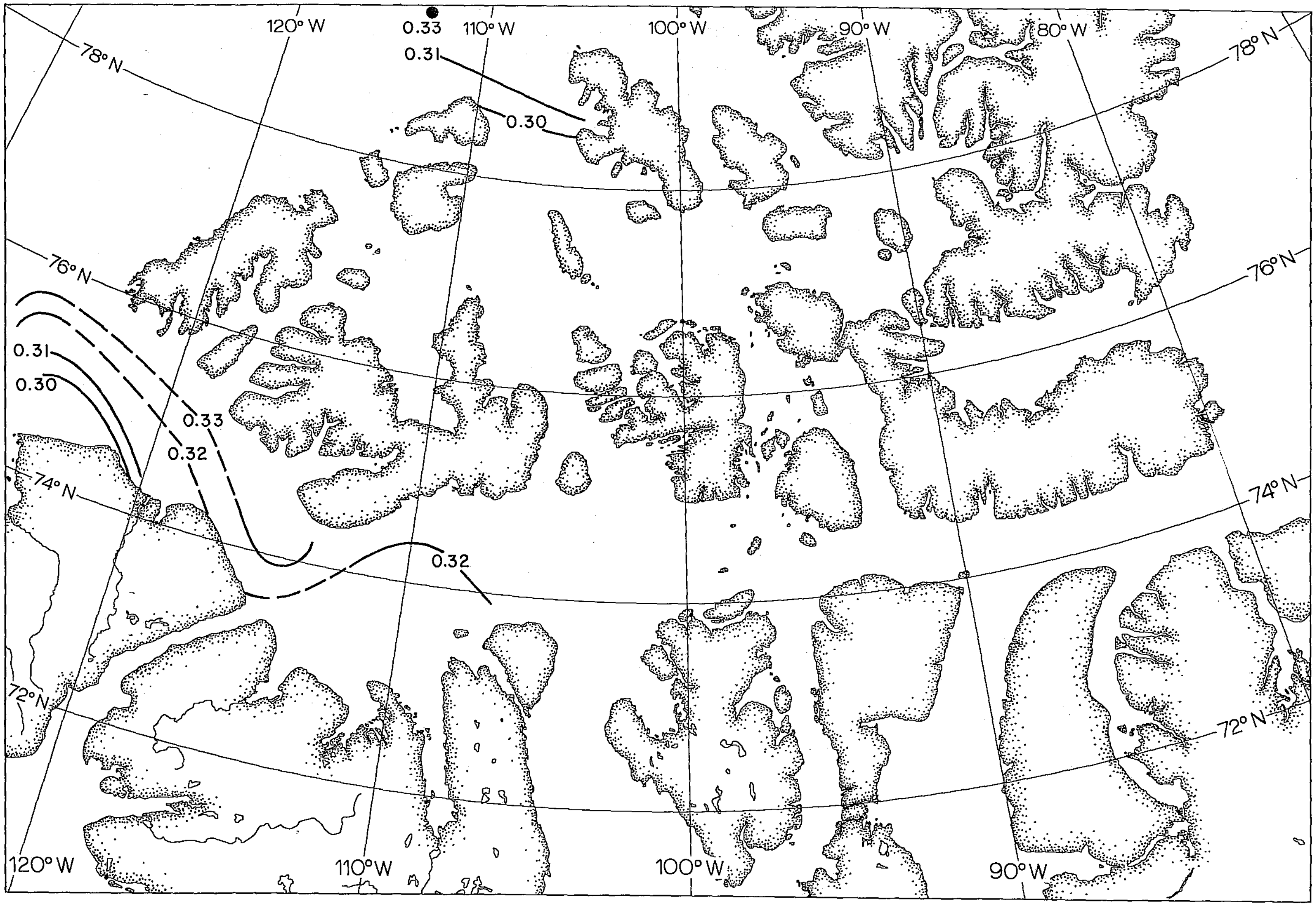

Figure 3.23 Temperature (degrees C) on the 34.83 isohaline surface in 1983, an indication of the Atlantic water temperature.

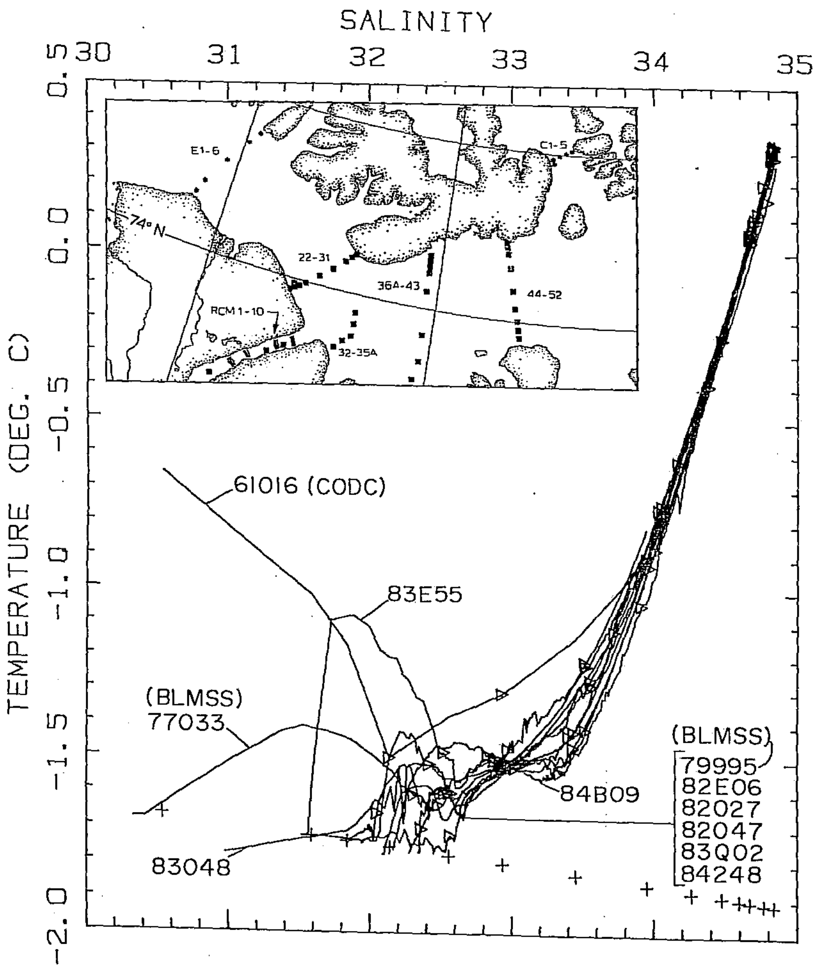

Figure 4.5 Composite temperature-salinity diagram for selected stations in western Parry Channel in summer (1961) and winter (1977, 1979, 1982, 1983, 1984).

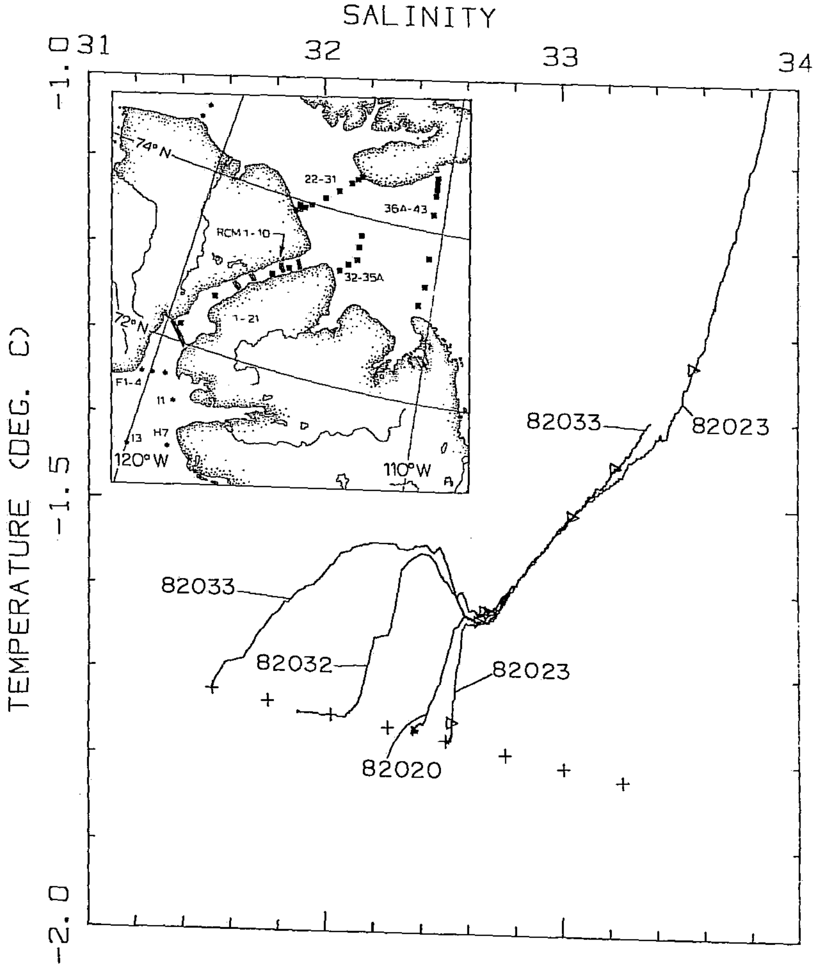

Figure 4.6 Comparison of temperature-salinity diagrams for stations in southeast M'Clure Strait, Northern Prince of Wales Strait and southwest Viscount Melville Sound.

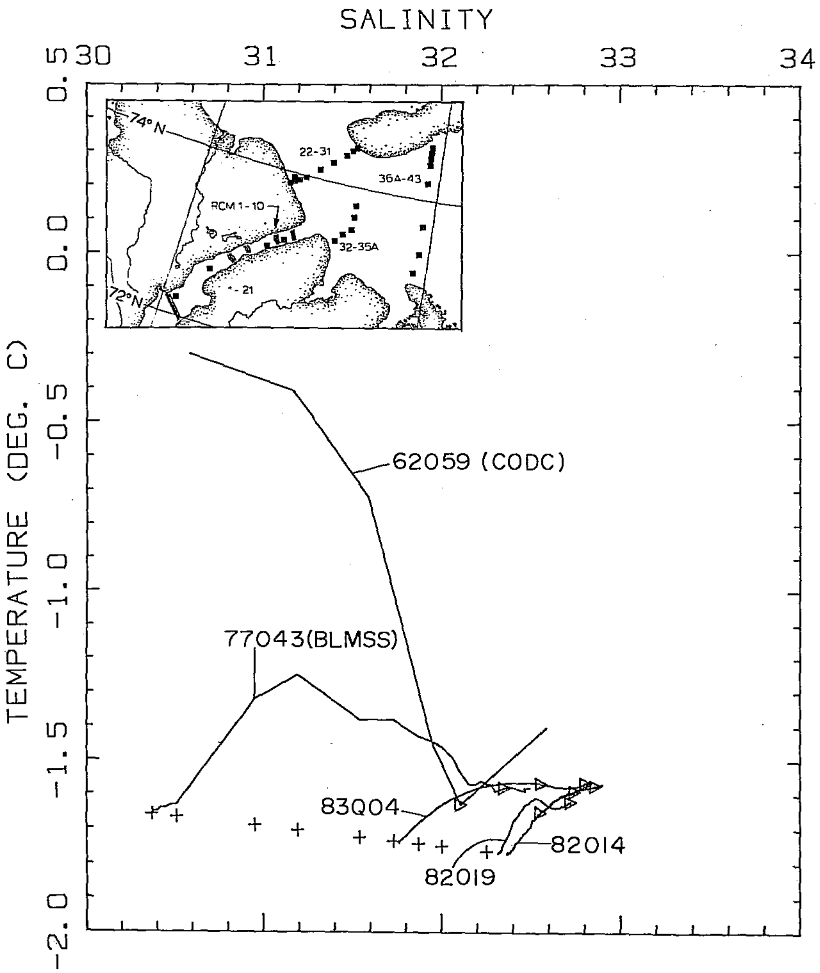

Figure 4.3 Temperature-salinity diagram variations for stations in northern Prince of Wales Strait: 1961 summer, 1977, 1982 and 1983 winter.

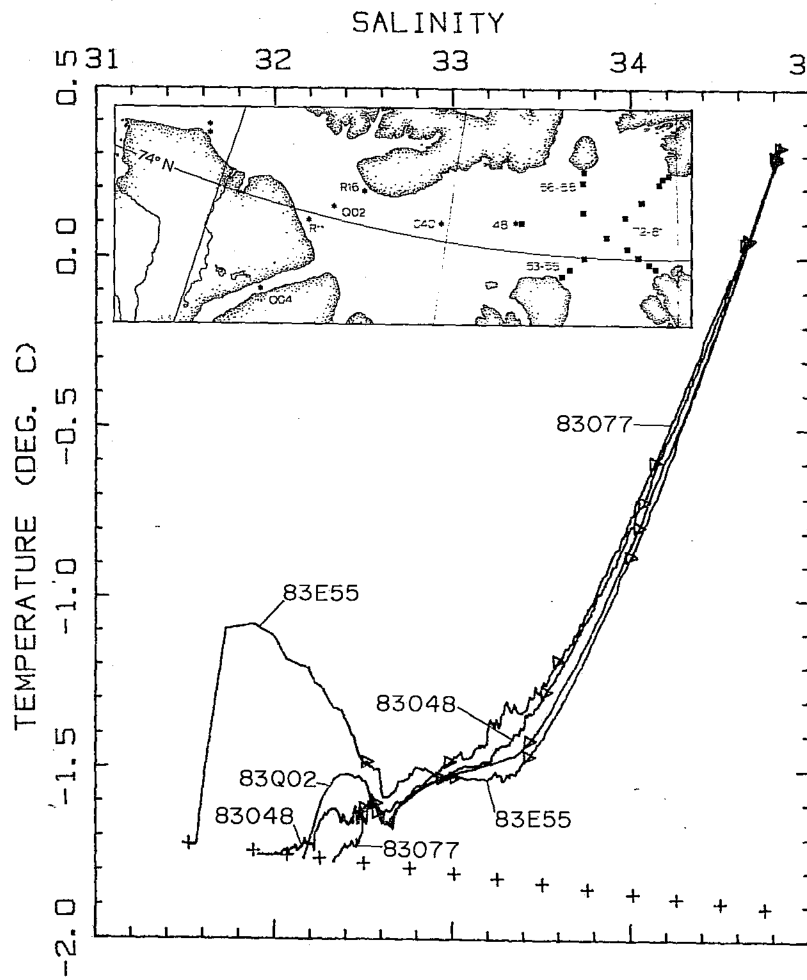

Figure 4.4 Temperature-salinity diagram for stations in western Parry Channel (1983).

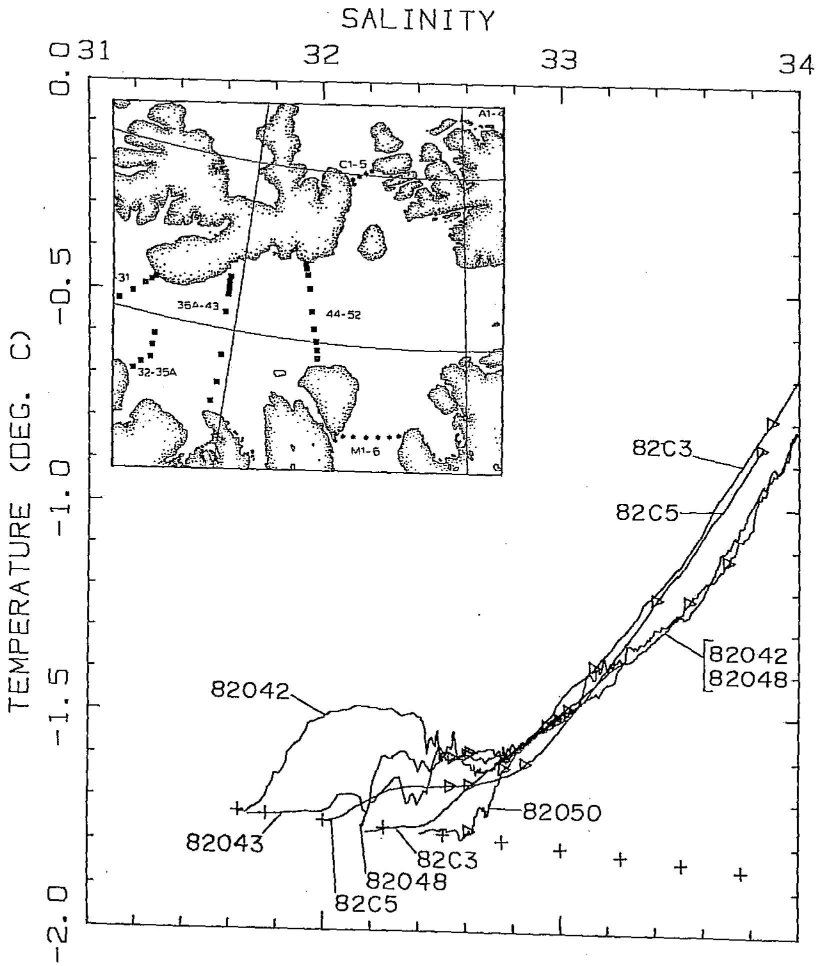

Figure 4.9 Temperature-salinity diagram for 1982 stations in northern Viscount Melville Sound and Byam Martin Channel.

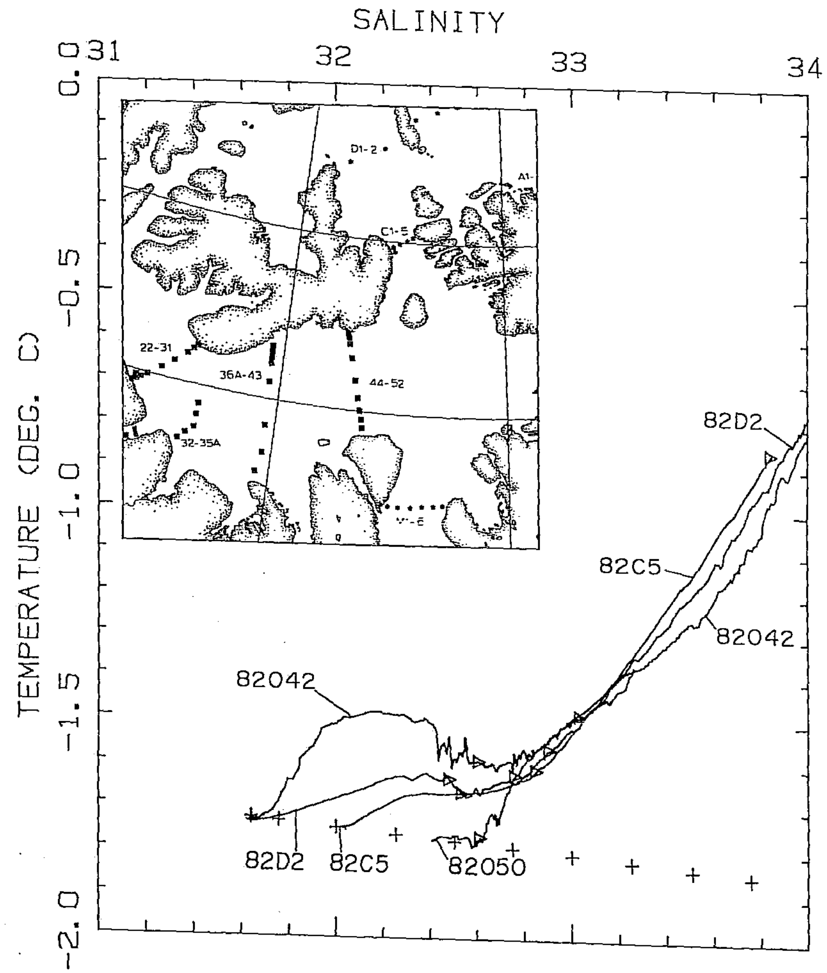

Figure 4.10 Temperature-salinity diagram for 1982 stations in northern Viscount Melville Sound, Byam Martin Channel and Hazen Strait.

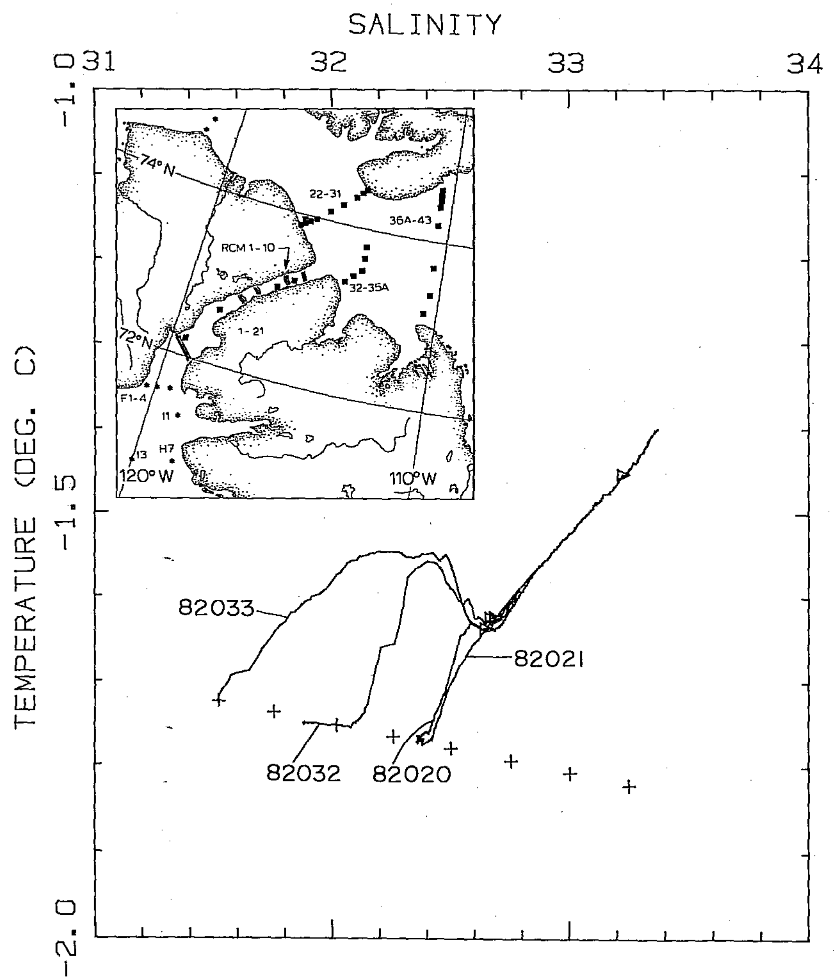

Figure 4.7 Comparison of temperature-salinity diagrams for station on the east side of the north end of Prince of Wales Strait with stations on southwest side of Viscount Melville Sound.

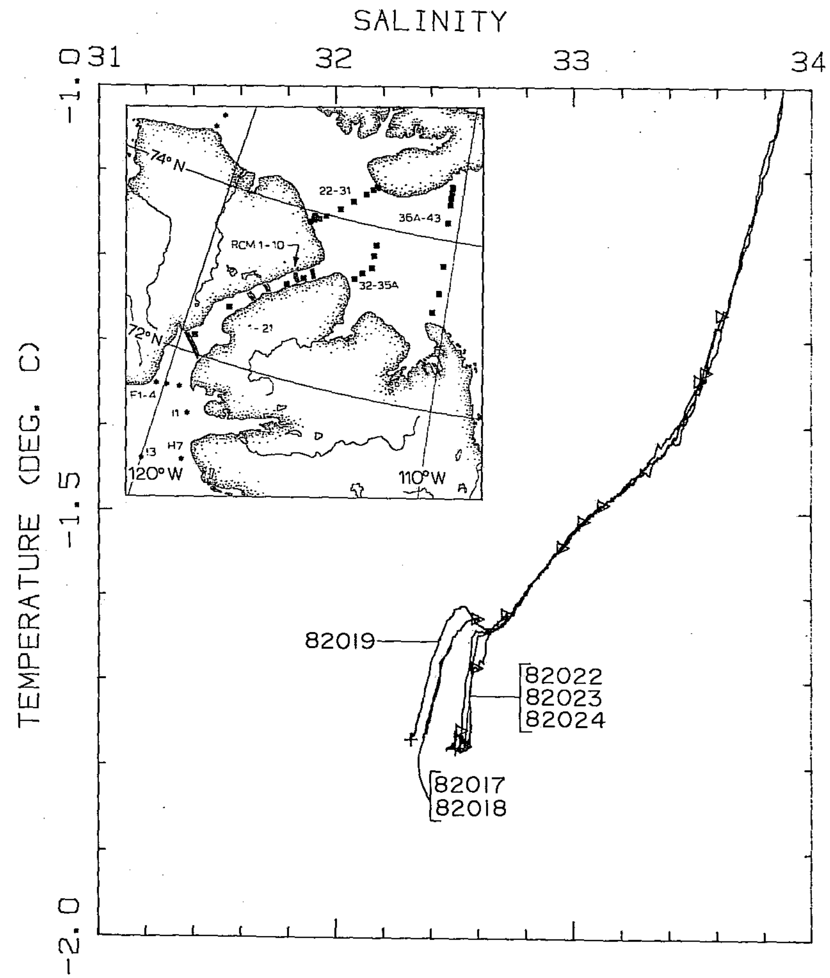

Figure 4.8 Comparison of temperature-salinity diagrams for stations on the northwest side of Prince of Wales Strait with stations on southeast side of M'Clure Strait.

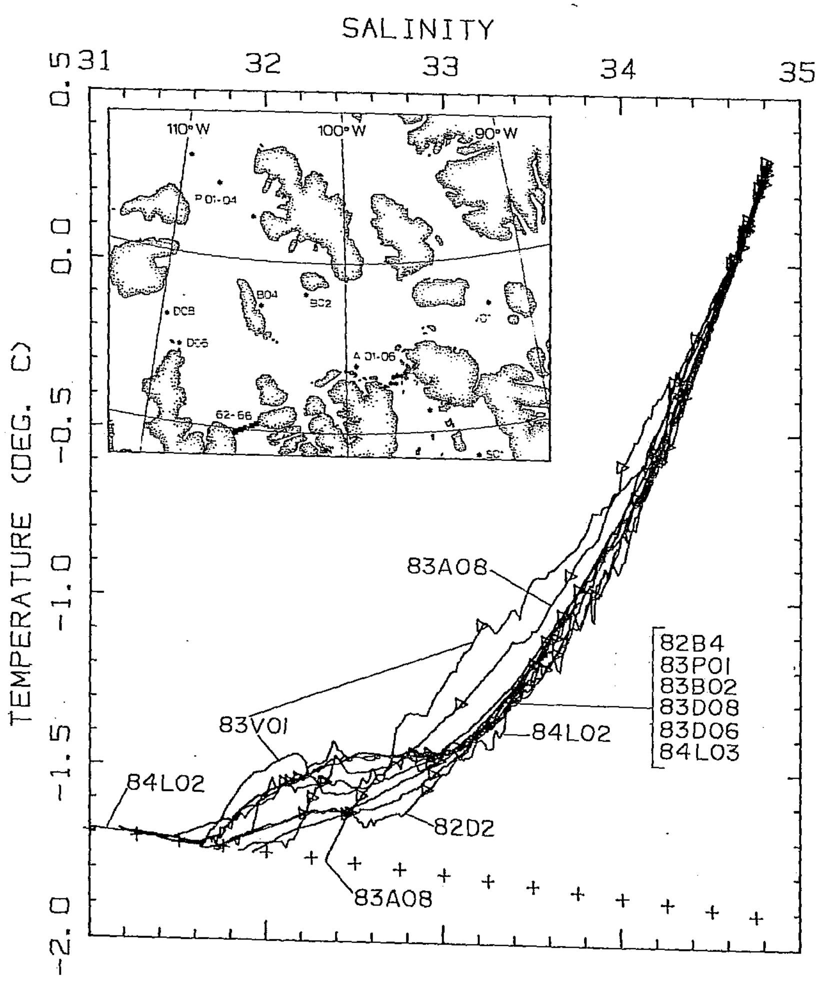

Figure 4.13 Temperature-salinity diagrams for stations in the Queen Elizabeth Islands from 1982 to 1984.

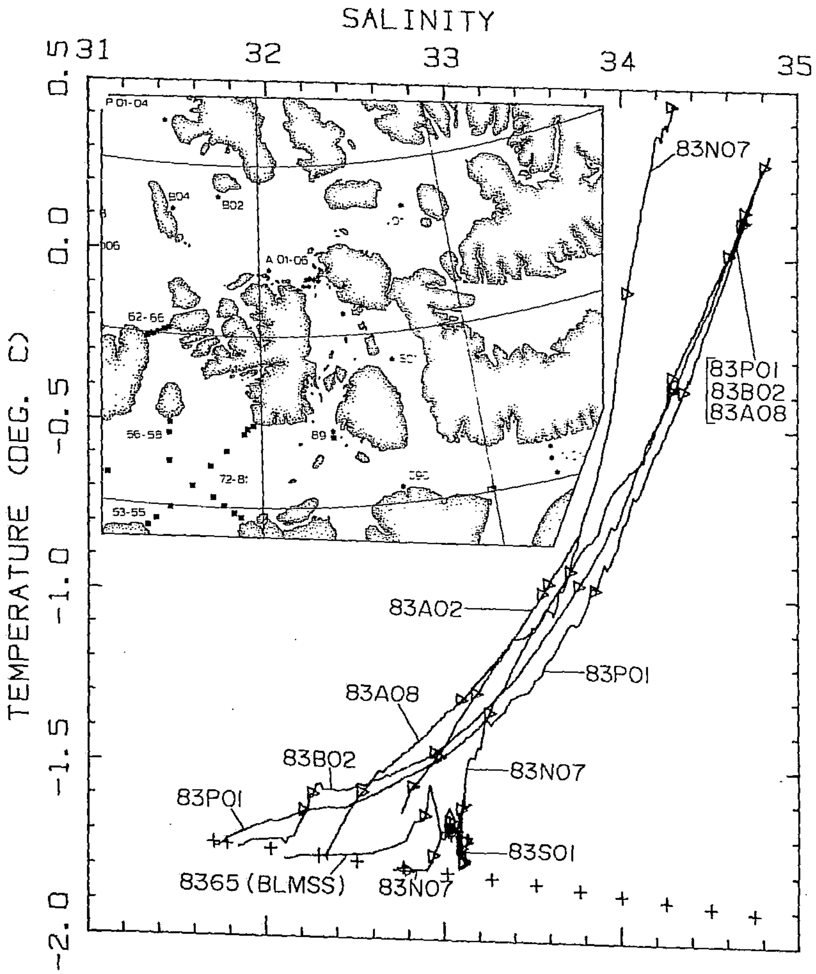

Figure 4.14 Temperature-salinity plots for stations on a section from Prince Gustaf Adolf Sea through Penny Strait and Wellington Channel to Lancaster Sound, 1983.

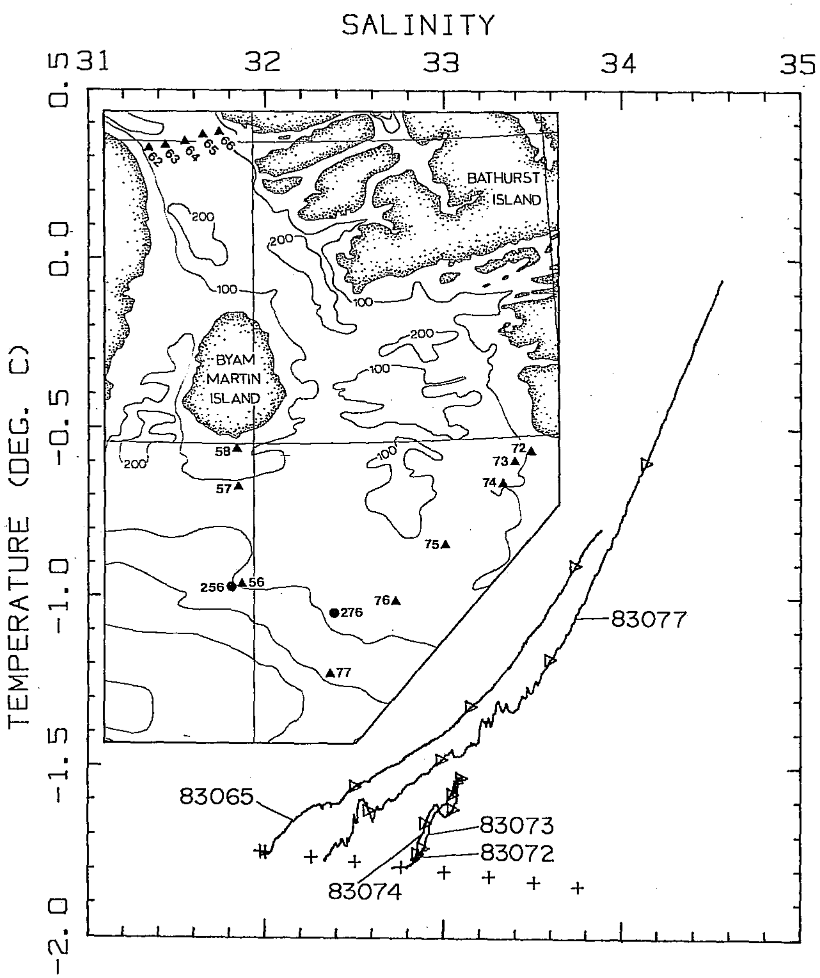

Figure 4.11 Comparison of temperature-salinity plots for 1983 stations in Byam Martin Channel and eastern Viscount Melville Sound.

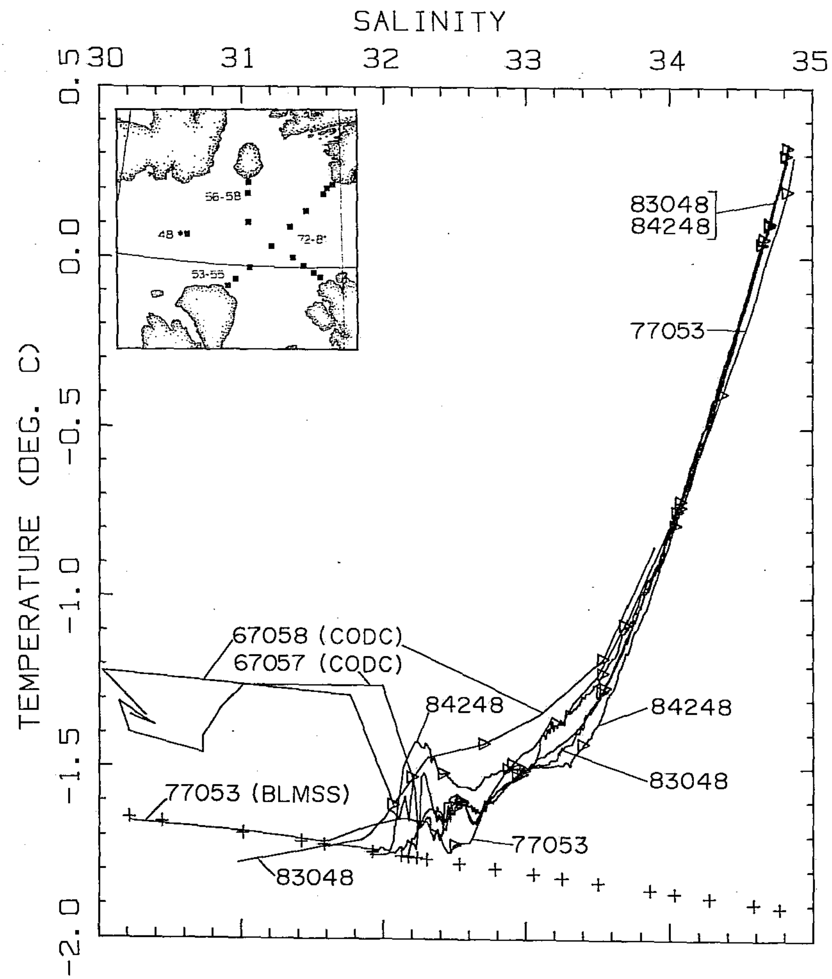

Figure 4.12 Comparison of temperature-salinity plots at one location in central Viscount Melville Sound for summer (1967) and winter (1977, 1982, 1983, 1984).

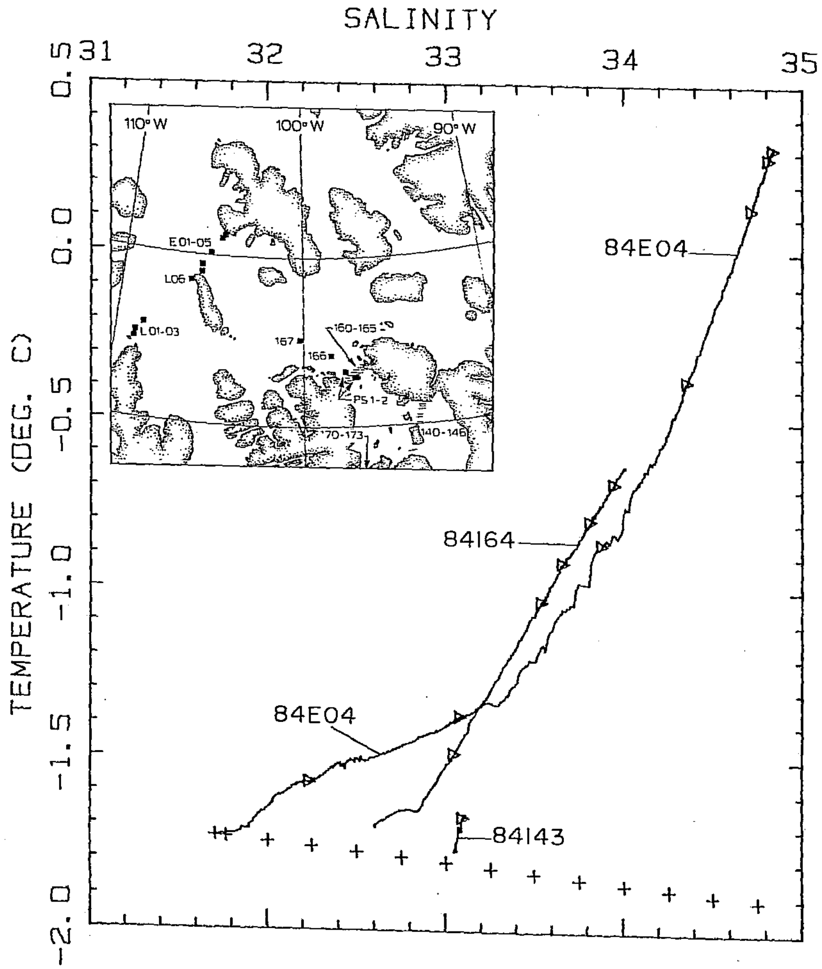

Figure 4.17 Comparison of temperature-salinity plots for 1984 stations either side of Penny Strait.

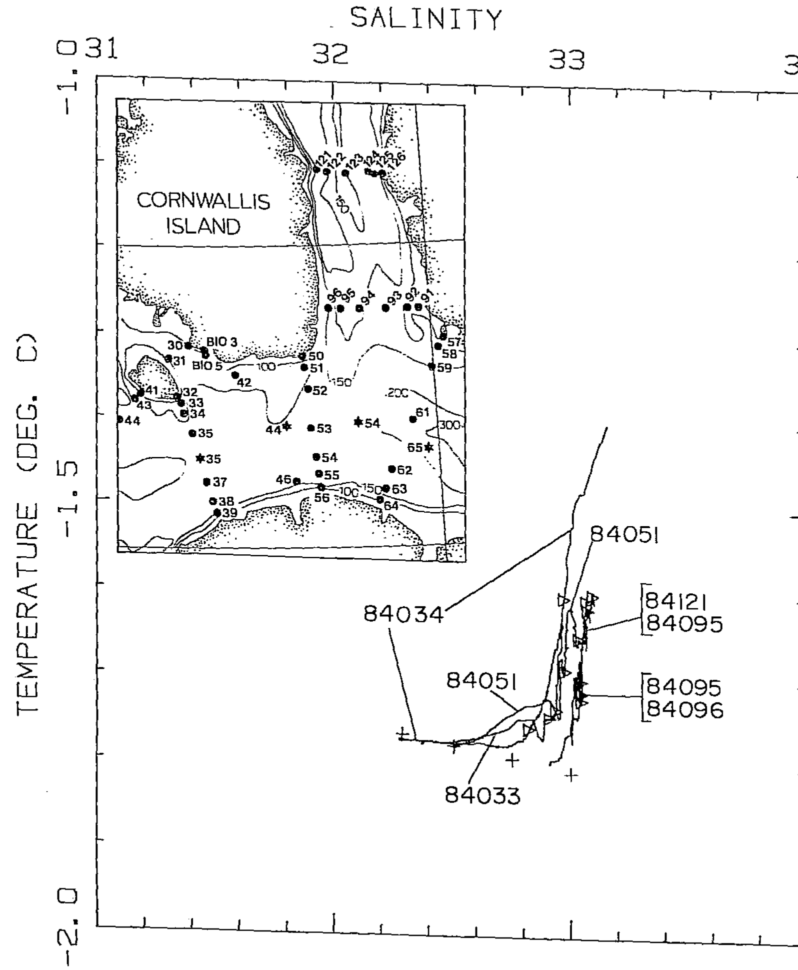

Figure 4.18 Comparison of temperature-salinity diagrams for Wellington Channel and the northeast portion of Barrow Strait, 1984.

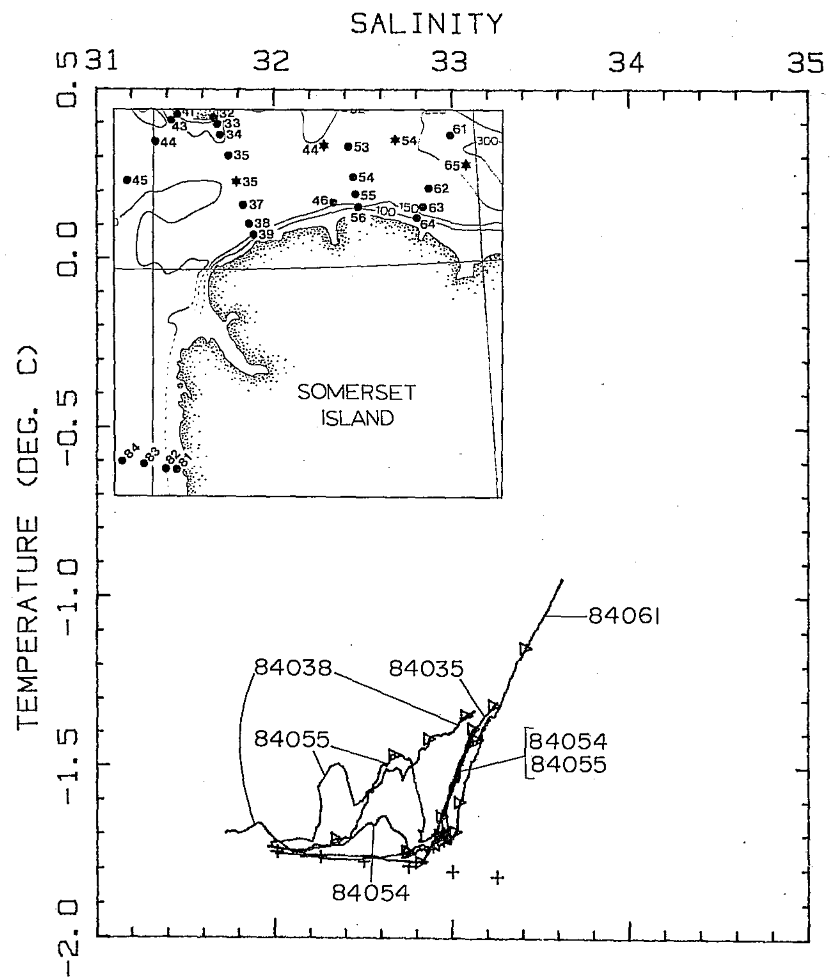

Figure 4.15 Temperature-salinity plots for 1984 stations from central to eastern Barrow Strait.

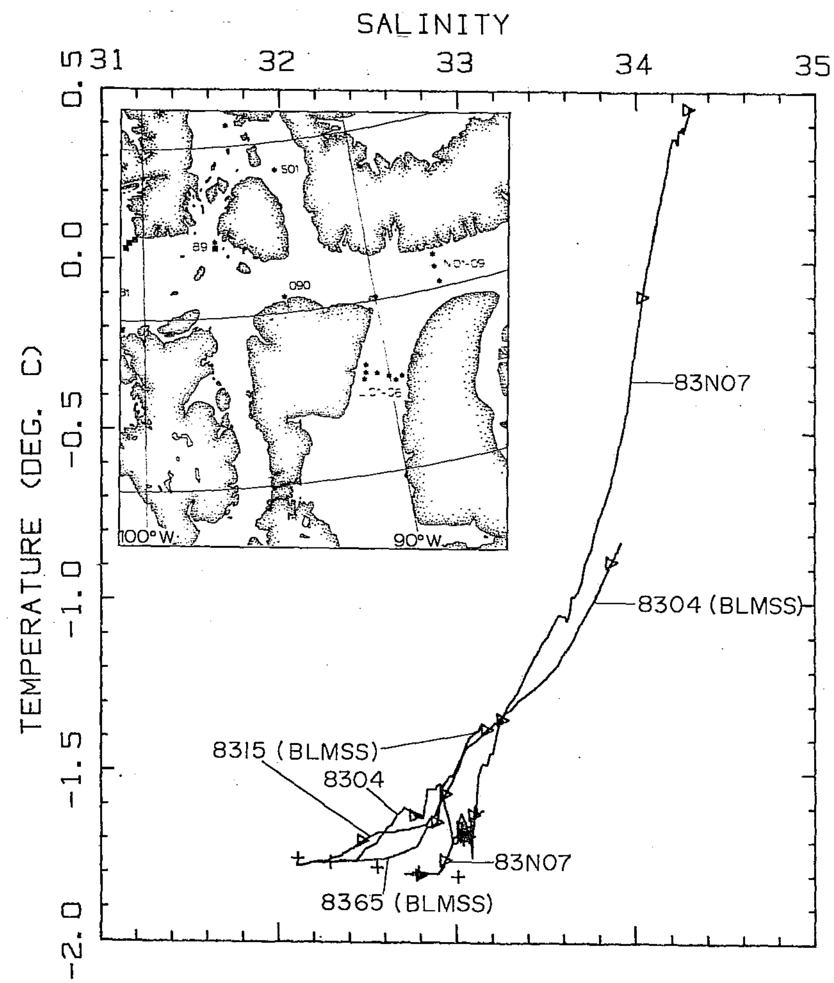

Figure 4.16 Temperature-salinity plots for 1983 stations from western Barrow Strait to Lancaster Sound.

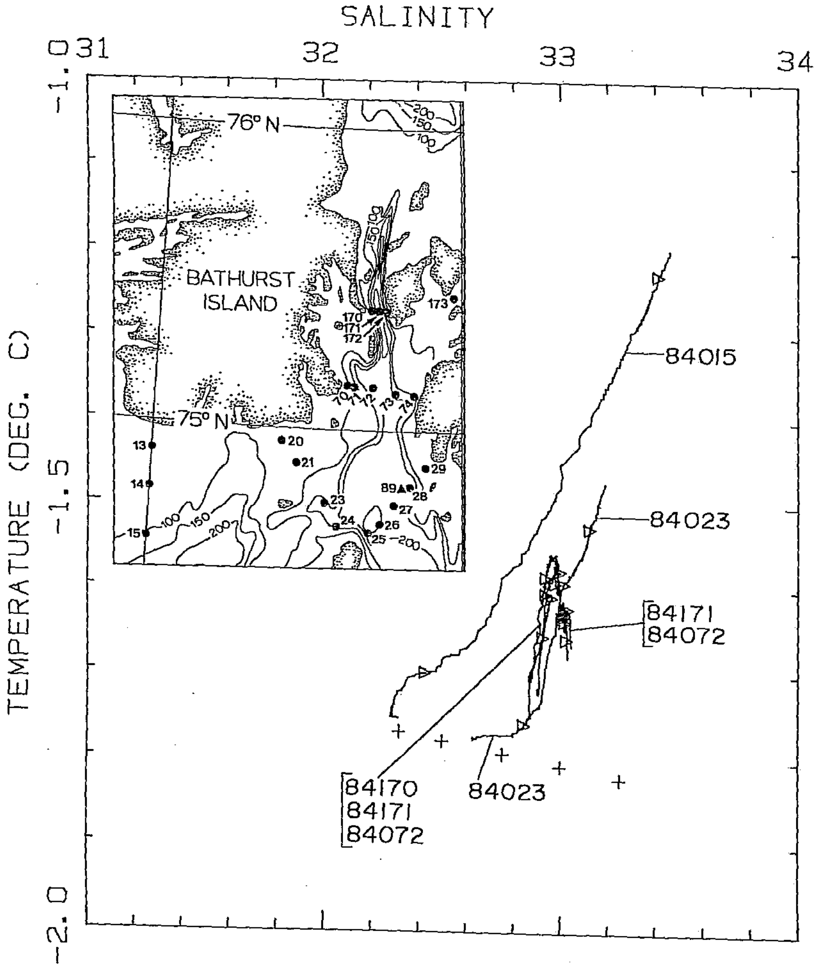

Figure 4.21 Comparison of temperature-salinity diagrams for Barrow Strait and McDougall Sound.

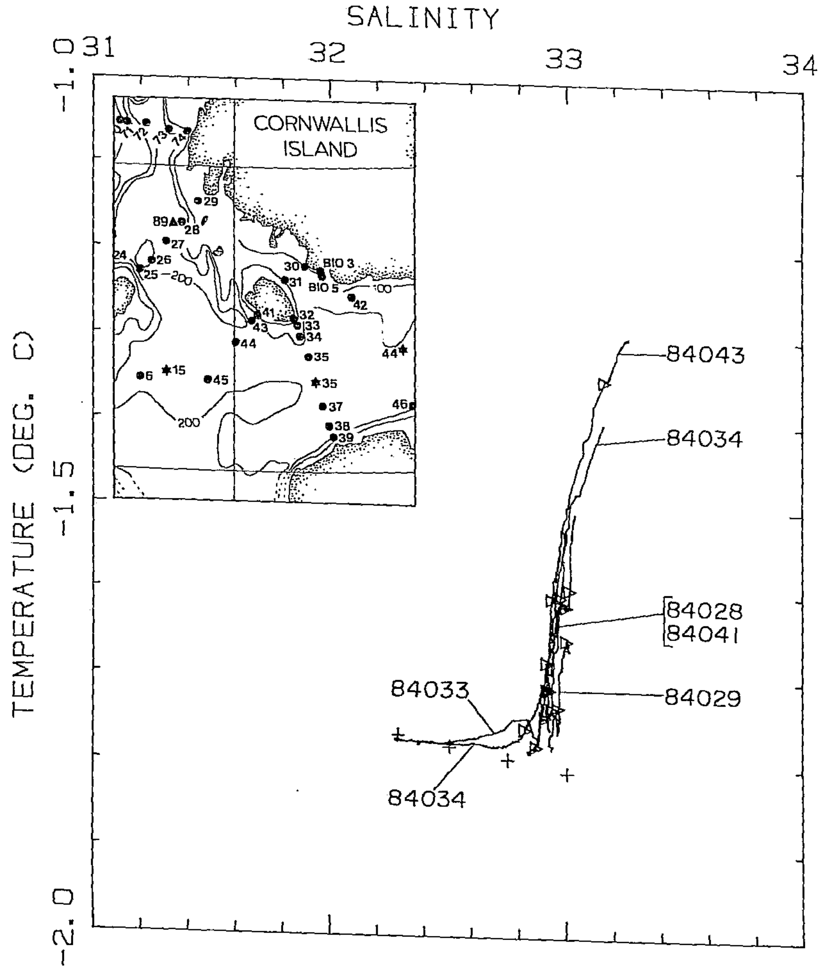

Figure 4.22 Comparison of temperature-salinity diagrams for stations in central Barrow Strait (north side).

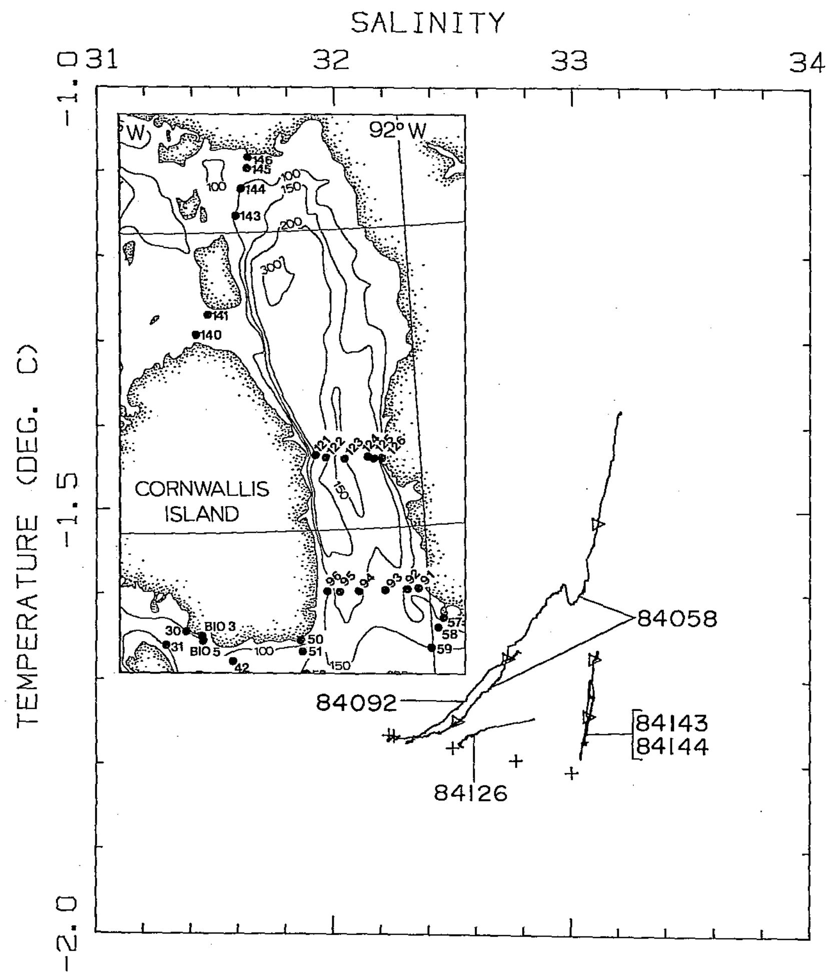

Figure 4.19 Comparison of temperature-salinity diagrams for the east side of Wellington Channel and Barrow Strait off Devon Island, 1984.

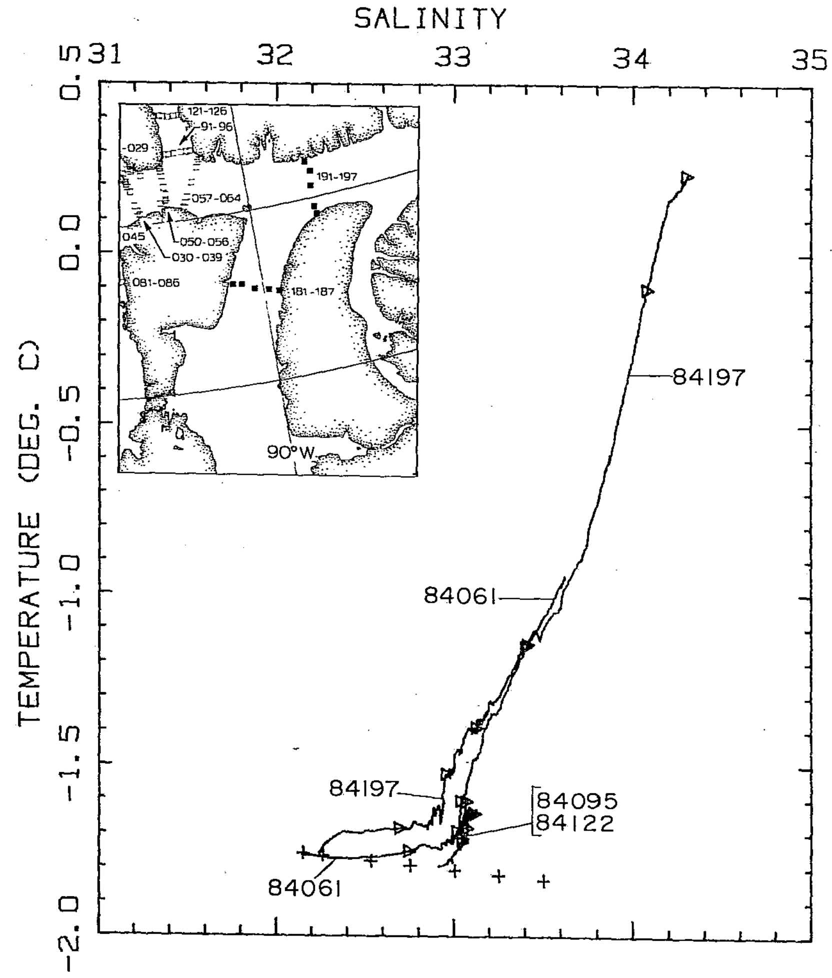

Figure 4.20 Comparison of temperature-salinity plots for 1984 stations in eastern Barrow Strait, Wellington Channel and Lancaster Sound.

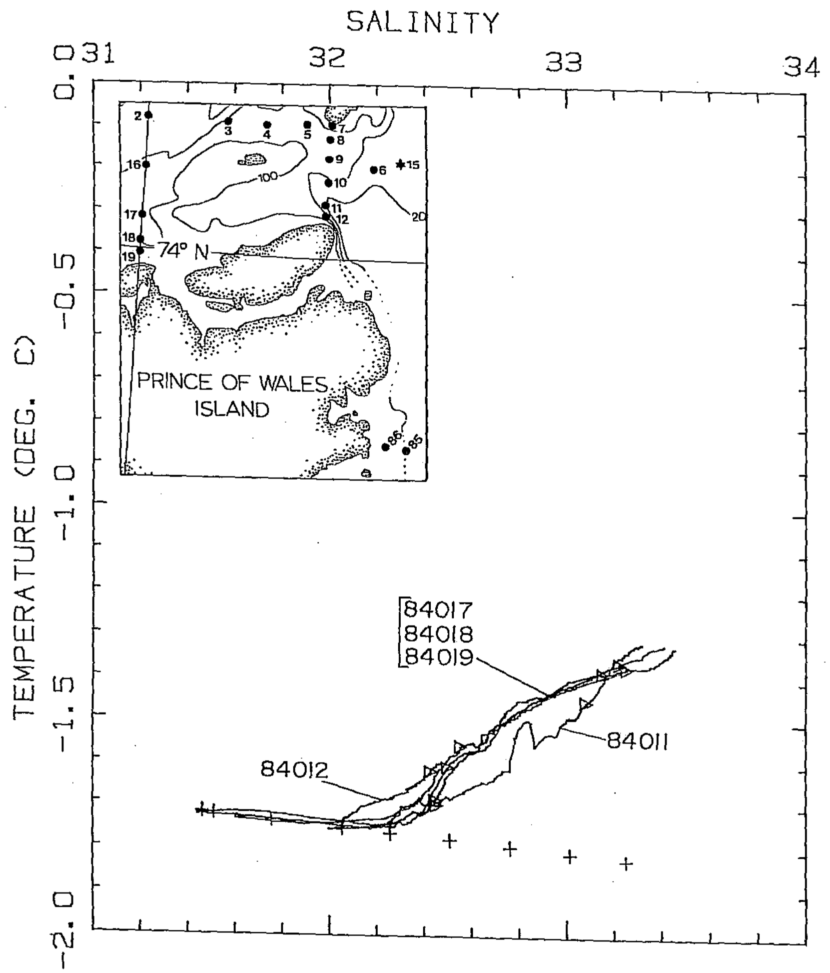

Figure 4.25 Comparison of temperature-salinity diagrams for stations north of Prince of Wales Island, 1984.

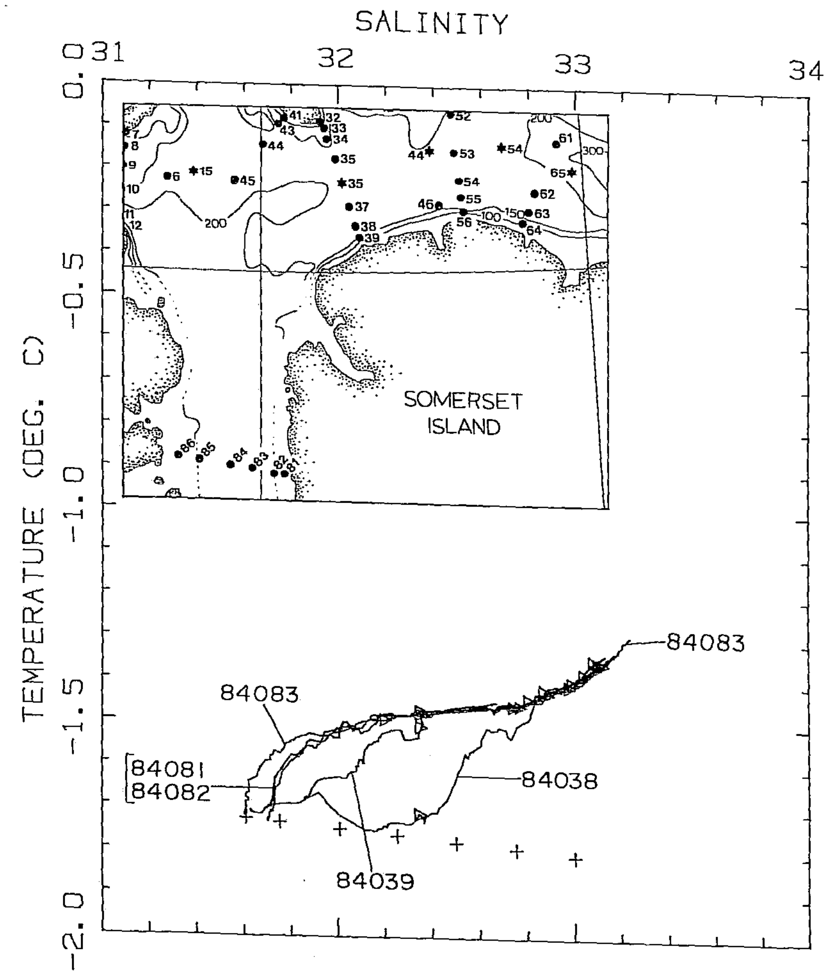

Figure 4.26 Selected temperature-salinity plots for 1984 stations in Peel Sound and northwest of Somerset Island.

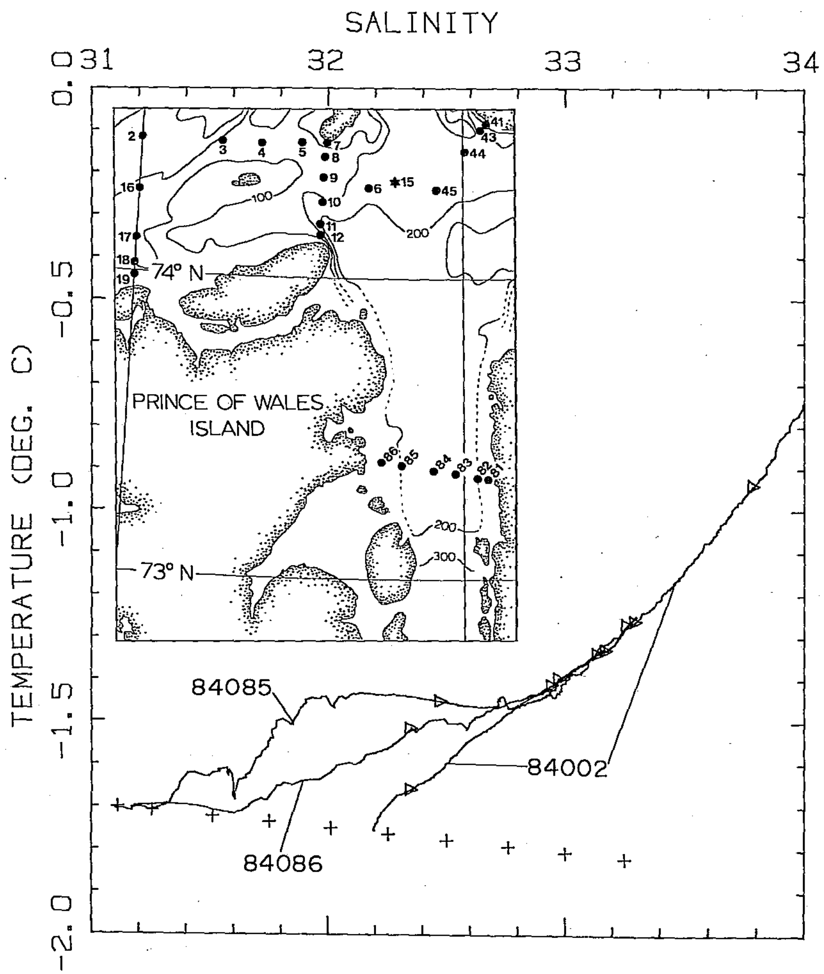

Figure 4.23 Selected temperature-salinity plots for 1984 stations Barrow Strait and Peel Sound.

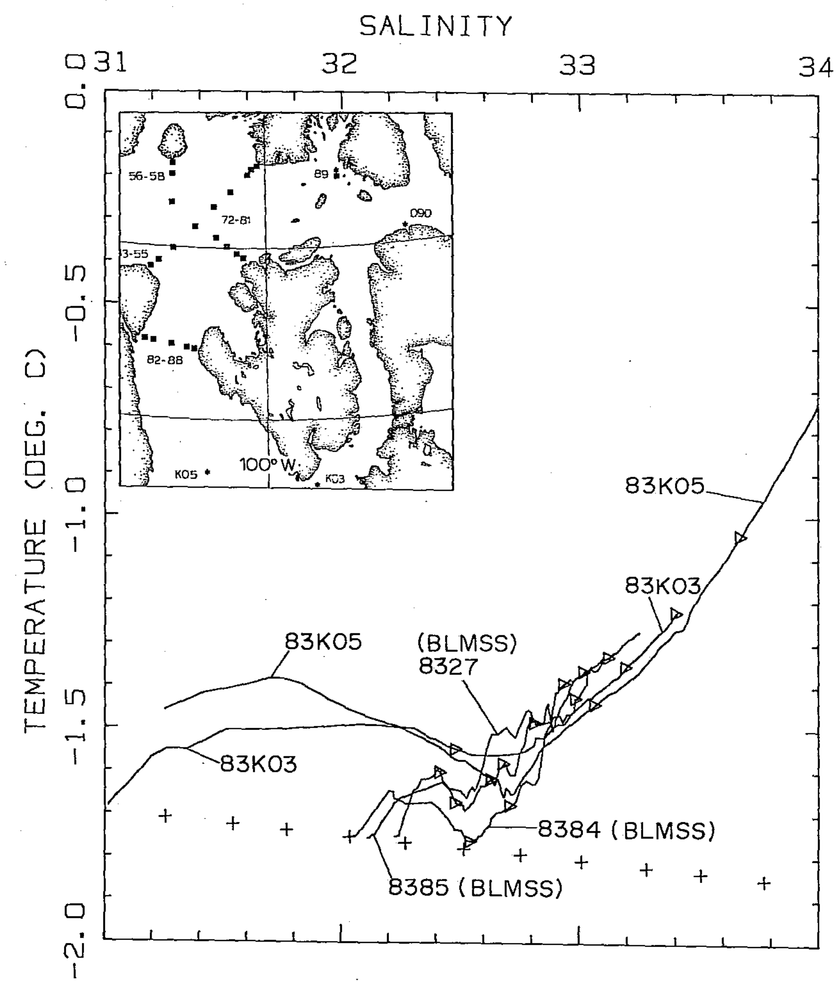

Figure 4.24 Comparison of temperature-salinity diagrams for stations in Larsen Sound, Peel Sound and Barrow Strait, 1983.

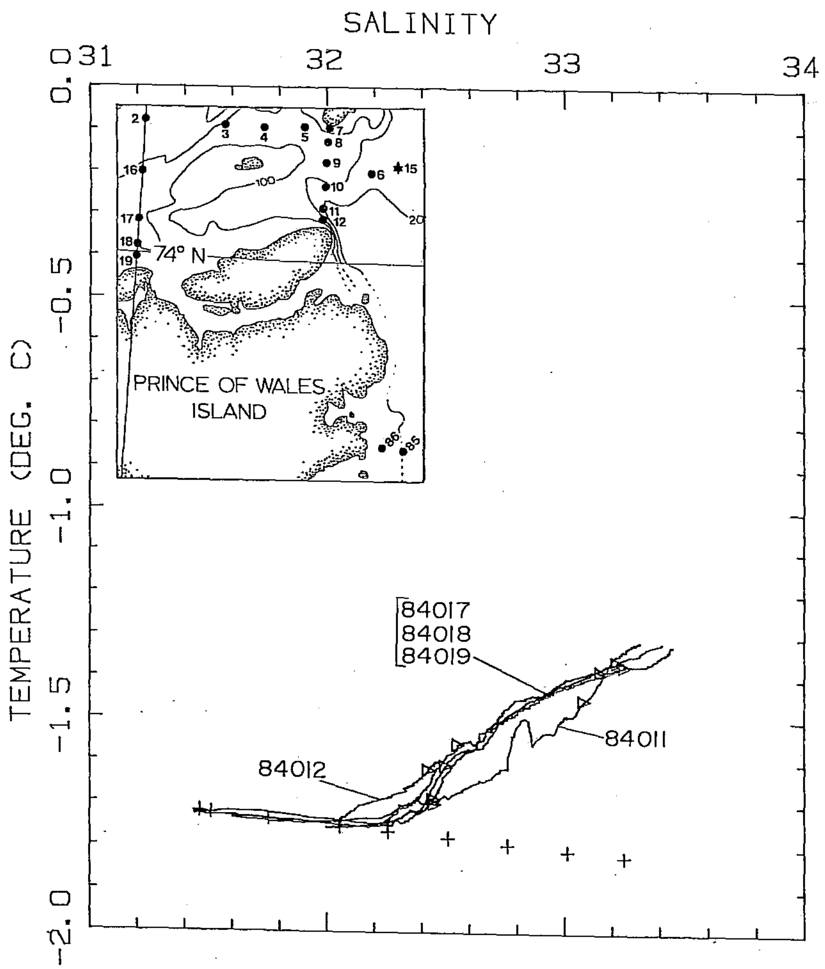

Figure 4.25 Comparison of temperature-salinity diagrams for stations north of Prince of Wales Island, 1984.

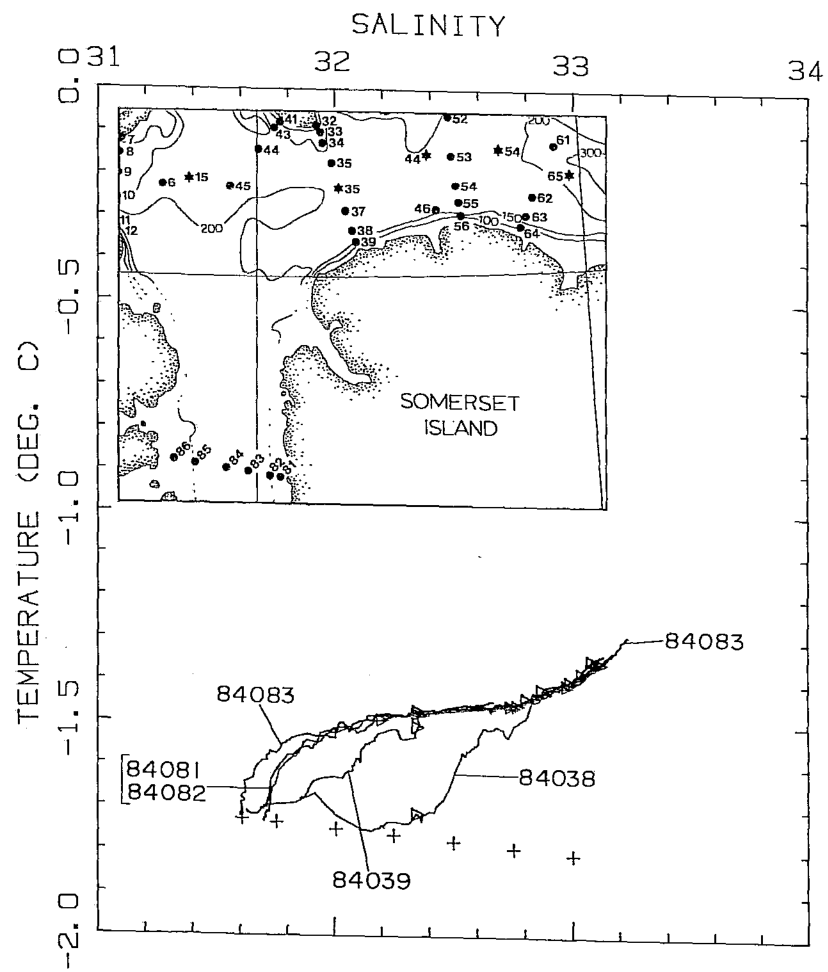

Figure 4.26 Selected temperature-salinity plots for 1984 stations in Peel Sound and northwest of Somerset Island.

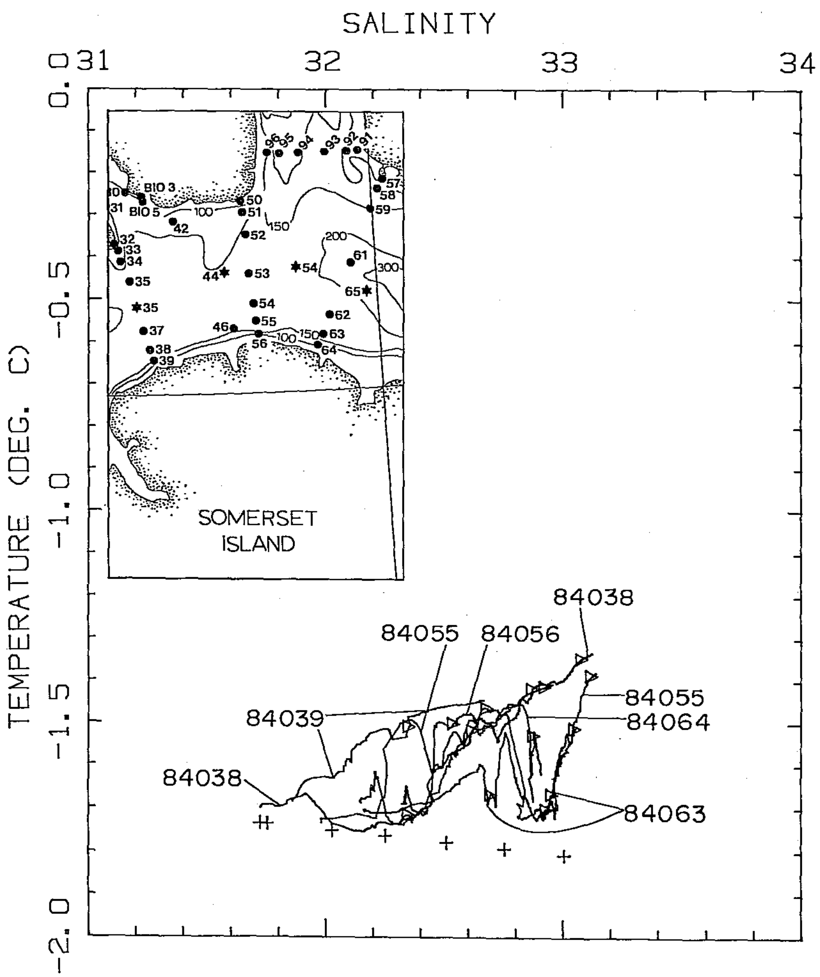

Figure 4.27 Comparison of temperature-salinity diagrams for stations along north shore of Somerset Island, Barrow Strait, 1984.

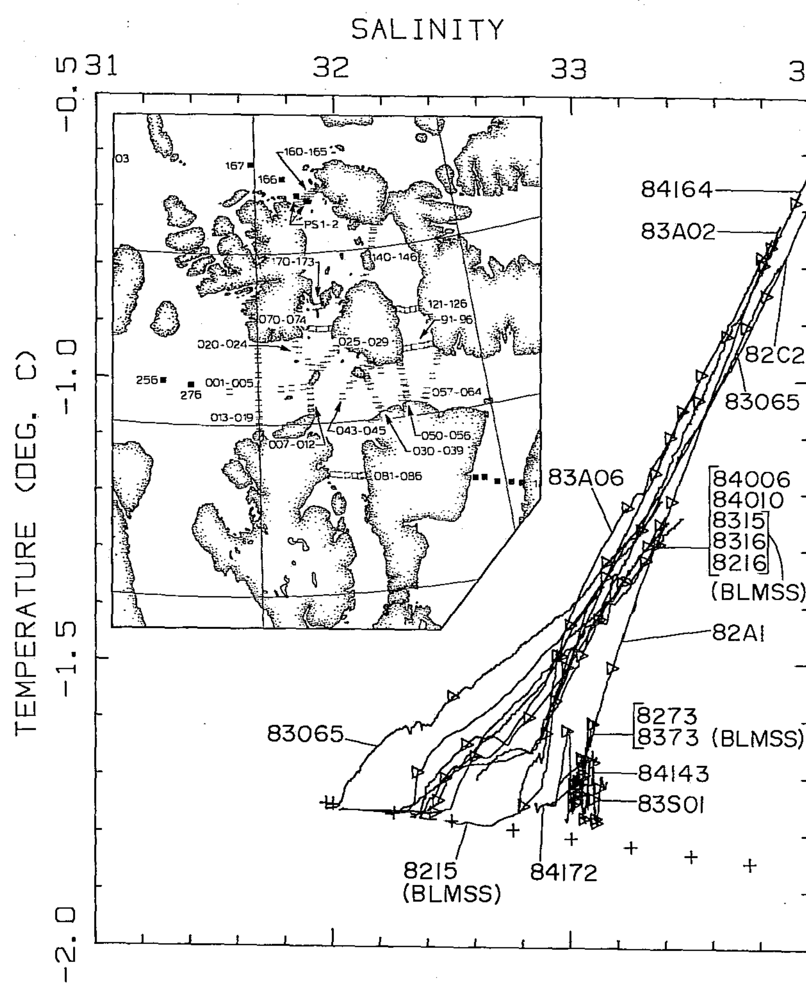

Figure 4.28 Representative temperature-salinity plots for stations in the central sills region for the years 1982 to 1984.

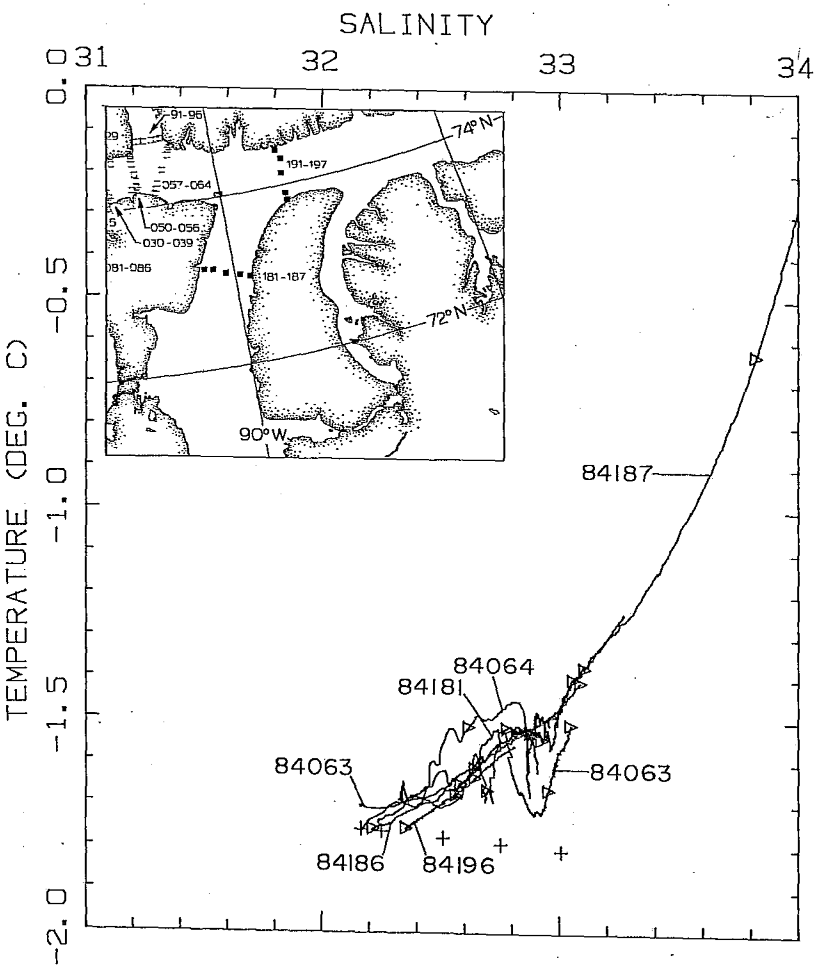

Figure 4.29 Comparison of temperature-salinity diagrams for stations in southeast Barrow Strait, Prince Regent Inlet and the south side of Lancaster Sound, 1984.

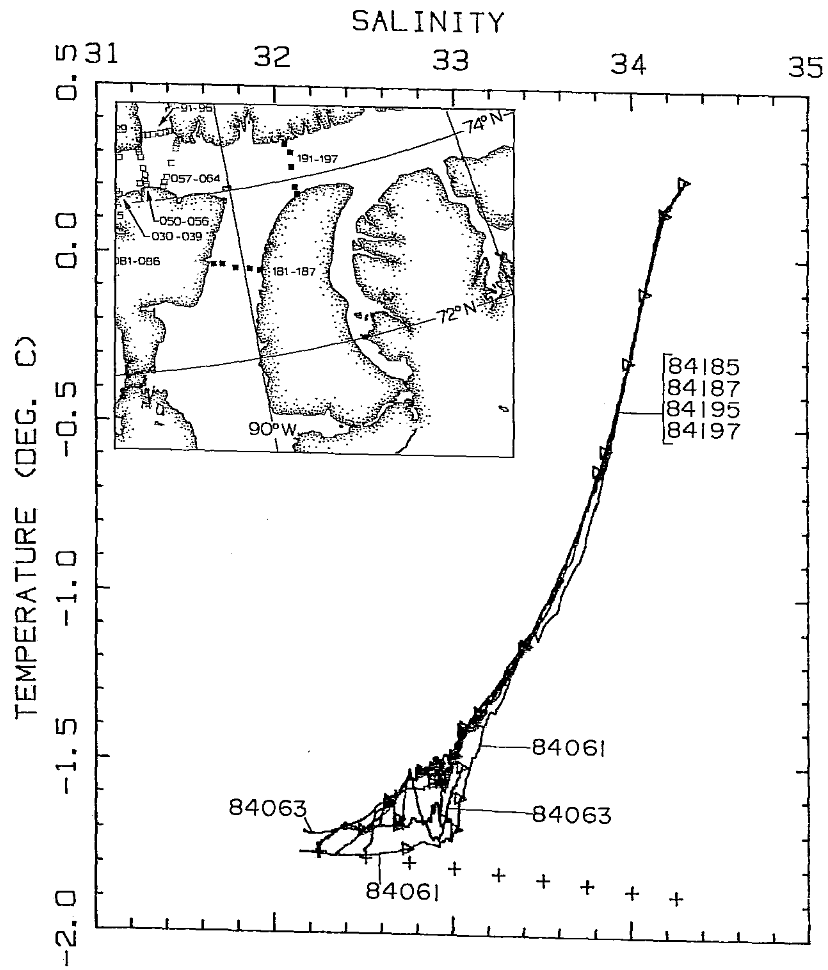

Figure 4.30 Representative temperature-salinity plots for stations in eastern Parry Channel in 1984.

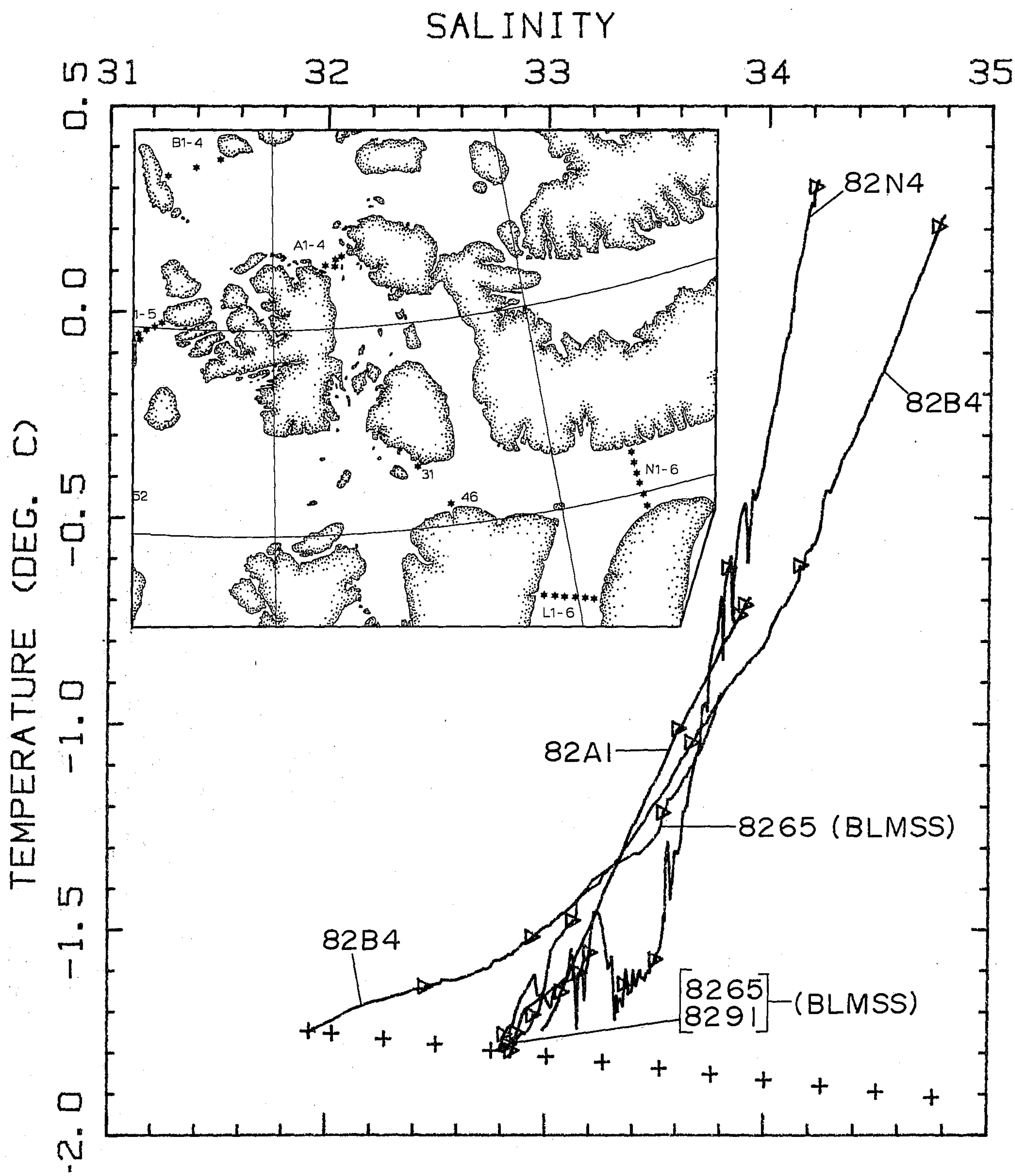

Figure 4.31 Selected temperature-salinity diagrams for stations on a transect from Maclean Strait through Wellington Channel to Lancaster Sound in 1982.

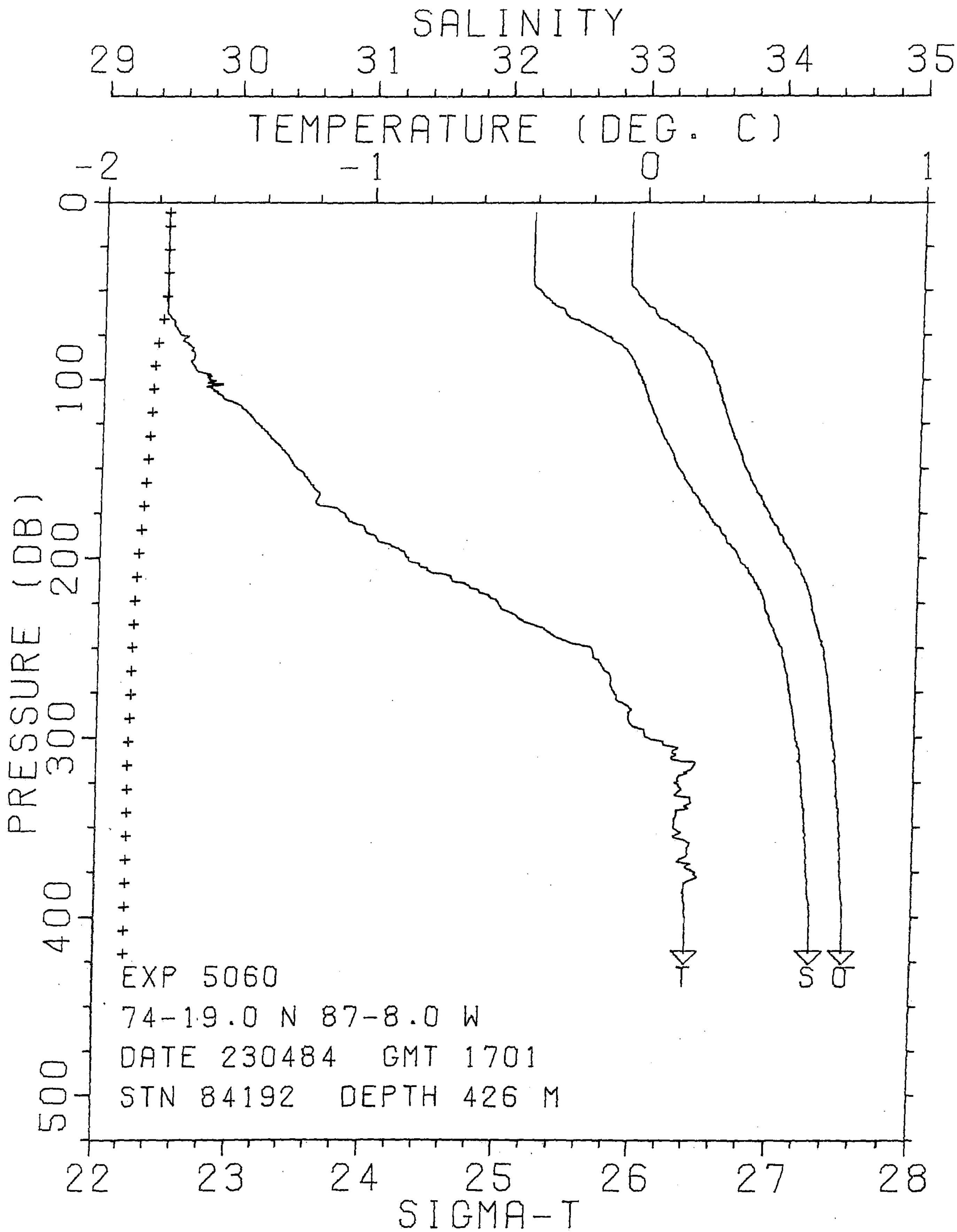

Figure 5.1 Example of "gradient reversal" fine structure from Lancaster Sound in 1984 (station 84192).

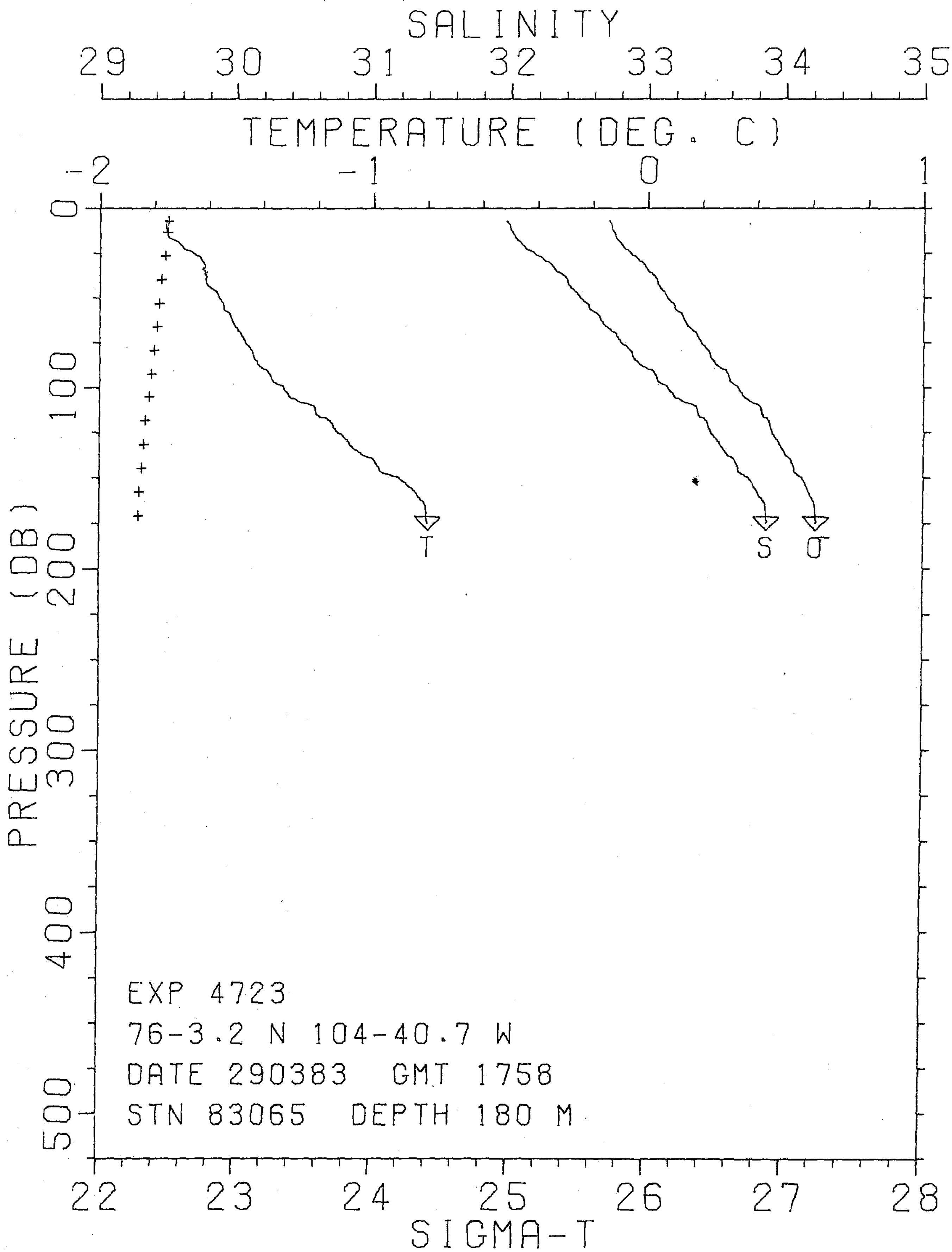

Figure 5.2 Example of "step-like" fine structure from Byam Martin Channel in 1983 (station 83065).

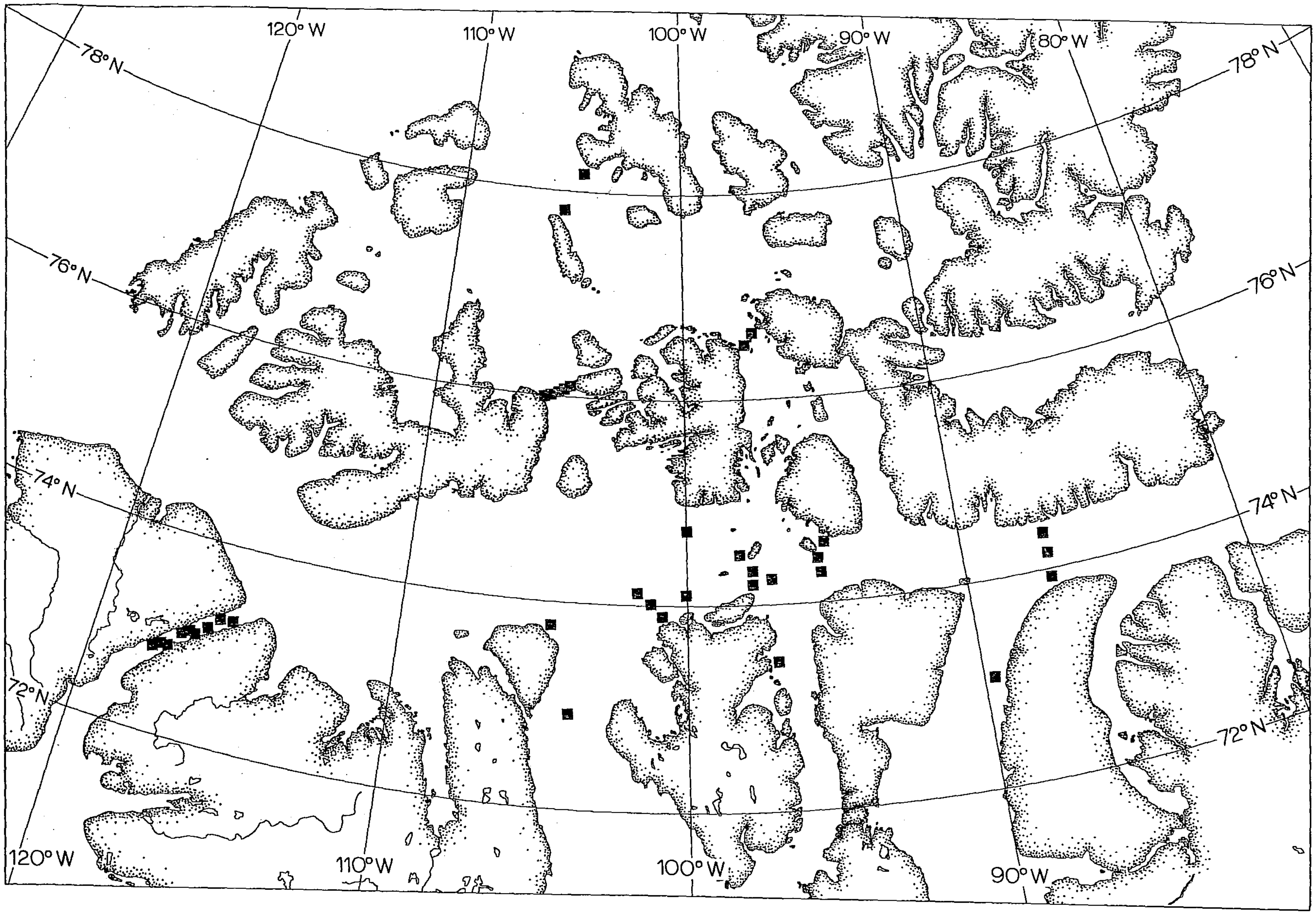

Figure 5.3 Stations with "step-like" fine structure, 1982 to 1984 (see text for explanation).

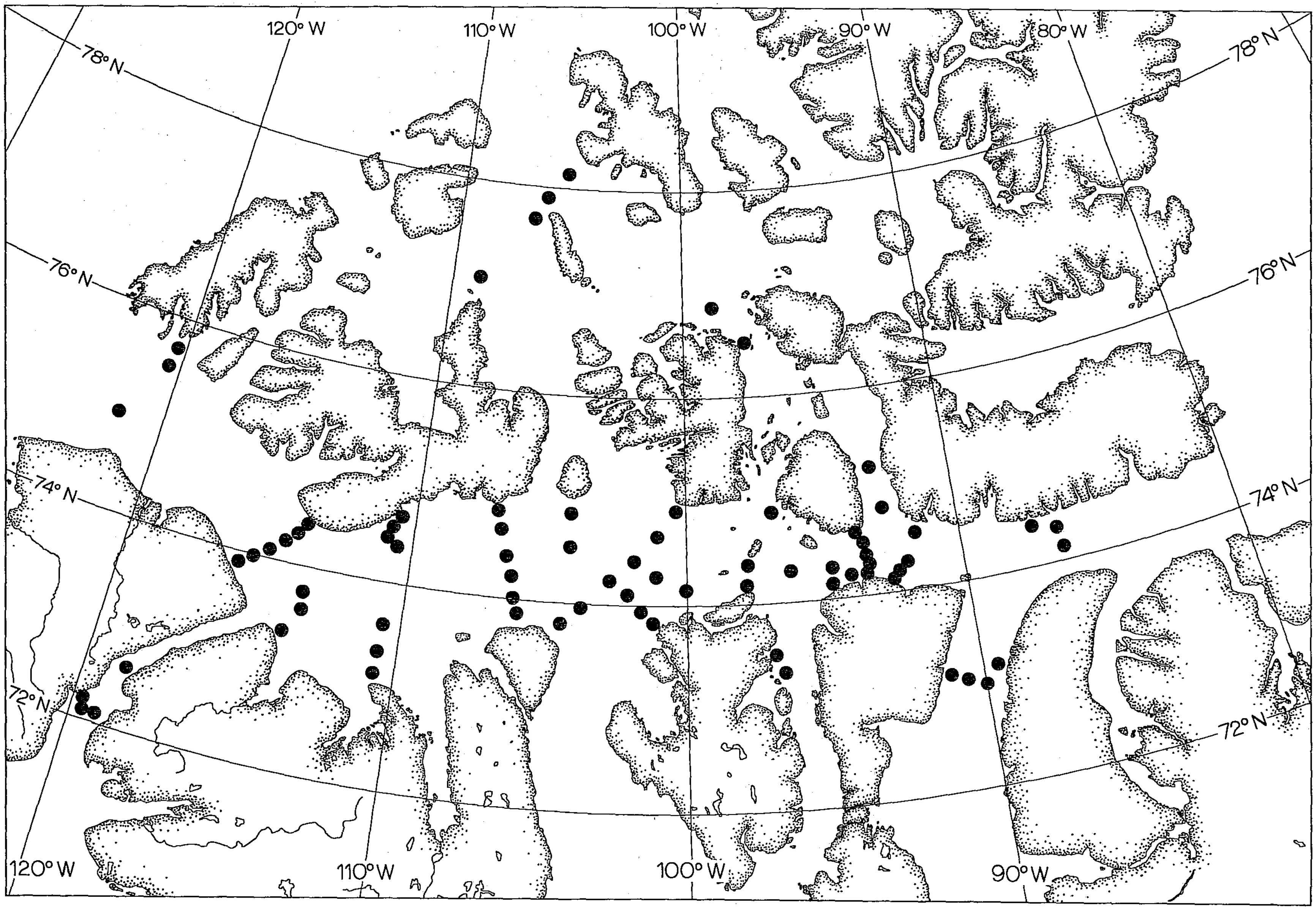

Figure 5.4 Stations with "gradient reversal" fine structure, 1982 to 1984 (see text for explanation).