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Considerations of coastal circulation and fish production on the Scotian Shelf and in the Gulf of Maine

Part 1. Coastal circulation and physical oceanography

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Abstract

Correlations between annual catch of coastal commercial species of fish and the environmental factors of sea temperatures and St. Lawrence River discharge have led to an investigation of the relationship between the latter. Examining year-to-year variability of monthly means, effects of the St. Lawrence River discharge can be traced by correlation analysis with sea temperatures to propagate from the Gulf of St. Lawrence, onto the Scotian Shelf and through the Gulf of Maine at known coastal current drift speeds. Seasonal salinity and transport data support such a flow at least to a section off Halifax on the Scotian Shelf. Within the Gulf of Maine seasonal salinities do not support continuity of flow, however possible reasons and mechanisms for this are discussed. Other factors such as local river runoff in the Gulf of Maine, Labrador Current, and large scale weather systems are briefly considered and discussed. It is proposed that the Gulf of St. Lawrence to the Gulf of Maine inclusive be considered as an oceanographic system and events occurring in the southern part on time scales of a month or more are not independent of more northerly events. It is not interpreted that the river discharge is the driving force of such an oceanographic system but rather influences the water properties within the source region of the flow, i.e., the Gulf of St. Lawrence. Some biological implications of the Gulf of St. Lawrence to Gulf of Maine pathway are pointed out.

Introduction

Fluctuations in commercial catches and in the year class strength of fish and shellfish are well known but their causes are poorly understood. These fluctuations are believed to be due in part to changes in the physical environment of the fish and have been the subject of many investigations (see for example ICNAF Special Publication No. 6: Environmental Symposium, 1965). Although temperature is the factor most often considered (e.g. Templeman and Fleming, 1953; Taylor *et al.*, 1957; Martin and Kohler, 1965; Flowers and Saila, 1972), river discharge (e.g. Sutcliffe, 1972, 1973), winds (e.g. Carruthers, 1951; Chase, 1955), residual drift (e.g. Redfield, 1939; Colton and Temple, 1961), water stability and eddies (e.g. Iselin, 1939) plus large scale weather systems (e.g. Dickson and Lee, 1972; Iles, 1973) also have been investigated. These investigations suggest that many environmental factors may be important; consequently, studies of associations between only one environmental factor and fish must be viewed with caution. However, such studies are useful in establishing that associations do or do not exist and, hopefully, detailed examination of such associations will lead to a better understanding of the actual connection between fish and their physical environment.

Our interest in the relations between commercial catches on the Continental Shelf and environmental factors and among the factors themselves was aroused when, studying Martin and Kohler's (1965) correlation between cod records from George's Bank and temperatures at St. Andrews, N.B., we found similar correlations with discharge records of the St. Lawrence River. Cursory examinations of a few other species were equally interesting. Were the temperatures and discharge somehow related?

Similarities in the year-to-year variations of sea temperatures on the continental shelves of northeastern North America are well known and have been discussed by several authors. Sterns (1965) finds agreement between signs of temperature anomalies among northeastern United States stations. Hachey (1961) and Lauzier (1965, 1967a) have described the marine climate of the Scotian Shelf area from the temperature variations at St. Andrews, N.B. Lauzier (1965) in discussing long-term variability states:

> "Trends of surface water temperatures at St. Andrews, N.B., are representative of surface temperature variations over a large area of the Atlantic Seaboard from, at least, Halifax, N.S., to Atlantic City, N.J."

He further notes:

> "Bottom temperature variations and trends on the Scotian Shelf, in the Bay of Fundy area, and the deeper layer temperature variations in the Laurentian Channel, are similar to those of surface temperatures at St. Andrews with minor differences."

Most recently Lauzier (1972) has reviewed these similarities and has also shown that like variations occur within the Gulf of St. Lawrence. Colton (1968a) examined offshore temperature conditions in the Gulf of Maine during specific months in the years 1953, 1964, 1965 and 1966, and in reference to year-to-year variability wrote, "The trends in offshore temperatures at the surface and within water masses paralleled trends in the surface temperatures at Boothbay, Maine." Colton (1968b) also found that for the years 1955 to 1966 the monthly and annual mean temperature trends at 200 m showed similar behaviour to surface temperatures. These similarities existing throughout the entire continental shelf areas from the Gulf of St. Lawrence south to at least the Gulf of Maine suggest a common factor or factors influencing the year-to-year temperature variability.

The similarity in the year-to-year fluctuations in river discharge from the St. Lawrence River and temperatures at St. Andrews, New Brunswick, was first pointed out by Elizarov (1965). Plots of these signals plus Boothbay Harbour sea surface temperatures are shown in Figure 1. The likenesses in the plots are not so surprising if one considers the following: the mean annual freshwater discharge of the St. Lawrence system into the Gulf of St. Lawrence is 424 km^3/yr (Trites, 1970a), a quantity greater than the sum of the entire fresh water discharge of the eastern United States between Canada and southern Florida, 353 km^3/yr (Meade and Emery, 1971). The fresh water from the St. Lawrence moves seaward mixing with sea water to form a low salinity surface layer in the southwestern portion of the Gulf of St. Lawrence stretching from the St. Lawrence estuary to Cabot Strait. This layer eventually flows out through Cabot Strait and is the major contributor to the total outflow, approximately 3×10^4 km^3/yr (MacGregor, 1956). An outflow of this magnitude, flowing southwest, is enough to replace much of the volume of water on the continental shelf between the Laurentian Channel and Cape Cod, 3.6×10^4 km^3 (Barinov and Bryantsev, 1971), in one year. Indeed water from the Gulf of St. Lawrence was found by McLellan (1954) to cover the major part of the Scotian Shelf.

It is recognized that temperature and discharge variations must be viewed as part of a larger system of complex interactions and feedback networks. Thus reasons are not sought for the variability in the sea temperatures beyond those waters directly affecting the Gulf of Maine or the Scotian Shelf nor in meteorological or hydrological phenomena. The possible role of large scale weather systems on the hydro-meteorological climate of the northwestern Atlantic coast is briefly discussed in a later section.

Although this study mainly considers temperature variations, mostly because of availability of data, transport and salinity data have been used where possible. We are concerned with the area of the Shelf from the Gulf of St. Lawrence to the Gulf of Maine and the primary investigative tool used is correlation analysis.

General circulation pattern

In any area, year-to-year changes in the sea temperature or any physical characteristic are determined in part by changes in the character or relative volumes of the constituent water masses. Before examining these details and to emphasize the continuity of the coastal regime the gross circulation along the northwestern Atlantic coast is presented below and summarized in Figure 2.

The Labrador Current is a southward-moving continental shelf current containing, relative to its surrounding waters, cold, low salinity water (Iselin, 1927). It is comprised of two streams, one inshore on the continental shelf and one offshore over the continental slope and the outer edges of the shelf (Smith *et al.*, 1937). The inshore stream contains the greater volume of cold water, being a mixture of waters of the Baffin Land Current and water flowing eastward through Hudson Strait (Smith *et al.*, 1937). The offshore stream contains waters characteristic of the warmer West Greenland Current (Smith *et al.*, 1937).

On the journey southward some exchange of Labrador Current and Gulf of St. Lawrence waters occurs at times through Belle Isle Strait (Hachey *et al.*, 1954). As the Labrador Current reaches the latitude of St. John's it splits, the slope branch continuing down the eastern edge of the Grand Banks while the inshore branch follows the Avalon Channel past Cape Race (Smith *et al.*, 1937). The inshore branch loses its Arctic characteristics in the vicinity of St. Pierre Bank and, although the current may reach into the Gulf of St. Lawrence (Hachey *et al.*, 1954), the origin of the cold intermediate layer in the Gulf is in situ cooling (Banks, 1966). In general, the waters in the Gulf flow northward along the west coast of Newfoundland, are eventually deflected towards the Quebec shore and flow towards Anticosti (Hachey *et al.*, 1954). Some exchange occurs through Belle Isle Strait as previously mentioned. After the waters pass around the northern end of Anticosti they join with the easterly, fast-flowing Gaspé Current. These waters flow out onto the Magdalen Shallows, slowing down in the process (Hachey *et al.*, 1954). There is an outflow through Cabot Strait named the Cape Breton Current by Dawson (1913). These waters flow southeasterly along Cape Breton and out onto the Scotian Shelf. The waters on the Scotian Shelf are a mixture of Gulf of St. Lawrence water and more saline waters from offshore called Slope Water (McLellan, 1954). In general, there is a southwesterly flow parallel to the Atlantic coast of Nova Scotia which was called the Nova Scotian Current by Bigelow (1927). This current rounds the tip of Nova Scotia and at times enters into the Gulf of Maine (Bigelow, 1927). Here there is an area of relatively intense and constant upwelling (Lauzier, 1967b). The circulation in the Gulf of Maine is mainly cyclonic, moving into the Bay of Fundy on the Nova Scotia side and out on the New Brunswick side then along the coast of the U.S. as far as Cape Cod (Bigelow, 1927). Part of these waters then flow northeasterly to Georges Bank where they again split, part towards the Bay of Fundy and part turning southward to flow back along the U.S. coast (Bigelow, 1927). The amount of flow in each direction depends on the time of year (Bumpus and Lauzier, 1965). Tidal mixing is a prominent feature of the Bay of Fundy (Hachey and Bailey, 1952). Slope water is a major constituent of Gulf of Maine waters (Hachey *et al.*, 1954). Effects of the Gulf Stream on the Gulf of Maine and Scotian Shelf waters are felt through its effects on the Slope Water characteristics.

Hachey *et al.* (1954) have an excellent summary of the general oceanic circulation. Bumpus (1973) provides a more detailed review of the circulation in the Gulf of Maine as well as the circulation southward over the continental shelf to Florida.

In summary, then, the oceanic influences which directly affect the waters in the Gulf of Maine or on the Scotian Shelf include the Nova Scotian Current, the Cape Breton Current, and the Slope Water. These in turn are directly affected by the Labrador Current, the Gulf Stream and the fresh water discharge from the St. Lawrence River. The Gulf of Maine waters are influenced by local fresh water run-off but the small contributions from rivers in Nova Scotia are not felt far beyond their point of discharge. Also affecting the Gulf of Maine and Scotian Shelf waters are on-site meteorological factors, such as solar heating, wind patterns, cloud cover, precipitation, and evaporation.

Data

All data series used in this report are listed in Table 1 with the sources. Some require comment. For locations see Figure 3.

Sea temperatures

Continuous daily sea temperature records of various years duration have been collected at a number of coastal stations within the area of interest. The stations range from Entry Island, Quebec, to Boston, Mass., (see Fig. 3). All except the lightships contain only surface temperatures. Missing data have been linearly interpolated in time from adjacent points.

Temperature data are also available at Station 27 off St. John's, Newfoundland, and Prince 5 off St. Andrews, New Brunswick, on a monthly or bimonthly basis for over 20 years. Temperature data taken at irregular intervals exist at five stations across Cabot Strait (between Cape Ray, Newfoundland, and Cape North, Nova Scotia) and at eight stations on a line southeast of Halifax to the edge of the Shelf called the Halifax Line.

Salinity data

Continuous long term records of salinity are scarce. Records of greater than 10 years duration exist for each of Portland, Boston and Nantucket lightships. Twenty years of data are available for Station 27 and Prince 5 on a once or twice a month basis and data on an irregular time basis are also available for Cabot Strait and the Halifax Line stations. Short series of salinity data are also used in the Gulf of St. Lawrence within this report.

Fresh water run-off St. Lawrence River System

The gauged run-offs from the St. Lawrence River (Lake Ontario outflow), the Ottawa River (Grenville Gauge #22B1 to 1962, Carillan Dam Guage #3118), and the Saguenay River (Centrale d'Ille Maligne) are combined into a river signal called RIVSUM.

Within this study the monthly averages from the three gauges are considered to be synoptic. For the years 1960 to 1970 inclusive when comparisons can be made with estimates of the total years fresh water discharge (Jordan, 1973), RIVSUM accounts for, on the average 79% of the total to Tadoussac (including the Saguenay) and 68% of the total to Pointe-des-Monts. The minimum amounts represented by RIVSUM for that same period are 75% in 1970 to Tadoussac and 64% in 1962 to Pointe-des-Monts. The monthly means of RIVSUM from 1914 to 1973 appear in Table 2.

Bay of Fundy and Gulf of Maine rivers

The gauged run-offs from the Saint John River in New Brunswick, the Penobscot, Kennebec and Androscoggin Rivers in Maine, and the Merrimac River in Massachusetts are considered synoptic for this study (after Meade and Emery, 1971) and are combined into a river signal called MAINER. It covers the years 1931 to 1970.

Air temperatures

Long records of air temperatures are available for many locations. Eastport, Maine; Halifax and Sable Island, Nova Scotia; Fredericton, New Brunswick; and Ottawa, Ontario are used in this study.

Methods and Statistical Considerations

To investigate the relationship between sea temperatures and St. Lawrence River discharge, correlation coefficients are calculated between monthly values of RIVSUM and sea temperatures. The data are separated by month and each of the monthly sets are averaged using three-year equally-weighted running means. The monthly river signals (RIVSUM) are progressively lagged behind the monthly temperature signals one month at a time (e.g. Dec. RIVSUM *vs* Jan. temperature) beginning with no lag and proceeding to a 12-month lag.

Correlation coefficients are calculated for each lag and each temperature month. This results in a 12 x 13 matrix of correlation coefficients where each of the horizontal rows represent correlations at one particular lag time of river for each month of temperature, the vertical columns represent correlations between one temperature month and lags of RIVSUM from 0 to 12 months and the left to right diagonals represent correlations between one set of monthly river data and each of the 12 months of temperature.

Table 3 is the correlation matrix between Boothbay Harbour sea surface temperatures and RIVSUM. The horizontal line corresponds to a three-month lag between the temperature and river signals, the vertical line corresponds to lags in the river signal from 0 to 12 months behind the October temperatures and the diagonal line follows the progressive lags of May rivers. These examples have been chosen for demonstration purposes only.

Oceanic, hydrological and meteorological data exhibit marked persistence, a feature which influences the significance levels of correlation coefficients using data of this type. The common procedure of determining significance levels using the total number of pairs of datum points and statistical tables of correlation coefficients is based on the assumption that each observation within the individual time series is independent. Statistically this requires the autocorrelation coefficients to be small for lag periods greater than zero. The time series being considered have not met such requirements. However, for autocorrelated time series, Bayley and Hammersley (1946) defined "the effective number of independent observations", n^*, by

$$\frac{1}{n^*} = \frac{1}{n} + \frac{2}{n^2} \sum_{j=1}^{n-1} (n-j)\,[\rho(j\tau)]^2 \tag{1}$$

where n is the total number of observations and $\rho(j\tau)$ is the autocorrelation coefficient of the j^{th} lag of period τ. As $j \rightarrow n-1$ the error in $\rho(j\tau)$ can become large due to the decreased number of points included in calculating $\rho(j\tau)$. As a result calculations of n^* within the present study are summed up to and including

$$j = \frac{n}{10} \tag{2}$$

This value is suggested in Blackman and Tukey (1958) on empirical grounds. For power spectra analysis they require that j should not exceed 10% of the total record in order to hold the rms deviation of each estimate below one-third of its average value. Small changes in n^* were found by increasing j from n/10 to n/2. On this basis a termination point of j = n/10 was considered reasonable. For series being cross-correlated, the effective number of independent pairs is taken to be equal to the lower n^* of the two series being considered. Correlation significance is then determined with (n^*-2) degress of freedom. This procedure produces a more representative value of the true statistical significance than estimates solely based on the total number of datum points. Hence the significance levels quoted within this paper are based on Bayley and Hammersley's (1946) procedure subject to the conditions in (2).

Results and Discussion

The temperature stations containing daily records are divided into two groups. The first, referred to as Group I, contains records of 30 or more years duration which necessarily limits the data to surface values. The stations include Entry Island, Sambro Lighship, St. Andrews, Boothbay Harbour, and Boston Harbour. The second group (Group II) has the years 1956-69 in common and include mostly subsurface values. The stations are Halifax Line (Inshore), Lurcher Lightship, Portland Lightship and Boston Lightship. Entry Island sea surface temperatures are included in the latter group as subsurface records for these years are not available in the Gulf of St. Lawrence. Correlation coefficients are calculated between these stations and RIVSUM. Several river months (the diagonals in the matrix) showed relatively high values. For Group I the highest correlation coefficients occurred with March RIVSUM but at various months of temperature depending on the particular station. Sambro Lightship was the exception with the highest value occurring with April river but relatively high values were also observed with March river. For comparison purposes the correlation coefficients for March river (the third diagonal in the matrix) are plotted for all Group I stations (Fig. 4a). For Group II the July river signal showed consistently high values and is therefore plotted (Fig. 4b). This difference in the river months between the two groups is discussed later.

The temperature month at which the maximum coefficient occurs varies from station to station but in general increases in time from the river month with distance from the river. The lag times between the river and temperature month at which the peak occurs are listed in Table 4. The peak is chosen as the month of the maximum coefficient unless several months of nearly equally high values occur. For example, at Boothbay Harbour the peak is chosen at a lag corresponding to the middle year of the crest. (Peaks occurring at lags of 1 or 0 months have been disregarded but will be discussed later). The statistical significance of the peaks, as determined by the method in the preceding chapter, is also given in Table 4. Enough data are not available for Group II stations to make meaningful estimates of their statistical significance.

If the lag times in Table 4 are assumed to represent the travel time for the effects of a particular month's discharge to be felt at the temperature station in question, then speeds can be calculated knowing the distance between the river and the temperature station. Figure 5 plots distance from the river, assumed to begin at the mouth of the Saguenay, against the lag times for the Group I and II stations as per Table 4. The slopes therefore represent average speeds. For Group I stations the speed is 6.5 km/day (3.5 naut. mi./day). For Group II stations the speed is approximately 9.2 km/day (5.0 mi./day). Combining all the data gives a speed of 7.8 km/day (4.2 mi./day). Ocean drift speeds calculated by Bumpus and Lauzier (1965) from drift bottle recoveries in the areas under consideration are between 2 and 7 miles per day but vary with the time of year. Forrester (1971) following oil from a spill originating near Canso, Nova Scotia, observed drift speeds of 4.3 miles/day southwestward along the coast of Nova Scotia. Comparison of the average speeds calculated from the correlation analysis with these measured values show good agreement.

Before proceeding to salinity and transport data some questions arise from the correlation analysis. The effect on sea temperature is found not to be the same for all months. The highest correlations at all temperature stations occur with one particular river month, which does not coincide with the maximum discharge month. It is known from drift bottle recoveries (Bumpus and Lauzier, 1965; Trites and Banks, 1958) that the surface flow on the Scotian Shelf varies with the season. Seasonal variation in the surface circulation of the Gulf of Maine is also well documented (Bigelow, 1927) being closely associated with the wind pattern (Bumpus and Lauzier, 1965). These seasonal variations suggest that the effect may be transported only at certain times of year, thereby offering one explanation for differences in the correlation coefficients between various river months.

Covariances have also been calculated between the temperature stations and RIVSUM. Covariances are the unnormalized products of two time series and therefore represent a rough measure of the magnitude of the fluctuations involved. A low convariance means small amplitude fluctuations. Covariances determined for both Group I and Group II sea temperature stations show relatively high values at the peak times listed in Table 4. High covariances were also generally observed with May or April river months. These large amplitude fluctuations, although not coupled together with sea temperatures as closely as for some river months may produce as much effect.

Comparisons in peak lag times made between Groups I and II show reasonable consistency considering the relative crudness of the technique. The station with the poorest fit in the progressive peak lag times in either group is Sambro Lightship. This may be due to local wind-induced upwelling during the summer (Longard and Banks, 1952), which is the season the effects might be expected from a March RIVSUM. Boston Lightship, also located in an area of wind-induced upwelling during summer (Kangas and Hufford, 1974), is observed to follow the pattern of correlation results. However, the effect of the river does not arrive during the season of expected upwelling.

Peaks are observed to recur at all stations if the river is either lagged or advanced beyond a time of 12 months. An example is plotted in Figure 6. The period of recurrence is 12 months and is believed due to the high autocorrelation or persistence within each particular monthly series. Persistence means one year's discharge is much like the next year's and so with temperature three year averaging contributes to this. Recurrence of the peaks is therefore not unexpected. Persistence can also explain why the correlations are high at small lag times.

To determine the effect of lowest frequency trends in the data a least squares line was calculated for all data and subtracted out. Correlations run with this trend removed produced similar results to correlations with the trend included. Slight decreases in both the height of the peaks and the depth of any troughs were observed.

The difference in peak lags between stations cannot be accounted for through progression of the seasonal temperature signals. For all stations the minimum temperatures occur in February or March while the maxima occur in August-September for surface values and October-November for subsurface values.

Correlation analysis with salinity, similar to that used for temperature, is desirable but long series of continuous data are not available. Seasonal salinities at several locations are available, however, factors important in determining seasonal patterns are not necessarily important in determining the year-to-year variations. With these reservations in mind the seasonal salinities are plotted in Figure 7. The salinity minima progress in time from June-July at Grand Rivière in Quebec to November off Halifax. If these minima are assumed to represent the effects of the peak discharge from the St. Lawrence River (May, see Fig. 8) then it takes four months to travel through the Gulf of St. Lawrence to Cabot Strait (also see Lauzier, 1957b) and another two months to reach Halifax. These times agree very favourably with the temperature correlation results. Beyond Halifax this simple picture no longer applies. Two reasons might explain this: the flow at this time of year may not reach into the Gulf of Maine or it may be masked by local effects.

Bailey (1957) found a seasonal pattern in the relative amounts of 'coastal' to 'Slope Water' at two stations in the Bay of Fundy (one station being Prince 5) between the years 1950 and 1955. 'Coastal water was assumed to have a salinity of 30.8°/oo and 'Slope Water' a salinity of 35.2°/oo. The percentage of coastal water was minimal between October and December and maximal in April to June. If the salinity minimum in November at Halifax progressed into the Gulf of Maine and Bay of Fundy it would be expected sometime between January and March. In fact, a sharp increase in coastal water (thus lower salinity) was observed between December and March by Bailey (1957) preceding the April-June minimum during the two years when data was collected at a central station in the mouth of the Bay of Fundy. At a station closer to the New Brunswick coast (Prince 5) local run-off was more influential and although the amount of coastal water increased at about the same time the rise was not as steep as at the central station. It is therefore possible that some of the salinity minimum at Halifax does reach Prince 5 but does not appear as a minimum due to masking by local conditions. The departure southward of this former Nova Scotia Current water may explain Bailey's observations that the salinity minimum usually occurred at both stations prior to the peak discharge of the nearby Saint John River, the largest river entering the Gulf of Maine.

The question therefore arises--if salinity trends are masked, why is the temperature signal observed (Figure 4a and 4b)? The answer to the question may lie with Colton (1968b). He observed that during warm years in the Gulf of Maine the boundary between the Slope Water and 'coastal' water is nearer the coast than in cold years. Analysis of the Halifax Line data shows that the vertical cross sectional area of the layer between the warmer upper and lower layers called the cold intermediate layer by McLellan, *et. al.* (1952) is reduced in years of high St. Lawrence discharge, i.e. warm years (see Figure 9). During these warm years the Slope Water is warmer and saltier than normal being comprised of larger proportions of North Atlantic Central Water and lesser of 'coastal' water (Colton, 1968b). At the same time our findings suggest the Nova Scotian Current is delivering warmer and fresher water than normal to the Gulf of Maine. Gulf of Maine water is a mixture of 'coastal' and Slope Water (Bailey, 1954) thereforein a warm year temperature of the two water masses will aid while their salinities will oppose. The relative volumes of each type of water would determine whether the salinity would increase or decrease. The mechanisms proposed explain how the temperature signal may be enhanced within the Gulf of Maine and the salinity signal may be masked. This explanation is, however, speculative.

Some attempts at long term relationships between salinity and temperature exist within the literature. Lauzier (1964) has observed that with the deep waters in the Bay of Fundy (175 m) and Emerald Basin (200-240 m), in the long term, increases in temperature are accompanied by increases in salinity and decreases in temperature by decreases in salinity. Colton (1968b) observed similar temperature-salinity relationships at 200 m in the Gulf of Maine. Colton also found that these variations occurred due to the composition of the offshore waters as well as the volume of their indraft through the Northeast Channel. It has not yet been shown that such a long-term relationship between salinity and temperature exists in the surface layers in the Gulf of Maine or on the Scotian Shelf.

In the Gulf of St. Lawrence, Lauzier (1957a) found that on the Magdalen Shallows during the years 1945-49 the minimum salinity varied inversely with the estimated April to June run-off from the St. Lawrence watershed. The correlation analysis shows high temperatures associated with high run-off. Combining these results with Lauzier's, it suggests that increases in temperature are associated with decreases in salinity and decreases in temperature with increases in salinity in the Gulf of St. Lawrence.

More work is required to establish the long term temperature-salinity relationships in the surface layer in the area under consideration.

Seasonal transport data is even more restricted, especially in spatial coverage, than salinity data. Cabot Strait and the Halifax Line provide sufficient temperature and salinity information to calculate geostrophic transports. For Cabot Strait the condition of zero net salt flux as discussed by Forrester and El Sabh (1972) is used. For Halifax Line a level of no motion coincident with the seabed is chosen. The net transport through the Halifax Line and the southward transport through Cabot Strait are plotted in Figure 8, along with the seasonal pattern of RIVSUM. If the maximum transports are assumed to correspond to peak discharge, the lag time between the river and Cabot Strait is about four months with another two to three months to Halifax. The maximum transports occur at times of minimum salinity (see Figure 7).

Figure 10 shows the seasonal transports for three layers of depths, 0-50 m, 50-175 m, and 175-450 m, as well as the total transport for both the northern and southern halves of Cabot Strait. The amount of data used to calculate these bimonthly transports are not equal for each grouping. The number of transects used for each two-month group, except January-February, is equal to or exceeds five. This number is given by El Sabh (1974) as the minimum required to average out short term density fluctuations for Cabot Strait. For the period January-February only four transects are available.

From Figure 10, the outflow through Cabot Strait is shown to be mainly concentrated in the upper layers of the southern half and is mostly balanced by an inflow on the northern half at intermediate depths. Most of the outflow is derived from over the Magdalen Shallows (Hachey *et al.*, 1954; Trites, 1970a; El Sabh, 1974). These waters were in turn influenced by the river discharge (Lauzier, 1957a, 1957b). The northern inflow will be discussed later within the paper. The maximum and minimum transports, both inflow and outflow, for both halves of the Strait occur in September-October and March-April respectively, agreeing reasonably well with MacGregor (1956) and El Sabh (1974).

In summary, the effects of the discharge from the St. Lawrence River can be traced by correlation analysis to travel at certain times of the year at approximate ocean speeds from the Gulf of St. Lawrence, along the Scotian Shelf and through the Gulf of Maine. Insofar as analogies can be made between seasonal and year to year fluctuations, the salinity and transport data support the correlation results as far as Halifax.

Other considerations - Gulf of Maine Rivers

Correlation coefficients between temperature stations in the Gulf of Maine and the combined discharge of the five largest rivers entering the Gulf (MAINER) were also calculated. Three temperature stations containing more than 30 years of records (St. Andrews, Boothbay Harbour, Boston Harbour), and three stations with subsurface records during the years 1956-69 (Lurcher Lightship, Portland Lightship, and Boston Lightship) were used. For the stations containing the longer temperature records, the correlations are relatively low and produce no apparent patterns. The correlation coefficients seldom reach about 0.7 and only at St. Andrews is a coefficient of 0.8 achieved. This occurs with April temperatures and February river discharge, that is at a lag of 2 months. At subsurface stations for the years 1956-69 high correlation coefficients are observed with the highest occurring with July MAINER. The highest values appear at lags of 4 months for Portland Lightship and 3-4 months at Boston Lightship. The statistical significance of these latter values cannot be determined due to the shortness of the record.

These correlation results suggest that year-to-year changes in sea temperature are not determined to a large extent by fluctuations in local discharge. During certain periods of years it may become important but this is not well established. Based only upon correlation analysis, the variations of the St. Lawrence River discharge are more closely associated with the year-to-year changes in the sea temperatures in the upper layers in the Gulf of Maine than are the variations of the local run-off. This is a general result applied to the Gulf as a whole and does not necessarily hold at all locations.

Labrador current

Some discussion of the Labrador Current is required. It, as well as wind stress (Murty and Taylor, 1970), may be a major influence on the circulation pattern within the Gulf of St. Lawrence controlling inflowing transport through Cabot Strait. Water entering Cabot Strait, mostly on the north side (Figure 10), is probably in part of Labrador origin.

Long term records on the properties of the Labrador Current are scarce. The U.S. Ice Patrol collected oceanographic data between 1935 and 1965 but these observations are primarily limited to the months of June and July. The longest continuous series of data has been collected by the Canadian Fisheries Research Board Biological Station at St. John's, Newfoundland, which since 1946 has taken measurements at least once per month at Station 27, 2 nautical miles off Cape Spear, Newfoundland.

Correlations between Station 27 temperatures at 30 m depth and Group II stations for the years 1956-69 show coefficients greater than 0.8 at various lags depending on the temperature station used. However the peak lags do not show any progressive pattern, being greatest at Entry Island (10 months) and least at Halifax (7 months) with the remaining stations exhibiting peak lags between these values.

The trend of the transports of the inshore Labrador Current between 1948 and 1963 calculated from measurements by the U.S. Ice Patrol (Annual Report, 1964) shows little resemblance to temperature or salinity variations in the Gulf of St. Lawrence or southward. Indeed the transports change little from year to year being slightly less than one sverdrup, a value supported by the measurements of Kudlo (1973). However, the Ice Patrol transports are based on only one to three cruises each year and usually occur during the late summer. Such limited data make meaningful comparisons difficult.

Within the literature several authors have observed or postulated that the Labrador Current affects the waters of the Scotian Shelf or Gulf of St. Lawrence. Labrador water has been observed on the Scotian Shelf off Banquereau Bank (McLellan and Trites, 1951) but no documented evidence exists to suggest it reaches farther south (see also Bumpus, 1973). The idea that the inshore Labrador Current may reach into the Gulf of St. Lawrence through the north side of Cabot Strait was originally suggested by Bjerkan (1919). This has been supported by observations of cold water moving westward along the northern side of the Laurentian Channel (see Lauzier and Trites, 1958; Kudlo and Burmakin, 1972; Lenz, 1973). The inflow observed through the northern side of Cabot Strait (see Figure 10; Trites, 1970a; El Sabh, 1974) is therefore generally considered to be in part of Labrador origin. Whether this inflow is responsive to other driving forces or whether it in itself is a driving force of the water movements in Cabot Strait has yet to be determined.

Lauzier and Trites (1958) observed that long term changes in the temperature of the deep layer in the Laurentian Channel from its mouth to Cabot Strait followed similar patterns to changes in the temperature of the Labrador Current. They also noted that the proportion of Labrador water in the core of the deep water was consistent between years.

The offshore branch of the Labrador Current has not been discussed in detail within this paper as it cannot be traced, as a current, much beyond the Tail of the Grand Bank (Lee, 1970). It is shown by Lee (1970) to influence the composition of the Slope Water below 200 m. Any effect of the offshore branch is therefore thought to be indirect through its effect on the Slope Water.

Many questions remain on the role of the Labrador Current in the year to year variability on sea temperatures in the Gulf of St. Lawrence and on the Scotian Shelf. The data to date do not allow for detailed analysis of its role.

Large scale weather systems

Changes in large scale weather systems with subsequent effects on winds, precipitation, and air and sea temperatures in the North and Northwest Atlantic have been discussed by Bjerknes (1963), Namais (1966), Dickson and Lamb (1972) and Rodewald (1972). In particular, shifts in the intensities and in the relative positions of the Icelandic Low and the Bermuda-Azores High (Bjerknes, 1963) have been connected with the trends in sea surface temperatures. Worthington (1964), Stern (1965), Colton (1968a) and Lauzier (1972) have considered large scale weather systems in their interpretations of oceanographic results in the Gulfs of St. Lawrence or Maine or on the Scotian Shelf.

Air temperatures within the region being discussed respond to large scale weather systems. This is suggested on the strength of the similarities between the annual air temperatures of Ottawa, Ontario; Fredericton, New Brunswick; Eastport, Maine and Sable Island, Nova Scotia, as shown in Figure 11. The diverse locations and the distances involved between these stations require large scale phenomena to explain the similarities.

Long term similarities have been observed between Boothbay Harbour sea temperatures and Eastport, Maine, air temperatures by Taylor, *et al.* (1957) and between St. Andrews sea temperatures and Halifax and Sable Island air temperatures by Lauzier (1972). Similar results are shown in Figure 12 using annual means of St. Andrews sea temperatures and Eastport air temperatures. In view of these correlation coefficients between Group I and II sea temperature stations and Eastport air temperatures were calculated in a manner similar to that described earlier for sea temperatures and river discharge. No large differences were observed between the various sea temperature stations and no trends appeared. Coefficients at small lag times for all stations were generally below .7 These results suggest that air temperatures alone cannot account for sea temperature variability.

What then is the role of the large scale weather systems? As stated previously these systems will control wind patterns, air temperatures and precipitation. These several factors combine with the waters within the Gulf of St. Lawrence and the resultant effects, which are associated with the levels of the St. Lawrence River discharge, flow southward over the Scotian Shelf and into the Gulf of Maine. The waters within these geographic areas are therefore affected by this flow as well as by direct hydrological and meteorological factors which in turn are linked back to the large scale systems. One can view the water properties on the Scotian Shelf and in the Gulf of Maine as being made up of a stationary signal and a moving signal, both signals being determined by the large scale weather systems. The stationary signal is due to direct effects of hydrological and meteorological factors and the moving signal is due to the oceanic pathway which appears to begin within the Gulf of St. Lawrence and transports more northly events into the southern sections. The relative importance of the two signals cannot at present be resolved but the existence and the possible importance of the oceanic pathway should be recognized.

Biological implications

As noted in the introduction, there are a number of correlations between various environmental factors and fish production as exemplified by commercial catch statistics on the Continental Shelf. It is difficult to say whether these are direct effects (say optimal temperatures for larvae) or whether the environmental parameters are mere indicators in the local oceanographic climate accompanying other factors, so far scantily measured and poorly documented in terms of long time series. Surely, however, some of the relationships must originate in primary production and thus in the availability of nutrients.

From the descriptive oceanography of earlier sections one can point to a number of areas where subsurface water would be mixed with or transferred to the surface. In the Cabot Strait transports, as shown by Trites (1970a) with current meters, El-Sabh (1974) and in Figure 10 it is seen that most of the flow into the Gulf of St. Lawrence is below 50 meters while the outgoing flow is above 50 meters. This indicates translation of subsurface water to the surface somewhere in the Gulf surely accompanied by nutrient enrichment of surface layers to some degree. Specific areas of upwelling in the Gulf of St. Lawrence have been noted to occur: along the north shore (Lauzier *et al.*, 1957) thought to be wind induced (P. Vandall, pers. comm.), the St. Lawrence estuary influenced by the River (Neu, 1970; Steven, 1971), the New Brunswick shore of the western Gulf (Lauzier, 1967b), a possibility of upwelling just southwest of the Magdalen Islands (Lauzier, 1967b) and anti-clockwise gyres both stationary (west of Anticosti Is.) and moving (Blackford, 1967; Trites, 1968; El-Sabh, 1974). Internal tides in the area of the estuary (Forrester, 1970) may contribute to mixing of deeper with upper water there (Forrester, 1974). Several of these areas can be identified in appropriate infra-red satellite photographs we have examined and the correspondence with some of the maps of fish stocks in the Gulf of St. Lawrence (Kohler, 1968), especially herring, is interesting.

Any or all of these areas could be affected to a greater or lesser degree by river discharge as well as the other factors involved in Gulf circulation and could be enhanced or dampened. No attempt is made here to assign relative biological importance to these areas, although highest productivity and phytoplankton concentrations as presented by El-Sabh (1974) were in the northwest Gulf near the mouth of the estuary.

Continuing from Cabot Strait to the southwest along Nova Scotia, at least three phenomena have been identified which could be affected one way or another by coastal flow of varying quantity and quality. Lauzier (1967b) has noted two upwelling areas as indicated by seabed drifters. One is at the northeast corner of Nova Scotia next to Cabot Strait, the other to the southwest adjacent to the Gulf of Maine. The latter, in particular, is highlighted by biological activity, being the area of greatest lobster catch in all Nova Scotia, and is the largest non-breeding concentration in the northwest Atlantic of surface feeding sea birds (R.G. Brown, pers. comm.). In between, along the Nova Scotia coast, is apparently a narrow band of intermittent upwelling promoted by favourable winds (Hachey, 1937; Longard and Banks, 1952). Its influence must be quite nearshore otherwise Lauzier's drifters would have been affected - the lack of returns between the two areas mentioned above is striking. A short discussion of possible mechanisms in the two areas mentioned by Lauzier is included in the Appendix.

It is difficult to see how coastal flow from Cabot Strait exerts influence in the Gulf of Maine except indirectly or through the temperature correlations already discussed. However, mixing from the Bay of Fundy and coastal upwelling (Graham, 1970; Kangas and Hufford, 1974) are certainly nutrient sources and might respond in some faction. We will enlarge upon at least the effects of some environmental factors on commercial fish catch in Part II.

Summation

Within this paper, attention has been focussed on several individual factors, in particular the St. Lawrence River system, which are thought to be related to year-to-year fluctuations in temperature on the Scotian Shelf and in the Gulf of Maine. It is recognized that other factors may also affect and indeed may be as important as those already discussed. Limitation of data and availability of time have restricted detailed examination to the factors considered previously. The relative importance of the many factors in determining changes in the temperature and salinity in the Gulf of Maine or on the Scotian are unknown. In many cases some knowledge of a factor's effect on the seasonal patterns are known but their role in the year-to-year fluctuations and long-term trends, for example on a decadal time scale, remain obscure. Some associations are observed but lack of understanding in the processes involved makes it difficult to judge their relative influence. Some basic problems are not yet solved, for instance the importance of changes in the physical parameters of the constituent waters compared to changes in their relative volumes in determining fluctuations in the Gulf of Maine temperatures and salinities.

In the light of the above, the importance of the role of the St. Lawrence River discharge on the year-to-year sea temperature fluctuations on the Scotian Shelf and in the Gulf of Maine must be viewed cautiously. It cannot be considered as uniquely determining these fluctuations or indeed of being the major contributor. Correlation analysis does suggest that it is in part influential in determining these sea temperature changes.

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Table 1 - Data Series and Sources

	Ref. No.
[1]Boothbay Harbour surface water temperatures	39
Boston Lightship, surface and 30 m	5-14, 18-21
Boston Harbour surface water temperatures	32-35
Cabot Strait	25
[1]Entry Island surface water temperature	23
Eastport, Me., air temperatures	15-16, 31
Grande Rivière surface water temperatures	23-24
Fredericton, N.B. air temperatures	2
Halifax Line	25
Lurcher Lightship bottom water temperatures	23
Gulf of Maine rivers (MAINER), after Meade & Emery (1971)	36-38
Ottawa Air temperatures	1,3
Portland Lightship temperatures - 30 m	5-14, 18-21
St. Lawrence, Ottawa and Saguenay Rivers (RIVSIM)	17, 26-30
[1]St. Andrews, N.B. surface water temperatures	23
Sable Island air temperatures	4, 15-16
Sambro Lightship	23
Station 27, off St. John's, Newfoundland	40
Prince 5 off St. Andrews, N.B.	25
Gulf of St. Lawrence surface salinities	22

[1]Recent unpublished data, where needed, were obtained through the kindness of author or agency indicated by reference no.

Table 2 - Monthly means of RIVSUM (see Table 1 for source, $10^3 ft^3/sec.$)

14	293	267	268	334	421	430	355	297	281	266	271	265
15	256	241	262	302	464	383	338	300	290	309	305	280
16	277	278	268	427	625	515	393	326	299	321	335	315
17	279	276	274	376	487	574	482	398	357	339	362	315
18	280	267	288	384	517	485	432	358	361	374	410	350
19	326	294	318	417	654	550	376	337	328	374	373	334
20	279	256	275	365	493	448	342	312	290	277	289	290
21	279	261	316	420	547	378	316	292	283	322	305	292
22	266	243	273	430	536	425	366	329	316	302	286	257
23	239	229	227	296	431	431	324	291	320	283	267	279
24	271	253	266	353	541	529	400	356	346	341	300	301
25	244	246	297	396	473	442	360	324	276	266	278	292
26	256	226	228	286	439	463	381	322	283	290	377	343
27	287	281	321	357	441	456	401	376	301	331	385	351
28	327	317	317	453	683	447	426	365	377	426	410	356
29	335	311	351	451	615	582	439	407	374	354	355	304
30	320	324	344	392	520	507	484	422	361	323	302	285
31	274	261	264	346	365	372	315	282	271	287	305	285
32	286	293	308	405	426	340	374	351	409	453	406	310
33	292	273	283	438	520	413	304	278	260	246	242	233
34	223	227	235	365	497	408	331	280	270	284	263	269
35	234	237	264	328	381	348	334	308	267	279	278	249
36	237	222	303	385	592	469	308	274	268	329	316	267
37	301	304	313	363	549	383	318	303	318	332	358	312
38	270	279	328	464	504	394	342	386	324	308	299	278
39	268	257	270	346	505	433	402	349	304	311	304	272
40	257	255	252	325	462	503	378	317	310	289	287	286
41	288	276	280	392	379	317	295	286	316	350	354	310
42	293	285	319	398	513	401	330	316	305	325	344	299
43	288	301	318	394	574	529	399	371	366	345	348	336
44	318	306	315	340	397	367	337	325	320	318	309	301
45	270	282	349	447	478	442	379	342	355	388	367	344
46	345	328	377	401	442	427	372	336	322	333	339	336
47	331	340	343	460	624	693	445	385	374	362	341	336
48	307	301	347	425	454	398	366	355	339	326	316	306
49	309	316	345	311	458	426	355	318	313	310	292	298
50	319	322	326	407	436	456	388	343	329	323	332	343
51	348	334	390	596	527	408	409	374	362	369	409	392
52	384	388	399	494	520	479	406	395	375	369	351	368
53	366	375	410	500	468	384	367	350	336	325	314	320
54	310	319	371	458	473	436	385	356	351	398	397	387
55	381	357	387	521	487	398	362	343	333	337	380	353
56	324	316	319	401	444	436	398	385	401	389	352	346
57	338	335	340	356	364	372	434	333	354	343	370	357
58	353	340	359	400	370	396	356	317	346	335	327	320
59	277	290	299	411	451	398	345	320	313	306	334	339
60	324	330	335	462	606	431	421	416	333	294	312	299
61	300	302	296	347	398	398	372	358	353	326	299	311
62	306	308	300	350	367	334	293	302	293	289	287	279
63	276	274	269	343	333	316	304	291	284	282	288	299
64	299	292	295	311	358	341	310	303	284	296	277	272
65	270	265	272	296	333	299	284	326	332	389	351	353
66	339	334	364	396	376	363	306	327	316	361	361	393
67	348	358	318	399	422	389	343	323	312	407	472	408
68	359	362	353	434	370	339	348	350	364	347	342	338
69	325	347	354	420	467	456	404	396	362	338	385	355
70	327	333	334	383	459	430	416	388	363	360	359	351
71	324	335	357	433	485	379	338	331	338	354	322	315
72	306	316	338	419	516	449	433	424	416	420	422	374
73	358	389	465	516	583	521	477	443	415	426	396	371

Table 3 - Correlation Matrix Between Boothbay Harbour Temperatures and RIVSUM.

MONTH FOR TEMPERATURE

# of Months Lagged	1	2	3	4	5	6	7	8	9	10	11	12
0	.783	.755	.711	.6	.017	.211	.306	.217	.388	.484	.374	.666
1	.59	.755	.64	.579	.486	.358	.478	.376	.317	.7	.424	.412
2	.409	.577	.653	.494	.345	.414	.507	.46	.343	.599	.655	.418
3	.43	.43	.473	.507	.242	.182	.343	.594	.214	.492	.49	.517
4	.549	.488	.281	.324	.27	.071	.081	.494	.202	.216	.367	.359
5	.447	.607	.37	.149	.107	.128	-.164	.297	.692	.094	-.001	.171
6	.295	.447	.475	.291	-.032	.003	-.097	.045	.589	.73	.169	-.177
7	-.134	.338	.34	.46	.07	.082	-.198	.062	.463	.843	.644	-.265
8	-.234	-.095	.289	.267	.322	.141	-.197	-.03	.474	.786	.809	.633
9	.591	-.201	-.028	.297	.134	.268	-.14	-.033	.397	.781	.82	.794
10	.773	.665	-.135	.014	.241	.331	.004	-.019	.243	.649	.805	.821
11	.799	.795	.647	-.068	.067	.482	.124	.056	.276	.5	.662	.795
12	.778	.797	.726	.667	.052	.367	.401	.151	.396	.475	.455	.659

Table 4 - The lag times (from Fig. 4a and b), their statistical significance of the correlation and the distance from the Saguenay River for temperature stations (see text).

	Station	Years	Distance from RIVSUM (in Km)	Lag Time (in months)	Statistical Significance
Group I	Entry Island	30-72	700	4	.02
	Sambro	36-66	1300	9	.04
	St. Andrews	21-72	1800	8	.001
	Boothbay Harbour	21-72	2100	9	.01
	Boston Harbour	22-66	2325	11	.01
Group II	Entry Island	56-69	700	4	-
	Halifax Line	56-69	1300	5	-
	Lurcher Lightship	56-69	1600	5	-
	Portland Lightship	56-69	2175	6	-
	Boston Lightship	56-69	2325	9	-

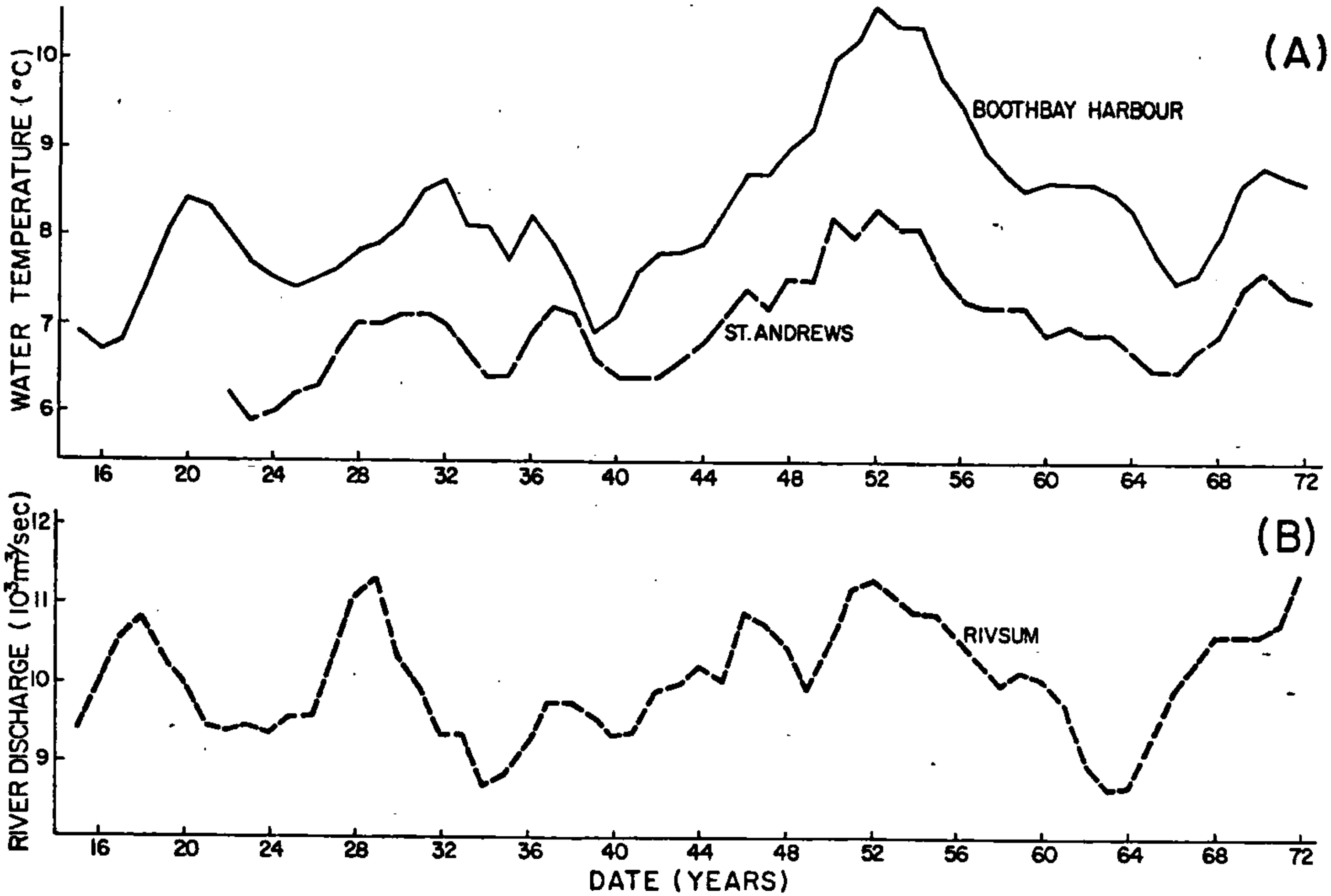

Figure 1 - Three year equally-weighted running means of the annual averages of sea surface temperatures at Boothbay Harbour, Maine and St. Andrews, New Brunswick, and the discharge representative of the St. Lawrence River (RIVSUM).

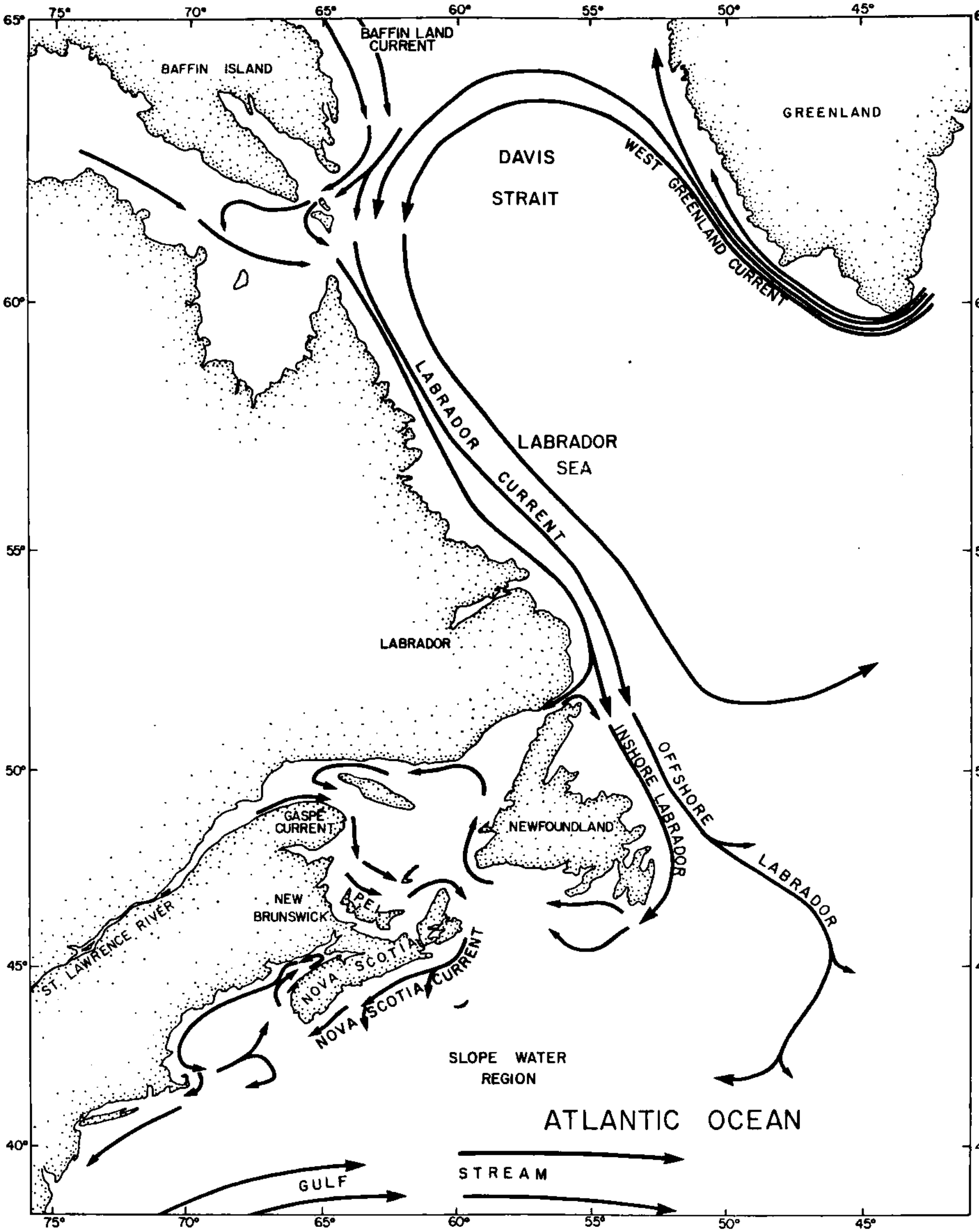

Figure 2 - Map of the Northwestern Atlantic coast showing general circulation patterns.

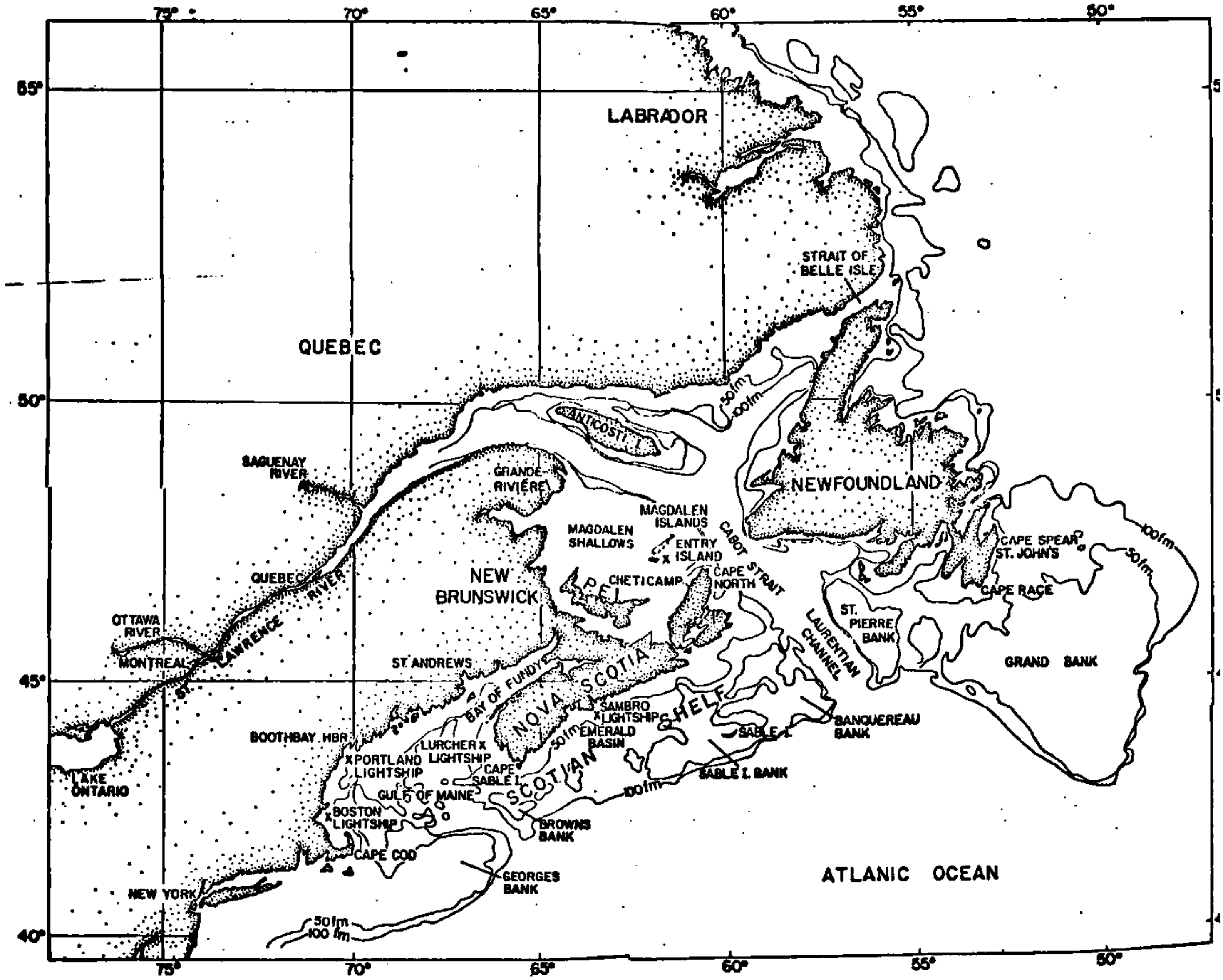

Figure 3 - Map showing location of temperature stations.

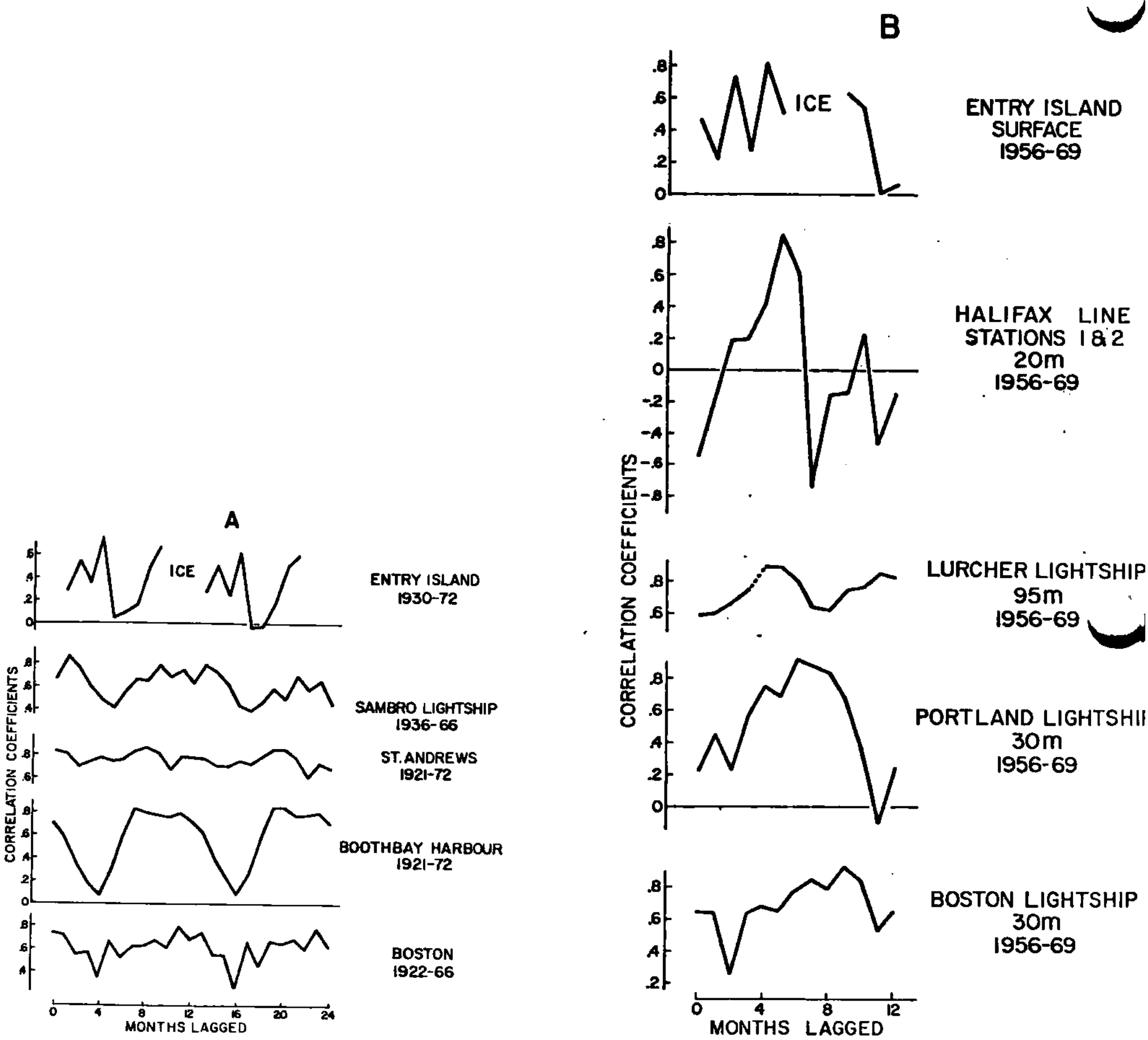

Figure 4 - a. Plots of the correlation coefficients from Group I stations using March RIVSUM, 3 yr running means.

b. Plots of the correlation coefficients for Group II stations using July RIVSUM, 3 yr running means. Halifax Line (broken data), June and July RIVSUM, no running means. Dotted lines indicate scanty data, annual ship refit.

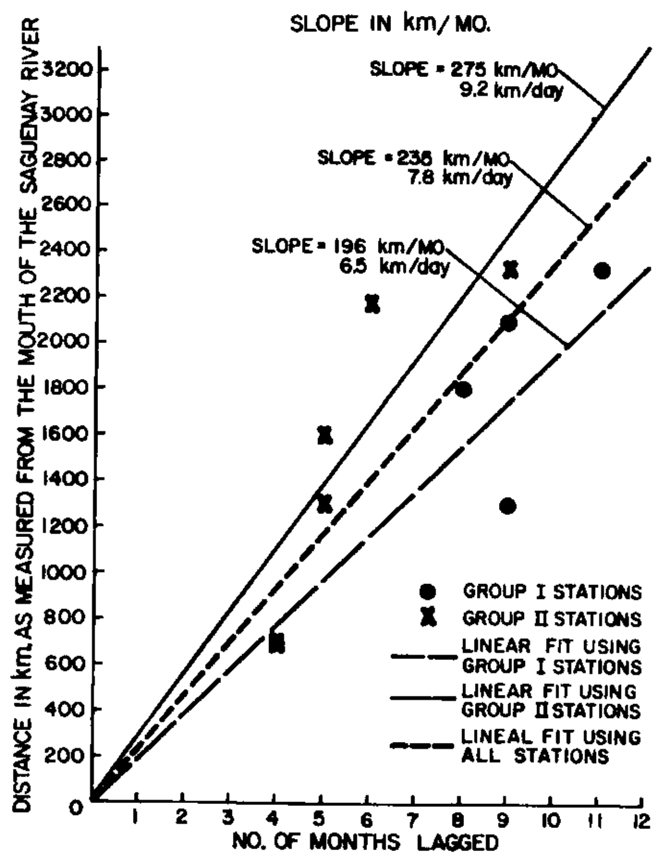

Figure 5 - Plot of distance versus lag time as determined by correlation analysis (see text).

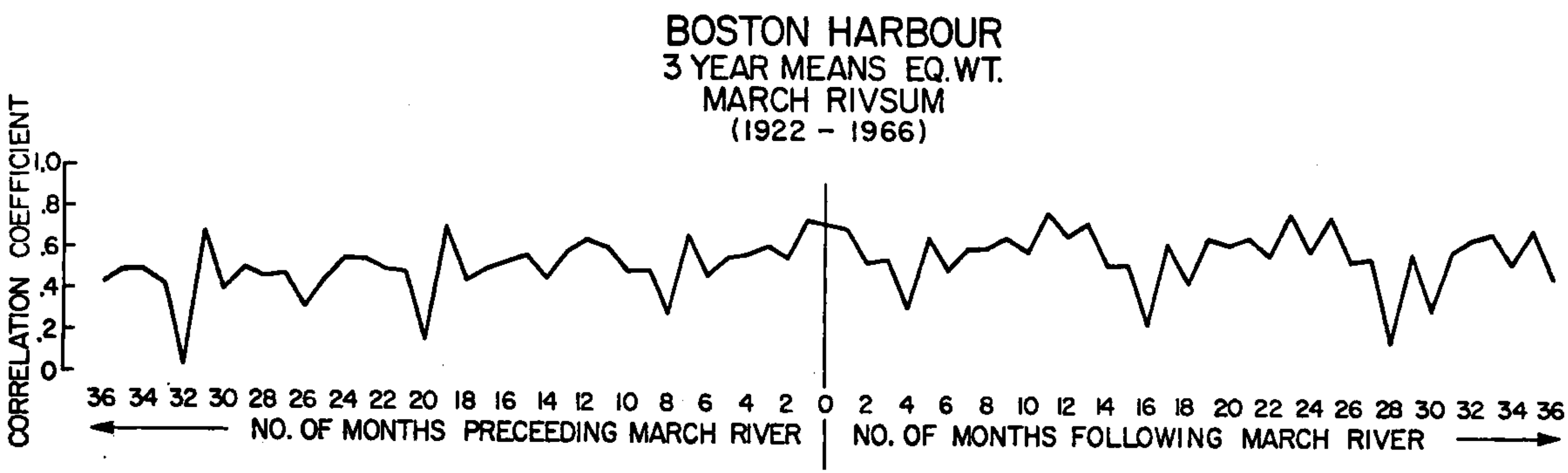

Figure 6 - Plot of Boston Harbour - RIVSUM correlations showing recurring patterns with the sea temperature preceding or lagging the river signal.

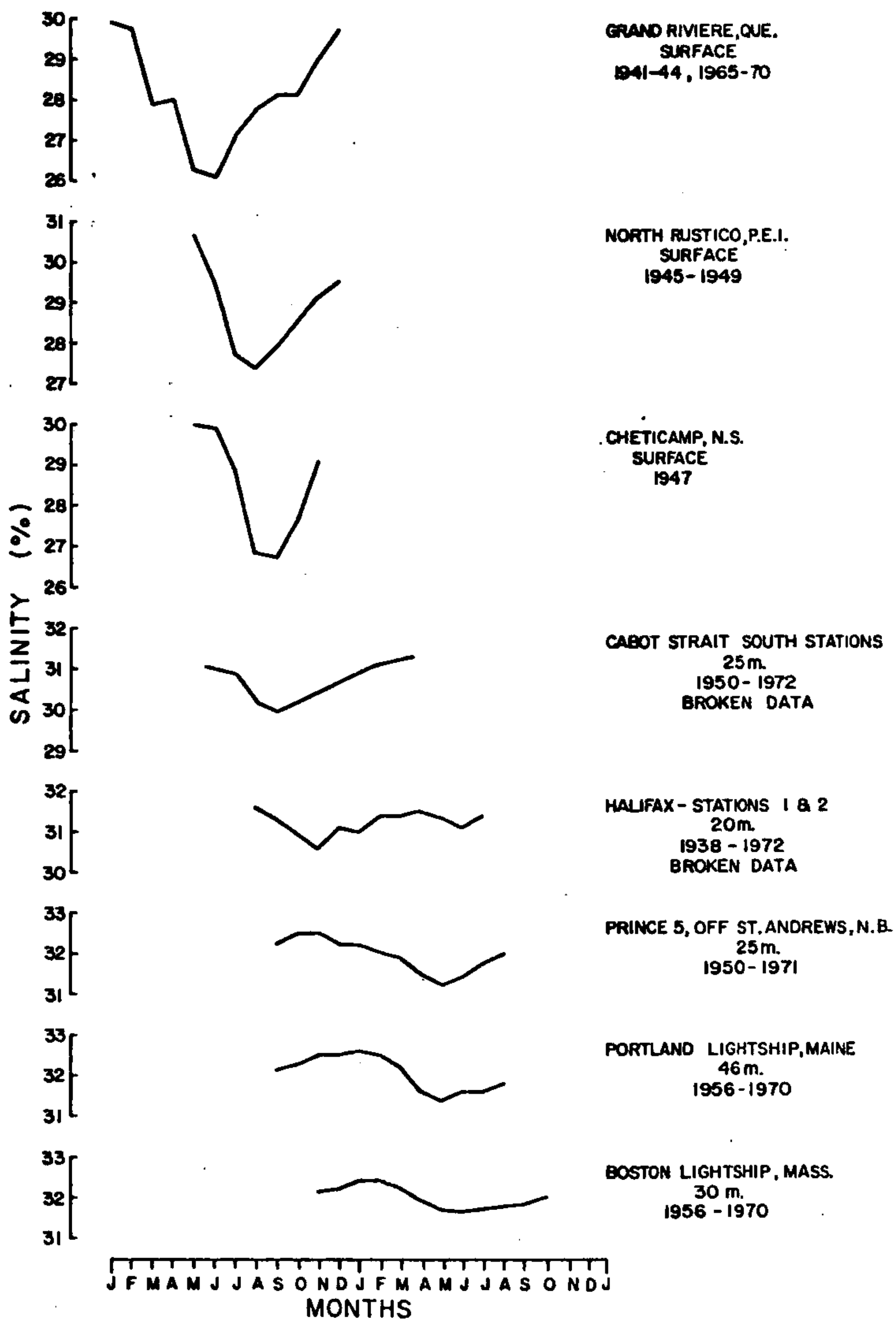

Figure 7 - Seasonal salinity curves at several locations in the Gulf of St. Lawrence, on the Scotian Shelf and in the Gulf of Maine; Cabot and Halifax Lines 2 month means (broken data).

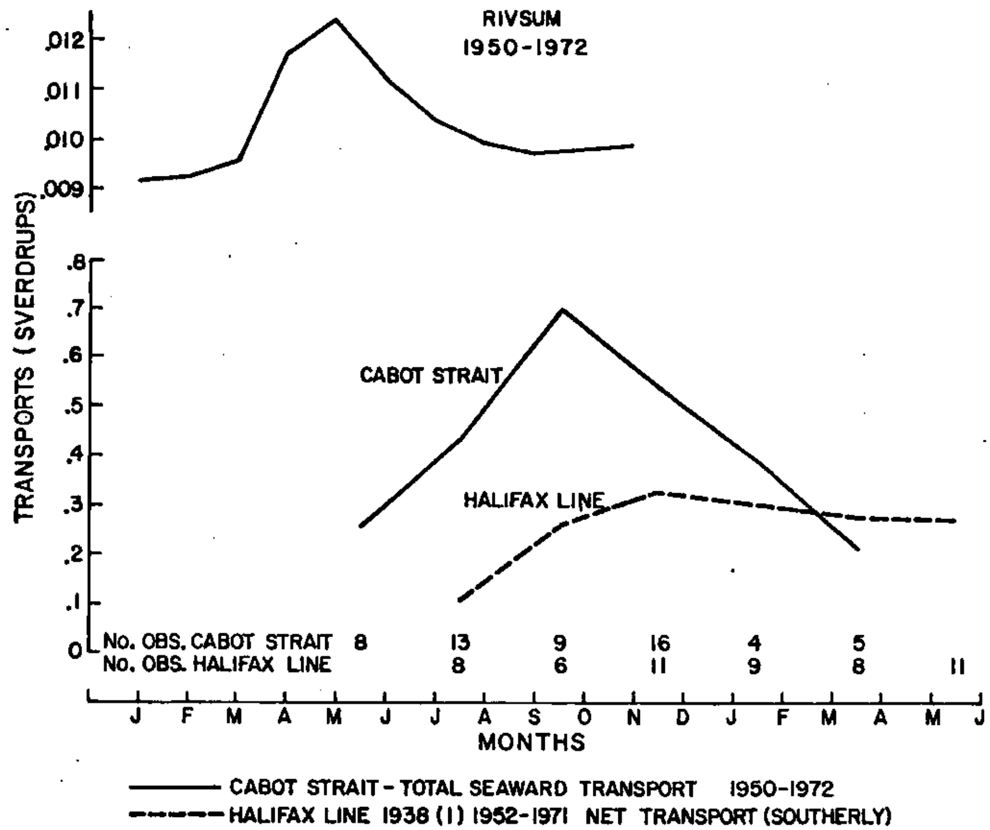

Figure 8 - Seasonal transport curves of RIVSUM, the southerly flow through Cabot Strait and the net flow through the Halifax Line.

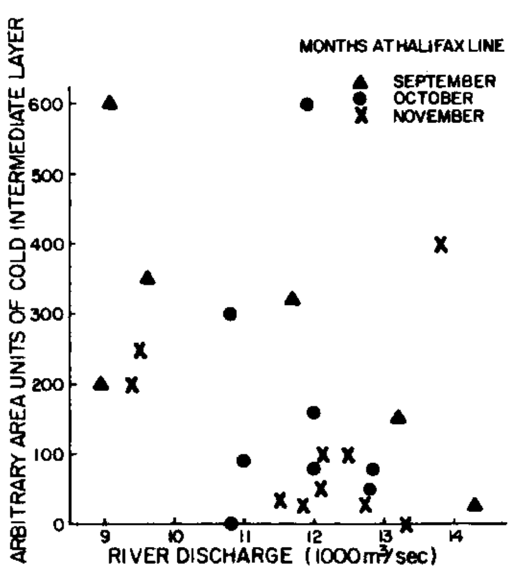

Figure 9 - Plot of the vertical cross sectional area of the 'Cold Intermediate Layer' ($T<4°C$; $S>31.5$ °/oo) determined from the temperature-salinity data along the Halifax Line versus the average discharge from the St. Lawrence River (RIVSUM) for lags of 4, 5 and 6 months.

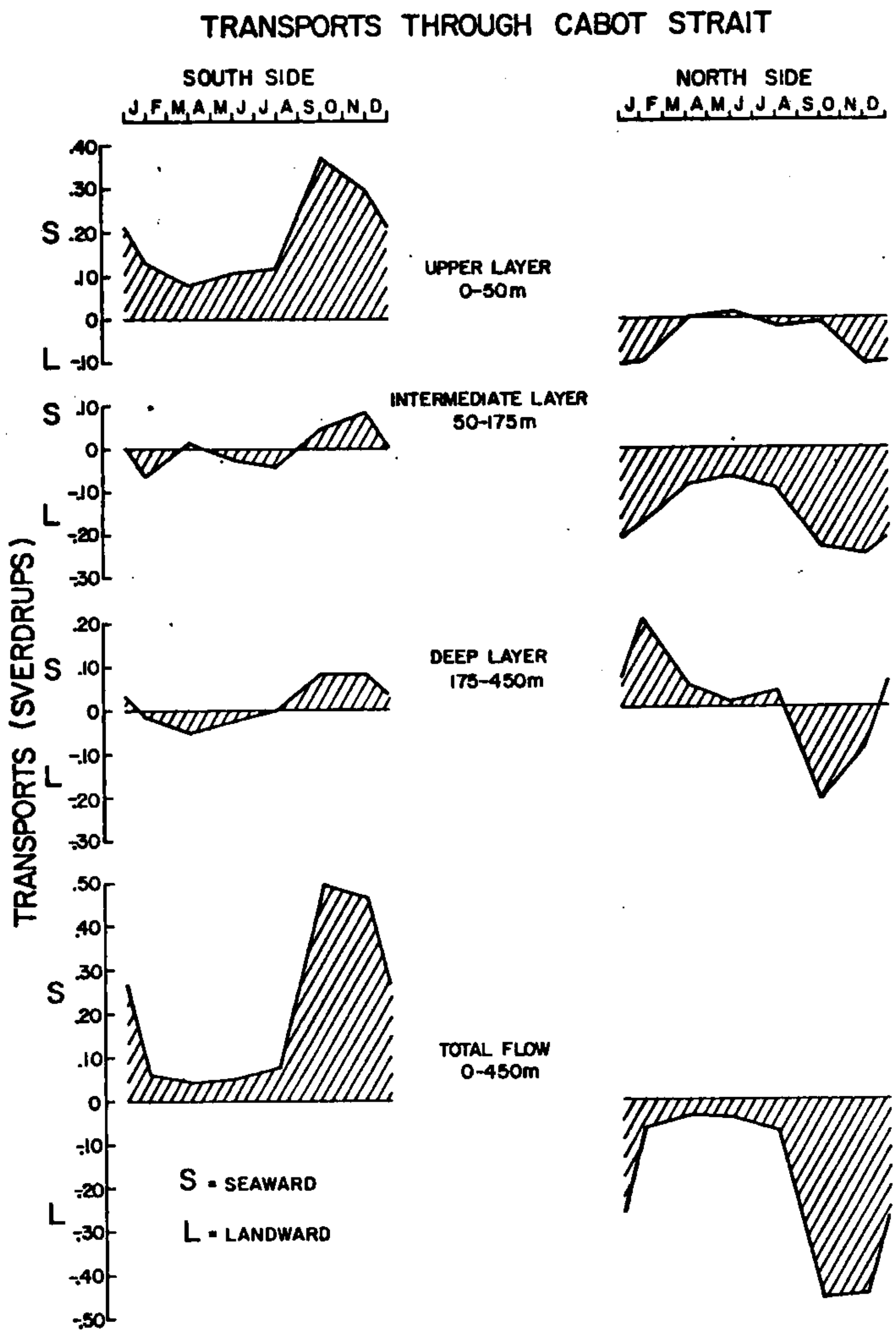

Figure 10 - The seasonal transport through the north and south halves of Cabot Strait with depth.

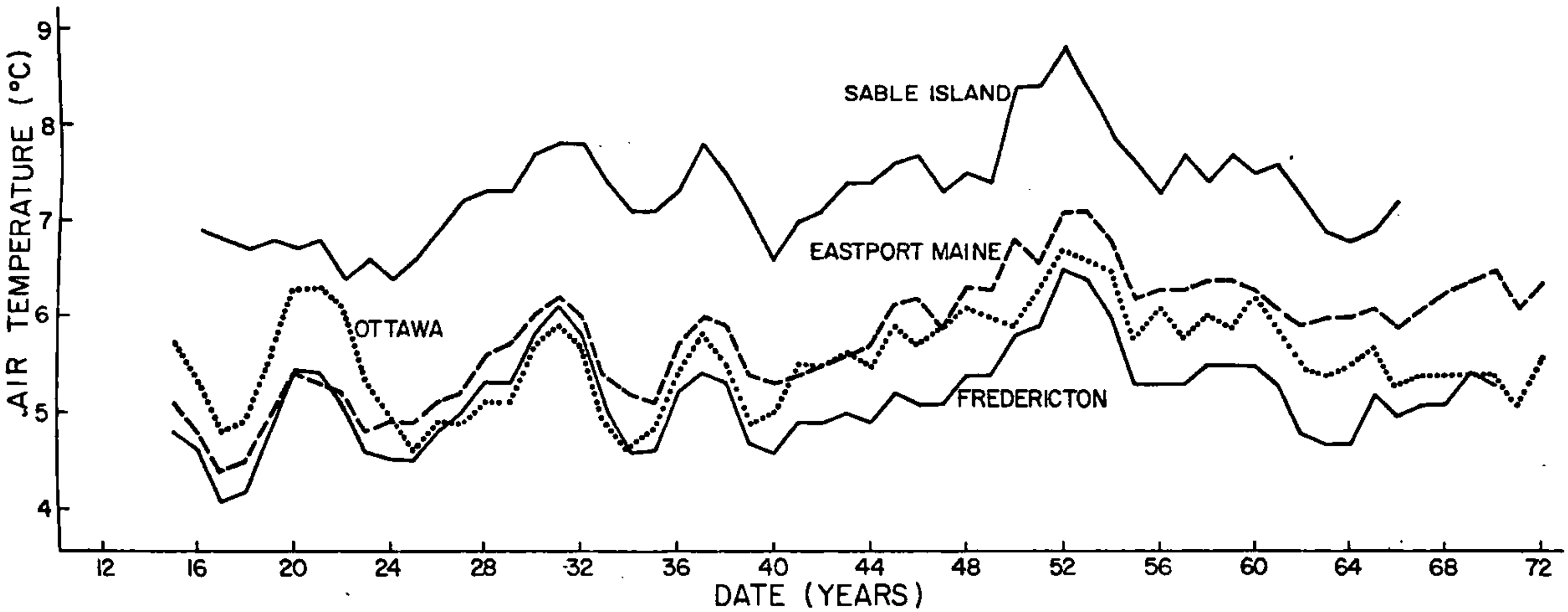

Figure 11 - Plots of the three year equally-weighted means of the annual averages of air temperatures at Eastport, Maine; Fredericton, New Brunswick; Ottawa, Ontario; and Sable Island, Nova Scotia.

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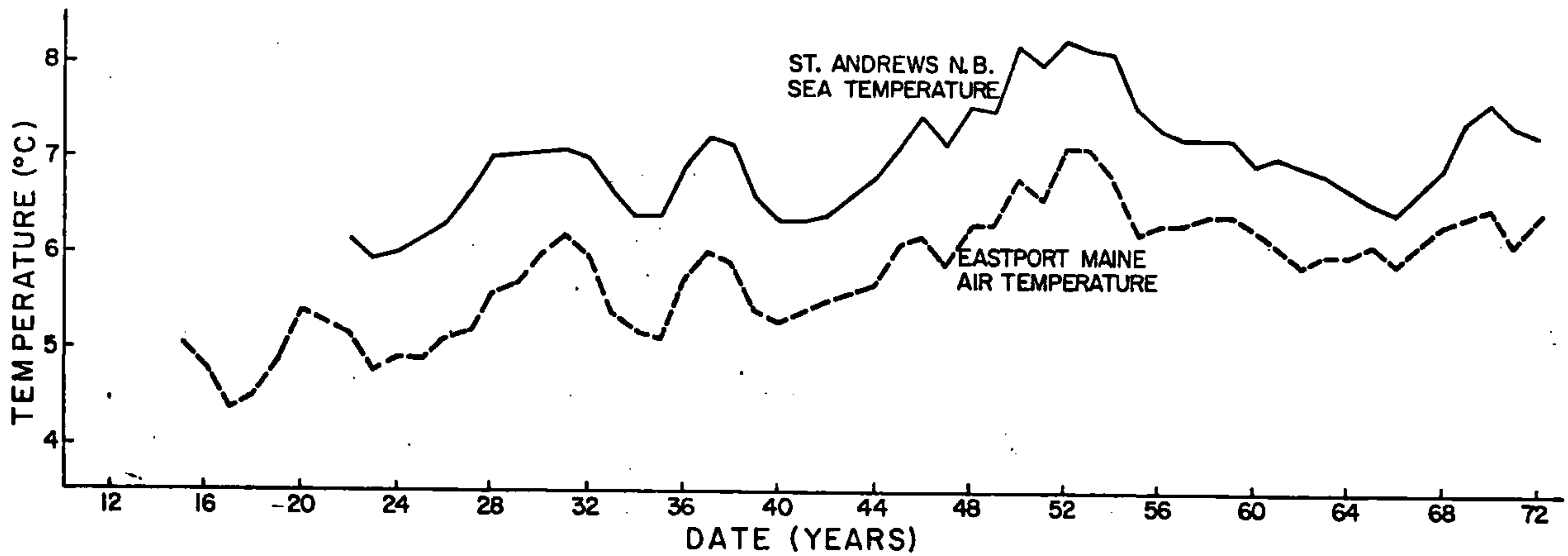

Figure 12 - Plots of the three year equally-weighted means of the annual averages of St. Andrews' sea surface temperatures and Eastport air temperatures.

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APPENDIX

Upwelling Along the Yarmouth Shore of Nova Scotia

by

C.J.R. Garrett[1] and R.H. Loucks

Introduction

The waters off Yarmouth and the southwestern tip of Nova Scotia support lobster, herring and the largest concentration of non-breeding surface feeding sea-birds in the northwest Atlantic (R. Brown, personal communication). Relatively low surface temperatures are observed in summer in this are and small surface-to-bottom temperature differences year-round (Lurcher Lightship temperature records, Lauzier and Hull, 1969). Sea-bed drifters indicate a shoreward bottom current of approximately one nautical mile per day (2 cm sec $^{-1}$) consistently throughout the seasons (Lauzier, 1967).

These biological and physical facts suggest that the waters off Yarmouth may be an area of upwelling. High productivity also occurs off the Louisburg shore of Cape Breton Island, and this is the only other region of the Atlantic coast of Nova Scotia for which sea-bed drifters move onshore. However, we shall restrict our attention to the area off Yarmouth and discuss possible mechanisms for such site-selective, seasonally consistent upwelling.

Possible mechanisms

Upwelling is often thought of as being wind-driven (e.g. Garvine, 1969 and Cushing, 1971). However, it is difficult to see how wind-driven upwelling could be either site-selective or seasonally consistent since the winds along the Atlantic coast of Nova Scotia have large spatial scales and shift in prevailing direction from winter (northwesterly) to summer (southwesterly).

Upwelling or downwelling may also be induced by currents parallel to the coast (Hsueh and O'Brien, 1971). The mechanism is illustrated in Fig. 1 for the case where, facing in the direction of the current, the coastline is to the right (this is the correct sense for the Yarmouth shore, Bumpus and Lauzier, 1965).

The Coriolis force acting on the current is onshore and causes a set-up of the sea surface. This in turn produces an offshore pressure gradient equal and opposite to the Coriolis force. In a boundry layer near the sea floor, the longshore current is reduced by bottom friction, with an equal reduction in the onshore Coriolis force. The offshore pressure gradient, however, is unchanged and so there is a net force offshore. This generates a frictionally-balanced offshore drift in the bottom boundary layer, and induces downwelling at the coast.

For a coastal current in the opposite direction upwelling occurs, and, of course, signs are reversed in the Southern Hemisphere. But off Yarmouth, the sense of the average coastal current is such as to induce downwelling by this mechanism, and fluctuating currents would presumably produce alternate upwelling and downwelling with no net effect.

Centrifugal upwelling

The Yarmouth coastline is convex to the ocean. Thus a current parallel to the coast will experience a centrifugal as well as a Coriolis force. As for the Coriolis force, the centrifugal force must be balanced by a pressure gradient, and this in turn drives an onshore flow in the bottom boundary layer, with consequent upwelling. This "centrifugal upwelling" is just the well-known secondary circulation induced as a river flows round a bend.

For a steady current U flowing in a radius of curvature R the ratio of the centrifugal force U^2/R to the Coriolis force fU is the Rossby number Ro = U/fR, which is generally small (Ro = 10^{-2} for U = 0.1 m sec $^{-1}$, f = 10 $^{-4}$ sec $^{-1}$, R = 10^5 m), so that Coriolis upwelling or downwelling dominates centrifugal upwelling.

However, the centrifugal effect produces upwelling independently of the current direction, so that strong fluctuating currents (such as tidal currents) will produce consistent upwelling.

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Application to the Yarmouth Shore

The average current U into the Gulf of Maine past Yarmouth is about 6 cm sec $^{-1}$ (Bumpus and Lauzier, 1965). This induces Coriolis downwelling with strength determined by the Coriolis force fU. Centrifugal upwelling associated with the average current is negligible, but there is a strong alternating tidal current with maximum strength $U_t \approx 1$ m sec $^{-1}$, and this induces centrifugal upwelling with strength determined by $1/2\ U_t^2/R$, where R is the radius of curvature of the coastline.

The Coriolis downwelling associated with the mean current, and centrifugal upwelling associated with the tidal currents are then in opposition; for values of the parameters as aboce and with R = 50 km the centrifugal upwelling is greater than the Coriolis downwelling by a factor of about 2.

For centrifugal upwelling we now attempt a roung estimate of the shoreward bottom drift velocity V in the bottom boundary layer where the pressure gradient is unbalanced. Assuming a quadratic bottom friction $\gamma\ u\ |u|$ where $\gamma = 0.003$, and a boundary layer thickness h, the force balance in the bottom boundary layer thickness h, the force balance in the bottom boundary layer is roughly between a pressure gradient $1/2\ U_t^2/R$ and a frictional force $\gamma\ U_t V/h$. Hence $V = h\ U_t/(2\gamma R) \approx 2$ cm sec^{-1} for h = 5 m.

This is of the right order of magnitude, but is a very rough calculation dependent on factors such as h, or rather, the profile of the tidal current U_t, which are unknown. Further work is proceeding on a more rigorous theoretical formulation, but there is also a need for moored current meter observations in the area, both to determine tidal current profiles as an input to a more detailed mathematical model, and also to resolve, if possible, the onshore bottom drift.

We have not discussed here the vertical mixing due to the strong tidal currents, and the density driven flows that would result. These clearly require further investigation. For the moment, we are content to find that centrifugal upwelling due to tidal currents can apparently account for many of the observations, and to remark that this upwelling would be reduced by an increase in the coastal current into the Gulf of Maine.

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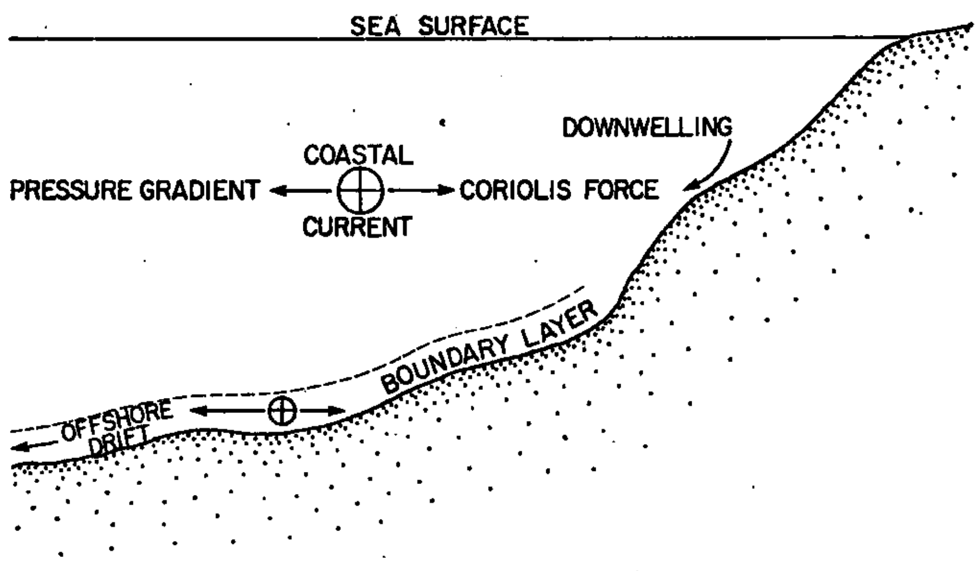

Fig. 1. Downwelling induced by a coastal current.