5.0 OCEANOGRAPHY

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The Hudson Bay marine ecosystem consists of two oceanographically distinct marine regions (Dunbar 1988)(Figures 1-1 and 1-2). The water properties of these regions depend mainly on exchanges with Foxe Basin and Hudson Strait and the large freshwater input from both runoff and melting sea ice in the spring and summer (Ingram and Prinsenberg 1998). An understanding of their differences is critical to the design and integration of coastal zone management initiatives.

The northern area, or **Hudson Bay marine region**, is characterized by the presence of Arctic marine water and biota, complete winter ice cover and summer clearing, moderate semidiurnal tides of Atlantic origin, a strong summer pycnocline, greater mixing and productivity inshore than offshore, and low biological productivity relative to other oceans at similar latitudes. Hudson Bay lacks the typically subarctic species that are found in Hudson Strait but does support some of the relict warm-water species found in James Bay.

The southern area, or **James Bay marine region**, is closely coupled oceanographically to the Hudson Bay marine region but its waters are typically shallower and more dilute, being modified to a much greater extent by freshwater runoff from the land. Its species composition reflects these Arctic and freshwater influences and it supports a variety of warm-water species that are relicts of an earlier connection with the Atlantic and Pacific oceans. These plants and animals have disjunct distributions and are rare or absent elsewhere in Canada's Eastern Arctic waters. Southeastern Hudson Bay is included in this region with James Bay largely on the basis of biogeography (Dunbar 1988). Strong density stratification limits mixing and leads to considerable surface warming by insolation in both marine regions.

Because of its remote location and the noncommercial nature of its marine resources, relatively few oceanographic field programs have been undertaken in this area (Martini 1986a; Ingram and Prinsenberg 1998). Seasonal ice cover effectively prevents most year-round research and the shallow coastal waters make it very difficult to conduct bay-wide research from a single research platform. Consequently, characteristics of the circulation and water mass are not well known, especially outside the open water period.

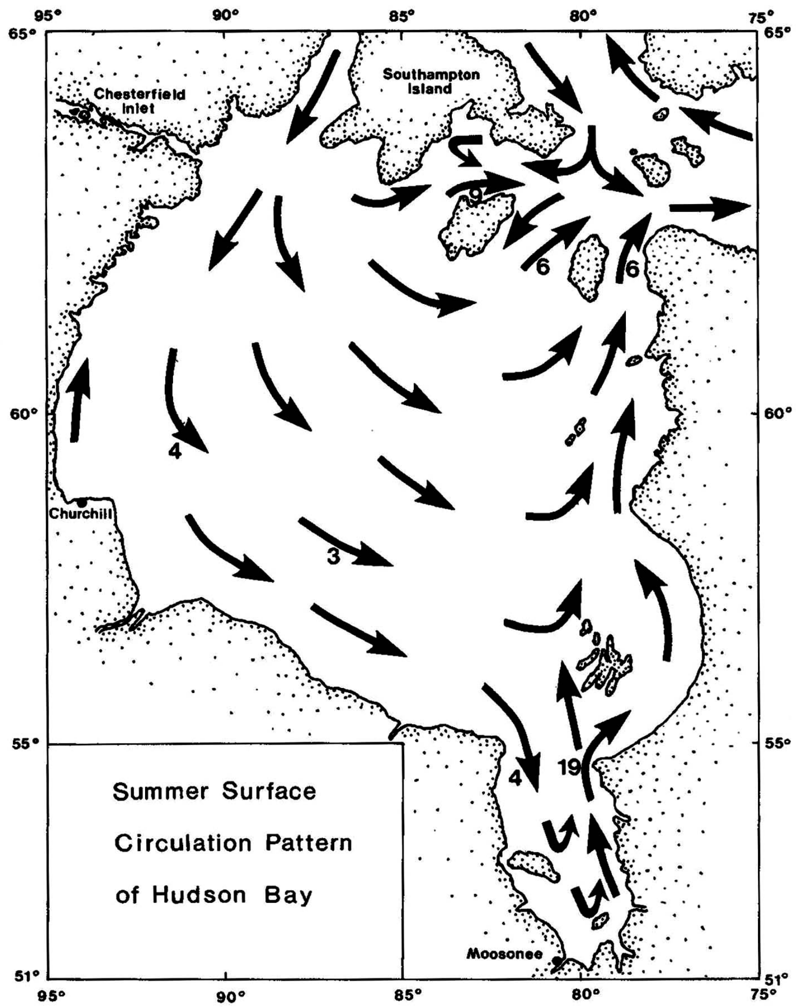

Figure 5-1. General surface layer circulation pattern for the summer condition of Hudson Bay and James Bay. Numbers are observed velocity values in cm•s^{-1} (from Prinsenberg 1986a).

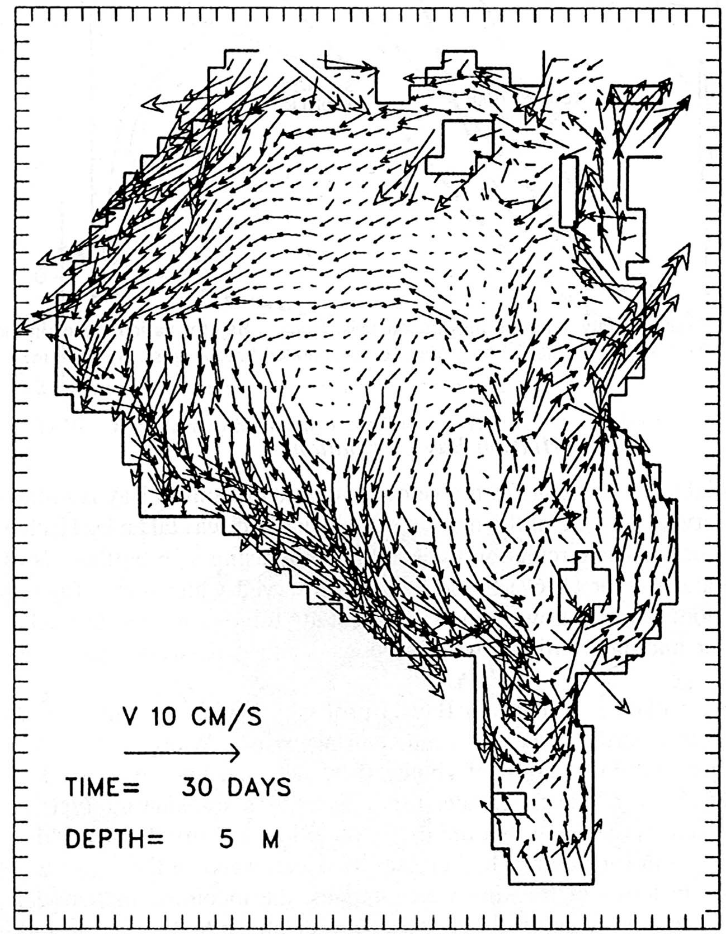

Figure 5-2. Surface circulation in Hudson Bay and James Bay determined from model results (from Ingram and Prinsenberg 1998, p. 852 as modified by J. Wang from Wang 1993).

Table 5-1. Oceanographic research expeditions to the Hudson Bay (H) and James Bay (J) marine regions, with selected references[1].

Vessel(s)	Year	Region	Selected Reference(s)
Burleigh	1914	J	Lower 1915; Melvill 1915; Comeau 1915.
Acadia	1929, 1930, 1931	H	Bell and MacFarlane 1933.
Loubyrne	1930	H, J	Davidson 1931; Hachey 1931, 1933, 1935; Huntsman 1931; Willey 1931; Vladykov 1933.
Haida	1948	H	Bailey and Hachey 1951.
Calanus	1953-1961	H, J	Dunbar 1958; Grainger and Hunter 1959; Grainger 1960, 1982; Bursa 1961; Trason 1964; Barber 1967, 1972; Squires 1967; Powell 1968; Wacasey et al. 1976; Rochet and Grainger 1988.
Lemming	1953	J	Edwards 1961.
Labrador	1955, 1956, 1959, 1967	H, J	Campbell 1958, 1959; Barber 1967.
Theta	1961	H, J	Barber 1967, 1968; Grainger 1963; Pelletier et al. 1968; Pelletier 1969, 1986.
John A. MacDonald	1962	H, J	Barber 1967, 1968.
Theron	1965	H	Pelletier et al. 1968; Hood 1969; Pelletier 1986.
Hudson	1965, 1982, 1987, 1992, 1996, 2003	H, J	Pelletier et al. 1968; Hood 1969; Pelletier 1986; Drinkwater and Jones 1987; Henderson 1989; Bilodeau et al. 1990; Josenhans and Zevenhuizen 1990; Drinkwater et al. 1991; Jones and Anderson 1994; Saucier et al. 1994; this volume Sedtion 16.
Narwhal	1972, 1973, 1974, 1975, 1988	H, J	Grainger and McSween 1976; Wacasey et al. 1976; El-Sabh and Koutitonsky 1977; Anderson 1979; Anderson and Roff 1980 a+b; Gerrath et al. 1980; Anderson et al. 1981; Prinsenberg 1976, 1977a, 1982, 1986a+b; Pett and Roff 1982; Josenhans and Zevenhuizen 1990; Josenhans et al. 1991.
Petrel	1976, 1978	H, J	Budgell 1976, 1982; Brooks 1979; Legendre and Simard 1979; Pett and Roff 1982.
Techno-Richelieu	1976	J	Legendre and Simard 1978, 1979.
L'Epinoche	1979, 1980	J	Grenon 1982.
Aiviq, Rocky Point	1990	J	Morin 1991.
Baffin	1990	J	Josenhans et al. 1991.
Fogo Isle	1993	H, J	Saucier et al. 1994; Harvey et al. 1996, 2001; Simard et al. 1996; this volume Chapter 16
Des Groseillers	2003	H	Saucier et al. 2004b

[1] See also general works by Barber (1967) and Dunbar (1982).

Oceanographic exploration began in the Hudson Bay marine region in 1929 but did not begin in earnest in the James Bay region until 1955 (Table 5-1; Dunbar 1982; Martini 1986a). Since the 1970's, most of this research has been directed toward predicting and assessing the effects of existing and proposed large-scale hydroelectric developments. Coastal conditions downstream of affected river systems, in particular the Eastmain, La Grande, Grande Baleine, Nottaway, Broadback, and Rupert rivers in Quebec and the Churchill and Nelson rivers in Manitoba, have received most attention.

Oceanographic bibliographies have been compiled for southeastern Hudson Bay (CSSA 1991; Hydro Quebec 1991a) and James Bay (Hydro Quebec 1991b). An annotated general bibliography of research on Hudson and James bays has also been prepared for DFO (Stewart 2001). The cumulative impacts of development on Hudson and James bays have been considered at workshops organized by the Department of Fisheries and Oceans (e.g., Bunch and Reeves [ed.] 1992; Gilbert et al. 1996) and the Hudson Bay Programme (Sallenave 1994). The latter initiative by the Rawson Academy of Aquatic Sciences, CARC, and the Environmental Committee of the Municipality of Sanikiluaq also includes a broad review of human impacts on the bays (Sly 1994, 1995) and a comprehensive study of traditional ecological knowledge (McDonald et al. 1995a+b, 1997).

While the descriptions that follow use the present tense, the data on which they are based were often collected in the 1970's, before extensive hydroelectric development. Oceanographic changes effected by these developments, which are considered stressors of the Hudson Bay marine ecosystem, are discussed in Chapter 15. Extensive multi-year, multidisciplinary research on the marine ecosystem is planned over the next decade under the MERICA (étude des MERs Intérieures du Canada) and ArcticNet programs (Saucier et al. 2004b; D. Barber, U. Manitoba, Winnipeg, pers. comm.; http://www.arcticnet-ulaval.ca/).

5.1 CIRCULATION

In summer, surface water circulates cyclonically (counterclockwise) around Hudson Bay, and the deep water moves in the same general direction but is influenced by bottom topography (Figure 5-1 and Figure 5-2; Hachey 1935, 1954; Barber 1967; Prinsenberg 1986a,b; Wang et al. 1994a). Cold, saline Arctic water from Foxe Basin enters Hudson Bay in the northwest via Roes Welcome Sound (Tan and Strain 1996). As it flows eastward along the southern coast of Hudson Bay some of this water enters James Bay while the remainder is deflected northward to exit northeastward into Hudson Strait. A westward, wind-driven return flow across the top of Hudson Bay has been predicted by modelling studies (Murty and Yuen 1973; Wang et al. 1994), and there is a small--perhaps intermittent, intrusion of Atlantic water from Hudson Strait at the northeastern corner of Hudson Bay. The extreme southerly incursion of Arctic waters creates Arctic oceanographic conditions much further south than elsewhere along the North American continent, and is a key feature of the Hudson Bay marine ecosystem.

Mathematical modelling suggests that the main reasons for this stable cyclonic circulation are the relatively weak coastal currents with limited coastal development to cause mixing, a relatively strong Coriolis effect that stabilizes the flow pattern by turning the freshwater outflow from rivers cyclonically around Hudson Bay, and strong density stratification due to intense freshening in summer (Wang et al. 1994a). This circulation is maintained by inflow/outflow forcing that likely occurs year round, and reinforced during the open water season by wind and buoyancy forcing. The total basin wide volume transport is an estimated 0.55 Sv, of which 0.2 Sv is inflow/outflow induced transport, 0.23 wind-driven transport, and 0.12 Sv buoyancy-driven transport. There are narrow coastal jets along all but the northern coast that appear to driven by freshwater inputs from runoff, ice melt discharge, and precipitation that create buoyancy driven circulation.

The observed monthly mean residual currents that are independent of tides are in the range of 4 to 6 $cm \bullet s^{-1}$ (Prinsenberg 1986b; S. Prinsenberg, DFO, Dartmouth, NS, pers. comm.). They are more variable at the surface than at greater depths, and are stronger at all depths during summer than winter. Seasonal variations reflect the effects of both wind stress and density-driven components of the circulation. Winds are generally weaker and more variable in summer than they are in the fall when strong northwesterly winds occur. The density-driven component is greatest in early summer when the surface freshwater input through runoff and ice melt is high. It is weakest between February and May, when surface freshwater input by runoff is reduced and is offset by the salt rejected by the growing ice, and when the ice cover limits wind stress and insulates the surface water.

Passing weather systems generate 5 to 6 day periodic motions that dominate the mean daily currents (Prinsenberg 1986b, 1987). These long-period motions are pressure-driven (barotropic) and decrease slightly with depth. They have amplitudes of up to 25 $cm \cdot s^{-1}$ and occur throughout the year, regardless of the ice condition. The mean hourly currents are dominated by pressure-driven tidal components of up to 28 $cm \cdot s^{-1}$ in amplitude. Storm winds can generate inertial currents that are as strong as the tidal currents but rotate clockwise, opposite to the tidal current direction. These wind-generated currents decrease with depth and are absent when Hudson Bay is ice-covered.

The mean residence time for water in Hudson Bay is uncertain because the contributions of runoff, ice melt and surface water to the surface layers of Hudson Bay are difficult to quantify (Ingram and Prinsenberg 1998). It has been estimated at 3 to 4 years (Jones and Anderson 1994) and 6.6 years (Prinsenberg 1984, 1986a), based on summer river runoff and salinity data. However, these estimates do not consider other sources of fresh water, so the true residence time may be on the order of one to two years (Ingram and Prinsenberg 1988).

The presence of a thin mud veneer and an absence of current bedforms indicates that modern (erosive) seafloor currents are limited or absent throughout much of Hudson Bay (Josenhans and Zevenhuizen 1990). However, erosional furrows in modern muddy sediments do evidence localized bottom currents at depths of 175 m offshore the La Grande River (Josenhans et al. 1991).

The circulation of water in the James Bay marine region is closely coupled with that in the adjacent Hudson Bay marine region. James Bay has a two-layer system of circulation consisting of an upper layer 20 to 50 m thick with a net outward flow, and a lower layer with a net inward flow (Prinsenberg 1982b). The summer circulation at the mouth of James Bay is characterized by a relatively slow mean inflow in the bottom layer and in the western half of the surface layer (2 to 5 $cm \cdot s^{-1}$), with a faster outflow (10 to 20 $cm \cdot s^{-1}$) concentrated in the eastern half of the surface layer (Figure 5-3; El-Sabh and Koutitonsky 1977; Prinsenberg 1982b, 1986b). Currents are generally faster in October than August. The net outward water transport varies between 65,000 $m^3 \cdot s^{-1}$ in August and 167,000 $m^3 \cdot s^{-1}$ in October, most of it along the east side (El-Sabh and Koutitonsky 1977). The estimated summer flushing time for James Bay is 10 months, with residence times of water in the surface (upper 10 m) and bottom layers of 3 and 7 months respectively (Prinsenberg 1982b, 1984). The winter residence time of 20 months is a reflection of the slower winter circulation.

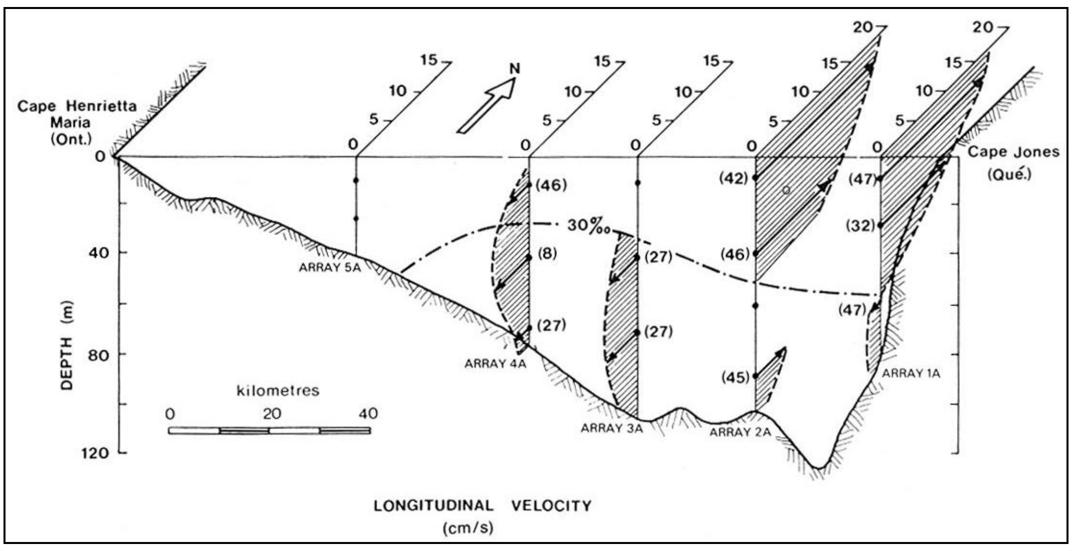

Figure 5-3. Mean longitudinal velocity distribution during the summer of 1975 at the entrance to James Bay (from Prinsenberg 1982b, p. 828). Numbers in brackets indicate the length, in days, of the useable current meter record.

The cyclonic gyre in northern James Bay continues year-round, driven partly by wind stress and partly by density differences due to runoff, with strong coastal jets on each side of the entrance (Peck 1976; Prinsenberg 1976, 1982b; 1986b; Wang et al. 1994a). Seasonal variations reflect the effects of both components of the circulation. As in Hudson Bay, passing weather systems can superimpose cyclic variations in the circulation at 4 to 6 day periods during the open water season. The magnitude of these variations depends upon the strength and directional persistence of the wind. A persistent southerly wind of about 8 $m \cdot s^{-1}$ (29 $km \cdot h^{-1}$) can reverse the surface outflow from James Bay. Once the Bay is ice-covered, the effects of wind forcing on the circulation are limited.

Runoff contributes to the dynamics of the James Bay circulation by diluting the salt water and creating density-driven currents. These currents are restricted to coastal waters (up to 50 km wide), and are greatest in early summer when runoff is high (Prinsenberg 1982a). Indeed, from summer to winter the magnitude of the density-driven currents at the inflow to James Bay decreases from 3.5 to 1.3 cm·s^{-1} and at the outflow from 15 to 5 cm·s^{-1} (Prinsenberg 1982a). Because the magnitude of density-driven currents is proportional to the runoff rate, hydroelectric developments that increase winter runoff will also increase winter circulation (Prinsenberg 1982a, 1991). Completion of the planned hydroelectric developments on rivers that flow into James Bay would double the total freshwater input into James Bay in winter (Prinsenberg 1980, 1991).

The pattern of water circulation in southeastern Hudson Bay is not well known, but there is a general movement of surface water east and north, following the cyclonic gyre that persists throughout Hudson Bay (Prinsenberg 1986b; McDonald et al. 1997). The area is affected by the part of the Hudson Bay water mass that continues eastward and northward instead of flowing southward into James Bay, and by the well-defined, surface current that flows northward along the eastern shore of James Bay. Part of the latter current crosses the mouth of James Bay to return southward with incoming water from the west; the remainder continues northward into southeastern Hudson Bay. Little is known of currents in Richmond Gulf.

Under-ice circulation has been estimated in early June near Kuujjuarapik using ice drift (Larouche and Dubois 1988, 1990). When there is less than 90% ice cover and floes are free to move, changes in ice floe movement are strongly correlated with the amplitude and direction of the wind, with a time lag that depends on the strength of the changing winds (Larouche and Dubois 1988, 1990). These wind-effects can be identified and removed to estimate the spring surface currents that are otherwise difficult to measure in the presence of moving ice. This method provided an estimate of 28 cm·s^{-1} for current near the southeastern coast of Hudson Bay in early June, but it has not been used to study under-ice currents elsewhere in Hudson Bay.

Knowledge of marine currents is important to Inuit, who travel on the water and ice and harvest birds and marine mammals that depend upon currents for access to food resources (McDonald et al. 1997). Maps based on their traditional knowledge of surface currents in eastern Hudson Bay and James Bay are closely similar to those prepared by oceanographers. Inuit recognize that the current strength varies from year-to-year but believe that there has been a general weakening of the surface currents since the 1950's. In support of this view, they have described a progressive increase in the ice cover in southeastern Hudson Bay that reduced the number of polynyas that remain open year round in the Belchers from 35 in the 1970's to 3 in the early 1990's. They are now able to cross Roes Welcome Sound, between western Southampton Island and the mainland, during the summer spring tide and, in 1992, a large polynya that seldom froze in January began freezing over in November and December. In some areas the floe edge has moved offshore and the strength and behaviour of the ice edge has also changed. Inuit wonder whether changes in the seasonality of river flows, caused by hydroelectric developments, have weakened the currents.

5.2 TIDES

Powerful tides surge into Hudson Bay twice daily via Hudson Strait (Dohler 1968; Drinkwater 1988). These semidiurnal tides originate in the Atlantic Ocean and overshadow local tides and any tidal influence from the Arctic Ocean. The main tidal constituent, the M_2 (principal lunar), is a Kelvin wave that propagates counterclockwise around Hudson/James Bay following the contour of the shoreline (Figure 5-4; Godin 1972; Freeman and Murty 1976; Prinsenberg and Freeman 1986). Part of the wave enters James Bay while the remainder, which is reduced in amplitude, continues northward along the east coast. As it passes south of the Belcher Islands, reflection from the shallow bar to the south produces a standing wave pattern. Offshore, in west-central and east-central Hudson Bay, components of the wave interfere and cancel each other, creating areas where there is little if any change in water level. When it joins the incoming tide in northern Hudson Bay, some 25 hours after it first entered, the wave is 10% of its original amplitude.

The damped progressive wave that enters James Bay is slowed in the vicinity of Akimiski Island, where it divides (Manning 1950; Godin 1972; Martini and Grinham 1984). One segment swings east around the island while the other is funnelled through Akimiski Strait. The Ekwan Shoal area of Akimiski Strait experiences the

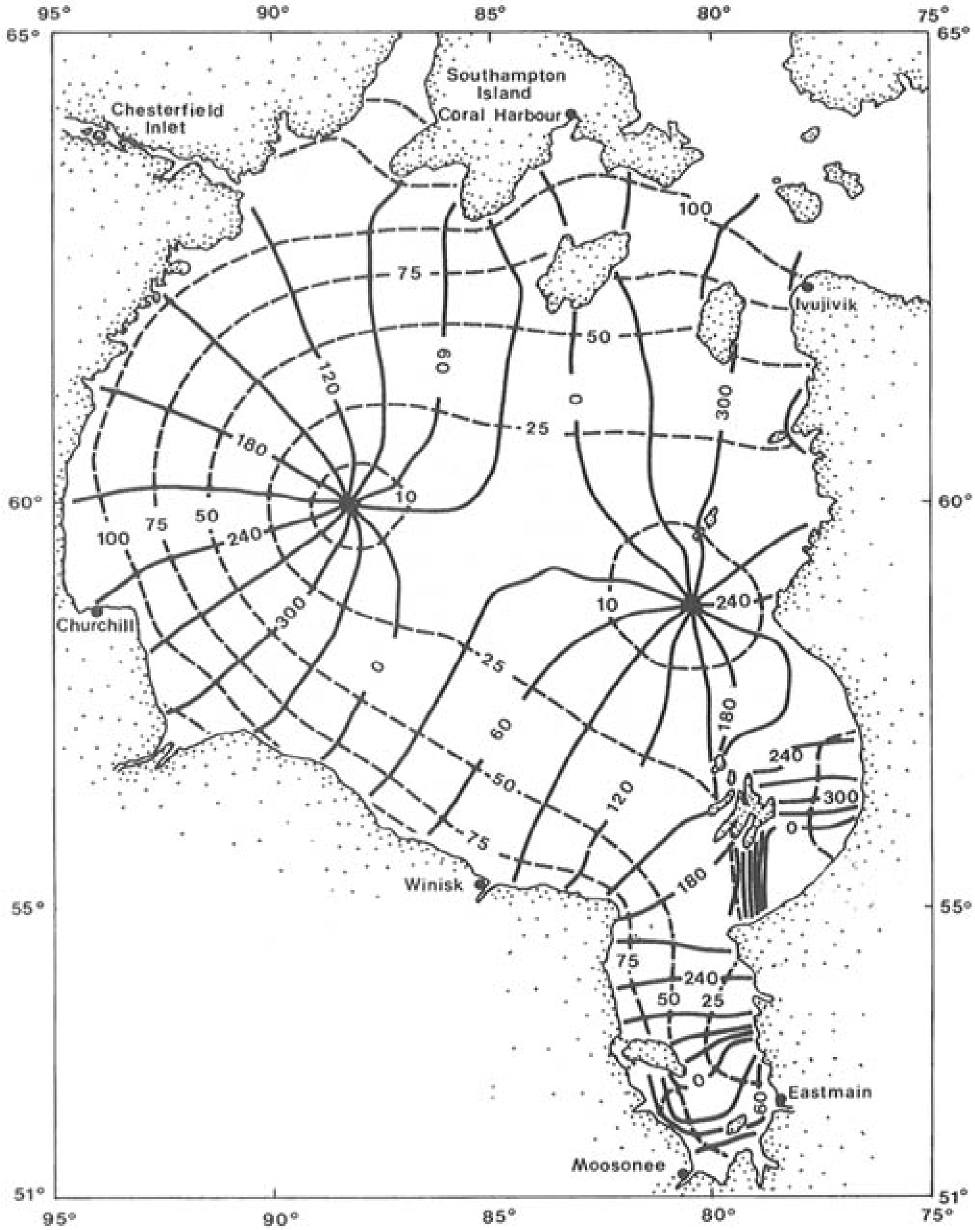

Figure 5-4. Lunar semidiurnal (M2) tide in Hudson Bay and James Bay calculated by Freeman and Murty (1976) (from Ingram and Prinsenberg 1998 p. 854). Solid lines represent cophase lines in GMT + e (degrees); dashed lines represent co-amplitude lines (cm).

highest tides (2.5 to 3.0 m) and fastest alongshore tidal currents (220 $cm{\cdot}s^{-1}$) of western James Bay (Martini and Grinham 1984). The low areas between the shoals and mainland are swept and reworked by relatively powerful reversing currents created by strong flood tides that ebb out both ends of the Strait. Ice blocks the passage of tides through the Strait from November through late June (Martini 1981b). There is a degenerative node in east-central James Bay, opposite Akimiski Island, where the incoming and outgoing tides partially cancel one another (Godin 1972).

The tide reaches Rupert Bay seven hours after it rounds Cape Henrietta Maria (Godin 1972). Tidal flow in the shallow Rupert Bay estuary is strong, and there are numerous frontal zones wherein large gradients in physical properties occur (Veilleux et al. 1992). Typical tidal speeds are 50 to 150 $cm{\cdot}s^{-1}$ (Ingram 1977). The average tidal range is about 2 m at the mouth of the bay decreasing to 1.3 m at its head--2.9 and 1.9 m respectively for spring tides (d'Anglejan 1980). The daily tidal excursion is about 10 m. Intense mixing creates homogeneous conditions in many areas (Veilleux et al. 1992). The tidal fronts are parallel to the principal channels or around small downstream islands, with the arrangement influenced by topography. Seasonally, the extent of salt intrusion does not appear to differ between the river flood stage and the low-water winter stage, since tidal friction under the ice cover approximately balances the reduction in freshwater discharge (Michel 1978 in d'Anglejan 1980). Salt is not introduced very far by the average tide--never beyond Stag Rock (d'Anglejan 1980).

Tides in the Hudson Bay marine ecosystem are classified as semidiurnal but there is also a weak diurnal tide (Prinsenberg and Freeman 1986). In general, both tidal amplitude and range decrease progressively moving counterclockwise along the coast (Dohler 1968; Prinsenberg and Freeman 1986). The extreme tidal ranges and velocities found in eastern Hudson Strait are not found in either bay. The semidiurnal tidal amplitude ranges from 1.50 m along the western shore at Churchill to 0.10 m along the eastern shore near Inukjuak, while the diurnal tide is only 0.08 and 0.03 m at the respective locations (Figure 5-4; Prinsenberg and Freeman 1986; Prinsenberg 1988a). The range in height between high and low water also changes around the bays, increasing from 3 m near the entrance of Hudson Bay to a high of 4 m along the west coast at Churchill Harbour and then decreasing to about 2 m along the east side of James Bay and in the Belchers and to 0.5 m at Inukjuak (Inoucdjouac) (Dohler 1968; Godin 1974). Tidal currents of 90 to 100 $cm \cdot s^{-1}$ have been measured at the entrance to Hudson Bay; within the bay they are generally less than 30 $cm \cdot s^{-1}$ (Dohler 1968; Prinsenberg and Freeman 1986). The tidal streams are strongest in the western part of the bay. Tidal currents of 50 $cm \cdot s^{-1}$ have been measured at the entrance to James Bay (Prinsenberg and Freeman 1986), and they can reach 80 $cm \cdot s^{-1}$ near Sand Head in the Moose River estuary at spring tide (Godin 1972). Unlike wind-generated currents the tidal currents recur daily and decrease slowly with depth.

Using a two-dimensional model of the semidiurnal tide, Griffiths et al. (1981) predicted that the tidal currents in central Hudson Bay are too weak to disrupt water column stratification during the summer. They also predicted vertical mixing of the water column and increased biological productivity on a small-scale off islands, headlands, and promontories where there is locally increased tidal streaming. Areas they identified included Frozen Strait, the entrance and exit to Roes Welcome Sound; the entrance to Wager Bay; waters offshore Cape Kendall and Cape Low on Southampton Island, Cape Acadia on Mansel Island, Cape Churchill, and Cape Henrietta Maria; Rankin Inlet; the Nelson River estuary; and western James Bay, particularly downstream from Akimiski Island. However, their model did not take into account wind mixing or the stabilizing effects of runoff, and thus may only be applicable to offshore regions of Hudson Bay (Prinsenberg and Freeman 1986).

Three factors may affect the regularity of the tides, the spring freshet, ice cover, and weather disturbances. During April-May ice breaks up in the rivers surrounding James Bay and their discharge increases abruptly (Godin 1972). This increase inhibits tidal penetration into the rivers and a wall of freshwater may extend well into James Bay, where the tide is reduced in range and the mixing of salt and fresh water intensifies. Ice cover damps the tide and advances the time of arrival of high water and flood current (Godin 1980, 1986; Godin and Barber 1980; Ingram 1982; Prinsenberg and Freeman 1986; Prinsenberg 1988a; Lepage and Ingram 1991; Prinsenberg and Ingram 1991). Semidiurnal tides in western Hudson Bay arrive 20 minutes earlier in winter than summer and show a 7 to 10% reduction in current and height; at Chisasibi, in eastern James Bay, they also arrive 20 minutes earlier in winter than summer but show a 30% reduction in amplitude (Godin 1986). The seasonal fluctuations are even greater at Inukjuak--where the tides arrive 40 minutes earlier in winter, probably due to displacement of the point of amphidromy by ice formation. Diurnal tides are similarly but less predictably affected. Weather induced tidal storm surges are discussed below.

Recent modelling of the ice-ocean seasonal cycle suggests that tides are an essential control on the regional pack ice and ocean climate of Hudson Bay (Saucier et al. 2004a). They suggest that tidal mixing controls the sensible heat transfer, thereby reducing sea ice formation, and is responsible for maintaining the coastal polynyas.

Chesterfield Inlet is an estuary with stronger tidal currents, higher tidal amplitudes, and a greater degree of mixing than elsewhere in Hudson Bay (Dohler 1968; Budgell 1976, 1982). It has tidal ranges of 5 m at spring tides and 2 m at neap tides in the lower third of the estuary, and of 3 m and 1.5 to 2 m over the remainder of the estuary (Budgell 1982). There is a strong tidal influence up to the entrance of Baker Lake where the combination of tidal dissipation at a shallow sill and the sharp increase in cross sectional area reduce the tidal range to less than 0.10 m. Tidal current in the inlet is reversing. Amplitudes of over 100 $cm \cdot s^{-1}$ are observed throughout the channel, with current speeds of 150-200 $cm \cdot s^{-1}$ near channel constrictions. The tidal streams at Chesterfield Narrows are strongly influenced by the freshwater outflow from Baker Lake (Dohler 1968) and the tides in turn likely contribute salt water to Baker Lake, which has a bottom layer of very dilute seawater (Johnson 1965).

5.3 WAVE CLIMATE AND STORM SURGES

Data on wave heights and periods in Hudson Bay and James Bay are scant. They suggest median August and September wave heights in Hudson Bay of 1 to 2 m with periods of 5 to 6 seconds (Maxwell 1986). During the open water season, wave heights over 3 m occur about 10 % of the time in northern Hudson Bay, with most of the largest waves originating from the northwest (Cohen et al. 1994). Wave heights of 8 m with periods of 10 seconds have been recorded in September in northern Hudson Bay (Maxwell 1986).

Strong storm surges sometimes occur in southern James Bay (Manning 1951; Godin 1972, 1975). These surges are most likely to occur during the storm seasons, namely between September to December and April to June. They pose a significant hazard to travellers who may be unprepared for a tide that can extend kilometres inland beyond the normal high water mark (Godin 1975). Murty (1972) predicted that if there were an 89 $km \cdot h^{-1}$ (51 mph) wind along the north-south axis of James Bay the amplitude of these surges could reach up to 6 m (18.8 ft) along its southern shores.

5.4 SEA ICE

The presence of winter ice cover is a very important feature of the Hudson Bay marine ecosystem. Its southern extent and the presence of extensive areas of fast ice are unusual and strongly affect the physical and biological oceanography, the surrounding land, and human activities. The reliance of Inuit and coastal Cree on sea ice for travelling and hunting is reflected in their detailed knowledge of its processes, characteristics, and annual cycles (McDonald et al. 1997). The sea ice determines the ecology of the ice biota and it also influences pelagic systems under the ice and at ice edges (Melnikov 1980; Legendre et al. 1992a). As the interface between air, ice, and water, ice edge habitats are areas of mixing that attract biota to feed. These areas are important sites of energy transfer within the ecosystem. Unlike marine regions to the north, Hudson Bay and James Bay are ice free in the summer.

During ice growth, most of the original seawater is rejected from the ice. The rejected salt increases the density of the surface water, which then enhances the deepening of the surface mixed layer by tidal current mixing (S. Prinsenberg, DFO, Darmouth, NS, pers. comm.). The surface layer deepens at places to 100 m, thus distributing the rejected salt to most of the water column. In contrast during the spring, the melting of the ice cover constitutes a freshwater flux to just the ocean surface layer, increasing the vertical stability of the water column and decreasing the vertical nutrient flux. If the pack ice did not move, the ice growth and decay would represent a salt flux to the total water column in the fall and winter, and a freshwater flux to the surface layer in the spring and early summer. However under predominantly northwest winds, the pack ice continually removes the ice from northwestern Hudson Bay and moves it to southeastern Hudson Bay where it rafts and ridges. These pack ice processes increase the mean spatial ice thickness in southeastern Hudson Bay (Prinsenberg 1988b), which in the spring represents a large freshwater input from the decaying ice cover. In northwestern Hudson Bay, in contrast, the ice is continually removed, encouraging new ice growth and continual salt input to the water column. Numerical simulations duplicate these pack ice properties (Saucier et al. 2004a). Sea ice begins to form in northwestern Hudson Bay and ice generally moves towards the southeast in early winter where it ridges to mean a thickness of 1.5 to 2.0 m by April. Thinner ice is simulated for northeastern Hudson Bay. Monthly mean ice drift velocities range from 10 $cm \cdot s^{-1}$ in northwestern Hudson Bay to 1 $cm \cdot s^{-1}$ in southeastern Hudson Bay where higher ice concentration and ridging occurs. The strong northwest winds and tidal mixing encourage the reoccurring leads and polynyas along the western shore of Hudson Bay.

Recent advances in satellite technology have facilitated detailed studies of winter sea ice in Hudson Bay and James Bay. Unless otherwise noted, the description that follows is based on two informative publications by W.E. Markham (1986, 1988), with updated information from Cohen et al. (1994). Markham's latter publication, an "Ice atlas of Hudson Bay and Approaches", is based on the interpretation of at least 20 years of summer and 6 years of winter ice data that were collected by satellite on a weekly or biweekly basis. Some useful earlier studies include: Low (1906), Hare and Montgomery (1949), Larnder (1968), and Danielson (1971).

Differences exist between conventional and passive microwave sea ice datasets for Hudson Bay (Etkin and Ramseier 1993). The conventional datasets, which are based on direct observation, are more accurate during the spring melt when meltwater ponding on the ice is mistaken by the remote sensing algorithms for open water. They are less accurate during freezeup when it is difficult to observe the presence of new ice. Remote sensing has also tended to underestimate the density of older pressure ridges that are obscured by snow cover (Hudier et al. 1993).

5.4.1 Terminology

During the cycle of ice formation and decay, sea ice may pass through a number of stages. From the time ice crystals begin to form slush at the surface until they coalesce the sea ice is termed **new ice**. Once the crystals coalesce into ice with a definite form the ice is termed **young ice**. Initially it forms a 5 to 10 cm thick elastic layer that tears rather than breaks when penetrated by a ship, and that bends on waves. The young ice loses its elasticity as it thickens from 10 to 30 cm. The final stage of ice growth in the area is **first year ice**. It can be thin (30-70 cm), medium (70-120 cm), or thick (120-200 cm) depending on the temperature regime. Ice that has survived one or more summers after its formation is termed **old ice**. Old ice is often 2 or 3 m thick and is harder than first year ice because it is nearly salt-free. It poses a greater hazard to ship navigation but seldom penetrates southward into Hudson Bay and James Bay.

Normally, ice forms first in shallow water areas. If it remains attached to shore it is termed **fast ice** (or **landfast ice**) and if it is carried away by wind, tide or current, **drift ice**. Drift ice and **pack ice** are synonymous, but pack ice is generally reserved for situations when 7 to 10 tenths of the water surface is covered by ice. When ice floes are pressed together by wind or current their edges may overlap. This **rafting** is most common on young ice. Ice fragments are forced upward and downward when thicker first year ice grinds together. When this occurs along a continuous edge it forms **pressure ridges**, and when it occurs in one area a **hummock**. A pressure ridge that is 1.0 m in height may extend 4 or 5 m below the ice floe, and poses a hazard to surface and submarine navigation.

This is a much simpler system of describing ice conditions than that used by Inuit, who use 71 distinct terms to describe each different ice condition through five stages of development (McDonald et al. 1997). Inuit knowledge of ice development is important for the assessment of how predicted changes in climate are affecting sea ice, locally and in the Hudson Bay marine ecosystem as a whole (see Chapter 17).

Ice cover is determined mainly by the amount of heat exchanged between ice, water, and air. These processes are complex and often difficult to measure. The number of degree days when the mean air temperature is greater than (melting degree days) or less than (freezing degree days) 0°C are useful indicators of the heat exchange. The ice conditions in Hudson Bay and James Bay are very variable early in the season when warm or cold spells can have a relatively great impact on the number of freezing or melting degree-days. Wind and the presence of cold incoming Arctic waters in the northwest are also important determinants of the ice cover.

5.4.2 Seasonal Changes

Markham used median rather than mean ice concentrations to describe ice cover. This has the advantage of representing the mid-point in the range of observed conditions rather than the mathematical average, which may rarely occur.

Median ice concentrations in the Hudson Bay marine ecosystem begin to increase in October in the Repulse Bay area and during November the ice cover spreads rapidly southward along the western coast of Hudson and James bays and then, more slowly, eastward (Figure 5-5). From November through late June ice impedes circulation and tides in Akimiski Strait (Martini 1981b). It causes the Strait to behave as an ice-walled bay, largely forcing circulation and tides outside Akimiski Island.

By mid-December the region is between 9 and 10 tenths ice-covered. Fast ice fills bays along the west coast of Hudson Bay south to Arviat and along the west coast of James Bay from Akimiski Island south to Rupert Bay, and forms around the Ottawa and Belcher islands. Ice growth has reached 65 cm at Chesterfield Inlet, Coral Harbour, and Churchill. There is a steady growth in ice thickness from January until April, with the ice cover remaining in the range of 9 to 10 tenths. In southeastern Hudson Bay, the conolidation of the pack into fast ice normally occurs between mid-January and mid-March (Larouche and Galbraith 1989).

The development of extensive areas of fast ice is an important feature of the ecosystem. By the end of April the fast ice edge south of Rankin Inlet has moved further offshore and extends in a narrow strip southward from Arviat to Churchill (Figure 5-6). There is a band of fast ice along the south coast of Hudson Bay from York Factory to Cape Henrietta Maria and along the entire coast of James Bay; the Ottawa, Belcher and King George archipelagos are surrounded; and there is a wide band of fast ice along the eastern shore from Cape Smith south to Pointe Despins. It is common for fast ice to cover much of southeastern Hudson Bay and surround the Belcher and King George archipelagos for part of the season. Indeed, in some cold winters the whole area from Cape Jones to the Belchers to the Ottawas to Cape Smith becomes one consolidated mass for a short period. The western extent of fast ice cover in southeastern Hudson Bay, between Kuujjuarpik and the Belcher Islands, appears to be controlled mainly by wind and temperature (Larouche and Galbraith 1989). During the freezing period, strong southwesterly winds can reduce the consolidation of the pack ice into fast ice in this area and, on rare occasions, melting conditions can largely prevent fast ice from extending away from the mainland coast.

Maximum ice cover occurs in April and May when the median ice concentrations are high and the ice thickness nears its peak. Depending upon the year and location, the maximum ice thickness can occur between late February and early June and range from 71 cm at Moosonee to 285 cm at Inukjuak (Table 5-2). Modelling studies suggest that changes in runoff have a greater effect on the interannual variability of the ice cover than do temperature changes associated with the North Atlantic Oscillation, especially in southeastern Hudson Bay (Saucier and Dionne 1998). However, most of the variability in ice cover is likely related to differences in the summer and autumn winds, air temperature (which control heat loss and winter preconditioning), spring cloud cover (which controls heat gain), and snow cover (which controls the winter insulation). By the end of April the fast ice edge south of the Belchers is beginning to recede.

The ice floes are kept in constant motion by the wind. Leads develop when the winds blow offshore and are quickly covered by new and young ice. These leads are important habitat for species such as the Hudson Bay eider and longtailed duck that overwinter in the region, and to migratory birds and mammals that arrive early in the spring (Prach et al. 1981; Nakashima 1988; Stirling 1997; Gilchrist and Robertson 2000). There are recurring leads along the west shore of Hudson Bay from Churchill to Chesterfield Inlet and Coral Harbour, and elsewhere along the fast ice edges. Open water that occurs along fast ice edges in the Belchers is not present during severe winters (Gilchrist and Robertson 2000).

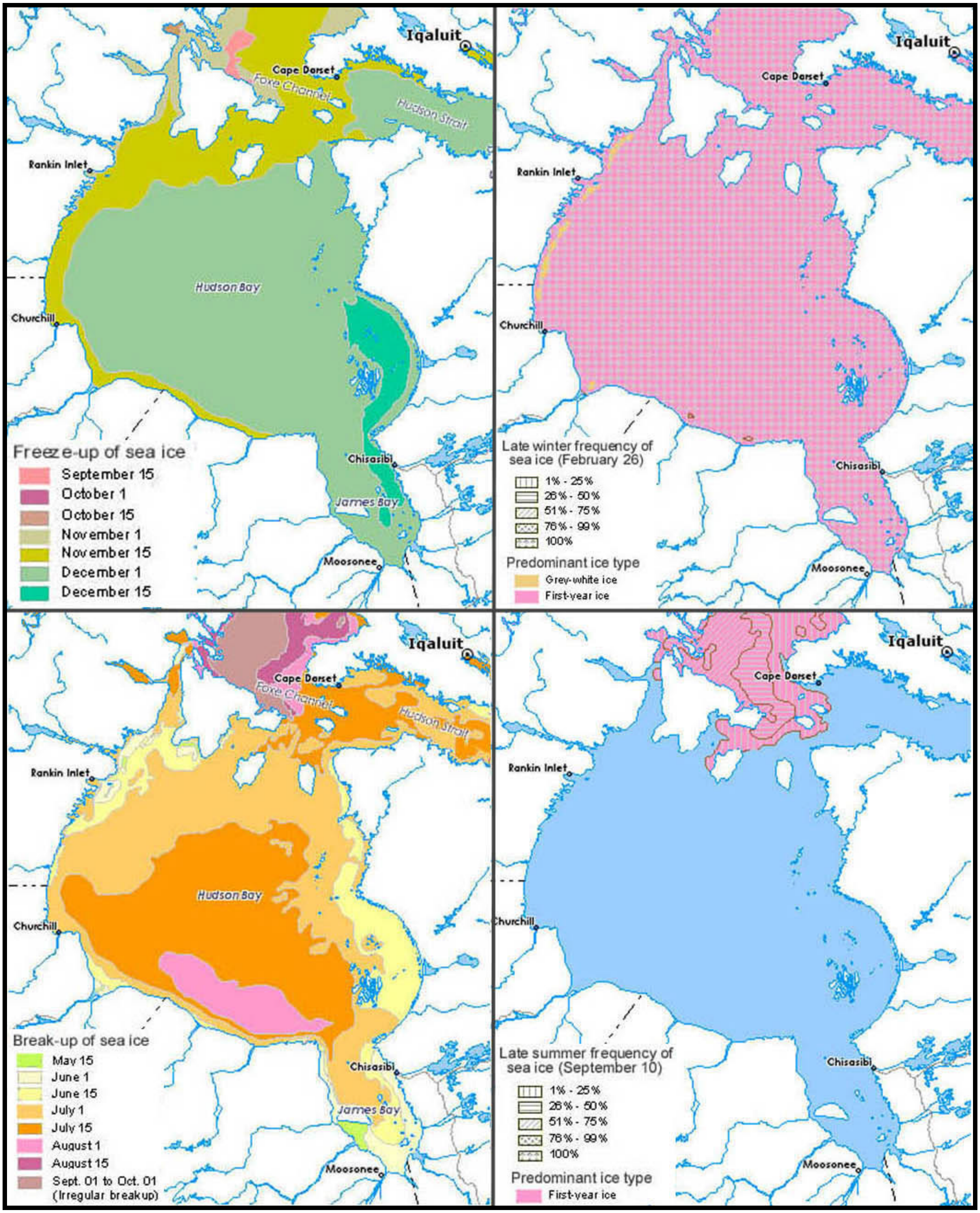

Figure 5-5. **Patterns of sea ice freeze-up (top left) and break-up (bottom left) and frequency and type of late winter (top right) and late summer (bottom right) sea ice, based on 30 years of data (adapted from National Atlas of Canada 2003).**

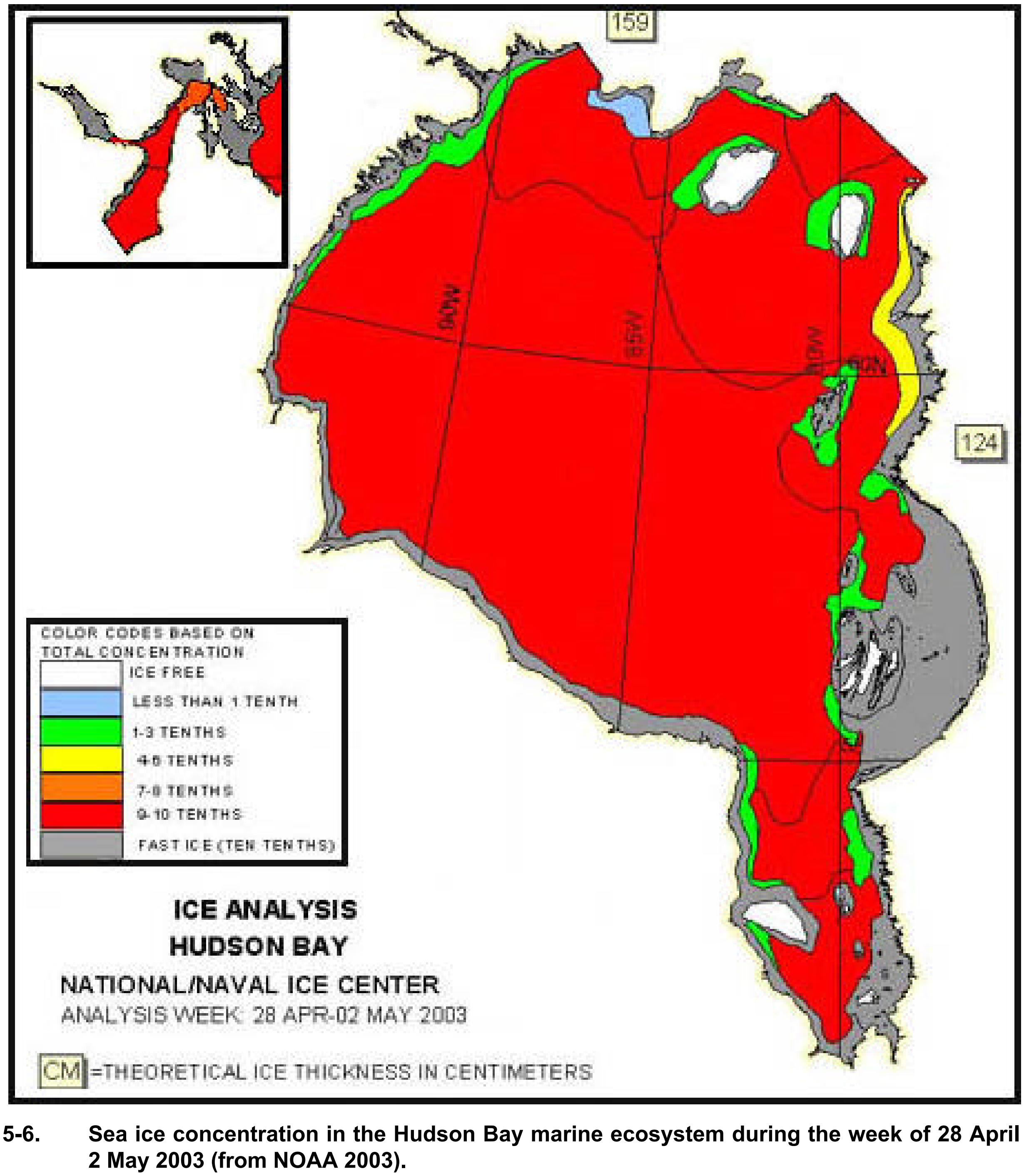

Figure 5-6. **Sea ice concentration in the Hudson Bay marine ecosystem during the week of 28 April to 2 May 2003 (from NOAA 2003).**

Table 5-2. Range of maximum ice thickness (augmented by 10% of snow cover) over the period 1963-83, corresponding range of dates, and range of dates of open water (from Loucks and Smith 1989).

Community	Maximum ice thickness (cm)	Date (Julian days)	Date of open water (Julian days)
Coral Harbour	143-208	114-162	177-213
Chesterfield Inlet	163-226	101-158	178-210
Inukjuak	176-285	71-157	156-189
Churchill	138-207	49-151	122-200
Kuujjuarapik	113-220	84-124	141-154
Moosonee	71-130	114-162	177-213

Small recurring polynyas are present in the Belchers and near islands along the coast of southeastern Hudson Bay (Figure 5-7), in Roes Welcome Sound, at the northern tip of Coats Island, near Digges Island, and just off the southwest tip of Akimiski Island (Martini and Protz 1981; Stirling and Cleator [ed.] 1981 Nakashima 1988; Gilchrist and Robertson 2000). The latter polynya is one of the most southerly in Canadian seas. These openings in the sea ice are vitally important to overwintering species and to early spring migrants. They are often areas of increased biological productivity.

Ice ridges and hummocks are formed in pack ice regions where its drift is restricted by shores. Ice ridges are common in northern and southern Hudson Bay and in James Bay, with 4 to 10 ridges$\cdot$km^{-1} (see also Prinsenberg 1988a). Twenty-six percent are over 1 m in height and they can reach heights of 3 to 3.5 m. Ridges are less common and smaller in northwestern Hudson Bay due to the offshore drift.

In early June the coastal leads become broader and more persistent. Median ice concentrations fall to the open water state in Chesterfield Inlet and there is open water in southern James Bay. Lower median ice concentrations are also evident along the east shore of Hudson and James bays and in Roes Welcome Sound (see also Hare and Montgomery 1949). With the lengthening days the ice is melting everywhere by late June, and extended warm or cold spells or periods with clear skies can determine whether breakup is early or late. Sea ice in James Bay is often sediment-laden, and this discolouration hastens melting in the spring.

The same melting pattern continues in July. Reduced ice concentrations and partial clearing are apparent in northwestern Hudson Bay where winds tend to move the pack ice offshore, and in eastern Hudson Bay from the Belcher Islands to Mansel Island where the ice is melted in place by the northward flow of spring runoff from James Bay (see also Markham 1976).

By mid-July the pack is located in the area from Churchill to the Belcher Islands to Mansel Island. While the median concentration is still 6 to 8 tenths, it has decayed to the point where it no longer poses a hazard to navigation. The actual amount of ice and its location varies widely at this time of year depending on the winds. Indeed, in some years the pack has been located off Chesterfield Inlet. Because ice in the floes is relatively flat there is extensive puddling or ponding on the surface during the melt period and, because the thickness is relatively uniform, large areas of ice melt within a very few days. The median ice concentration reaches open-water conditions in the first week of August and continues until freeze-up in the fall except in the Repulse Bay area, which receives ice from Foxe Basin (Figure 5-5).

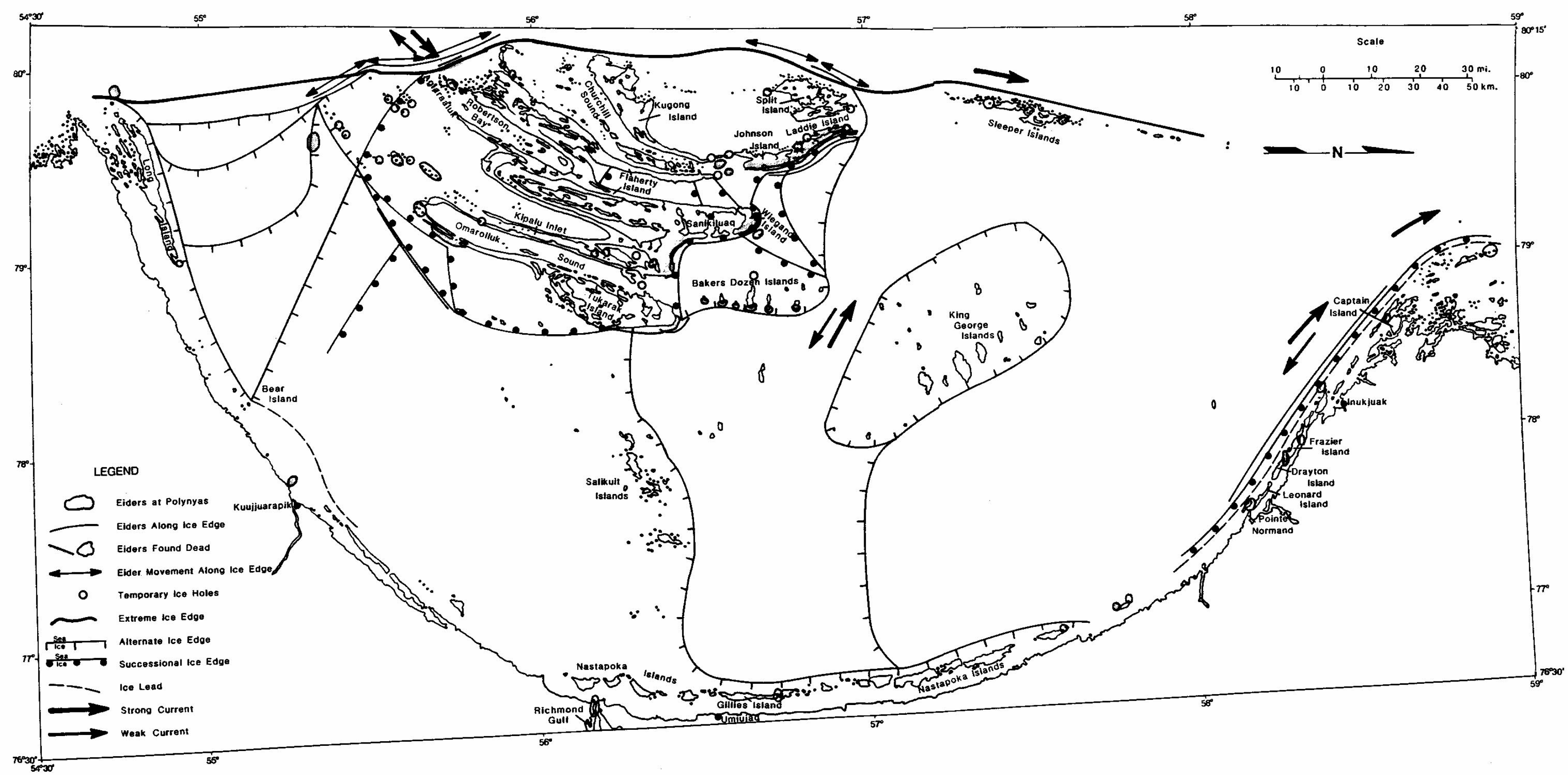

Figure 5-7. Winter ice formation and eider distribution in southeastern Hudson Bay (from Nakashima 1988).

Depending on weather conditions, the timing of freeze-up or breakup may be retarded or advanced by up to a month, but the basic pattern of ice formation remains similar (see also Barber 1972). In 1816, for example, open ice lingered a month longer than usual in northern James Bay and northwest of the Belchers (Catchpole and Faurer 1985). The extent of the fast ice also varies. In some winters, the whole area from Cape Smith to the Ottawa Islands to the Belcher Islands to Cape Jones can be covered by fast ice. The inter-annual variability of sea ice in Hudson Bay is related to large-scale atmospheric circulation changes (Wang et al. 1994c; Mysak et al. 1996). During strong winter westerly winds of the North Atlantic Oscillation and Low/Wet summer episodes of the Southern Oscillation, the sea ice grows thicker and breakup is delayed. While ice area has been directly correlated with runoff volume the previous year (Manak and Mysak 1989; Wang et al. 1994c), a direct cause-effect relationship between the two parameters has not been well established (see also Saucier et al. 2004a). Throughout Hudson Bay and James Bay, the melt period is longer and more variable in its timing than the freezeup period (Figure 5-8; Cohen et al. 1994).

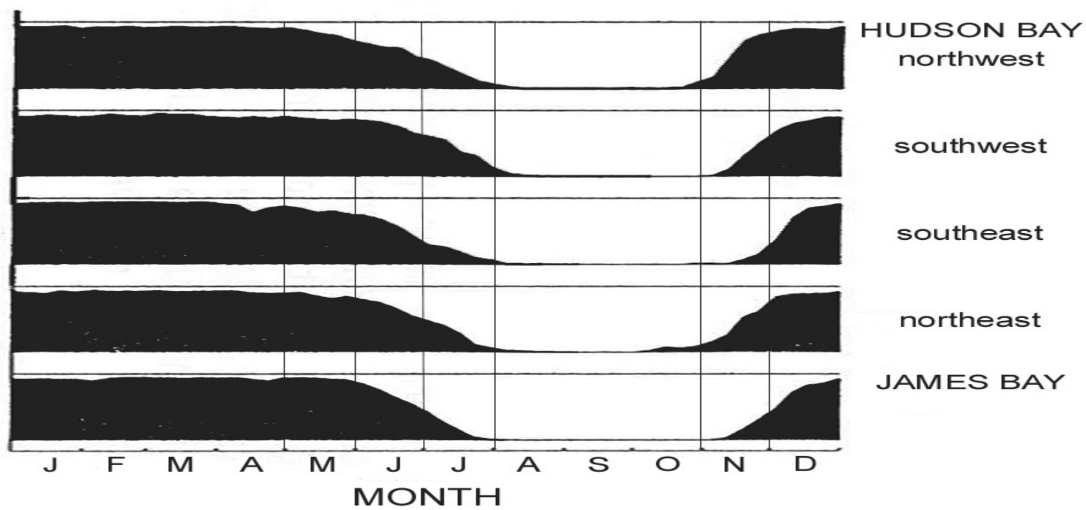

Figure 5-8. Mean ice concentrations (1972-88) by month for locations in Hudson Bay and James Bay (modified from Cohen et al. 1994).

Old ice and icebergs are rare in Hudson Bay and rare or absent James Bay. There are incursions of old ice from Foxe Basin to just south of Coats and Mansel islands between 1% and 25% of the time in October, and on rare occasions it penetrates southward to Arviat (see also Fleming and Newton 2003). An iceberg has been reported in the Mansel Island area.

The importance of sea ice to the Hudson Bay marine ecosystem and its vulnerability to climatic warming have spurred efforts to develop a mathematical model that accurately simulates the region’s sea ice dynamics (Wang et al. 1994b, 2003; Saucier and Dionne 1998; Senneville et al. 2002; Gachon and Saucier 2003; Saucier et al. 2004a). This work is enabling oceanographers to test the relative importance of strong westerly winter winds created by the North Atlantic Oscillation, anomalous high runoff, hydroelectric regulation of runoff, strong autumn winds or low autumn surface air temperatures, climatic warming, tidal forcing, and other factors that affect the sea ice. The modelling results suggest that autumn winds and air temperatures may have a greater effect on the thickness of sea ice than either strong winter westerlies, or regulated runoff conditions (Saucier and Dionne 1998). They also suggest that tidal mixing, which controls sensible heat transfer, is an important factor limiting winter sea ice volume and creating polynyas (Saucier et al. 2004a). The location of the sea-ice margin in turn appears to affect the deepening and tracking of polar lows that form over Hudson Bay (Gachon et al. 2003).

5.5 SALINITY, TEMPERATURE, AND MIXING

The distributions of salinity and water temperature in Hudson Bay and James Bay vary seasonally with changes in the freshwater runoff, ice cover, and surface heat flux (Prinsenberg 1986a). These changes are not well understood since there is no complete set of temperature-salinity transects that covers the entire area in any season, most sampling has been conducted during the open water season, and few studies have been repeated in subsequent years.

Roff and Anderson (1980a; n = 158 sites) and Prinsenberg (1986a; n = 200 sites) prepared salinity and temperature distributions for Hudson Bay using data collected in August and September 1975 (Figure 5-9). These are the most comprehensive single-season distributions available and form the basis for much of the following discussion. They are somewhat different from the distributions prepared by Barber (1967), which are composites based on data collected by various studies during the open water period in the 1950's and 1960's (Figure 5-10). The year-to-year variability of the surface salinity and temperature distributions is unknown but differences between studies suggest that it may be substantial (see Barber 1967; Anderson and Roff 1980a; Prinsenberg 1986a; Drinkwater et al. 1994). Some under-ice data are available from coastal waters near the mouths of the Eastmain, La Grande, and Grande Baleine rivers and from northern James Bay and Manitounuk Sound.

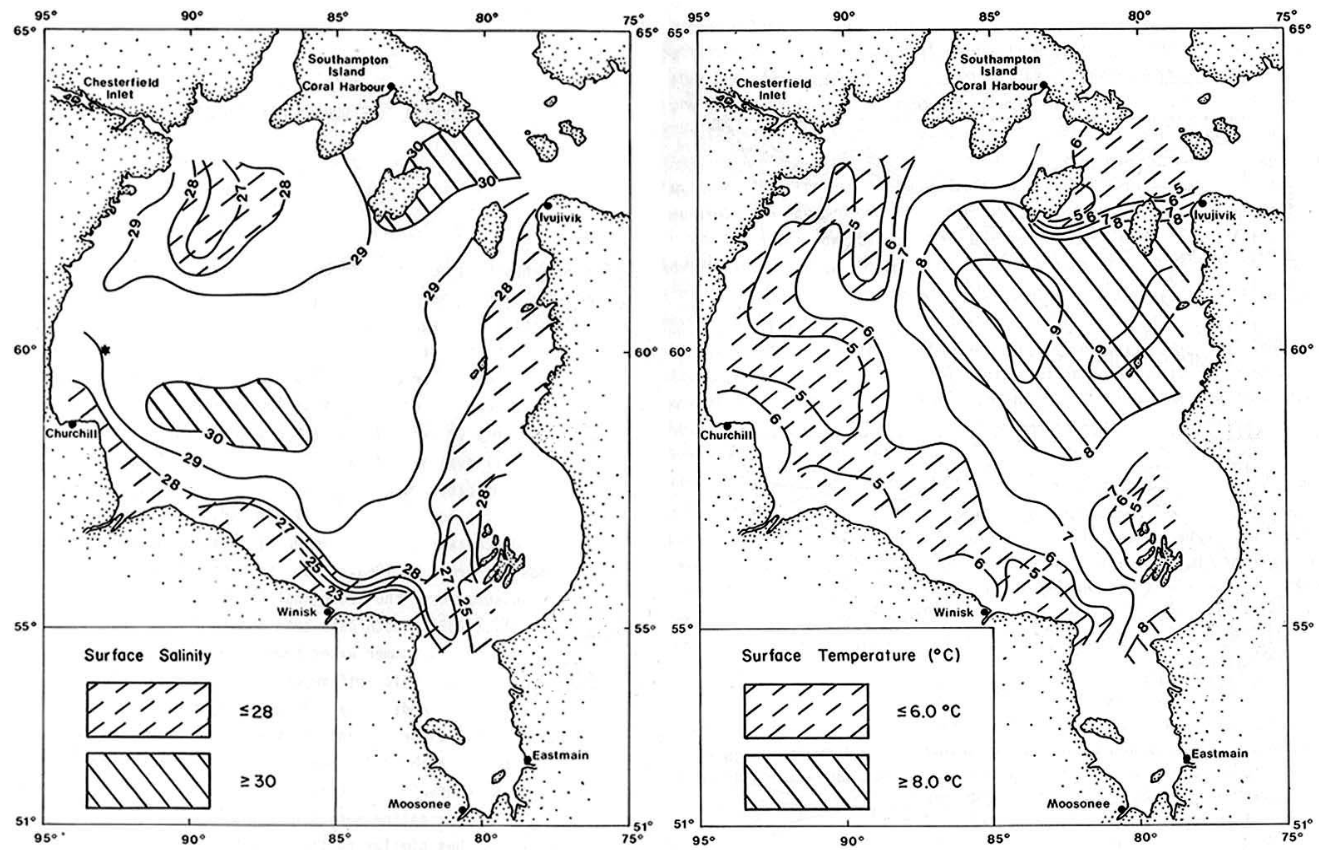

Figure 5-9. Surface salinity and temperature distribution of Hudson Bay in August-September 1975 (adapted from Prinsenberg 1986a).

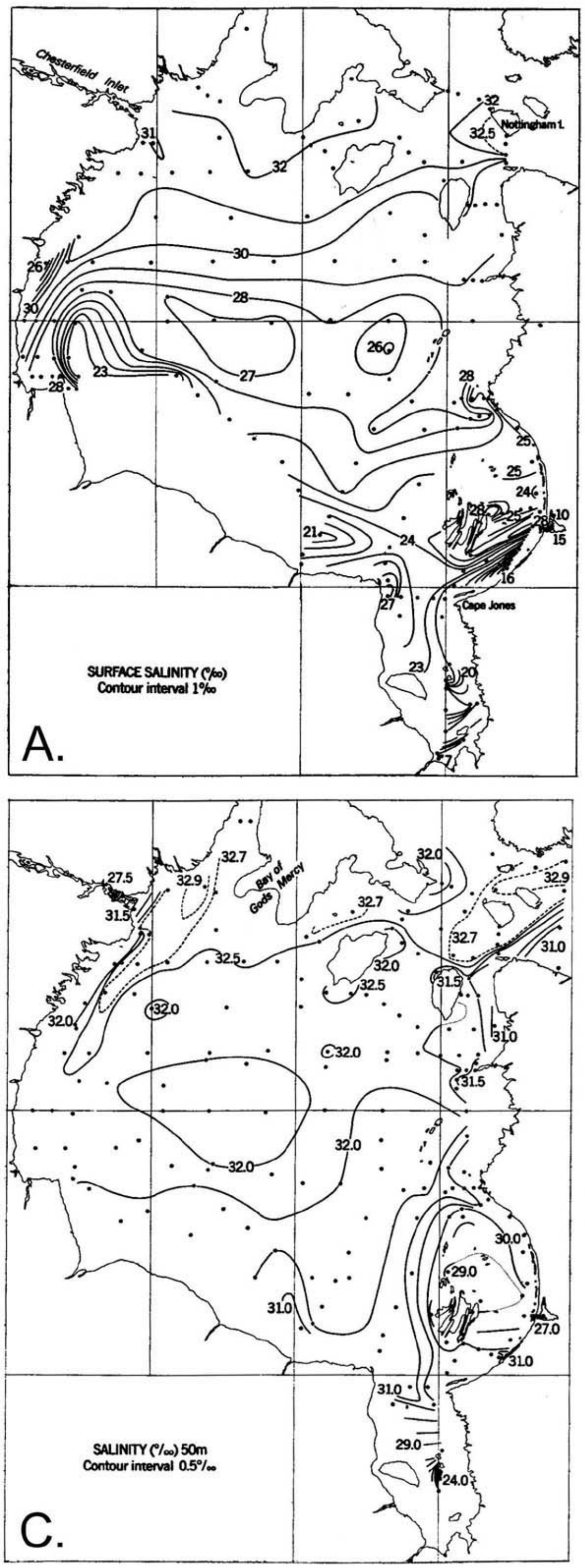

Figure 5-10. Distribution of salinity (A) and temperature (B) at the surface and salinity (C) and temperature (D) at 50 m depth from data observed during the open water season by a number of sources (adapted from Barber 1967). Dots indicate sampling stations.

5.5.1 Surface Distributions

During the summer, cold, saline surface water that enters Hudson Bay from the north is diluted by runoff and heated by solar radiation as it circulates cyclonically around Hudson Bay--some of it enters James Bay (Prinsenberg 1986a). This dilution and warming creates vertical density gradients that effectively prevent mixing of the surface and deep waters and thereby the transport of nutrients from the bottom waters into the surface waters. Strong vertical stratification is characteristic of the offshore waters of both Hudson and James bays during the summer and must limit biological productivity (Roff and Legendre 1986). In winter, lower runoff, salt rejection from the growing ice cover, and surface cooling weaken the vertical stratification and permit very slow vertical mixing. There is little coastal development or bottom relief to promote mixing or upwelling that might increase the availability of chemical nutrients in the surface waters.

In the Hudson Bay marine region, there is a general increase in temperature and salinity with distance offshore in the summer (Figure 5-9; Anderson and Roff 1980a; Prinsenberg 1986a; S. Prinsenberg, DFO, Dartmouth, NS, pers. comm.). In western Hudson Bay, the lower temperature inshore is attributed to the northwest wind, which causes upwelling and along with the strong tidal current mixing brings colder deeper water to the surface. The lower surface salinities there are due to dilution by freshwater runoff. In southern Hudson Bay the colder and low salinity inshore conditions are attributed to the pack ice that lingers there well into the summer months. In August and September 1975, salinities at 3 m depth ranged from about 23 to 30 ‰ (‰≈psu) and temperatures from 4 to 11°C. Salinities were highest offshore near the middle of Hudson Bay and between Coats and Mansel islands where the region borders on Hudson Strait. Inshore, the salinities were lower except along the west coast. The lowest salinities were found very close to shore adjacent to the Nelson River and distinct plumes were created by runoff from Chesterfield Inlet and the Churchill and Nelson rivers. Low temperatures were observed in the coastal waters of southern and western Hudson Bay and around the Belcher Islands. Higher temperatures, generally 8°C or greater, extended over a large area from the Quebec coast westward into the middle of the bay. They are attributed to solar heating of more stable surface waters.

Under the ice, salinity and temperature distributions in the Hudson Bay marine region are largely unknown. Surface distributions likely follow a similar pattern to those seen in the summer but with higher salinities and lower temperatures, and more extensive surface dilution by river plumes (Prinsenberg 1986a, 1987; Wang et al. 1994a; Ingram and Prinsenberg 1998). Below 50 m depth there should be little seasonal change in these distributions seasonally or from year to year.

Currents transport the relatively cool, low salinity surface water of southern Hudson Bay into northwestern James Bay in the summer (Figure 5-1 and Figure 5-2; Barber 1967, 1972; El-Sabh and Koutitonsky 1977; Prinsenberg 1986a). As it circulates counterclockwise around the bay the water is warmed by solar radiation and further diluted by runoff. This warming continues as the water exits to follow the eastern coast of Hudson Bay northward, and mixes with the more saline water of southeastern Hudson Bay. In the Belcher Islands area, offshore, the summer surface temperature is lower and the salinity greater than elsewhere in the James Bay marine region (Figure 5-10 and Figure 5-11). In contrast, the relatively sheltered surface water of Richmond Gulf is warmer, up to 17°C, and more dilute (Rochet and Grainger 1988). Similar high temperatures may occur locally in other sheltered areas.

The summer surface salinity values over most of James Bay and southeastern Hudson Bay are low relative to the rest of Hudson Bay, which itself is low relative to other seas. They range from less than 10 ‰ (‰≈psu) in southern James Bay and along the Hudson Bay coast south of Richmond Gulf to 28 ‰ around the Belchers, offshore the mouth of Richmond Gulf, and near Inukjuak (Figure 5-10 and Figure 5-11; Barber 1967, 1968; Prinsenberg 1986a). There is a relatively steep surface salinity gradient south and east of the Belchers where the more saline waters surrounding the Belchers meet the less saline waters exiting from James Bay. There are also steep salinity gradients in the vicinity of the major river outlets.

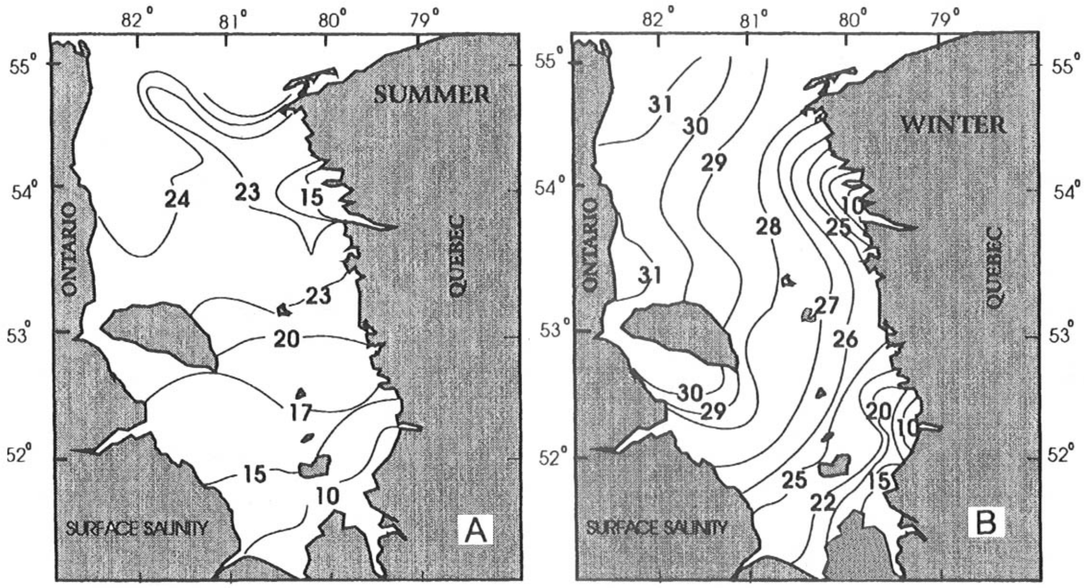

Figure 5-11. **Surface salinities in summer (A) and winter (B) in James Bay (adapted from Ingram and Prinsenberg 1998).**

Under the ice, the surface salinity pattern in James Bay resembles that in the summer but with higher values due to salt rejection due to ice formation and decreased runoff (Figure 5-11; Peck 1976; Prinsenberg 1986a). During the winter, surface water enters from Hudson Bay with a salinity of 31 ‰ (‰≈ psu), is diluted by runoff as it circulates around the bay, and leaves with a salinity of 28 ‰. In winter, the lowest offshore surface salinity values are still found off the major river systems. In late spring and early summer, the surface water temperature distribution is patchy, with pockets of warm water where ice cover has dissipated and water temperatures near freezing where it has not (Barber 1967).

Extensive freshwater plumes with steep surface salinity gradients are observed off river mouths in Hudson Bay and James Bay year-round (Figure 5-11; Prinsenberg and Freeman 1986). There is considerable variation in these gradients depending upon the runoff volume (Ingram et al. 1986) and other factors such as ice cover (Ingram 1981, 1982) and the tides (Baker 1989; Lawrence 1996). The expansion and compression of the estuarine plume with each low and high tide creates an environment of rapid fluctuations in depth, salinity, temperature, and current to which sedentary estuarine biota must be adapted (Baker 1989).

In winter, ice cover inhibits wind-induced mixing, leaving just tidal current mixing, and allows the plumes to spread further and deeper than under the ice-free conditions of summer, despite runoff rates that can be an order of magnitude lower (Ingram 1981, 1982; Freeman et al. 1982; Ingram and Larouche 1987a+b; Lepage and Ingram 1991; Messier and Anctil 1996). The Nastapoka River plume, for example, is about 40 times larger when there is ice cover than when there is not, despite river runoff that is 2.5 times smaller (Messier and Anctil 1996). Tidal dissipation is the main source of plume mixing during the period of ice cover. This mixing is slow under the relatively smooth landfast ice and faster offshore the ice edge under the rougher pack ice or in leads. Some plumes, such as that from the La Grande River, remain coherent up to 100 km from their source and reach widths of 20 to 30 km (Figure 5-12; Messier et al. 1986, 1989). Under the Coriolis force all currents and plumes are deflected to the right causing them to move counterclockwise along the coast.

The effects of high runoff are most pronounced along the south coast of Hudson Bay east of the Nelson River, in eastern James Bay and along the south coast of the Hudson Bay Arc and, perhaps, in Richmond Gulf. While hydroelectric developments on the Churchill/Nelson and Eastmain/La Grande rivers may not be altering the overall runoff, they are changing its timing and spatial distribution. The effects of these developments on the plumes are discussed further in Section 15.1

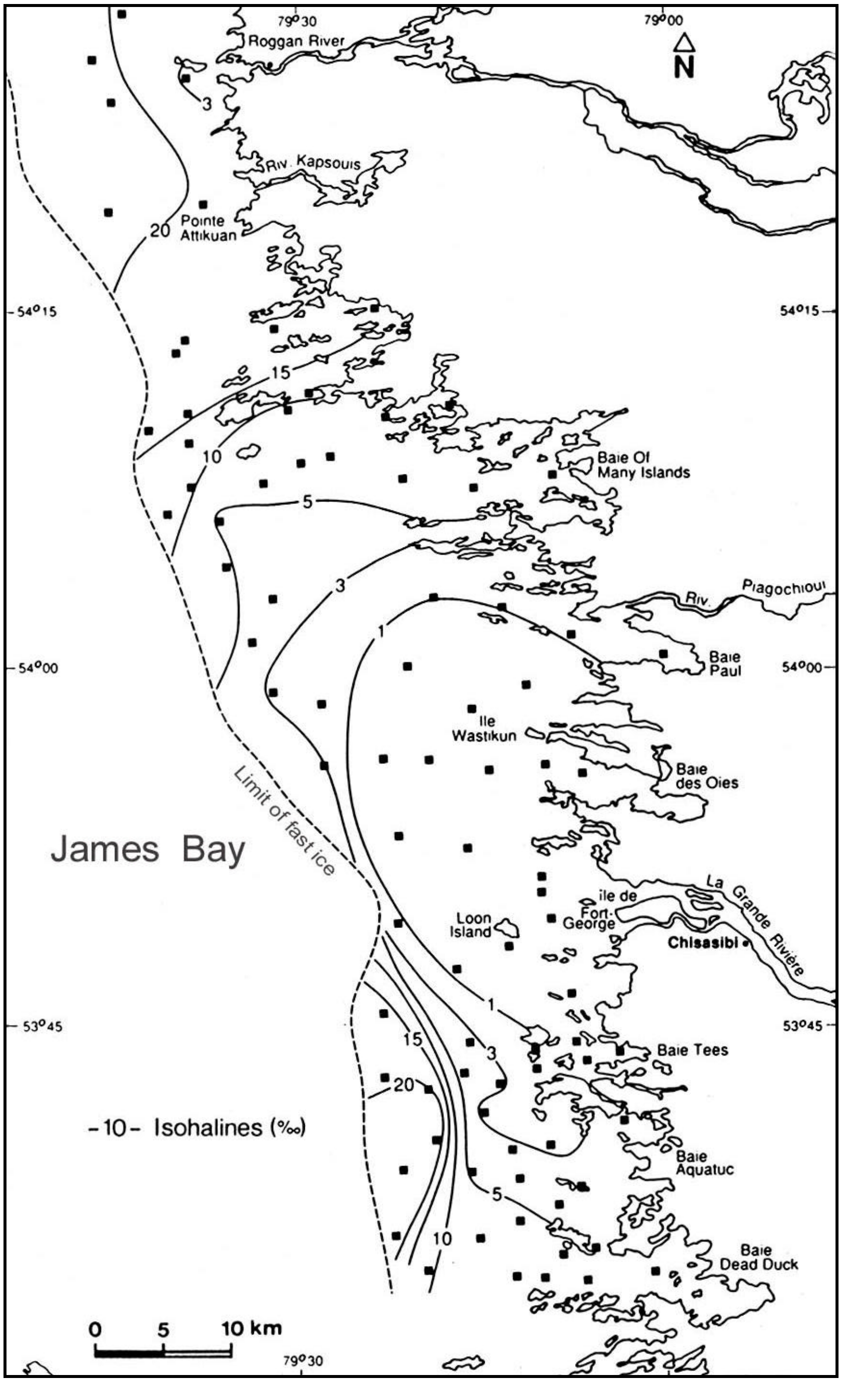

Figure 5-12. Surface isohalines of La Grande River plume between 20 February and 2 March 1987 (adapted from Messier et al. 1989, p. 280).

5.5.2 Vertical Profiles

There is a seasonal cycle to the vertical temperature and salinity distributions in Hudson Bay and James Bay (Barber 1967, 1968; El-Sabh and Koutitonsky 1977; Prinsenberg 1982b, 1986a, 1991; Jones and Anderson 1994; Wang et al. 1994a; Ingram and Prinsenberg 1998). In general terms, it involves the dilution of surface waters each spring and summer by fresh water inputs from melting ice and runoff. There is subsequent downward mixing with reduction of runoff and salt rejection from the growing pack ice in fall and winter. Seasonal heating and cooling, tides, and bathymetry contribute to this cycle, which is illustrated in Figure 5-13.

The specifics of this seasonal cycle are not well known, but it likely proceeds as follows (Barber 1967, 1968; El-Sabh and Koutitonsky 1977; Prinsenberg 1982b, 1983, 1986a, 1991; Jones and Anderson 1994; Wang et al. 1994a; Ingram and Prinsenberg 1998). In the spring, runoff and melting ice create a thin layer of low salinity water immediately beneath the ice cover. There is a strong density gradient where this water layer meets the saltier seawater below. This pycnocline limits mixing of the surface and bottom water layers and stabilizes conditions in the water column. As the season progresses, tidal mixing entrains seawater from below and the under-ice surface layer gradually deepens (Figure 5-14 and Figure 5-15; Prinsenberg 1987; Lepage and Ingram 1991). It is mixed by the wind and warmed by the sun once the ice cover disappears but remains relatively shallow (<20 m) through August. The stability of the water column increases through the summer and into September. In late summer and fall, cooler temperatures, lower runoff and stronger winds cause the surface layer to deepen rapidly. When ice begins to form it releases salt into the upper water that further weakens the density gradients and destabilizes the water column. Through the winter the surface layer grows progressively deeper and more saline, with mixing to a depth of 60 to 100 m (Pett and Roff 1982; Prinsenberg 1986a; Jones and Anderson 1994). However, below a depth of 50 m these seasonal changes appear to be small. When the cycle repeats itself in the spring, the weak salinity gradient that separates the surface and bottom water layers is slowly eliminated by vertical diffusion (Prinsenberg 1986a).

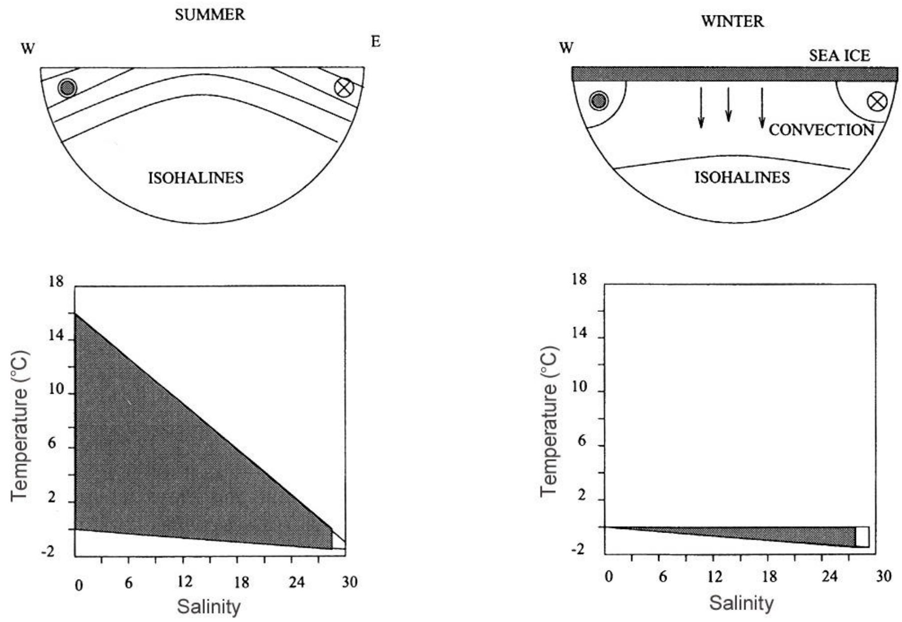

Figure 5-13. **Schematic diagram of seasonal differences in the temperature-salinity relations in Hudson Bay (adapted from Ingram and Prinsenberg 1998). Similar current directions are indicated; however, surface flows are stronger in summer than winter. Summer stratification is characterized by a warm and fresher shallow upper layer.**

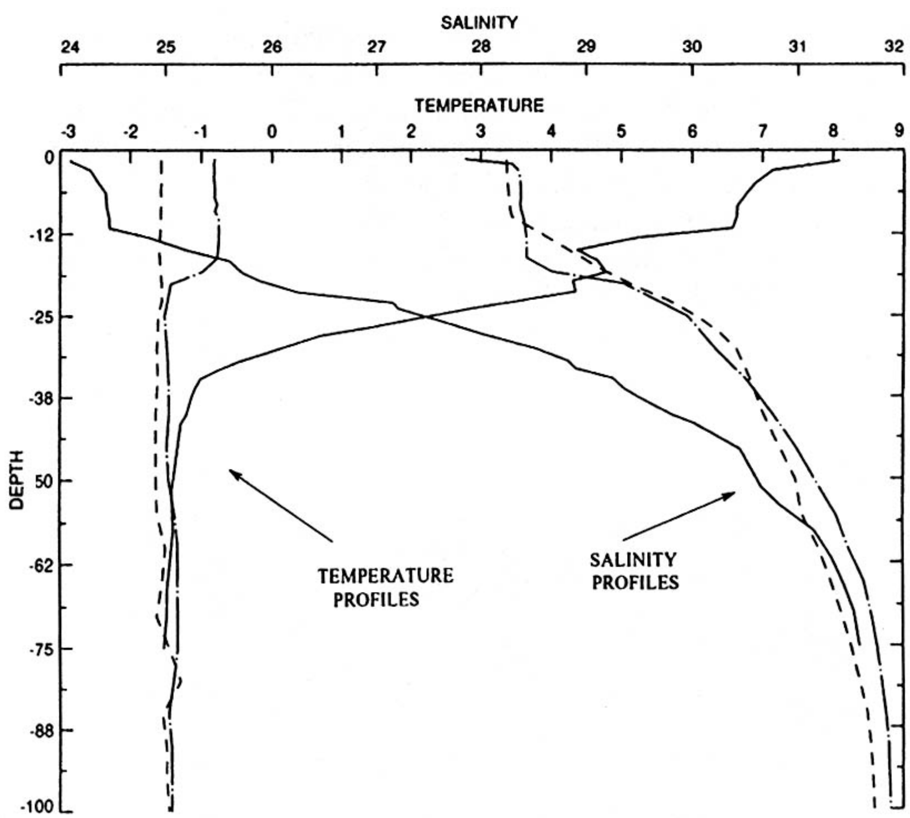

Figure 5-14. **Representative vertical profiles of temperature and salinity in southeastern Hudson Bay at various times of the year (different years); April 15, 1982 (dashed line), May 16, 1982 (dashed-dotted line), August 15, 1976 (solid line) (from Ingram and Prinsenberg 1998, p. 851).**

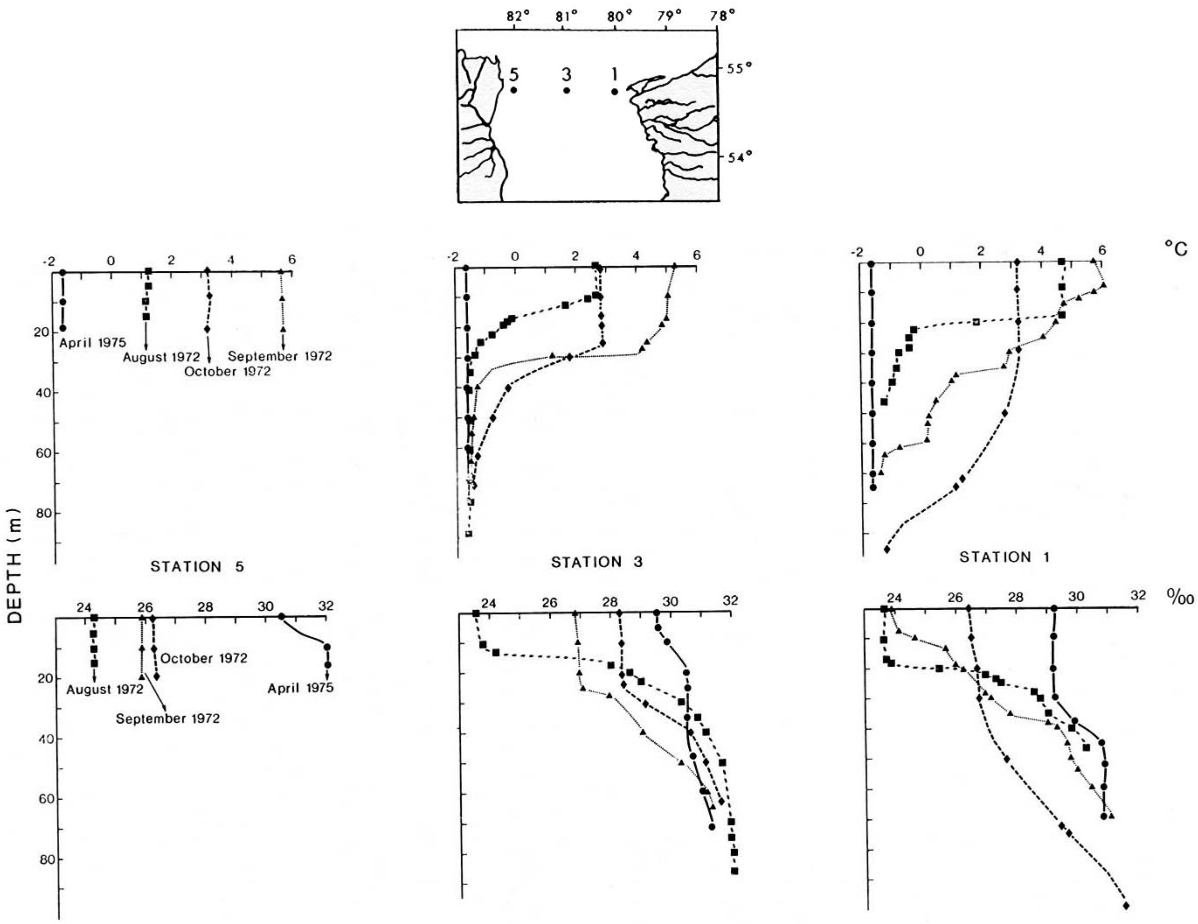

Figure 5-15. Seasonal vertical profiles of temperature (top) and salinity (bottom) at three oceanographic stations across the entrance of James Bay (see inset) (adapted from El-Sabh and Koutitonsky 1977).

During the summer months there is a strong pycnocline at 15 to 25 m that effectively prevents vertical exchange between surface and deep waters of Hudson Bay (Anderson and Roff 1980a; Prinsenberg 1986a). This strong vertical stratification is characteristic of Hudson Bay waters in summer. The water becomes progressively colder and more saline with depth, approaching the same deep water type at about 100 m where the mean temperature is less than -1.4^{o}C and salinity greater than 33 ‰ (‰≈psu) (Figure 5-16; Hachey 1933; Barber and Glennie 1964; Barber 1967, 1968; Pelletier et al. 1968; Anderson and Roff 1980a; Prinsenberg 1986a; Drinkwater et al. 1991; Jones and Anderson 1994; Simard et al. 1996). The deep water layer in James Bay is subject to considerable seasonal and interannual variation in temperature and salinity, due in part to the relative shallowness of the bay (El-Sabh and Koutitonsky 1977). Limited sampling suggests that the deep water in Richmond Gulf may be less saline than that found elsewhere in the region (27 ‰ at 90 m; Rochet and Grainger 1988). Mixing of water below the pycnocline is much slower than that above, with a turnover time of 3 to 14 years—longest in the deeper water of central Hudson Bay (Barber 1967; Pett and Roff 1982; Roff and Legendre 1986). Below 130 m—the sill depth between Foxe Basin and Hudson Strait, the Hudson Bay bottom water receives higher salinity overflow from Foxe Basin (Jones and Anderson 1994).

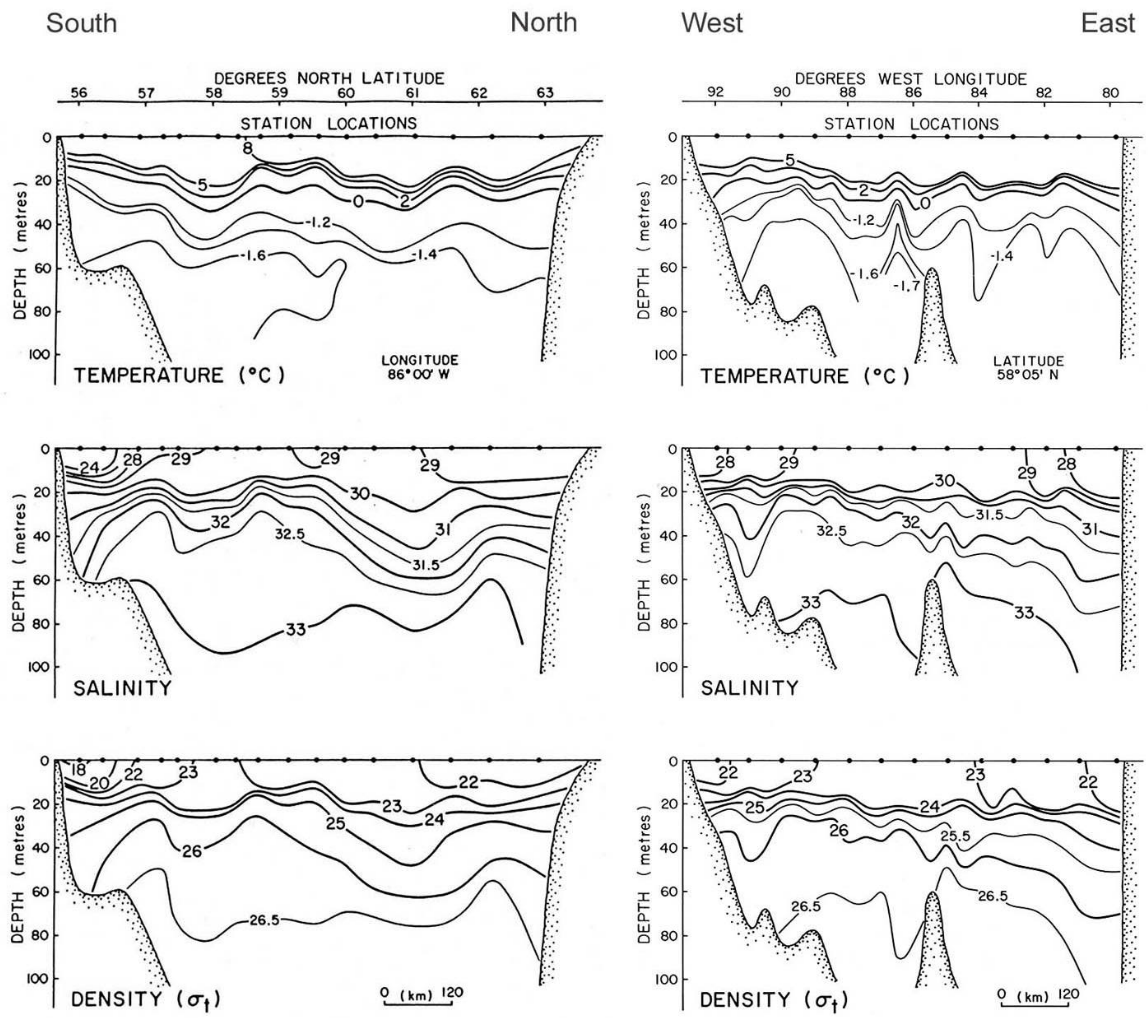

Figure 5-16. **South to north (86°00' W) and west to east (58°55'N) temperature, salinity and density sections from Hudson Bay during the summer of 1975 (from Prinsenberg 1986a, p. 166 and 168).**

Depth and strength of the pycnocline change with the seasons (Prinsenberg 1986a). Profile data from the centre of Hudson Bay show a gradual deepening of the pyncoline as the summer season progresses (Figure 5-17). On 4 August 1975, after the large freshwater input of the ice melt and runoff and during peak solar heating, the pycnoline was observed at 13 m--above that the water was relatively warm and dilute. As summer progressed the pycnocline deepened in response to increasing wind stress and decreasing inputs of heat and runoff. By 28 September it was at 28 m, above which the water was very well mixed. While surface temperature decreased over the summer, the surface layer heat content actually increased, and the highest surface salinity appeared to occur in mid-summer. Changes to the vertical profiles during the rest of the year in Hudson Bay have not been well documented. Data from a year-round monitoring site for temperature and salinity at depths of 18.5, 53.5, and 93.5 m northwest of Churchill show that the pycnocline does deepen due to the fall cooling and salt rejection from the growing pack ice throughout the winter and can reach a depth of 95 m (Figure 5-18)

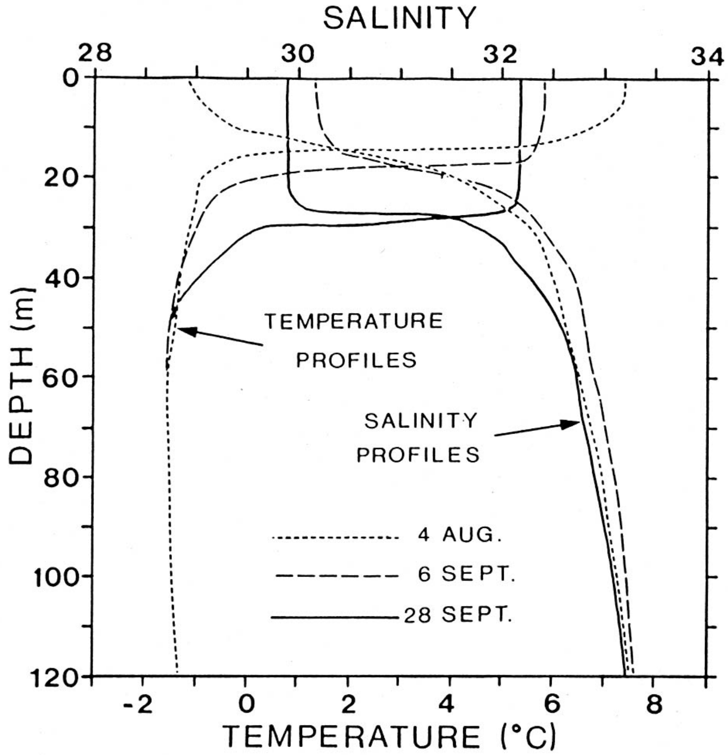

Figure 5-17. Profile data from the center of Hudson Bay (1975) (from Prinsenberg 1986a, p. 173).

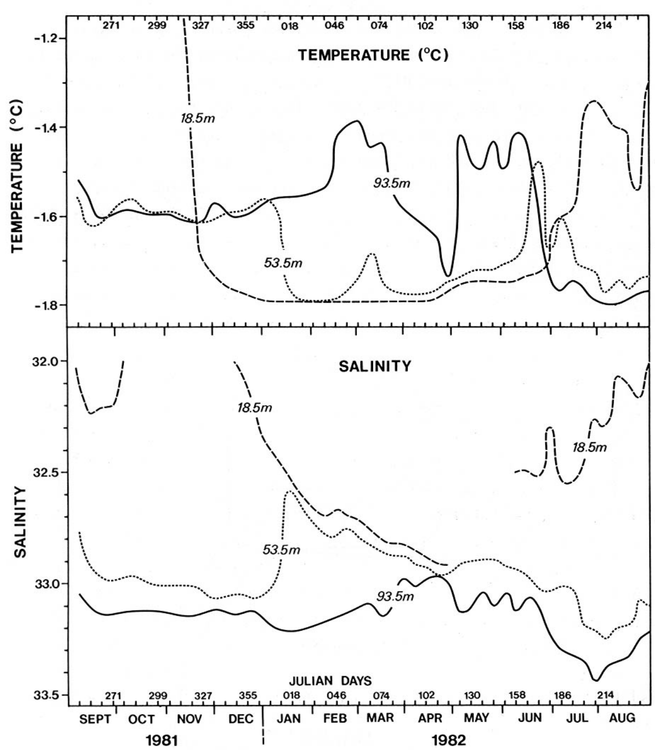

Figure 5-18. Weekly averaged temperature and salinity values as measured by current meters moored 150 km northeast of Churchill at 18.5 m, 53.5 m, and 93.5 m depths (from Prinsenberg 1986a, p. 174).

Summer and winter measurements have been taken across northern and central James Bay (Figure 5-19 and Figure 5-20). They reveal seasonal conditions that are similar to those described above but modified by bottom topography and runoff. Strong vertical density stratification is characteristic of the water column in summer. Depending upon the date of observation and on weather and bottom conditions, there is a well-defined surface layer that ranges in depth from 15 to 30 m (e.g., Barber and Glennie 1964; Barber 1967, 1972; Prinsenberg 1976; El-Sabh and Koutitonsky 1977; Rochet and Grainger 1988). It is separated from the deep-water layer by a pycnocline that is weaker along the west coast of the bay, where there is greater tidal mixing over the shallow topography and less runoff (El-Sabh and Koutitonsky 1977; Prinsenberg 1986a). In March, the waters of northwestern James Bay are relatively saline (>30 ‰; ‰≈psu) and unstratified, while those of northeastern James Bay are less saline (about 27-31 ‰) and in some areas show a halocline (Peck 1976). Across central James Bay the salinities decrease progressively from over 30 ‰ along the west coast to less than 27 ‰ along the east coast, with little vertical stratification apparent. The salinity profile changes as a result of winter mixing and could give salinities close to 33 in near surface waters (Prinsenberg 1988b).

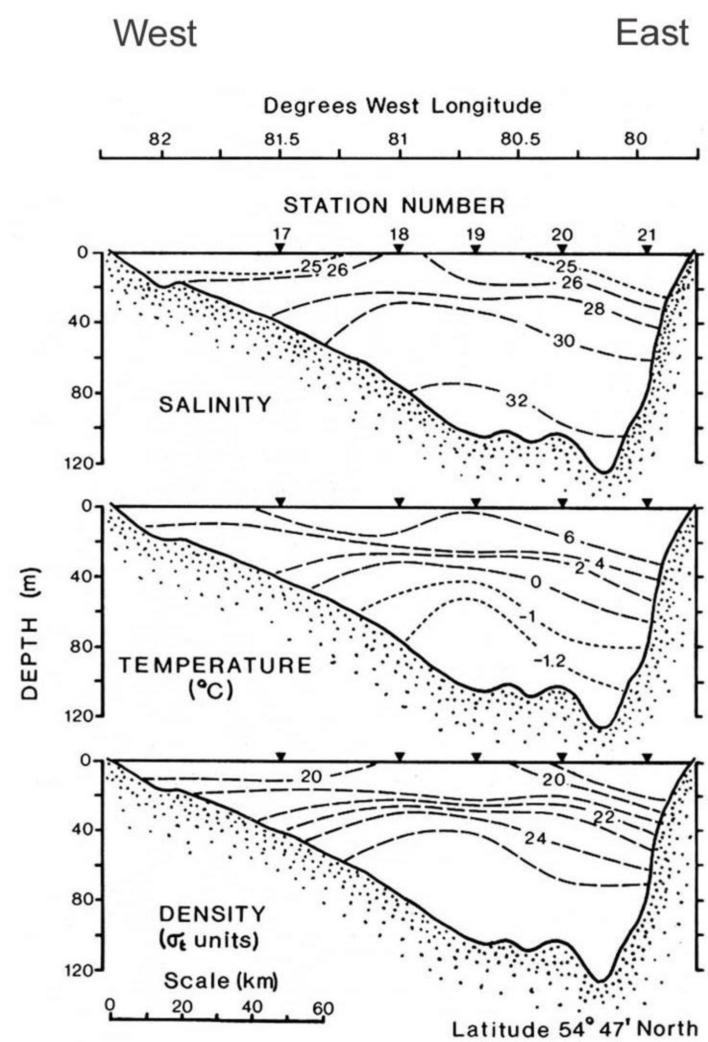

Figure 5-19. Salinity, temperature, and density sections from the entrance of James Bay along 54°47'N latitude on September 16, 1975 (adapted from Prinsenberg 1986a, p. 170).

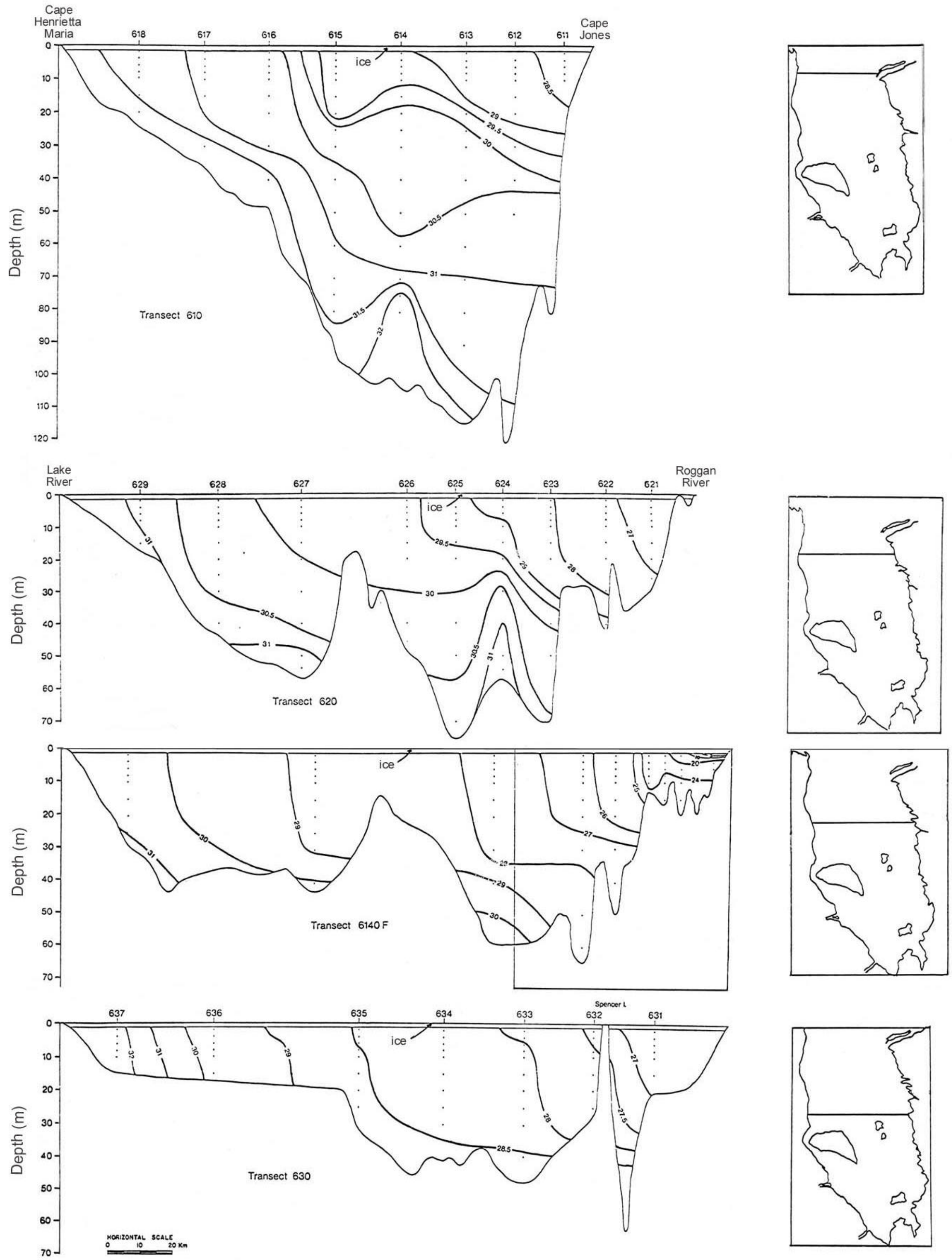

Figure 5-20. Vertical salinity transects across northern and central James Bay in March 1976 (adapted from Peck 1976, pages 136-139).

Vertical stratification in the vicinity of the large river plumes varies seasonally and with distance from the river mouth (Legendre et al. 1981; Ingram and Larouche 1987a+b; Ingram et al. 1989; Lepage and Ingram 1991). It is directly related to the runoff volume and inversely related to the tidal kinetic energy (Freeman et el. 1982). The vertical temperature and salinity gradients weaken with increasing distance from the river mouth. This weakening is most pronounced when wind mixing is greatest, during the open water period and in winter outside the fast ice.

Fortnightly tidal variations can have a greater effect on vertical mixing and plume dynamics than changes in runoff (Lepage and Ingram 1986; Messier et al. 1989). This is particularly apparent in Rupert Bay, where the tidal amplitude to water depth ratio is large (0.625) and intense mixing in the lower two-thirds of the estuary causes vertically homogeneous conditions to prevail (Veilleux et al. 1992). The Nelson estuary is also subject to intense tidal mixing such that its water column is vertically homogenous except in the deep central channel (Baker 1989; Baker et al. 1993). The La Grande and Grande rivière de la Baleine estuaries are less affected by tidal mixing and better stratified.

Under the landfast ice, the upper 10 m of the water column in the area of the Grande rivière de la Baleine plume is highly stratified (Figure 5-21; Ingram and Larouche 1987b; Lepage and Ingram 1991). With the onset of ice breakup these salinity gradients are strengthened by the addition of a 2 m layer of fresh meltwater and runoff between the ice and more saline seawater. These stratified conditions persist, with a strong pycnocline between 2 and 4 m depth, until the ice has decayed sufficiently to allow movement of the flows and increase turbulent mixing by the tides and winds. This mixing destroys the shallow pycnocline and collapses the river plume to an area comparable to its open water dimensions. Strong vertical stratification continues to prevail under open water conditions in the reduced area of the plume.

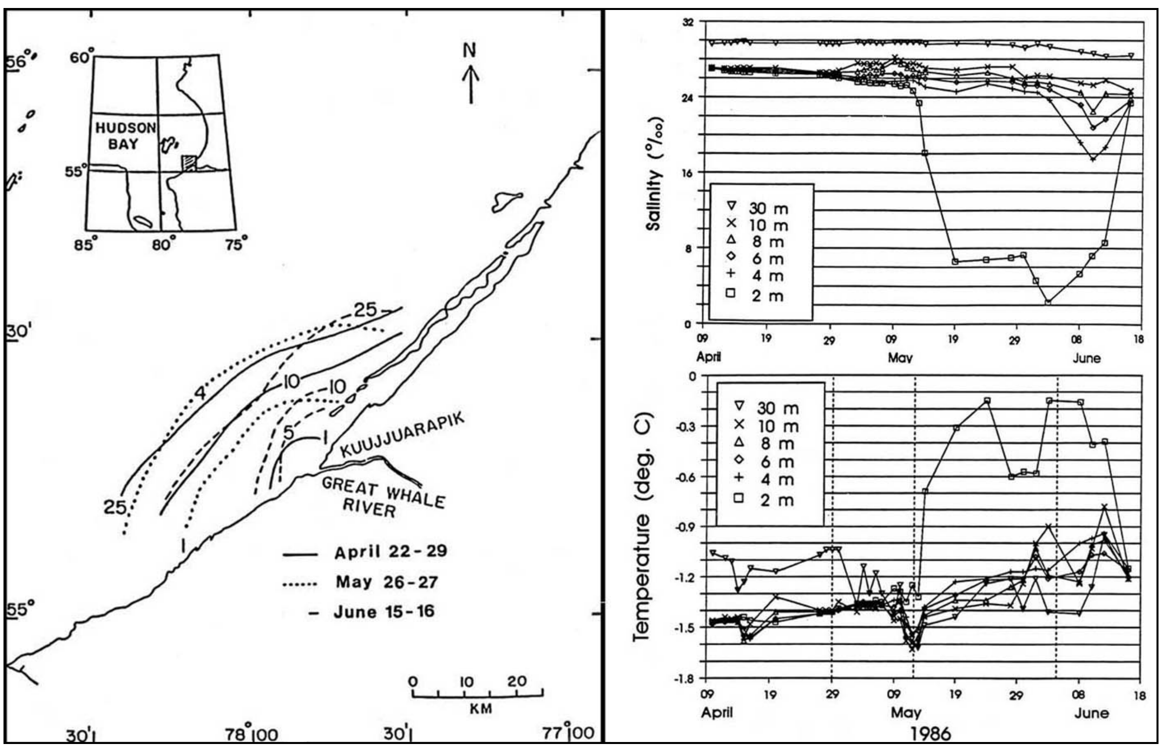

Figure 5-21. Surface isohalines and salinity and termperature profiles offshore Grande rivière de la Baleine in the spring of 1986 (adapted from Lepage and Ingram 1991, p. 12,714-5).

The strong pycnoclines in summer, and under the ice near river mouths, slow vertical mixing between surface and deep water layers (Hachey 1933, 1954; Barber 1967, 1968; Roff and Legendre 1986). This vertical stability of the water column is a key factor limiting the concentrations of nutrients that are available to phytoplankton in the surface waters of Hudson Bay during the summer, particularly nitrates (Harvey et al. 1997; Figure 5-22). The historical presence of a large population of bowhead (Ross 1974; Reeves et al. 1983) strongly suggests that there is an area of high productivity and therefore vertical mixing in northwestern Hudson Bay. However, its existence has not been proven.

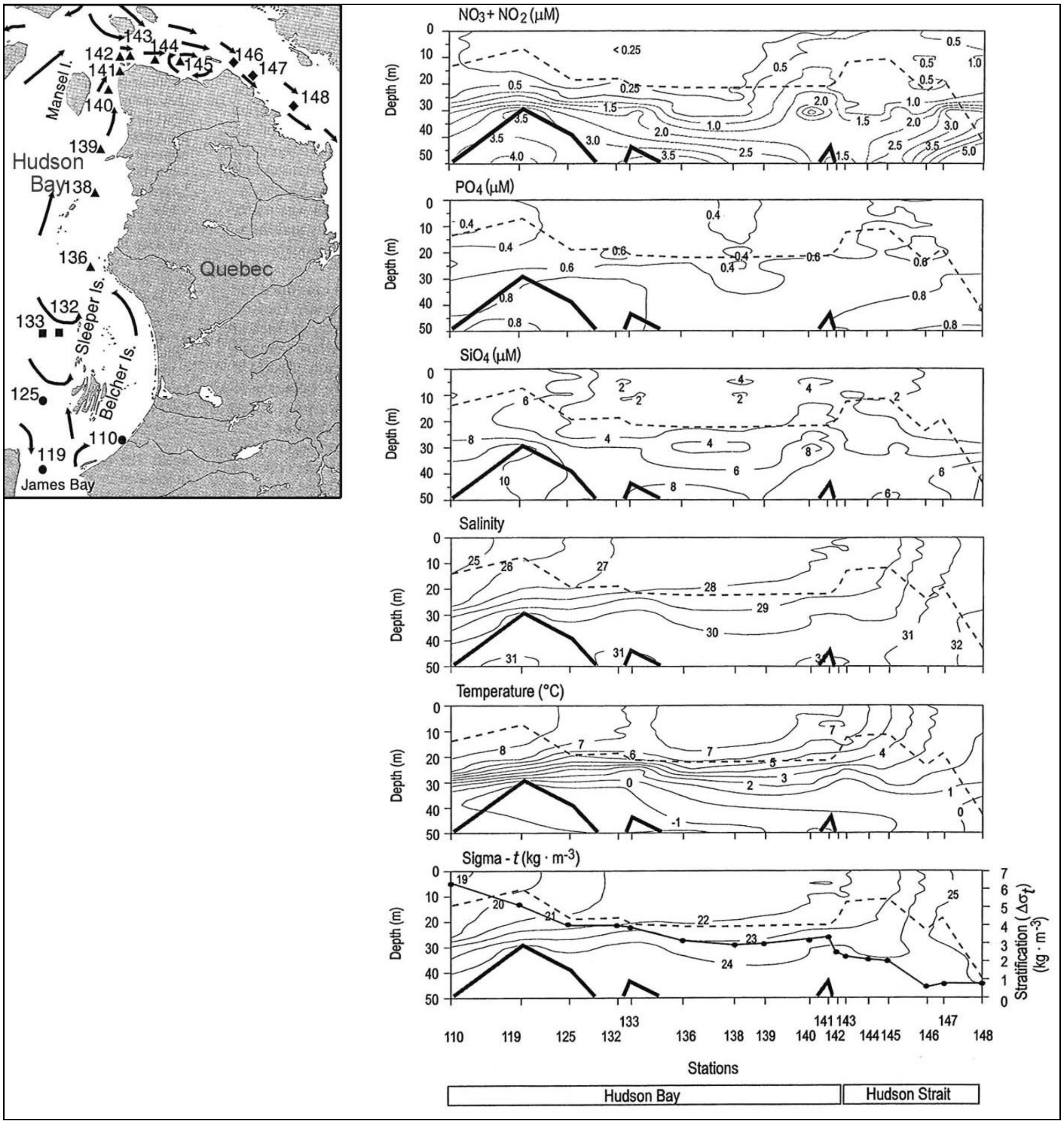

Figure 5-22. **Horizontal and vertical distribution of nitrates, phosphates, and silicates in relation to salinity, temperature, and density (σ_t) in the upper 50 m of the water column along a sampling transect in Hudson Bay and Hudson Strait in September 1993 (adapted from Harvey et al. 1997). Also shown are: the depth of the mixed layer (----), which was defined as the depth where the temperature was 1°C less than at the surface; the lower depth limit of the pycnocline (—); and an index of vertical stratification (●—●), which was defined as the difference in density between the bottom and surface layers ($\Delta\sigma_t$). The map symbols indicate distinct phytoplankton assemblages that were identified by cluster analysis (● group A; ■ group B, ▲ group C, ◆ group D).**

Prinsenberg (1991) summarized the importance of the vertical stability of the water column on the biological production as follows:

> *"The vertical stability of the water column is important in determining the magnitude of the vertical nutrient flux required for sustaining surface layer ice algae and phytoplankton spring blooms. Vertical nutrient fluxes are proportional to turbulent current energy and inversely proportional to the density stratification. High production of ice algae and phytoplankton only occur in stable stratification for a short period as nutrients are quickly depleted. Sustained production requires a constant or intermittent nutrient flux without increasing the mixed layer depths beyond the euphotic zone for too long (Gosselin et al. 1985). In the plumes studied in the Hudson and James bays, high biological activity occurs in areas where the vertical nutrient flux is maintained intermittently by tidal mixing at the time when tidal currents reach their maximum amplitude (Gosselin et al. 1985)."*

5.6 WATER CLARITY AND QUALITY

Water clarity as measured by Secchi disk depth during the navigation period in 1959 and 1961, is shown in Figure 5-23. Secchi values ranged from 1.2 m at the head of James Bay to 10 m at the mouth, with a sharp increase in Hudson Bay (Barber 1967; Kranck and Ruffman 1982). The water in Hudson Bay is relatively clear. In August-September 1975, Secchi disk measurements of surface water clarity in the bay averaged 18.2 m offshore and were generally 11-12 m inshore (Anderson and Roff 1980b). Clear photographs of bottom sediments and biota in Barber (1968), Grainger (1968), and Barber et al. (1981) also demonstrate the clarity of some deep bottom waters. In the southern and western areas of James Bay, which are shallow and receive a great deal of sediment laden runoff, the water clarity is low relative to other parts of the marine ecosystem and to other Arctic marine regions generally. There is a progressive increase in water clarity moving northward and also a general increase in water clarity moving offshore.

Clarity of the surface water depends on many factors, particularly those that affect sediment inputs and phytoplankton production, and will vary with location and season. Inshore water clarity is likely least during periods of high runoff, strong onshore winds, and peak primary productivity. Bank erosion during periods of high runoff and from new diversion channels contributes large quantities of sediment to the James Bay marine region (Ingram et al. 1986).

In September 1974, water at the mouth of the La Grande River was turbid, with Secchi depths ranging from <1 m in the river mouth to 2.6 m in the estuary (Grainger and McSween 1976). At the river mouth 1% of the light incident at the surface penetrated less than 2 m into the water, whereas about 8 km offshore it penetrated 7 m. In September 1976, Secchi disk readings at Manitounuk Sound decreased from 10.8 m at the mouth, to 5.9 m at the head of the sound (Legendre and Simard 1979). Water in the Belcher Islands area is clear and appears to be relatively free of suspended sediments. Grainger (1982) estimated that 1% of the surface illumination reached depths of 26 to 44 m during the summers of 1958-59.

During the open water season, high turbidity prevails in Rupert Bay, with a pronounced streakiness in the flow direction and stable fronts at the boundaries of the river plumes (d'Anglejan 1980). In July 1976, suspended sediments averaged 50 $mg{\cdot}L^{-1}$ throughout the water column, with values as high as 200 $mg{\cdot}L^{-1}$ at some stations, and near-bottom levels that may be much higher. During the period of ice cover the suspended sediments were much lower, averaging <10 $mg{\cdot}L^{-1}$, in March 1977. Because of the large number of islands and turbid waters, Rupert Bay is ideal for flow visualization in a natural setting (Ingram and Chu 1987). Turbidity plumes also show clearly the reversing tidal currents in Akimiski Strait (Martini and Grinham 1984).

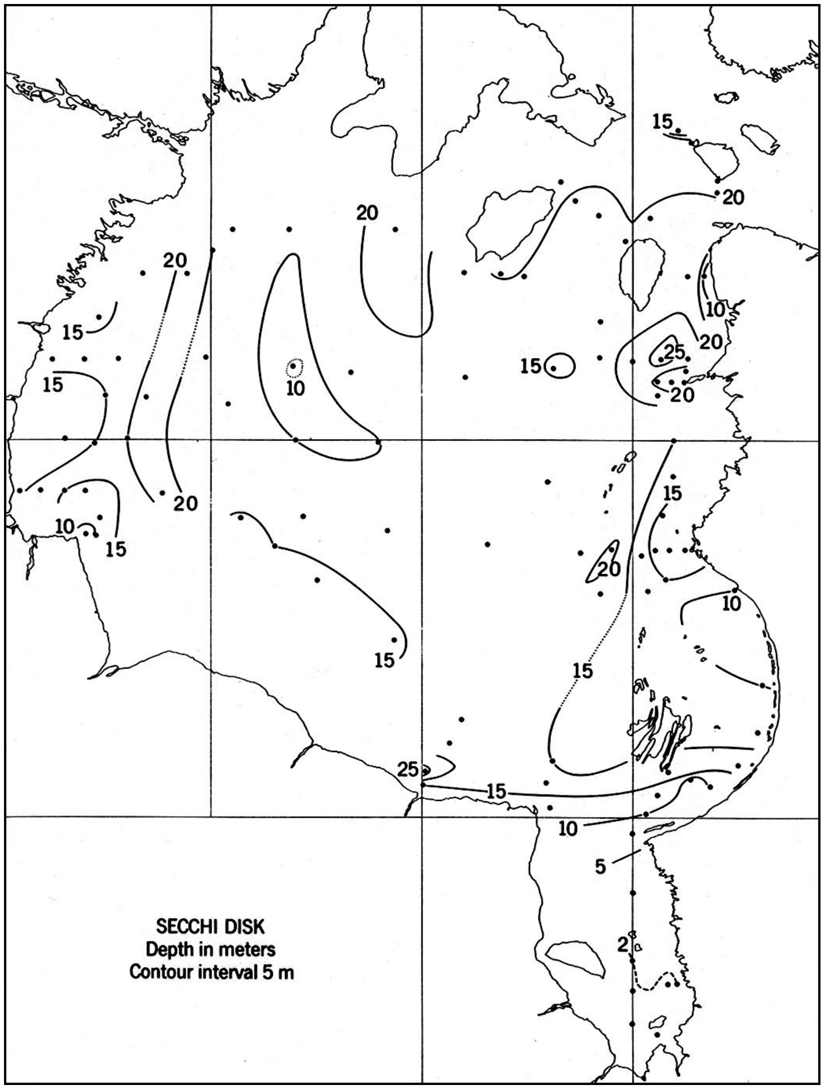

Figure 5-23. Water clarity in Hudson Bay and James Bay based on Secchi disk observations taken during the navigation period in 1959 and 1961 (from Barber 1972, p. 42). Dots indicate measurement locations, the dashed contour line is in addition to the regular contour interval, and dotted contours indicate a doubtful interpretation.

5.7 BIOLOGICAL PRODUCTIVITY

Nutrient chemistry, biomass, and productivity of the Hudson Bay marine ecosystem are not well-documented. Anderson and Roff (1980a) conducted the only bay-wide survey of pelagic biomass and it only covered the ice-free summer months in a semi-synoptic survey. Shallow marine waters can be biologically productive if their surface waters receive nutrients from the seafloor. This does not appear to be the case with Hudson Bay, which has a larger area of shallow water than the Gulf of St. Lawrence and Grand Banks combined (780,000 cf. 519,000 km^2 less than 200 m; Pilson and Seitzinger 1996). Several factors, particularly the strong density stratification and relatively low nutrient concentrations in surface waters, suggest that the area is unproductive. Based on extremely limited data, the average annual productivity appears to be low and comparable to that of other seasonally ice-free Arctic marine waters (Roff and Legendre 1986). It appears to be greatest in coastal waters, particularly at embayments and estuaries, and near islands where there is periodic entrainment or upwelling of deeper, nutrient-rich water (Anderson and Roff 1980a; Legendre et al. 1982; Percy

1992). However, the likelihood that maximum primary production occurs in late spring has yet to be measured (Dunbar 1982), and the historical presence of many bowhead and beluga whales suggest that production in some areas may be greater than expected.

The structure of the food web in the Hudson Bay marine ecosystem is not well known, nor is the flow of energy through that web. Inuit and Cree have identified some of the species' interrelationships, based on their observations of predation and stomach contents (Figure 5-24); scientific studies have yet to provide a broad overview of this food web and its energy flow.

5.7.1 Plants

Phytoplankton, ice algae, benthic algae and benthic macrophytes contribute to primary production in the marine ecosystem. Their contributions are difficult to measure. One of the main obstacles to determining the annual primary production is the difficulty of sampling at breakup when the main phytoplankton bloom likely occurs.

Incomplete vertical mixing and the resultant low regeneration rates of nutrients, particularly nitrogen, appear to limit primary production in Hudson Bay (Dunbar 1970; Pett and Roff 1982; Drinkwater and Jones 1987). Deep water mixing and freshwater runoff are important sources of nitrate and total nitrogen, with atmospheric deposition accounting for about 10% of the inputs (Pett and Roff 1982; Roff and Legendre 1986). Nitrate and phosphate levels in surface waters are very low during the summer (see also Grainger and McSween 1976) and relatively high in sea ice and snow cover. Melting snow and ice cover may be an important nutrient source during the spring phytoplankton bloom (Freeman et al. 1982).

Freshwater runoff affects the primary productivity negatively by increasing vertical stability of the water column, and positively through nutrient additions--either direct or due to deep-water entrainment (Anderson and Roff 1980a; Pett and Roff 1982). While river runoff carries large quantities of carbon and nutrients into the marine ecosystem, particularly during ice-breakup, the river waters are less concentrated in nutrients than Hudson Bay coastal waters (Grainger and McSween 1976; Roff et al. 1980; Freeman et al. 1982; SEBJ 1990; Schneider-Vieira et al. 1993; Hudon et al. 1996; Figure 5-25). Overall, the concentration of dissolved organic carbon in waters of rivers flowing through tundra to Hudson Bay is about half that of rivers flowing through forested basins to the Gulf of St. Lawrence (Hudon et al. 1996). The annual exportation of particulate inorganic matter (PIM), particulate organic matter (POM), and dissolved organic carbon (DOC) to Hudson Bay from the Great Whale River were estimated at 135,000, 21,000, and 90,000 t, respectively in 1990-91. Concentrations were positively related to discharge. Of three size-fractions of particulate organic matter examined, the finest (0.7-53 μm) exhibited the highest heterotrophic activity, the lowest C:N ratio, and the highest chlorophyll ***a*** concentration (1 $\mu g \cdot L^{-1}$) during the summer. Primary production by phytoplankton in the estuaries of Hudson Bay may be limited by low nutrient concentrations and turbulence, which mixes cells out of the photic zone (Schneider-Vieira et al. 1993).

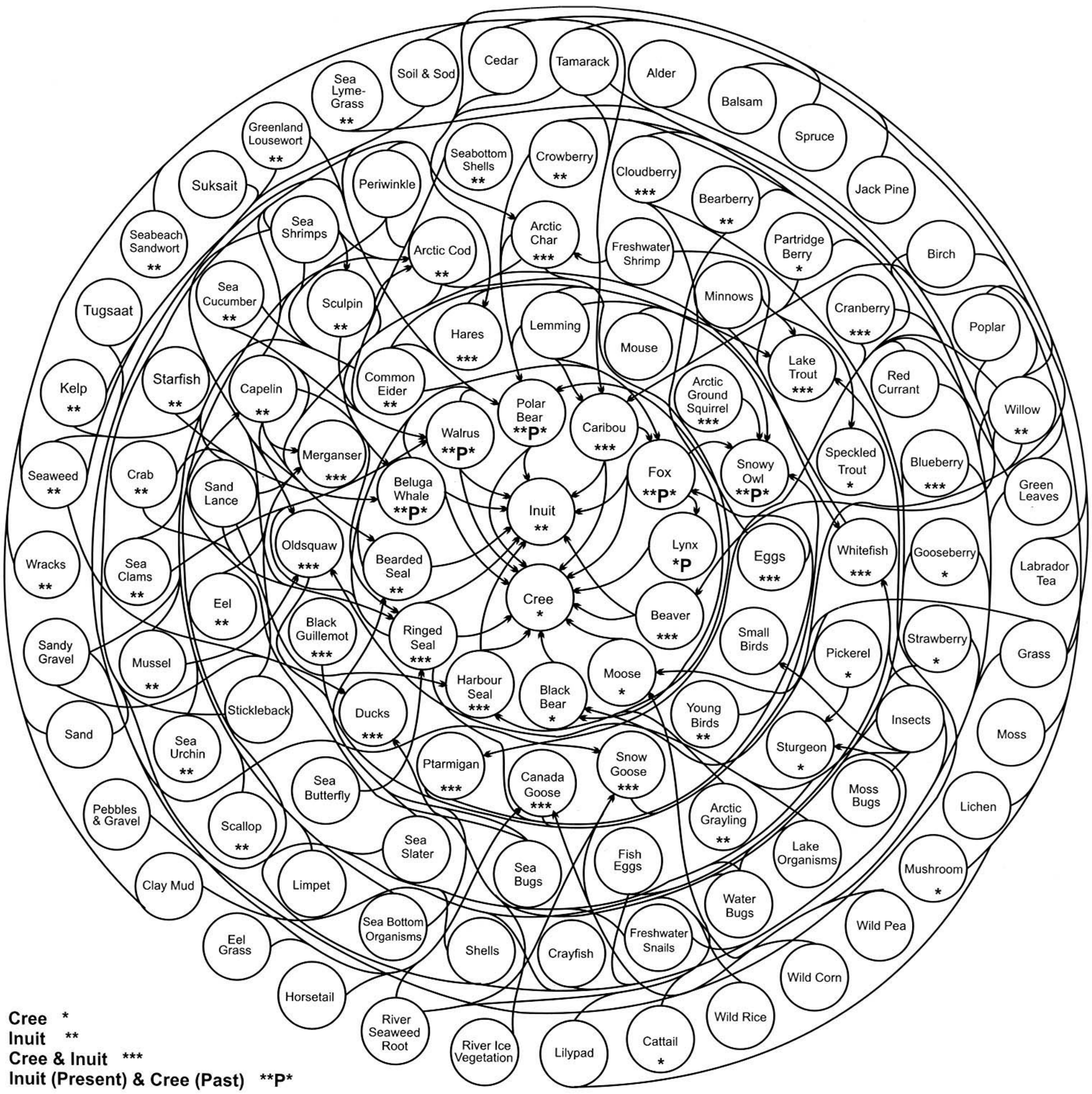

Figure 5-24. Traditional knowledge of the structure of the Hudson Bay food web (from McDonald et al. 1997, p. 20).

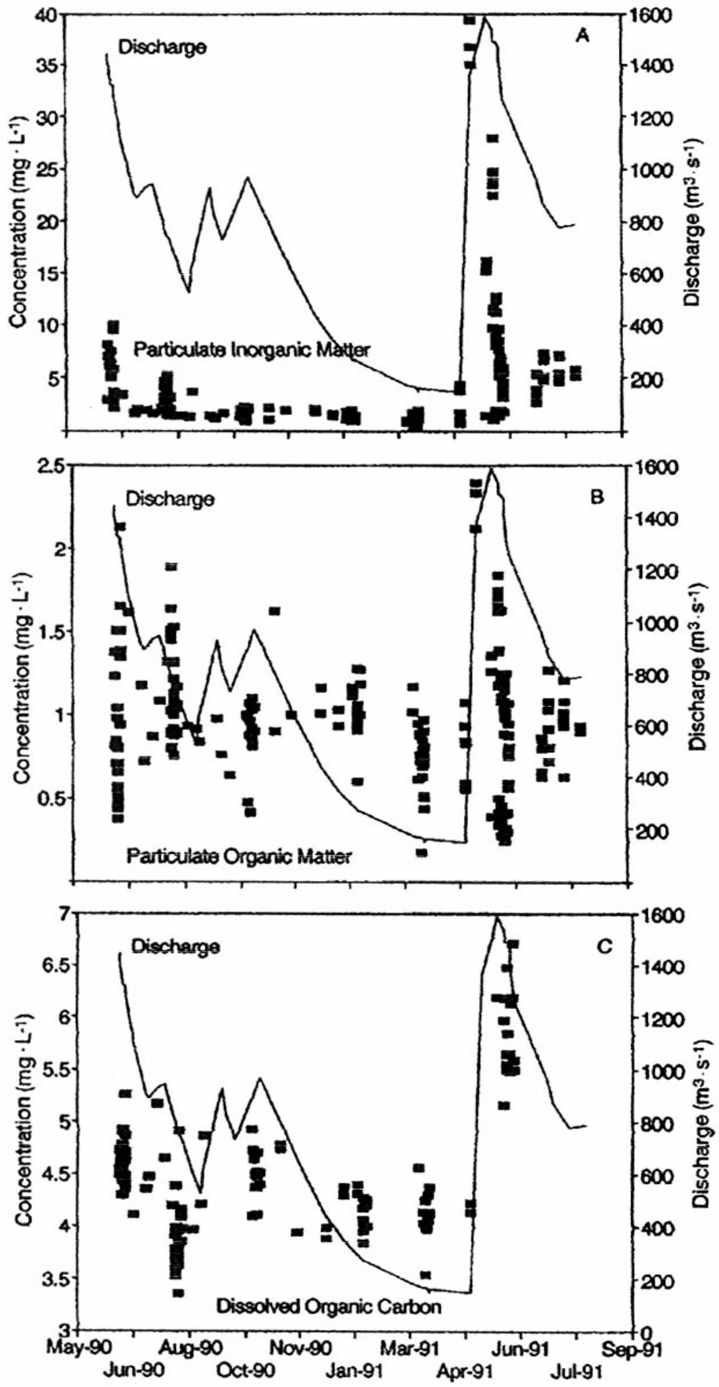

Figure 5-25. **Seasonal variations in the concentrations of (A) inorganic and (B) organic particulates, (C) dissolved organic carbon, and runoff discharge from Grande rivière de la Baleine (Great Whale River) (From Hudon et al. 1996, p. 1517).**

During the summer, primary productivity in the Hudson Bay marine ecosystem appears to be greater inshore than offshore (Figure 5-26). In August and September 1975, the primary production by phytoplankton southwest of the Belchers was similar to that observed in coastal areas of Hudson Bay (Anderson and Roff 1980a). The mean concentrations of particulate organic carbon, chlorophyll ***a***, and ATP were generally 2 to 3 times higher in these inshore areas than in offshore areas of Hudson Bay (Anderson 1979; Anderson and Roff 1980a). High chlorophyll ***a*** levels were found west of the Belcher and Sleeper archipelagos and near the entrance to James Bay, with maximum concentrations generally at or above the pycnocline (Anderson and Roff 1980b). Offshore in Hudson Bay the maximum summer chlorophyll concentrations were found below the pycnocline where nutrient concentrations are higher and the clear waters allow sufficient light penetration for photosynthesis (Figure 5-27; Anderson 1979; Anderson and Roff 1980b; Roff and Legendre 1986). This is one of the most northerly reports of a subsurface maximum and one of the most highly developed. It may contribute significantly to the annual production of Hudson Bay. Harvey et al. (1997) also found a well-developed subsurface chlorophyll ***a*** maximum at stations in northeastern Hudson Bay and western Hudson Strait (Figure 5-22; group C).

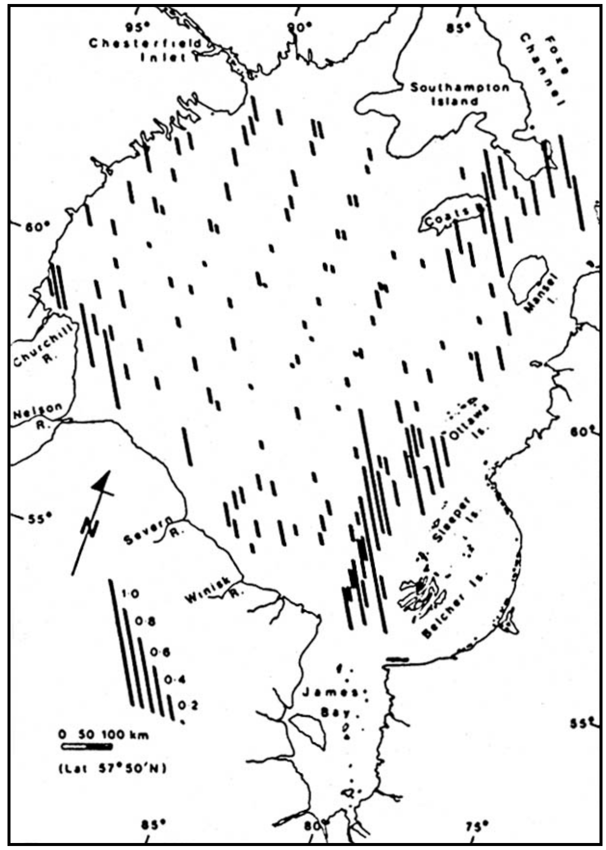

Figure 5-26. Surface chlorophyll *a* (mg$\cdot$m^{-3}) distribution in Hudson Bay, August-September, 1975. Station location is base of bar (from Anderson and Roff 1980a, p. 2247).

In summer, most primary productivity occurs in the upper 10 m of the water column, at or above the pycnocline (Legendre and Simard 1979; Anderson and Roff 1980b; Grainger 1982). At the Belcher Islands in the summer of 1959, the maximum-recorded rate of primary production was about 3 mg C$\cdot$m$^{-3}\cdot$h^{-1} (Figure 5-28). This was closely similar to the rates measured by Legendre and Simard (1979) in Manitounuk Sound (about 2.5 mg C$\cdot$m$^{-3}\cdot$h^{-1}), and notably low in the range of recorded marine values (Grainger 1982). These hourly rates translate to an annual primary production rate of about 35 g C$\cdot$m$^{-2}\cdot$a^{-1}, not including the main spring diatom bloom or ice algal production (Roff and Legendre 1986). In Manitounuk Sound, the summer blooms of phytoplankton occurred following stabilization of a previously unstabilized water column (Legendre et al. 1982).

Primary production under ice-covered conditions is not well known, but there is probably a well-developed microalgal community at the bottom of the ice in spring throughout most of Hudson Bay and James Bay (Freeman et al. 1982). The production of sea ice microalgae depends upon the available light energy, fortnightly tidal mixing which replenishes nutrients and removes accumulated algal biomass from the ice/water interface, and the heat flux responsible for the maintenance or destruction of energetic interfaces (e.g., the ice/water ergocline) (Gosselin et al. 1985; 1990; Demers et al. 1989). In southeastern Hudson Bay, production by sea-ice microalgae appears to be nitrogen limited (Maestrini et al. 1986; Demers et al. 1989). The salinity of surface waters underlying the ice is a major determinant of the standing crop and taxonomic composition of the ice algae, both of which are lower near freshwater outflows such as the Grande rivière de la Baleine (Poulin et al. 1983).

In 1982 at Manitounuk Sound, Gosselin et al. (1985) observed the maximum concentrations of ice-algal cells and chlorophyll ***a*** on 7 May at the ice water interface and also in the bottom 20 cm of ice (12 x 10^8 cells$\cdot$m^{-2} and 0.85 mg Chl ***a***$\cdot$m^{-2}). Poulin et al. (1983) sampled the same station in May 1978 and reported similar values.

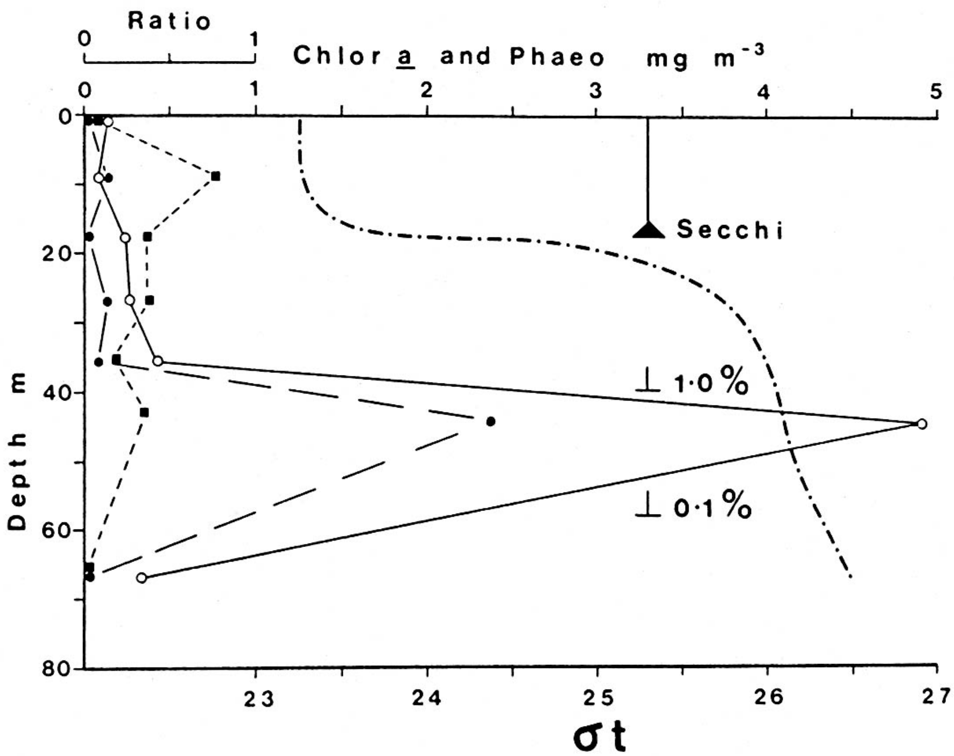

Figure 5-27. Vertical profiles of chlorophyll *a* (○—○), phaeopigments (●—●), phaeopigment to chlorophyll *a* ratio (■ – ■), and density (σ t; ● – ●) representative of the offshore region of Hudson Bay (from Anderson and Roff 1980b, p. 211). The 1.0 and 0.1% light levels were calculated from Secchi disc readings.

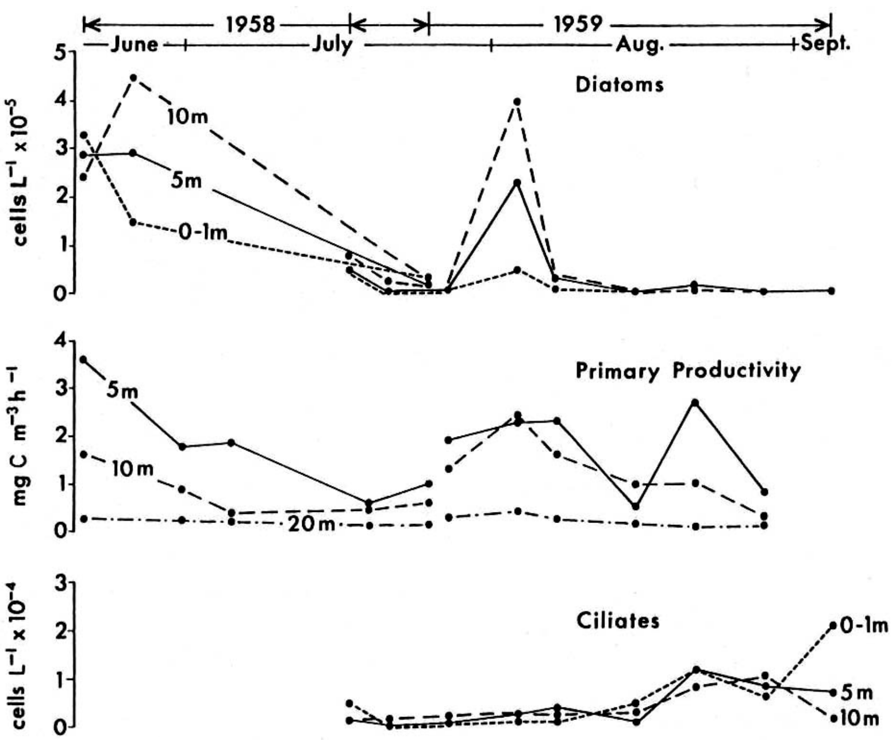

Figure 5-28. Seasonal variations of diatom counts, primary productivity rates, and ciliate counts at selected depths near the Belcher Islands (from Grainger 1982, p. 790).

These chlorophyll concentrations are an order of magnitude lower than many of those reported from other Arctic seas (e.g., Apollonio 1965; Dunbar and Acreman 1980; Hsiao 1980; Grainger and Hsiao 1982). It is not known whether they are representative of southeastern Hudson Bay.

Jones and Anderson (1994) estimated the total biological productivity in the upper 40 m of the Hudson Bay watercolumn to be at least 24 g $C{\cdot}m^{-2}{\cdot}a^{-1}$. Roff and Legendre (1986) estimated that the production of all planktonic and epontic communities in productive inshore areas of Hudson Bay may reach 70 g $C{\cdot}m^{-2}{\cdot}a^{-1}$. These estimates are comparable to those of other Arctic marine waters where primary production estimates range from 12 to 98 g $C{\cdot}m^{-2}{\cdot}a^{-1}$ --based on a 120 day growth season, with the higher values generally observed in the coastal bays and fjords (Subba Rao and Platt 1984).

Little is known of the primary production by benthic algae in this region but in Rupert Bay and Manitounuk Sound, where benthic algae have been studied (Breton-Provencher and Cardinal 1978), the low biomass observed suggests that productivity is also low.

Eelgrass beds in James Bay appear to be productive relative to their surroundings but no data were found on their rate of primary production. Depending on conditions the eelgrass, *Zostera marina*, grows at a density ranging from 88 to 1,761 shoots$\cdot m^{-2}$ and produces a dry biomass of 34 to 472 $g{\cdot}m^{-2}$ (SEBJ 1990). The density and biomass of the beds vary with location and year in response to the effects of waves and currents, ice, turbidity, water temperature, and illumination. While less dense and productive than beds in Alaska, which have densities of 788 to 5,033 plants$\cdot m^{-2}$ and biomasses of 186 to 1,840 g dry weight$\cdot m^{-2}$ (McRoy 1970), the communities do form the base of major food chains in the James Bay marine ecosystem (Curtis 1974/5; Ettinger et al. 1995).

Coastal salt marshes may be very productive relative to other areas of this region (see Section 4-5), but their contribution of food energy to the marine ecosystem is unknown.

5.7.2 Invertebrates

Secondary productivity may be of the same order as seasonally ice-free Arctic marine waters but significantly lower than that of temperate oceans as a consequence of longer generation times in the colder water, particularly offshore where the numbers and biomass are lower (Roff and Legendre 1986). Manitounuk Sound, for example, had an estimated standing biomass of copepod zooplankton of 0.7-0.9 g ash free dry weight$\cdot m^2$ and productivity of 80 $kJ{\cdot}m^{-2}{\cdot}a^{-1}$ in summer (Simard pers. comm. in Roff and Legendre 1986). In summer, Rochet and Grainger (1988) found the greatest copepod densities in southeastern Hudson Bay above the pycnocline near shore. Grainger and McSween (1976) found the zooplankton of James Bay to be present in moderate quantity and fairly high diversity for northern waters.

Small concentrations of commercially attractive benthic macroinvertebrate species have been located in the Belchers (Jamieson 1986; Giroux 1989). Individuals of some of these species, such as the green sea urchin *Strongylocentrotus droebachiensis* and blue mussel *Mytilus edulis*, are smaller as a rule than their temperate counterparts (Lubinsky 1980; Jamieson 1986). Test fisheries for invertebrates along the western and northeastern coasts of Hudson Bay have not identiified commercially attractive concentrations of benthic invertebrates (Morin 1991; Doidge 1992b; Doidge and Prefontaine 1997; Stewart 1994).

There are very few data on production by pelagic and benthic invertebrates in the Hudson Bay marine region but what there are suggest low productivity--despite the presence of large numbers of shorebirds and former presence of large numbers of bowhead. The *Calanus* Expedition found a good standing crop of Euphasiids in the Coats Island area but these data were never published (E.H. Grainger, DFO, Ste. Anne-du-Bellevue, pers. comm.). During summer, limited data suggest that the numbers and biomass of metazoan zooplankton in inshore and estuarine areas are similar to temperate oceans, while in offshore areas they are much lower--presumably as a consequence of low primary productivity (Roff and Legendre 1986). The biomass and abundance of total zooplankton along a north-south transect from the mouth of James Bay to Hudson Strait were low in the south (averaging 1.6 g $DM \cdot m^{-2}$ and 9432 $ind \cdot m^{-2}$) and increased sharply at the upper end of the bay and in Hudson Strait (averaging 6.0 g $DM \cdot m^{-2}$ and 40,583 $ind \cdot m^{-2}$) (Harvey et al. 2001).

In James Bay, the maximum diversity of the zoobenthos recorded from a single sampling site by Wacasey (1976) was 35 species, and the maximum biomass was 13 $g \cdot m^{-2}$. This is less than in other areas of the Eastern Arctic, but comparable to known values from bottoms with similar conditions in the southern Beaufort Sea. Along the west coast of James Bay, high densities of the clam *Macoma* and gastropod *Hydrobia* have been observed in sandy, sheltered flats or bays (Martini et al. 1980b). Nearby, coastal areas influenced by river mouths were found to have very low densities of invertebrates.

5.7.3 <u>Vertebrates</u>

As with plants and invertebrates, little is known of vertebrate productivity in the Hudson Bay marine ecosystem. But, species abundance provides indirect evidence for higher productivity in some areas.

Anadromous species excepted, commercially attractive fishes have not been found in sufficient quantity to warrant fishery development (e.g., Hachey 1931; Vladykov 1933; Hunter 1968; Dunbar 1970; M. Allard, Makivik, Corp., Lachine, pers. comm.). This suggests that inshore and estuarine areas may be more productive than offshore areas, and that the offshore areas may be less productive than other oceans at similar latitudes--the extent to which it reflects the difficulty of sampling the deeper offshore waters is unknown.

The numbers of some mammal and bird species have been estimated, but most of them are migratory. The contribution food resources in the waters of James Bay and southeastern Hudson Bay make to their biological productivity is unknown. The seasonal abundance of shorebirds and waterfowl in coastal saltmarsh habitats, and of waterfowl at eelgrass beds, suggests that these may be relatively productive habitats. So too, the historical summer presence of large numbers of belugas in the Churchill, Nelson, Nastapoca, Grande Baleine, and Petite Baleine estuaries; concentrations of walrus at Cape Henrietta Maria and on Coats Island; and bearded seal in the Belchers. The historical presence of a large population of bowhead (Ross 1974; Reeves et al. 1983) strongly suggests that there is an area of high productivity in northwestern Hudson Bay, but its existence has not been proven. The apparent historical rarity of bowhead in James Bay and southeastern Hudson Bay suggests that the production by copepod zooplankton may be insufficient to support them in these areas.

5.8 SUMMARY

The Hudson Bay marine ecosystem consists of two oceanographically distinct marine regions. The water properties of these regions depend mainly on exchanges with Foxe Basin and Hudson Strait and the large freshwater input from both runoff and melting sea ice in the spring and summer. An understanding of their differences is critical to the design and integration of coastal zone management initiatives.

The northern area, or **Hudson Bay marine region**, is characterized by the presence of Arctic marine water and biota, complete winter ice cover and summer clearing, moderate semidiurnal tides of Atlantic origin, a strong summer pycnocline, greater mixing and productivity inshore than offshore, and low biological productivity relative to other oceans at similar latitudes. Hudson Bay lacks the typically subarctic species that are found in Hudson Strait but does support some of the relict warm-water species found in James Bay.

The southern area, or **James Bay marine region**, is closely coupled oceanographically to the Hudson Bay marine region but its waters are typically shallower and more dilute, being modified to a much greater extent by freshwater runoff from the land. Its species composition reflects these Arctic and freshwater influences and it supports a variety of warm-water species that are relicts of an earlier connection with the Atlantic and Pacific oceans. These plants and animals have disjunct distributions and are rare or absent elsewhere in Canada's

Eastern Arctic waters. Southeastern Hudson Bay is included in this region with James Bay largely on the basis of biogeography. Strong density stratification limits mixing and leads to considerable surface warming by insolation in both marine regions.

Because of its remote location and the non-commercial nature of its marine resources, relatively few oceanographic field programs have been undertaken in the Hudson Bay ecosystem, where seasonal ice cover effectively prevents most year-round research and the shallow coastal waters make it very difficult to conduct bay-wide research from a single research platform. Consequently, characteristics of the circulation and water mass are not well known, especially outside the open water period.

In summer, surface water circulates cyclonically (counterclockwise) around Hudson Bay, and the deep water moves in the same general direction but is influenced by bottom topography. Cold, saline Arctic water from Foxe Basin enters Hudson Bay in the northwest via Roes Welcome Sound. As it flows eastward along the southern coast of Hudson Bay some of this water enters James Bay while the remainder is deflected northward to exit northeastward into Hudson Strait. A westward, wind-driven return flow across the top of Hudson Bay has been predicted by modelling studies, and there is a small--perhaps intermittent, intrusion of Atlantic water from Hudson Strait at the northeastern corner of Hudson Bay. Mathematical modelling suggests that the main reasons for this stable cyclonic circulation are the relatively weak coastal currents with limited coastal development to cause mixing, a relatively strong Coriolis effect that stabilizes the flow pattern by turning the freshwater outflow from rivers cyclonically around Hudson Bay, and strong density stratification due to intense freshening in summer. This circulation is maintained by inflow/outflow forcing that likely occurs year round, and reinforced during the open water season by wind and buoyancy forcing. The extreme southerly incursion of Arctic waters creates Arctic oceanographic conditions much further south than elsewhere along the North American continent, and is a key feature of the Hudson Bay marine ecosystem.

There is little Atlantic influence except in terms of the tides which enter Hudson/James Bay twice daily via Hudson Strait. These semidiurnal tides move as a Kelvin wave counterclockwise around the coastline and overshadow other tidal influences. They do not attain the extreme ranges in height found in Hudson Strait. Dangerous storm surges do occur in southern James Bay.

The Hudson Bay marine ecosystem is unusual among the world's oceans in that it is nearly covered by ice in winter and is free of ice in summer. In spring and summer, the cold saline surface water that enters the region is diluted by meltwater and runoff from the land, warmed by the sun, and mixed by the wind as it circulates through Hudson Bay and James Bay. This produces the strong vertical stratification of the water column that is characteristic of the ecosystem in summer, particularly offshore. This stratification slows vertical mixing, thereby limiting nutrient additions to surface waters and biological productivity. In winter, lower runoff, ice cover, and surface cooling weaken the vertical stratification and permit very slow vertical mixing. There is little coastal development or bottom relief to promote mixing or upwelling that might increase the availability of chemical nutrients in the surface waters. Temperature and salinity are relatively stable below a depth of 50 m, but small changes related to the seasonal disappearance of the pycnocline have been observed to 65 m in James Bay and 100 m in Hudson Bay. The water becomes progressively colder and more saline with depth, approaching the same deep water type at about 100 m where the mean temperature is less than -1.4°C and salinity greater than 33 ‰ (‰≈psu). The deep water layer in James Bay is subject to considerable seasonal and interannual variation in temperature and salinity, due in part to the relative shallowness of the bay. Seasonal oceanographic variations are not well known. There is no complete set of temperature-salinity transects that covers the entire area in any season, and most sampling has been conducted during the open water season.

The extreme southern presence of nearly complete ice cover with extensive areas of fast ice and polynyas strongly affects this region's physical and biological oceanography, the surrounding land, and human activities. Depending on weather conditions, the timing of freeze-up or breakup may be retarded or advanced by up to a month, but the basic pattern of ice formation remains similar. The reliance of Inuit and coastal Cree on sea ice for travelling and hunting is reflected in their detailed knowledge of its processes, characteristics, and annual cycles. The sea ice determines the ecology of the ice biota and it also influences pelagic systems under the ice and at ice edges. As the interface between air, ice, and water, ice edge habitats are areas of mixing that attract biota to feed. These areas are important sites of energy transfer within the ecosystem.

In winter and early spring the ice floes are kept in constant motion by the wind. Leads develop when the winds blow offshore and are quickly covered by new and young ice. These leads are important habitat for species such as the Hudson Bay eider that overwinter in the region and to migratory birds and mammals that arrive early in the spring. Recurring polynyas are present in the Belchers and near islands along the coast of southeastern Hudson Bay, in Roes Welcome Sound, at the northern tip of Coats Island, near Digges Island, and just off the southwest tip of Akimiski Island. The latter polynya is one of the most southerly in Canadian seas. These openings in the sea ice are vitally important to overwintering species and to early spring migrants. They are often areas of increased biological productivity. Old ice and icebergs are rare in Hudson Bay and rare or absent James Bay.

The importance of sea ice to the Hudson Bay marine ecosystem and its vulnerability to climatic warming have spurred efforts to develop a mathematical model that accurately simulates the region’s sea ice dynamics.

The volume of freshwater runoff to this region from the land is very large and has an even greater effect on the oceanography of the James Bay marine region than is seen in the Hudson Bay marine region. It has a strong influence on the timing and pattern of the breakup of ice cover, the surface circulation, water column stability, species distributions, and biological productivity. Summer surface salinity values over most of this region are low relative to other marine regions. Extensive freshwater plumes are observed off its river mouths year-round. They spread further and deeper under the ice than under the ice-free conditions of summer, despite runoff rates that are an order of magnitude lower. The effects of high runoff are most pronounced in eastern James Bay, along the southeastern coast of Hudson Bay and, perhaps, in Richmond Gulf. In southern and western James Bay, which are shallow and receive a great deal of sediment laden runoff, the water clarity is low relative to other parts of the marine ecosystem and to other Arctic marine regions generally.

In summer, there are distinct physical and biological oceanographic differences between inshore and offshore areas of this region. Inshore areas generally have lower water temperatures, salinities, and clarities and higher chlorophyll ***a***, ATP, and pelagic biomass. These differences may be attributable to mixing processes which bring colder, deeper, relatively nutrient-rich water to the surface, and to dilution and nutrient addition by freshwater runoff. Vertical density stratification is particularly strong offshore in central Hudson Bay, where it effectively prevents mixing of the surface and deep waters and thereby replenishment of nutrients above the pycnocline.

Freshwater runoff affects the primary productivity negatively by increasing vertical stability of the water column, and positively through nutrient additions--either direct or due to deep-water entrainment. While river runoff carries large quantities of carbon and nutrients into the marine ecosystem, particularly during ice-breakup, the river waters are less concentrated in nutrients than Hudson Bay coastal waters.

Biological productivity appears to be low relative to other oceans at the same latitude and comparable to that of seasonally open-water areas of Canada's Arctic Archipelago. It appears to be greatest in coastal waters, particularly at embayments and estuaries, and near islands where there is periodic entrainment or upwelling of deeper, nutrient-rich water. Productivity above the pycnocline and under the ice may be limited by the availability of nutrients, particularly nitrogen. In summer there is a layer of maximum primary productivity below the pycnocline in Hudson Bay. The historical presence of large numbers of bowhead whales suggests that there is an area of higher productivity in northwestern Hudson Bay. The structure of the food web in the Hudson Bay marine ecosystem is not well known, nor is the flow of energy through that web.

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