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Coastal Fresh Water Discharge in the Northeast Pacific

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Very high annual rates of precipitation in the coastal mountains that border the northeast Pacific Ocean produce large fresh water discharges (23000 m^3 s^{-1}). This discharge has been ignored previously since it does not enter the ocean in the form of large rivers, but, instead, the water enters by way of numerous small rivers and streams. Thus, it acts as a line source instead of a point source. This coastal discharge contributes at least 40% of the fresh water that enters the northeast Pacific from the atmosphere. The discharge is comparable to the mean annual discharge of the Mississippi River system. The fresh water creates a cross-shelf density gradient that drives an alongshore baroclinic jet. The width of this jet is less than 25 km with speeds in excess of 100 cm s^{-1}. It extends along the coast from southeast Alaska to at least Kodiak Island. Apparently, the flow is maintained as a narrow current adjacent to the coast by wind stress that causes downwelling conditions here throughout most of the year.

INTRODUCTION

The importance of fresh water to the ocean circulation of the northeast Pacific has been recognized since *Tully and Barber* [1960] treated it as an estuary. More recently, coastal fresh water discharge has been identified as being a primary driving mechanism of local coastal circulation in the northwest Gulf of Alaska [*Schumacher and Reed*, 1980] and throughout the northern Gulf of Alaska [*Royer*, 1981]. Baroclinic flow controlled by salinity distributions is possible here because of the relatively low water temperatures and high rates of fresh water discharge.

Previous discussions of the availability of the fresh water in the northeast Pacific are based either on river discharges or precipitation rates. *Roden* [1967] addresses the discharge of major river systems into the northeast Pacific and Bering Sea. The major river mean annual discharges are the Fraser River (2690 m^3 s^{-1}) in British Columbia and the Copper River (1050 m^3 s^{-1}) in Alaska. Roden also included six other minor rivers, which have a combined average discharge of less than 3000 m^3 s^{-1}. Another method of assessing the availability of fresh water in the northeast Pacific is through the use of oceanic precipitation estimates. However, as can be seen by two recent estimates of maximum precipitation rates for the northeast Pacific, the calculated fresh water input can differ considerably. *Reed and Elliott* [1979] report a precipitation rate of 100 cm yr^{-1}, while *Dorman and Bourke* [1979] show a rate of 180 cm yr^{-1} for the same oceanic area. This discrepancy is primarily caused by Dorman and Bourke correcting through the use of coastal station data and because coastal rainfall rates are high in the northeast Pacific their oceanic rates are enhanced.

Indirect evidence for high rates of runoff and advection in the northeast Pacific is given by *Reed and Elliott* [1973]. Using the assumption that upwelling in the central Gulf of Alaska causes on upward salt flux that must be balanced by fresh water to maintain steady state conditions, they obtain a fresh water influx of 125 cm yr^{-1} based on *Tully and Barber*'s [1960] estimate of a vertical velocity in the Gulf of Alaska of 20 m yr^{-1}. The excess precipitation over evaporation rate is 50 cm yr^{-1} so that the remaining 75 cm yr^{-1} must be made up by runoff and advection. By using the area of the northeast Pacific from 60°N to 50°N, 150°W to 140°W at 60°N and 150°W to 130°W at 50°N to be about 1.1×10^{12} m^2, then about 26000 m^3 s^{-1} of fresh water through runoff and advection is required to balance this upward salt flux and the total influx of fresh water must be approximately 43000 m^3 s^{-1} (17000 m^3 s^{-1} from precipitation-evaporation).

In this paper, the sources of this runoff and advection for the northeast Pacific are discussed. Seasonal changes of the runoff and their effects on the coastal circulation are considered.

DISCHARGE COMPUTATIONS

The high precipitation rates suggested by *Dorman and Bourke* [1979] are consistent with the rates used by *Royer* [1979, 1981] to obtain coastal fresh water discharge in the northern Gulf of Alaska. The coastal discharges were determined by using a 150×600 km drainage area to represent the coastal region. To calculate discharges, the monthly mean U.S. Weather Service divisional precipitation rates for southcoast Alaska were used. In this work however, two U.S. Weather Service climatic divisions are used: southcoast Alaska (approximately 140°W to 150°W) and southeast Alaska (approximately 130°W to 140°W) (see Figure 1). Depending on the monthly mean air temperature, the precipitation is allowed either to runoff during the month or to be stored as snow. The snow is released later when the air temperature is above freezing. It is released gradually over a period of several months, with a pattern that closely approximates the river discharge of those rivers that drain the coastal mountain ranges. An approach involving indirect computations is required here because direct measurements of the discharges of the myriad of rivers and streams is not possible.

To improve the estimate of coastal fresh water discharge into the northeast Pacific, the simple computations have been modified to approximate actual drainage areas better. This improved method still uses the monthly mean divisional precipitation and air temperatures as its input, but incorporates more realistic drainage areas and allows the interannual ablation or growth of the glacial fields that are found in these coastal mountains. This type of response is necessary since glaciers occupy approximately 20% of this coastal drainage area. Abnormally high summer air temperatures are permitted to cause a higher than normal fresh water discharge with abnormally low temperatures causing low fresh water drainages.

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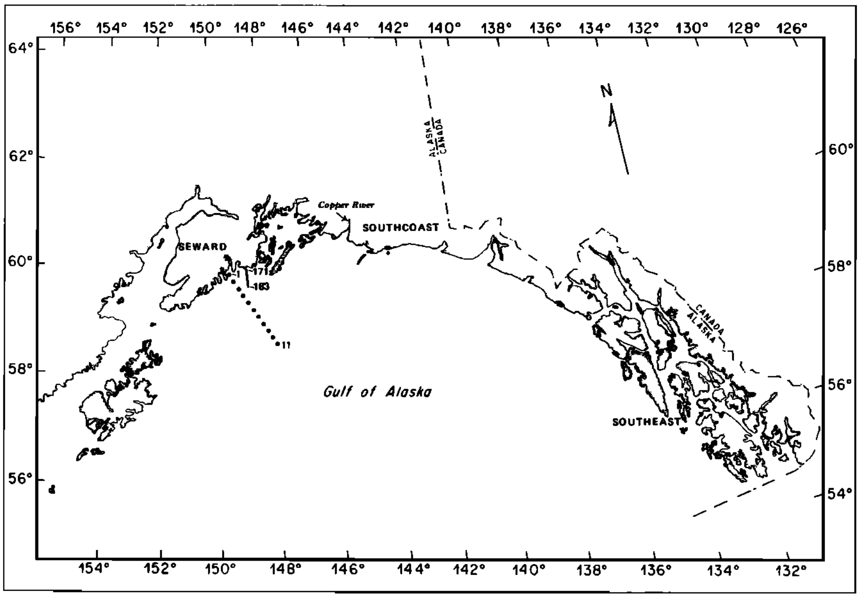

Fig. 1. Coastal region of northeast Pacific Ocean.

Two separate drainage areas, southeast and southcoast Alaska, are used in the computations (Figure 1). Precipitation and air temperatures are available for each of the two divisions. Transit times in the form of phase shifts in the discharge are incorporated in this method. A northwestward coastal flow averaging about 30 cm s^{-1} from southeast to southcoast Alaska is approximated by lagging the southeast discharge by 1 month. (Because monthly means are being used, choices for flow speeds are limited to increments of 30 cm s^{-1}, ~ 800 km per month.). The discharges from streams and rivers that are gauged are not included separately. The contribution from the Copper River, which drains a portion of interior Alaska, is not included because its records do not cover the same time period as the precipitation and air temperature records. As will become more evident later in this paper, its contribution (1000 m^3 s^{-1}) is less than 5% of the total coastal discharge and is insignificant in comparison with other errors in the hydrology model.

The coastal topography and the precipitation distributions here are similar to other high latitude regions, such as the coast of Scandinavia [*Bergeron*, 1949]. Along that coast the moist marine air masses impinging on the coastal mountain ranges are elevated abiabatically, and precipitation takes place. Analogous processes occur at the northeast Pacific Coast where mountains with heights exceeding 4 km are common in the Alaska Coastal Range. The orographic control of precipitation causes higher rates at higher elevations. Thus, the 180 cm yr^{-1} precipitation rate measured at sea level probably translates into a much greater rate at the higher elevations. The location of meterological observing sites in coastal communities, therefore, leads to an underestimate of regional precipitation rates. However, other areas on the leeside of the mountains will have lower rates. It is beyond the scope of this paper to evaluate the magnitude of these errors, since precipitation rates over these sparsely inhabited areas are not well known. Though detailed seasonal variations in rainfall are not well documented, the most complete representation of the spatial distribution of precipitation is given by *Selkregg* [1974]. Annual precipitation rates in excess of 240 inches (610 cm) are present for the glacial areas, with one area in southeast Alaska having a 320 inch (813 cm) contour. Thus, while the use of the divisional precipitation averages (about 240 cm) might be an underestimate it will be used in lieu a suitable substitute. The high rate of coastal precipitation extends to the south along the British Columbia coast. *Kendrew and Kerr* [1955] indicate that the 100-inch (254 cm) precipitation contour is continuous along the British Columbia coast from Alaska to Washington. The effects of the British Columbia discharge will not be included in this study, though they are undoubtedly important.

Concerning the proportion of precipitation, *Sellers* [1965] indicates that nearly all the precipitation in the northwest Pacific runs off. No mention is made of Alaskan coastal precipitation runoff, but similar runoff conditions will be assumed for this region.

FRESH WATER DISCHARGE

The addition of the monthly fresh water discharges for southcoast and southeast (lagged by 1 month) Alaska from 1931 through 1979 (Figure 2) demonstrates a large seasonal signal (Figure 3). This seasonal cycle in the fresh water discharge closely resembles the discharge reported in *Royer* [1979]. The minimum in February–March coincides with oceanographic winter. The submaximum in May represents spring runoff followed by a general increase to the October maximum. This increase is a result of the meltwater discharge and increased seasonal precipitation rate. The sharp

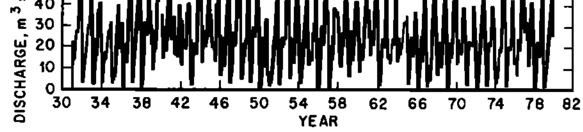

Fig. 2. Monthly fresh water discharge for northeast Pacific with southeast Alaska discharge lagging southcoast Alaska discharge by 1 month.

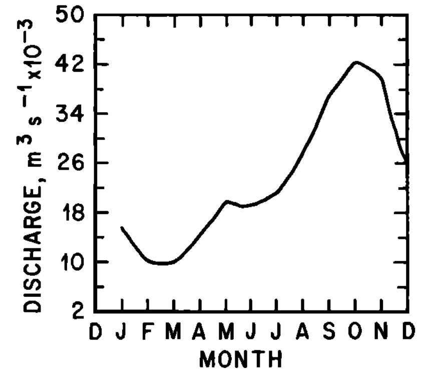

Fig. 3. Mean monthly fresh water discharge determined in same manner as Figure 2, using data from 1931 to 1979.

decline in November–December is a consequence of air temperatures becoming less than 0°C.

The total discharge (Figure 2) varies from nearly zero to greater than 60000 $m^3 s^{-1}$. Its 1931–1979 mean of 23000 $m^3 s^{-1}$ is similar to the runoff advection required by the salt balance (26000 $m^3 s^{-1}$). To illustrate better long-term trends, the monthly air temperatures, precipitation, and discharges are used to determine annual means (Figure 4). As expected, the air temperatures for southcoast Alaska are always less than those for southeast. The curves of mean annual temperature and precipitation are quite similar for southcoast and southeast Alaska, confirming that the same atmospheric system influences both regions. The below freezing annual mean southcoast air temperatures for 1934 (−0.31°C) and 1935 (−0.16°C) are especially interesting since they were more than 2° below any others and were accompanied by a subnormal precipitation rate. The discharges for 1934 and 1935 were 16000 $m^3 s^{-1}$ and 18000 $m^3 s^{-1}$, respectively, which are well below the 1931–1979 average of 23000 $m^3 s^{-1}$. The minimum annual computed discharge occurred in 1950 when the average was slightly less than 16000 $m^3 s^{-1}$. This decreased discharge was the result of subnormal temperatures and precipitation rates in November 1950, yielding a discharge of only 16000 $m^3 s^{-1}$ compared with the mean November rate of 40000 $m^3 s^{-1}$. Throughout 1950, the monthly discharges were slightly below normal also. The maximum discharge occurred in 1940 with 33000 $m^3 s^{-1}$, after which there was a general decline in discharge until about 1971. The precipitation and air temperatures contain similar patterns. Because oceanographers commonly deal with transports of the order of 10^6 $m^3 s^{-1}$, these discharges seem to be insignificant. However, the mean annual discharge of the Mississippi River is about 18000 $m^3 s^{-1}$

Oceanic Response

The availability of hydrographic data for the Seward line (Figure 1) from 1974 through 1979 permits the comparison of the fresh water discharge with the computed along-shore baroclinic flow for this period (Figure 5). The along-shore transport (0/100 dbar) between station pairs 1 and 2 and 1 and 7 (Figure 1) are used for the correlation. Alongshore is defined here as being orthogonal to the section line (see Figure 1). Lags for 0–3 months for the southeast discharge relative to the southcoast discharge are also used to determine the linear 'best fit' response. On the basis of 22 samples, the best correlation ($r = 0.763$) between baroclinic

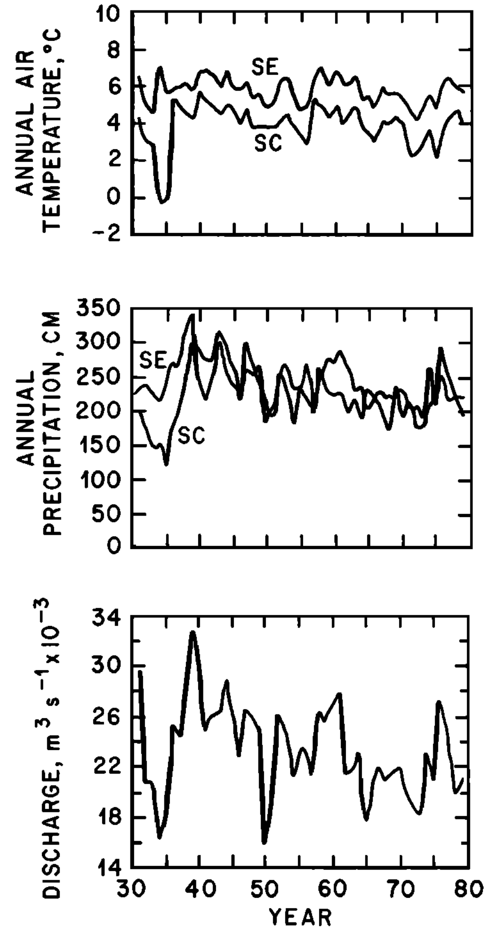

Fig. 4. Annual mean air temperature (top) for southeast Alaska (SE) and southcoast Alaska (SC), precipitation (middle), and fresh water discharge (bottom).

transport and fresh water discharge occurs where the southeast discharge is lagged by 1 month. The confidence interval for this correlation is greater than 99.9%. The reduction in the correlation depending on lags is not sharp since the autocorrelation of discharge decreases slightly 0.847 at two months.

The above correlation between (0/100 dbar) transport and fresh water discharge contains a very large seasonal signal, so that the high correlation could be simply due to both

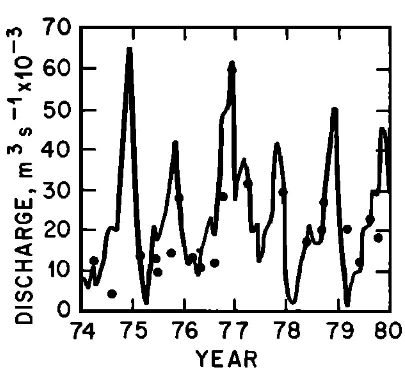

Fig. 5. Monthly fresh water discharge (same as Figure 2) for 1974 to 1979 (line) with baroclinic transport (dots) for 1–7 0/100 dbar superimposed. Range of the baroclinic transport is 0 to 1.5 × 10^6 $m^3 s^{-1}$.

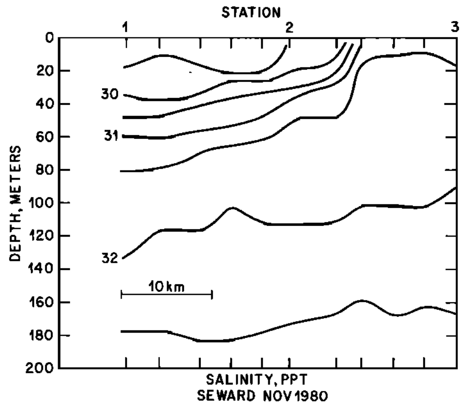

Fig. 6. Salinity cross section for Seward line looking eastward with additional stations between stations 1 and 3 (see Figure 1), November 1980.

responses having this annual signature. A better test of the relationship between fresh water discharge and baroclinic flow is done on the anomalies, that is, the time series remaining after the removal of each of their respective annual signals. The cross correlation of the anomalies of (0/100 dbar) baroclinic transport for stations 1–7 and fresh water discharge with southeast lagged by 1 month is 0.603, which has a confidence interval of greater than 99.5%. This correlation is slightly higher than that determined previously for the simple runoff computations, which was 0.580 [Royer, 1981]. The more realistic methods used in this work probably better estimate the actual discharge.

The representation of the coastal circulation throughout the northern Gulf of Alaska by flow across the Seward line might not be valid in light of some more recent temperature and salinity cross sections taken upstream from this line. In November 1980, the coastal current can be seen as a lens of low salinity water (< 31 ‰) near the surface between stations 1 and 2 on the Seward line (Figure 6). The width of this lens is approximately 25 km. Because the flow is adjacent to the coastline, its direction will be strongly influenced by it. Islands and peninsulas to the west divert the flow southward across the Seward line, making that transect oblique to the current. This is verified by a section at Cape Fairfield (Figure 7) taken a few hours prior to the Seward transect. At Cape Fairfield, the 31 ‰ salinity band is only about 15 km wide. The difference could be a consequence of cross-shelf spreading of the coastal current. However, for another section to the west of the Seward line, the coastal current is approximately 18 km wide. The conclusion is that the Seward line does not intersect the coastal current normal to the flow but rather obliquely. Thus, while the transports are valid, the width appears greater than it actually is, and the baroclinic current speeds here are underestimated. These current widths are similar to the internal Rossby radius of deformation, which ranges from 4 to 10 km for the Cape Fairfield line.

The dynamics of fresh water coastal currents have been investigated in the Norwegian Sea by *Heaps* [1980]. He uses a two-layer analytic model on a deep shelf similar to that of the Gulf of Alaska. The major difference between the two situations is that his fresh water discharge per unit length of coastline is about half of that for the Gulf of Alaska. As expected, his transports are about half of those for the Seward line. One feature that he predicts but is not observed in our situation is a fairly rapid cross-shelf dissipation of the coastal flow. He predicts a cross-shelf velocity at the surface that is approximately 20% of the along-shore component. Thus, the coastal current should spread to encompass the entire shelf after traveling several hundred kilometers downstream. As can be observed in Figure 6, this does not occur here. The narrow current could be a result of flow being constricted as it exits Prince William Sound. However, other observations in the northern Gulf of Alaska [*Royer et al.*, 1979; *Schumacher and Reed*, 1980; R. Muench, personal communication, 1981) verify a narrow flow elsewhere.

A mechanism that could concentrate this flow at the coast is wind stress. The wind stress, here expressed as upwelling

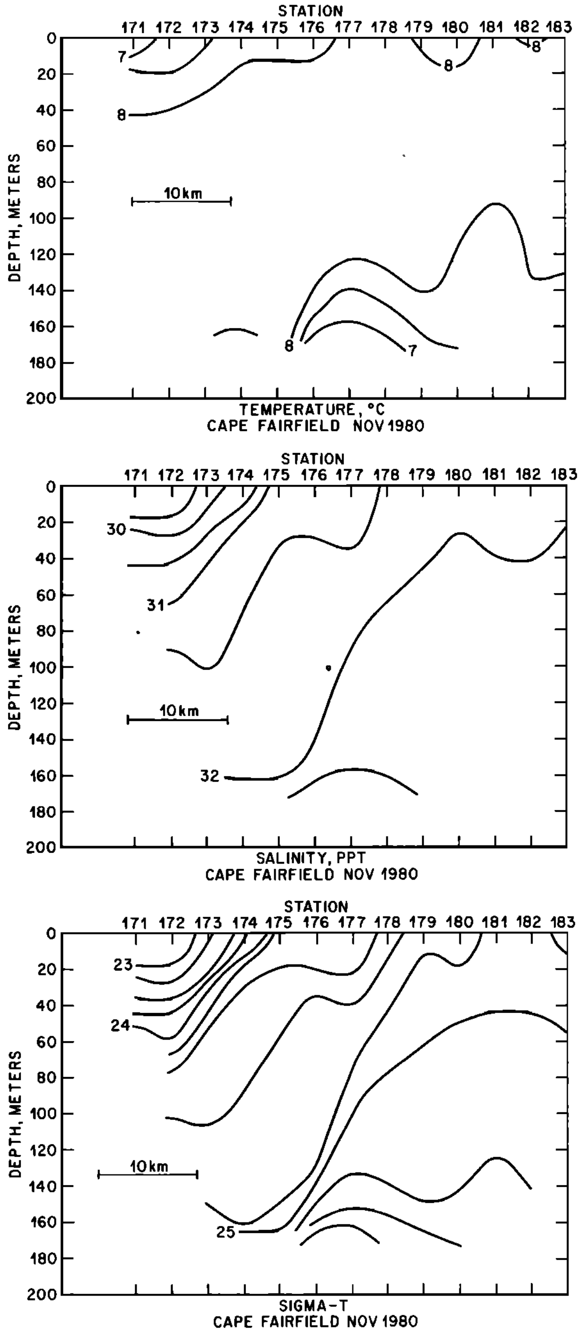

Fig. 7. Cross sections of temperature (top), salinity (middle), and density (bottom) at Cape Fairfield, November 1980 (see Figure 1 for locations).

indices, changes by an order of magnitude from summer to winter in the norhtern Gulf of Alaska [*Royer*, 1975]. Throughout the year, easterly winds are common so that the northern Gulf of Alaska usually has downwelling conditions [*Livingstone and Royer*, 1980]. During the November 1980 cruise the mean downwelling index was 69 $m^3 s^{-1}$ (100 m of coastline)$^{-1}$ (A. Bakun, personal communication, 1981), which is typical for that time of year [*Royer,* 1979]. The offshore transport in the upper layers as predicted by the *Heaps* [1980] model can be compensated by the onshore Ekman transport. Each of these processes has a lower layer of comparable thickness that moves in the opposite direction of the upper layer. Downwelling is phased so that it lags the maximum fresh water discharge by approximately 3 months; however, the maximum baroclinic transport remains in phase with the fresh water discharge [*Royer*, 1979]. The narrow, intense coastal current can be considered to be created by the fresh water discharge and then modified by the winds.

Eddies are predicted for salinity-induced flows under similar conditions [*Elliott and Reid*, 1976], but in this case the available hydrographic data are inadequate to address this problem. Eddies might form if there is cross-isobathic flow. However, repeated hydrographic sections in February–March 1976 had only about 10% variation in the 0/100 dbar transport between stations 1 and 7 [*Royer*, 1981]. A more complete temporal signature of this coastal flow is not available, such as its response to nonsteady forcing.

CONCLUSIONS

Through the use of runoff computations, the coastal fresh water discharge for southeast and southcoast Alaska is estimated as about 23000 $m^3 s^{-1}$. This computation does not include the Copper River input or discharge from British Columbia, though the latter is probably significant. This runoff rate is close to the 26000 $m^3 s^{-1}$ required by a salt balance over the entire Gulf of Alaska. The water enters as a line source at the coast. This is in contrast with other entry mechanisms such as a point source, which is typical of large river input or as a uniform surface distribution, which is characteristic of precipitation over the ocean. This line source of fresh water creates a cross-shelf horizontal density gradient driving an alongshore baroclinic flow which can exceed 1.3×10^6 $m^3 s^{-1}$. The winds, which are typically easterlies, converge this upper Ekman layer water at the coast and maintain the flow as an intense narrow current, generally less than 20 km wide.

The coastal mountain ranges bordering the Gulf of Alaska act as a barrier to the storms that move easterly across the North Pacific. Adiabatic elevation of these moist air masses cause very high rates of precipitation (>8 m yr^{-1}) in the form of rain and snow. The precipitation can be retained for months or even years in the glacial fields, which occupy approximately 20% of the region. Better estimates of the growth or ablation of these ice sheets would improve this method of determination of fresh water discharge. For example, what effect would be observed in the coastal current under glacial advance or retreat? With any increase in the fresh water discharge, the baroclinic transport would probably increase, what would be the current response? Will it become wider, deeper, or simply faster? These types of questions should be answered through the application of the two-layer baroclinic model.

While the volume of fresh water discharge is significant in comparison with large river discharges, it is equally important to stress the significance of this source of fresh water to the northeast Pacific circulation. The area of the northeast Pacific is about 1.1×10^{12} m^2, and the annual precipitation rate is between 0.9 m yr^{-1} [*Reed and Elliott*, 1979] and 1.1 m yr^{-1} [*Dorman and Bourke*, 1979]. Thus, the coastal fresh water discharge in the northeast Pacific contributes between 38 and 43% of the total amount of fresh water that enters from the atmosphere. Oceanic evaporation will increase this percentage contribution to more than 50%. More importantly, this large coastal discharge can affect the dynamics since it creates a sharp horizontal density gradient that might drive along-shore flows at distances offshore.

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