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An Oceanographic Flip-Flop: Deep Water Exchange in Fortune Bay, Newfoundland

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Observations are presented of deep-water exchange in a large coastal embayment on the south coast of Newfoundland. The bay has several outer sills not unlike those found in fjords. Outside the bay, two submerged channels on the continental shelf are reservoirs for two different deep-water masses which may flow over separate sills to replace deep water within the bay. The deep water is renewed approximately twice a year. On average, exchange occurs both during the half-year period centered on midwinter and during the half-year period centered on late summer, with the water properties of the inflow in each period being dominated by one of the two source water masses in the shelf channels. The system therefore occupies one of two renewal states for much of the year. The transition between states appears to be induced largely by seasonal variations in the wind forcing over the shelf.

1. INTRODUCTION

Ventilation of semienclosed basins by replacement of deep water occurs in coastal environments and in the ocean basins themselves. A common feature of many deep-water renewal problems is the presence of a sill at the basin entrance. Sills have been shown to represent barriers to exchange at small scales in fjords [see *Gade and Edwards*, 1980; *Farmer and Freeland*, 1983] and at ocean basin scales for the outflows from the Mediterranean and Norwegian seas into the North Atlantic, and from the Red Sea and Persian Gulf into the Indian Ocean [see *Warren*, 1981].

Conceptually, the simplest system consists of two basins connected by a channel with a sill. During renewal, one basin serves as a reservoir of deep water for the other. Deep-water replacement is initiated when water which is denser than the deep water in the receiving basin appears above sill depth, and can in principle continue until the density contrast is removed. At this point the system is stable, and exchange will not occur again unless processes act either to increase the density of the deep water in the reservoir or to decrease that in the receiving basin. In reality, of course, both types of process occur, and in most cases of interest, exchange occurs repeatedly over time.

The density of the reservoir water cannot normally be increased indefinitely, however, and so reducing the deep-water density in the receiving basin is fundamentally necessary for repeated exchange in a two-basin system [*Gade and Edwards*, 1980; *Farmer and Freeland*, 1983]. This process of density reduction we call preconditioning. The process is diffusive, using this term in its broadest sense to include turbulent and other fluxes, and with the net effect being a downward transfer of buoyancy. In fjords this is most often the downward diffusion of freshwater or, equivalently, the upward diffusion of salt. In the ocean it is commonly the downward diffusion of heat.

This article is concerned with the timing of exchange in a three-basin system, in which the receiving basin is connected to two reservoir basins instead of one. We show that repeated exchange need no longer be contingent upon a preconditioning type of process. This is because in three-basin systems, exchange can occur from either reservoir, and provided the deep-water properties in the two reservoirs differ appropriately, exchange can in principle occur from either basin at any time. This immediately places the focus on the variability of water properties outside the basin, in this case on the continental shelf, rather than on interior processes associated with preconditioning.

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2. THE STUDY AREA

Our observations are from Fortune Bay, on the south coast of Newfoundland (Figure 1). It is a basin intermediate in scale between typical fjords and ocean basins, with several outer sills, and is one of many large unsheltered coastal embayments on the Newfoundland coast. A prominent feature of the topography in the region is the deep submerged channels on the shelf. These will be shown to represent the two reservoir basins in the system: the Hermitage Channel to the west, and the Saint-Pierre/Avalon Channel system to the south and east.

Compared to typical fjords, Fortune Bay is wide: 22 km at the widest point. It is about 130 km long, giving a length-to-width ratio of about 6, which is lower than that for most fjords [*Pickard and Stanton*, 1980]. There are three outer sills. The Saint-Pierre sill lies between the island of Saint-Pierre and the Burin Peninsula; the Miquelon sill just northeast of the island of Miquelon; and the Sagona sill between Sagona Island and Newfoundland proper. Their limiting depths are nearly the same: 125, 115, and 110 m, respectively. The maximum depth in Fortune Bay is 526 m in Belle Bay, which is separated from the rest of the bay by an inner sill 195 m deep. The maximum depth in the main basin is 420 m, at a point northeast of the central 180-m-deep bank.

3. METHODS

The stations occupied during the course of this study are shown in Figure 1*b*. Fourteen cruises to the area were conducted over the period June 1981 to August 1985 on CSS *Dawson*, CSS *Hudson*, CGS *Shamook*, CGS *Marinus*, and MV *Pandora II*. The conductivity-temperature-depth (CTD) data were collected using a Neil Brown MK III-B probe mounted on a General Oceanics rosette. Bottle sample salinities were determined using a Guideline Autosal. Salinities and density anomalies were computed using the UNESCO 1978 practical salinity formula [*Millero and Poisson*, 1981]. Data are also presented from moored Aanderaa RCM4 current meters.

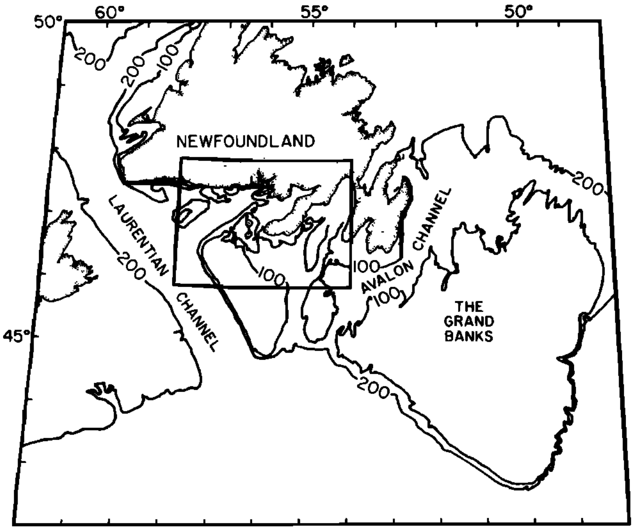

Fig. 1*a*. The Newfoundlana shelf, showing Avalon Channel and Laurentian Channel. The study area is enclosed by the box. Contours are in meters.

4. RESULTS

Deep Water Masses on the Shelf

Very different water masses are present in each shelf channel below the depth of the Fortune Bay sills. Figures 2*a* and *b* are temperature-salinity (TS) diagrams corresponding to station 49 in Hermitage Channel and station 11 in Saint-Pierre Channel, respectively. The solid curves are from December 1981, and the dashed curves from May 1982. In Hermitage Channel the water is warmer and more saline: 1°–6°C; 32.8–34.8 salinity below 150 m depth (Figure 2*a*). In Saint-Pierre/Avalon Channel below the depth of the Saint-Pierre sill (125 m), the water is colder and relatively fresh: −1°–1°C; 32.2–33.8 salinity (Figure 2*b*). This water mass is derived from the inshore branch of the Labrador Current, a major feature of the oceanography of the Newfoundland shelf [*Smith et al.*, 1937; *Petrie and Anderson*, 1983]. The deep water in Hermitage Channel flows into the area from the continental slope region via the Laurentian Channel (Figure 1), mixing at intermediate depths with overlying cold Labrador Current Water and winter mixed layer water. It has been called Modified Slope Water [*McLellan*, 1957]; and this term will be used here.

Comparing the dashed and solid curves in Figures 2*a* and 2*b*, it is seen that the properties of the deep water in each shelf channel are not static. At shallower depths (< 125 m) the effects of surface cooling and mixing in winter are pronounced. This is especially apparent in Figure 2*c*, where the dashed curve corresponds to station 53 in February 1982 and the solid curve to station 49 in December 1981 (the same station 49 data as in Figure 2*a*; station 49 was not sampled in February 1982). The nearly uniform water properties at depths less than 100 m can be attributed to surface cooling and vertical mixing. The low temperature "knee" between 50 and 150 m depth in spring, summer, and fall is therefore due at least in part to the presence of residual winter-cooled water. Labrador Current Water also contributes to this feature [*de Young*, 1983].

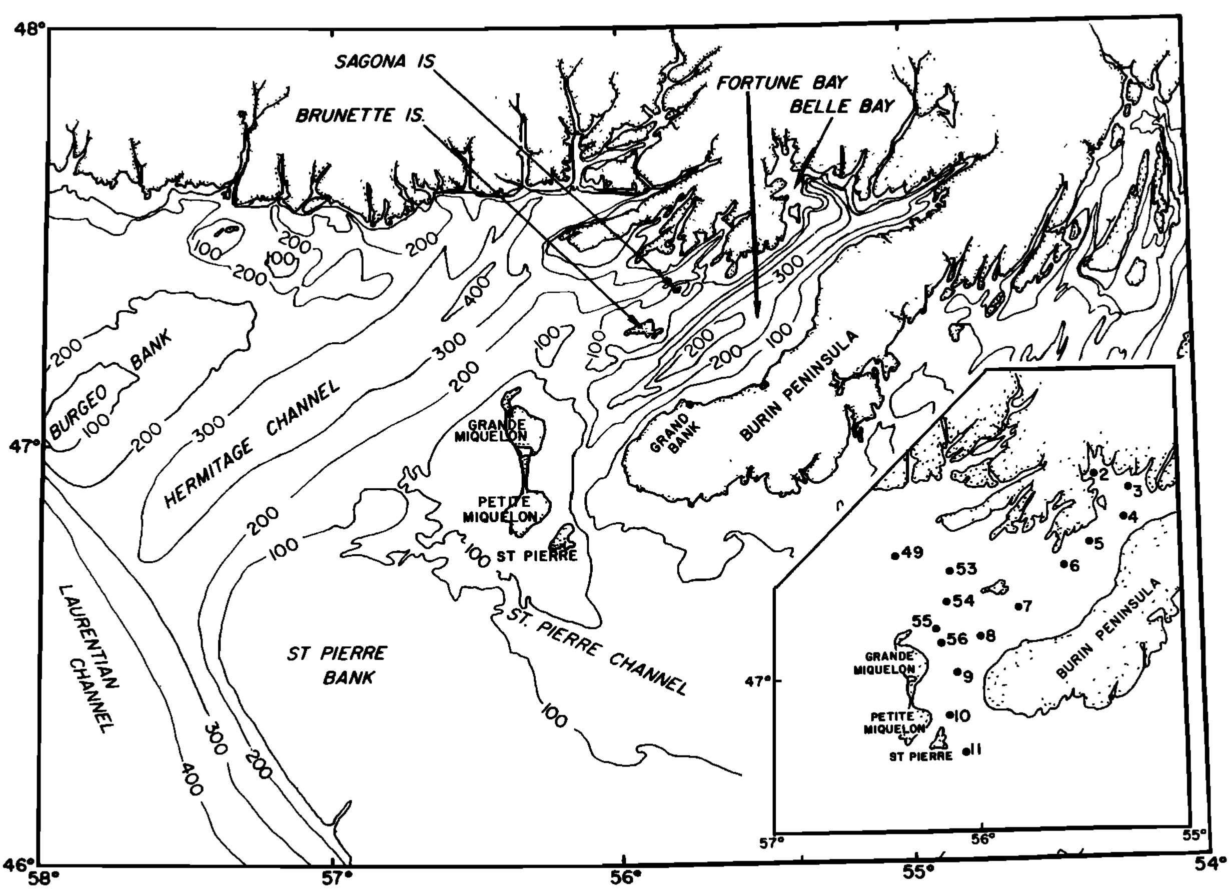

Fig. 1*b*. Fortune Bay and Hermitage Channel study area. Contours are in meters. Station locations are shown in the inset.

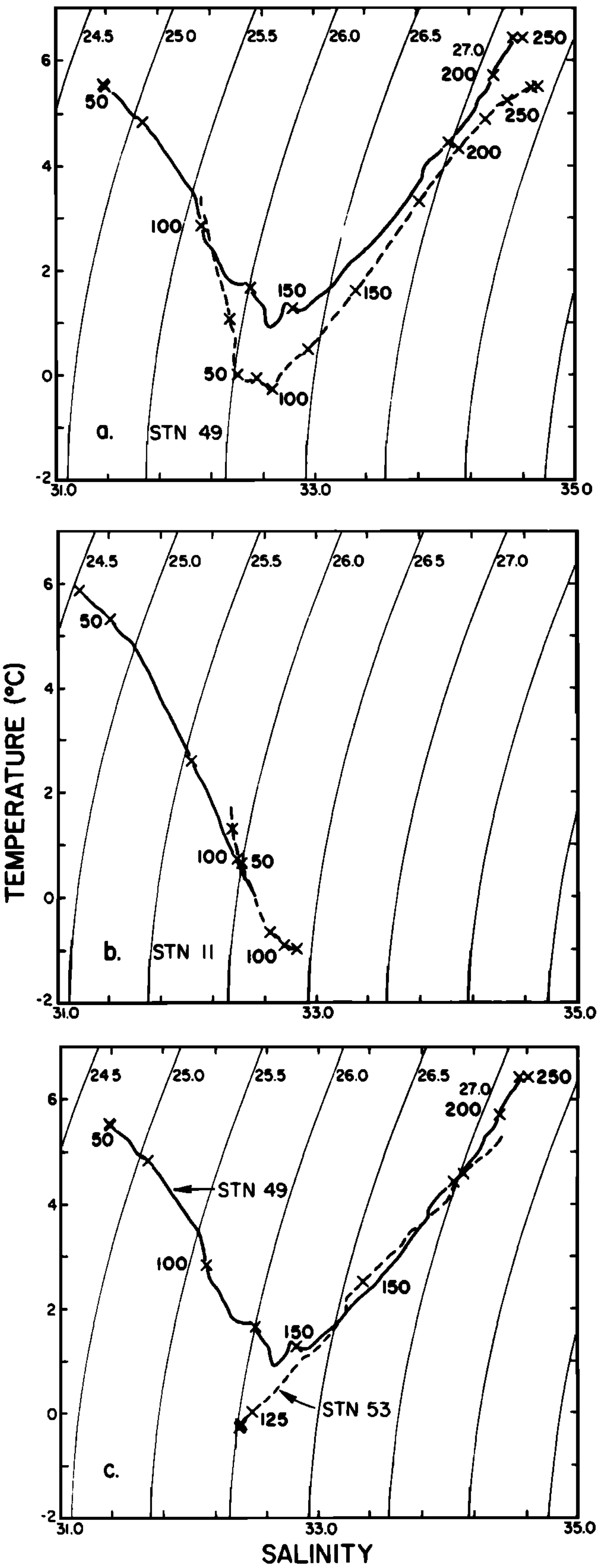

Fig. 2. Temperature-salinity diagrams: (a) in Hermitage Channel at station 49 in December 1981 (solid line) and May 1982 (dashed line); (b) in Saint-Pierre Channel at station 11 in December 1981 (solid line) and May 1982 (dashed line); (c) in Hermitage Channel at station 49 in December 1981 (solid line) and at station 53 in February 1982 (dashed line).

Observations of Exchange

Figure 3 shows temperature and σ_t sections in early winter, midwinter, and spring in 1981/1982 along a single transect from Hermitage Channel across the outer part of Fortune Bay into Saint-Pierre Channel. Note that for the temperatures and salinities observed here, density is primarily a function of salinity. Because density is the fundamental quantity controlling exchange and temperature is a good tracer, salinity sections are not presented.

In December 1981 the temperature at the bottom of the bay was less than 0.5°C (Figure 3*a*). Two months later, in February 1982 (Figures 3*c* and 3*d*), the bottom water temperature had increased to 1.5°C and σ_t by about 0.1. Warm (5.3°C) and very dense (27.0) water was observed near the bottom at station 56 on the Miquelon sill. These data clearly demonstrate that an influx of warm water from Hermitage Channel across the Miquelon sill had taken place and, furthermore, that water with properties typical of those found at depths of 200 m or more in Hermitage Channel had risen to the level of the Miquelon sill.

By May 1982 (Figures 3*e* and 3*f*) the temperature of the bottom water at station 8 had decreased to −0.27°C and σ_t to a value 0.15 less than that observed in February. Water outside the Saint-Pierre sill was very cold (−1°C). The figure indicates that cold water was flowing over the Saint-Pierre sill into Fortune Bay as a dense bottom current.

Sections along the axis of the bay during June and November 1982 are presented in Figure 4. During cold water exchange in June, the inflow over the Saint-Pierre sill is evident (Figures 4*a* and 4*b*), as is the neutrally buoyant character of this inflow in the central part of the bay. Complete replacement of the warm bottom water had not yet occurred: the warmer (>0°C) water at the bottom from station 7 headward is the residue of the previous warm water exchange (Figure 3*c*). By November 1982 (Figures 4*c* and 4*d*) the temperatures in the near-bottom zone were colder toward the head, in contrast with the cold water exchange shown in Figure 4*a*. This indicates that cold water exchange continued after the June/July observations. Furthermore, the presence of warmer (> 0°C) water at depth in the vicinity of stations 7 and 8 (Figure 4*c*) suggests the onset of a second period of warm water renewal.

Semiannual Exchange

The variations in deep-water temperatures within the bay during the 4-year duration of this study are shown in Figure 5. Near-bottom temperature data are presented from three stations in the main basin. Similar patterns are observed at each location, with differences in absolute values due to differences in position and depth, and some differences in detail because all three stations were not always sampled. Making allowance for their coarse temporal resolution, the data are consistent with a semiannual pattern of renewal. Maximum temperatures occurred during the 6- to 7-month period centered about midwinter, that is, December through June. Minimum temperatures were observed during the 6- to 8-month period centered about late summer, that is, May through December.

Better temporal resolution is provided by records from moored instruments deployed in the bay in 1983 and 1984. Figure 6 shows time series of temperature and velocity 5 m above bottom at a location between stations 9 and 10 inside the Saint-Pierre sill for the period late June to early September 1984. The velocity components are oriented such that the

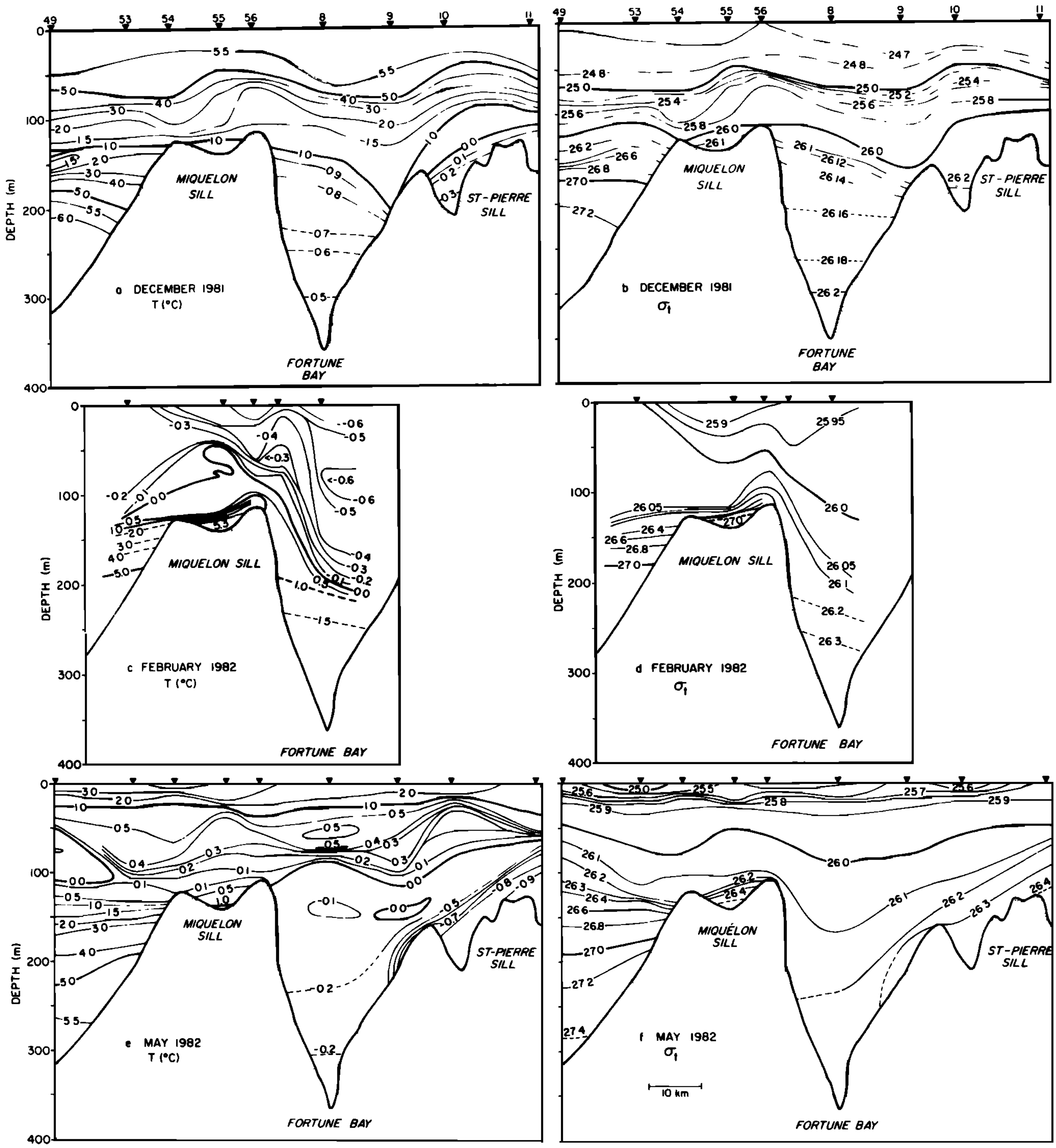

Fig. 3. Temperature and σ_t in the Fortune Bay area in December 1981, February 1982, and May 1982. The station numbers at the top refer to those shown in Figure 2. The plots run from the northwest on the left to the southeast on the right, over the Miquelon sill, across Fortune Bay, and out into Saint-Pierre Channel.

"alongshore" component is directed into the bay along 26° true, and the "onshore" component is in the orthogonal direction along 116° true. Inflow of cold water occurs throughout the record, producing a general trend toward lower temperatures. The inflow is modulated at semidiurnal and lower frequencies, but the velocity averaged over a tidal cycle is directed into the bay essentially for the entire record.

Figure 7 shows two deep-water temperature records at 10 m above bottom near stations 6 and 7 during the 1983–1984 winter from early December to early April. The striking feature in these records is the occurrence of abrupt temperature increases of up to 2°C at 15- to 50-day intervals. The most pronounced of these are present in both records, which either lead or lag each other depending upon the event. Each abrupt increase is followed by a slow decay, but the overall trend is toward higher temperatures. During the decay period after each event there is very little semidiurnal tidal modulation of the temperature signal. Significant modulation is present at

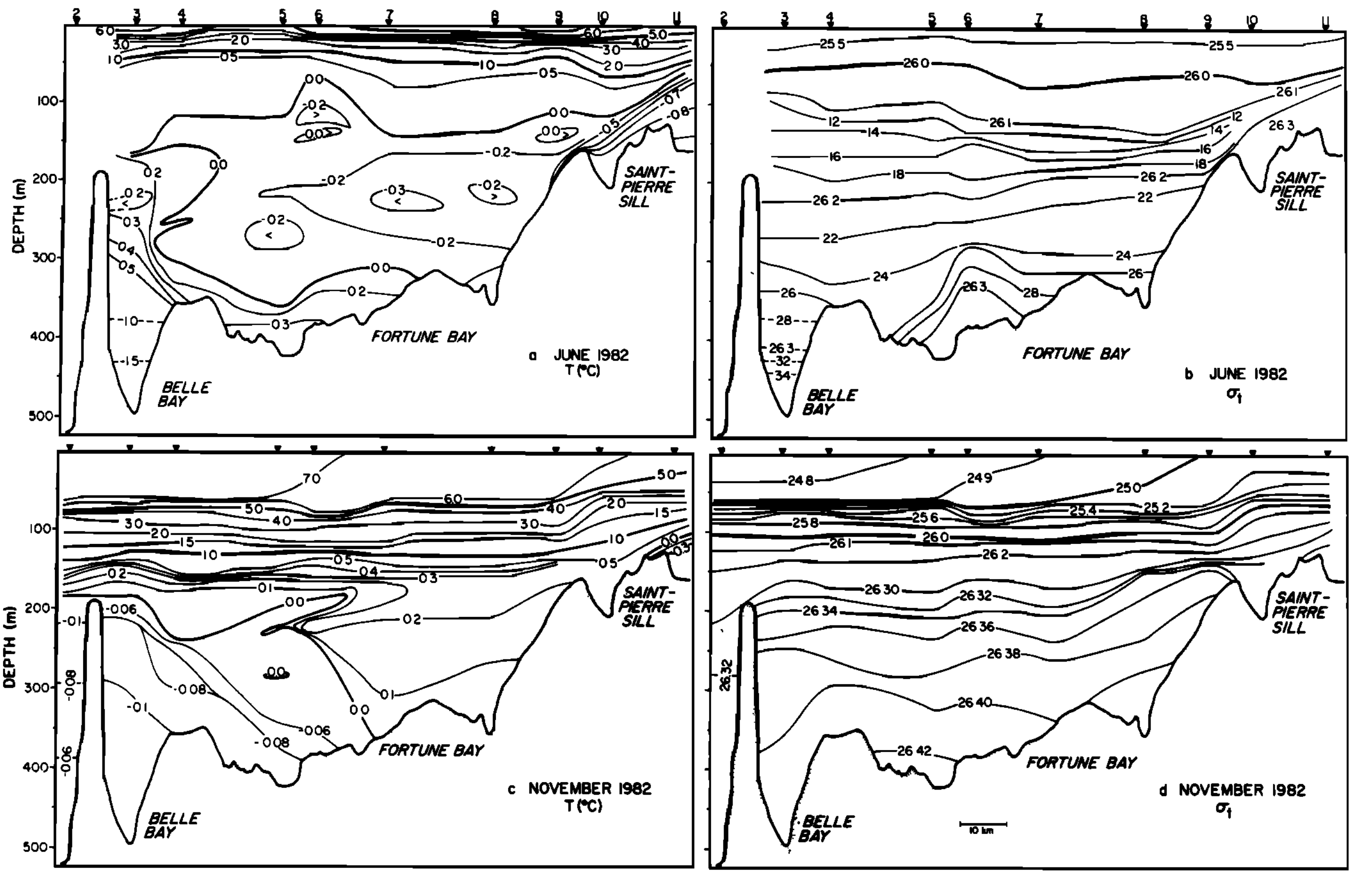

Fig. 4. Temperature and σ_t along the axis of Fortune Bay in June and November 1982.

periods of 1.4 to 2.1 days, particularly in the upper record after the events beginning on Julian day 347, 1983, and Julian day 72, 1984. These periods are in the range expected for internal seiches in Fortune Bay.

Variability in the Wind Field

The stress components calculated from the winds measured at Saint-Pierre are shown in Figure 8 for the period 1982–1984. The wind stress was calculated using the quadratic drag law, $\tau_i = \rho_a C_D U U_i$, where ρ_a is the air density, C_D is the drag coefficient, U is the wind speed at 10 m, and the subscript represents the directional component. The drag coefficient was chosen to be 1.5×10^{-3} [*Large and Pond*, 1981]. The alongshore stress is parallel to the Hermitage Channel axis and positive in the downchannel direction along 215° true. The onshore stress is in the cross-channel direction and is positive along 305° true.

During the summer months, stresses are generally low and directed toward the northeast. Stresses in winter are high, and the cross-channel component is generally negative. The axial stress is also high in winter and in these data may be either positive or negative. These data therefore indicate high stresses in winter due to winds from the north quadrant (northwest to northeast), directions favorable to upwelling at the head of Hermitage Channel and at times along its southeastern margin.

These results are generally consistent with other wind data for the region. The climatological means for Grand Bank, on the southeastern shore of Fortune Bay (Figure 1*b*), show a strong seasonal shift in wind direction and speed [*Atmospheric Environment Service*, 1975]. Summer winds (July-August-September) are from the southwest with a mean speed of 4.5 m/s. Winter winds (December-January-February) are from the northwest and northeast with a mean speed of 7.6 m/s. De

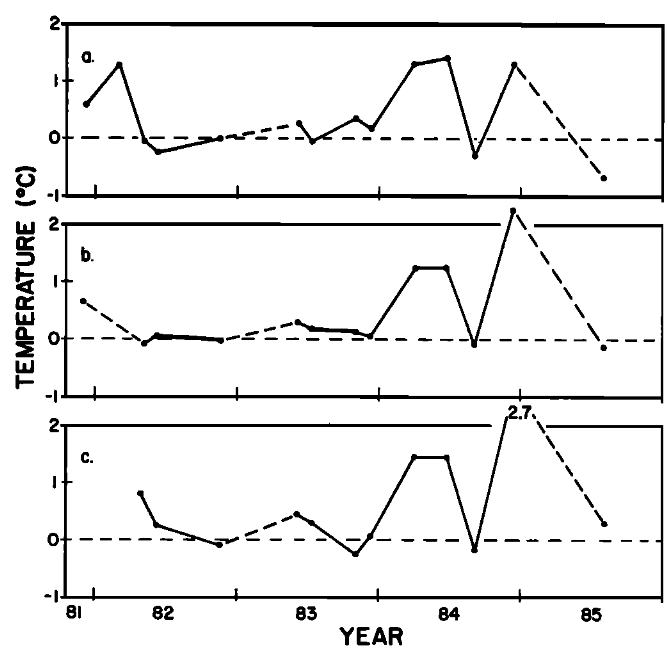

Fig 5. Deep-water temperatures at three locations in Fortune Bay during 1981–1985; (*a*) station 8, 270 m, (*b*) station 7, 320 m, and (*c*) station 6 (420 m).

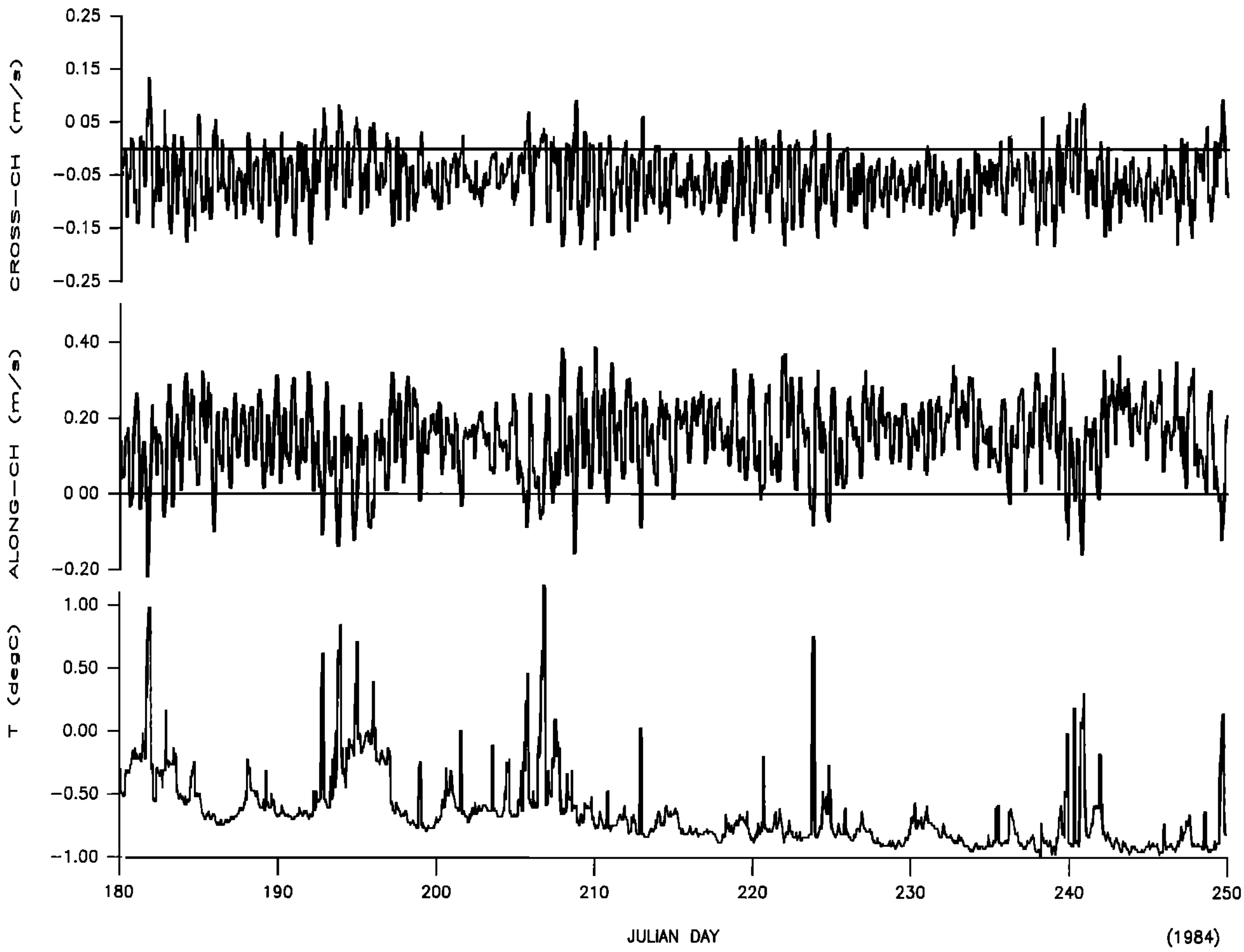

Fig. 6. Time series of temperature and velocity in the cold water inflow at 154 m depth (5 m above bottom) between stations 9 and 10 during the period Julian day 180 to 250, 1984. The alongshore current is directed into the bay along 26° true; the onshore component is positive along 116° true.

Young [1983] analyzed wind stress derived from 3 years of sea level atmospheric pressure data from the region. Maximum stresses were again found to occur in winter and to be directed on average from the north and northeast, the directions favoring upwelling at the head and along the southeastern margin of Hermitage Channel.

5. Discussion

The Renewal Cycle and Wind Forcing Over the Shelf

The data presented in section 4 suggest that the wind-forced baroclinic response on the shelf and in the shelf channels in particular is a plausible mechanism for warm water exchange in winter. Upwelling at the head of Hermitage Channel and along its southeastern margin should occur in winter in response to the surface stresses associated with strong winds generally from the north quadrant. Furthermore, northeasterly winter winds would tend to drive a downwelling response outside the Saint-Pierre sill in Saint-Pierre Channel, reducing the likelihood of cold water inflow. Finally, with the summertime shift to southwesterly winds, the opposite should occur. Figure 9 illustrates the process. During cold water exchange (Figure 9*a*), weak but persistent southwesterly winds produce a stress at the sea surface which is upwelling favorable in Saint-Pierre Channel but downwelling favorable in Hermitage Channel, and cold water flows into the bay over the Saint-Pierre sill. During warm water exchange, on the other hand (Figure 9*b*), northwesterly to northeasterly winds drive upwelling in Hermitage Channel and at times downwelling in Saint-Pierre Channel, and the opposite occurs. It is in essence a push-pull system driven by seasonal variations in the wind stress: hence the use of "flip-flop" in the title.

A test of this model at short time scales is provided by the deep-water temperature records for the 1983–1984 winter in Figure 7. The Saint-Pierre wind stress data for the same period are plotted in Figure 10. Those events in Figure 7 which are most pronounced and present in both records occurred in 1983 on Julian days 347 and 362 and in 1984 on days 28 and 74. On examining Figure 10 it is seen that strong cross-channel stresses of about −0.5 Pa occurred on days 345 and 354 in 1983 and on days 26, 70, and 88 in 1984. Therefore high-speed winds which were upwelling favorable at the head of Hermitage Channel occurred 2–3 days prior to the events on days 347, 28, and 74. This phase lag is consistent with that expected from baroclinic adjustment to an applied coastal upwelling wind stress. It should also be noted in this respect that the observations in Figures 3*c* and 3*d* were made following a 7-day storm during which the wind, as observed from *Shamook* in harbor a few kilometers to the west of Grand Bank, on the Burin Peninsula, persisted at 20–25 m/s from the west. This storm can be seen in the wind stress record from Saint-Pierre (Figure 8) as the persistent event beginning on Julian

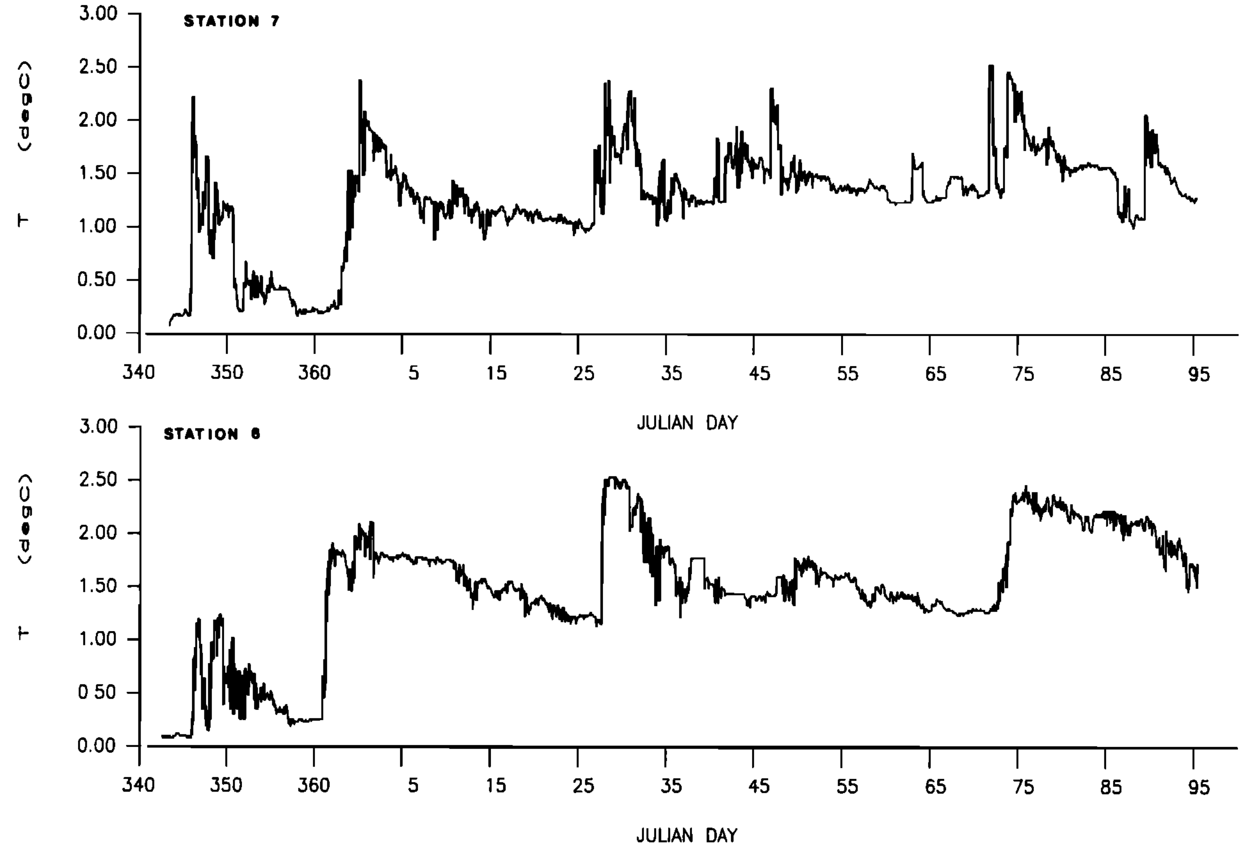

Fig. 7. Temperature time series during warm water exchange at 346 m depth near station 7 (top panel) and at 411 m near station 6 (bottom panel) from Julian day 343, 1983, to Julian day 98, 1984.

day 54, 1982. There was a strong negative cross-channel component associated with the storm, which is consistent with upwelling at the head of Hermitage Channel.

It is clear from Figure 3 that the 5°C water in the depression on Miquelon sill in Figure 3c must have come from depths of 200 m or more, requiring a vertical displacement of about 100 m. The stratification in Hermitage Channel in winter can be represented by a two-layer system: an upper mixed layer about 125 m thick and a lower layer about 200 m thick with an excess density of about 1 kg/m^3 (see Figures 2c and 3). The internal Rossby radius a' is then about 9 km, and since the channel itself is about 40 km across between the 100-m isobaths, the responses on opposite sides are essentially isolated from each other at short time scales. A northeasterly wind will therefore generate the classical upwelling response on the southeastern side and a northwesterly wind upwelling at the head. According to the usual two-layer coastal upwelling model, the displacement of the interface will be given by [*Gill*, 1982, p. 404]

$$\eta = th'\tau/[\rho f a'(h + h')]$$

where h and h' are the upper and lower layer thicknesses, respectively, f is the Coriolis parameter, and t is the time. Substituting the above values for a', h, and h', setting $f = 10^{-4}$ s^{-1} and the wind stress to 0.5 Pa (Figure 10) gives $t = 2 \times 10^5$ s, or about 2 days for a vertical displacement η of 100 m.

The events in Figure 7 on day 347, 1983, and days 28 and 74, 1984 (and that in February 1982), are therefore reasonably well explained by the model. However, events do not appear in the temperature records on days 357 and 90 which one might have expected on the basis of the wind stresses 2–3 days earlier. Furthermore, the event on day 362 occurred 2 days after a period of strong northeastward stress which should have generated a downwelling response in Hermitage Channel. These departures from the simple model above are attributed to factors not considered here, such as the history of the wind forcing, the Saint-Pierre winds sometimes not being representative of conditions over Hermitage Channel, and the complicated topography of the area.

Preconditioning and the Renewal Cycle

It was stated in the introduction that preconditioning, the diffusive decrease in deep-water density within the receiving basin, may exert less control over the frequency of renewal in a three-basin system than in a two-basin system. This is illustrated for Fortune Bay in Figure 11, a TS diagram of the water masses involved in the exchange. Consider cold water exchange first. The temperature and salinity of the source water outside the Saint-Pierre sill, and of the water at middepth in the bay, are shown for June 1982. The resultant water is that observed in July 1982. Assuming that the middepth Bay Water is representative of the time- and depth-averaged properties of the fluid entrained into the inflowing density current, Figure 11 indicates that about 55% of the resultant water was entrained during inflow. This accords roughly with accepted entrainment rates for dense bottom currents [*de Young and Hay*, 1987]. Similarly, the data for the wintertime warm water exchange in February 1982 show that 65% of the resultant water contains entrained middepth Bay Water.

Because of mixing during renewal the difference in density

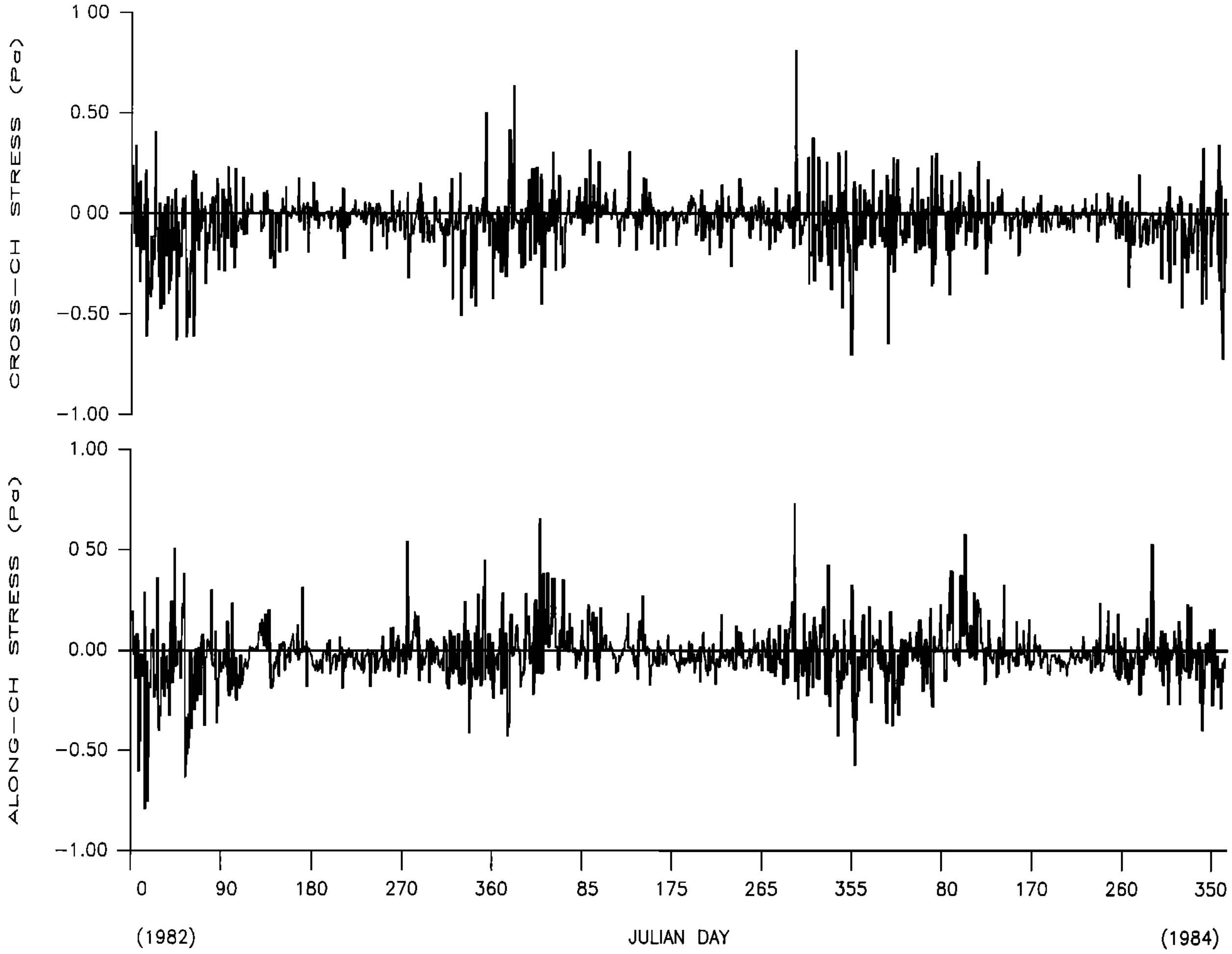

Fig. 8. Time series of wind stress components computed from winds measured at Saint-Pierre during the period 1982–1984. Onshore and longshore directions are defined in the text.

between the two resultant water masses is only about $0.1\sigma_t$, much less than that between the two source water types ($0.6\sigma_t$). Furthermore, the density of the resultant bottom water after cold water renewal must always be less than the density of the warm source water in Hermitage Channel. This implies that after cold water renewal, no preconditioning of the deep water is necessary before warm water renewal can recur. However, the resultant bottom water density after warm water renewal in 1982 was comparable to and clearly may be greater than the density of the cold source water in Saint-Pierre Channel. This means that cold water replacement must normally take place at intermediate depths first as in Figure 4*a*, especially because the density of the inflowing cold water is further reduced by entrainment of overlying fluid. After sufficient time the underlying residual warm water is eroded, and the cold water inflow can penetrate to the bottom.

A time scale for total replacement by cold water can be estimated roughly from the transport of cold ($< -0.3°C$) water over the Saint-Pierre sill into Fortune Bay: approximately 10^4 m^3/s in June 1982 [*de Young and Hay*, 1987]. This transport would require over 300 days to completely replace all of the water in the bay below 150 m, a volume of about 3.4×10^{11} m^3. It is therefore unlikely that total replacement, by which we mean that all the water below sill depth would then have properties identical to those of the source water in the reservoir basin, can occur in a single year. This is consistent with the the fact that near-bottom temperatures approaching $-1°C$ were never observed in the course of this study. However, the 300-day time scale for total replacement is quite consistent with the observed partial replacements over several months and cyclic renewal of the bottom water at roughly 6-month intervals.

It is seen that prior to cold water renewal at the bottom, there may be a period of downward diffusion of buoyancy due to density current inflow to intermediate depths. This process is not really preconditioning, however, since the bottom water is renewed the moment it is "preconditioned." Deep-water erosion is a more appropriate term. Downward buoyancy transfer during erosion is accelerated because the density current both penetrates to intermediate depths and represents a source of kinetic energy for mixing at these depths. We note that a similar effect may occur in two-basin systems because of fluctuations in the deep-water properties in the reservoir basin. *Gade and Edwards* [1980, pp. 466 and 470], for example, refer to this as a special vertical diffusion process in fjords associated mainly with large-volume intrusions of water less dense than the bottom water. *Farmer and Freeland* [1983, p. 179] also mention vertical mixing by gravity currents in fjords. Fortune Bay provides in our view a case in which the erosion process is a dominant mechanism. Finally, we note that downward buoyancy transfer during cold water exchange is aided by the fact that the densities of the two source water types are not very different. Indeed, if the source water densities were the same, there would be no need for an erosion phase at all.

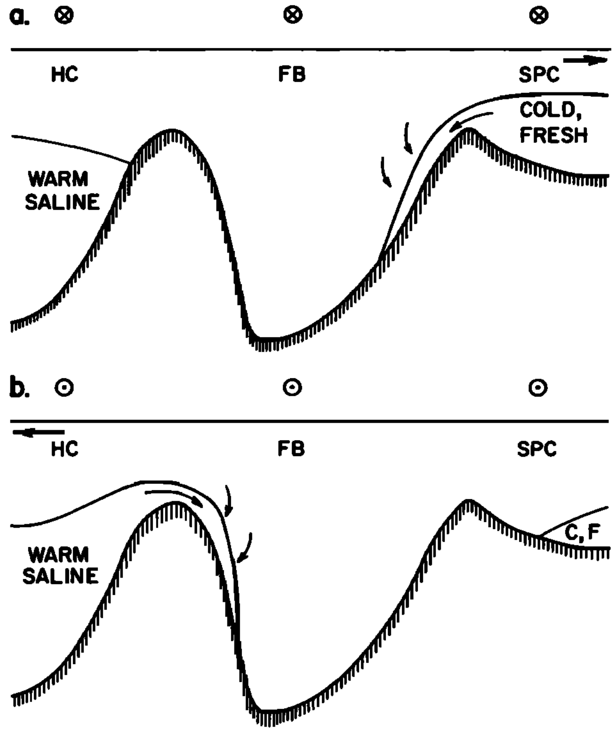

Fig. 9. Conceptual model of the exchange cycle in Fortune Bay (FB): (*a*) cold water exchange from Saint-Pierre Channel (SPC); (*b*) warm water exchange from Hermitage Channel (HC).

Other Factors

The conceptual model of the exchange process sketched in Figure 9 is an oversimplification. Other factors, notably low-frequency variations in the strength of and water properties in the inshore branch of the Labrador Current must contribute. The results presented here are not sufficient to test the existence of seasonal or annual Labrador Current transport cycles. Instead they add to the catalog of transport variability [*Smith et al.*, 1937; *Petrie and Anderson*, 1983], including a single observation of the inshore branch penetrating the western end of Saint-Pierre Channel into Hermitage Channel [*de Young*, 1983].

There is evidence in the historical data base for seasonal variations in the near-bottom water properties in the Saint-Pierre and Avalon channels. *Petrie and Anderson* [1983] have shown that a seasonal salinity cycle exists at 50 and 100 m depth on the shelf near Saint-Pierre. Maximum salinities were found to occur from July through September, and minimum salinities in March-April-May. Table 1 shows the average temperatures and salinities at 100 m depth in Saint-Pierre Channel at station 11 computed from our data for May-June and November-December over the 4-year period from 1981 to 1985. These are the months in which most of our cruises took place. Temperatures are lower on average in early summer than in early winter. The mean salinities are the same in both periods. The important conclusion to be drawn from these and the historical data is that maximum density (minimum temperature, maximum salinity) water should appear at the

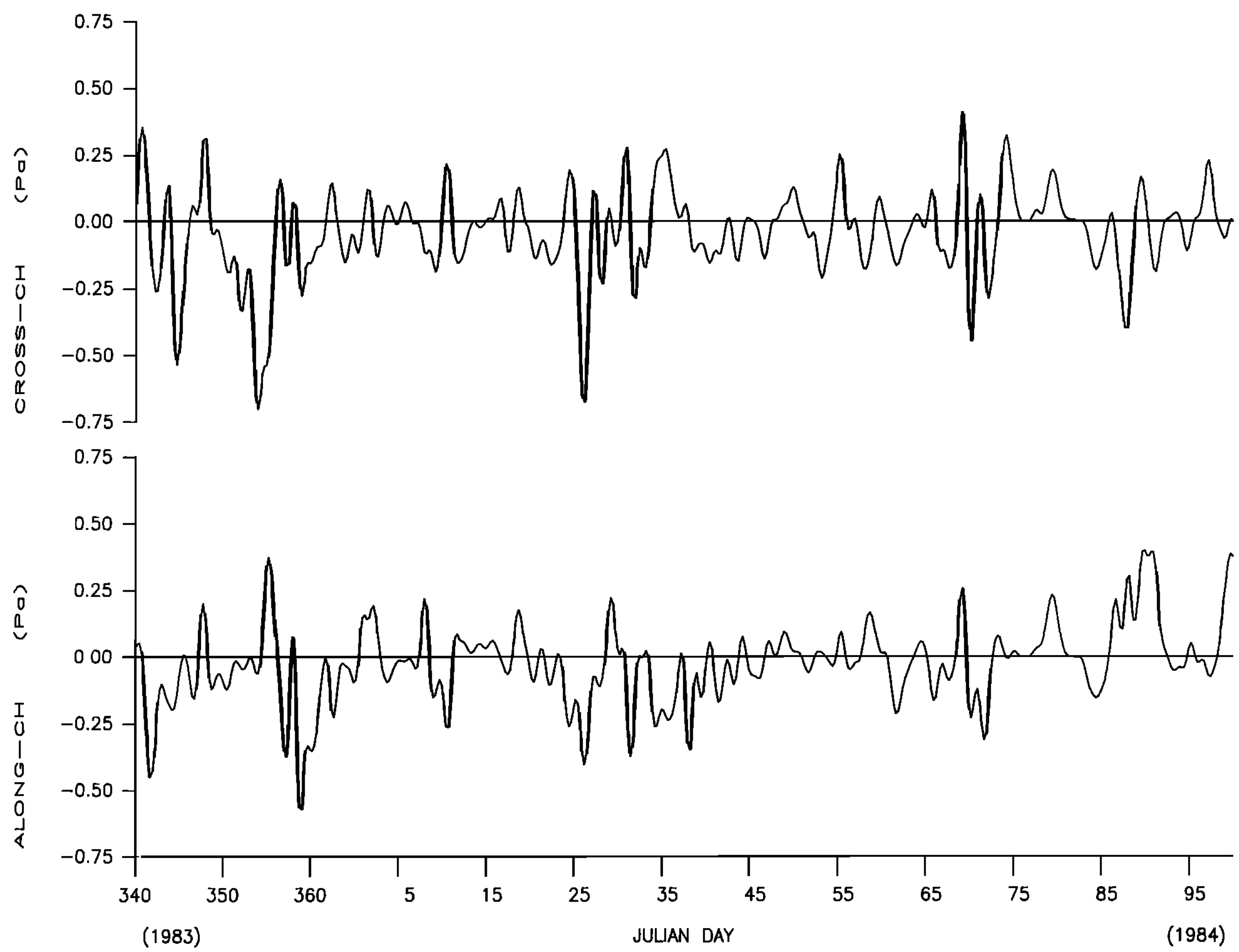

Fig. 10. Wind stress components computed from the Saint-Pierre winds for the period Julian day 340, 1983, to Julian day 100, 1984, spanning the period of the temperature records in Figure 7.

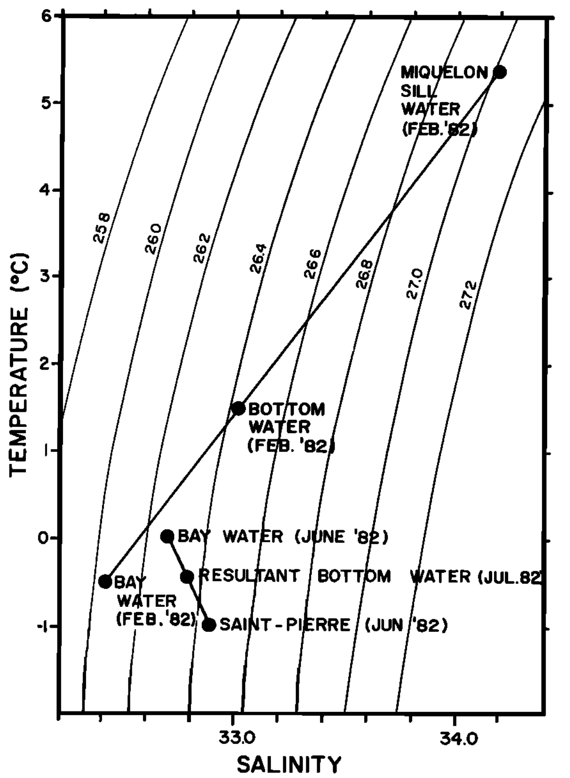

Fig. 11. Temperature-salinity diagram showing mixing lines and relationships between sources, middepth water in Fortune Bay, and the resultant bottom water during both cold and warm water exchange in 1981–1982.

Saint-Pierre sill from late spring through fall. This is when we observe cold water inflow into the bay, and the pattern is consistent with our wind-driven model (Figure 9).

The mean water properties in Hermitage Channel from our data are also given in Table 1 for 100 m and 200 m depth at station 49, again for the periods May–June and November–December. The only clear seasonal difference is in the temperature at 100 m: a roughly 2°C increase from early summer to early winter, with a large standard deviation. This is consistent with wind-driven upwelling in Hermitage Channel in winter. It should be noted that studies of deep-water properties in the Laurentian Channel and in the Gulf of St. Lawrence [*Trites*, 1972] (see also *Dickie and Trites* [1983]) found no seasonal signal. This is consistent with our 200-m results, which exhibit no significant seasonal differences.

Although the focus of this paper has been on driving mechanisms external to the bay itself and the role of density currents in vertical diffusion, there are nevertheless several additional dynamical processes acting within the bay as part of the exchange process which we mention here for completeness. The decay of the wintertime high-temperature pulses implies active mixing, possibly through the internal wave field including the apparent internal seiche. Furthermore, since the incoming warm water necessarily lies beneath colder water (see Figure 3*c*), the interesting possibility exists that double-diffusive instabilities of the layered type [*Turner*, 1973] may develop after warm water inflow. This could also happen during cold water inflow when cold fresh Labrador Current Water overrides the warmer more saline residual winter bottom water. Strong evidence of layering was not found in the CTD profiles, suggesting that sufficiently quiescent conditions did not exist at the times of these observations.

TABLE 1. Mean Temperatures and Salinities at Stations in Saint-Pierre Channel and Hermitage Channel in Early Summer and Winter During 1981–1984

Station	Months	N	Depth, m	T, °C	S
11	May–June	4	100	−0.51 (0.37)	32.59 (0.11)
11	Nov.–Dec.	5	100	0.54 (0.90)	32.59 (0.20)
49	May–June	4	100	0.05 (0.26)	32.57 (0.14)
49	Nov.–Dec.	4	100	1.98 (1.01)	32.64 (0.14)
49	May–June	4	200	4.49 (1.88)	34.12 (0.46)
49	Nov.–Dec.	4	200	4.11 (0.83)	33.93 (0.15)

The numbers in parentheses are the standard deviations from the mean. N is the number of samples.

6. Summary and Conclusions

The renewal of deep water in Fortune Bay at roughly half-year intervals has been documented. Warm water exchange over the sills bordering Hermitage Channel occurs primarily in winter, driven by upwelling of Modified Slope Water in Hermitage Channel, mainly in response to winds from the north quadrant and on at least one occasion from the west. Cold water exchange takes place in summer, when Labrador Current Water flows over a sill bordering Saint-Pierre Channel and when the winds are dominantly from the southwest. Conceptually, the bay and adjacent shelf channels represent a push-pull system driven by the seasonally varying winds, and exchange occurs as part of the baroclinic response on the open shelf.

There are two critical topographic elements in the system: the several outer sills of nearly equal depth at the bay mouth, and the submerged channels on the shelf. The channels result in there being two distinct water masses outside different sills available as sources for deep-water replacement.

This is a three-basin system, with one receiving basin (the bay) and two reservoir basins (the two shelf channels). In two-basin systems the frequency of deep-water renewal is usually limited by the rate of preconditioning: the reduction of deep-water density by downward diffusion of buoyancy. In Fortune Bay, however, no preconditioning is required prior to warm water exchange because the source water density is always greater than the bottom water density in the bay. Prior to cold water renewal at the bottom, a period of deep-water erosion takes place during which the cold water inflow descends as a density current from the Saint-Pierre sill to intermediate depths. There it represents both a source of buoyancy for the underlying deep water and a source of kinetic energy for mixing.

Finally, we note that three-dimensional effects, similar to those found in deep-ocean renewal problems, are expected to be important during exchange in Fortune Bay because it is for the most part wider than the internal Rossby radius of deformation.

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