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Oceanographic Conditions in the Deeper Waters of the Gulf of St. Lawrence in Relation to Local and Oceanic Forcing

by

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Changes of up to 2 ^{o}C in the temperature of the deeper waters of the Gulf of St. Lawrence have been observed over the past several decades. Temperatures were very low during the middle 1960's and have recently reached what appear to be the highest consistent values attained over the last half century. These temperature variations are very coherent along the length of the Laurentian Channel from Cabot Strait to the St. Lawrence Estuary, a distance of over 1000 km. Observed along channel gradients and phase lags are consistent with forcing which originates at the edge of the continental shelf, rather than local effects.

These variations, along with those in other parameters such as salinity and oxygen, are examined using a simple model and results indicate that they are related to variations in mixing which may be monitored by the position of the Shelf-Slope water boundary giving some predictive capability.

SPECIAL SESSION ON ENVIRONMENTAL CONDITIONS

1.0. Introduction

The Laurentian Channel (Fig 1.1) is a deep trough which extends from the continental shelf through the Gulf of St. Lawrence and into the St. Lawrence Estuary, separating the Grand Banks of Newfoundland and the Scotian Shelf. This trough, which maintains depths of over 300 m. throughout most of its 1100 km. length, provides the only deep access to the Gulf for waters of oceanic origin.

The waters of the Laurentian Channel may be roughly divided into three layers: a surface layer which displays large variations in temperature and salinity in response to vernal changes in surface heat flux and freshwater discharge, an intermediate cold layer with lesser seasonal variations which merges with the surface layer in winter, and a deeper warmer layer which undergoes only longer term changes. (Bugden, 1981) The deeper layer generally exhibits a small temperature maximum at a depth of approximately 250 m. (Fig. 1.2)

Variations in the value of the temperature maximum on time scales of several years were first noted by Lauzier and Trites, (1958) at Cabot Strait, the entrance to the Gulf of St.Lawrence. Using data from scattered cruises, some extending back as far as 1915, they described changes of order 2°C in the maximum temperature over three decades from the 1920's to the early 1950's. They assumed the deeper waters of the Laurentian Channel to be formed of a mixture of Labrador and Slope Water in essentially constant proportions, steadily supplied to the Gulf from the edge of the continental shelf. On the basis of T-S analysis they attributed the observed temperature changes to changes in the temperature of the Labrador Water.

2.0 Temperature, Salinity and Density Variations

2.1 *Temporal Variations*

To examine these temperature variations in more detail, using the additional data now available, all bottle and bathythermograph data for the Gulf of St. Lawrence were obtained from the Marine Environmental Data Service in Ottawa. For recent years, these observations were supplemented by CTD data from files at the Bedford Institute of Oceanography. Eight rectangles along the Laurentian Channel, named for nearby geographic features, were selected for data abstraction. (Fig 2.1) The rectangles were chosen as much as possible to encompass traditional hydrographic sections to ensure the presence of data. For stations within each of these rectangles, the temperature at 250 m.for each month for the years 1950 to the present was calculated.

Figures 2.2 through 2.4 show the resulting time series of temperature, salinity and sigma-theta from two rectangles separated by approximately 450 km. along the axis of the Laurentian Channel. To prepare the filtered series data gaps were first filled by linear interpolation and then a 37 month boxcar filter applied. Several significant and well resolved temperature excursions are evident in Figure 2.2.. The temperature at Cabot Strait rises from a low of about 4.1 °C in mid 1966 to what appears to be an all time high over the last quarter century of about 6.2 °C in late 1985. Also apparent in Figure 2.2 is an along channel temperature gradient of about 1.8 x 10^{-3} °C / km, the difference in the average of the two time series being 0.76 °C. The temperature signal is seen to be very coherent along the channel with phase lags suggesting an advection velocity of about 0.5 cm/s.

The salinity data shown in Figure 2.3 is not of as high a quality as the temperature data for several reasons. Salinity is more difficult to measure accurately than temperature and, in addition, a profiling instrument for salinity (conductivity) similar to the bathythermograph for temperature did not become generally available until the early 1970's. This reduces the amount and vertical resolution of the salinity data as only discrete bottle observations are available for earlier years. However, it is evident from the figure, especially the more recent measurements, that significant changes in salinity occur and that they are positively correlated with temperature. The along channel phase lag and gradient indicated in the temperature series are also clearly discernible in the salinity series.

The density, displayed as sigma-theta in Figure 2.4, suffers from the same quality problems as the salinity because of the intimate relationship between these two parameters. In spite of this, it is evident from Figure 2.4 that the density at this depth is essentially constant both in time and along the channel. The mean value of all data points shown in Figure 2.4 is 27.26 and the scatter can be explained by vertical isopycnal displacements of only about 15 m. given the average vertical density gradient at this depth. On the other hand, at the Cabot Strait average 250 m. salinity of 34.49 ‰ the previously described temperature change of 2.1 °C implies a sigma-theta change of 0.24 which is not apparent in the data. Both temporal and along channel temperature changes appear to be density compensated.

It is interesting to note that there is no apparent along channel phase lag in the density signal as is present in the temperature and salinity. This seems to suggest that the small density changes which are observed at this depth may be the result of low mode baroclinic disturbances which are poorly sampled in time and space.

2.2 *Along Channel Variations*

Figures 2.5, 2.6 and 2.7 show along channel vertical sections of temperature, salinity and density from data obtained in late November through early December 1987. The average temperature at Cabot Strait at 250 m. at this time was about 5.9 °C. Figures 2.8, 2.9 and 2.10 show along channel vertical sections of the same parameters from data obtained in November 1967. The temperature at Cabot Strait at 250 m. at this time was 4.6 °C. Comparison of Figures 2.7 and 2.10 illustrates the density compensation which accompanies this temperature change. The sigma-theta at a pressure of 250 decibars is about 27.25 in both years. Figure 2.6 shows a large volume of water with a salinity greater than 34.75 ‰ occupying the deeper portions of the Laurentian Channel in 1987, none of which was present in 1967.

There is no consistent inclination to the isopycnal lines below about 150 m. Table 2.1 shows the results of a linear least squares fit of the level of the 27.20 isopycnal to distance along the channel for several cruises when the characteristics of the deeper waters were quite different. The pressure level of the 27.20 isopycnal is consistent from year to year but the slope is inconsistent, the scatter being attributable to baroclinic activity and the non-synoptic nature of the observations. The surface layers, however, are significantly lighter in 1987, a year of relatively normal freshwater runoff, than in 1967 which followed a number of years with low freshwater runoff.

2.3 *Cross Channel Variations*

Figures 2.11 and 2.12 show typical temperature sections across the Laurentian Channel at Cabot Strait and Port Menier respectively. Apparent in these diagrams is the cross channel temperature gradient at about 250 decibars at Cabot Strait with the warmer water on the northeastern side of the channel. This temperature has been essentially erased by cross channel mixing at Port Menier (Figure 2.12).

3.0. Simple Channel Model

3.1 *Temperature Data*

To minimize the effect of vertical motion of the isotherms on the temperature observations and to provide data suitable for comparison with a vertically and laterally integrated model the average monthly temperature from 200 - 300 m. was calculated for each of the rectangles. The resulting time series from the Cabot Strait rectangle is displayed in Figure 3.1. The filter used in this case was only a 13 month boxcar filter reflecting the smoothing effect of the vertical averaging. Figure 3.2 shows the filtered time series from two other rectangles in addition to that from Cabot Strait. The gap in the South Point series reflects an absence of data over about five years. The along channel coherence, phase lag and gradient are even more evident in the smoother vertically averaged data than in the data from 250 m..

3.2 *Simple Slab Model*

Under certain conditions the temperature in a layer of constant dimensions may be described by

$$\frac{\partial T}{\partial t} + U\frac{\partial T}{\partial x} = K_H \frac{\partial^2 T}{\partial x^2} + \frac{1}{h} K_V \frac{\partial T}{\partial z} \Big/_{\text{top}} \qquad 3.1$$

Here T is the average temperature over the width and thickness of the layer , t is time, x the along channel coordinate, z the vertical coordinate, h the layer thickness and K_H and K_V the horizontal and vertical diffusion coefficients respectively. The average along channel advection velocity U, the diffusion coefficients and the vertical temperature gradient at the top of the layer are assumed to be constant and diffusion at the bottom of the layer has been ignored because of the smaller vertical gradients which exist there.

Postponing until later an examination of when Equation 3.1 might be valid the equation was cast in finite difference form and a fully explicit, space centered ,forward time step scheme (Richtmyer and Morton, 1967) used for numerical integration. The model grid was initialized using a quadratic approximation to temperatures observed near August 1952 and was subsequently driven by the Cabot Strait observations with data gaps filled by linear interpolation. An IMSL non-linear least squares routine then used the model output to determine the parameters U, K_H and K_V by fitting the model output to the observed temperatures at 6 of the other rectangles.

The parameters selected by this least squares fitting procedure are shown in Table 3.1. The vertical eddy diffusivity K_V was obtained using a vertical temperature gradient of 2.5°C/100 m.at 200 m.. A linear least squares fit to the temperature data in the depth range 150-250 m. yielded an average gradient of 2.4 °C/100 m. for the Cabot Strait rectangle and 2.5 °C/100 m. for the South Point rectangle. The root mean square error between the model output and the 164 temperature observations fitted is 0.21 °C. The success of this simple model in duplicating at least the longer time scale temperature variations over several hundred kilometers is indicated in Figures 3.3 and 3.4.

3.3 Observed Mean Current Structure in the Channel

Figure 3.9 shows average vectors over the available record length for all current meter positions from Pointe des Monts eastward. A strong cross channel shear is apparent with inward flow on the northeastern side of the channel and outward flow on the southwestern side. Recirculation appears to occur in the region of Pointe des Monts.

Figure 3.10 shows the mean currents from the observations east of Pointe des Monts plotted as average current speed over the length of the record vs distance from the axis of the Laurentian Channel. Speed, rather than along channel velocity, was chosen to eliminate the effect of local topography.Inward directions are plotted as negative numbers. The most obvious feature is the cross channel shear which is now seen to be essentially linear. A linear least squares fit of current speed vs distance from the axis of the channel using all nine available data points yields an average speed of 0.13 cm/s inward with a cross channel shear of 0.13 cm/s/km. The correlation coefficient was 0.86. Dropping the one outlying point gives an average speed of 0.71 cm/s inward with a shear of 0.10 cm/s/km. The correlation coefficient rises to 0.94. There is no obvious reason to drop this data point except that the short period of data obtained may not be as representative of mean conditions as the other points.

The mean speed from the observations agrees reasonably well with that obtained from the model, being of appropriate sign and magnitude. It is perhaps appropriate to note here that, taking the average width of the channel at the 250 m. isobath to be 75 km., the volume transport implied by a 0.5 cm/s current over a depth of 200 m. is only 0.08×10^6 m^3/s which is less than one third the upwelling from below 50 m. as calculated by Bugden (1981) to balance near surface salt and heat budgets in the Northwestern Gulf and Estuary.

This amount of inward transport presents no problems with continuity. However, the mean currents at these depths in the channel are dominated by the cross channel shear ,and the conditions when Equation 3.1 might be considered to be valid are not immediately obvious.

3.4 *Applicability of the 1-D Advection - Diffusion Model*

Since the publication of G. I. Taylor's classic papers on the dispersion of dissolved tracers in shear flows in 1953 and 1954 the concept of 'shear dispersion' has been applied to a variety of environmental flows. (e.g. Young, Rhines and Garrett (1982) The essence of shear dispersion is ; the rate of separation of fluid parcels by turbulent diffusion is greatly exceeded by the rate of separation caused by the different advection velocities experienced by the parcels as they are randomly diffused across the velocity profile. This enhanced separation may sometimes be expressed as an effective diffusion coefficient.In simple two dimensional bounded shear flow of the type which appears to occur in the deeper portions of the Laurentian Channel away from the channel ends, and ignoring for the moment the effect of vertical diffusion the analysis of Fischer et. al 1979 is applicable.

A sufficient length of time after the addition of the tracer into the shear flow (usually given as some fraction of the cross channel mixing time scale W^2/K_L , where W is the channel width and K_L is the lateral turbulent diffusion coefficient, e.g. Chatwin 1970) a balance is established between longitudinal advective transport and cross channel diffusive transport . *Subsequent to this time, an advection* - diffusion equation like 3.1 becomes applicable to the cross channel averaged tracer concentration. The effective dispersion coefficient which is generally much larger than the turbulent diffusion coefficient is given by

$$K_E = - {}^1/_W \int_0^W u' \int_0^y {}^1/_{K_L} \int_0^y u' \, dy^3 \qquad 3.2$$

where u' is the deviation of the velocity profile from the mean and y is the cross channel coordinate. For a simple linear velocity profile this becomes

$$K_E = \frac{U^2 W^2}{120 K_L} \qquad 3.3$$

where U/W is the cross channel shear.

Using the effective dispersion coefficient from the model fitting process and the shear from the linear fit to the velocity observations we obtain a lateral diffusion coefficient $K_L \simeq 322\ m^2/s$. This in turn implies a cross channel mixing time scale of the order of 200 days. At the fitted mean advection velocity of 0.5×10^{-2} m/s , the distance a tracer would be advected before 3.1 became appropriate would be less than 87 km. which is comfortably less than the 385 km. from the edge of the continental shelf to Cabot Strait. If we assume that the tracer, in this case temperature perturbations, are introduced at the edge of the continental shelf then Equation 3.1 is applicable to the cross channel averaged temperature from Cabot Strait inward and several hundred kilometres to seaward.

The actual situation is not as simple as indicated above. First of all recirculation occurs somewhere near the head of the channel. Strictly, Equation 3.1 will apply only to a semi-infinite channel. However, we will assume that the inward propagating variations are sufficiently damped out by the time Pointe des Monts is reached and thus that the channel is effectively semi-infinite. This is not strictly true, and this fact coupled with the problem that Equation 3.1 describes the evolution of the cross channel average and the data are sometimes not very representative of this average probably account for some of the discrepancies between the model results and the data. A two dimensional model is presently under development to correct some of these problems. In addition, to calculate the effective dispersion coefficients properly one must consider both vertical and horizontal shear and their variations with time. Young et al (1982) have expressed the effective dispersion coefficient in bounded shear flow as a weighted integral of the shear spectrum. For the purposes of the present work it is sufficient to note that the vertical shear is unimportant because of scaling and that the presence of low frequency variability in the cross channel shear will change some of the calculated coefficients. The essential results remain unchanged. An equation like 3.1 is applicable to the along channel low frequency variability and this, as will be shown later, permits some interesting predictive capability.

4.0 Origin of the Variations in the Deeper Waters

The Gulf of St. Lawrence receives an average annual freshwater discharge from the St. Lawrence River system of 424 km^3. This annual discharge alone exceeds the freshwater runoff of the entire eastern United States between Canada and Florida. (Sutcliffe et al, 1976) This discharge has been considered to be one of the main influences in the determination of the marine environment in the Gulf, and concern has been raised that natural and anthropogenic changes in the discharge pattern may exert less than desirable influences on the Gulf ecosystem. (Neu, 1975) Correlations have been found between freshwater discharge and landings of commercial fish species with lags appropriate to the species age at maturity. (Sutcliffe, 1973)

The phase lags displayed in Figure 3.2 leave no doubt that the temperature-salinity variations documented in this paper propagate inward from the edge of the continental shelf. The success of Equation 3.1 in reproducing these variations and the physics behind the applicability of this equation seem to indicate that, in the deeper waters at least, oceanic influences dominate on longer time scales. The variations in freshwater discharge might be expected to exert an influence on the deeper waters through entrainment which would change the mean inward velocity in the channel.

The establishment of horizontal pressure gradients through variations in the density of the surface waters might also change the pattern of the cross channel shear and changes in vertical mixing because of changes in stratification would also be expected. However, the expected magnitude of these effects is not capable of reproducing the observed property variations with their phase lags and horizontal gradients. Rather, these variations result, at most, in changes of the values of the various parameterizations in Equation 3.1 about their means.

As mentioned previously, Lauzier and Trites(1958) believed the deeper waters of the Gulf to be composed of a mixture of Labrador and Slope waters in relatively constant proportions. Variations were attributed to changes in the temperature of the Labrador water. Subsequent work has disputed the existence of Slope Water with a constant T-S correlation. (Lee,1970; Gatien,1976) Rather, in the region east of the Scotian Shelf, appreciable variations in T-S properties occur, generally related to location. The waters found in the Slope Water region below about 150 m. appear to be formed of a mixture of Labrador and Atlantic water in varying proportions, the proportion of Labrador Water generally decreasing with distance from the Grand Banks as well as from the Scotian Shelf. (Lee,1970)

Figure 4.1 shows T-S curves from two stations in Cabot Strait for three different years. One set of observations, obtained with a CTD, is from November 1985 when the average temperature from 200-300 metres from the stations displayed as well as others in the Cabot Strait rectangle was 6.1 $^{\circ}$C. (see Figure 3.1) The other two sets are from November 1966 and 1967 when the temperatures were 4.0 and 4.4 $^{\circ}$C respectively and were obtained by bottle observations. The two heavy curves shown in figure 4.1 are T-S curves for the Western North Atlantic (Armi and Bray, 1982) and the Labrador Shelf and Slope (Lazier,1982). The 200-300 metre layer generally lies between σ_T values of 26.9 and 27.4. At σ_T

values below about 27.2, the approximate value of the temperature maximum in the 1985 observations, the curves for all stations are seen to tend toward the intermediate cold layer T-S values of about 0 °C and 32.5 °/oo.(Forrester,1964)

Table 4.1 shows the fraction of Labrador Water necessary to produce the T-S values of the average of the two stations in each set assuming that mixing takes place between Labrador and Western North Atlantic waters at equal initial σ_T. At σ_T's greater than about 27.2 the fraction is quite constant within a particular year but very different between cold and warm years, ranging from slightly over 30% to more than 50%. The apparent increase of the Labrador water fraction at lower σ_T's is probably due to mixing with the intermediate cold layer, the effect of vertical mixing below this level being reduced because of smaller temperature and salinity gradients.

The effect of vertical diffusion on the mixed water as it is advected along the channel from the edge of the continental shelf to Cabot Strait has been ignored in these calculations. Assuming the same parameterizations found for the Gulf are appropriate to the Laurentian Channel outside the Gulf, this effect is expected to be negligible in the present context, amounting to less than 0.5 °C in temperature.

The Western North Atlantic water is known to exhibit an oxygen minimum at a sigma-T of about 27.15. (Fuglister, 1963) Labrador water exhibits no such minimum. The Gulf of St. lawrence also displays an oxygen minimum in the deeper waters. Data not shown here indicate that the density at which the oxygen minimum occurs and the value of the minimum in the Gulf change in the manner expected if the minimum was derived from that in the North Atlantic water and changes in the characteristics of the Gulf waters were the result of their being composed of a mixture of Labrador and North Atlantic waters in varying proportions supplied from the edge of the continental shelf.

5.0 Larger Scale Relationships

The region near the edge of the continental shelf at the mouth of the Laurentian Channel is known to be a very dynamic area. Here the waters of the Gulf Stream meet waters of Labrador origin flowing around the southeastern corner of the Grand Banks. Gulf Stream meanders and warm core eddies shed by the Gulf Stream produce large variations in exchange across the continental slope.(Smith and Petrie, 1982) Sharp fronts whose onshore-offshore position can vary over more than 150 km. on a time scale of months are known to exist in the area.(Horne,1978) Figure 5.1 shows an analysis of satellite derived sea surface temperatures which are produced at weekly intervals. The position of the surface front between the "shelf water" and" slope water" in relation to the 200 m. isobath within the region indicated on the diagram has been digitized from this and other sources for 1978 through to the present. This data, reduced to monthly averages and filtered with a 7 point boxcar filter is shown in Figure 5.2. A large excursion of about 300 km on a time scale of 5 years is apparent. Also shown on Figure 5.2 is the distance from the 200 m .isobath to the Gulf Stream prepared from the same data. The signals appear to be strongly related, but some interesting differences are clear. The distance to the front is again shown on Figure 5.3 plotted on an inverted scale and lagged 2.5 years, the advective lag using the velocity from the model and the distance from Cabot Strait to the shelf break. The cold dip observed in 1982 at Cabot Strait appears to be related to the offshore excursion of the shelf-slope water boundary. Although the 1982 cold dip is defined by only one data point at Cabot Strait, it is apparent propagating up the channel in the data from other rectangles .(see Figure 3.2) This relationship, if discovered to be robust, permits some interesting predictive capability in the sense that perhaps 1987 and 1988 should be relatively cold years at Cabot Strait.

It is unfortunate that reliable satellite imagery is unavailable prior to about 1975 and even then for regions further southwest than the Laurentian Channel entrance. Given the convoluted form of the various fronts and the path of the Gulf Stream, traditional oceanographic sections provide little information on the large scale motion of these features. Even large surveys such as "Gulf Stream '60 " take several weeks to perform and provide only one isolated data point of use in the present context. One regularly occupied section which gives at least some indication of larger scale relationships is the Halifax Section shown on Figure 5.1.(Drinkwater and Taylor,1982). This section was run southeastward from Halifax across the Scotian Shelf on an approximately seasonal basis from 1950 to 1977. Figure 5.4 shows the temperature data from the first station after the shelf break treated in the same manner as the Cabot Strait data shown in Figure 3.1. The similarity between the two filtered series is remarkable. The correlation between the two filtered data sets is 0.78 at zero lag. Although the large variance in the Halifax Section data, to be expected in this region of high horizontal gradients, precludes any firm conclusions about phase lags, this similarity does lend some support to the idea that variations in the temperatures of the deeper waters of the Gulf may be related to onshore-offshore movements of the Gulf Stream system which are reflected on a wider geographical scale.

Figure 5.5 shows the discharge from the St. Lawrence and other rivers into the St. Lawrence Estuary as monthly averages filtered with a 13 weight boxcar filter. Again the striking similarity between this time series and the Cabot Strait temperatures shown in Figure 3.1 is readily apparent. The correlation between the unfiltered Cabot Strait data points and the filtered river discharge reaches a maximum of 0.77 with the river discharge lagged 27 months, very close to the 30 month advective lag predicted from the model velocity.

As variations in advection velocity and/or vertical mixing of a magnitude sufficient to result in the observed temperature variations have been effectively ruled out by the phase lags and horizontal gradients along the Laurentian

Channel, the relationship between the river discharge and temperatures cannot be due to an estuarine circulation type response. Another possibility is that both signals are responding to large scale atmospheric forcing which involves both precipitation in the Gulf of St. Lawrence drainage basin and the vagaries of the Gulf Stream-Labrador Current system.

6.0 Summary and Conclusions

The essential results of the present study are that conditions in the deeper waters of the Gulf of St. Lawrence are the result of a slow but relatively steady advection of water from outside the Gulf in addition to some vertical diffusion. The effects of varying shear flow and lateral diffusion may be represented as an effective horizontal diffusivity. The mean advection velocity in the deeper Laurentian Channel is relatively slow and constant, the main variations in the characteristics of the deeper waters being determined by mixing at the mouth of the channel. This puts the Gulf somewhat in the role of a sampler equipped with a low pass filter sampling conditions on a constant density surface at the edge of the continental shelf.

The deeper waters of the Laurentian Channel appear to be made up of a mixture of Labrador and North Atlantic waters in varying proportions. The proportions vary on time scales of several years and may be monitored by remote sensing through the position of the Gulf Stream and/or the Shelf-Slope water boundary. The applicability of a simple advection-diffusion equation to the propagation of parameter anomalies provides some forecasting ability as the disturbances move slowly inward at the mean advection velocity.

The as yet unexplained relationship between the position of the Shelf-Slope water boundary and fresh water discharge from the St. Lawrence drainage basin puts various correlations between freshwater discharge and other biological parameters in a new perspective. The phase lag between the Cabot Strait temperatures and the freshwater discharge (within a few percent of that predicted by the objectively fitted model) admit no other explanation but that both the position of the Gulf Stream system and the *precipitation* over the *St. Lawrence* drainage basin are responding to large scale meteorological forcing. This indicates that further thought is necessary in interpreting various lagged correlations in the Gulf of St. Lawrence system.

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CRUISE	DATE	T_{250}	$P_{27.20}$	$\frac{\partial P_{27.20}}{\partial x}$
66-029	Nov 1966	4.2	225	-.030
67-008	Nov 1967	4.5	228	0.106
83-038	Dec 1983	5.7	248	0.062
87-045	Dec 1987	5.9	222	-.019

Table 2.1 Mean pressure and along channel change in pressure level of the 27.20 sigma-theta surface from. years when the temperature at 250 m. was very different.

Parameter	Symbol	Value
Advection Velocity	U	4.9×10^{-3}m/s
Horizontal diffusion	K_H	$8.2\times10^{2}m^{2}/s$
Vertical Diffusion	K_V	$2.2\times10^{-5}m^{2}/s$

Table 3.1 Parameters obtained by a least squares fit of model parameters to temperature data.

Initial σ_T	Fraction of Labrador Water		
	66-029	67-007	85-039
27.10	0.66	0.63	0.43
27.20	0.61	0.58	0.39
27.30	0.57	0.55	0.35
27.40	0.52	----	0.33
27.50	0.54	----	0.33

Table 4.1 Fraction of Labrador Water necessary to produce the average T-S characteristics for each cruise assuming mixing takes place at equal initial σ_T.

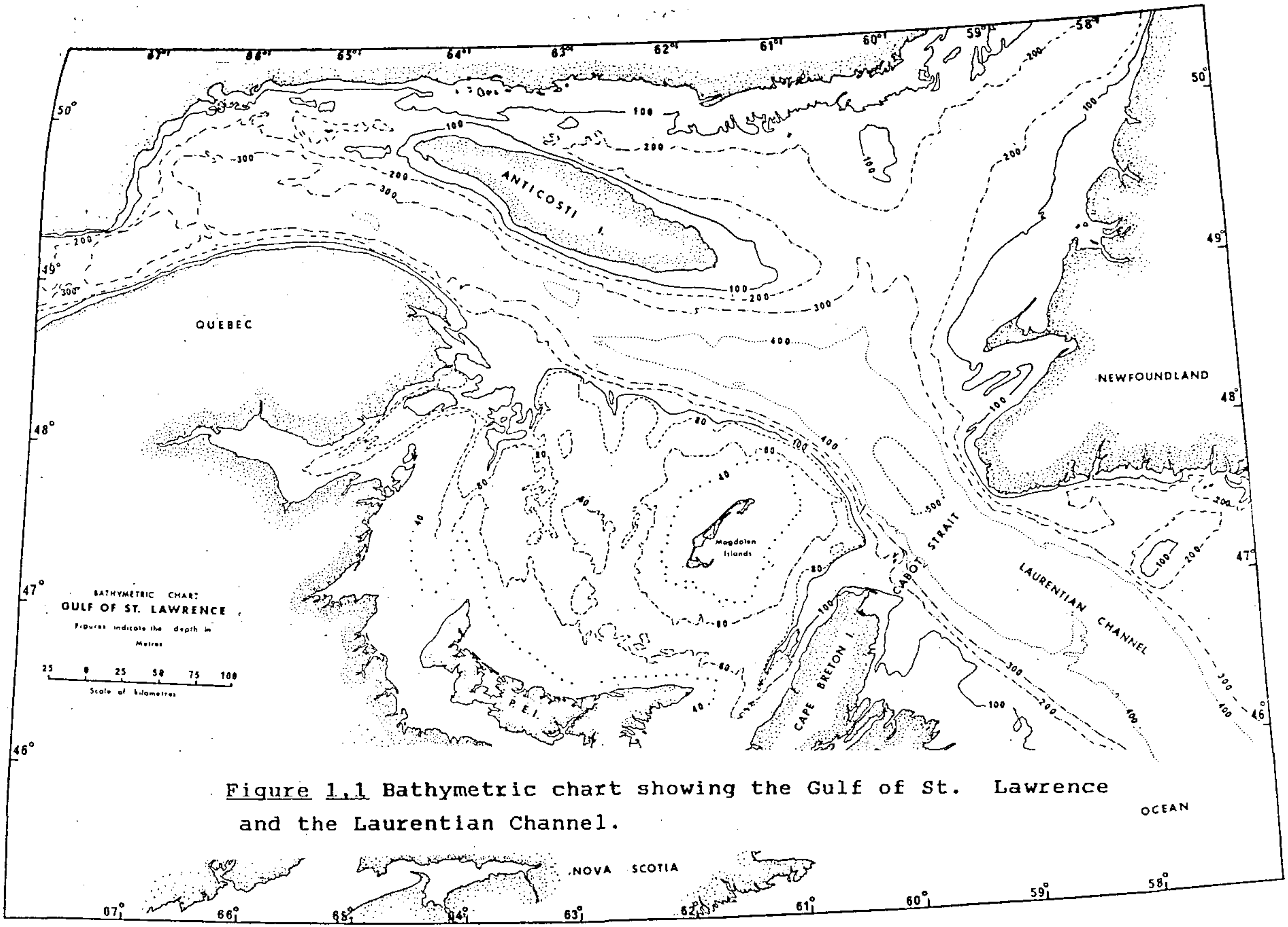
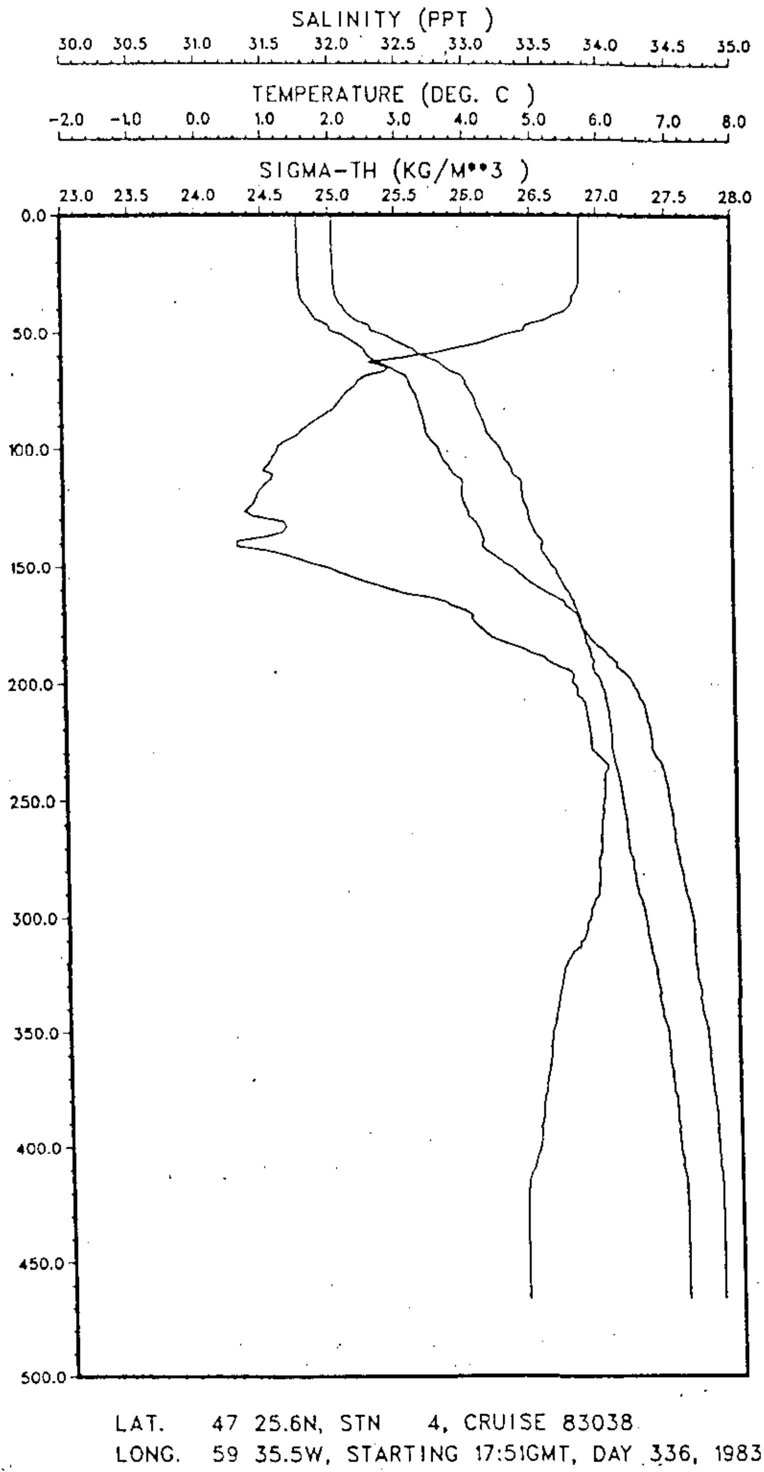
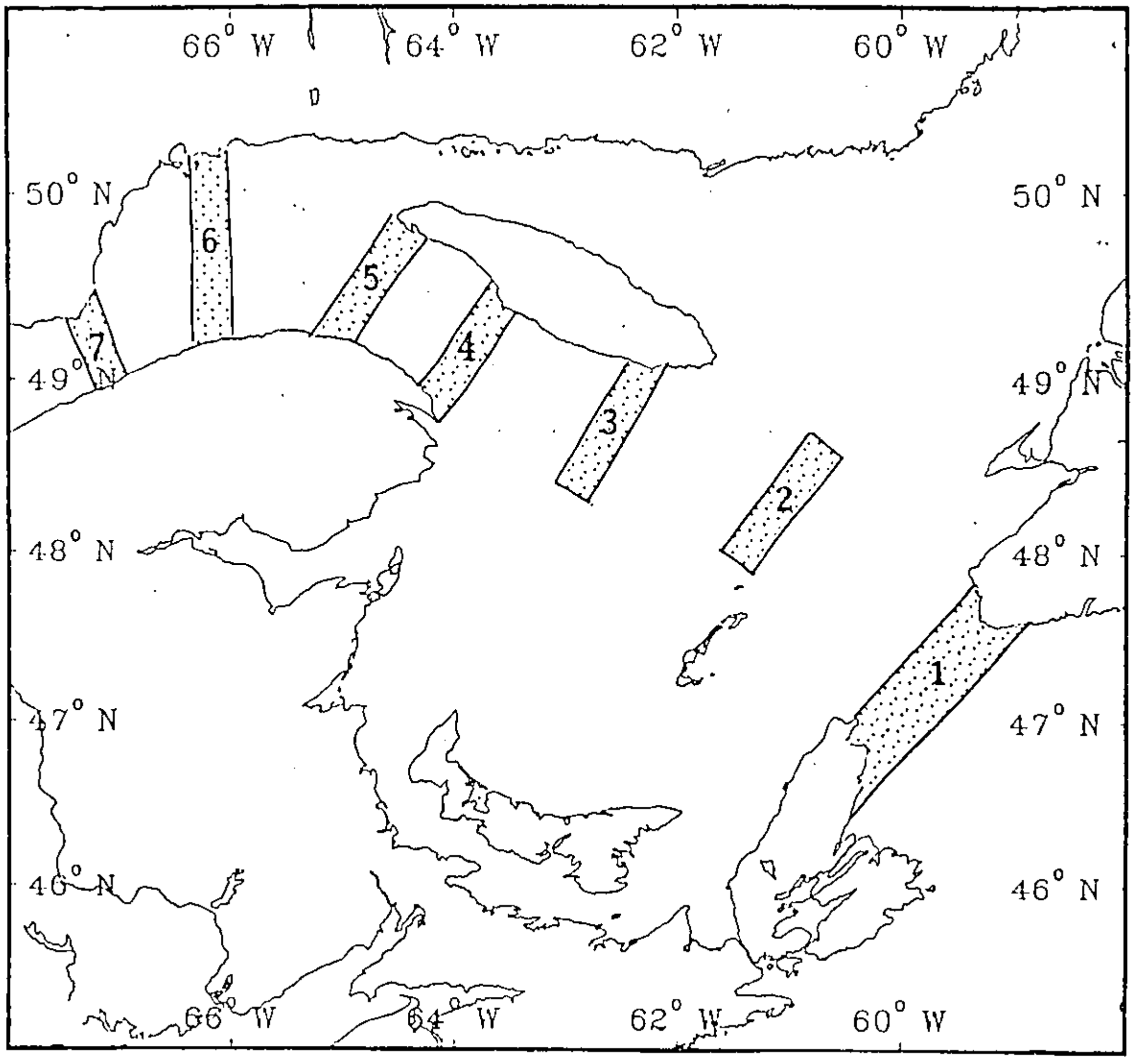


Figure 1.1 Bathymetric chart showing the Gulf of St. Lawrence and the Laurentian Channel.

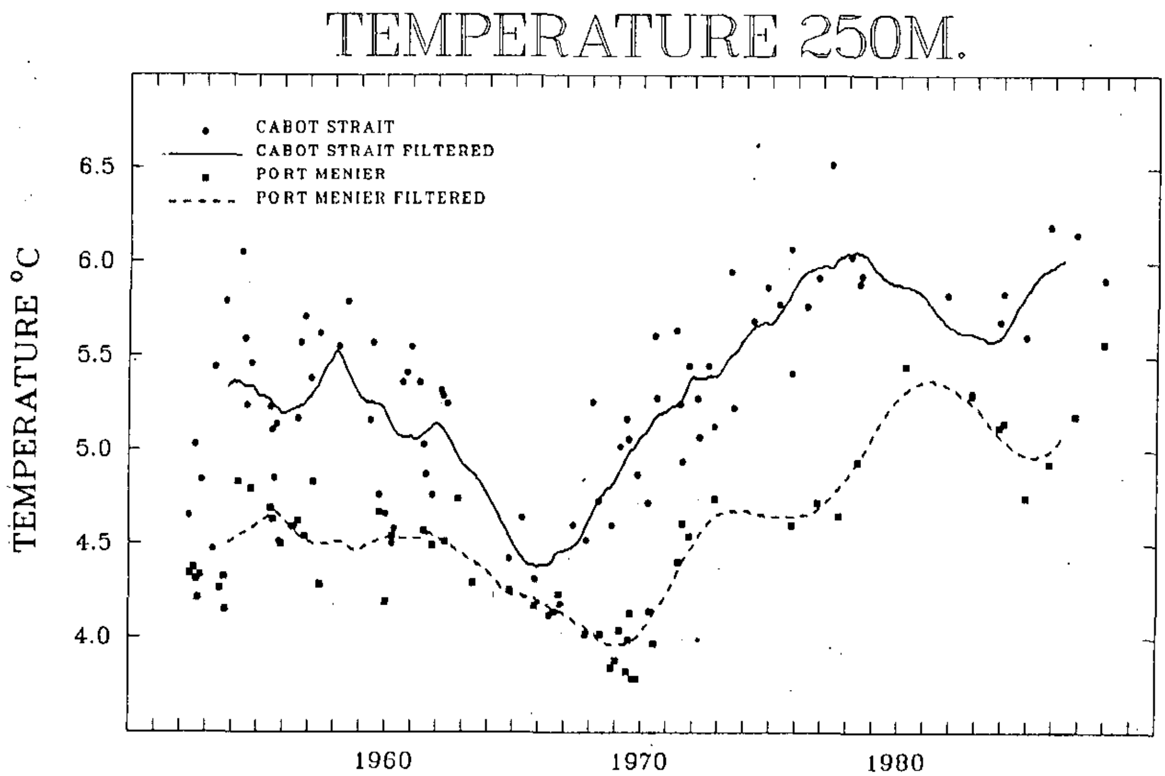


<u>Figure 1.2</u> Typical temperature, salinity and density profiles from the Laurentian Channel.

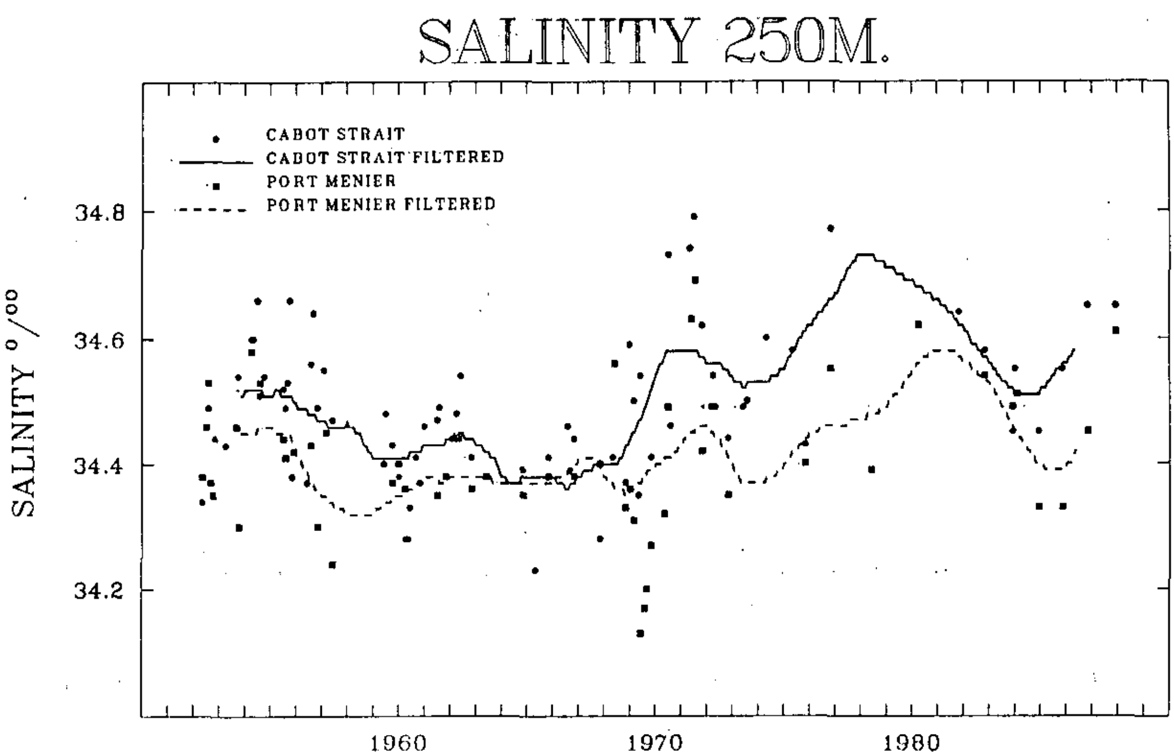


1.	Cabot Strait	385 km.
2.	Esquiman	530 km.
3.	South Point	652 km.
4.	Southwest point	776 km.
5.	Port Menier	837 km.
6.	Sept Iles	957 km.
7.	Pointe Des Monts	1163 km.

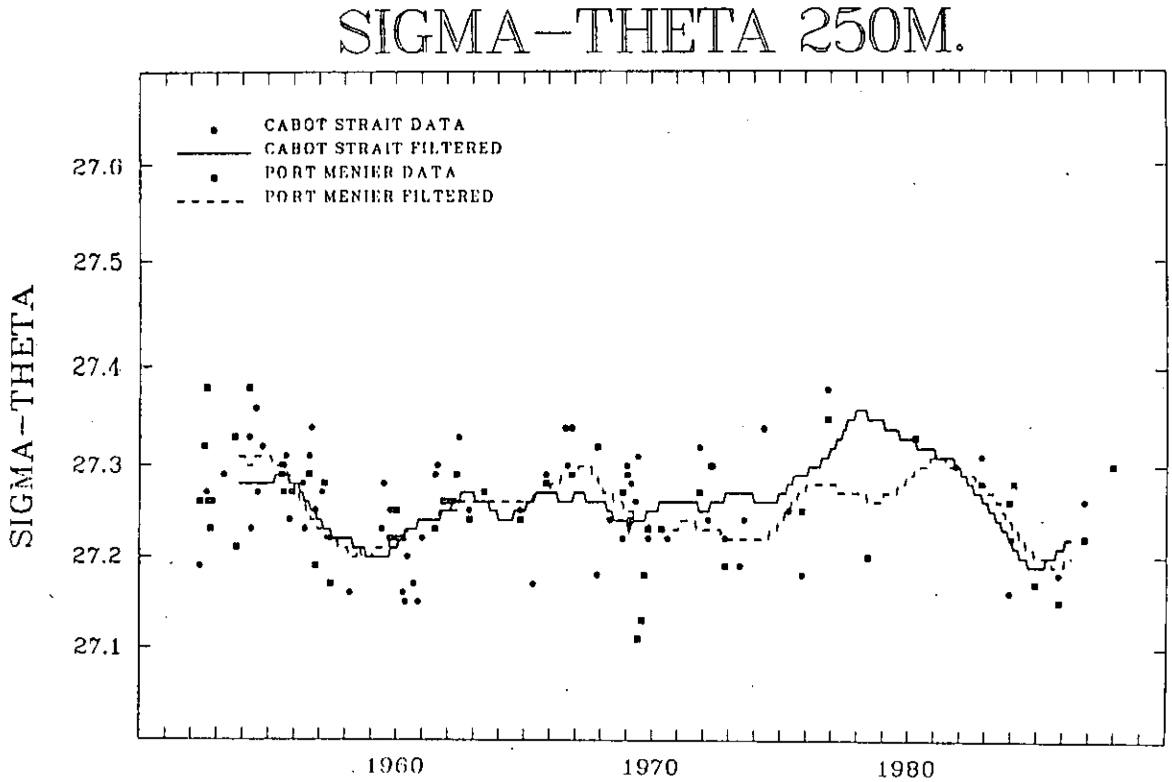
Figure 2.1 Data abstraction rectangles and distances from the edge of the continental shelf



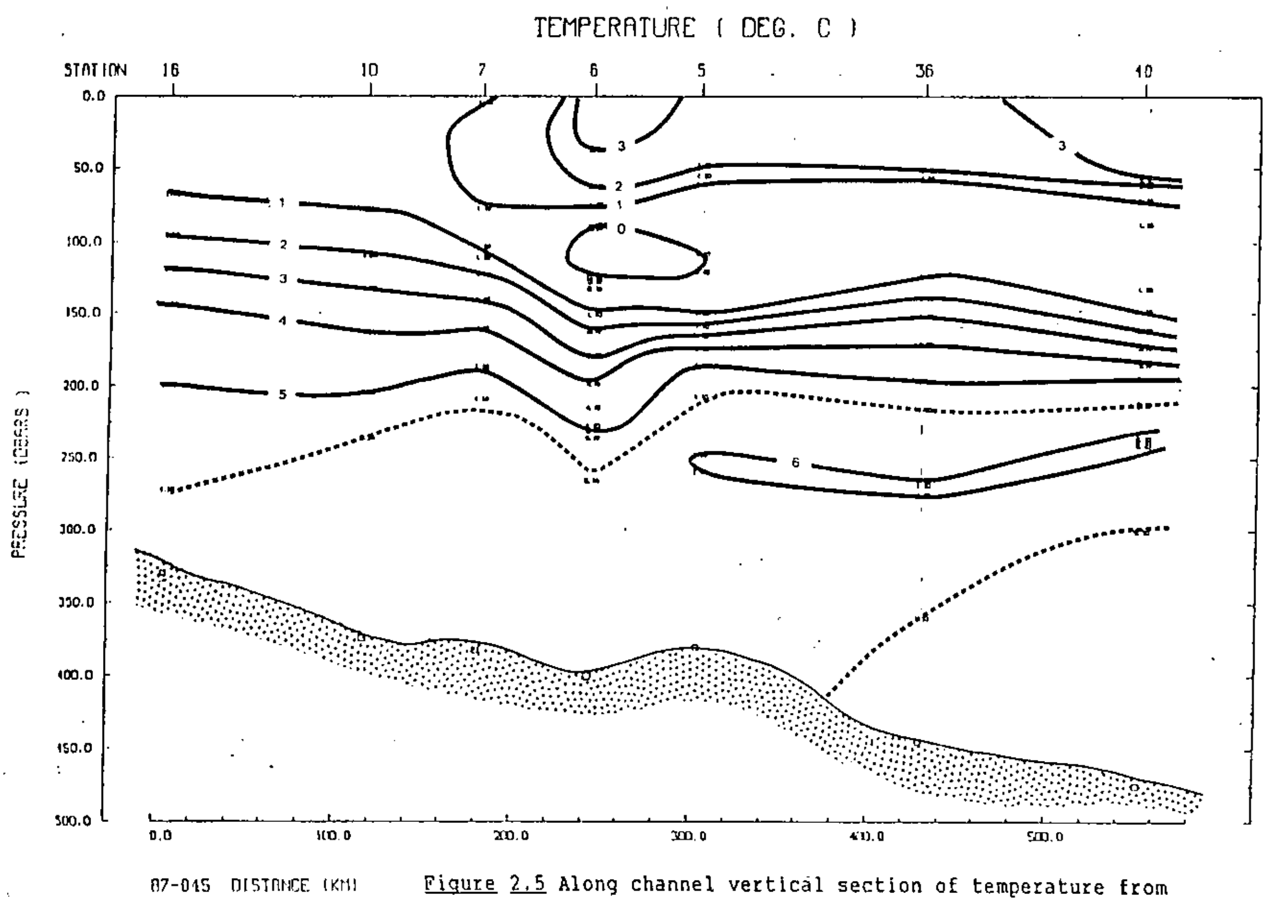
<u>Figure 2.2</u> Temperature at a depth of 250 metres from two rectangles in the Gulf of St. Lawrence separated by 430 km.. Filter is described in the text.



<u>Figure 2.3</u> As Figure 2.2, but salinity.



<u>Figure 2.4</u> As Figure 2.2, but potential density. The steps in the filtered data are caused by the fact that these data are considered accurate to only two decimal places.



<u>Figure 2.5</u> Along channel vertical section of temperature from November-December 1987.

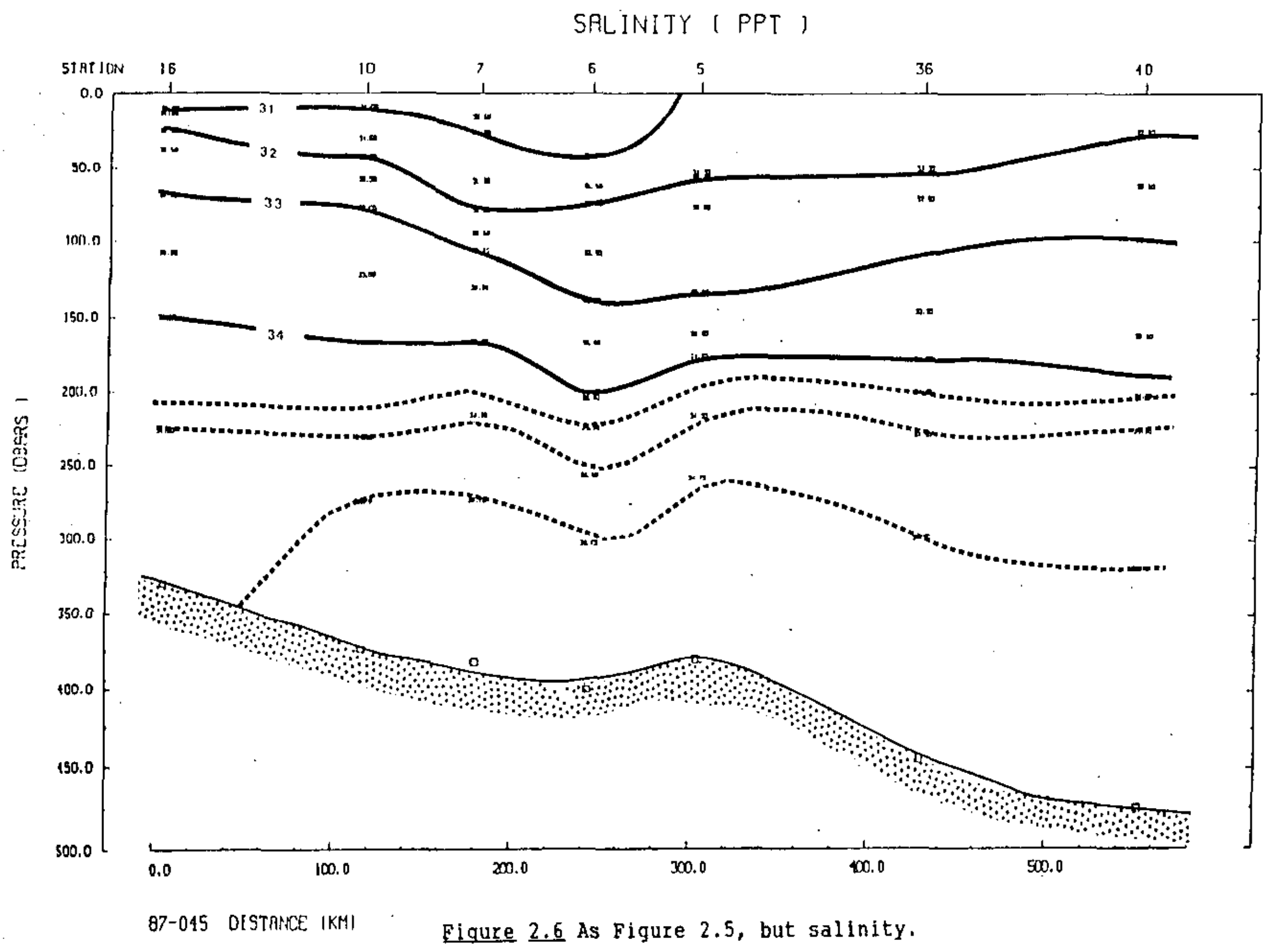


Figure 2.6 As Figure 2.5, but salinity.

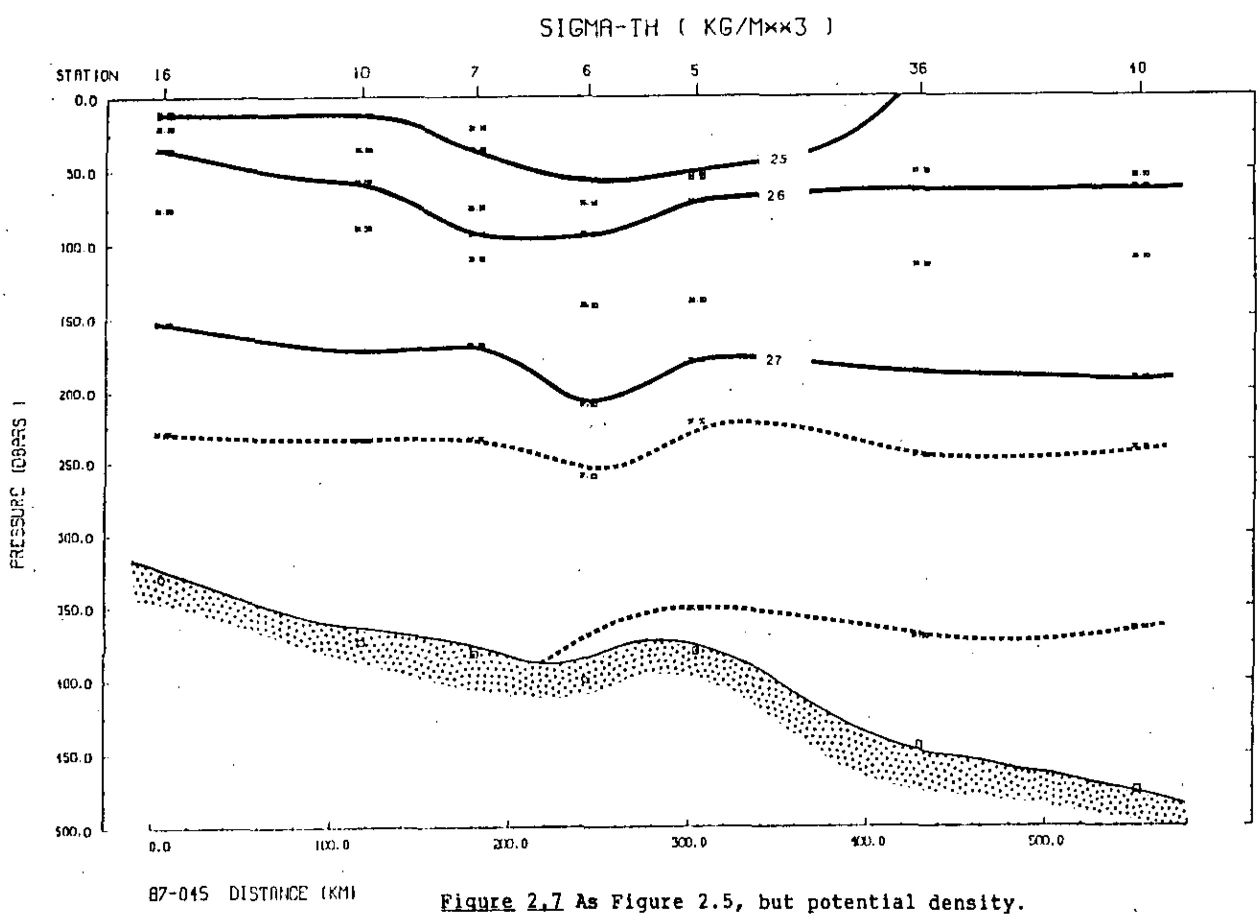


Figure 2.7 As Figure 2.5, but potential density.

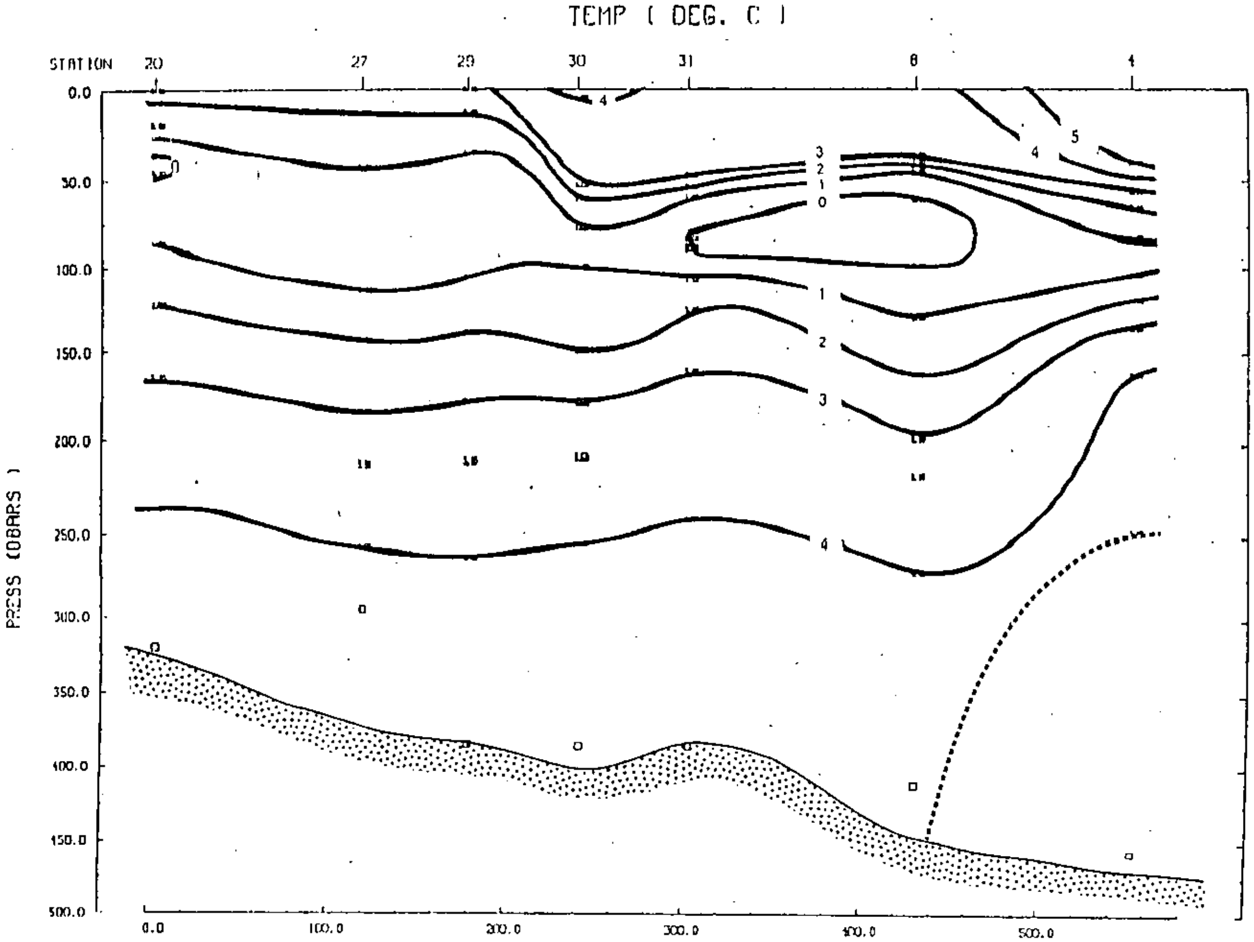
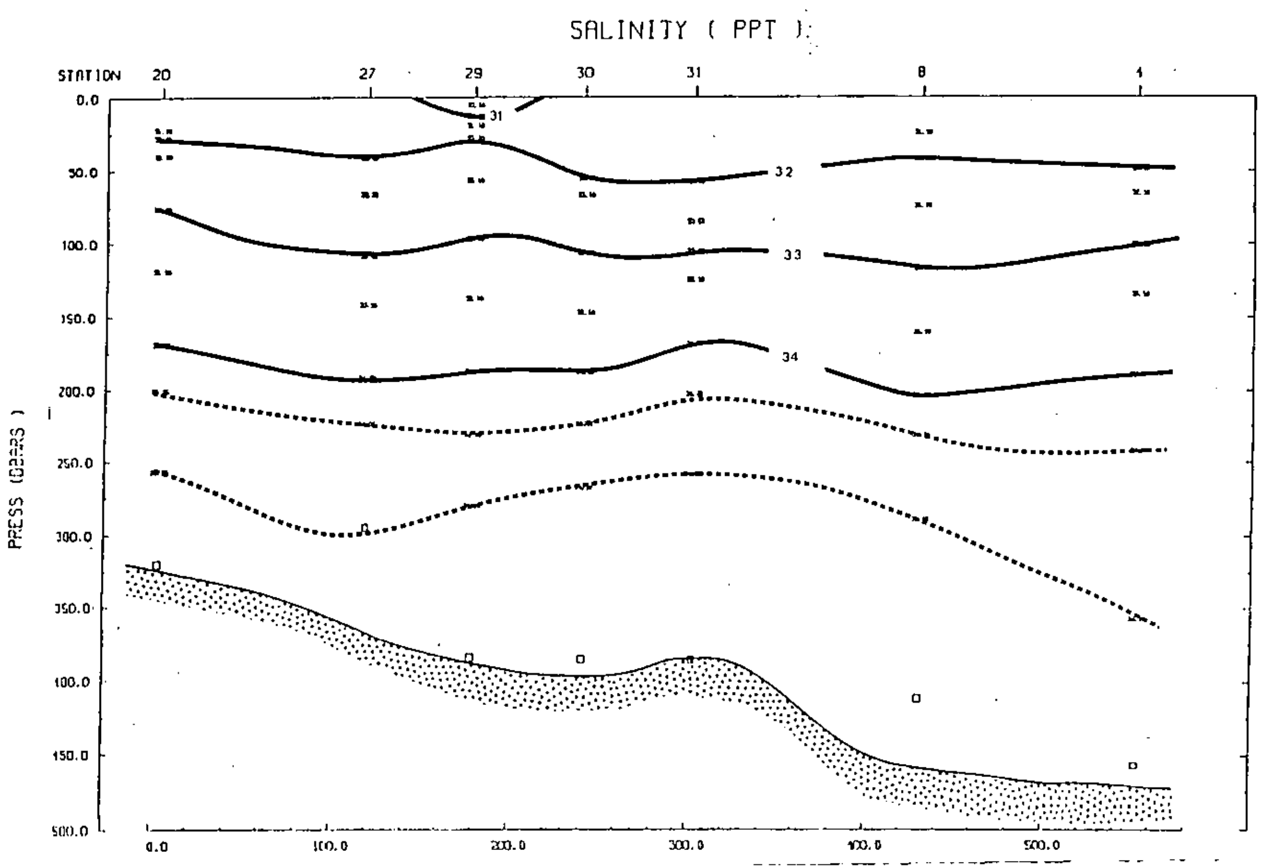
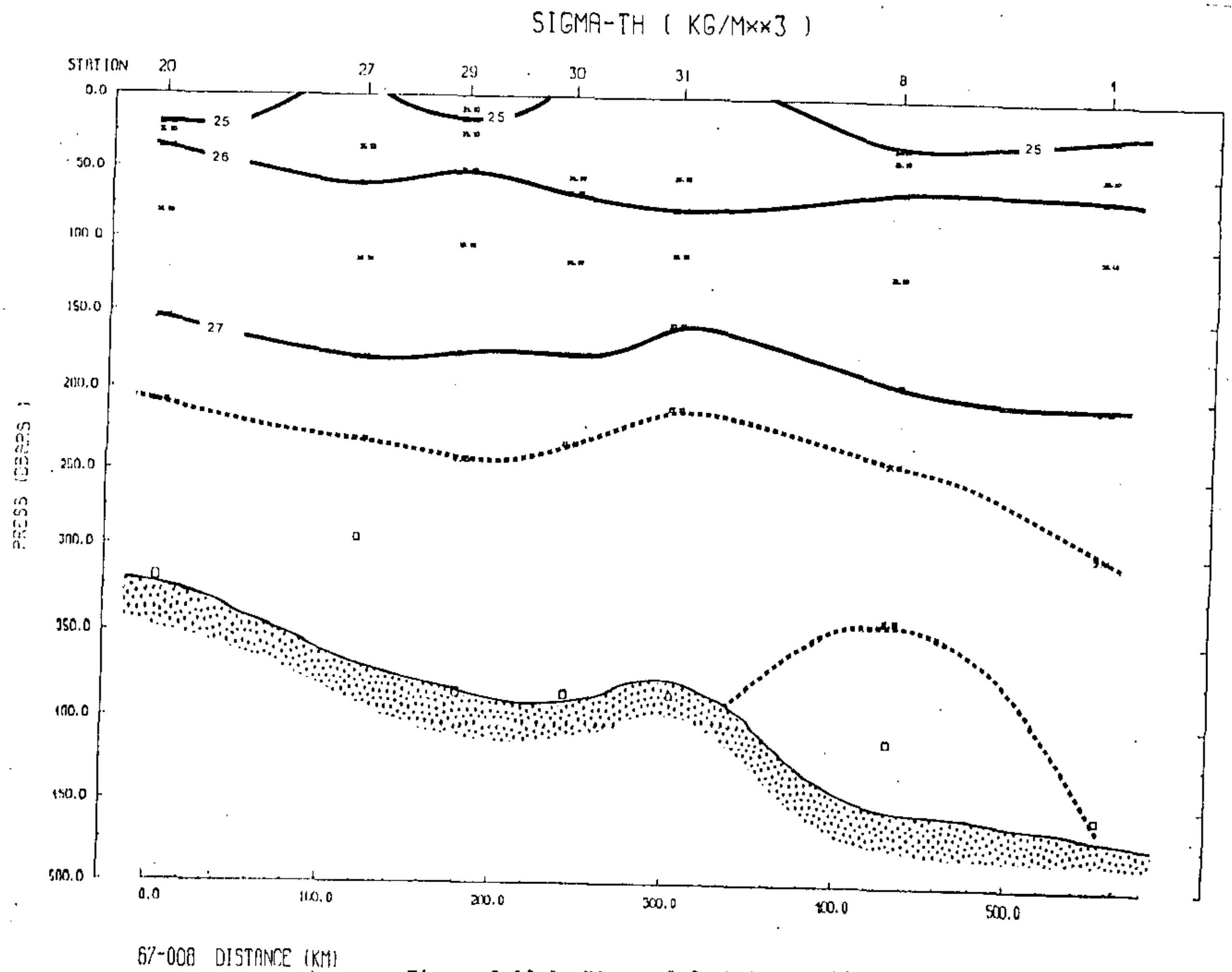


Figure 2.8 Along channel vertical section of temperature from November 1967.





<u>Figure</u> <u>2.10</u> As Figure 2.8, but potential density.

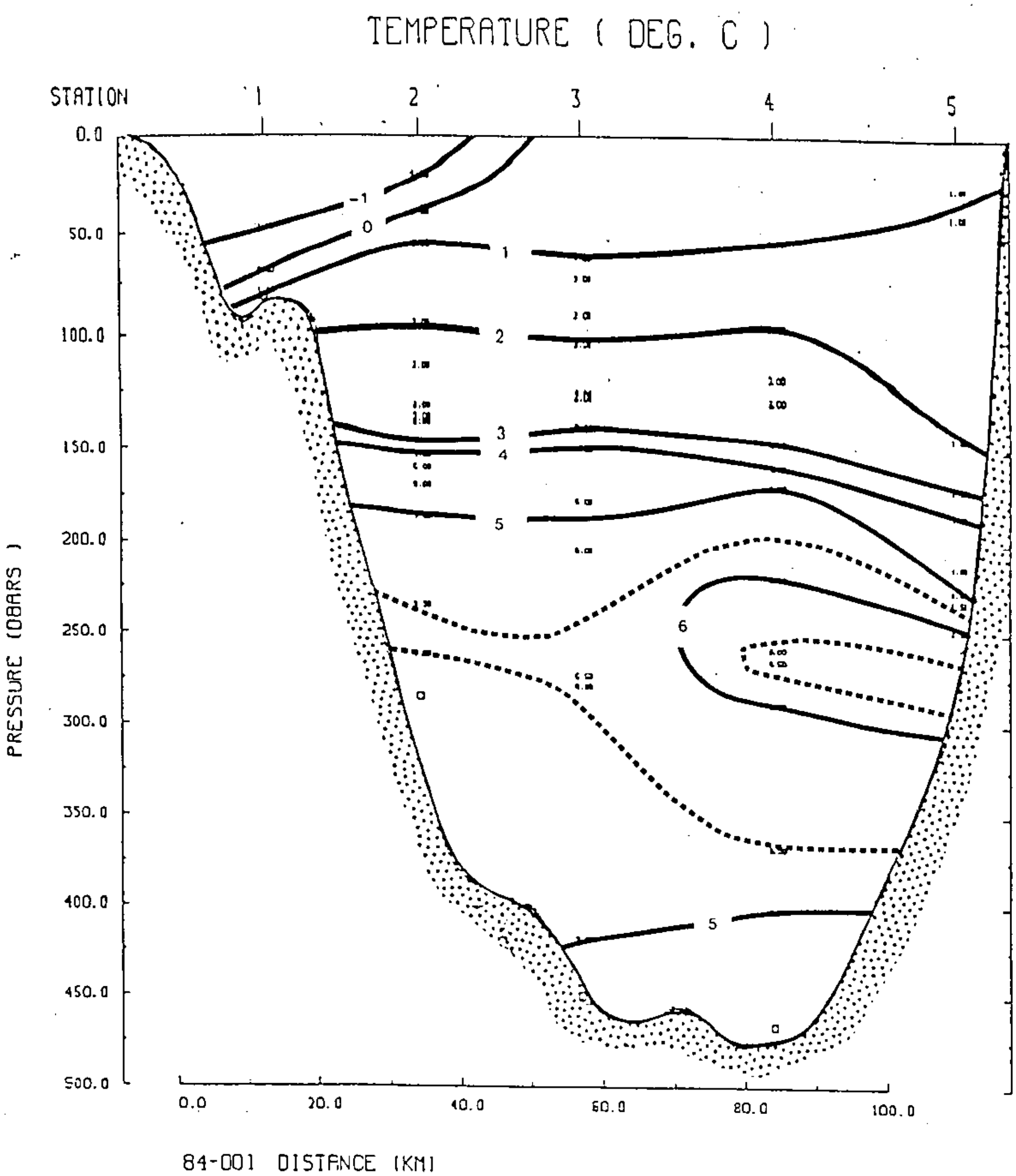
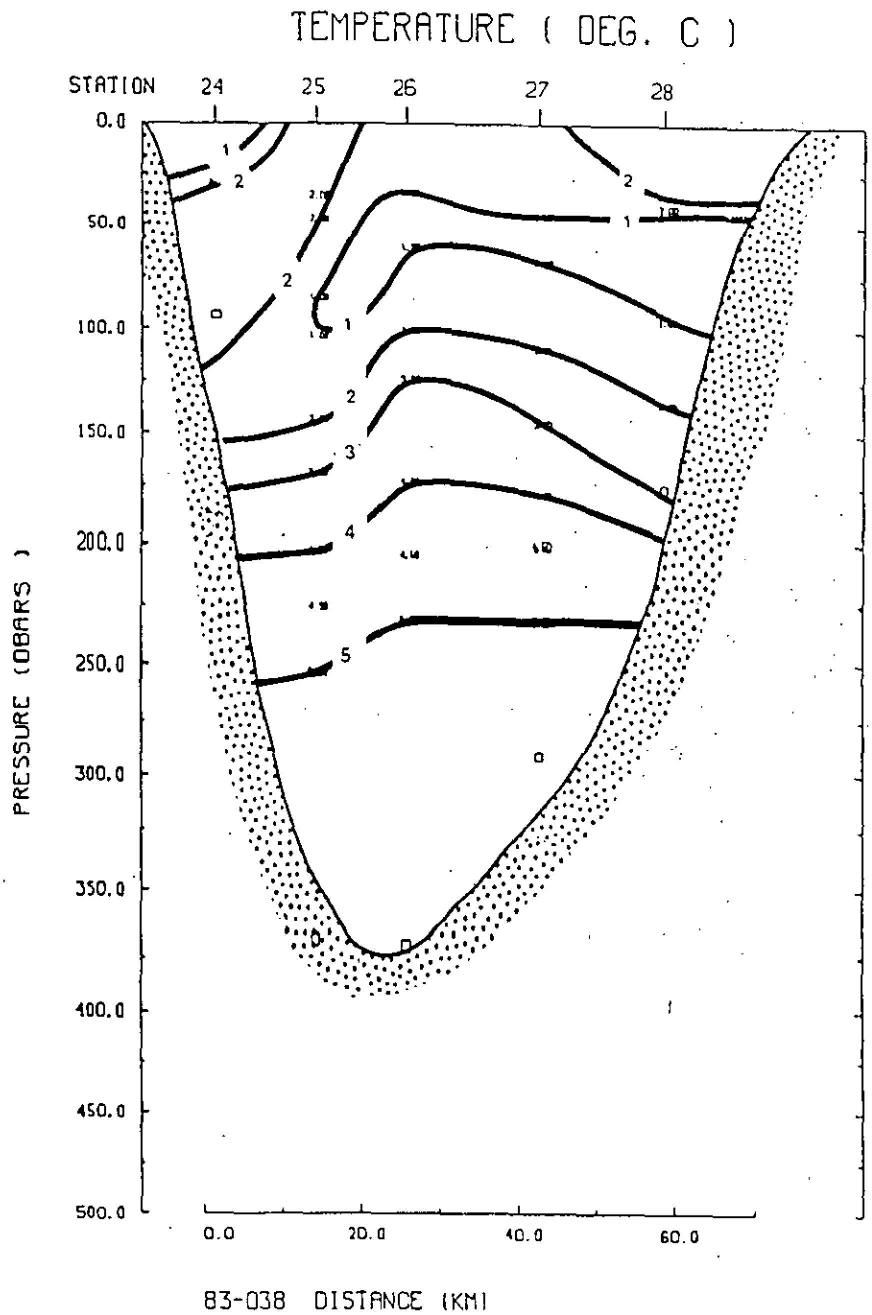


Figure 2.11 Typical cross channel vertical temperature section at Cabot Strait from data obtained in February 1984.



<u>Figure</u> <u>2.12</u> Typical cross channel vertical temperature section at Port Menier from data obtained in November 1983.

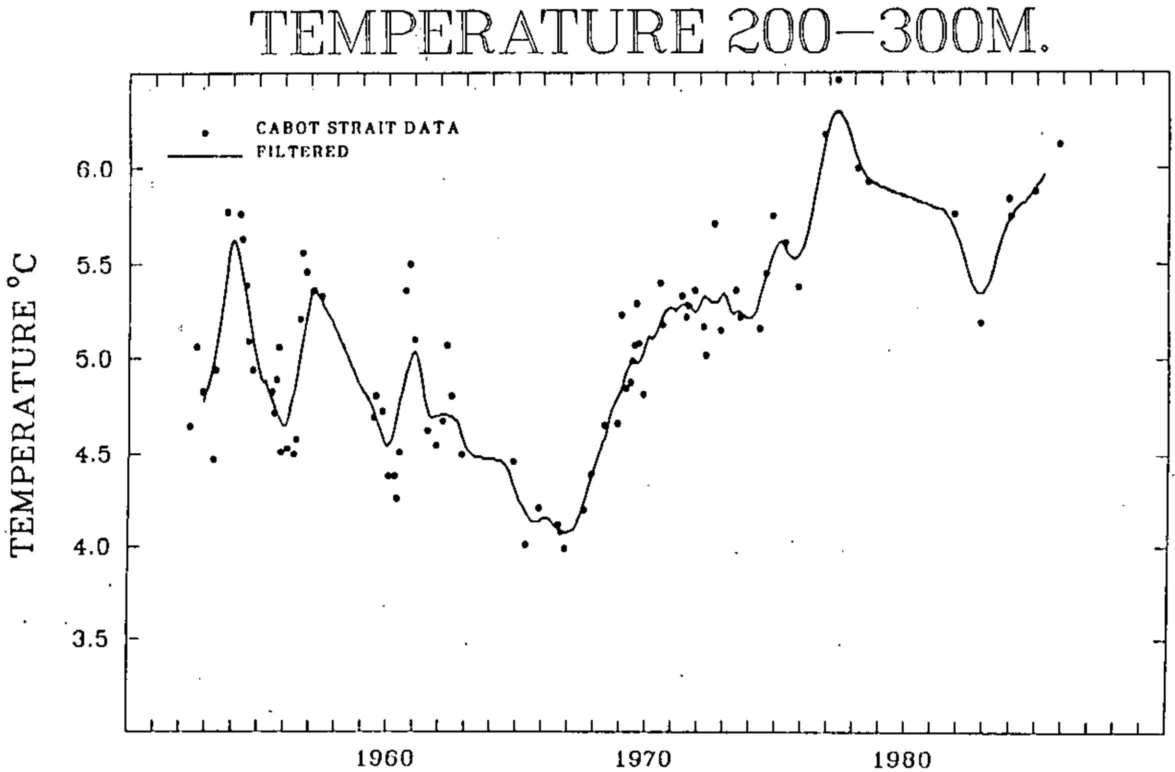


Figure 3.1 Temperature averaged from 200 to 300 m. at Cabot Strait. See text for explaination of filtering.

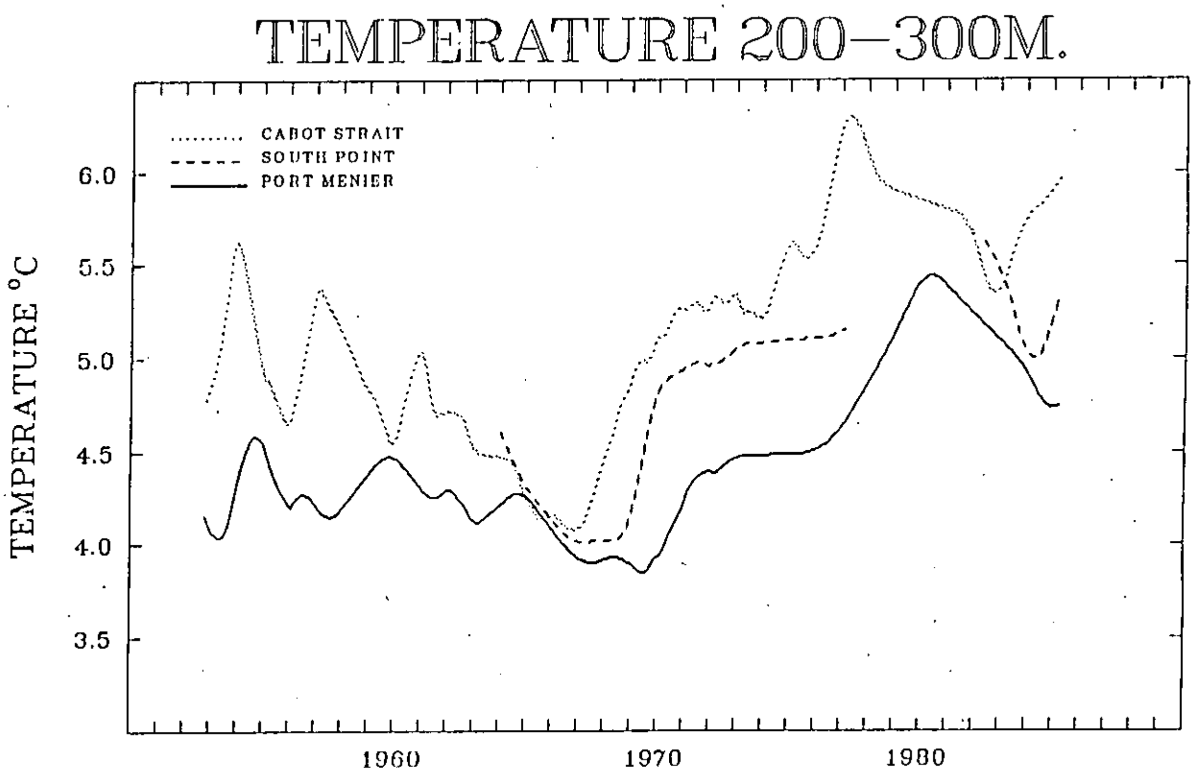


Figure 3.2 Temperature averaged from 200 to 300 m. from three rectangles along the Laurentain Channel filtered as described in the text.

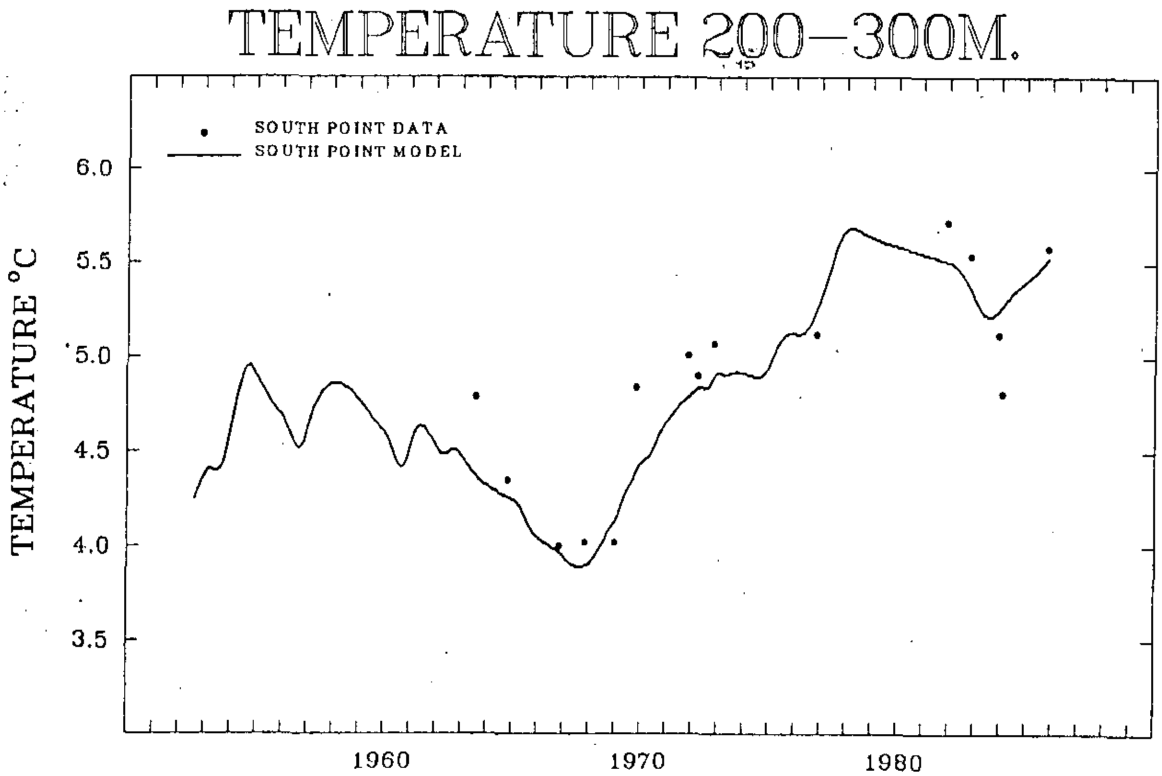


Figure 3.3 Temperature observations and model output for the SouthPoint rectangle, 267 km. from the driven model boundary at Cabot Strait.

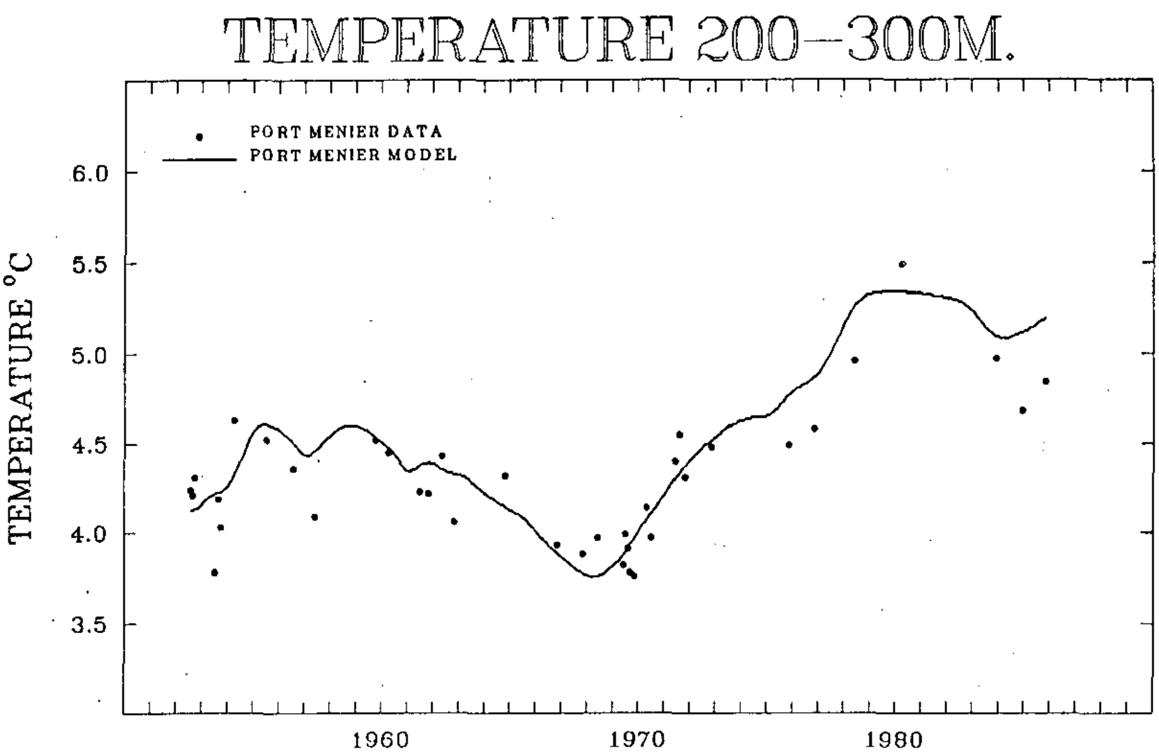


Figure 3.4 Temperature observations and model output for the Port Menier rectangle, 452 km. from the driven model boundary at Cabot Strait.

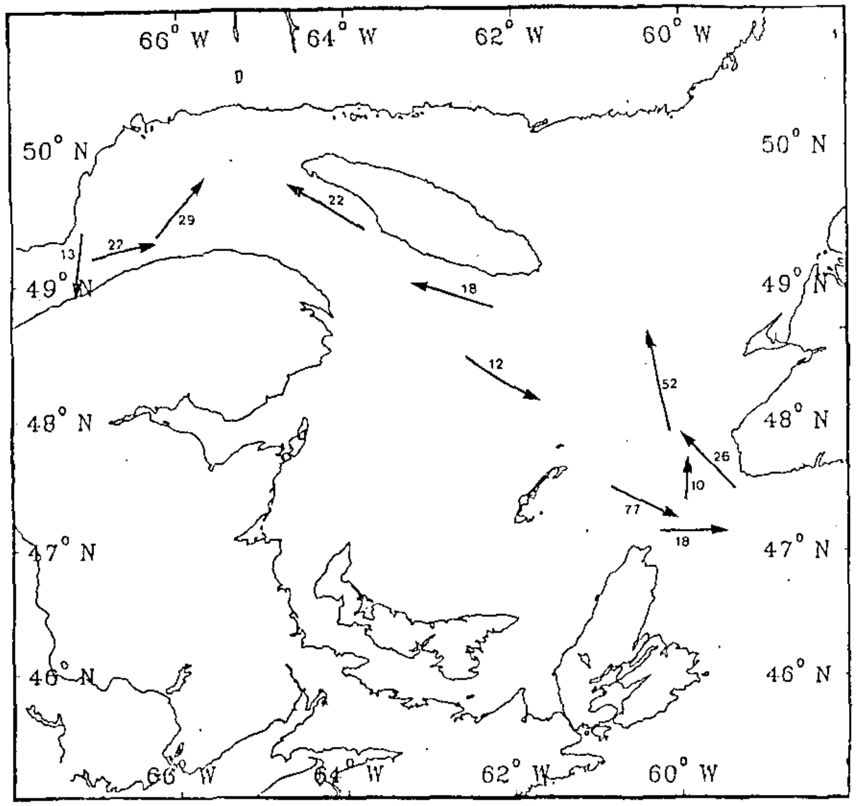


Figure 3.9 Current vectors averaged over the available length of record for stations East of Pointe des Monts. Current speeds are in mm/s.

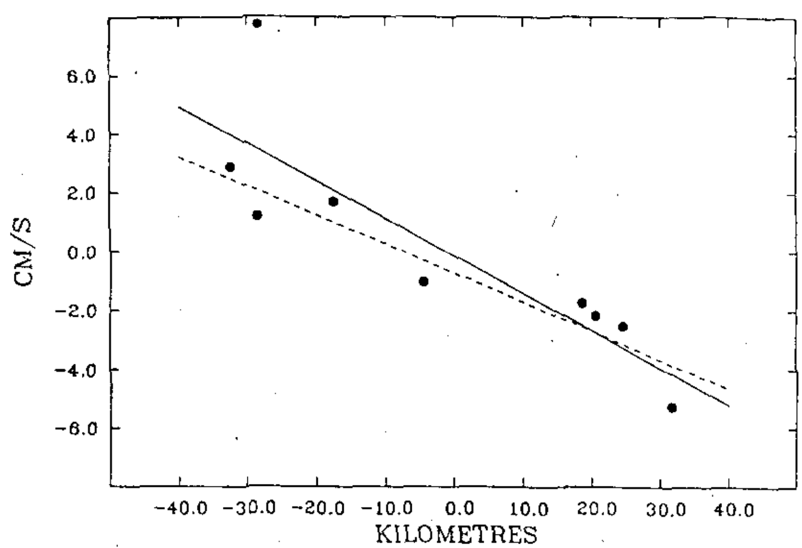


Figure 3.10 Average current speed vrs direction from the axis of the Laurentain Channel. See text for explanation of fitted lines.

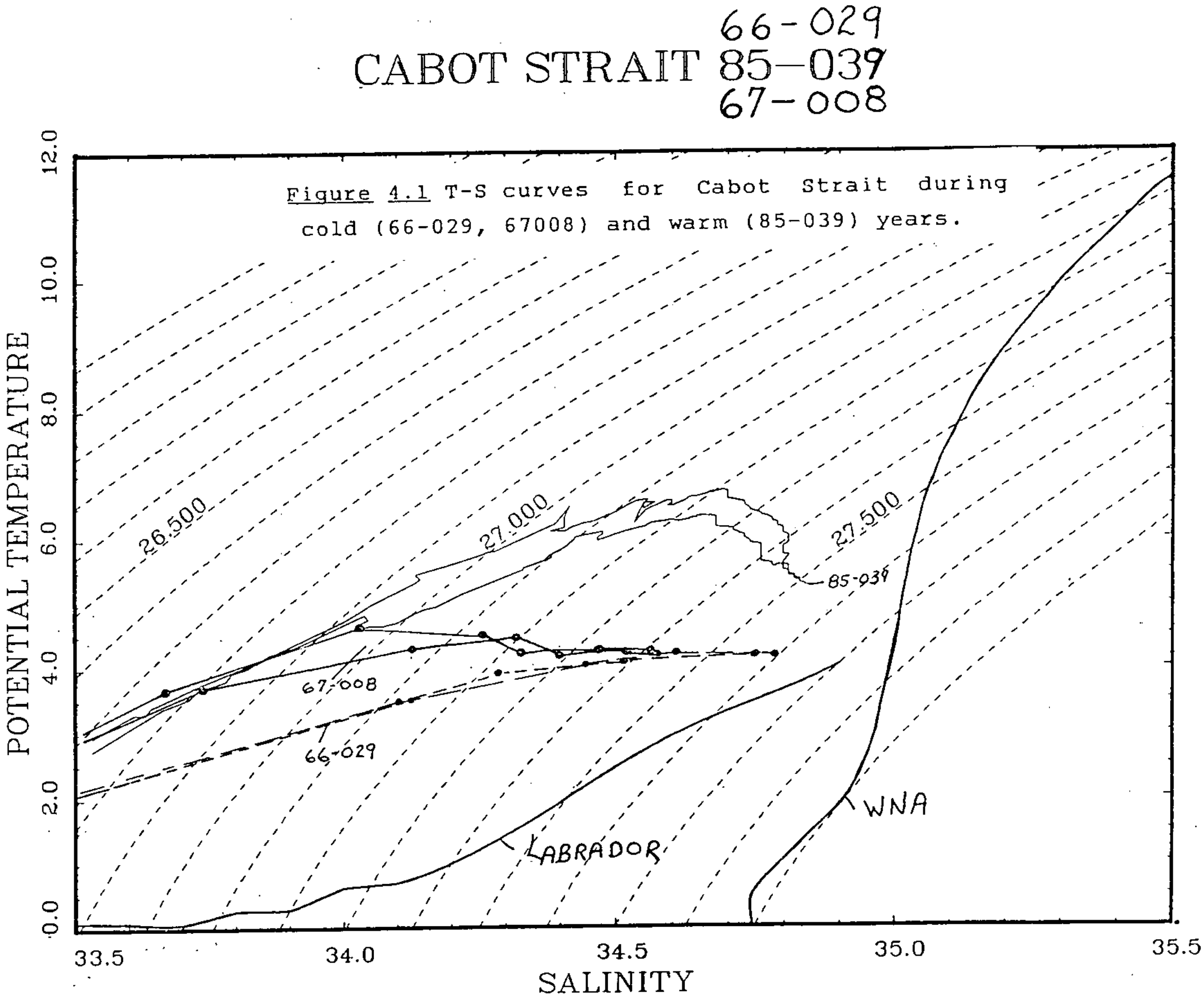


Figure 4.1 T-S curves for Cabot Strait during cold (66-029, 67008) and warm (85-039) years.

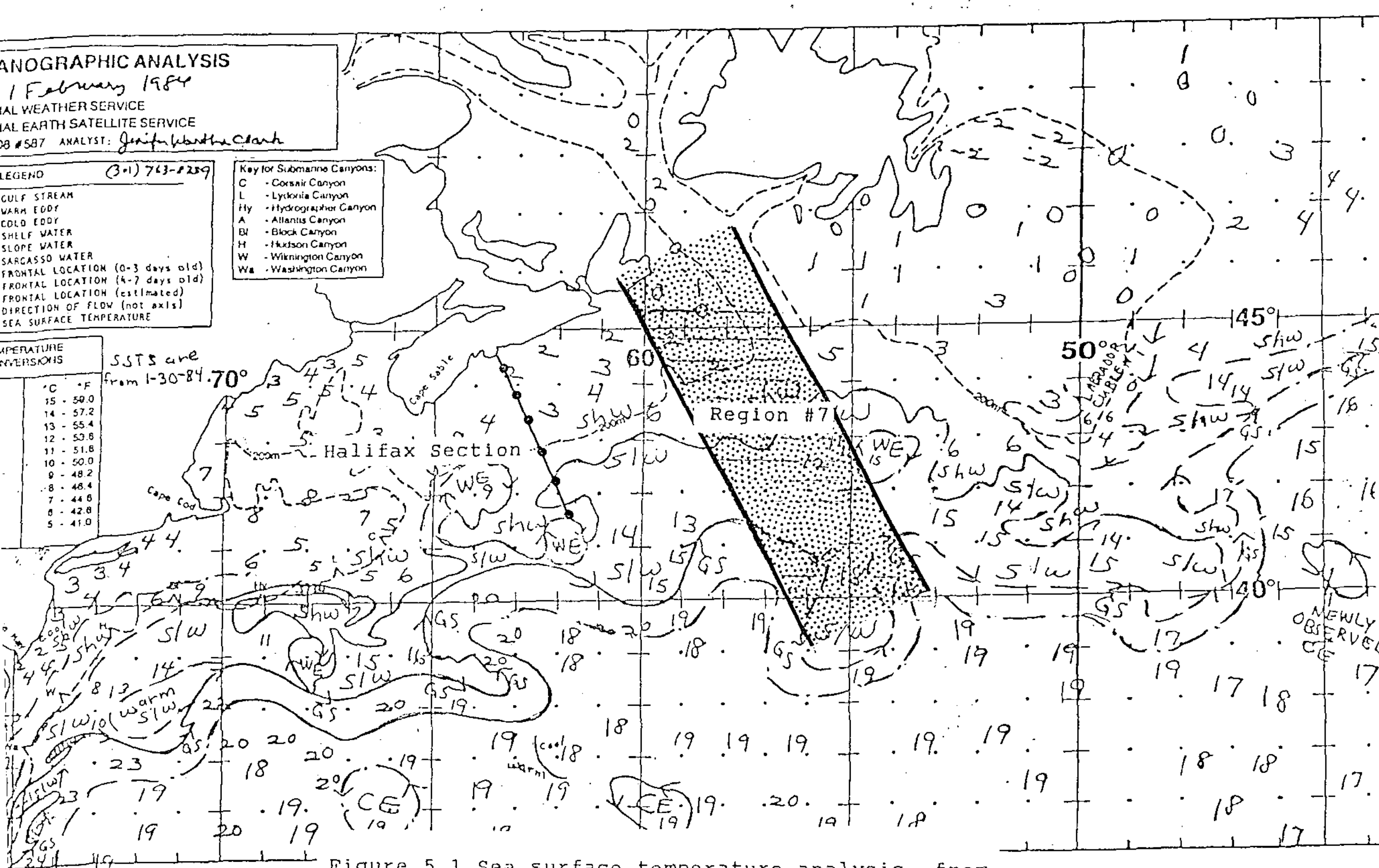


Figure 5.1 Sea surface temperature analysis from satellite derived information.

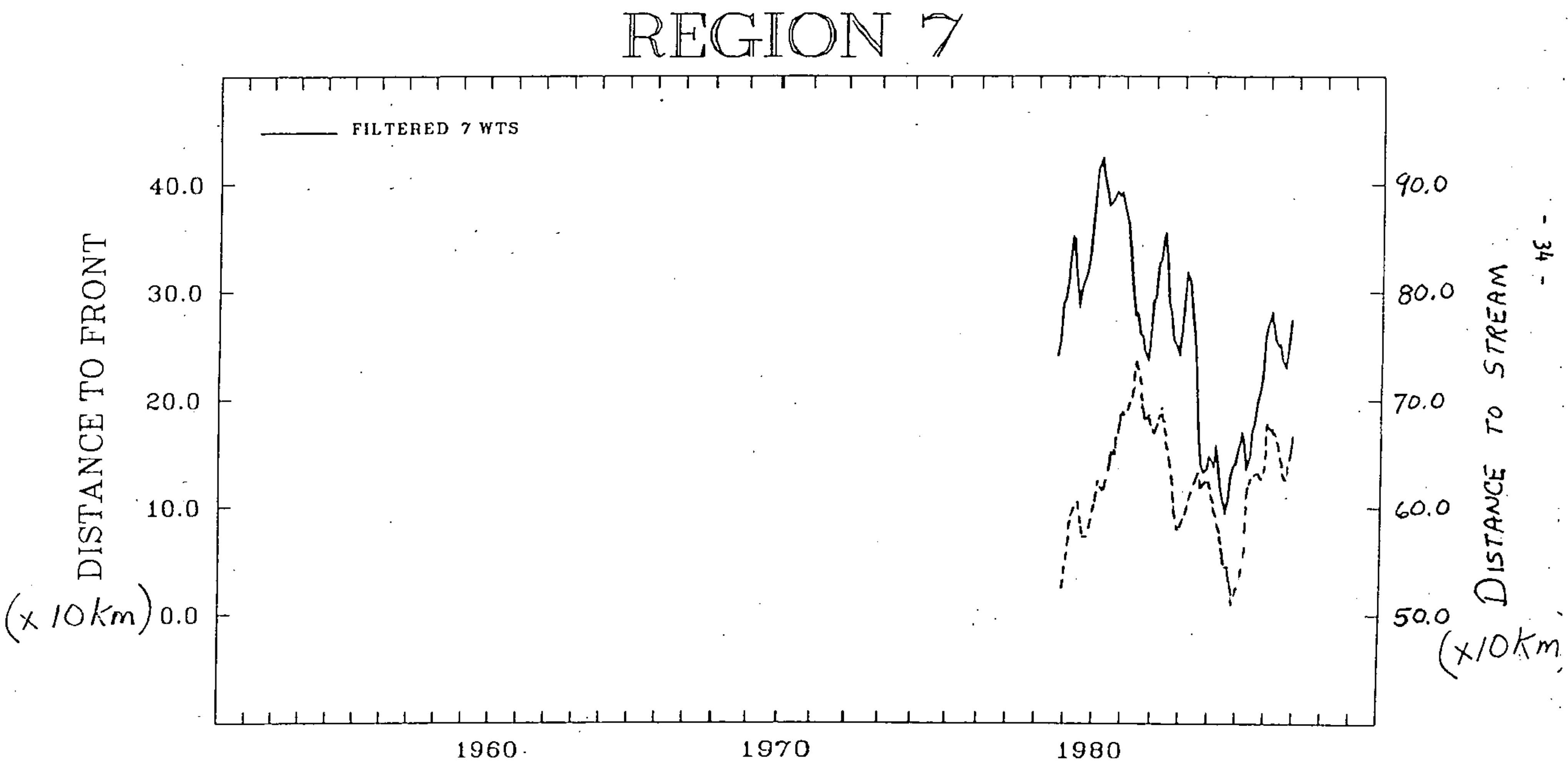


Figure 5.2 Average distance from the 200 m. isobath to the shelf-slope water boundary, reduced to monthly averages and filtered. Also shown is the distance to the Gulf Stream.

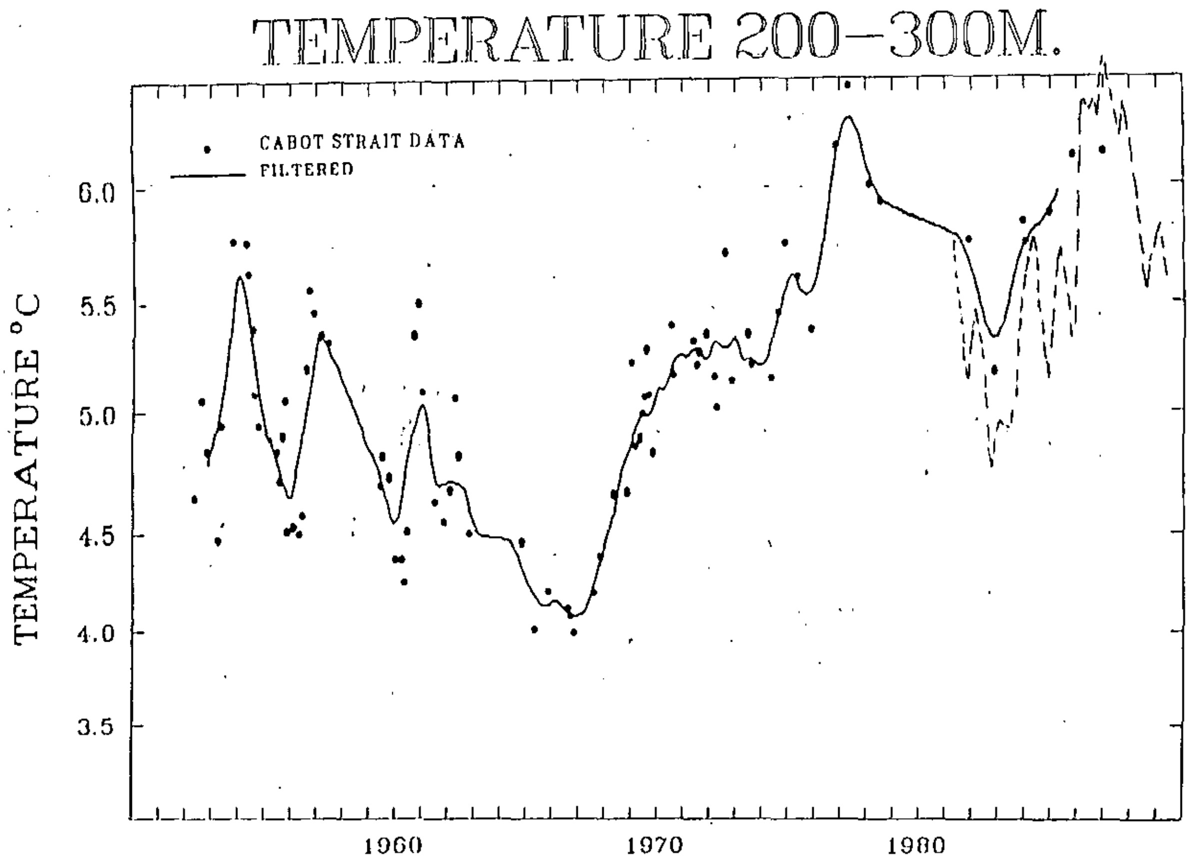


Figure 5.3 Cabot Strait temperatures along with distance to the shelf-slope water boundary lagged 2.5 years.

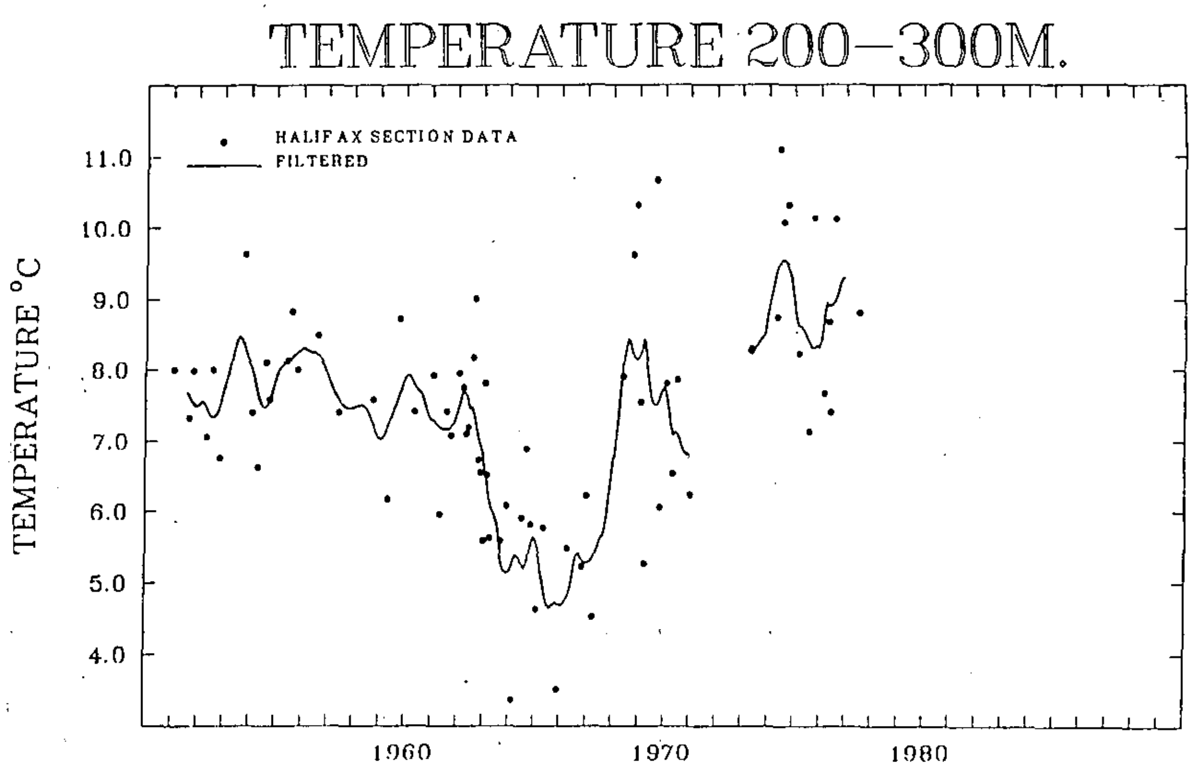


Figure 5.4 Temperature at the first station of the Halifax Section beyond the shelf break

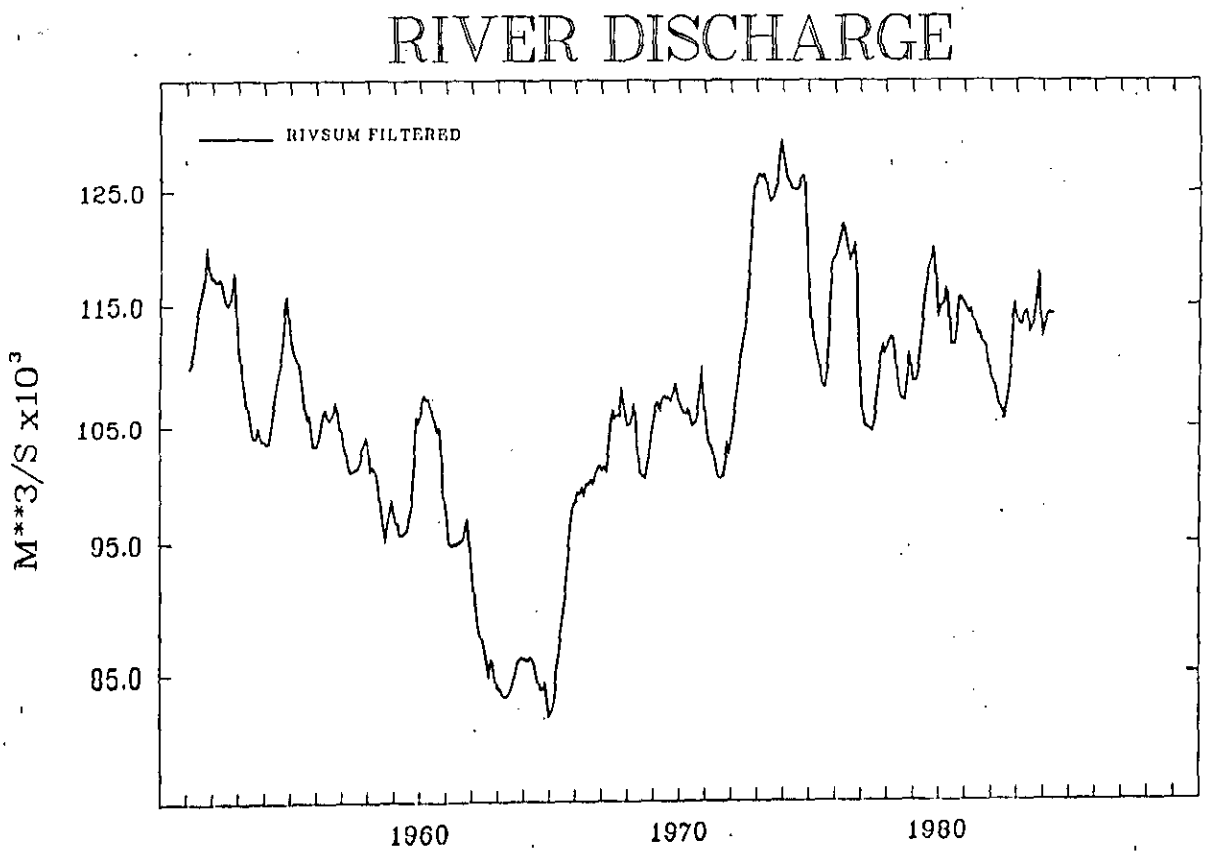


Figure 5.5 Discharge from the St. Lawrence and other rivers into the St. Lawrence Estuary as monthly averages filtered with a 13 weight boxcar filter.