JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 94, NO. C10, PAGES 14,485–14,498, OCTOBER 15, 1989

The Role of Sea Ice and Other Fresh Water in the Arctic Circulation

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Salinity stratification is critical to the vertical circulation of the high-latitude ocean. We here examine the control of the vertical circulation in the northern seas, and the potential for altering it, by considering the budgets and storage of fresh water in the Arctic Ocean and in the convective regions to the south. We find that the present-day Greenland and Iceland seas, and probably also the Labrador Sea, are rather delicately poised with respect to their ability to sustain convection. Small variations in the fresh water supplied to the convective gyres from the Arctic Ocean via the East Greenland Current can alter or stop the convection in what may be a modern analog to the halocline catastrophes proposed for the distant past. The North Atlantic salinity anomaly of the 1960s and 1970s is a recent example; it must have had its origin in an increased fresh water discharge from the Arctic Ocean. Similarly, the freshening and cooling of the deep North Atlantic in recent years is a likely manifestation of the increased transfer of fresh water from the Arctic Ocean into the convective gyres. Finally, we note that because of the temperature dependence of compressibility, a slight salinity stratification in the convective gyres is required to efficiently ventilate the deep ocean.

INTRODUCTION

The role of fresh water in ocean circulation and climate change is presently of increasing interest. Two particular points of inquiry have been the role of the precipitation-evaporation imbalance in the North Atlantic in driving the large-scale thermohaline circulation [*Weyl*, 1968; *Broecker et al.*, 1985], and the impact on that circulation of hypothesized rapid glacial melting in the distant past [*Rooth*, 1982], the so-called halocline catastrophe. With respect to the latter, *Bryan* [1986] has demonstrated with a general circulation model that the response of the global circulation to such an event can occur in a century or less. We here introduce another component in the hydrologic cycle, in this case a purely oceanic one, namely, freezing. Later in this paper we shall include its effects in a modern analog to the possible halocline catastrophes of the past.

Most of the Arctic receives a net surplus of fresh water from the hydrologic cycle, including a large amount of runoff discharged into the Arctic Ocean. Furthermore, much of this latter ocean in particular is permanently and strongly stratified, a prerequisite to significant ice formation in deep oceans. The importance of runoff in particular has encouraged consideration of an estuarinelike circulation for the Arctic Ocean, and several models have been proposed, e.g., by *Stigebrandt* [1981] and *Björk* [1989].

Another mode of arctic thermohaline circulation is associated with brine rejection during freezing (see *Schumacher et al.* [1983] for a regional example and *Aagaard et al.* [1985] for a more general discussion]. Thus far, investigation of the freezing process has centered on the production and dispersion of the brines and their mixtures, together with their influence on dynamically passive tracers, and a great variety of work now points to convection forced by freezing over the adjacent shelves as pivotal in ordering the hydrographic structure of the Arctic Ocean [e.g., *Aagaard et al.*, 1981; *Melling and Lewis*, 1982; *Moore et al.*, 1983; *Jones and Anderson*, 1986; *Wallace et al.*, 1987]. However, we will here focus on a complementary issue which we believe is of major importance but has been largely ignored (an exception is the seminal paper by *Rooth* [1982]). This issue is the role in the general circulation of the fresh water which has been distilled during freezing, which process we consider to be a high-latitude analog to evaporation.

The importance of fresh water to high-latitude circulation follows from the properties of the equation of state for seawater at low temperatures. In particular, because the thermal expansion coefficient for seawater at low temperatures is so small (so that temperature stratification has very little effect on the density structure), the role of fresh water is critical to considerations of vertical motion at high latitude. For example, at the freezing temperature and 34.5 psu (practical salinity units) [*Lewis and Perkin*, 1978] the ratio of the haline and thermal coefficients is about 30:1, and even at +2°C and 34.5 psu it is 10:1. The introduction of small amounts of fresh water can therefore prevent convective overturn even in the case of substantial surface cooling. It is therefore somewhat surprising that relatively little attention has been paid to the details of the fresh water cycle in the Greenland, Iceland, and Norwegian seas (hereinafter collectively referred to as the GIN Sea), despite the fact that both the Greenland and Iceland seas contain convective regimes of major importance to the global thermohaline circulation [*Aagaard et al.*, 1985].

In this paper we examine the fluxes of fresh water in the Arctic Ocean and its connections to the North Atlantic through the GIN Sea (Figure 1), with a particular eye toward the control of the stratification through these fluxes. We are especially interested in the augmentation of the effects of the hydrologic cycle through the freezing, transport, and melting of sea ice. Some of our conclusions will have applicability to

Paper number 89JC01375.
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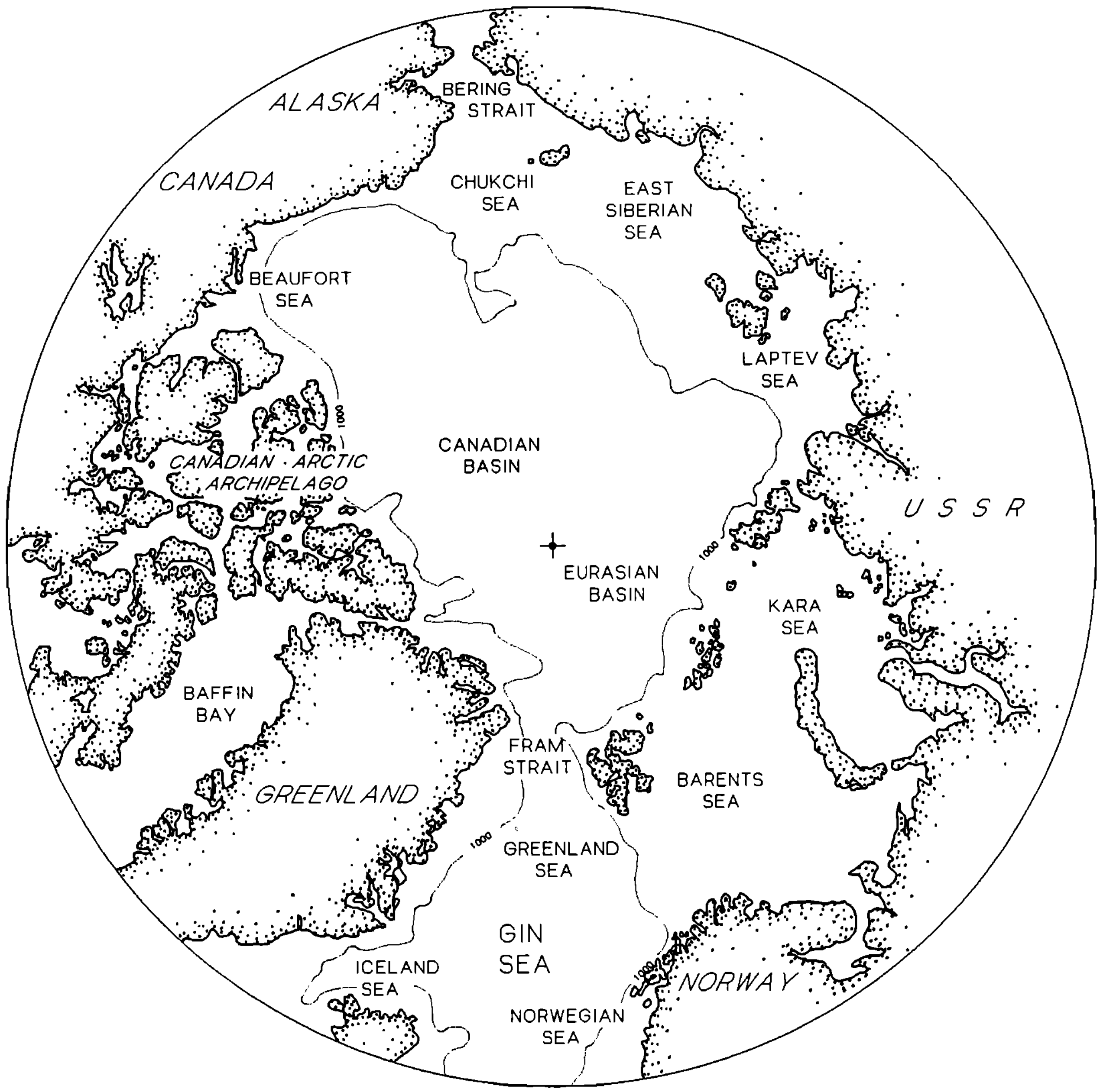

Fig. 1. The Arctic Ocean and the GIN Sea. The 1000-m isobath delineates the deep basins.

other parts of the Arctic, particularly the Labrador Sea, and probably also to portions of the Antarctic, especially the Weddell Sea.

Freezing as a Large-Scale Distillation Process

The amount of salt expelled from sea water as it freezes depends strongly on the ice growth rate, but typically two thirds or more of the salt is rejected initially. Most of the salt remaining in the ice is subsequently released to the ocean through a combination of processes which by summer results in ice with only 5–10% of its original salt content (see *Maykut* [1985] for a review). Since several meters of ice are typically formed annually in the polar regions (ranging from ~0.5 m under permanent equilibrium-thickness Arctic Ocean ice [*Maykut*, 1985] to ~10 m in persistent polynyas [*Martin and Cavalieri*, 1989], the distillation rates from freezing are fully comparable to those from evaporation in such highly evaporative basins as the Red Sea (~2 m yr^{-1} [*Bunker et al.*, 1982]). If the ice is subsequently exported from its production area, or alternatively, if the brines produced are exported so that a net local distillation occurs, then the freezing and melting cycle becomes the oceanic equivalent of the hydrologic cycle in the atmosphere, i.e., evaporation and precipitation. (Since high-latitude evaporation rates are small, freezing is in fact the only effective distillation process operating in the polar regions.) In the Arctic, the major ice outflow from the polar basin occurs east of Greenland, where the exodus represents a fresh water transport of about 2800 km^3 yr^{-1} (compare below). This is a discharge close to twice that of North America's four largest rivers combined (the Mississippi, St. Lawrence,

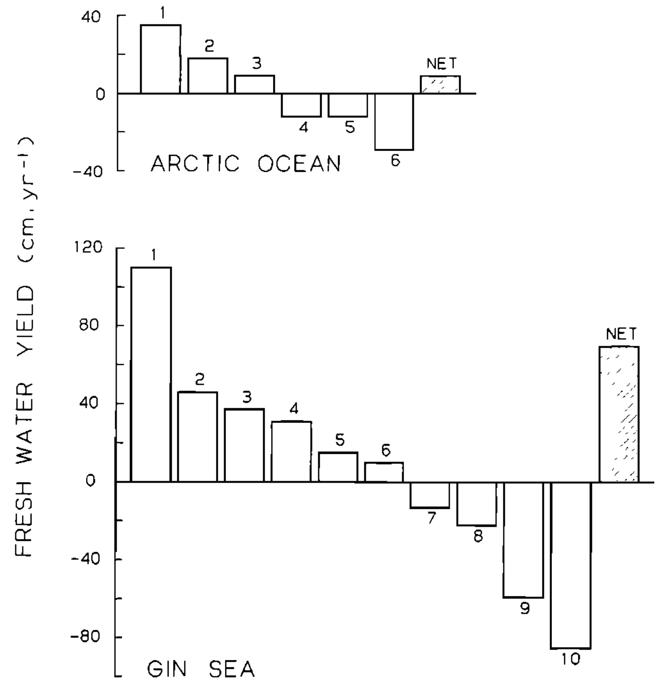

Fig. 2. Fresh water sources and sinks for the Arctic Ocean and the GIN Sea. Calculations for the Arctic Ocean are relative to a base salinity of 34.80, and those for the GIN Sea are relative to 34.93. The individual terms are independently estimated with no attempt to balance the budget. Sources and sinks corresponding to a yield less than 9 cm yr^{-1} are not shown. Arctic Ocean plots are as follows: 1, runoff; 2, import through Bering Strait; 3, precipitation less evaporation; 4, water export through Fram Strait; 5, export through the Canadian archipelago; 6, ice export through Fram Strait. GIN Sea plots are as follows: 1, ice import through Fram Strait; 2, water import through Fram Strait; 3, import from Skagerrak with the Norwegian Coastal Current; 4, precipitation less evaporation; 5, runoff; 6, saline water export to Barents Sea; 7, export to Barents Sea with the Norwegian Coastal Current; 8, ice export through Denmark Strait; 9, water export through Denmark Strait; 10, saline water import from the North Atlantic.

Columbia, and Mackenzie), and in the world it is second only to that of the Amazon [*Holland*, 1978, p. 86]. Furthermore, as we shall demonstrate, this fresh water is transported with very little dispersion at least as far as Denmark Strait, over 1500 km from its exit point in Fram Strait and still farther from the principal ice formation areas within the Arctic Ocean. In fact, a reasonable interpretation of the recent work by *Dickson et al.* [1988] is that the fresh water initially carried southward by the East Greenland Current can subsequently be followed around the subpolar gyre of the North Atlantic. The point is that the formation of sea ice in the Arctic Ocean and its transport into the North Atlantic represent a fresh water flux comparable to that of continental runoff and a basin-scale translation of the fresh water. In the Antarctic, a comparable phenomenon may be represented by freezing in the southern Weddell Sea and ice transport northward along the Antarctic Peninsula.

THE FRESH WATER BUDGET

The Arctic Ocean

Figure 2 and Table 1 show the individual contributions to the fresh water budget of the Arctic Ocean. All fresh water fractions are relative to a salinity of 34.80, which we estimate as the mean salinity for the Arctic Ocean based on the compilations of *Codispoti and Richards* [1968], *Hanzlick and Aagaard* [1980], *Gorshkov* [1983], *Pfirman* [1985], *Treshnikov* [1985], and *Macdonald et al.* [1987]. Despite our relatively poor knowledge of the Arctic Ocean hydrography, the uncertainty in this estimate is probably only about 0.04, and corresponding changes in the reference salinity will not materially influence our conclusions. For example, with a change of 0.04 in the reference salinity, the largest change in an individual term in the fresh water budget would be 110 km^3 yr^{-1} and only 20 km^3 yr^{-1} in the net budget. These are both less than the uncertainty in the corresponding transport estimates.

The various sources and sinks have been calculated independently with no attempt to balance the budget; the latter issue is addressed below. We report the results both as annual flux (in cubic kilometers per year) and as yield (flux per unit area, expressed in centimeters per year), with the area of the Arctic Ocean taken as 9.55×10^6 km^2. The individual terms for the Arctic Ocean were calculated as follows:

Runoff. Inflow tabulations from the major rivers entering the Arctic Ocean have been compiled by *UNESCO* [1978], *Milliman and Meade* [1983], and *Treshnikov* [1985]. We accept the latter values, which total 3300 km^3 yr^{-1} (35 cm yr^{-1}), as these are based on the most recent and extensive compilations. Individual contributions, shown schematically in Figure 3, include the Yenisei (603 km^3 yr^{-1}), Ob (530 km^3

TABLE 1. Fresh Water Budget for the Arctic Ocean

Source or Sink	Transport, km^3 yr^{-1}	Yield, cm yr^{-1}
Ice export through Fram Strait	−2790	−29
Water export through Fram Strait	−820	−9
Runoff	3300	35
Precipitation less evaporation	900	9
Water import through Bering Strait	1670	18
Water export through Canadian archipelago	−920	−10
Import with Norwegian Coastal Current	250	3
Saline water import through Barents Sea	−540	−6
Saline water import with West Spitsbergen Current	−160	−2
Net	890	9

Fresh water fractions are relative to the salinity 34.80. Yield calculated for an area of 9.55×10^6 km^2. Values are positive for sources and negative for sinks.

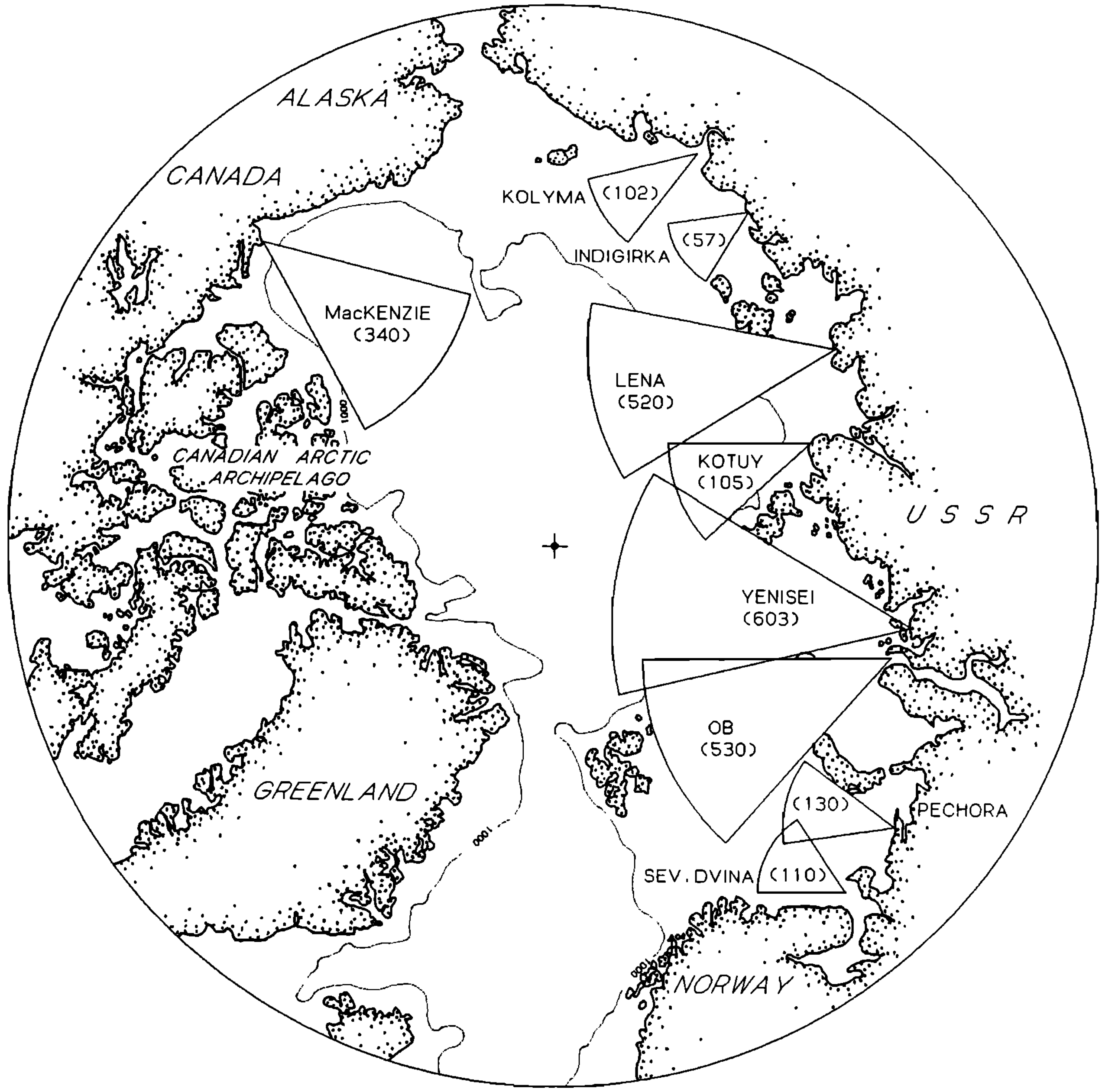

Fig. 3. Mean annual runoff to the Arctic Ocean in cubic kilometers per year. Only the nine largest rivers are shown.

yr^{-1}), Lena (520 km^3 yr^{-1}), Pechora (130 km^3 yr^{-1}), North (Severnaya) Dvina (110 km^3 yr^{-1}), Kotuy (105 km^3 yr^{-1}), Kolyma (102 km^3 yr^{-1}), Pyasina (86 km^3 yr^{-1}), and Indigirka (57 km^3 yr^{-1}) rivers in the U.S.S.R., the Mackenzie River (340 km^3 yr^{-1}) in Canada, and numerous other smaller rivers (totaling 720 km^3 yr^{-1}) surrounding the basin.

There are significant annual and interannual variations in these flows [*Cattle*, 1985]. For example, the Yenisei and the Lena show a fortyfold increase from very low winter values to the peak flows of June and July. The corresponding change for the Mackenzie is much less, but still large, about fivefold. Interannual flow variability in individual rivers is typically 5–20% of the annual mean. In this paper we use long-term means and thus ignore such variability, both in this and other budget terms.

Precipitation less evaporation. There is considerable uncertainty regarding this flux. Estimates range from 400 km^3 yr^{-1} [*Baumgartner and Reichel*, 1975] to 1400 km^3 yr^{-1} [*Burova*, 1981]; we accept an intermediate value of 900 km^3 yr^{-1} (9 cm yr^{-1}), close to that of *Vowinckel* and *Orvig* [1970]. While this term is small compared to others in the budget, we note that its effects might vary markedly under slightly different climatic conditions. For example, winter snow accumulation could either primarily enter the ocean as meltwater, or return to the atmosphere through sublimation, depending on spring atmospheric conditions.

Liquid fresh water import through Bering Strait. The most recent transport estimates for Bering Strait [*Coachman and Aagaard*, 1988] show an annual cycle of amplitude 0.3 Sv superimposed on a long-term mean flow of 0.8 Sv, with a maximum in summer and a minimum in winter. The salinity of the inflow is generally 31–33, and the long-term mean is

TABLE 2. Fresh Water Budget for the GIN Sea

Source or Sink	Transport, $km^3\ yr^{-1}$	Yield, $cm\ yr^{-1}$
Ice import through Fram Strait	2790	110
Water import through Fram Strait	1160	46
Ice export through Denmark Strait	−560	−22
Water export through Denmark Strait	−1520	−60
Precipitation less evaporation	790	31
Runoff	420	16
Import from Skagerrak, Norwegian Coastal Current	950	37
Export to Barents Sea, Norwegian Coastal Current	−330	−13
Saline water import from North Atlantic	−2160	−85
Saline water export to Barents Sea	260	10
Net	1800	70

Fresh water fractions are relative to the salinity 34.93. Yield calculated for an area of 2.55×10^6 km^2. Values are positive for sources and negative for sinks.

probably near 32.5 [*Aagaard and Greisman*, 1975; *Coachman et al.*, 1975]. We therefore calculate the fresh water import from the Pacific as 1670 $km^3\ yr^{-1}$ (18 $cm\ yr^{-1}$).

Liquid water export through the Canadian arctic archipelago. The Canadian archipelago is a large and complex system of channels through which upper waters from the Arctic Ocean enter Baffin Bay. *Fissel et al.* [1988] have recently synthesized the results of a major current monitoring program for the archipelago. They found a net transport of 1.7 Sv, about 20% less than an earlier estimate by *Muench* [1971] which has been widely cited. We combine the recent Canadian transport results with the corresponding mean salinity estimate of 34.2 by *Aagaard and Greisman* [1975] to get a fresh water outflow of 920 $km^3\ yr^{-1}$ (−10 $cm\ yr^{-1}$).

Import of fresh water with the Norwegian Coastal Current, import of saline water through the Barents Sea and with the West Spitsbergen Current, and export of ice and liquid water through Fram Strait. These are discussed below under the GIN Sea budget and represent gains of 250, −540, −160, −2790 and −820 $km^3\ yr^{-1}$, respectively (3, −6, −2, −29, and −9 $cm\ yr^{-1}$).

Omissions. Several fresh water sources and sinks have been omitted from our budget calculations. First, we have neglected the import of ice through Bering Strait. C. Pease (personal communication, 1989) has estimated this to be about 30 $km^3\ yr^{-1}$, and if this ice has a bulk salinity of 7, the associated annual fresh water inflow is only 24 km^3 (0.3 $cm\ yr^{-1}$), which is negligible for our purposes.

Second, we have neglected the export of ice through the Canadian archipelago. We estimate the total cross section of the major passages through the archipelago as 34 km^2, which for a transport of 1.7 Sv [*Fissel et al.*, 1988] yields a characteristic outflow speed of 5 $cm\ s^{-1}$. If we take a mean ice thickness of 2 m and an outflow duration of 3 months (the ice being landlocked the other 9 months), the total ice export is 155 km^3. Much of this ice is frozen locally, rather than representing outflow from the Arctic Ocean, so that the net ice export from the polar basin through the Canadian arctic archipelago is smaller than any of the terms retained in Table 1.

Third, we have neglected the export of fresh water south of Spitsbergen. This flux is discussed in some detail under the omissions in the GIN Sea budget and corresponds to a yield for the Arctic Ocean of only about −2 $cm\ yr^{-1}$, even if its recirculation (see the GIN Sea discussion) is ignored. The latter effect reduces the yield still further.

The net surplus. Our fresh water budget for the Arctic Ocean shows a surplus of 890 $km^3\ yr^{-1}$ (9 $cm\ yr^{-1}$), i.e., about the same as the estimated contribution of precipitation less evaporation. Considering the uncertainties in the various terms in the budget, this imbalance is probably indistinguishable from zero. For example, the excess is only about 25% of the estimated fresh water export through Fram Strait, which by itself could be in error by that amount. For present purposes, therefore, our Arctic Ocean budget can be considered balanced.

The GIN Sea

Figure 2 and Table 2 show the individual contributions to the fresh water budget of the GIN Sea. The yield is based on an area of 2.55×10^6 km^6. All fresh water fractions are relative to the salinity 34.93, which we estimate as the mean salinity for the GIN Sea based on composite calculations from *Carmack* [1972], *Swift* [1980], and *Dietrich* [1969], and which is about 0.13 greater than that of the Arctic Ocean. We note that the various calculations are not very sensitive to slightly different selections of reference salinity. For example, even if the reference is changed by 0.02, which is probably the maximum error in the mean salinity estimate for the GIN Sea, the most sensitive individual fresh water flux in the GIN Sea would change by less than 130 $km^3\ yr^{-1}$ and the net flux by less than 180 $km^3\ yr^{-1}$. Such changes are indistinguishable from zero in this budget. The various sources and sinks were calculated as follows:

Ice import through Fram Strait. This flux appears to represent the largest contribution of fresh water. Primarily because of difficulties in determining ice thickness, present flux estimates probably have a very large margin of uncertainty, although the two most recent studies [*Wadhams*, 1983; *Vinje and Finnekåsa*, 1986] agree to within about 20%. The divergence of ice flow in Fram Strait, leading to the wintertime production of new ice in open water [cf. *Scientific Committee on Oceanic Research Working Group 58, (SCOR)*, 1979], introduces additional uncertainty in the estimates. For the moment we accept the later and more extensively based of these ice transport estimates (namely, 0.16 Sv, i.e., that of *Vinje and Finnekåsa* [1986]) but note that this estimate represents the southward flux of ice across the parallel 81°N, which is at the northern extremity of Fram Strait. *Untersteiner* [1988] has pointed out that in the eastern part of this region the ice is rapidly melted by the warm water flowing northward with the West Spitsbergen Current; on the basis of a steady heat-balance model, he has estimated that about 0.06 Sv of the southward ice flux melts in the northeastern part of Fram Strait and is incorporated into the mixed layer beneath the ice. This would then leave 0.10 Sv to be exported to the GIN Sea as ice. If we assume the mean salinity of this ice to be 4 [*Östlund and Hut*, 1984], the ice flux represents a fresh water addition to the GIN Sea of 2790 km^3 yr^{-1} (110 cm yr^{-1}).

The fate of the meltwater produced in *Untersteiner's* [1988] model is not clear, although a number of recent studies [*Aagaard et al.*, 1987; *Bourke et al.*, 1988; *Gascard et al.*, 1988] suggest that a majority of it should recirculate southward and join the East Greenland Current. If all of it were to recirculate, its fresh water contribution would represent a southward flux of 1680 km^3 yr^{-1}, which is considerably larger than our estimate of the total liquid fresh water load carried by the East Greenland Current (compare below). The latter also includes the contribution from the low-salinity upper layer in the Arctic Ocean, and while the uncertainties in these various estimates and arguments are too large to allow more than speculation, there is some suggestion that if the ice melt model is approximately correct, then a significant portion of the Fram Strait ice melt is carried farther into the Arctic Ocean, rather than recirculating in Fram Strait.

Ice export through Denmark Strait. *Moritz* [1988] has studied the areal (two-dimensional) ice budget of the Greenland Sea and from that has estimated (R. E. Moritz, personal communication, 1989) that on an annual average about one-half the ice import through northern Fram Strait melts north of 73°N. If we extrapolate this melt rate, then about 80% of the original ice import melts north of Denmark Strait, leaving 560 km^3 yr^{-1} (-22 cm yr^{-1}) to be exported still in the form of ice.

Liquid fresh water import through Fram Strait. On the basis of long-term moored measurements, *Foldvik et al.* [1988] have estimated a transport of Polar Water ($T < 0$°C) from the Arctic Basin through Fram Strait near 79°N of 1.0 Sv; the transport temperature is -1.49°C. From Table 2 of *Aagaard and Greisman* [1975] the latter is seen to correspond to a salinity less than 34.0. Examination of various sections across the northern East Greenland Current [e.g., *Paquette et al.*, 1985] suggests 33.7 as representative. The corresponding fresh water addition to the GIN Sea is 1110 km^3 yr^{-1}. The deeper water has a mean salinity close to the reference value of 34.93 and will not contribute much to the fresh water flux. For example, *Foldvik et al.* [1988] estimate the transport of Arctic Intermediate Water as 2.0 Sv with a transport temperature of 1.3°C. Reference to *Aagaard and Greisman* [1975] and *Paquette et al.* [1985] shows this to correspond to a salinity very near 34.90, yielding a fresh water flux with this water mass of only 50 km^3 yr^{-1}. Adding this to the upper water flux gives 1160 km^3 (46 cm yr^{-1}) as the annual liquid fresh water import with the East Greenland Current.

Liquid fresh water export through Denmark Strait. Using a combination of dynamic sections and direct current measurements, *Malmberg et al.* [1972] have estimated that 1.6 Sv exit Denmark above 300 m. Waters in this depth interval would include most of the liquid fresh water flux, since the salinity deeper than 300 m is generally well above 34. The four sections taken by *Malmberg* [1972*b*] in August 1971 suggest a mean salinity in the upper 300 m at the western end of about 34.1. Figure 8 of *Aagaard and Coachman* [1968] suggests mean summer salinities at least that high and mean winter values several tenths of a psu higher. At this point we can find no persuasive evidence for assigning to this outflow an annual mean salinity less than 34, which is coincident with *Mosby's* [1962] estimate; hence we adopt a value of 34.1 together with the volume transport of *Malmberg et al.* [1972]. In addition, there is a small contribution from the outflow of dense waters. For the latter, *Ross* [1978] gives a transport of 2.5 Sv, and *Swift et al.* [1980], give a salinity of 34.85 or slightly more. Combining these various values yields a liquid fresh water flux through Denmark Strait corresponding to a loss to the GIN Sea of 1520 km^3 yr^{-1} (-60 cm yr^{-1}).

Precipitation less evaporation. From *Gorshkov's* [1983, pp. 68–69] atlas we estimate the mean annual precipitation to exceed evaporation by 31 cm yr^{-1}. This is about 15 cm yr^{-1} greater than the estimate by *Vowinckel and Orvig* [1970] if their value, which is for the combined GIN and Barents seas, is adjusted to include only the GIN Sea. On the other hand, it is about 20 cm yr^{-1} less than estimated by *Mosby* [1962], again originally for the combined GIN and Barents seas, but here adjusted to include only the GIN Sea. We therefore accept the intermediate value of 31 cm yr^{-1} computed from *Gorshkov* [1983]. This represents an annual fresh water flux of 790 km^3.

Runoff. Three land masses contribute runoff to the GIN Sea: Norway, Greenland, and Iceland. *Mosby* [1962] has estimated the annual runoff along the Norwegian coast to be 350 km^3. This includes the amount discharged directly into the Barents Sea as well as that entering the Skagerrak from southern Norway (the latter subsequently to be imported with the Norwegian Coastal Current). About 70% of the total runoff along the Norwegian coast can be considered as discharging directly into the eastern GIN Sea. For the western side of the GIN Sea, *Reeh* [1985] has calculated that 54 km^3 of glacial ice is discharged annually along east Greenland north of Denmark Strait, and if two thirds of this melts in the area [*Mosby*, 1962], 36 km^3 can be considered to actively enter the fresh water budget. Calculations of meltwater runoff from east Greenland have a meager observational base, but Weidick's unpublished estimate (quoted by *Reeh* [1985]) that for the entire Greenland ice sheet the runoff of meltwater exceeds the loss from calving by 40%, applied proportionally, yields an annual meltwater addition north of Denmark Strait of about 75 km^3. Finally, *Stefansson* [1962] has estimated the annual runoff from the north coast of Iceland to be 62 km^3. A total fresh water addition to the GIN Sea from all sources of runoff is therefore probably near 420 km^3 yr^{-1} (16 cm yr^{-1}).

Fresh water import with the Norwegian Coastal Current. From inverse calculations of the geostrophic circulation, *Gammelsrød and Hackett* [1981] have determined the transports of fresh water (relative to the salinity 35.2) and total mass through a section from the southern Norwegian coast to Denmark during spring and fall, and through immediately adjacent sections during summer. The spring and fall sections were single occupations during different years, but the summer sections represent 10-year means. We have recalculated their results relative to a reference salinity of 34.93 and have extended their estimates to an annual mean transport by taking the winter transports to be the mean of the fall and spring values. In this manner we estimate the contribution of fresh water to the GIN Sea by the Norwegian Coastal Current to be 950 km^3 yr^{-1} (37 cm yr^{-1}).

Fresh water export with the Norwegian Coastal Current. The hydrographic and current measurements of *Blindheim* [1989] suggest that the transport into the Barents Sea within the coastal wedge of low-salinity water off northern Norway is about 0.7 Sv, and comparison of his salinity data with those of *Dickson and Blindheim* [1984] indicates that the mean salinity is not less than about 34.4. These values represent a loss of fresh water to the GIN Sea of 330 km^3 yr^{-1} (−13 cm yr^{-1}).

Saline water import from the North Atlantic. *Dooley and Meincke* [1981] have described the saline inflow as composed of two water masses with slightly different history: Atlantic Water, with a salinity near 35.4, which flows northward over the Scottish continental slope; and Modified Atlantic Water, with a salinity about 0.2 less, which has crossed the Iceland-Faeroe Ridge, flowed southward immediately east of the Faeroes, and then recirculated in the Faeroe-Shetland channel. They estimated that 2.0 Sv enters the GIN Sea as Atlantic Water, and another 1.3 Sv as Modified Atlantic Water. Their total inflow of 4.1 Sv also includes 0.8 Sv of a less saline recirculated arctic water mass which will not enter our calculations. These estimates were based on measurements over a 1-month period during late summer. *Gould et al.* [1985] have shown that there is a large seasonal variation in the inflow, with summer transports considerably less than the annual mean. However, the total northward flow of 4.1 Sv reported by *Dooley and Meincke* [1981] is consonant with the summer values of *Gould et al.* [1985], so that the two data sets appear compatible. We have therefore extended *Dooley and Meincke's* [1981] 1-month calculations for the two separate saline water masses to an annual value by increasing them proportionally, so that the total annual mean northward flow through the Faeroe-Shetland Channel coincides with the estimate of *Gould et al.* [1985]. This yields an Atlantic Water influx of 3.7 Sv and one of 2.4 Sv for the Modified Atlantic Water. The combined inflow represents a fresh water deficit for the GIN Sea of 2160 km^3 yr^{-1} (−85 cm yr^{-1}).

Saline water export to the Barents Sea. Based on moored current measurements, but of limited duration, *Blindheim* [1989] has estimated the total outflow to the Barents Sea to be 3.1 Sv. About 0.7 Sv is accounted for by the low-salinity water carried by the Norwegian Coastal Current (compare above), and the remaining 2.4 Sv then represents the transfer of saline Atlantic Water to the Barents Sea. The total outflow of 3.1 Sv is about 40% larger than that indicated by *Blindheim and Loeng's* [1981] mean dynamic sections and is thus of reasonable magnitude, considering the likelihood of an additional barotropic contribution to the flow. However, *Blindheim's* [1989] salinities represent the anomalous low-salinity conditions of the late 1970s, and therefore we have turned to *Dickson and Blindheim*'s [1984] sections of long-term mean salinity to calculate the transport salinity, namely 35.05. The corresponding salt flux into the Barents Sea is equivalent to a fresh water gain by the GIN Sea of 260 km^3 yr^{-1} (10 cm yr^{-1}).

Omissions. Several fresh water sources and sinks have been omitted in our budget calculations for the GIN Sea. First, we have neglected the import of saline water through Denmark Strait with the Irminger Current. Much of this water recirculates in Denmark Strait, and *Stefansson* [1962] has estimated that only 0.36 Sv of Atlantic water with a core salinity of 35.15 actually rounds the northwest coast of Iceland. This corresponds to a fresh water deficit of only about 70 km^3 yr^{-1}.

Second, we have neglected the import south of Spitsbergen of low-salinity water from the Barents Sea. It appears that most of this water is carried northward along the west coast of Spitsbergen in a layer perhaps 100 m deep and 20 km wide (see, for example, Figure 3 of *Aagaard et al.* [1987]). Unpublished calculations from the latter data set suggests that this layer has a mean salinity near 34.2, and if we take the mean speed to be 15 cm s^{-1} (Figure 14 of *Hanzlick* [1983]), the fresh water flux corresponds to an addition of about 200 km^3 yr^{-1} to the GIN Sea. This may be compared with *Blindheim's* [1989] 0.4-Sv estimate of the outflow from the Barents Sea of low-salinity water south of Bear Island, based on 6 weeks of moored current records. The mean salinity of the latter water is probably near 34.4 (his Figure 10), representing a fresh water flux of 190 km^3 yr^{-1}. The agreement is somewhat deceptive, for some of this water is probably recirculated north of Bear Island, and additional low-salinity waters flow in from the northeast immediately south of Spitsbergen. Nevertheless, because the combined flow appears to be trapped along the west coast of Spitsbergen, it is likely that its fresh water component (which in any case is of second order in the budget) is largely carried back into the Arctic Ocean and does not in the net contribute significantly to the GIN Sea fresh water budget.

Third, we have neglected the export to the Arctic Ocean of saline water carried by the West Spitsbergen Current. *Hanzlick* [1983] has estimated that the total transport by this current above 600 m is 3.7 Sv, but much of this flow recirculates in Fram Strait. For example, *Bourke et al.* [1988] suggest that only 20% of the baroclinic flow in northern Fram Strait continues into the Arctic Ocean; this would principally represent the inshore branch of the West Spitsbergen Current [cf. *Aagaard et al.*, 1987]. If a similar proportion is also applicable to the total flow, then about 0.7 Sv leaves the Greenland Sea. Alternatively, Figures 4 and 5 of *Aagaard et al.* [1987] suggest the width of the inshore branch to be no more than 20 km, and if we assume a mean depth of 400 m and a mean speed of 15 cm s^{-1} [*Hanzlick*, 1983], the transport is at most 1.2 Sv. Taking a mean salinity of 34.98 (Figure 5 of *Aagaard et al.* [1987] or Figures 5 and 11 of *Bourke et al.* [1988]) and an intermediate transport value of 1 Sv, the contribution to the fresh water budget is less than 50 km^3 yr^{-1}. Even if half the West Spitsbergen Current transport estimated by *Hanzlick* [1983] were lost to the Arctic Ocean (instead of recirculating) and its mean salinity were as much as 35.00, this would represent a fresh water gain to the GIN Sea of less than 120 km^3 yr^{-1}.

Finally, we have neglected the effects on the fresh water budget of deep exchanges through Fram Strait and the Faeroe-Shetland passage, as well as the inflow of saline water from the Barents Sea resulting from brine rejection. The salinity of the deep waters actually being exchanged through Fram Strait and the Faeroe-Shetland passage is probably within 0.01–0.02 of our reference salinity of 34.93, and the fresh water flux equivalents are therefore negligible. Similarly, while at least a part of the dense drainage from the Barents Sea is of high salinity, the transports are very small [*Quadfasel et al.*, 1988], so that the effects on the fresh water budget are again negligible, of the order of 20 km^3 yr^{-1}.

The net surplus. Our fresh water budget estimate for the GIN Sea shows a surplus of 1800 km^3 yr^{-1} (70 cm yr^{-1}), which imbalance would be sufficient to reduce the mean salinity of the GIN Sea by 0.015 yr^{-1}. We have examined the individual term calculations with an eye toward how much their credible adjustment could contribute toward a balanced fresh water budget, and the largest advective terms in particular have sufficient uncertainty in their determination that a balanced budget is easily conceivable without obviously violating observational constraints. For example, if we assume a mean outflow speed north of the sill in Denmark Strait of 7 cm s^{-1} (which is compatible with *Ross*'s [1977] observations for a 5-week period in August 1973) and combine it with a summer section of low salinity (with a mean in the upper 500 m near 33.75, based on *Ross*'s [1982] observations for the same period), the fresh water flux would be about 2600 km^3 yr^{-1}. If we further assume that this fresh water transport represents summer conditions and apply it to 6 months of the year, letting the other 6 months be represented by our earlier estimate of 1.6 Sv with a mean salinity of 34.00, then the annual fresh water flux corresponds to a loss to the GIN Sea of 2140 km^3 yr^{-1} (-84 cm yr^{-1}); this includes the small contribution from the deep outflow. Under these assumptions, the fresh water excess in the GIN Sea would be reduced to 1120 km^3 yr^{-1} (44 cm yr^{-1}); i.e., over one third of the excess in Table 2 would be eliminated by this adjustment of a single advective term.

FRESH WATER STORAGE

The Arctic Ocean

Except for portions of the shelf seas, particularly during winter, the Arctic Ocean is generally strongly salinity-stratified and therefore allows only shallow local convection. However, within the Arctic Ocean there are significant differences between the fresh water content of the Eurasian and Canadian basins, with stratification in the former being significantly less, despite its proximity to the very large runoff from the Eurasian land mass (Figure 3). It is therefore conceivable that climatic changes might be differently manifested in the two major basins.

The hydrographic data base for the Arctic Ocean is extremely small, particularly with respect to high-quality deep stations. Nonetheless, a useful estimate of fresh water storage is possible. To this end, we have used the atlases of *Gorshkov* [1983] and *Treshnikov* [1985] for the deep basins. *Gorshkov* [1983] defined seven domains over the basins within which *T-S* correlations are similar, and he determined mean correlations for each domain. From these correlations we have estimated the volume of water within coarse (0.5 psu) salinity intervals, and then converted the estimates to fresh water content for each domain. We have also calculated the amounts of fresh water in the shelf seas (for depths less than 500 m), using *Treshnikov*'s [1985] volume tabulations together with salinity distributions from *Codispoti and Richards* [1968], *Hanzlick and Aagaard* [1980], *Aagaard et al.* [1981], *Gorshkov* [1983], *Pfirman* [1985], *Treshnikov* [1985], and *Macdonald et al.* [1987]. The calculations for the deep basins and for the shelf seas are all relative to 34.93, which is close to the mean salinity of the deep water masses, and the storage thus calculated is a measure of the total salinity stratification. Note that for computing salinity stratification, we do not use the basin-wide mean salinity (34.80) appropriate to computing the fresh water budgets for the Arctic Basin.

The results are shown in Figure 4. We estimate the mean liquid fresh water storage in the Arctic Ocean to be 80,000 km^3. Of this, 22,000 km^3 occurs on the shelves and 58,000 km^3 in the deep basins. Of the latter, the Canadian Basin contains 45,800 km^3, and the Eurasian Basin contains 12,200 km^3. We note two important features of the fresh water distribution. First, the storage varies greatly across the deep basins of the Arctic Ocean, progressing from the largest values in the Beaufort Sea to the smallest in the southwestern Eurasian Basin. Second (but not apparent in Figure 4), the distribution of fresh water over the range of salinities is bimodal, with maxima near 33 and 34. These volume maxima represent the modal characteristics of upper waters in the Canadian and Eurasian basins, respectively. (In the Eurasian Basin, the salinity increases rapidly with depth, reaching 34.9–35.0 at about 200 m, while the temperature remains below −1.5°C to 150 m before increasing. In the Canadian Basin the halocline is deeper, and the surface salinity is lower.) Mechanistically, the bimodal distribution must of course reflect the principal sources and sinks of fresh water, and in fact in the eastern Canadian Basin, where there is a temperature minimum and nutrient maximum centered on the salinity 33.1, the volumetric mode probably reflects inflow through the western Bering Strait, which occurs in a relatively narrow salinity range.

In addition to liquid storage, fresh water is also stored in sea ice. Figures 4–45 and 4–75 of *Parkinson et al.* [1987] suggest that the annual mean ice covered area of the Arctic Ocean (including the Barents and Kara seas) is about 6.5×10^6 km^2 but varies seasonally by $3.5–4 \times 10^6$ km^2, about one third of the variability occurring in the Barents and Kara seas. From Figure 18 of *Hibler* [1979] we estimate a mean ice thickness for the Arctic Ocean of 3 m, taking into account the thinner ice typical of the shelf areas. If as before we assume the bulk salinity of the ice to be 4 (compare also Figure 3-1 of *Parkinson et al.* [1987]), the mean fresh water volume stored in sea ice is about 17,300 km^3, which is over 20% of that stored in liquid form.

The Convective Gyres

The convective gyres of the Greenland and Iceland seas are major windows on the deep ocean through which are transmitted properties acquired at the sea surface. For example, dense waters formed in the Iceland Sea during winter are a major component of the Denmark Strait overflow which ventilates the deep North Atlantic, and the burden of these dense waters has included large amounts of bomb tritium [*Swift et al.*, 1980].

Within the GIN Sea, deep convection is restricted to the cyclonic gyres because of their very low stratification; elsewhere, the density structure is too pronounced to allow effective ventilation. Except for a seasonal temperature gradient in the upper ocean which is easily broken down in winter, the stratification in the cyclonic gyres depends on a

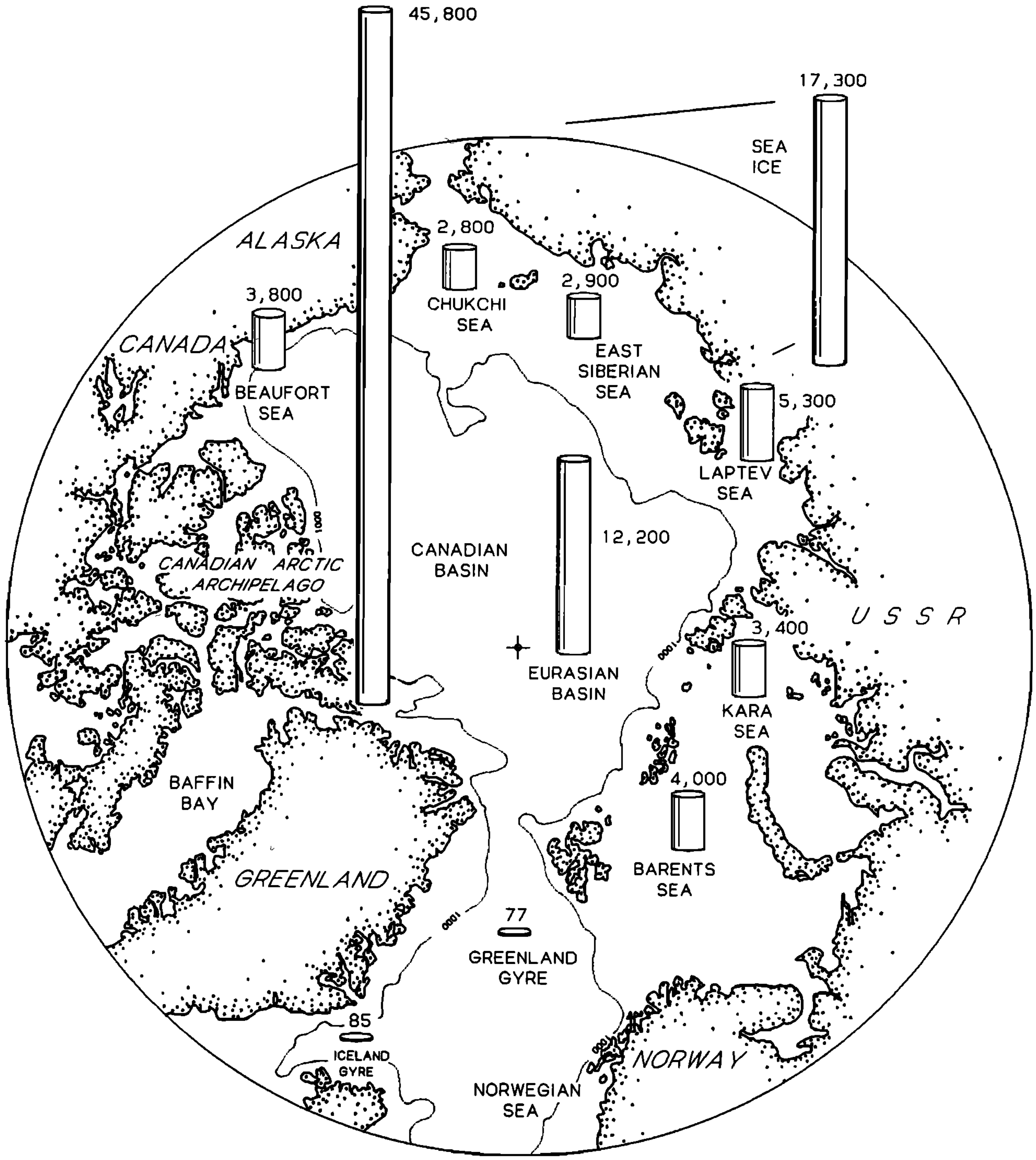

Fig. 4. The distribution of fresh water storage in the Arctic Ocean and the GIN Sea. The placement of each bar indicates the region to which it is applicable.

slightly reduced salinity in the upper ocean. To examine the origin of this stratification, we have calculated the freshwater content of the gyres as follows. For the Greenland Sea, the region of very low stratification can be estimated from *Carmack* [1972, Figures 3, 40, and 41] to encompass 135,000 km^2. The same data sets suggest the salinity stratification to be contained in the upper 200 m or less, with a salinity deficit relative to the deeper water (which has a density approaching 28.1 in σ_θ) of not more than 0.1 [e.g., *Carmack*, 1972, Figures 4, 31, and 53]. The fresh water content in the convective region does therefore not exceed 77 km^3, corresponding to 57 cm or less over the 135,000 km^2 of the central gyre. For the Iceland Sea, we estimate from Figure 45 of *Swift* [1980] that the convective area covers about 140,000 km^2. If we take the base of the stratified water column to be represented by 28.05 in σ_θ [*Swift et al.*, 1980, Figure 9], which in the convective area lies at a depth near 265 m and is overlain by water with a salinity deficit of about 0.08 [*Swift*, 1980, Figures 69 and 82], the fresh water content in the area is 85 km^3, corresponding to 61 cm over the 140,000

km^2. The fresh water content of the two gyres is therefore comparable.

There are two possible sources for this fresh water: local excess precipitation over evaporation, and inflow of ice and low-salinity water from the East Greenland Current. Waters to the east, representing Atlantic influence, are more saline than those in the gyres. We now assume that winter convection annually stirs the water column in the central gyres sufficiently to remove the salinity deficit in the upper few hundred meters, i.e., we assume that the salinity stratification is renewed annually by the external fresh water sources. From *Gorshkov*'s [1983, pp. 68–69] atlas, we estimate the excess precipitation in the Greenland Sea gyre to be 5 cm yr^{-1} and that in the Iceland Sea as 35 cm yr^{-1}, corresponding to an annual fresh water addition over the respective areas of 7 km^3 and 49 km^3, leaving 70 km^3 and 36 km^3 to be supplied by the boundary current. We therefore deem it likely that the principal source of stratification in the convective region of the Greenland Sea is fresh water inflow from the East Greenland Current, while in the Iceland Sea precipitation is at least equally significant. (An alternate calculation, based on *Swift*'s [1980] seasonal salt budget for an area of 120,000 km^2 in the eastern Iceland Sea, suggests that the fresh water balance there might be maintained by the excess of precipitation over evaporation alone.) In any case, the maximum suggested annual fresh water flux into the convective gyres from the boundary current of 106 km^3 is less than 3% of its initial fresh water burden in Fram Strait.

In the discussion thus far we have not considered the effect on the density stratification of extreme cooling, i.e., to the freezing point. *Carmack*'s [1972, Figure 31] calculations for the Greenland Sea gyre suggest that if the water were cooled to freezing, an increase in the mean salinity of 0.04 in the upper 200 m would be sufficient to allow deep overturn. From this perspective, the excess fresh water content of the gyre is only about 31 km^3, which can be offset by the freezing of 29 cm ice annually with a bulk salinity of 7.

For the Iceland Sea gyre, the salinity corresponding to a σ_θ value of 28.05 at freezing is 34.812, i.e., very close to the mean salinity above that density surface. This gyre can therefore overturn thermally as long as the upper ocean salinity is not reduced below 34.81. This is consonant with *Malmberg*'s [1972*a*] observation in the Iceland Sea of convection being absent during the years when the surface salinity fell to 34.7.

We therefore arrive at three conclusions. First, the upper layers of the East Greenland Current are nearly isolated from the interior convective regions of the Greenland and Iceland seas, with only a few percent of the current's upper waters presently penetrating into the interior. We do not know the means by which the fresh water transfer from the East Greenland Current into the convective gyres is effected. However, *Foldvik et al.* [1988] have shown that the turbulent heat flux across the northern part of the Polar front is small. We therefore suggest that in the Greenland Sea the fresh water transfer primarily occurs in the recirculation in the southern part of the gyre (the Jan Mayen Current).

Second, if the flux of fresh water from the boundary current were to increase slightly, convection would likely cease. For example, if we consider that the Greenland Sea can overturn thermally with the neutralization of 31 km^3 of fresh water by freezing and that the Iceland Sea can overturn thermally under present conditions with no freezing, then even if 1 m of sea ice were formed in these gyres, an influx from the boundary current of just over 250 km^3 fresh water (contrasted with the annual influx of 106 km^3 which we estimate to be representative presently) would shut down convection. Such an increased amount would still represent only about 6% of the annual fresh water load entering through Fram Strait.

Third, through its export of fresh water, the Artic Ocean ultimately controls the ocean ventilation which occurs in the Greenland and Iceland seas. The actual mechanism by which the control is exercised is the release of fresh water from the boundary current into the interior of the convective gyres, and such control is therefore intimately tied to the dynamics of the boundary current.

DISCUSSION

The importance of fresh water, including that formed in the Arctic Ocean during freezing, to the convective processes in the GIN Sea has largely been overlooked in the literature. Instead it has been argued, for example, by *Worthington* [1970], that the principal water mass transformation in the GIN Sea is the cooling of saline waters drawn in from the North Atlantic and that this transformation is the necessary precursor to the renewal of the North Atlantic Deep Water. The essence of the argument is that the initially stratified water column brought into the GIN Sea from the south, which is stable because of its large temperature gradient but is unstable with respect to salinity, is cooled sufficiently to convect and then flows back into the North Atlantic, being replaced by new warm and saline waters from the Atlantic. While this in some respects may be a useful conceptualization, it is nevertheless an incomplete one, for it ignores the fact that the dense outflows to the North Atlantic are significantly fresher than the warm inflows. Thereby it also ignores the implications of that freshening on water mass transformation and convection within the GIN Sea. (The substantial expansion of *Worthington's* [1970] arguments by *McCartney and Talley* [1984] does include a salt balance calculation for the GIN Sea but is unable to distinguish between the effects of surface fresh water exchange and advection with the boundary current.)

We have shown that the present small salinity (and density) stratification in the convective gyres in the Greenland and Iceland seas is likely maintained in part by the local precipitation excess (at least in the Iceland Sea) and in part by a lateral influx of fresh water from the East Greenland Current. When the upper waters in the gyres are cooled during winter, their slight salinity deficit is transferred to the deep water by convection. Small variations in the surface salt deficit may also be transferred downward at least to intermediate levels, but if the surface layers are freshened too much, cooling even to the freezing point will be insufficient to initiate convection: the convective gyres will be capped by a fresh water lid. Such a sequence is essentially what has been proposed in the halocline catastrophe scenarios, in which runof from rapid continental deglaciation may have diminished or even halted North Atlantic Deep Water production, with major consequences for climate and the global circulation [*Broecker et al.*, 1985]. (Note, however, that deep convection may be possible even if the entire upper ocean does not turn over, either because the convection is highly localized, for example, in chimneys [cf. *Kill-*

worth, 1979] or is driven through double-diffusive fluxes [*Carmack and Aagaard*, 1973; *McDougall*, 1983].)

We here suggest that the present-day GIN Sea and probably also the Labrador Sea are rather delicately poised with respect to their ability to sustain convection and that we have in fact during the past several decades seen a small-scale analog of the halocline catastrophe proposed for past deglaciations. A major difference is that the present situation does not require dramatic increases in fresh water flux to effect a capping of the convection, nor does it depend on deglaciation. Rather, very modest changes in the disposition of the fresh water presently carried by the boundary current can alter or stop the convection, and the principal source of fresh water is sea ice, rather than glacial ice. The essence of the present situation is that the large fresh water output from the Arctic Ocean passes perilously close to the very weakly stratified convective gyres and that the stratification in these gyres is easily perturbed, either by variations in the discharge from the Arctic Ocean or by leaks or recirculation from the boundary current. Such changes in the stratification during winter typically result in anomalous local ice formation.

It is just such variations and perturbations which have occurred in the last few decades. *Malmberg* [1972*a*] (see also *Dickson et al.* [1975]) showed that during winter and spring of most of the years from 1965 to 1971, extremely heavy ice conditions prevailed in the Iceland Sea, with the entire north and east coasts of Iceland enveloped by ice during the extreme years of 1965 and 1968. Furthermore, the severe ice years were characterized by low upper water salinities north of Iceland, which *Malmberg* [1972*a*] showed were advected into the Iceland Sea from the northwest. The critical salinity appeared to be about 34.7, at which surface value the water column would not overturn, thus allowing the formation and preservation of sea ice. This is therefore an example of fresh water capping at least a portion of the convective region in the Iceland Sea, with the anomalously strong fresh water influx originating in the East Greenland Current. In effect, the polar hydrographic domain of the boundary current expands into the interior during such events.

A somewhat similar situation has been described for the Labrador Sea by *Lazier* [1980], who found that during 1968–1971 the near-surface salinity in the convective region fell to 34.4–34.6, a change of about 0.2, thereby limiting convective renewal to the upper 200 m of the ocean. However, since the convective gyre in the Labrador Sea is considerably warmer than those in the GIN Sea, with temperatures greater than 3°C, its invasion by water of anomalously low salinity does not result in local ice formation.

Recently, *Dickson et al.* [1988] have examined these events in a larger perspective, as part of the so-called "great salinity anomaly" which freshened much of the upper northern North Atlantic during the past 25 years. From its first observation in the Iceland Sea in the mid-1960s, this large salt deficit can be traced as it circulated around the subpolar gyre, passing through the Labrador Sea by 1972, then propagating back across the North Atlantic and into the GIN Sea in the mid-1970s through the Faeroe-Shetland channel. The bulk of the anomaly appears to have passed through a given area within 2–3 years. *Dickson et al.* [1988] estimated that the salt deficit being advected through the Labrador Sea was about 72×10^9 tons, which is equivalent to a fresh water excess of 2000 km^3, i.e., about one-half the annual fresh water transport of the East Greenland Current as it enters the GIN Sea from the Arctic Ocean. The "great salinity anomaly" can therefore be accounted for by a moderate perturbation of the outflow from the Arctic Ocean, for instance, a 2-year period of fresh water flux 25% above normal. Since the fresh water storage within the Arctic Ocean approaches 100,000 km^3, the effect on the Arctic Ocean fresh water reservoir of such a withdrawal is negligible and could conceivably be maintained for decades.

The apparent feasibility of the Arctic Ocean as the source of the North Atlantic salinity anomaly can be contrasted with an origin north of Iceland, as hypothesized by *Dickson et al.* [1988]. If we assume that 10^5 km^2 of the Iceland Sea which does not normally freeze is stabilized so that it does not convect, and if 1 m ice of bulk salinity 7 forms there, about 40 km^3 fresh water will be distilled out of solution. If we further assume that the excess of precipitation over evaporation (35 cm yr^{-1}) is allowed to accumulate in this area, adding another 35 km^3 annually, and allow these events to persist over 2 years, a total of 150 km^3 of fresh water will be segregated. However, even such an extreme scenario would provide less than 8% of the fresh water excess estimated to have passed through the Labrador Sea, making the Iceland Sea an unlikely source of the North Atlantic salinity anomaly. This calculation also implies that the feedback effect on the overall fresh water budget of local freezing in the convective gyres, when they are stabilized by an anomalous influx of fresh water from the boundary current, is of second order or less relative to the Arctic Ocean outflow. This of course does not diminish the local importance of increased ice cover during periods of increased water column stability, such as has recently been suggested by the ice anomaly analysis of *Mysak and Manak* [1989], in which they point out the coincidence of heavy ice years and the passage of the North Atlantic salinity anomaly.

We think it likely that the control of the convective gyres by the fresh water flux from the boundary current has a range of manifestations. In the extreme case of a relatively large flux into the interior, the convection will cease; this is the scenario of the halocline catastrophe. In the case of more modest fresh water additions, the convective products will either be maintained at their normal reduced salinity relative to the Atlantic inflow (about 0.5 psu) or freshened slightly further, as appears to have happened recently to the Denmark Strait overflow into the North Atlantic. The latter situation resulted in a significantly fresher deep northwestern Atlantic [*Brewer et al.*, 1983; *Lazier*, 1988]. Presumably, intermediate situations are possible, in which middepth convection (which is the main source of the Denmark Strait overflow) occurs, albeit involving waters of reduced salinity, while the deeper convection which renews the densest waters in the system is shut down.

Thus far our discussion has emphasized the suppressive effects of salinity stratification on convection. We emphasize, however, that a small amount of such stratification is necessary to renew the deep water most efficiently (compare Figure 5). Essentially, this is because a salinity gradient requires the temperature of a downward-convecting water parcel to be colder, and hence more compressible, than the underlying warmer water. For example, consider a stably stratified water column which is warmer and more saline near the surface. As the surface water cools, it will convect,

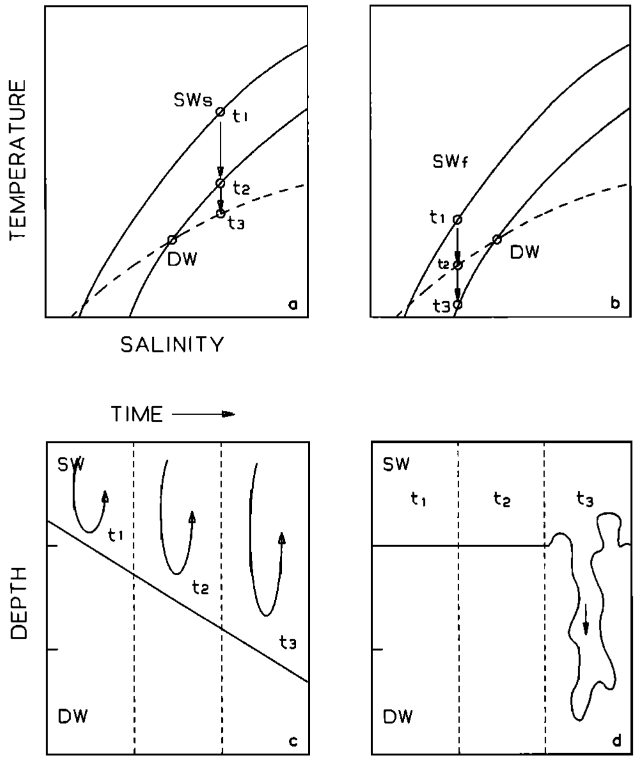

Fig. 5. Schema comparing the convective regimes associated with progressive cooling of surface water that is slightly more saline (SW_s, Figure 5*a*) and slightly less saline (SW_f, Figure 5*b*) than the ambient deep water (*DW*); also shown in Figures 5*a* and 5*b* are isolines of density relative to near-surface pressures (solid) and near-bottom pressures (dashed). As SW_s cools, it eventually reaches a temperature (t_2) at which its density is equal to that of the underlying water, and convection will ensue. However, at near-bottom pressures, the SW_s will still be less dense than DW, and further cooling (to t_3) is required to drive the convection deeper. Hence the water column will be ventilated by a progressive deepening of the upper layer (Figure 5*c*). When SW_f is cooled to the temperature (t_3) where its density at near-surface pressures matches that of DW, it has already surpassed the density of DW at all greater pressures and thus will continue to sink (Figure 5*d*).

and the water column will be ventilated by a gradual and progressive deepening of the surface mixed layer. We can contrast this situation with a water column which is stratified by a slight salinity gradient. The first stages of cooling will decrease the mixed-layer temperature but will not increase the depth of the mixed layer. Eventually, however, the surface water will be cooled sufficiently to convect, and because of the salinity stratification, this water will be significantly colder than the underlying water. Because the cold water is more compressible, it will with increasing pressure be increasingly denser than the ambient water and will continue to sink. The net effect is that the deep water will be renewed episodically without requiring the entire water column to overturn progressively.

Admittedly, the dependence of compressibility on temperature is a small effect, and its importance becomes evident only when one considers the extremely weak vertical density gradient in the convective gyres in winter. For example, suppose the surface water starts its descent into the deeper layer when their potential densities referred to surface pressure are equal: if the surface water were 0.01 less saline than the underlying water, it would be 0.3°C colder. If the densities are again compared at 3000 dbar, the relative density of the surface water has increased about 0.03 kg m^{-3}. While this change may seem small, we note that an equal effect works against renewal by surface water that is 0.01 more saline than the underlying water. The combination of salinity stratification and temperature-dependent compressibility thus provides a catastrophic short circuit of the surface-driven convective process. This simple argument may also explain the characteristic negative temperature-salinity correlations observed in convective gyres and their water mass products.

Investigators have in the past had a major concern for the mechanistic details of convection in the GIN Sea. In the process, the importance of the very modest influx of fresh water to the surface layers has tended to be overlooked. We suggest that remedying this oversight will prove a productive endeavor, not only with respect to understanding the present convective situation in the GIN Sea but also in exploring the effects of possible future perturbations in climate on the northern hemisphere thermohaline circulation.

Acknowledgments. We appreciate helpful discussions with A. Foldvik, R. Macdonald, J. Swift, and N. Untersteiner. Partial support came from the Arctic Program, Office of Naval Research, and from the Department of Fisheries and Oceans, Canada. NOAA Pacific Marine Environmental Laboratory contribution 1114.

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(Received March 27, 1989;
accepted May 1, 1989.)