ICES 1989 PAPER C.M. 1989/C:16

SEASONAL AND INTERANNUAL VARIABILITY OF THE LABRADOR CURRENT AND WEST GREENLAND CURRENT

by

Ransom A. Myers, J. Helbig, and D. Holland
Science Branch
Department of Fisheries and Oceans
P.O. Box 5667
St. John's, Newfoundland, Canada A1C 5X1

ABSTRACT

Results are presented from an analysis of long time series of four standard hydrographic sections in the Labrador Sea: the Fyllas Bank line off West Greenland, the Seal Island section off Labrador, the Bonavista transect on the Northeast Newfoundland Shelf, and the Flemish Pass line across the Northern Grand Banks. Over 300 transects were analyzed. Time series of geostrophic transport for each section were computed and were analyzed for seasonal and interannual variations, both over the shallow continental shelf and over the slope. We tested hypotheses concerning the relationships between variability in seasonal and interannual transport, wind forcing, and buoyancy fluxes caused by ice melt and runoff. Baroclinic transport in the West Greenland and Labrador Currents in the summer is negatively correlated with strength of the westerlies in the North Atlantic in the previous winter.

Résumé

On présente les résultats d'une analyse de longues séries chronologiques portant sur quatre sections hydrographiques normalisées dans la mer du Labrador: la ligne du banc Fyllas au large de la partie ouest du Groenland, la section de l'île Seal au large du Labrador, le transect Bonavista sur le plateau de Terre-Neuve (nord-est), et le passage du Bonnet Flamand coupant le banc Northern Grand. Plus de 300 transects ont été analysés. Les séries chronologiques du transport géostrophique correspondant à chaque section ont été calculées et analysées pour les variations saisonnières et interannuelles, à la fois sur le plateau continental peu profond et sur la pente. Nous avons vérifié les hypothèses concernant la relation entre la variabilité dans le transport saisonnier et interannuel, le forçage éolien et les flux de flottabilité causés par la fonte des neiges et le ruissellement. Il y a en été une corrélation négative entre, d'une part, le transport barocline dans les courants de l'ouest du Groenland et du Labrador et, d'autre part, l'intensité des alizés dans l'Atlantique nord au cours de l'hiver précédent.

1 Introduction

Previous attempts to demonstrate fluctuations in currents that might be of real climatic significance have not been convincing because the signal has been obscured either by noise or by differences in methodology (Wunsch 1981). For example, the wide variability in estimates of the strength of the Gulf Stream north of Cape Hatteras has been attributed to varying methods of estimating reference levels (Wunsch 1981). Notwithstanding the past failures of such projects, we have attempted to examine data from the Labrador Sea for interannual and seasonal variation in transport. We did so for the following reasons. First, we were motivated by Myers *et al.* (1988) who showed that the interannual variation in the density structure off the West Greenland shelf appears to be related to large scale wind forcing. Second, on all sections, the data base was extensive enough to yield at least 5 transport estimates for each year. We thus were able to separate short and long term fluctuations. Finally, large scale interannual changes in deep water formation (Clark and Gascard 1983, Brewer *et al.* 1983), salinity (Dickson *et al.* 1988, Levitus 1989) and ice cover (Manak and Mysak 1989) in the Labrador Sea indicate that this may be a particularly dynamic region on both seasonal and interannual time scales. Moreover, Dickson *et al.* (1988) argued that a large salinity anomaly was advected around the North Atlantic during the decade beginning in 1968. This anomaly was first observed north of Iceland, then in the Labrador sea, and finally in the central and eastern Atlantic.

In this paper we examine transport, temperature, and salinity variations along four standard transects in the Labrador Sea. As well we consider the evidence in our data for the salinity anomaly reported by Dickson *et al.* (1988) in the Labrador Sea. In all cases we have returned to the original data so that common methods are used throughout.

2 Data

Data from four standard Northwest Atlantic Fisheries Organization (Stein 1988) oceanographic sections were considered (Fig. 1). These data consisted of temperatures and salinities collected by a variety of agencies over the past fifty years, and were compiled from files of the Marine Environmental Data Service (MEDS), National Oceanographic Data Center (NODC), and International Council for the Exploration of the Seas (ICES). Duplicates were identified and eliminated, and the data were interpolated onto standard depths. *Fig. 1 near here*

3 Methods

The seasonal cycle in salinity and temperature was calculated at each station and at each standard depth as the annual harmonic of the corresponding time series. The seasonal cycle was then removed to give time series of interannual variation. Standardized anomalies were created by dividing by the standard deviation of the residuals from the seasonal cycle. The mean or median of the resulting series for groups of stations and standard depths were used in the analysis. This allows sporadically collected data to be standardized for the analysis.

The baroclinic transport over the continental slope and shelf was calculated relative to 100 m. This reference level was chosen for two reasons. First, it was the deepest level that did not intersect the bottom over most of the shelf. Second, the effects that we are most interested in, namely baroclinic forcing by wind and ice-melt, are largely confined to the upper 100 m. In a later paper, we will report results obtained using the approach of Csanady (1979) in which the bottom is used as a level of no motion.

4 Results

4.1 Salinity Variation

Salinity variations in the upper 200 m in the Labrador Sea show evidence of three distinct freshenings at decadal time scale (Fig. 2): one around 1959, one around 1971, and one around 1984. The anomalies were also evident in the Station 27 time series, which is located in the inshore branch of the Labrador current near St. John's, Newfoundland. These three negative salinity anomalies correspond to periods of extended ice cover in the Labrador sea (Mysak and Manak 1989). However, only one of these freshenings, the early 1970's anomaly, appears to be correlated with salinity variations north of Iceland shown by Dickson *et al.* (1988), while the other two do not. The data thus do not support Dickson *et al.*'s hypothesis of a multiple passage of the fresh water pulse around the North Atlantic. They also sugggest that the 1970's anomaly differed in character from the other anomalies. We are continuing our investigation of this possibility.

Fig. 2 near here

The long term variations in the deep (greater than 750 m) salinity time series are coherent among stations and sections (Fig 2). There is a long term increase from 1950 to 1970, followed by a decrease until 1975, and then a gradual increase. The same pattern is seen in the Greenland regional atmospheric pressure anomaly (Rogers 1984). Thus, the deep salinity field may be responding to long term, large scale atmospheric patterns. It is interesting to note that the deep water temperature at the Panulirus station in the subtropical North Atlantic shows a similar pattern (Levitus 1989). Note that these salinity anomalies are not seen in the deeper waters in the West Greenland Current (Fig. 2a).

4.2 Transport Variation

4.2.1 Seasonal Transport

The variablility in the Labrador and West Greenland Currents seems to be greatest near the shelf break due to the presence shelf waves and/or eddies which have a time scale on the order of days and a spatial scale on the order of 30 km. We found that the short term variability in the spatially integrated transport estimates was significantly reduced if we extended our integration both inshore and seaward of the effects of these disturbances.

The baroclinic transport relative to 100 m on the continental slope off the Grand Banks of Newfoundland at 47 degrees latitude appears to peak in the spring and drop to a minimum in the fall. By contrast, the seasonal cycle of the current over the shelf appears to be the inverse of this pattern (Fig 3). That is, transport over the shelf peaks in the late fall with a minimun in spring. These cycles are illustrated further in Fig. 4 which shows the vertically integrated geopotential for four stations on the Flemish Cap section. More precisely, the quantity plotted is the geopotential integrated vertically over the top 100 m and divided by the Coriois parameter. It follows that the difference in value between stations gives the upper layer transport across the line joing the two stations. The stations are numbered sequentially from the coast: Station 2 ($47°00'N$, $51°00'W$; $102m$) lies over the inner Grand Banks of Newfoundland, but seaward of what is generally regarded as the inshore branch of the Labrador Current; Station 7 ($47°00'N$, $48°07'W$; $126m$) lies inshore of the shelf break and inshore of the core of the Labrador Current; Station 14 ($47°00'N$, $46°01'W$; $308m$) lies in the Flemish Pass and offshore of the typical location of the core of the Labrador Current; Station 19 ($47°00'N$, $43°45'W$; $674m$) lies far enough east of Flemish Cap to capture most of the portion of the Labrador Current that rounds the northern face of the Cap. Thus differences between Stations 7 and 14 and provide a measure of the baroclinic transport of the main branch of the Labrador

Fig. 3 near here

Fig. 4 near here

Current that flows through Flemish Pass, and to the extent the geostrophic relation is valid in shallow water, the differences between Stations 2 and 7 give the transport over the shelf and some measure of the transport of the inner branch of the Labrador Current.

Fig. 4 shows that the geopotential integrated over the upper layer varies over the year in a coherent manner over all the stations. This implies that the results from the inner stations are meaningful even though they lie in shallow water where friction and wind forcing might be expected to degrade the geostrophic balance. The most interesting feature in Fig. 4 is the phase relationship between the curves. The curves for Stations 2, 14, and 19 are phase locked so that the total upper layer baroclinic transport across the shelf and slope varies little throughout the year. That is the total upper layer transport due to the inshore and offshore branchs of the Labrador Current is nearly invariant over the year. However, as can be seen by comparing Stations 2 and 7 and Stations 7 and 14 or 19, the inshore and offshore transports vary in near-antiphase: the inshore transport peaks in the winter and declines to a minimum in spring whereas the offshore transport peaks in the spring and achieves a minimum in the winter.

4.2.2 Interannual Variability

Our goal was to identify major sources of interannual variability in transport on a basin wide scale. Here we report the clearest pattern of variability we have found so far, namely, that the baroclinic transport throughout the upper layer of the Labrador Sea is decreased in summers following winters with strong mid-latitude westerlies. *Fig. 5 near here*

Fig. 5 shows contours of the correlation coefficient between the NAO index and the salinity and temperature on the Fylla bank section the following July. Following winters of strong westerlies, the salinity and thus the density, of the water column offshore is reduced relative to inshore waters.

Consequently, the baroclinic pressure gradient is smaller and the currents are diminished giving an overall smaller transport. This finding also suggests that the shelf break front is weaker in summers following strong westerlies, a phenomenon which could be apparent in satellite infrared imagery.

Baroclinic transport relative to 100 m in the summer is negatively correlated with the strength of the winter mid-latitude westerlies in the North Atlantic as represented by the North Atlantic Oscillation Index (Rogers 1984, Myers *et al.* 1988). Significant negative correlations ($p < 0.05$) occurred on the Fyllas Bank section, the Seal Island section, and on the Flemish Cap section within Flemish Pass (Fig. 6). *Fig. 6 near here*

Note that the mean transports are smaller through Flemish Pass than through the Seal Island section. This is most likely because a portion of the Labrador current is diverted east of Flemish Cap and a portion flows inshore over the Grand Banks.

Negative correlations were also found with ice cover the previous winter (Fig. 7), but this is probably caused by the strong correlation with the North Atlantic Oscillation and ice cover in the Labrador Sea (Myers *et al.* 1988). The ice cover data was provided by John Walsh (Walsh and Sater 1981). This is evidence against the hypothesis that the baroclinic portion of the Labrador current is mainly driven by thermohaline forces. *Fig. 7 near here*

Finally, a simple analysis suggested that the baroclinic transport was not correlated with local wind forcing over the previous week.

5 Discussion

We have presented evidence that coherent variations occur in the water properties and upper layer baroclinic transport throughout the Labrador Sea. Our discussion has been primarily descriptive and statistical and the the more difficult task of explaining our results mechanistically is still underway. However, the three major findings that were presented here are briefly discussed.

First, we showed that salinity anomalies in the Labrador Sea of similar strength to the early 1970's anomaly described by Dickson *et al.* (1989) appear to occur on a decadal time scale in the Labrador Sea. What makes the 1970's anomaly unique in terms of the limited salinity time series available is its expression throughout the full North Atlantic. A more through study is underway to determine if the salinity anomalies observed in the Labrador Sea are coherent in any way with variations in water properties in the rest of the North Atlantic.

Second, we demonstrated an inverse relationship between winter westerly winds and the upper layer baroclinic transport in the Labrador Sea the following summer. Strong westerlies imply weakened shelf break fronts and diminished baroclinic transports. This pattern is consistent throughout the Labrador Sea and may have profound implications for understanding the interannual variability in the ocean climate in the region. This is one of the few examples where interannual variability within a major current has been demonstrated and related to large scale atmospheric forcing.

Finally, we found that the total upper layer baroclinic transport of the Labrador Current across the Flemish Cap line was nearly constant throughout the year as the individual transports due to the inshore and offshore branchs varied in anti-phase with maximum offshore transport in spring and maximum inshore transport in winter. Several plausible mechanisms can be advanced to explain this finding. The simplest is that the core of the Labrador Current simply shifts onshore. This is unlikey though, as Station 7 is over 30 km inshore from the shelf break. Moreover, there is no suggestion of any shift in mean water properties (Keeley, 1981). Another possibility is that in winter, a larger fraction of the Labrador Current is diverted towards the inshore branch when the Labrador Current leaves the Labrador shelf and flows onto the Northeast Newfoundland shelf. In this regard we note that a similar pattern does not seem to prevail over the Labrador shelf. Lazier

(1988) showed that both direct velocity measurements at 600 m over the Labrador continental slope and the difference in steric height between the Labrador shelf and the mid-Labrador Sea are at their minimum values in spring and peak in the fall. We have yet to analyze the Labrador shelf data for inshore/offshore seasonality. A more thorough examination of these results is one of the objectives of a numerical modelling study of the Labrador Current presently underway.

Acknowledgements

We are pleased to acknowledge many helpful conversations with Dr. Gordon Mertz.

6 References

Dickson, R. R., J. Meincke, S.A Malmberg, and A. J. Lee. 1984. The Great Salinity Anomaly in the Northern North Atlantic, 1968-82. Progr. Oceanogr. 20: 103-151.

Brewer, P.G., W. S. Broecker, W. J. Jenkins, P. B. Rhines, C. G. Rooth, J. H. Swift, T. Takahashi and R. T. Williams. 1983. A climatic freshening of the deep Atlantic north of 50°N over the past 20 years. Science 222: 1237-9.

Clarke, R. A. and J. Gascard. 1983. The formation of Labrador Sea Water, part 1. Large scale processes. J. Phys. Oceanogr. 13: 1764-1778.

Csanady, G. T. 1979. The pressure field along the western margin of North Atlantic. J. Geophys. Res. 84(C8): 4905-4915.

Keeley, J. R. 1981. Mean conditions of potential temperature and salinity along the Flemish Cap Section. Environment Canada, Marine Environmental Data Service Technical Report No. 9. 148 p.

Lazier, J. R. N. 1980. Oceanographic conditions at O. W. S. Bravo 1967-74. Atmosphere-Ocean 18: 228-238.

Lazier, J. R. N. 1988. Measurements from instruments moored in the Labrador Current 1978 - 1986. Intergovernmental Oceanographic Commission Technical Series 33, 1-10.

Levitus, S. 1989 Interpentadal variability of temperature and salinity at intermediate depths of the north atlantic ocean, 1970-74 versus 1955-59. In Press J. Geophys. Res. Oceans.

Mysak, L. A. and D. K. Manak. 1989. Arctic sea ice extent and anomalies, 1953-1984. Atmos. Ocean, 27, 376-405.

Myers, R. A., S. A. Akenhead, and K. Drinkwater. 1988. The North Atlantic Oscillation and the Ocean Climate of the Newfoundland Shelf. International Council for the Exploration of the Sea. C.M. 1988/C:12.

Rogers, J. C. 1984. The association between the North Atlantic Oscillation and the Southern Oscillation in the Northern Hemisphere. Mon. Wea. Rev. 112: 1999-2105.

Stein, M. 1988. Revision of list of NAFO standard oceanographic sections and stations. Northwest Atlantic Fisheries Organization SCR Doc. 88/01.

Sutcliffe, W. H., R. H. Loucks, K. F. Drinkwater and A. R. Coote. 1983. Nutrient flux onto the Labrador Shelf from Hudson Strait and its biological consequences. Can. J. Fish. Aquat. Sci. 40: 1692-1701.

Thompson, K. R., J. R. N. Lazier and B. Taylor. 1986. Wind-forced changes in Labrador Current transport. J. Geophys. Res. 91(C12): 14261-14268.

Walsh, J. E. and J. E. Sater. 1981. Monthly and seasonal variability in the ocean–ice–atmosphere systems of the North Pacific and the North Atlantic. J. Geophys. Res. 86(C8):7425-7445.

Wunsch, C. 1981. Low-frequency variability of the sea. *Evolution of Physical Oceanography* B. Warren and C. Wunsch, Ed., Massachusetts Institute of Technology Press, 342-374.

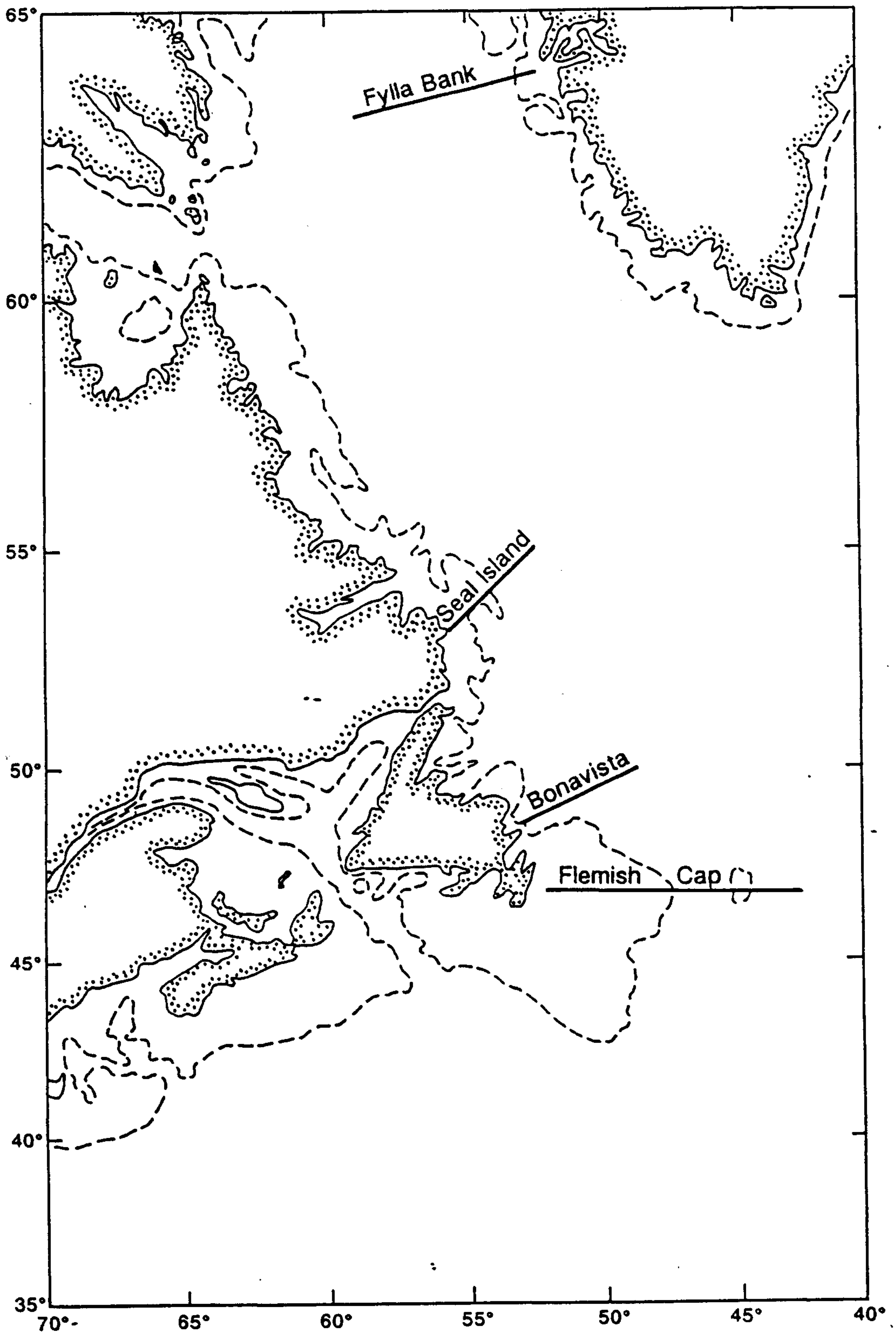


Fig. 1. Location of the four standard sections and Station 27. Station numbers are also indicated on each section.

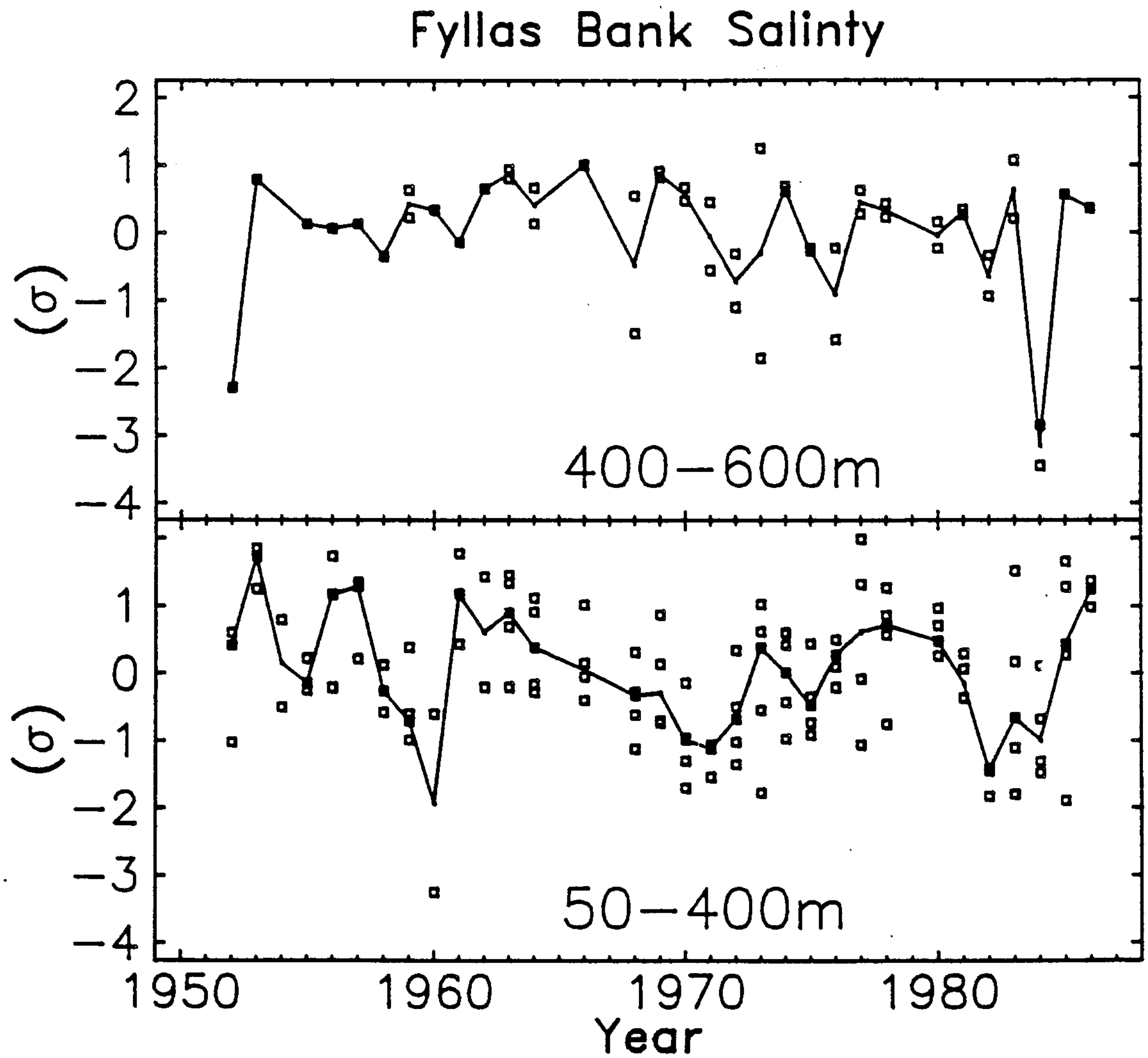


Fig. 2a. The normalized salinity at Fyllas Bank. The solid line for the Fyllas Bank data connects the median of the estimates for each depth zone.

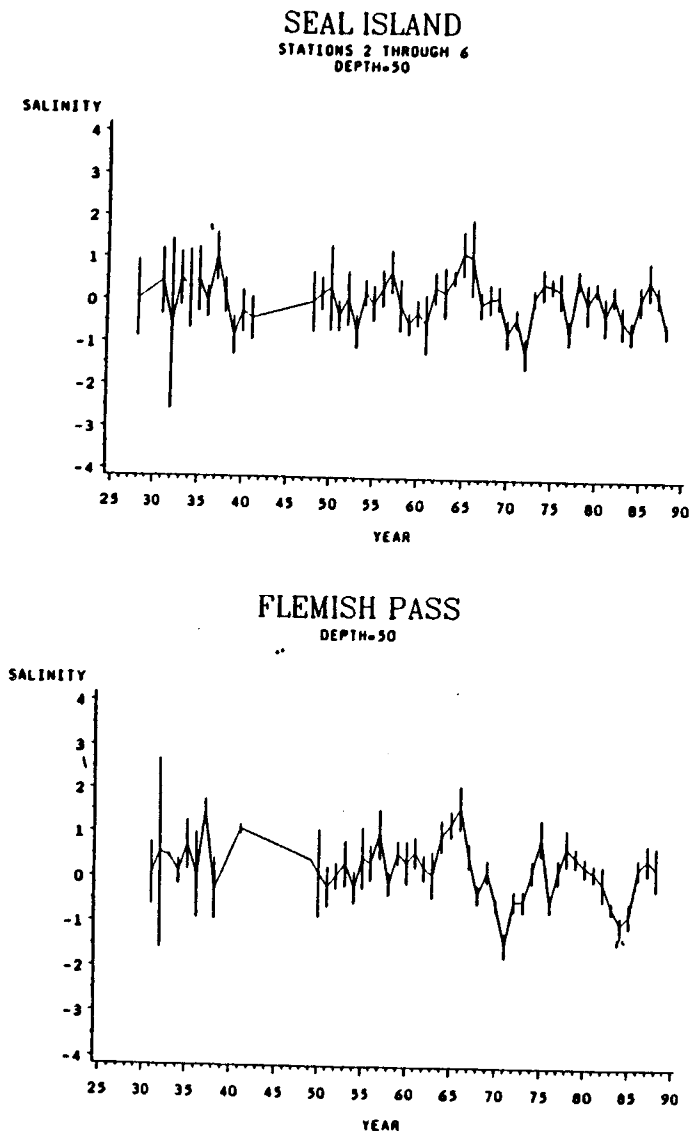


Fig. 2b. Time series of salinity residual for (a) Seal Island Line and (b) Flemish Pass Line at depth 50 m. Units of salinity residual are standard deviations. Approximate 95 % confidence limits are shown.

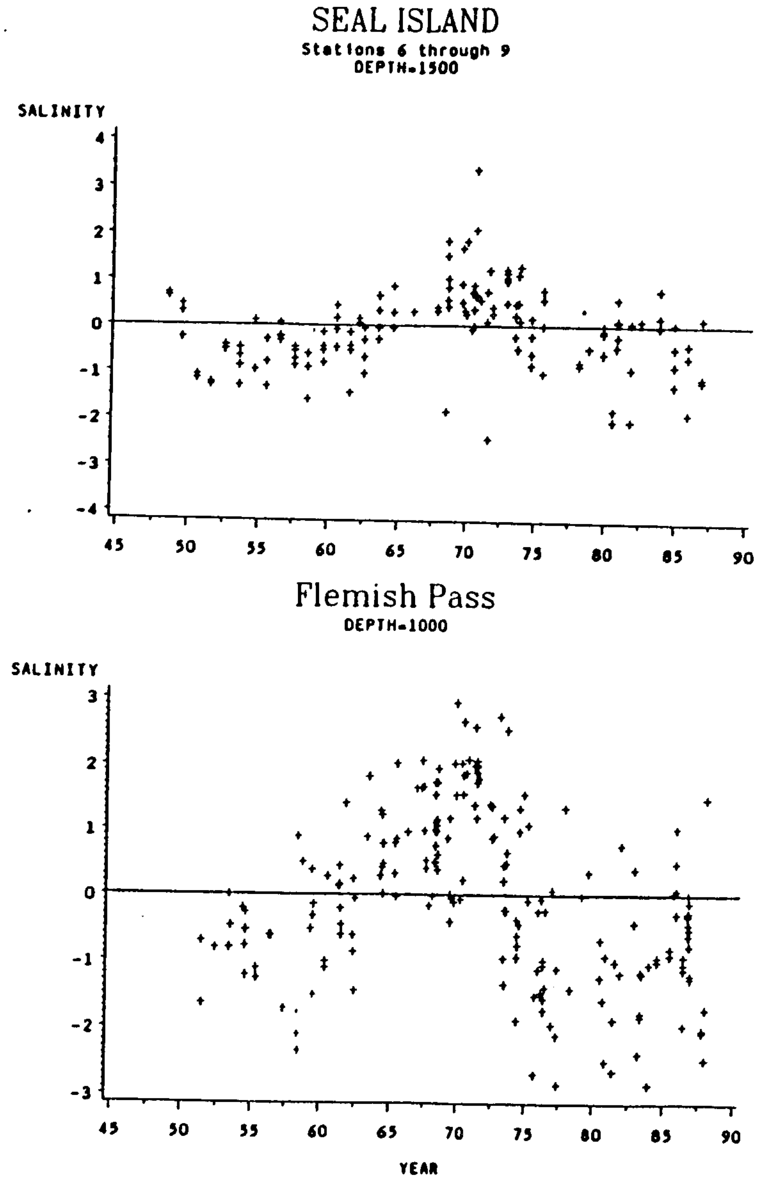


Fig. 2c. Time series of salinity residual for (a) Seal Island Line at depth 1500 m and (b) Flemish Pass Line at depth 1500 m. Units of salinity residual are standard deviations.

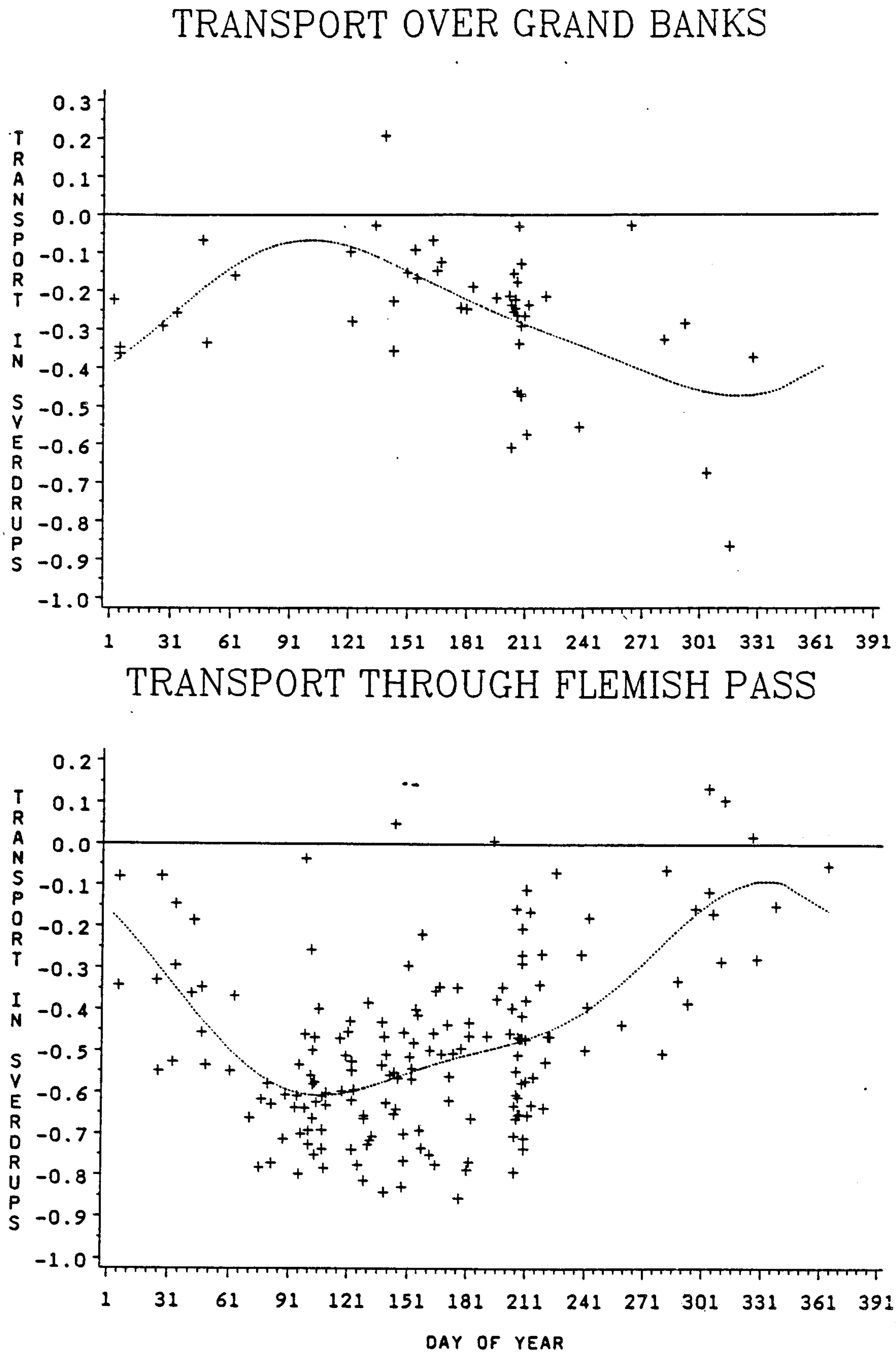
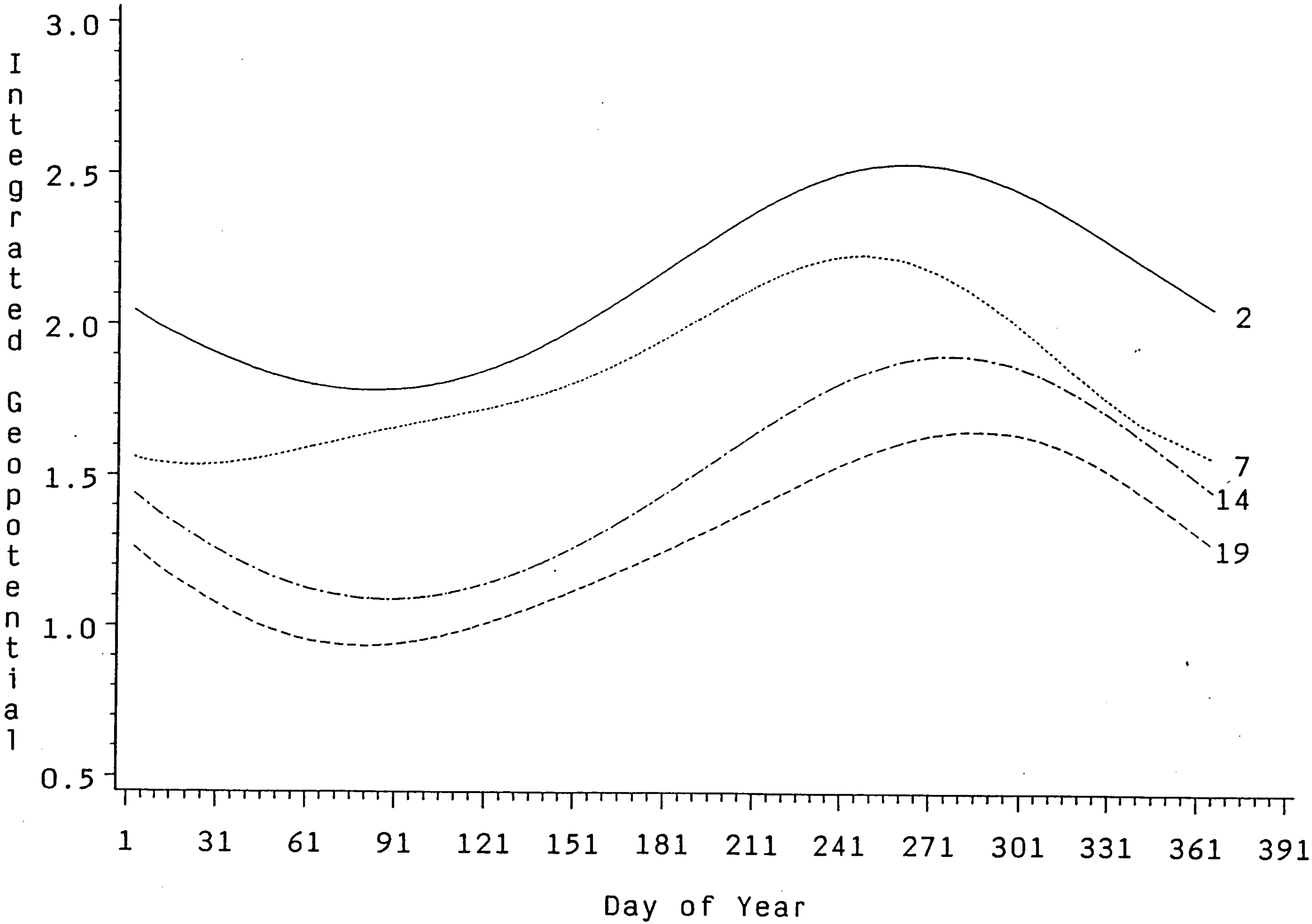


Fig. 3. Seasonal cycle of geostrophic transport on the Flemish Cap Line. (a) Transport between stations 2 and 7 over the Grand Banks, and (b) between stations 7 and 14 through the Flemish Pass. Each data point represents the transport calculated for a single transect. The seasonal cycle is shown by the dashed line. Transports calculated for top 100 m.

Fig. 4. Seasonal cycle in the geopotential vertically integrated over the upper 100 m and divided by the Coriois parameter for Stations 2, 7, 14, and 19 on the Flemish Cap line. The difference in values between stations gives the transport in Svedrups ($10^6\ m^3\ s^{-1}$).



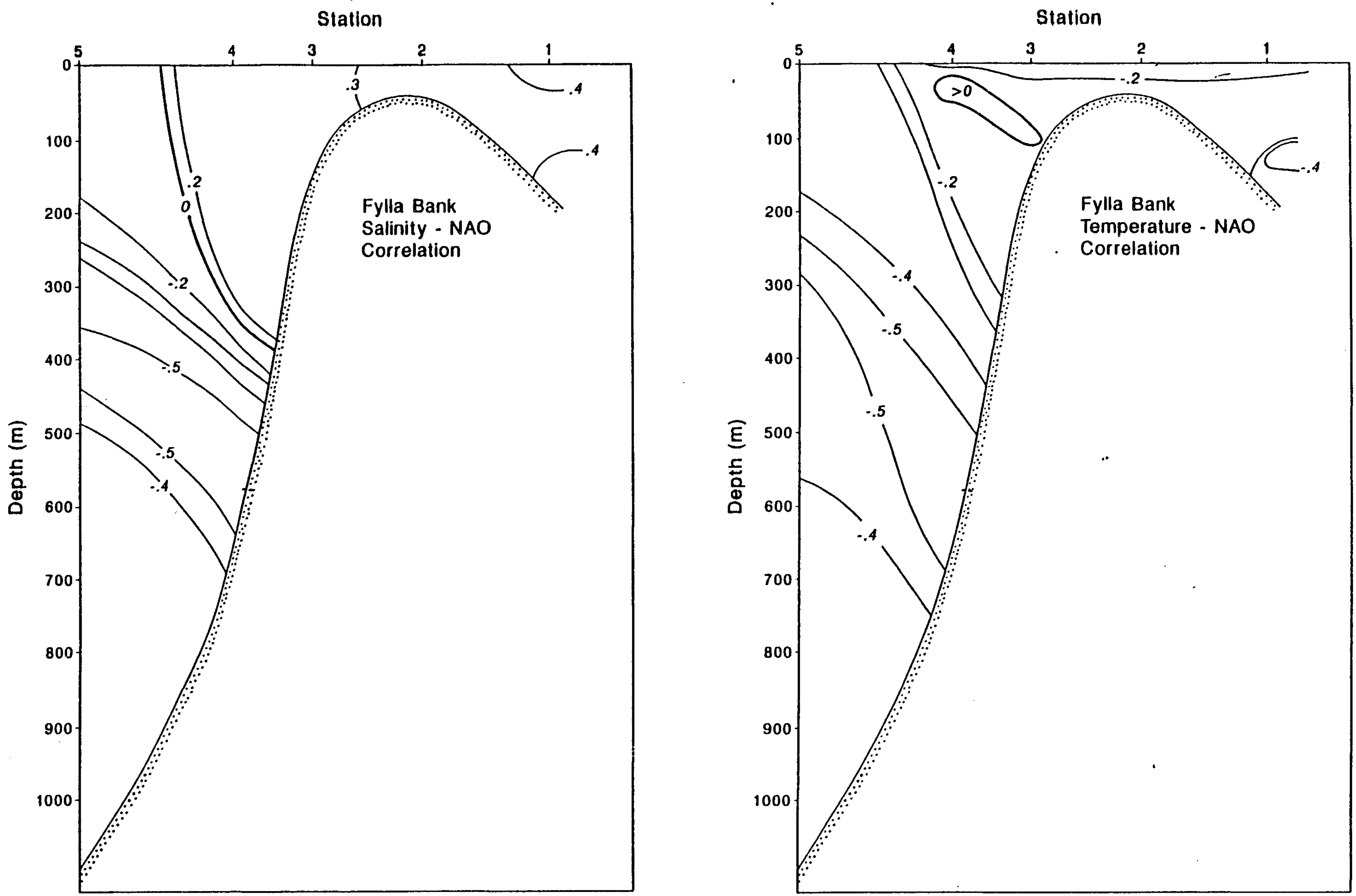


Fig. 5. Contours of correlation coefficient between the strength of the winter mid-lattitude westerlies in the North Atlantic Oscillation index (the NAO index) and the salinity and temperature on the Fylla bank section the following July.

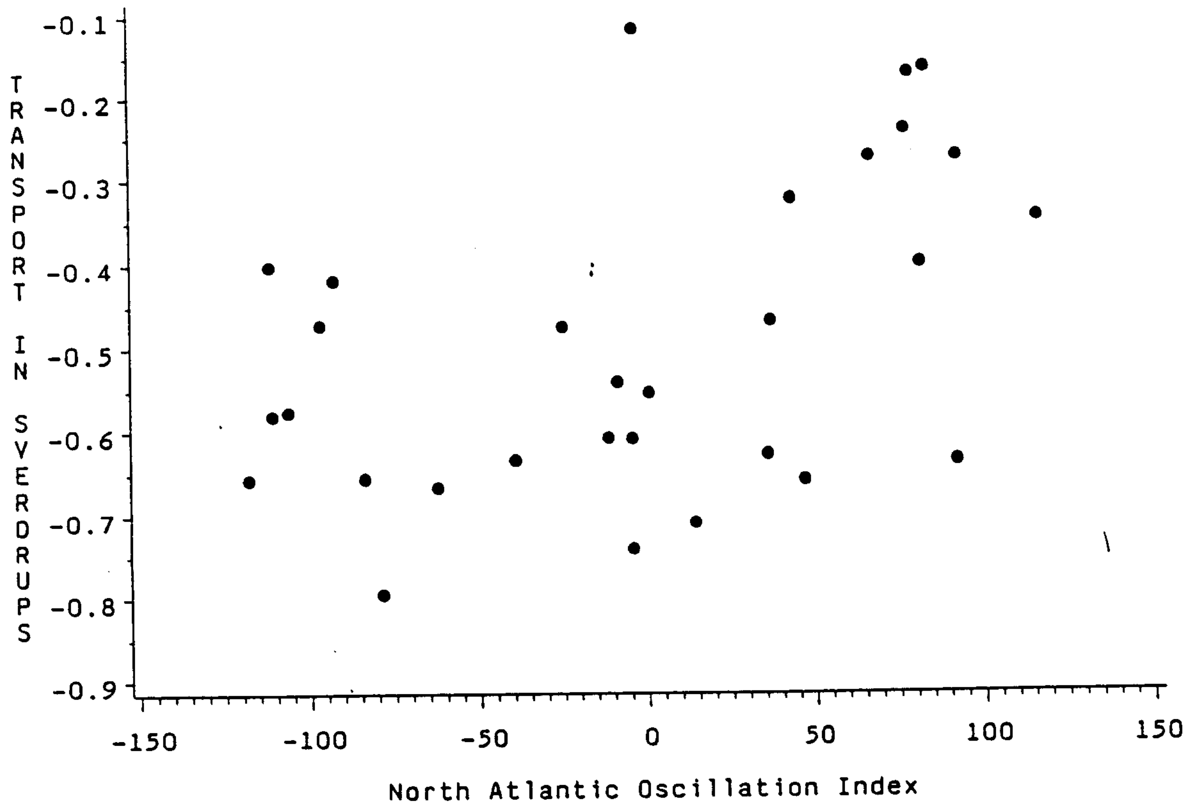


Fig. 6a. Summer baroclinic transport relative to 100 m through Flemish Pass versus the North Atlantic Oscillation index (the NAO index). Southward transport is negative.

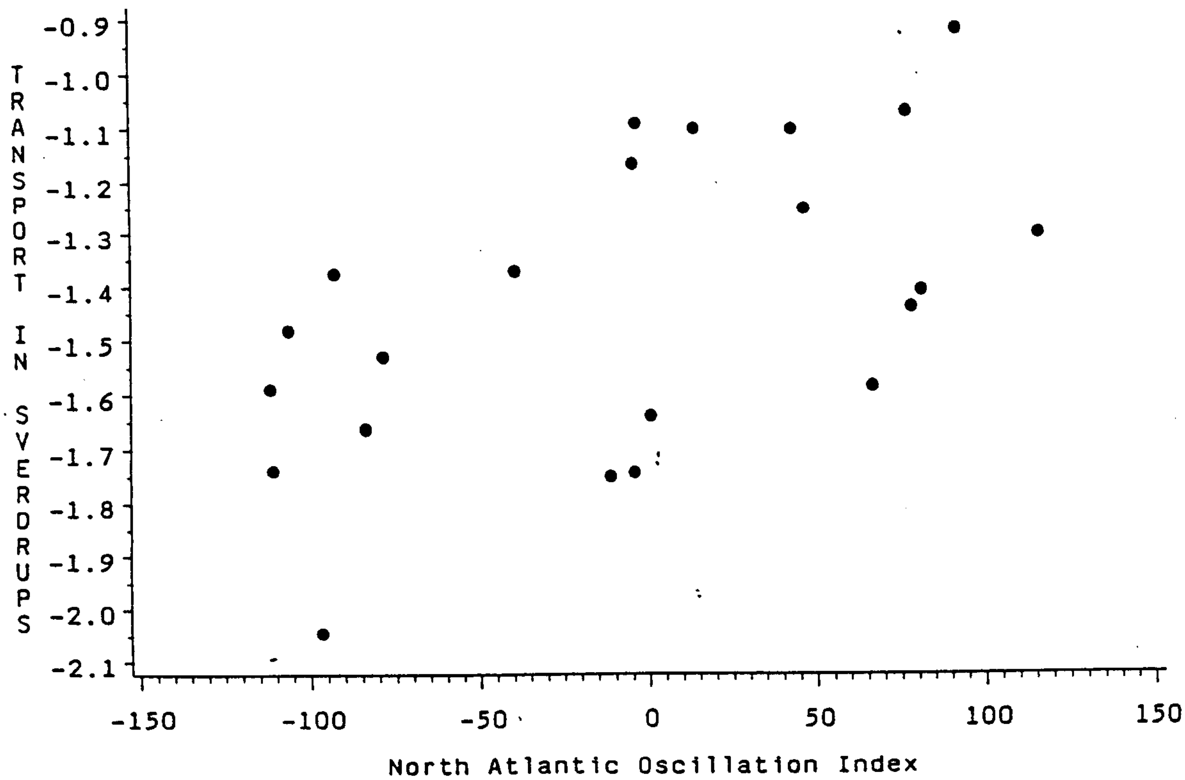


Fig. 6b. Summer baroclinic transport relative to 100 m through the Seal Island Line versus the North Atlantic Oscillation index (the NAO index). Southward transport is negative.

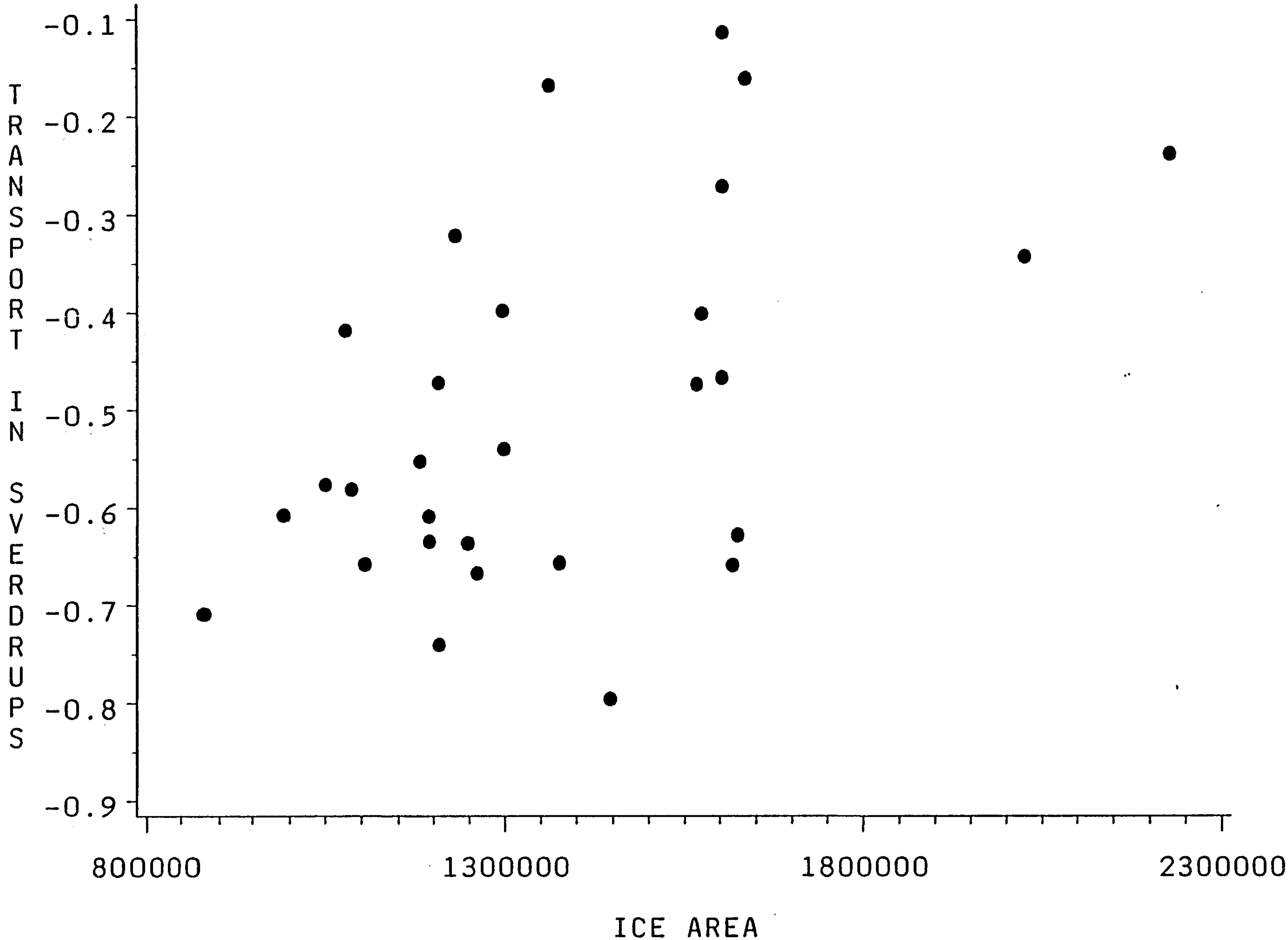


Fig. 7. Summer baroclinic transport relative to 100 m through Flemish Pass versus the area of ice in km^2 south of 65°N along the Baffin Island-Labrador coast from March to June. Southward transport is negative.