On the Freshwater Transport Through the Southwest Canadian Arctic Archipelago due to Buoyancy and Wind Forcing

by
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Abstract

The freshwater input from the Arctic into the North Atlantic is an important component of the global climate system through its effects on deepwater formation. Part of this freshwater is transported through the Canadian Arctic Archipelago (CAA) via sea ice and low density surface water, where it is able to set up buoyancy boundary currents (BBCs). To infer the existence of freshwater transport via BBCs in the southwest CAA, data are examined from summer cruises conducted in 1995, 1999, and 2000. The hydrographic data are supplemented with traditional knowledge relevant to this study. The presence, predominantly on the south side of channels, of driftwood originating from the Mackenzie River confirms an eastward transport through the region. The hydrographic data also show that the southwest CAA is relatively fresh compared to surrounding regions, and that the sources of buoyancy forcing are large and from different origins. The presence of BBCs on both sides of the channels appears to be a frequent occurrence with, as shown in previous work, the fresher water being more often present on the south shore. Some data from the summer 2000 show a different feature with much fresher water on the north side. A subsequent strong wind event creates a complete reversal of this situation, setting up a strong cross-channel horizontal salinity gradient and an amplified BBC on the south shore. In this region, buoyancy and wind forcing act together to force an eastward freshwater transport in the southwest CAA.

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Table of Contents

List of Tables

List of Figures

Glossary

ADCP:

Acoustic Doppler Current Profiler: Instrument measuring the current by measuring the Doppler effect of the current on an acoustic wave sent by the instrument.

Current meter:

Instrument measuring the strength and the direction of the ocean current.

Drifter:

Instrument that drifts with the current. Some are passive drifters that will give only starting point and final point (where it was left, and were it was found). Others are satellite tracked drifters and give the information along the path they follow.

Inertial period:

Period of the earth rotation.

Kelvin wave propagation:

A Kelvin wave can travel along the coast in one direction only, that is with the coast on its right (left) in the northern (southern) hemisphere.

O(1):

Of the order of one, notation used when the quantity is around or of the same order

Psu:

Practical salinity unit: reference unit to measure salinity of the water.

Pycnocline:

Region of the water column where there is a strong change in density. The pycnocline separates waters of very different density (upper layer and lower layer).

Pycnocline ventilation:

Due to the existence of the pycnocline, the exchange between the upper and lower ocean layers is damped. The ventilation is the exchange of water masses from the upper layer with the lower layer.

Rossby deformation radius:

During a classical experiment called the lock exchange experiment, a vertical barrier separates two fluids of contrasting densities. When the barrier is removed the denser fluid flows under the lighter one. After about one inertial period, the collapsing flow is deflected to the right. Through geostrophic adjustment this leads to an S-shaped density interface, which is always confined to about the same distance called internal Rossby deformation radius.

Supercritical and Subcritical flow:

Internal gravity waves tend to form near the interface of two layers, and these waves then propagate. When the flow has a speed much smaller than the speed of these internal gravity waves the flow is subcritical. When the flow speed is of the same order as the speed of the internal gravity waves the flow becomes critical and the energy cannot go away with the waves and tends to accumulate. The flow then tends to become unstable. When the flow goes much faster than the internal gravity waves, the flow is supercritical and instabilities tend to develop.

TS diagram:

Diagram with the temperature (T) is on the vertical axis, and the salinity (S) on the horizontal axis. This type of diagram helps to look at water properties (origins, mixing etc.).

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" I got the feeling that I had constantly to match wits with the moving pack ice. Many times did I head for an opening in the ice only to watch it crash together just ahead of me, as it were a living thing deliberately trying to keep me from reaching open water.
On other occasions the ice would snap shut behind me, as it held me in a trap. But it also happened that when things looked hopeless and I was almost resigned to giving up, the ice would suddenly open up and the leads would gradually get wider and wider and allow us to slide through the cracks for mile after mile."

- "Voyages of the RCMP schooner St.Roch" by H. Larsen

Chapter 1 — Introduction

The Arctic Ocean and its surrounding seas are fed by freshwater that enters these regions via atmospheric transport, ocean inflow, and river inflow. Sea ice formation in the winter, and associated brine rejection, as well as sea-ice melting during the summer together increase the capacity of the upper ocean layer to act as a freshwater reservoir. The Arctic Ocean can be seen as a huge reservoir of freshwater made of sea-ice and low-density surface water formed by the melting of sea ice and river input. The advection of this freshwater in the form of sea-ice and lower density surface water from the Arctic into the northern North Atlantic influences deep water formation, and thus influences the Earth's climate (Aagaard and Carmack, 1989). Therefore, understanding the advective process of freshwater from the Arctic Ocean into the North Atlantic is an issue of great importance.

The pathways of this export toward the northern North Atlantic are through Fram Strait, Barents Sea and the Canadian Arctic Archipelago (CAA). The main export is through Fram Strait in the form of sea-ice (2,790 km^3yr^{-1}) and ocean freshwater transport (820 km^3yr^{-1}; Aagaard and Carmack, 1989). The oceanic freshwater transport through the CAA (920 km^3yr^{-1} — Aagaard and Carmack, 1989) is the second largest source albeit certainly the most complex. As mentioned by Melling (2000), the CAA occupies about 20% of the combined area of the Arctic Ocean and its surrounding seas. Often, differences in steric height between the Pacific Ocean, the Arctic Ocean and the Atlantic Ocean are invoked for the advection of ocean water toward the Atlantic Ocean. Considering this approach, one would tend to expect a relatively simple two-layer flow from the Arctic Ocean to the North Atlantic through the CAA with no important horizontal changes of density and velocity in the cross channel direction. However, observations in the CAA (e.g. Leblond, 1980) have shown the presence of right-bounded surface currents in opposite directions on facing sides of the channels, with the low density waters (buoyancy forcing) concentrated on each side near the shore.

In the northern hemisphere (Leblond, 1980) buoyancy forcing balanced by the Coriolis force can lead to these right bounded coastal currents which transport a significant component of the freshwater in the upper layer of the ocean, and are termed buoyancy boundary currents (BBCs). Arctic BBCs can be thought of as rivers flowing along the coast of the ocean. Therefore, the mechanisms of the eastward advection of freshwater through the CAA, which is a substantial buoyancy input to the northern North Atlantic, are complex and need to be clarified.

This study will first briefly summarise in section 2 what is known about buoyancy driven currents at high and lower latitudes, including a discussion of some previous work that shows the importance of buoyancy forcing in the coastal ocean, and the importance of quantifying the fresh water input as well as its origin. This will also help to define more precisely the buoyancy driven currents. Then, an analysis of what is known about the importance of the fresh water input to the Arctic will be presented in sections 2.3 and 2.4, leading to a focus on the specific region of interest in section 3. After a short description of the southwest part of the CAA, the freshwater input to the study region will be discussed. In the fourth section some traditional knowledge and observations are described that point towards an eastward transport in this region. The fifth section is dedicated towards demonstrating the importance of the presence of freshwater in the upper and lower layers of the Coronation Gulf/Dease Strait/Queen Maud Gulf study region of the CAA. Here too, hydrographic sections will be used to show the importance of buoyancy forcing and wind forcing in driving an eastward transport. Then the flow is characterized using physical non-dimensional parameters for each example. These results are summarised in the sixth section.

Chapter 2 — Background

2.1. Buoyancy input

The input of freshwater to the coastal ocean, occurs either through the surface (sea-ice melt, precipitation) or along lateral boundaries (river input). This buoyancy input implies horizontal and vertical buoyancy gradients which are both dynamically significant (Hill, 1998). The latter increases the vertical stratification, limiting vertical motion and mixing, while the former creates a horizontal pressure gradient due to the horizontal variation in density. The change in the horizontal pressure gradient due to buoyancy input induces horizontal motion which, after about one inertial period, combines with the Coriolis effect, to create a geostrophic balance. At first, a plume of relatively fresh water exists which later reattaches to the coast to form a right-bounded coastal current (in the northern hemisphere), or BBC (Hill, 1998). As shown below and in other work (e.g. Münchow, 1992, Hill 1998), BBCs are not steady currents, as they can develop instabilities depending on the relative depth of the flow to the local ocean depth. They can also dissipate depending on both the shape of the coastline and the relative depth of the flow, as well as through direct external forcing of the BBC.

As noted earlier, when there is an input of freshwater or buoyancy, a horizontal variation in density may ensue, which also means that the isopycnals would no longer be horizontal; therefore, they would not be parallel to the approximately horizontal isobars and the state is referred to as being baroclinic. Melling (2000) mentioned that through baroclinic adjustment, the density stratification influences the width of the BBC. When the flow is stratified, baroclinic effects are important, and the BBC scales with the internal Rossby deformation radius R' (Hill, 1998).

At high latitudes, as in the case for the CAA, R' (~5–10 km) is an order of magnitude smaller than at lower latitudes where similar ensembles of channels exist (in the Indonesian Archipelago, for example). Most of the CAA channels are wide enough to allow for the presence of flows of opposite direction on each side of these channels (Leblond 1980), making the flow through the CAA quite complex. As well as underlining the importance of the width of the channels, Leblond (1980) argued that meandering and instabilities could be the reason for observed flows that are wider than R'. Stratified flows can develop barotropic and baroclinic instabilities, especially near discontinuities or sharp corners of the coast line. When instabilities and meandering occur, mixing occurs; the stratification is then weakened as is the baroclinicity, and hence the BBC is weakened as well (Melling, 2000).

The presence of sills in the complex topography of the CAA also affects the BBCs and throughflow. The shallow sills that the flows have to go through range from 20 m (e.g. Dolphin and Union Strait) to 220 m (Nares Strait). This restriction of the upper part of the ocean also affects the upper layer by turbulent mixing due to friction and tidal forcing. The turbulent mixing due to friction is increased during winter by the presence of the sea-ice cover, but the sea-ice cover prevents turbulent mixing due to wind forcing. The static winter sea-ice cover therefore affects the depth of the layer that carries the freshwater. Due to the presence of numerous shallow sills in the CAA, it is important to note that the freshwater transport occurs mainly in the upper layer where it is affected not only by the topography, but also by the coastline shape, and the sea-ice cover.

2.2. Examples of BBCs

At high latitudes, since at low temperatures the salinity dominates the equation of state (Gill,1982), buoyancy fluxes are dominated by salinization (brine rejection) and freshening (sea-ice melt and river run-off). At mid-latitudes, where temperature becomes more important, numerous BBCs forced by river input have been observed. Münchow (1992) reports corresponding salinity gradients that are three to nine times smaller than found in this study (3 – 5 psu over 5 km; see section 5). Drifter and ADCP data have also shown that the BBCs associated with these relatively weak gradients have speeds ranging from 10 to 30cm/s. For a three to five times weaker salinity gradient (1 psu over 5 km) than found in this study, Münchow and Garvine (1993b) have shown the existence of a BBC mean flow of about 10 cm/s despite non-favorable winds.

With stronger salinity gradients and also a stronger effect of salinity on the equation of state at high latitudes, one would expect quite swift BBCs to exist. If one considers buoyancy input from the coast as the driving force for a BBC, the flow direction is expected to be such that in the northern hemisphere, the coast is on the right (i.e., in the sense of Kelvin wave propagation). The presence of melting sea-ice can strongly influence the BBC through acting as a horizontally discontinuous source of buoyancy and hence a driving boundary. During the summer of 1995 Münchow et al. (1999), using deployments of 39 tracked surface drifters, tested the hypothesis that the Kolyma River forces an eastward BBC from the East Siberian Sea towards the Bering Strait. The drifters showed an unexpected drift towards the opposite direction and the westward flow correlated weakly with the local winds. The data showed a spatially-coherent flow of up to 50 cm/s with the spatial scale of the flow exceeding the nearshore internal Rossby deformation radius (R') by a factor of three, although it corresponded well with the internal Rossby deformation radius associated with the offshore ice edge. Hydrographic observations and oxygen isotope ratios of sea-water showed a high proportion of freshwater due to sea-ice melt relative to riverine freshwater in the East Siberian Sea. This data, along with the drifter data, showed that the BBC was driven by the freshwater from the melting of sea-ice. The retreating ice-edge was acting as a horizontally discontinuos source of buoyancy and hence as a driving boundary, forcing a westward surface flow in the East Siberian Sea. This westward flow contradicted prior measurements (Sverdrup, 1929, and U.S. Weather

Bureau, 1957), and dynamical expectations (Yankovsky and Chapman, 1997, and Garvine, 1999) of an eastward flow.

The fact that an eastward flow due to large river discharge was not observed in 1995 suggests that it is a transient rather than a permanent feature of the region. Rather, during times when the sea ice edge is retreating, a flow in the opposite direction is set up. Therefore the buoyancy sources have to be inferred and quantified properly to understand the freshwater transport in Arctic regions. At first, the importance, quantitatively speaking, of the freshwater input and freshwater storage in the Arctic has to be considered.

2.3. River runoff and drainage basins of Arctic rivers

The drainage basins of the Arctic rivers extend as far south as 45° N, although the total area of the combined drainage basins depends on the definition of a contributing drainage basin. As mentioned by Prowse and Flegg (2000), depending on this definition, the cumulative contributing area varies from 15.5×10^6 km^2 to 23.7×10^6 km^2. The former estimate only includes the drainage basins of rivers that flow directly into the Arctic Ocean. In this definition most of the CAA and all of the Hudson Bay regions are not considered. In the latter estimate, the CAA, Hudson Bay and also the rivers that flow into the Bering Sea are considered. Over the range of definition in Arctic river drainage basins, the estimated runoff varies from 3,299km^3 yr^{-1} to 5,250 km^3 yr^{-1}. The magnitude of this flux demonstrates that despite the fact that the Arctic Ocean represents only 5% of the total area and 1% of the total volume of all the world's oceans, it receives in excess of 10% of the total global river runoff.

A conservative estimate of Arctic Ocean river runoff is available from gauged river data obtained from the Global Runoff Data Centre (GRDC: http://www.bafg.de/grdc.htm) which gives a runoff of 2603 km^3/yr, but only approximately 60% of the total area is covered by hydrometric measurements. Also, Arctic rivers are technically very difficult to gauge and the drainage basins are often in quite isolated areas, so uncertainties are large. The runoff from each Arctic river shows a strong seasonal cycle (Grabs et al., 2000; GRDC data) with a minimum in late winter around April and a maximum around June. For most rivers, the runoff goes from its minimum to its maximum in only two months, and the magnitude of the maximum runoff is up to thirty times that during minimum runoff. Only the Mackenzie River (262 $km^3\ yr^{-1}$) and the Yenesei River (630 $km^3\ yr^{-1}$) have headwater lakes (Athabaska [7,936 km^2], Great Slave [28,930 km^2], and Great Bear [31,000 km^2] lakes for the Mackenzie River; Lake Baykal [31,500 km^2 — more than 1600 meters deep] for the Yenesei). The presence of these headwater lakes damps the seasonal variation in river runoff and also helps maintain a larger runoff magnitude compared to other rivers in the region.

The GRDC data reveal that the Mackenzie River shows the least variability with a maximum runoff (typically 26,800 m^3s^{-1}) thirteen times larger than the minimum runoff (typically 2,110m^3s^{-1}). The Yenesei, due to the extent of its drainage basin, and the Ob rivers have the aforementioned large variations from late winter to summer, but the late winter outflow of the Yenesei River is still 1.5 times bigger than the Mackenzie River winter outflow and the Ob River outflow is about the same as that of the Mackenzie River.

As will be seen below, the Mackenzie River has an important influence as a buoyancy forcing source on the CAA throughflow. The Mackenzie River, with a drainage basin of 1.8×10^6 km^2, represents the largest North American contribution to the freshwater input to the Arctic Ocean. Other smaller rivers flowing from the Canadian mainland and the CAA islands into the CAA waters contribute further to this freshwater input.

The combined area of the CAA islands is 1.1×10^6 km^2, while the rest of the Canadian mainland (i.e., without the Mackenzie River drainage basin), being the sum of drainage basins drawing freshwater into the CAA, yields a total area of 9.6×10^5 km^2. Thus, the mainland and the CAA islands form a total contributing area of more than 2.1×10^6 km^2, equivalent to the Mackenzie River drainage basin. Nevertheless, almost no data are available to confirm that this region gives an equivalent runoff to the Mackenzie River since these areas are almost entirely ungauged. The total area of contribution for these regions (mainland including Mackenzie drainage basin and CAA islands) composes 3.58×10^6 km^2 of land from which freshwater runoff to the CAA can greatly influence the dynamics of the currents there through buoyancy forcing. If we consider an average runoff of 145.5 km^3/yr per 10^6 km^2 of land (which is the Mackenzie River drainage basin average runoff estimated from GRDC data), we obtain an estimated freshwater runoff of 520 km^3/yr from the Canadian mainland and the CAA islands (with large uncertainties) into the western side of the CAA (for the Mackenzie River) and directly into the CAA waters.

2.4. Sea-ice

As mentioned earlier, sea-ice formation and melting represent substantial buoyancy sources/sinks in the Arctic region. Winter sea-ice formation and associated brine rejection lead to pycnocline ventilation (Melling and Lewis, 1982) and sea-ice melting in the summer lowers the salinity of the surface layer. The mean sea-ice volume over the Arctic Ocean is estimated to be 2.1×10^4 km^3 (Aagaard and Carmack 1989), with primary export through Fram Strait (estimated at 2,875 km^{3}yr^{-1}), and secondary export through the CAA towards the North Atlantic.

The CAA occupies about 2.5×10^6 km^2 of the continental shelf. This area minus the combined area of the islands mentioned above gives a total 1.6×10^6 km^2 of ocean. During winter, the ice thickness in the CAA ranges from 2 to 5 meters (Melling, 2000). Using a very conservative value of 2 m on average, one can estimate that 3,200km^3 of sea-ice forms during winter, equivalent to 3,060 km^3 of freshwater, which is at least ten times larger than the annual run-off from the Mackenzie River. As mentioned above, a crude estimate of the freshwater runoff from the mainland and the CAA islands is around 520km^3/yr. This is about 20% of the total fresh water input due to sea-ice melting, assuming all the sea-ice melts during the summer which is not always the case especially in the northern CAA (e.g. Berton, 1988, or Melling, 2000). The freshwater input from sea-ice is therefore very important in the CAA.

The sea-ice in the CAA channels is not moving for 6 months of the year, but during the other 6 months the sea-ice drifts at a speed between 5 and 20 cm/s (Melling, 2000). Drift of ice islands from the Arctic Ocean through the archipelago have also been observed (Jeffries and Shaw, 1993). Using low values for ice thickness and ice speed, Melling (2000) estimated that the flux of freshwater through the CAA from sea-ice drift to be 480km^3 for 6 months, considering 2.5m sea-ice passes through the key passages of the CAA at a moderate speed of 10cm/s. This amount for 6 months is almost 20% of the yearly sea-ice export through Fram Strait. Also, even in the northern CAA channels where all the sea-ice does not melt during summer, sea-ice is an important freshwater source during melting, and even more so in the southern part of the CAA where most of the time sea-ice completely melts during the summer. Considering all the sources of freshwater for the CAA, one would expect the formation of BBCs around each island and all along the coast, making the circulation through the CAA quite complex.

Chapter 3 — Description of the Region of Interest

3.1. Southwestern CAA

The south west region of the CAA (Figure 1) that is considered here is composed of Coronation Gulf, Dease Strait and Queen Maud Gulf. As mentioned earlier for the CAA in general, this area has a complex coastline and is bounded by shallow sills. On the west side of this region, Dolphin and Union Strait separates Coronation Gulf from Amundsen Gulf (600m deep, and 125km wide). Dolphin and Union Strait is a long strait stretching between the Canadian Mainland and Victoria Island, and it is only 20m deep and 30km wide in its shallowest part. Towards the east, the depth of Coronation Gulf increases to 400m and widens to 100km. On the east of Coronation Gulf, Kent Peninsula narrows the North West Passage to form Dease Strait, which is only 20km wide at its narrowest part and around 80 m deep. Still further east is Queen Maud Gulf where the North West Passage widens to 125 km and deepens to around 100m. Queen Maud Gulf is bounded on its north side by Victoria Strait, with a depth of 40m, and on the east by Simpson Strait, which is only a few hundred metres wide and about 12m deep. It should be mentioned that in this region of the CAA, tides are small and much less than one meter (Stronach et.al., 1987). Also the tidal currents are small and less than 5cm/s (Melling, 1998). The tidal currents are only strong in the very narrow straits like Simpson Strait as we observed in summer 1999.

3.2. Buoyancy forcing: River runoff and sea-ice

Many streams and rivers flow into the Coronation Gulf/Dease Strait/Queen Maud Gulf region of the CAA. Being typical Arctic rivers, their runoff varies from very low or nothing during winter to a maximum runoff in early summer. A few rivers are gauged in this region, with the two major ones (Burnside River and the Coppermine River) having gauges located far away from their mouth. Burnside River has a discharge close to zero in winter and 888 m^3/s, on average, in the summer, giving a mean annual runoff of $3.7 km^3/yr$ (GRDC data). The Coppermine River discharge varies from 22.8 m^3/s on average in winter to 476.6 m^3/s on average in the summer, giving a mean annual runoff of 3 km^3/yr. A few other rivers are gauged, but also far away from their mouth, and many other ungauged streams and rivers contribute to the total runoff into the region. Therefore, the combined total runoff from the mainland and the islands is hard to estimate. In addition, the southwestern CAA is completely covered with land-fast ice during the winter that disappears completely by the end of July, thereby creating a subsequent source of freshwater until the middle of the summer (and a sink of freshwater in the winter).

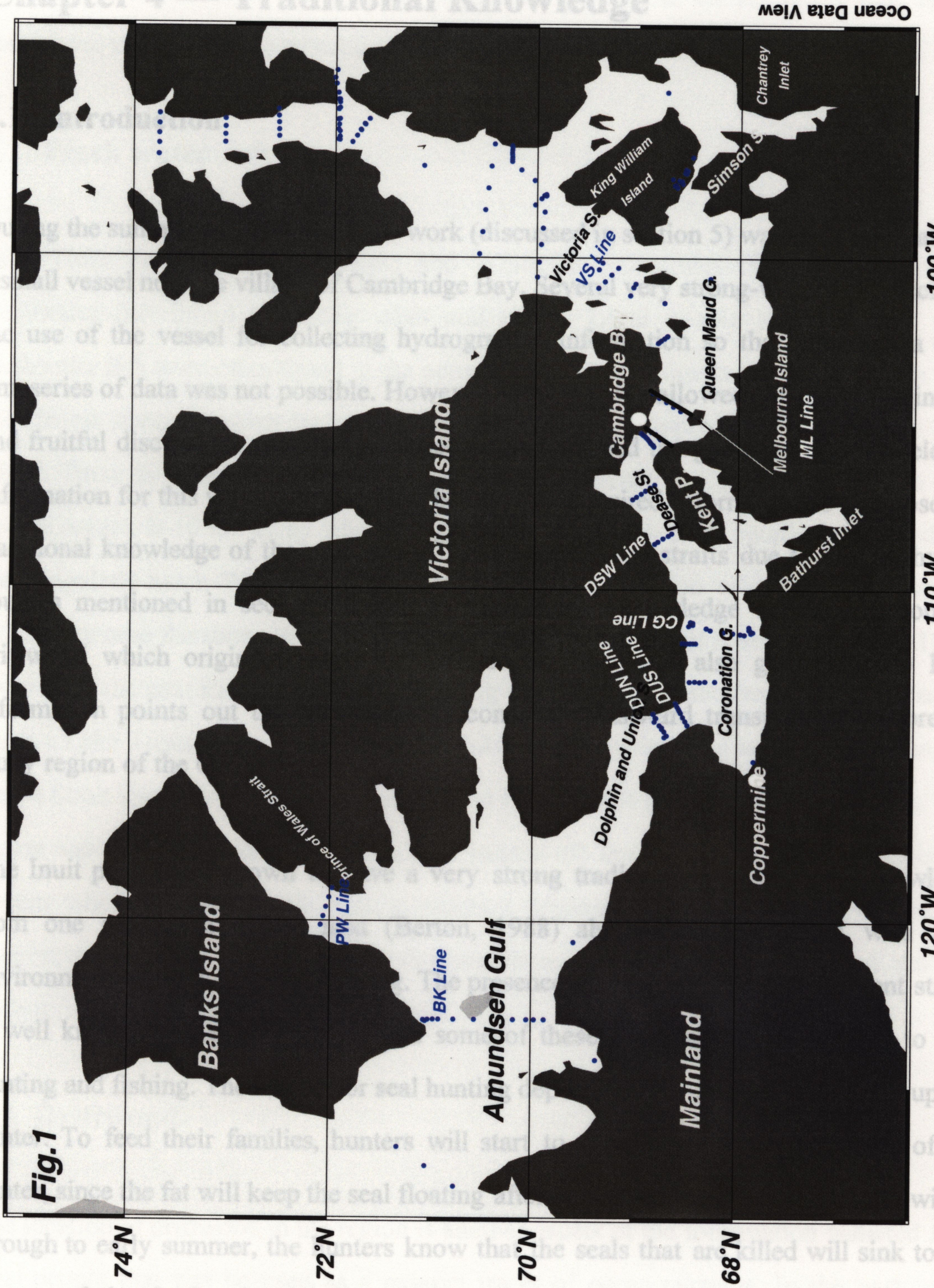

Figure 1: Map of the southwestern CAA containing the hydrographic stations of the summer cruises 9533, 9924, and 2000-33.

Chapter 4 — Traditional Knowledge

4.1. Introduction

During the summer of 2000, the field-work (discussed in section 5) was conducted aboard a small vessel near the village of Cambridge Bay. Several very strong-wind days precluded the use of the vessel for collecting hydrographic information so that obtaining a daily timeseries of data was not possible. However, these periods allowed for many fascinating and fruitful discussions with the local people and enabled the gathering of very relevant information for this work. As described below, this acquired information is composed of traditional knowledge of the presence of fresh water in the straits due to the freshwater sources mentioned in sections 2 and 3. In addition, knowledge of the locations of driftwood which originates from the Mackenzie River was also gathered. The latter information points out the presence of a consistent eastward transport in the present study region of the CAA.

The Inuit people are known to have a very strong tradition in passing oral knowledge from one generation to the next (Berton, 1988) about their interaction with their environment while fishing and hunting. The presence of freshwater in the different straits is well known to the local people and some of these observations are related to seal hunting and fishing. The season for seal hunting depends on the fat that seals stock up for winter. To feed their families, hunters will start to catch seals at the beginning of the winter, since the fat will keep the seal floating after it is shot. Near the end of the winter through to early summer, the hunters know that the seals that are killed will sink to the bottom of the fresher layer, being less buoyant than the freshwater but being more buoyant than the saltier lower layer. This fact shows that local people are aware of the presence of a more buoyant layer at the surface — a fresh surface layer that is also sometimes used by fish to travel far away from the river mouth.

4.2. Fresh water fish

Lake Trout (salvelinus namaycush) and Whitefish (coregonus cisco) are species of freshwater fish found in Arctic lakes and rivers. Local people reported that these freshwater fish were often found in the ocean or the CAA channels more than 10 km from the closest streams, which are not major rivers like the Coppermine River or the Back River. There is a location west of Cambridge Bay, named the *gravel pit* by the local people, which is 10 km away from the nearest river. A few local people mentioned that they regularly catch Whitefish there. This is also the location where logs occasionally washed up onto the shore. A second location east of Dease Strait is also more than 10 km away from the next stream, and a person going to his summer cabin was very surprised to catch a Lake Trout so far away from any lake. The third location which is southeast of Dease Strait was mentioned by a local guide to be a reliable place to catch Whitefish. This region was near the guide's summer cabin, located 12 km away from the mouth of the stream where he thought the fish population originated. These personal testimonies suggest the presence of relatively fresh water far away from any major stream and also distant from even relatively small streams.

4.3. Driftwood

The location of driftwood in the region is another aspect of the information gathered from local people that can be used as a passive tracer of ocean currents. In the past, other authors like Sverdrup (1929), Dyke et al. (1997) used observations of driftwood found on the beaches as passive tracers for ocean circulation. Through the analysis of radiocarbon dating of Holocene driftwood from the CAA, Dyke et al. (1997) discussed changes in

Arctic Ocean circulation and inferred an eastward transport through the CAA. The driftwood originated from boreal regions either in Russia or North America. The driftwood that washed up onto the shore was, and still is in some places, very important for local people. In these high latitude regions there are no trees so wood can become a combustible treasure, and the knowledge of where to find it is very important. During the summer cruises conducted in 1999 and 2000, driftwood was observed to be present all along the coast of the CAA. We now discuss its potential origin.

As mentioned by Dyke and Savelle (2000) and by local people, the only source of wood for the southwest of the CAA is Mackenzie River driftwood, since it is the only river that originates far enough south to carry wood. In the summer 2000 field season, the beaches at the mouth of the Mackenzie River were observed to contain much driftwood compared to the deserted shores at the mouth of the Coppermine River and in Chantrey Inlet (1999, cruise 9924). The two latter locations are located at the mouths of the next two largest rivers (after the Mackenzie River) that flow from the Canadian mainland. The logs at the Mackenzie River mouth, some more than 5m long, are from south of the tree line and are carried by the river into the Arctic regions. The Coppermine River carries some small logs that differentiate in size from the logs from the Mackenzie River. These small logs are only a few inches in diameter and are not usually longer than three feet. From the information gleaned from people of Bathurst Inlet, no logs are found in the inlet itself, and from our observations in summer 1999 the situation is the same in Chantrey Inlet. Thus, the driftwood originates from the Mackenzie River and is spread by ocean currents along the coast.

As observed in the summers of 1999 and 2000, and as described by local people, big logs, some more than 5 m long, can be found along the shore at quite specific locations. The density of driftwood on the beaches decreases from the Mackenzie River mouth towards the east, but from the information obtained during the summer of 2000, logs could still be found as far as Queen Maud Gulf, more than 1200km east of the Mackenzie River mouth, as well as on the west coast of Boothia Peninsula, which is even further east.

The distribution of the logs from the Mackenzie River mouth to the east is characterised by a very high density of logs on the west side of Cape Parry — beaches there were described as completely covered by logs, and were compared to the beaches in British Columbia by people who traveled down south. Cape Parry is already more than 400km east of the Mackenzie Delta. No information was collected by us for Amundsen Gulf and Dolphin and Union Strait, but Dyke and Savelle (2000) discussed the presence of much driftwood on the beaches of western Wollaston Peninsula on the southwestern side of Victoria Island. They reported wood deriving mostly from conifers originating from the Mackenzie River. To explain the transport of this driftwood more than 700 km east of the Mackenzie River mouth, the authors suggested the existence of an eastward-flowing coastal current on the inner continental shelf forced by buoyancy input from the Mackenzie River — they termed this the *Mackenzie Current*. When not disturbed by the opposing winds, the current flows eastward through Amundsen Gulf, Dolphin and Union Strait, and then Coronation Gulf.

The traditional knowledge acquired during the summer of 2000 suggests that logs are frequently found on the south shore of Coronation Gulf. In addition, logs are usually found on the western side of the islands north of Bathurst Inlet on the south side of the Gulf. This is also true for the western and northern coasts of Kent Peninsula, as well as the southern shore of Dease Strait, near Cape Alexander (for example). At the gravel pit near Cambridge Bay (where whitefish are spotted) located on the other side of Dease Strait and more than 1100 km from the MacKenzie River, single logs can be found quite regularly. Our observations during the summer of 2000 are consistent with traditional knowledge that suggests that driftwood is more often found on the south side of the Strait. For example, we observed and sampled a log on the south shore of Dease Strait at

68°49.8′ N, 105°37.5′ W, southwest of Cape Colburn, at the beginning of the CC hydrographic transect (see Figure 1 and below). From its elevation of less than 1m on the beach, it seems to be less than 100 years old (A. Dyke, Geophysical Survey of Canada, Ottawa, personal communication). Local people going to their summer cabin near 103°W observed other logs on the south shore of Queen Maud Gulf.

The local people also observed logs and bamboo sticks on the shore of Boothia Peninsula, but one could argue that multiyear sea-ice going through MacClintock Channel could have transported logs into these regions. Therefore, in this region the path of the transport is not certain. One interesting characteristic is that in the southwest of the CAA (Coronation Gulf, Dease Strait, and Queen Maud Gulf), logs are usually found on the south shore. One could argue that since the source of the driftwood found on the south shore is the Canadian mainland, which is also to the south, one would expect to find the driftwood there. Nevertheless, the traditional knowledge shows that there is no consistent cross-channel mixing from the south to the north. In addition, as will be shown in the next section, the wind forcing is predominantly from north to south. The presence of the driftwood in the noted regions demonstrates a prevailing eastward ocean transport and suggests there is little cross channel transport to the north. It is also consistent with the existence of a buoyancy forced BBC and the dominant winds from north to south.

The logs found in Dolphin and Union Strait were, however, on the northern shores. As noted earlier, the presence of a very shallow sill in the strait could create instabilities and meandering or recirculation of the eastward current, thereby allowing driftwood to be transported to these regions. Also, wind forcing over the ocean and sea-ice could have transported the driftwood to this particular location. This could also possibly explain the reason why driftwood was occasionally observed by local people at the aforementioned gravel pit on the north side of Dease Strait near Cambridge Bay.

Chapter 5 — Hydrographic Data

5.1. Field work description

Hydrographic data were collected during the summer (August and September) of 1995 (cruise 9533), 1999 (cruise 9924), and 2000 (cruise 2000-33). The vertical profiles of temperature and conductivity versus depth were made using Seabird 19 conductivity-temperature-depth profilers (CTD). Water samples were also collected for the determination of salinity and chemical tracers ($^{18}O/^{16}O$, Barium, and nutrients). The fieldwork in 1995 was done aboard the C.C.G.S. Arctic Ivik, while in 1999 it was conducted aboard the C.C.G.S Sir Wilfried Laurier, and in the summer of 2000 aboard the Saint Roch II. In addition, local guides from Cambridge Bay enabled the collection of data from repeat sections taken aboard the 25 ft boat Ice Dancer. This swift launch allowed higher sampling than in the preceding years, with 16 stations on the Cape Colburn line (CC line — see Figure 1 for location) where only four or five stations were permitted during the previous years aboard the ice-breakers.

The location in Dease Strait near Cambridge Bay for field work 2000-33 was chosen because this region represents the narrowest part of the strait, and as it revealed the presence of relatively fresh water. In addition, a section was taken twice during cruise 9924 and showed very strong variations in hydrographic properties (see below). The proximity of the village of Cambridge Bay also made the region more accessible and suitable to establishing a daily time-series of hydrographic conditions. The weather conditions were also easily assessed by remaining in one location during the field work and observing changes in atmospheric forcing supported by data from the weather station at the nearby airport. The airport data provide a very representative measure of wind conditions over the study area as the topography is so flat. As shown below through both direct observations and climatological data, the wind forcing over the region is quite important for the local hydrographic and current structure.

5.2. Results from hydrographic data

5.2.1. Introduction

The data from cruises 9533 and 9924 revealed the presence of buoyant (fresher) water along the coast, and in many cases showed the potential presence of a BBC. These buoyant fresher waters were found around almost every island. Two good examples of such salinity distributions for Banks Island (from the cruise 9924) are shown in Figures 2 and 3. The former region is part of the Amundsen Gulf section effectuated on September 5th 1999. Near the end of the section, a front was spotted from the upper deck of the icebreaker Sir Wilfried Laurier and a decision was made to add few stations at the end of the line (BK line — see Figure 1 for location).

The salinity distribution at the southern tip of Banks island (Figure 2) shows fresher water nearby the coast. During the following night the icebreaker Sir Wilfried Laurier made its way north through the Prince of Wales Strait with northerly winds bringing multi-year ice from McClure Strait. This tongue of ice was moving south sticking to the western side of the channel as would be expected from simple dynamical considerations. On the 6th of September at 72°N, a section was done from east to west in Prince of Wales Strait (PW line — see Figure 1 for location).

The salinity distribution for this section (Figure 3) shows the presence of fresher water on the western side, revealing the potential presence of a southward BBC accompanying the tongue of ice. Such structures with even weaker gradients, of the order of decimals of psu, are usually associated with BBCs (e.g. Leblond, 1980; Prinsenberg and Bennett, 1973), so that one would expect the presence of quite a strong BBC on each side of the channel. Unfortunately, no current measurements are available for the southwestern part of the CAA. This type of measurement using ship mounted ADCP or moored current meters could confirm and quantify the transport on each side of the channel, and would greatly help to evaluate the net transport. Even though this type of data is not available, the speed of the BBC on each side of the channel can be estimated, assuming a geostrophic balance (described below), from the density distribution to be 0.21 cm/s for the south coast of Banks Island, and 0.38 cm/s on the west side of Prince of Wales Strait.

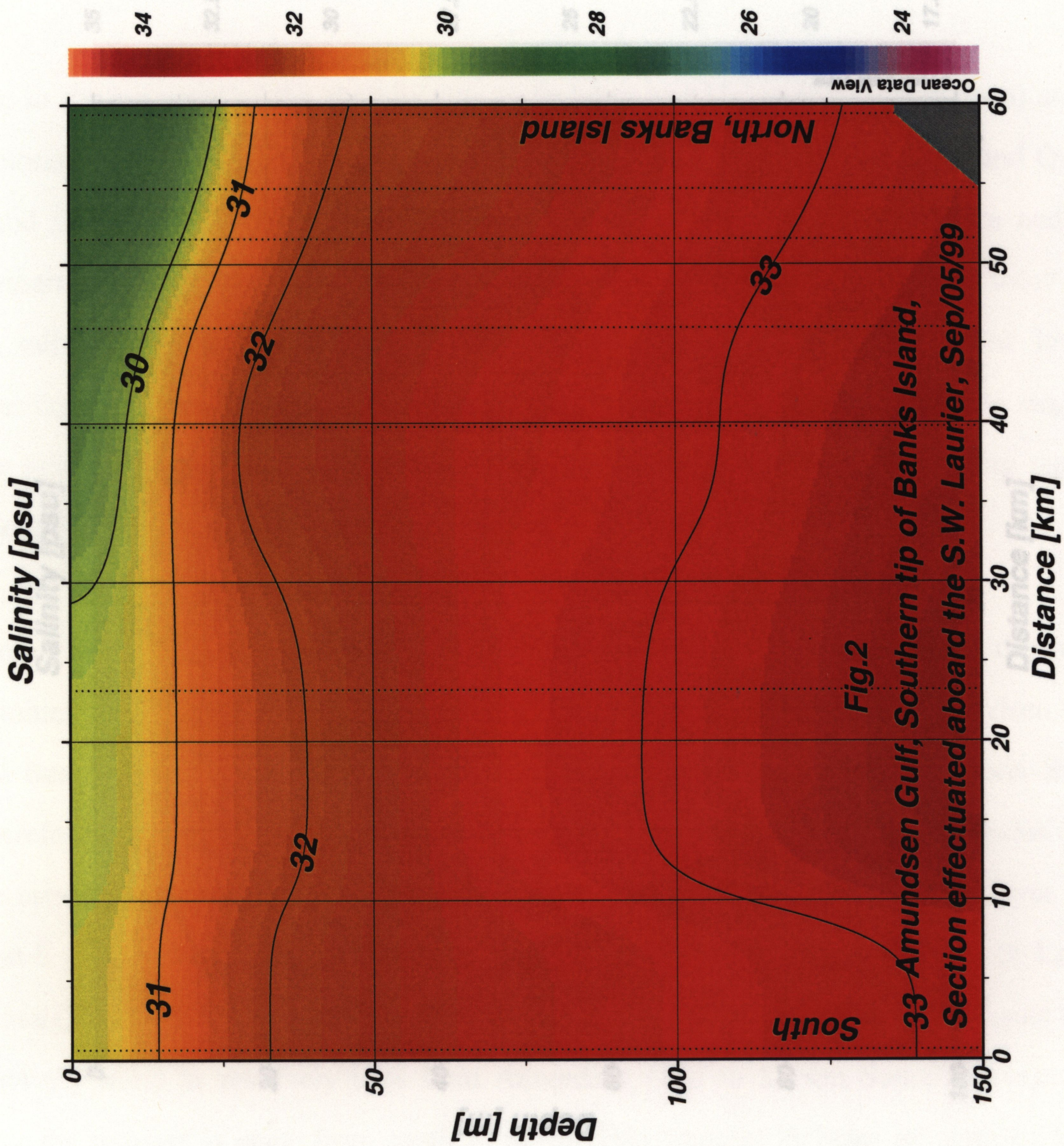

Figure 2: South Banks Island salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 5th, 1999

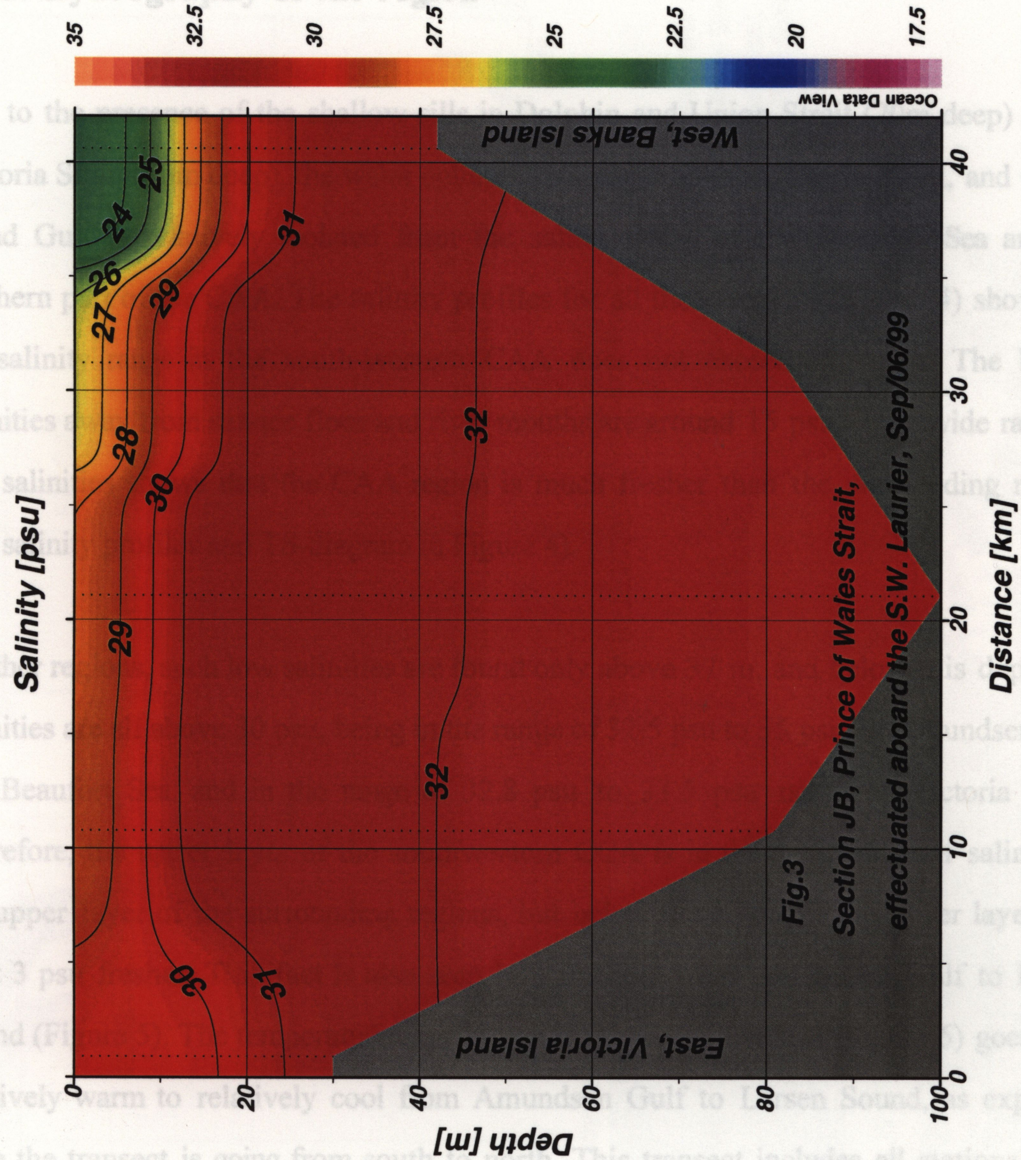

Figure 3: Prince of Wales salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 6th, 1999

5.2.2. Hydrography of the region

Due to the presence of the shallow sills in Dolphin and Union Strait (20m deep) and in Victoria Strait (40m deep), the water column in Coronation Gulf, Dease Strait, and Queen Maud Gulf is relatively isolated from the saltier water of the Beaufort Sea and the northern part of the CAA. The salinity profiles for all these regions (Figure 4) show that the salinity range in the south-western CAA does not exceed 29.5 psu. The lowest salinities away from sea-ice floes and river mouths are around 15 psu. This wide range of low salinities shows that the CAA region is much fresher than the surrounding regions (see salinity profiles and TS-diagram in Figure 4).

In other regions, such low salinities are found only above 37 m, and below this depth the salinities are all above 30 psu, being in the range of 32.5 psu to 35 psu in Amundsen Gulf and Beaufort Sea, and in the range of 32.8 psu to 33.5 psu north of Victoria Strait. Therefore, the upper layer of the southwestern CCA is in the same range of salinity as the upper layer of the surrounding regions, but below the halocline the lower layer is at least 3 psu fresher. This fact is also seen in a transect from Amundsen Gulf to Larsen Sound (Figure 5). The temperature distribution for the same transect (Figure 6) goes from relatively warm to relatively cool from Amundsen Gulf to Larsen Sound, as expected, since the transect is going from south to north. This transect includes all stations taken during cruise 9924 through this part of the Northwest Passage. To construct Figures 5 and 6 all data from cruise 9924 (summer 1999 only) were included, taking into account the cross-channel differences from the west to the east and thereby representing an average of the fresh water distribution. The results agree with those of Figure 4, showing that the lower layer is fresher in the south-west CAA. This points out that only the upper layer moves easily through the region, with the lower layer transport being restricted by the presence of sills.

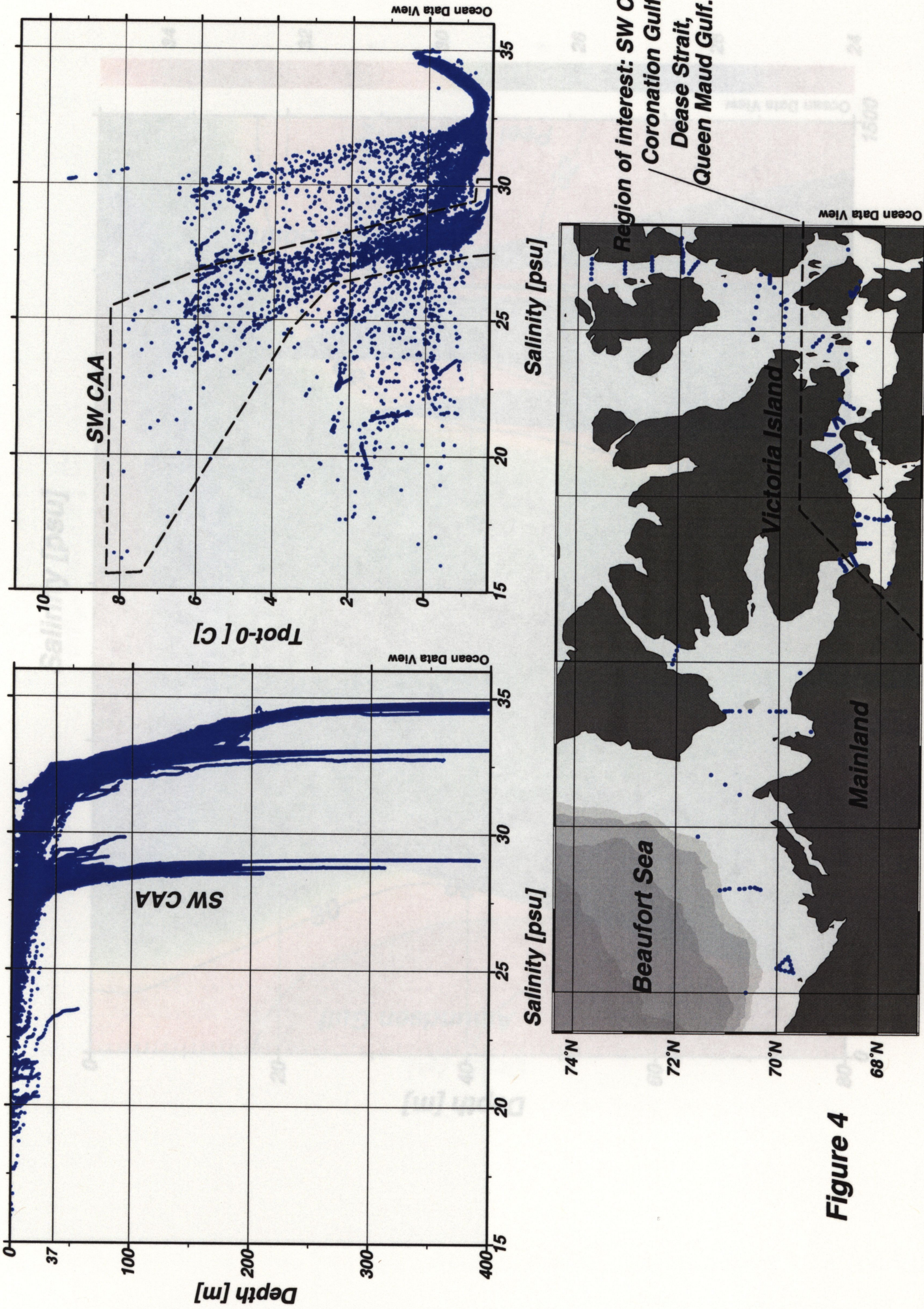

Figure 4: Salinity profiles and TS diagram for the southwestern CAA

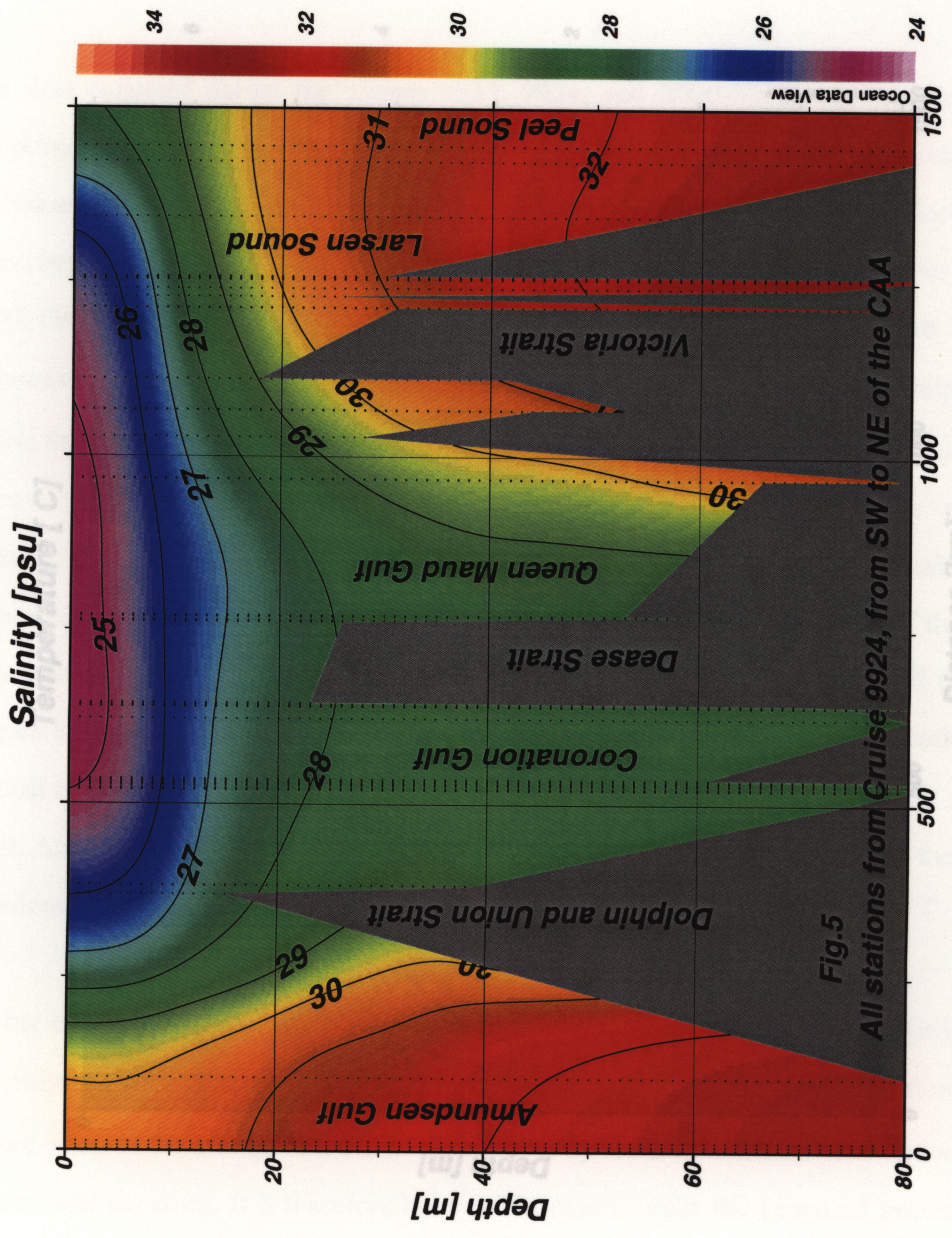

Figure 5: Salinity distribution over the southwestern CAA during the cruise 9924.

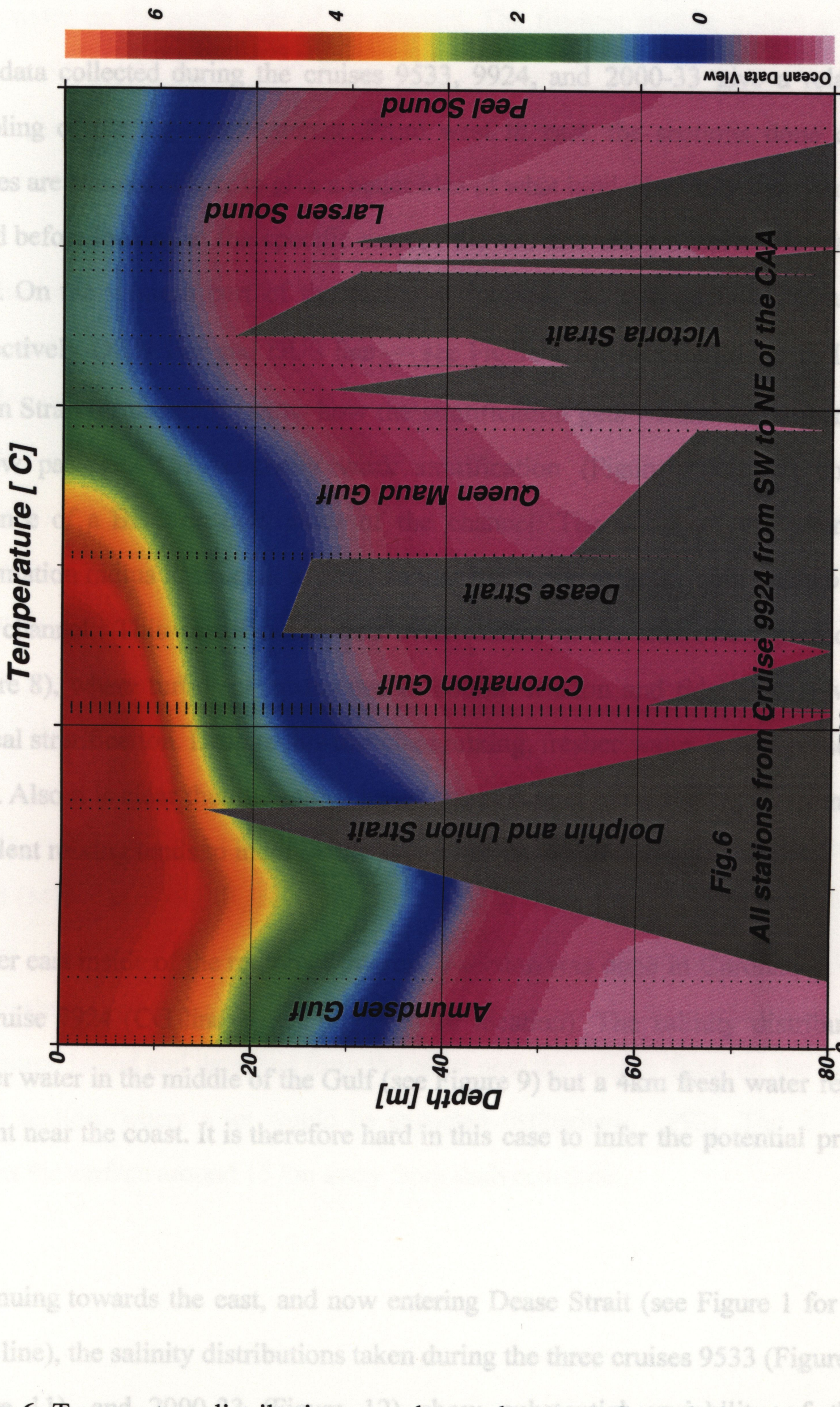

Figure 6: Temperature distribution over the southwestern CAA during the cruise 9924.

5.2.3. Results from hydrographic sections

The data collected during the cruises 9533, 9924, and 2000-33 give a relatively good sampling of the region of interest. From west to east, the sections done during these cruises are examined here to give a better idea of what kind of salinity distributions can be found before looking at the specific case of the CC line using observations from summer 2000. On the western part of the region of interest, the two sections (Figure 7 and 8, respectively DUN line and DUS line — see Figure 1 for location) done in Dolphin and Union Strait (cruise 9533) show how the stratification gets weaker when going through a narrow passage. The relatively weak stratification (Figure 7) shows the potential presence of a BBC on both sides of the channel. The width of the internal Rossby deformation radius R' is equal to 3km, so that BBCs can exist on both sides of the 30-km wide channel . This stratification gets even weaker in the shallowest part of the strait (Figure 8), where turbulent mixing due to bottom friction and tidal forcing weakens the vertical stratification. Despite this turbulent mixing, fresher water is still present near the coast. Also it is clear that the saltier lower water cannot enter the region even though the turbulent mixing tends to increase the salinity of the upper surface.

Further east inside of the region of interest, a section was done in Coronation Gulf during the cruise 9924 (CG line — see Figure 1 for location). The salinity distribution shows fresher water in the middle of the Gulf (see Figure 9) but a 4km fresh water region is also present near the coast. It is therefore hard in this case to infer the potential presence of a BBC.

Continuing towards the east, and now entering Dease Strait (see Figure 1 for location of DSW line), the salinity distributions taken during the three cruises 9533 (Figure 10), 9924 (Figure 11), and 2000-33 (Figure 12) show substantial variability of the salinity distribution from year to year, even though these data were not collected at the exact same location. During the summers of 1995 and 2000, the sections show the presence of fresh water on the south side of the channel. The freshest surface waters are around 23 psu in 1995 and 21 psu in 2000. During the summer of 1999, the freshest surface waters are on the north side and are around 23 psu. In the three cases, the scale of the BBC associated with the salinity distribution is around 15km.

Still further eastward, data were taken near Cambridge Bay during the summers of 1995, 1999, and 2000 near the narrowest part of Dease Strait. As mentioned earlier, the CTD profiles were taken a few days apart on the CC line (see Figure 1 for location) in 1999, and the results were very different (Figs. 13 and 14), showing strong variability in hydrographic properties. On the 7th of September the freshest surface waters were located to the north, near the shore of Victoria Island, with the 25.6 psu isoline reaching the surface 10km away from the coast. On the 11th of September, the freshest surface waters were located to the south, near the Mainland shore. The 25 psu isoline reached the surface about 5km away from the coast, scaling well with R' (see below).

Finally, in Queen Maud Gulf, a section done between Melbourne Island and Victoria Island (ML line — see Figure 1 for location) on August 12th 1995, during cruise 9533 (Figure 15), shows the presence of relatively fresh water on both sides of the channel. The freshest surface waters (around 23 psu) are located to the south near the shore of Melbourne Island and detached from the coast near the salinity front on the northern side of the section, where a jet would be expected (Münchow, 1992). The 27 psu isoline reaches the surface around 15 km away from each coastline.

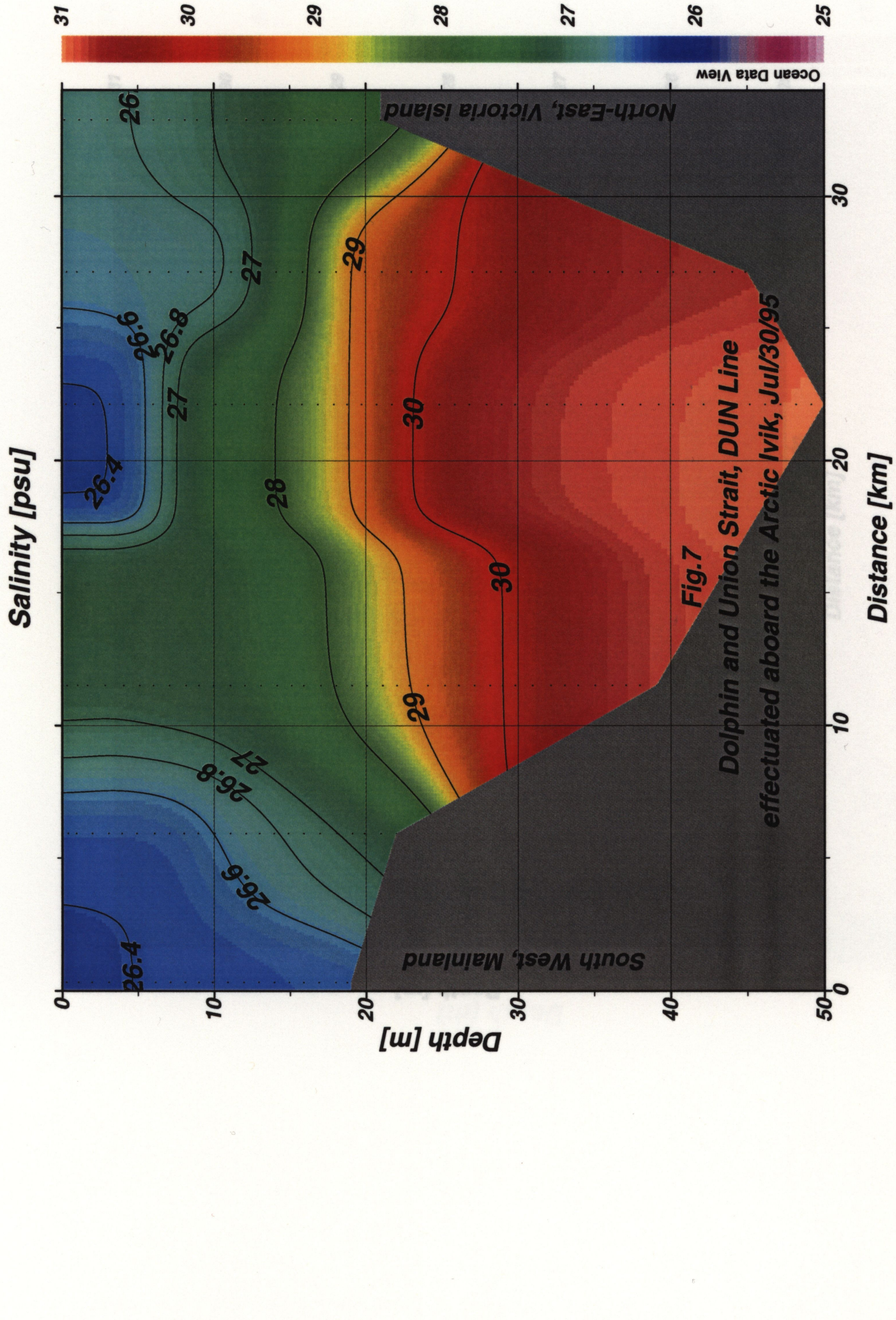

Figure 7: Dolphin and Union Strait northern salinity section collected aboard the C.C.G.S. Arctic Ivik on July 30th, 1995.

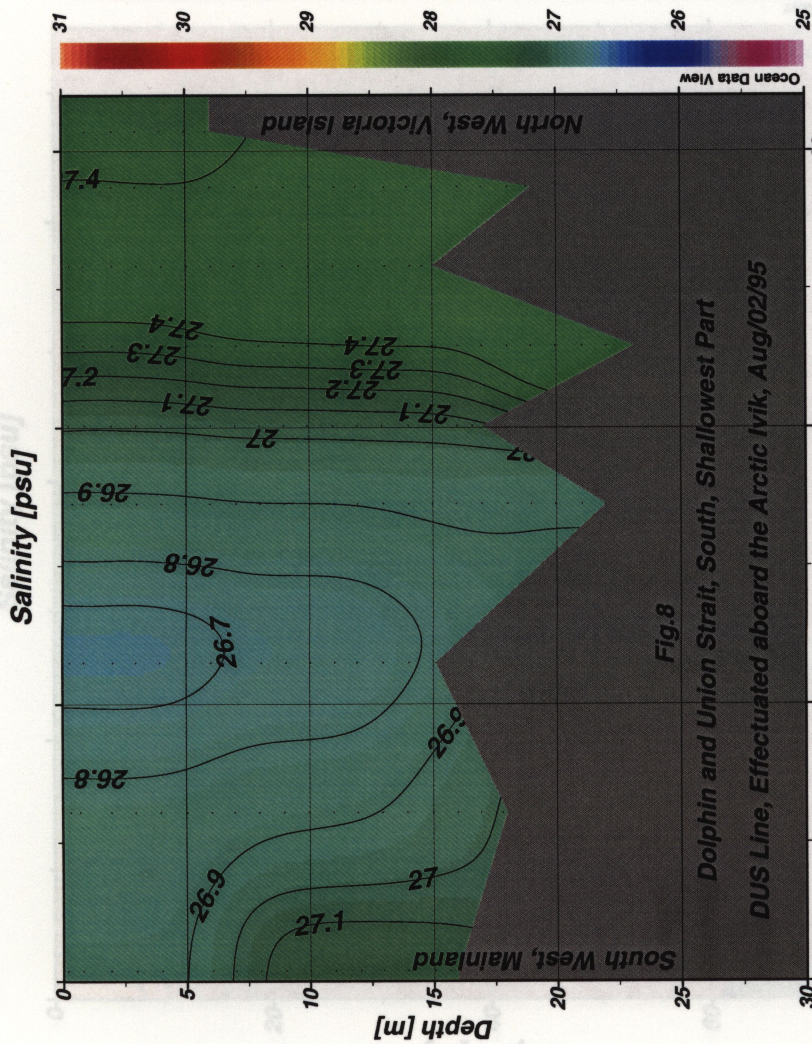

Figure 8: Dolphin and Union Strait southern salinity section collected aboard the C.C.G.S. Arctic Ivik on August 2nd, 1995.

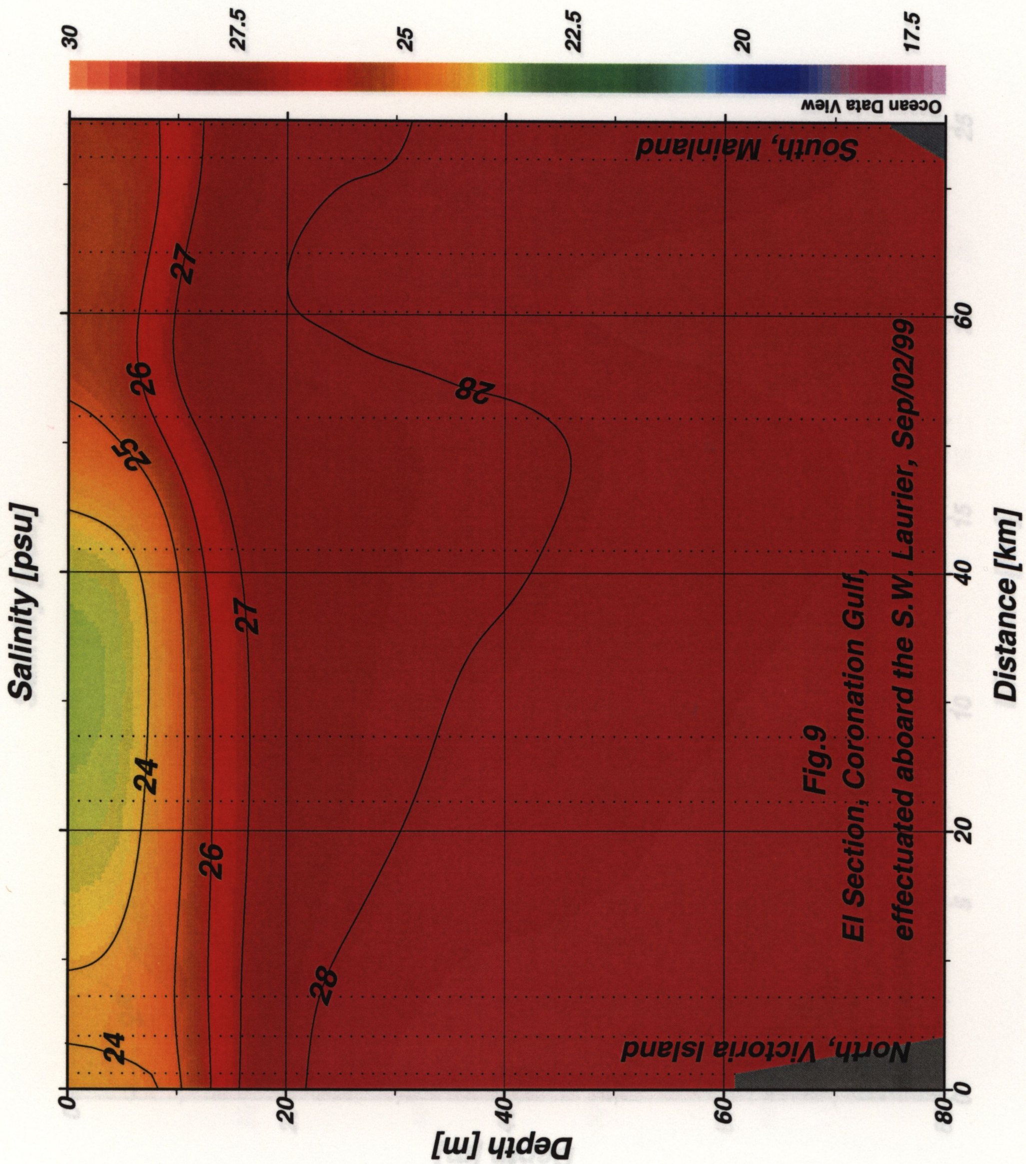

Figure 9: Coronation Gulf salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 2nd, 1999.

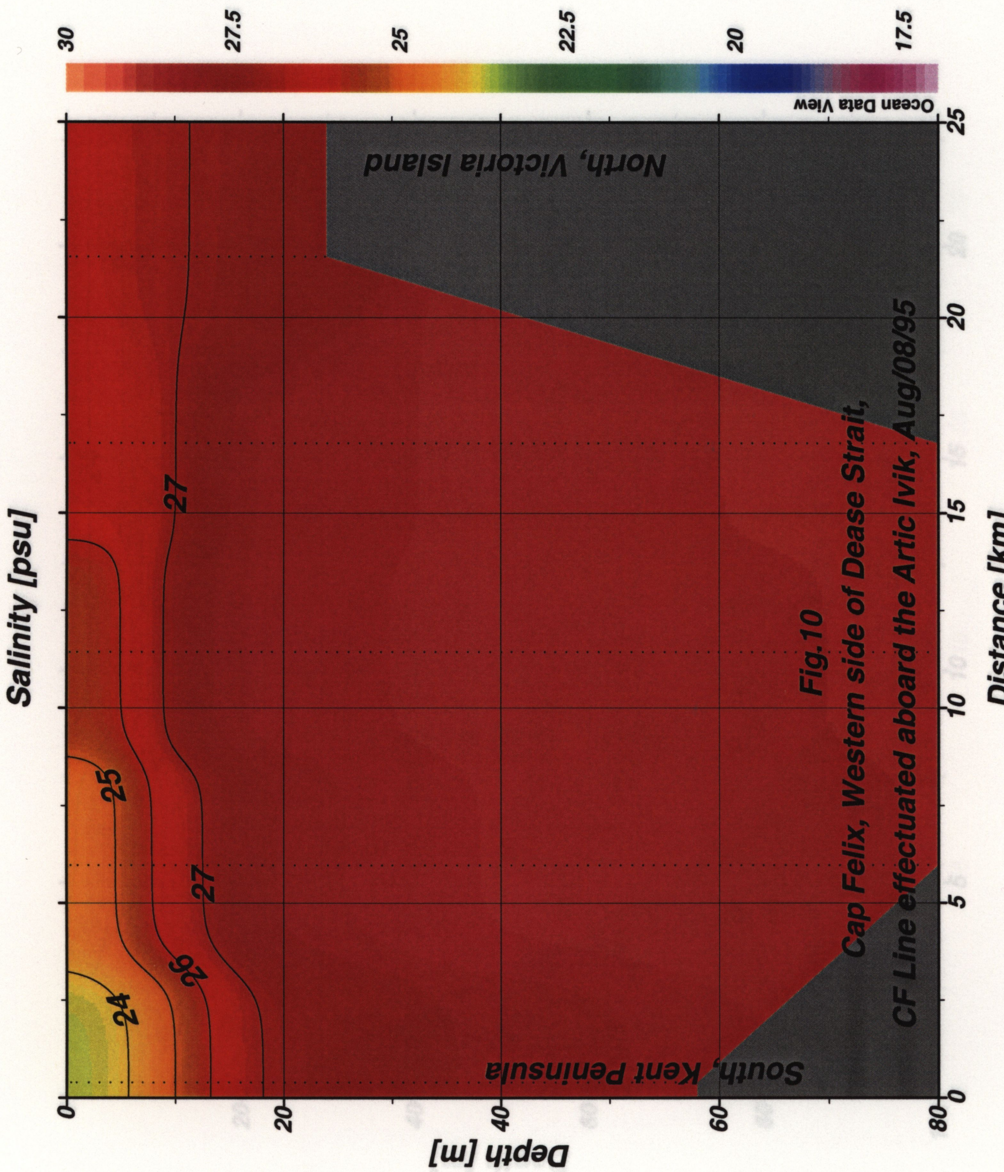

Figure 10: Dease Strait West salinity section collected aboard the C.C.G.S. Arctic Ivik on August 8th, 1995.

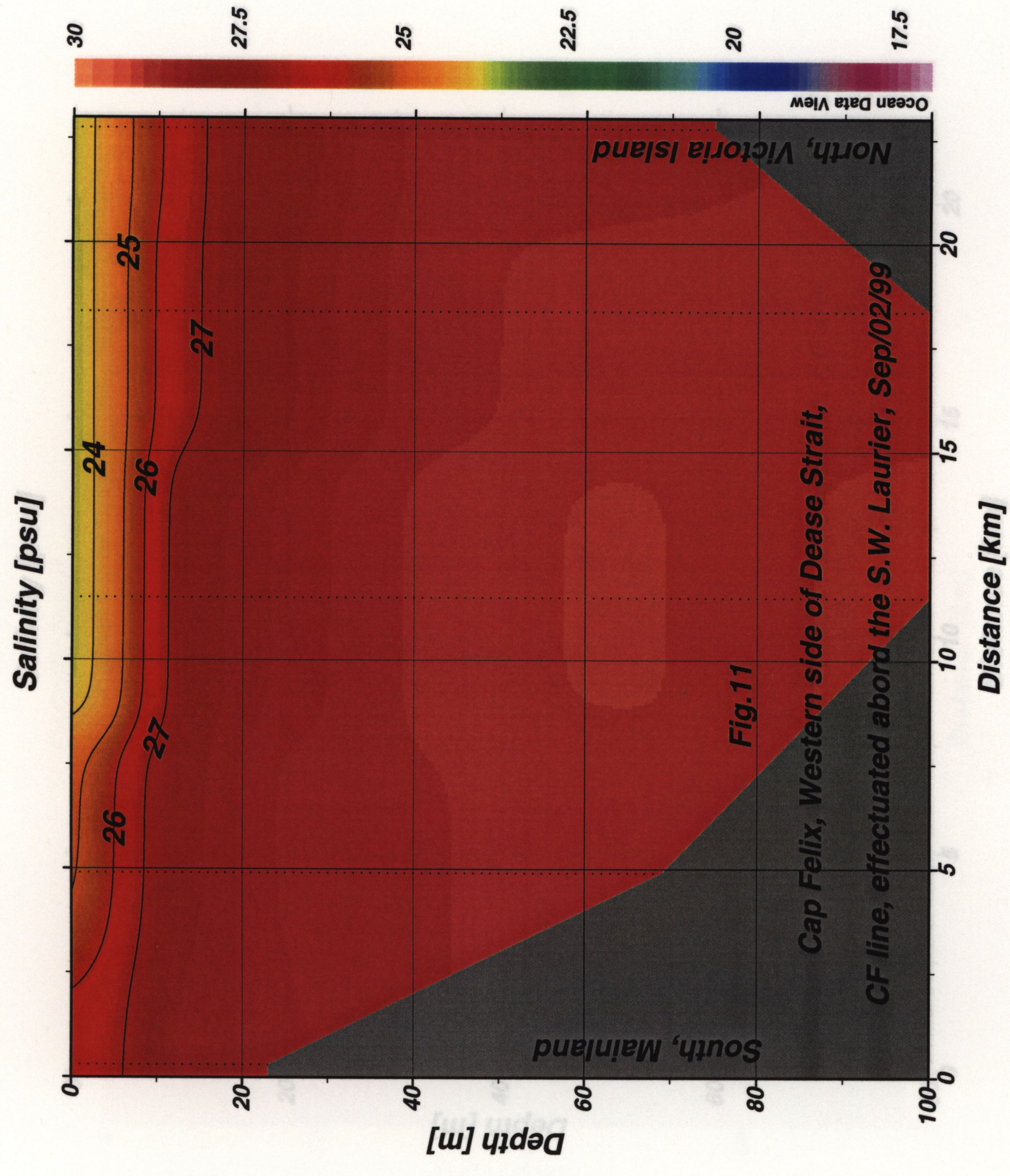

Figure 11: Dease Strait West salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 2nd, 1999.

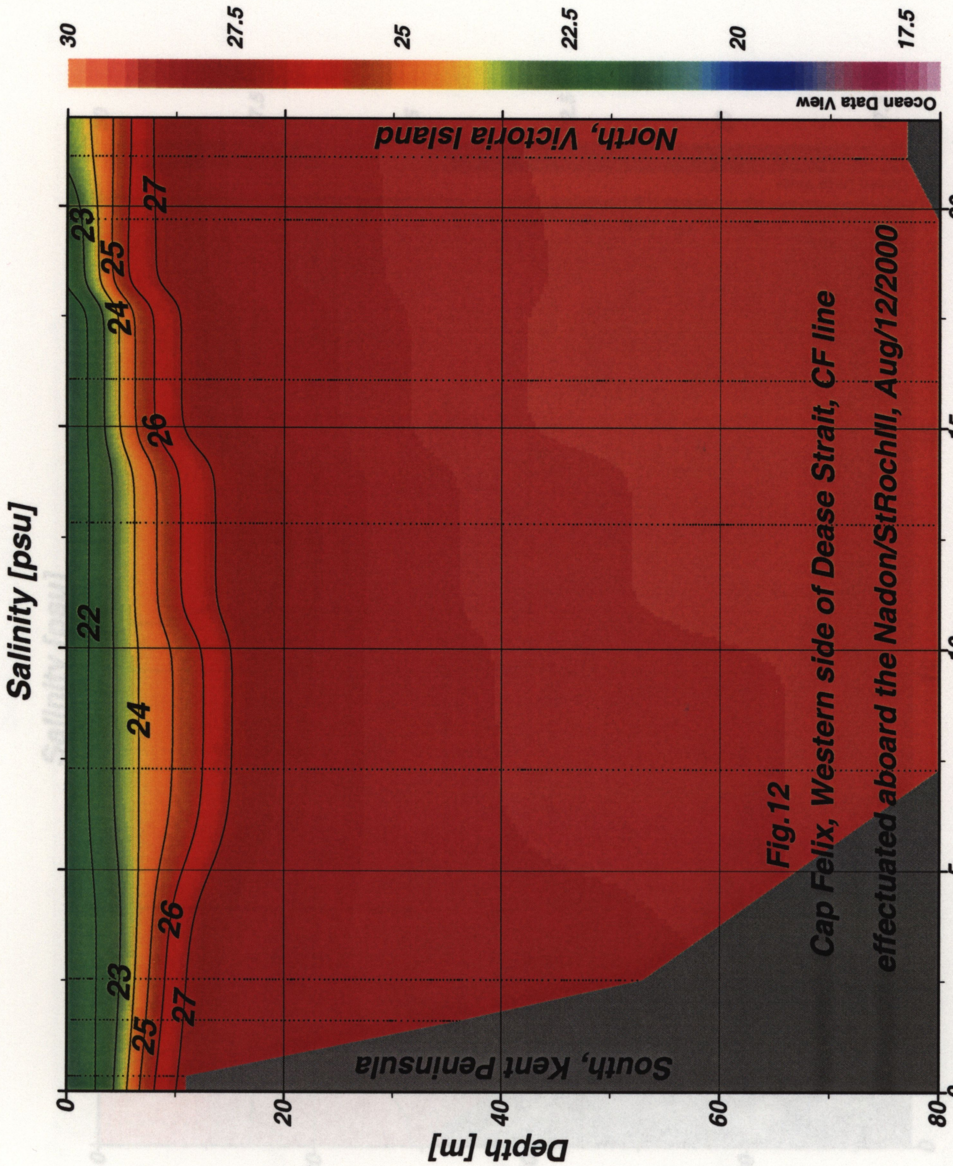

Figure 12: Dease Strait West salinity section collected aboard the S^{t} Roch II / Nadon on August 12th, 2000.

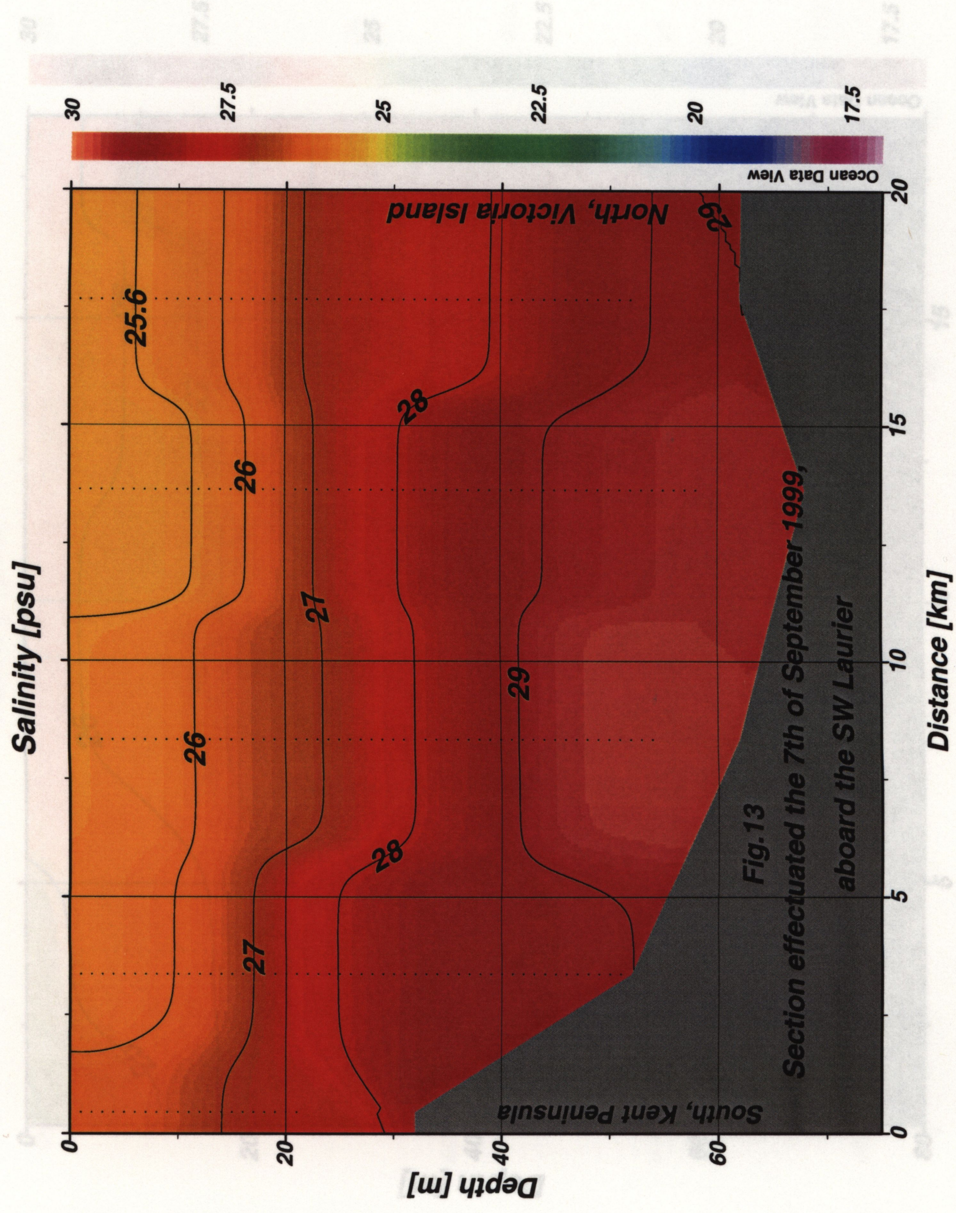

Figure 13: Cape Colburn salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 7th, 1999.

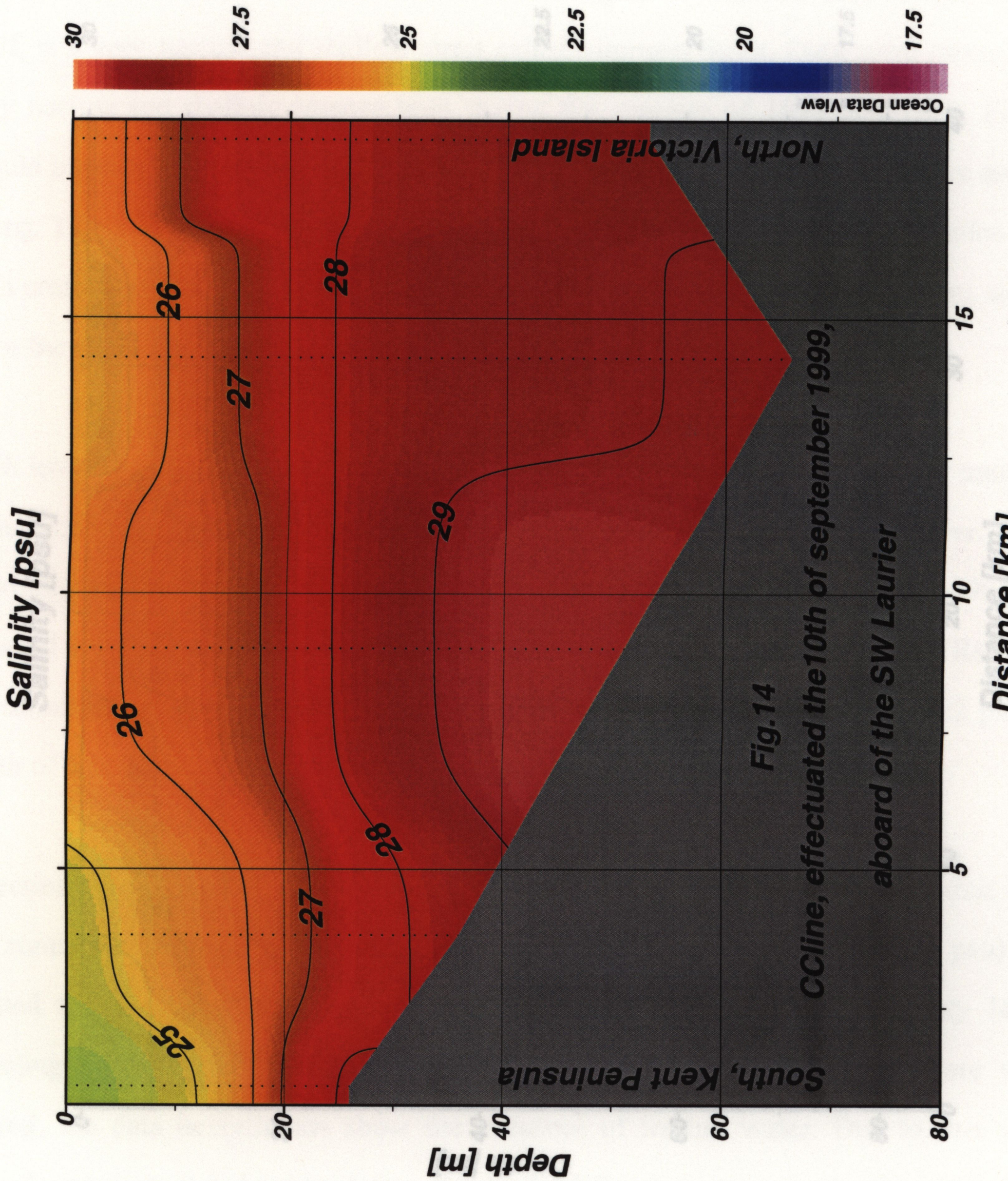

Figure 14: Cape Colburn salinity section collected aboard the C.C.G.S. Sir Wilfred Laurier on September 10th, 1999.

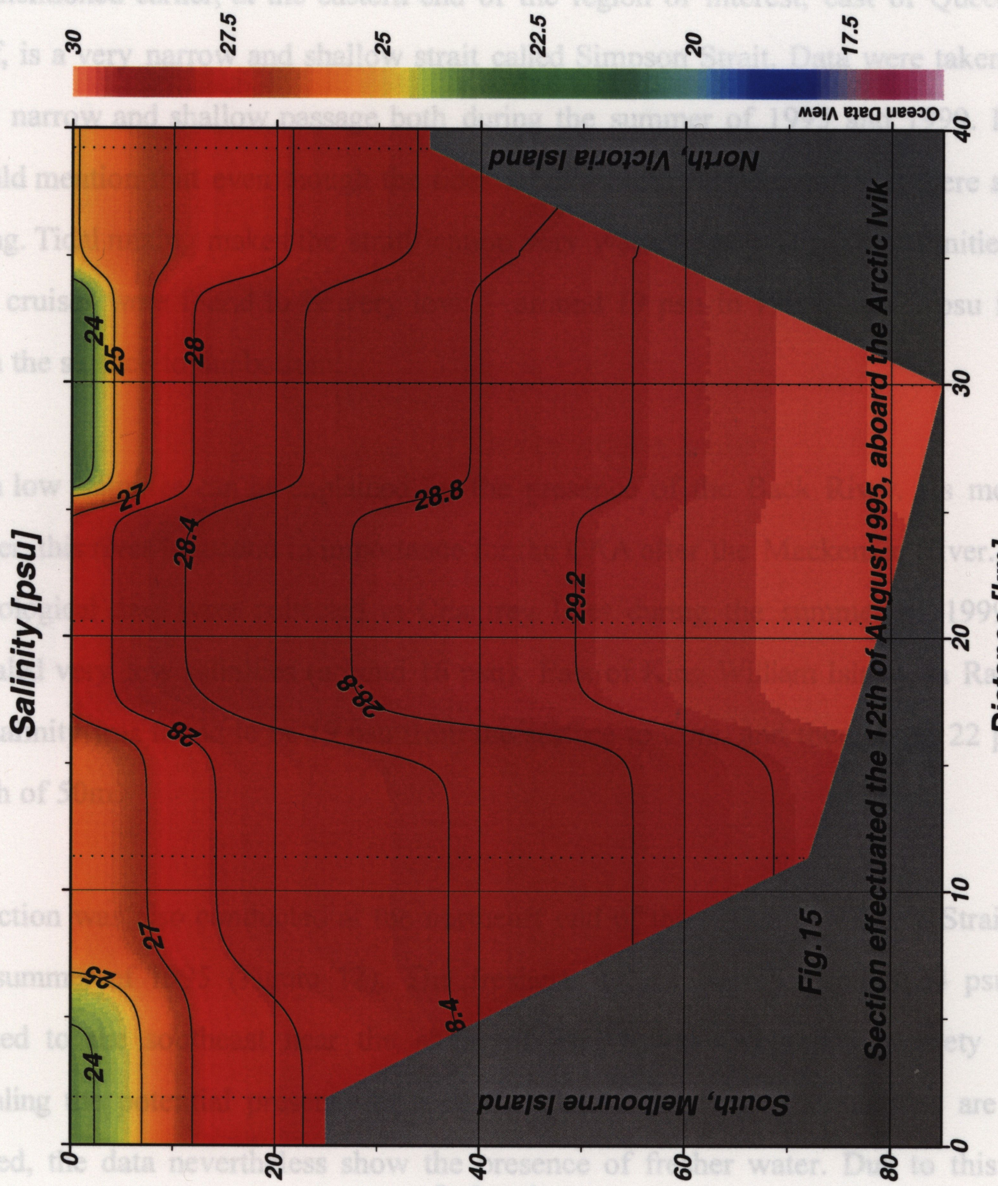

Figure 15: Melbourne Island salinity section collected aboard the boat Arctic Ivik on August 12th 1995.

As mentioned earlier, at the eastern end of the region of interest, east of Queen Maud Gulf, is a very narrow and shallow strait called Simpson Strait. Data were taken in this very narrow and shallow passage both during the summer of 1995 and 1999. First we should mention that even though the tides are not high, the tidal currents there are quite strong. Tidal mixing makes the stratification very weak (Figure 16). The salinities during both cruises were found to be very low — around 19 psu in 1995, and 21 psu in 1999, from the surface to the bottom.

Such low salinities can be explained by the presence of the Back River. As mentioned earlier, this river is second in importance for the CAA after the Mackenzie River. Further hydrological data were collected in Chantrey Inlet during the summer of 1999 which revealed very low salinities (around 16 psu). East of King William Island, in Rae Strait, the salinity was found to be 19 psu from the surface to 20m, and then about 22 psu to a depth of 50m.

A section was also conducted at the northeast end of the region in Victoria Strait during the summer of 1995 (Figure 17). The freshest surface waters (around 24 psu) were located to the southeast near the shore of The Royal Geographic Society Islands, revealing the potential presence of a BBC. While in this case the samples are widely spaced, the data nevertheless show the presence of fresher water. Due to this widely spaced sampling, it is hard to define clearly the BBC, but we estimate the width of the expected BBC (20km) to be 5 times bigger than the internal deformation radius (R'=4km). The problems associated with widely-spaced sampling motivated the field work done during the summer of 2000, where a swift launch allowed higher sampling.

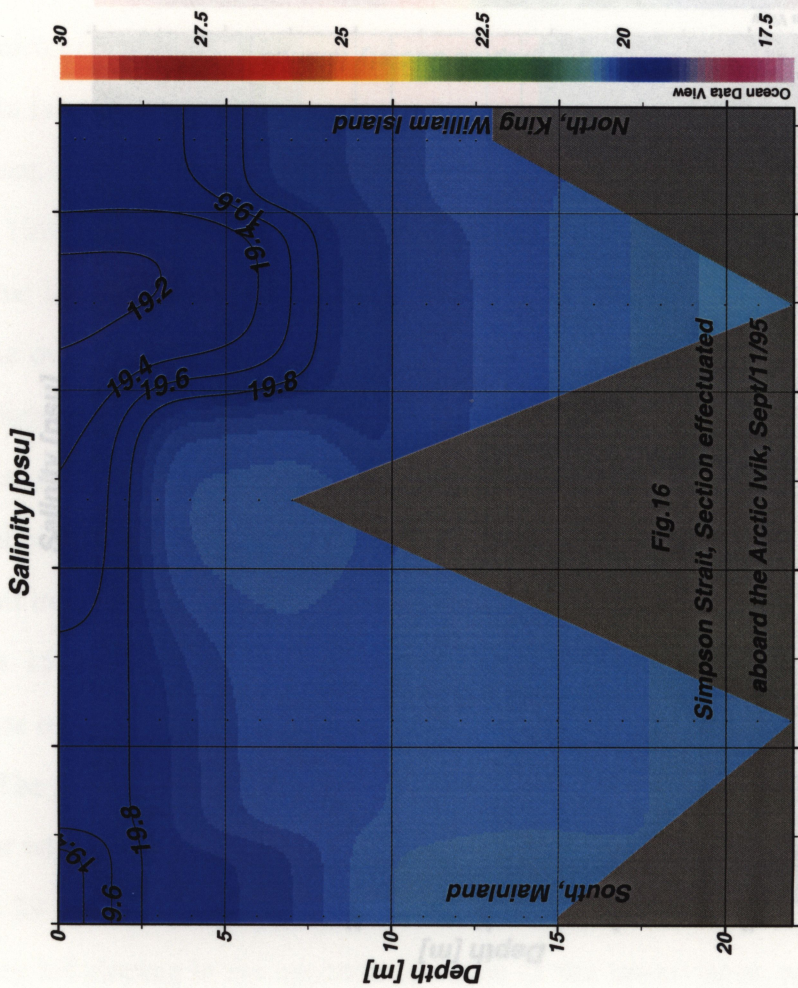

Figure 16: Simpson Strait salinity section collected aboard the boat Arctic Ivik on September 11th 1995.

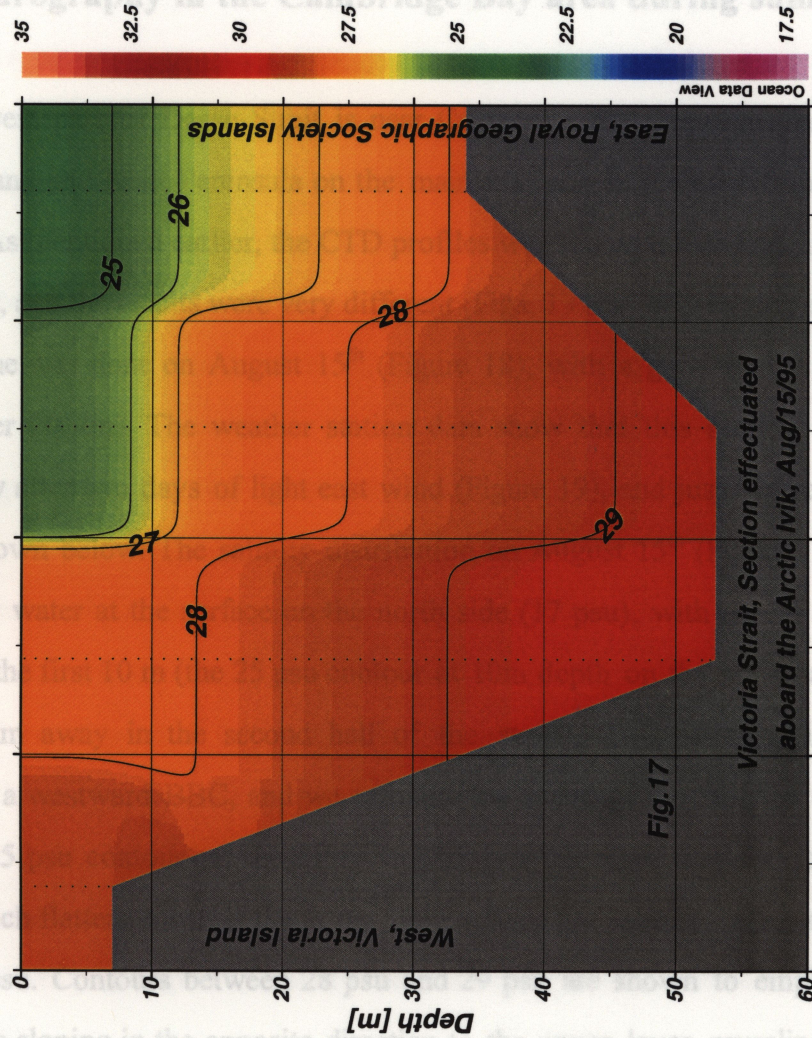

Figure 17: Victoria Strait salinity section collected aboard the boat Arctic Ivik on August 15th 1995.

5.2.4. Hydrography in the Cambridge Bay area during summer 2000

The narrowest part of Dease Strait is near Cambridge Bay between Cape Colburn on Victoria Island and Kent Peninsula on the mainland, and is oriented from northwest to southeast. As mentioned earlier, the CTD profiles were taken a few days apart on the CC line in 1999, and the results were very different (Figs. 13 and 14). During cruise 2000-33, the same line was done on August 15th (Figure 18), with a good spatial resolution (16 stations over 20 km). The weather station data show that this section was conducted immediately after two days of light east wind (Figure 19), and just before a strong wind event as shown below. The salinity distribution on August 15th (Figure 18) shows very low salinity water at the surface on the north side (17 psu), with a very strong salinity gradient in the first 10 m (the 25 psu contour at 10m depth on the north side reaches the surface 15km away in the second half of the strait). This stratification shows the presence of a westward BBC, and we estimate the speed of the BBC to be around 0.24 cm/s. The 25 psu contour on the other side slopes the other way as does the 27 psu contour which flattens towards the north. Below 20m, the salinity varies slowly from 28 psu to 29 psu. Contours between 28 psu and 29 psu are shown to emphasize the fact that they are sloping in the opposite direction to the upper layer, revealing a background state (i.e., the stratification at depth looks very similar before and after the wind event).

The next two days were very windy (Figure 19) and the strong west and northwest winds did not allow for field work to be conducted aboard the boat. The wind conditions were upwelling favourable on the north side and downwelling favourable on the south side of the channel. On August 18th the winds were light and the hydrographic data showed very different conditions in the upper layer from two days earlier (Figure 20, c.f., Figure 18). Below 20 m the stratification was, however, still similar. In the upper layer the wind forcing redistributed the salinity through turbulent mixing, lateral mixing, and changes in upwelling/downwelling such that the westward BBC disappeared. The new salinity distribution sloped the opposite way from two days earlier, showing the presence of a reversed, eastward BBC, with a horizontal salinity gradient of the same order as the one found in Figure 18. The freshest surface waters (around 22 psu) were now located near the middle of the strait and, as mentioned before, in the area near Melbourne Island —part of the section that would be expected to support the jet of the BBC. The 24 psu contour starts around 15 m deep on the south shore and reaches the surface almost at the other side of the strait. The 27 psu contour slopes the same way as before (Figure 18) but is less flat in the northern half of the section and, like the 28 psu contour, slopes the same way as in the background state mentioned earlier. The wind event weakened the buoyancy gradient through cross-channel mixing, but it did not mix the entire water column, leaving the deeper (below 20m) stratification nearly the same. It did, however, reverse the conditions favouring the formation of an eastward BBC, that we estimate to have a speed of 0.17cm/s which is, like the above estimated BBC speeds, of the same order of observations done at lower latitudes with weaker salinity gradients (e.g. Münchow, 1992). This kind of salinity distribution and horizontal salinity gradient is similar to what was observed in other parts of the CAA in the scientific cruises mentioned above (e.g. Figure 3) and in other work (e.g. Leblond, 1980 or Prinsenberg and Bennett, 1973).

Other wind events giving similar distributions happened again during the following two weeks. The Environment Canada climatological wind data for the 30-year period 1961-1990 (Table 1) clearly show that the dominant wind direction is northwest, with a yearly average of 21 km/h and extreme hourly peaks of more than 100 km/h in October. As mentioned earlier, the orientation of Dease Strait in this region is from northwest to southeast; therefore downwelling is favourable on the south side of the strait and upwelling is favourable on the north side. The presence of the same salinity stratification at depth before and after the wind events (i.e., the background state), together with the wind data showing the dominating northwest wind in this region, show that the situation observed on August 18th is certainly the most frequent one. Buoyancy forcing and wind forcing act together to force the formation of a BBC near the south shore and thus favour an eastward transport.

Wind Speed (km/h)	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**	**Oct**	**Nov**	**Dec**	**Year**
Monthly mean speed	23	22	21	20	21	20	20	22	23	23	21	21	21
Most frequent direction	NW	NW	NW	NE	NW	NW	N	W	N	NW	NW	NW	NW
Extreme hourly speed	89	89	84	80	80	93	71	79	87	101	82	97	
Direction	N	NW	NW	N	NW	NW	SE	N	NE	NE	NW	N	
Extreme gust speed	108	109	97	102	102	120	93	109	116	121	102	122	
Direction	N	NW	NW	N	NW	N	SE	W	NE	NE	NE	N	

Table 1: Climatological wind data for Cambridge Bay Airport (1961-1990) as obtained from the Meteorological Service of Canada.

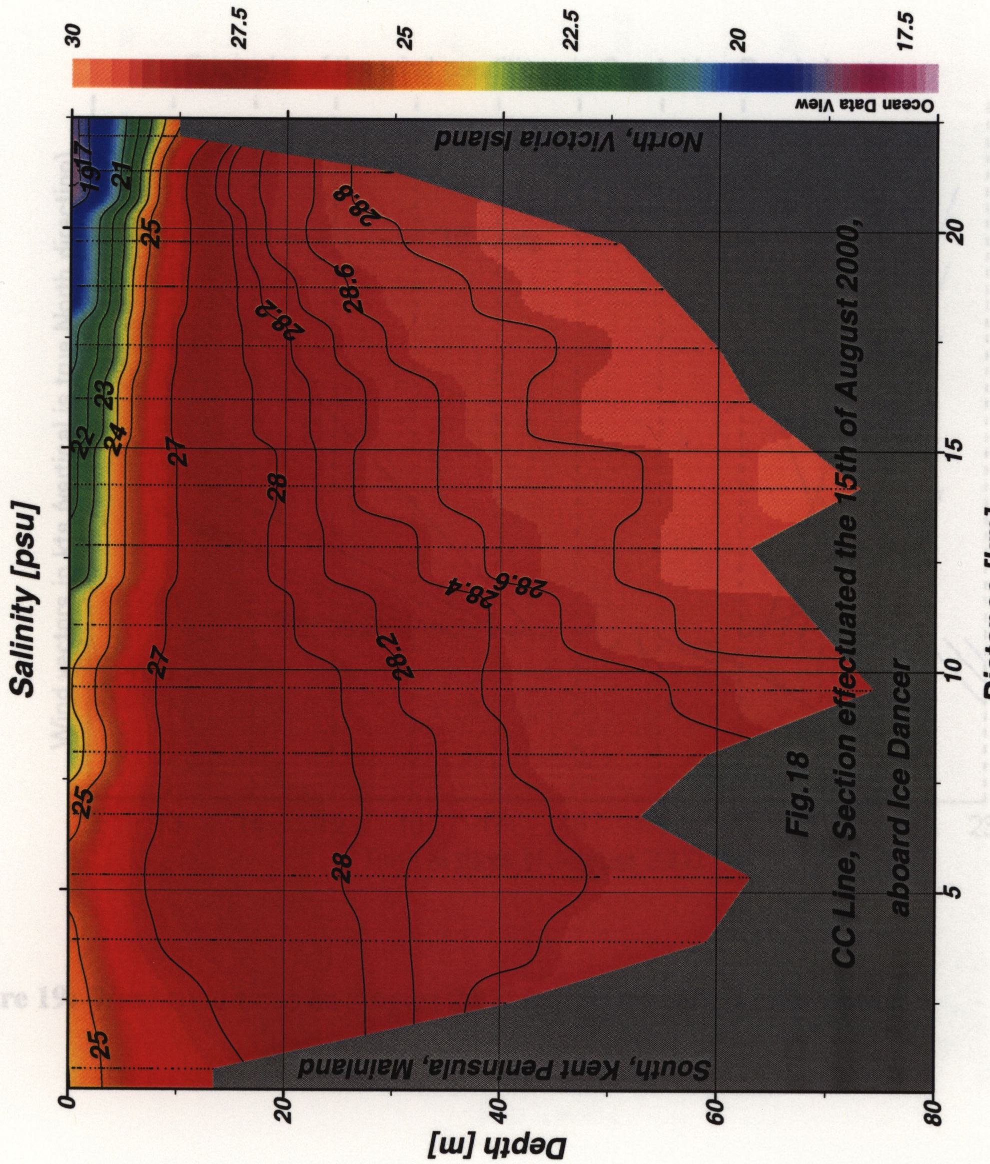

Figure 18: Cape Colburn salinity section collected aboard the boat Ice Dancer on August 15th 2000.

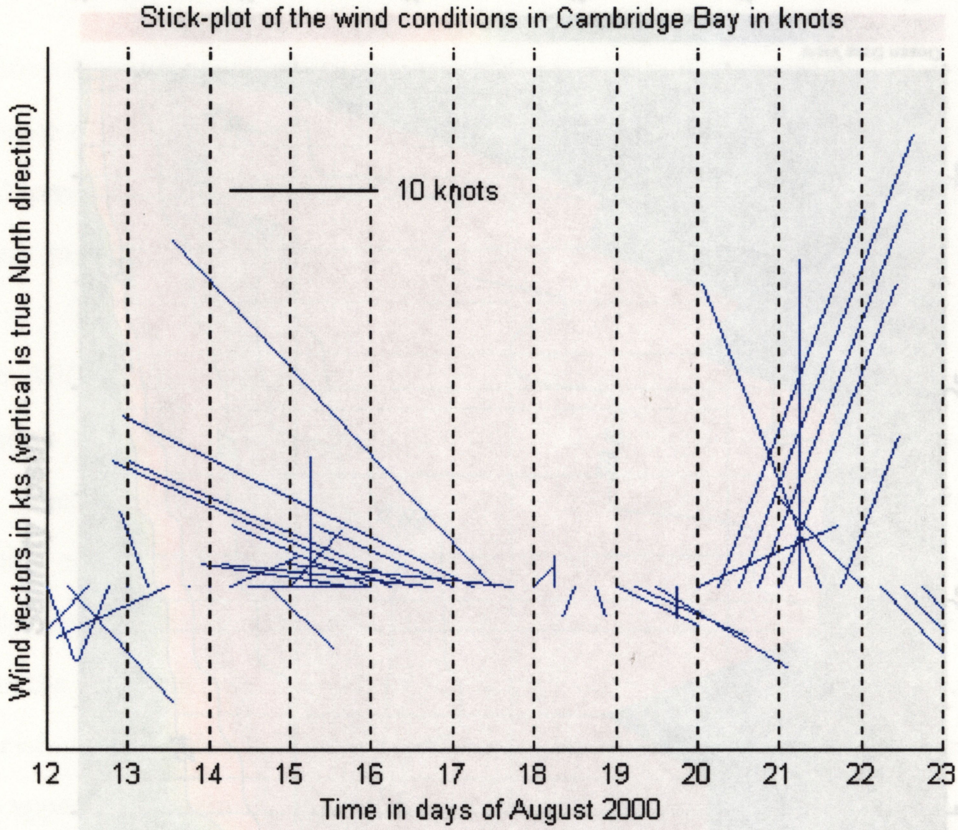

Figure 19: Wind conditions on August 2000 during the period of sampling.

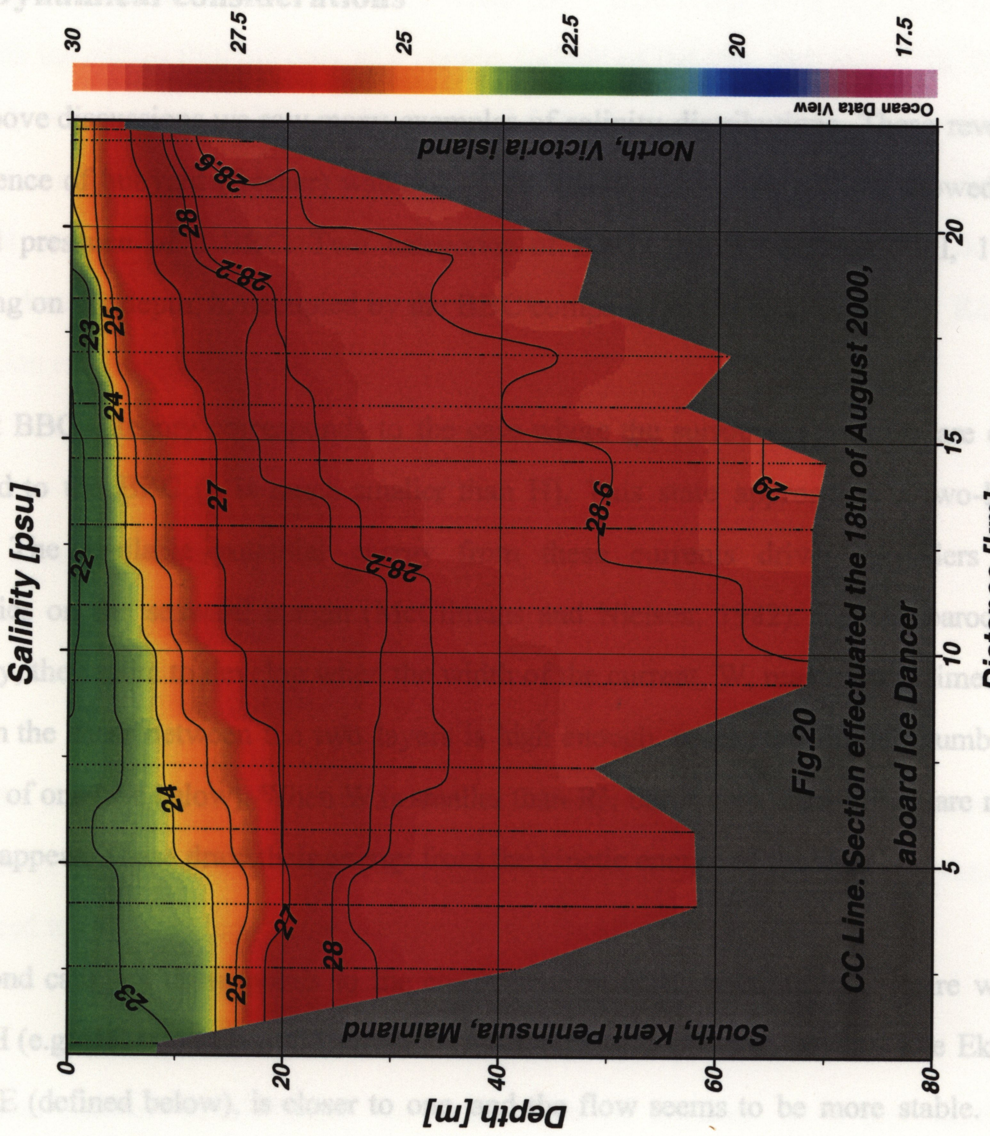

Figure 20: Cape Colburn salinity section collected aboard the boat Ice Dancer on August 18th 2000 after a strong wind event.

5.2.5. Dynamical considerations

In the above discussions we saw many examples of salinity distributions. These revealed the presence of buoyant (fresher) water along the coast, and in many cases showed the potential presence of a BBC. Two categories of BBCs can be defined (Hill, 1998) depending on the depth, h, occupied by the BBC compared to the total depth, H.

The first BBC category corresponds to the case where the surrounding waters are deep compared to the BBC (h is much smaller than H). This state approaches a two-layer system. The available potential energy from these currents drives meanders and instabilities on the seaward margin (McClimans and Nielsen, 1982). Due to baroclinic instability, they seem to develop when the width of the current, W, reaches two times R', and when the shear between the two layers is high enough, unless the Froude number is the order of one (see below). When W is smaller than R', barotropic instabilities are more likely to appear. These draw their energy from the kinetic energy of the flow.

The second category corresponds to the case where the BBC occupies the entire water column H (e.g. Delaware Coastal Current — Münchow and Garvine, 1993a). The Ekman number, E (defined below), is closer to one, and the flow seems to be more stable. The bottom friction certainly contributes to the stabilizing effect (Münchow and Garvine, 1993b). To evaluate any of the following quantities one must first evaluate the width W and depth h of the BBC. To evaluate these, density distributions were examined for each section. First the depth of the strongest change in density (defined as the pycnocline) was obtained and then h was chosen to be the depth at the coast of the isopycnal corresponding to the upper limit of the pycnocline. The characteristic width W was estimated by determining the distance from the coast to where this isopycnal met the surface.

Upwelling conditions and tides certainly make some BBCs flip from one type to the other, with the eddies forming only at a certain periods of the day and year (Hill, 1998). To specify which type of BBC can be expected from the present observations, we must define some qualifying numbers that help to classify the flow. As mentioned earlier, an important number is the internal Rossby deformation Radius R'. When the flow is stratified, baroclinic effects are important, and the BBC scales with the internal Rossby deformation radius R' (Hill, 1998):

$$R' = \frac{\sqrt{g'h}}{f}, \tag{1}$$

where g' is the reduce gravity

$$g' = g\left(\frac{\rho_0 - \rho}{\rho_0}\right), \tag{2}$$

with ρ the density in the BBC, ρ_0 a reference density and f the Coriolis parameter. Another important number is the Ekman number, E, taken here as being the importance of h compared to the total depth H as defined by Munchow (1992):

$$E = \left(\frac{h}{H}\right)^2. \tag{3}$$

The Ekman number (E) indicates the importance of friction. In a low frictional environment (E << 1), the coastal current corresponds approximately to a two-layer flow, and the BBC takes the form of a buoyant wedge that rests against the coast.

The Kelvin number K,

$$K = \frac{W}{R'}, \tag{4}$$

indicates how the BBC scales with the internal Rossby radius of deformation. For small W, the Earth's rotation has little effect on the fresh water plume, whereas when W is O(1) or bigger, the Coriolis effect is important and BBC is well bounded by the coast. In many cases similar to the present study, away from the source region W is two to three times R' (e.g. Münchow and Garvine, 1993b). Another important number is the Rossby number ε, which indicates the importance of the non-linear inertial terms relative to the Coriolis force:

$$\varepsilon = \frac{U}{fW}, \tag{5}$$

where U is the velocity scale of the BBC. If ε is much smaller one, and the Ekman number E is also much smaller than 1, then the flow can be considered to be nearly geostrophic. The two-layer flow resulting from inviscid Rossby adjustment is geostrophic (Hill, 1998), so that the velocity difference between the two layers is given by the Margules relation (Hill, 1998):

$$U_1 - U_2 = \frac{g'}{f}\frac{\partial h}{\partial y}, \tag{6}$$

where U_1 and U_2 are the upper and lower level velocities, and y is the across-shore direction. Considering a lower layer at rest (U_2=0), the BBC velocity scale U is given by the relation:

$$U = \frac{g'}{f}\frac{h}{W}, \tag{7}$$

The last parameter considered is the Froude (Fr) number, which is defined by:

$$Fr = \frac{U}{\sqrt{\dfrac{g'h_2h}{H}}}, \tag{8}$$

where $h_2 = H - h$ (Bracalari et al., 1989). The Froude number determines whether or not the flow is supercritical (Fr > 1) or subcritical (Fr < 1). Observations (Bracalari et al., 1989) have shown that when the flow is supercritical, the flow is robust (i.e. the current should not dissipate rapidly), with K around 2 to 3. When Fr < 1 the flow is irregular and may dissipate rapidly.

For the hydrographic sections described earlier, the flow parameters noted above were calculated (see Table 2 below). The BBC is considered to be represented by a two-layer flow, having a wedge-like shape as described in other work (e.g. LeBlond, 1980, Hill, 1998). We assume that the BBC moves over a lower layer at rest, since no data are available to confirm the velocity of the lower layer. For each section, the limit for the upper fresh layer was defined by the depth of the pycnocline (where the isopycnals are closest together — i.e., where the salinity gradient is strongest). The layer depth (h) was defined as the depth near the coast of the isopycnal representing the top of the pycnocline. The total depth (H) was defined as the near-coast total depth and the width of the BBC (W) was defined as the distance from the coast to where the isopycnal outcrops at the surface. As we consider a two-layer flow model, the salinity was depth-averaged from 0 to h and from h to H. The average salinity and density in each layer was then used to calculate the parameters mentioned above.

The fresh water transport associated with the BBC (T) can then be estimated from:

$$T = UWh\frac{(S_o - S)}{S_o}, \qquad (9)$$

where S_o and S are respectively the lower layer salinity (which is most of the time around 28 psu) and the upper layer salinity. While this calculation is admittedly crude, it provides an estimated fresh water transport ranging from 30 to 585m^3s^{-1}.

Several caveats concerning this calculation need to be addressed. The reference salinity S_o was taken as the lower layer salinity, which is already relatively low (around 28 psu). The estimations of h, W, and H were done by somewhat arbitrarily choosing the location where the isopycnals were closest together. The estimation of W also required knowledge of the position where a particular isopycnal outcropped at the surface, which is dependent on the interpolation of data between stations. Nevertheless, with these caveats noted, the estimations still give an idea of the order of magnitude of freshwater transport associated with the surface freshet. Direct measurements of current velocity at different depths would be necessary to quantitatively validate these estimated values.

In all the case R' ranges from 3km to 6km, allowing a BBC to form on both sides of the channels. In most of the cases E and ε are much smaller than one, confirming that the flow in most cases is of the two-layer type. Thus the geostrophic velocity can be estimated and it ranges from 0.07cm/s to 0.39cm/s with K~3 for most of the cases, as expected.

The geostrophic velocity estimates for the BBCs suggest the existence of slow subcritical ($Fr < 1$) mean flows. To estimate the velocity we made the assumption that the lower layer is at rest. This is not certain since in some cases the lower layer isopycnals are sloping in the opposite direction from the ones at the surface (as in the case for the CC line on the 15th of August 2000). If the Froude number were to be O(1) or bigger, then the flow would be expected to be supercritical with a quite robust BBC. The fact that the data from summer 2000 show that the wind forcing is quite important in this region, suggests that the mean geostrophic BBC flow is easily altered as $Fr \ll 1$ in all cases.

The estimations in Table 2 suggest that the flow is of two-layer type and that even though $\varepsilon < 1$ usually, it still is O(0.1) in most cases. This means that the turbulent terms could be important. The Ekman number is small in most of the cases but the region is relatively shallow. Therefore the above approximations are relatively crude and one would need observations such as current meter data, ADCP data or drifter data to confirm their validity.

Section	h (m)	W (km)	U (cm/s)	R' (km)	ε	E	Fr	K	T (m^3/s)
CC line, 10/9/99 Fig.14	17	6	0.39	4.2	0.48	0.080	0.008	1.4	80
CC line, 15/8/00; Fig.18	7	12	0.24	4.60	0.14	0.014	0.004	2.6	150
CC line, 18/8/00; Fig.20	8	17	0.16	4.50	0.07	0.018	0.003	3.8	103
ML line, south side; Fig.15	8	9	0.21	3.80	0.17	0.01	0.004	2.4	53
ML line, north side; Fig.15	7	14	0.13	3.60	0.07	0.007	0.003	3.8	41
PW line, 1999; Fig.3	10	10	0.38	5.24	0.27	0.001	0.005	1.9	197
BK line, 1999; Fig.2	30	30	0.21	6.90	0.05	0.011	0.002	4.40	586
CG line, 1999; Fig.9	14	45	0.07	5.00	0.01	0.008	0.001	5.28	150
VS line, 1995; Fig.17	14	20	0.10	3.80	0.03	0.096	0.002	5.28	60
DUN line, 1995; Fig.7	19	7	0.19	3.11	0.29	0.90	0.02	2.24	36
DSW line, 1995; Fig.10	10	10	0.21	3.9	0.15	0.028	0.004	2.55	71
DSW line, 1999; Fig.11	5	15	0.09	3.2	0.04	0.002	0.002	4.6	30
DSW line, 2000; Fig.12	5	15	0.09	3.2	0.04	0.03	0.002	4.6	44

Table 2: Dynamical parameters with estimated velocity for the sections showing fresher waters on one side of a channel.

Chapter 6 — Conclusions

Due to the existence of numerous freshwater sources, the presence of numerous sills, and a complex coastline, the CAA freshwater throughflow is a complex phenomenon. The freshwater input from the CAA islands sets up a cross-channel pressure gradient which, together with the Coriolis force, leads to the formation of a BBC around each island (e.g. Victoria Island [CC line], Melbourne Island [ML line]), creating counter currents on each side of the channels. The summer melting of sea-ice is another large and important source of freshwater in the CAA, that is also able to set up its own BBCs as seen in the East Siberian Sea, making the problem even more complex.

The presence of the freshwater in the CAA is known by local people. Driftwood originating from the Mackenzie River drainage basin is also regularly found by them on the south side of the channels, suggesting that the CAA supports a predominant eastward transport. The hydrographic data collected in the summer of 2000, as well as climatological wind data, showed that buoyancy forcing and wind forcing can act together to enhance this eastward transport from Coronation Gulf to Queen Maud Gulf. The salinity distribution before a strong wind event showed a westward BBC on the north side, and the lateral mixing due to the wind forcing redistributed the salinity to set up an eastward BBC on the other coast. The predominant occurrence of driftwood on the south shores, and the climatological wind data showing dominant northwesterly winds, underpin the idea of a mean eastward transport in the region. This follows since there is no persistent south to north cross channel mixing to bring the driftwood to the northern shores. The two places where driftwood was observed on the north side can be explained by wind forcing and the meandering or recirculation of the coastal currents, which are also due to the wind forcing and to the presence of shallow sills and the complex CAA coastline.

In this study only CTD data have been used and one would need data from current meters, onboard ADCP or even better satellite-tracked drifters to better understand and estimate the freshwater transport through the CAA. Also, the emphasis of this work has been on summer conditions when the hydrographic data were collected. During winter, the presence of land-fast ice adds friction at the top of the water column, but prevents the wind forcing from either aiding or counteracting the flow. Also, the buoyancy forcing from river-input is greatly reduced. Only the Mackenzie River gives a relatively large runoff in this region at this time of year, and since hydrographic data are nonexistent for the south west CAA during winter, it is hard to judge what happens to the eastward transport during winter which is much longer than summer in these high latitude regions.

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