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Freshwater balance and the sources of deep and bottom waters in the Arctic Ocean inferred from the distribution of $H_2{}^{18}O$

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Abstract – Data from sections across the Eurasian Basin of the Arctic Ocean occupied in 1987 and 1991 are used to derive information on the freshwater balance of the Arctic Ocean and on sources of the deep waters of the Nansen, Amundsen and Makarov basins. Using salinity, $H_2{}^{18}O$, and mass balances we estimate the river-runoff and the sea-ice melt water fractions contained in the upper waters of the Arctic Ocean and infer pathways of the river-runoff signal from the shelf seas across the central Arctic Ocean to Fram Strait. The average mean residence time of the river-runoff fraction contained in the Arctic Ocean halocline is determined to be about 11 to 14 years. Pacific water entering through Bering Strait is traced using silicate and its influence on the halocline waters of the Canadian Basin is estimated. Water column inventories of river-runoff and sea-ice melt water are calculated for a section just north of Fram Strait and implications of these inventories for sea-ice export through Fram Strait are discussed. Comparison of the $^{18}O/^{16}O$ ratios of shelf water, Atlantic water and the deep waters of the Arctic Ocean indicate that the sources of the deep and bottom waters of the Eurasian Basin are located in the Barents and Kara seas.

CONTENTS

1. INTRODUCTION

Freshwater export from the Arctic Ocean may have significant impact on the intensity of deep water formation in the Greenland, Iceland and Labrador Seas (AAGAARD and CARMACK, 1989). Although we know the mean fresh water export rates reasonably well, there are still many open questions concerning the relative proportions of sea-ice melt water and river-runoff in the Arctic fresh water, as well as the distribution of these components within the central Arctic Ocean. Understanding of the freshwater balance of the Arctic Ocean is also relevant for studies of pathways and storage times of pollutants (mainly those dissolved in seawater), because many of them are delivered to the Arctic Ocean by river discharge.

While we know the major trajectories of the drifting sea-ice (GORDIENKO and LAKTIONOV, 1958; COLONY and THORNDIKE, 1984), we know little about advection and mixing of the surface and halocline waters. It has been shown in earlier work that stable isotope studies are a valuable tool for investigation of sources and composition of freshwater in the Arctic halocline (e.g. ÖSTLUND and HUT, 1984; SCHLOSSER, BAUCH, FAIRBANKS and BÖNISCH, 1994). The reason for this is that sea-ice shows only a very small enrichment in its $^{18}O/^{16}O$ ratio relative to seawater. Arctic river-runoff, on the other hand, is strongly depleted in its oxygen isotope ratio.

In this study we extend earlier work (SCHLOSSER *et al*, 1994) from the Nansen Basin to the Amundsen and Makarov Basins of the Arctic Ocean. Because Pacific water entering the Arctic Ocean through Bering Strait has significant impact on the properties of the halocline waters in the Canadian Basin, we introduce an approach which allows us to quantify the contribution of Pacific Water to the Arctic halocline.

2. SAMPLE COLLECTION AND MEASUREMENT

The data used in this study were collected along the 1987 Nansen Basin section occupied by the German research icebraker *Polarstern* during the ARK IV/3 expedition (PSSP, 1988) and on sections covering the Nansen, Amundsen and Makarov Basins occupied by the Swedish icebreaker *Oden* in the framework of the ARCTIC 91 expedition (ANDERSON and CARLSON, 1991) (for geographic position of the stations see Fig.1). Additionally, we measured samples from the Norwegian and Greenland seas (stations 79 and 617 respectively). These stations were occupied by the German research vessel *Meteor* in 1985 (M71, station 79) and 1988 (M8, station 617). During each cruise hydrographic data and a variety of tracer data were collected (e.g. PSSP, 1988; ANDERSON, JONES, KOLTERMANN, SCHLOSSER, SWIFT and WALLACE, 1989; ANDERSON and CARLSON, 1991). Here we discuss ^{18}O and salinity data from the upper 300m of the water column obtained on sections occupied in 1987 (Barents Shelf to Gakkel Ridge, section B in Fig.1) and in 1991 (Barents Shelf to the Makarov Basin, section A in Fig.1; Yermak Plateau to Morris Jessup Plateau, section C in Fig.1). We also evaluate ^{18}O data collected on stations in the central basins of the Arctic Ocean and the central Greenland and Norwegian seas covering the entire water column (stations 358, 16, 26, 33, 617 and 79 in Fig.1).

Initially, a small number of the ^{18}O halocline samples collected in 1987 were measured at the Institut für Umweltphysik at the University of Heidelberg (SCHLOSSER *et al*, 1994). Precision of these $^{18}O/^{16}O$ ratios was typically ±0.07‰. Most of the ^{18}O samples were measured at the Lamont-Doherty Earth Observatory using a MAT 251 mass spectrometer after equilibration of the water samples with CO_2 following procedures described by ROETHER (1970) and FAIRBANKS (1982). ^{18}O results are reported in the δ notation where $\delta^{18}O$ is the permille deviation of the $H_2{}^{18}O/H_2{}^{16}O$ ratio of the sample from that of SMOW (Standard Mean Ocean Water). Precision of the L-DEO $\delta^{18}O$ data is about ±0.02 to ±0.03‰.

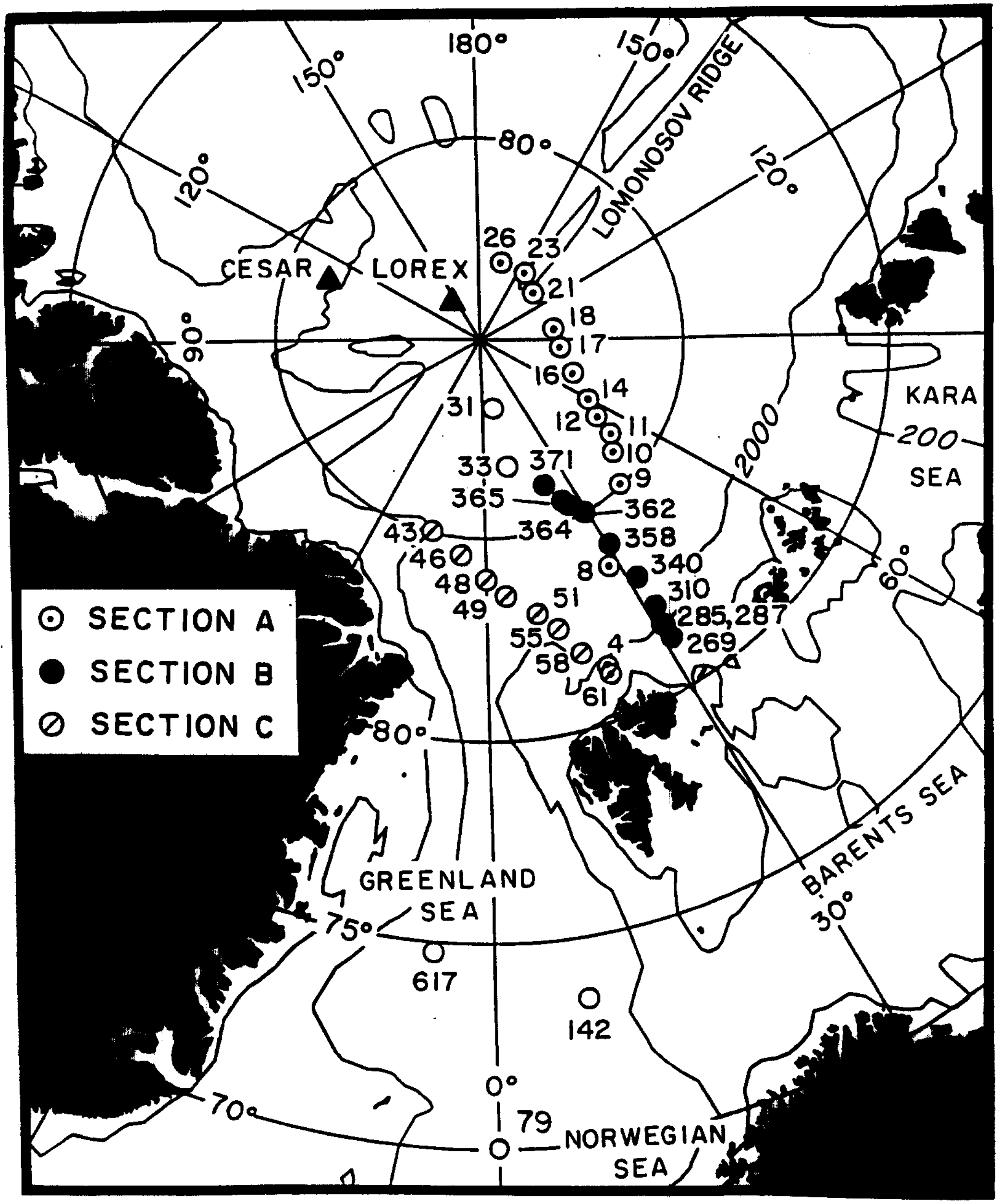

FIG. 1. Geographic positions of the stations used in this study. Section A shown in Fig.2 extends from station 61 (Nansen Basin) to station 26 (Makarov Basin), section B from station 269 (Barents Sea) to station 371 (Gakkel Ridge), and section C from station 43 (Morris Jessup Plateau) to station 61 (Yermak Plateau). Sections A and C were occupied during the ARCTIC 91 expedition, section B represents the 1987 *Polarstern* section across the Nansen Basin. Stations 617 (Greenland Sea) and 79 (Norwegian Sea) were occupied during cruises of R/V *Meteor* in 1988 and 1985, respectively.

3. HYDROGRAPHIC FEATURES OF THE UPPER LAYERS OF THE ARCTIC OCEAN

The upper layers of the Arctic Ocean are characterized by a shallow surface mixed layer (SML) with temperatures close to the freezing point of seawater overlying a strong halocline. The halocline is maintained by water advected from the shelf seas (AAGAARD, COACHMAN and CARMACK, 1981). The shelf waters in the Eurasian Basin of the Arctic Ocean consist primarily of Atlantic-derived water with addition of freshwater which is a mixture of river-runoff, sea-ice melt water and brine released during sea-ice formation. Atlantic water underlying the halocline enters the Arctic Ocean via the West Spitzbergen Current, as well as via the Barents and Kara Seas (AAGAARD, SWIFT and CARMACK, 1985; MIDTTUN, 1985; PFIRMAN, BAUCH and GAMMELSRØD, 1994). The core of Atlantic-derived water is reflected in a distinct temperature maximum below the halocline waters at about 300 to 500m depth throughout the Arctic Ocean (COACHMAN and BARNES, 1963).

Pacific water entering the Arctic Ocean through Bering Strait contributes to the surface and halocline waters, primarily in the Canadian Basin.

Chemical properties can be used to distinguish between lower and upper halocline water. Upper Halocline Water (UHW) is associated with a salinity of 33.1 and a nutrient maximum. Part of the nutrient maximum originates in the Pacific but most of this signal is imprinted by interaction of the Pacific inflow with the sediments of the Chukchi and East Siberian seas (KINNEY, ARHELGER and BURRELL, 1970; MOORE, LOWINGS and TAN, 1983; JONES and ANDERSON, 1986; WILSON and WALLACE, 1990; JONES, ANDERSON and WALLACE, 1991). Lower Halocline Water (LHW) has a salinity of about 34.25 and a pronounced NO minimum (NO=$[O_2]$+9 $[NO_3]$, BROECKER, 1974). It is formed on the shelves of the Barents and Kara Seas (JONES and ANDERSON, 1986; JONES *et al*, 1991). In the Eurasian Basin the presence of UHW is fairly restricted while LHW is typically observed throughout the Arctic Ocean (JONES and ANDERSON, 1986; JONES *et al*, 1991).

4. RESULTS

The general features of the $\delta^{18}O$ distribution in the upper waters of the Eurasian Basin (Fig.2a) are low values in the surface waters which increase with depth to about 0.25 to 0.3‰ at 300 to 500m depth, i.e. in the core of the Atlantic Water. The surface values are relatively high on the Barents Shelf (about 0.25‰ at station 269 and about 0.03‰ at station 61, i.e. similar to the Atlantic core; see Fig.2a) and decrease to the north reaching minimum values of about -2.54‰ at station 26 in the Makarov Basin. In section C north of Fram Strait, the minimum $\delta^{18}O$ value is -2.87‰ at station 43 at the northwestern end of the section. The trend of decreasing $\delta^{18}O$ values towards the surface and the north (or west) reflects the contribution of river-runoff to the surface waters of the central Eurasian Basin and in the Canadian Basin. This is also seen as a strong $\delta^{18}O$ gradient in the upper 100 to 150m in the central Nansen Basin with lower $\delta^{18}O$ values found in the northern part of the section.

The salinity distribution (Fig.2b) is similar to the $\delta^{18}O$ distribution with low values at the surface which increase to about 34.9 at 300m depth. The surface values are relatively high on the Barents Shelf and decrease toward the north. The surface salinities on the Barents Shelf do not reach the values of the Atlantic core as in the case of $\delta^{18}O$. They stay at about 34.0 or lower. The surface salinity distribution shows a relatively smooth south/north decrease in contrast to the $\delta^{18}O$ distribution.

Below the Atlantic water, $\delta^{18}O$ values observed at stations in the central Arctic Ocean and in the Greenland/Norwegian seas are more or less constant throughout the water column. There seems to be a slight increase in $\delta^{18}O$ of about 0.03 to 0.04‰ in Arctic Ocean waters below about 2,600m (Fig.3). The deep profiles in the Greenland and Norwegian Seas do not show this trend (Fig.3, see also Fig.11).

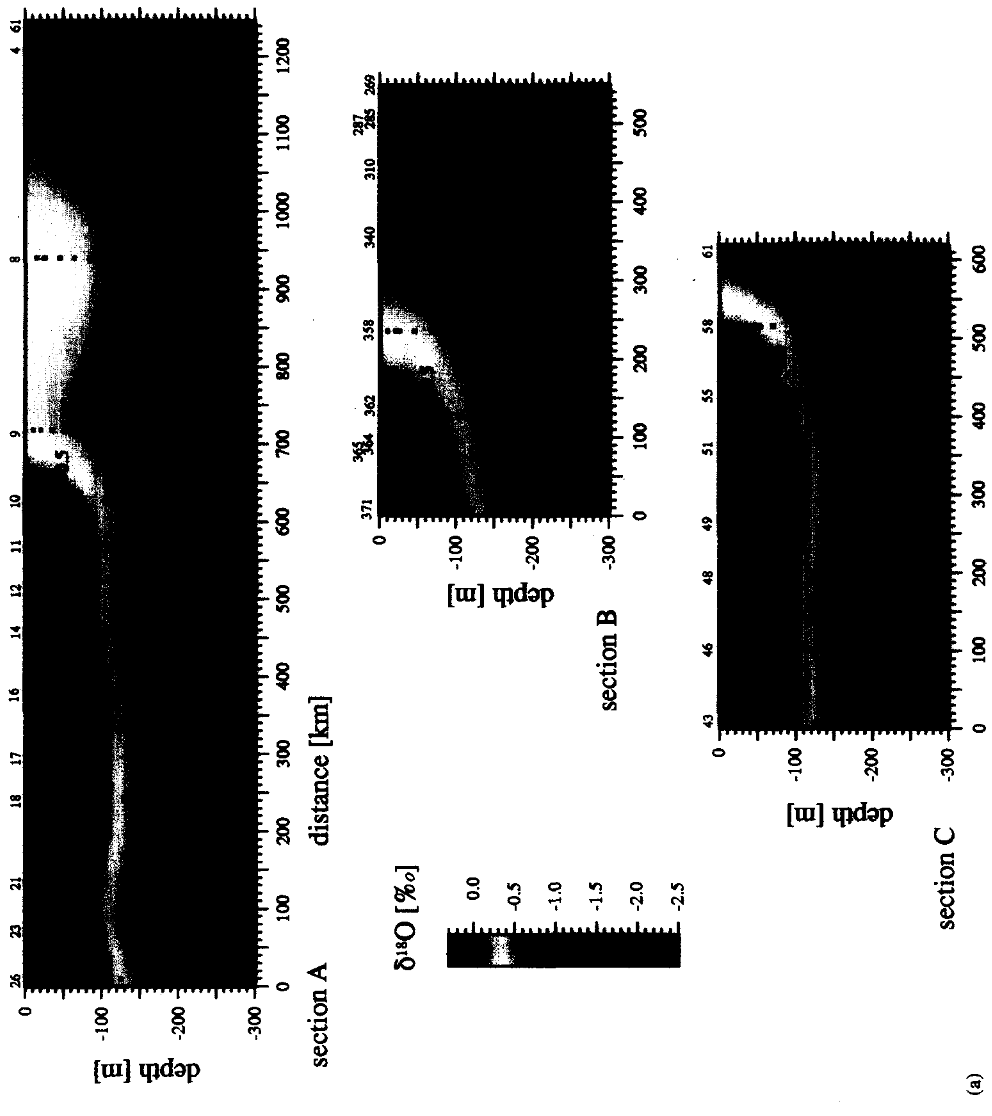

FIG.2. (a) $\delta^{18}O$ distribution in the upper 300m of the water column for three sections across the Eurasian Basin; for geographical position of the stations see Fig.1

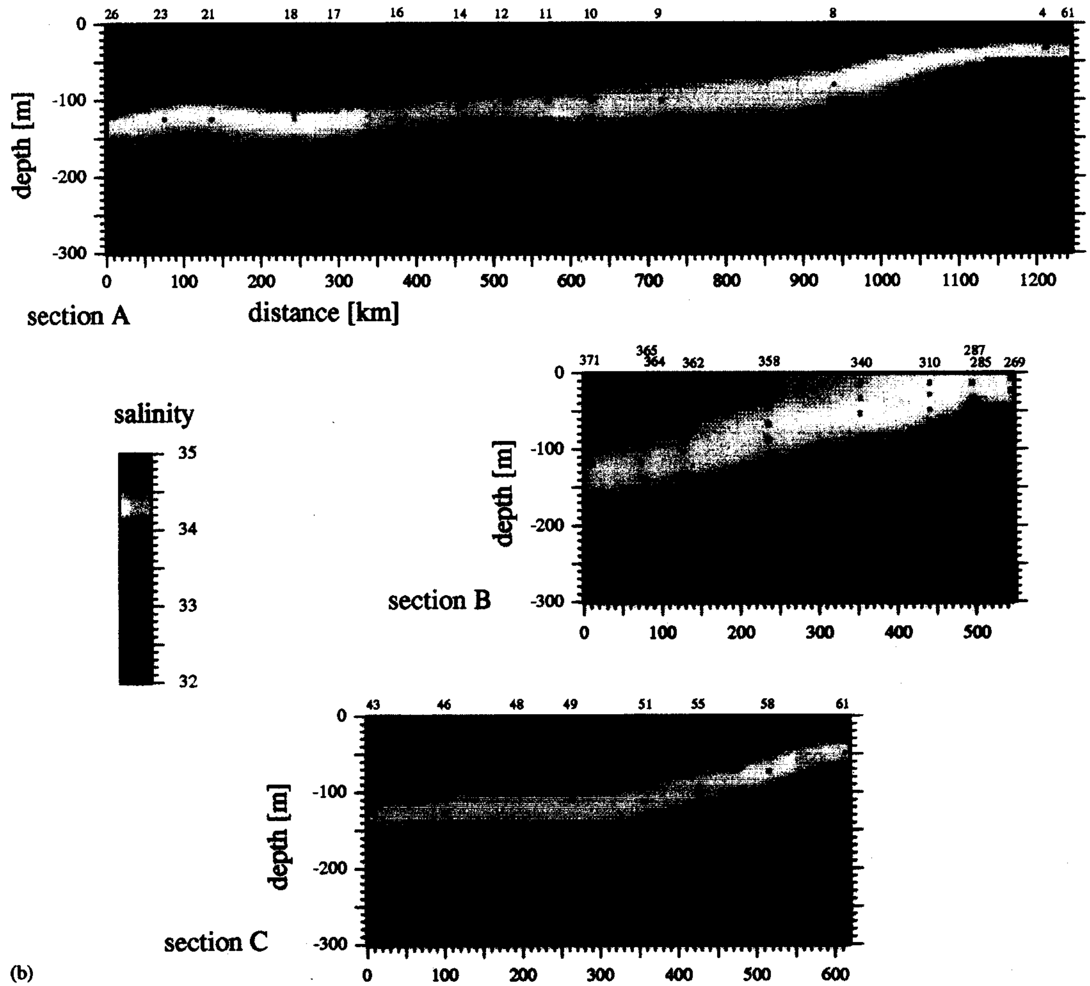

FIG.2. (b) salinity distribution in the upper 300m of the water column for three sections across the Eurasian Basin; for geographical position of the stations see Fig.1.

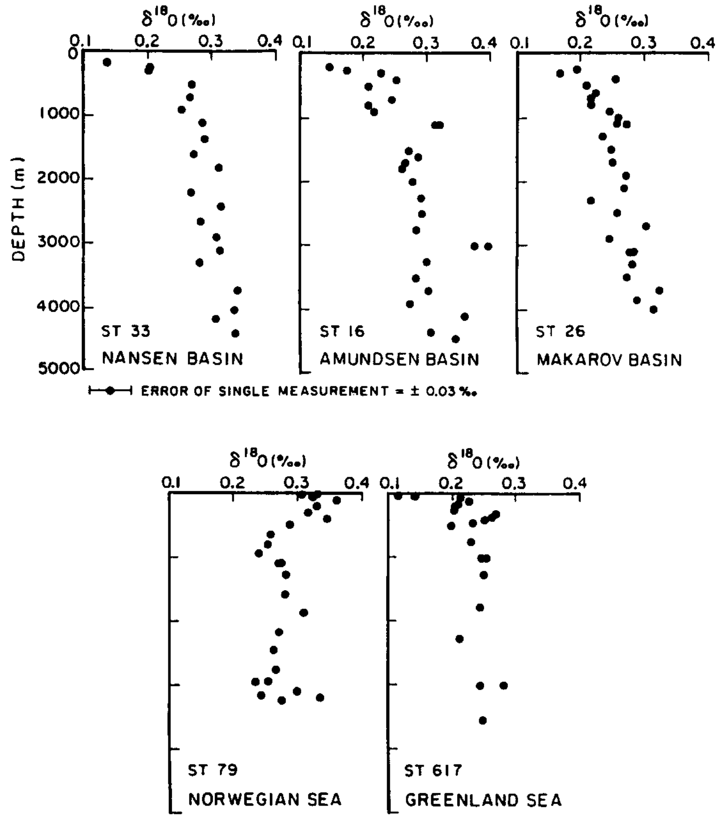

Fig.3. $\delta^{18}O$ versus depth plot of stations located in the central Arctic Ocean and the Norwegian and Greenland Seas. For geographical position of the stations see Fig.1.

5. DISCUSSION

5.1 Separation of river-runoff and sea-ice melt water in the Arctic Ocean surface waters

A combined salinity/^{18}O balance can be used to distinguish between river-runoff and sea-ice melt water in the freshwater component of the halocline (ÖSTLUND and HUT, 1984; SCHLOSSER *et al*, 1994). Here, we apply this concept to the sections obtained during the *Polarstern* 1987 cruise and ARCTIC 91 Expedition. Equations and choice of the parameters are discussed in SCHLOSSER *et al* (1994). For Atlantic water, the salinity is set to 34.92 and $\delta^{18}O$ to 0.3‰. The $\delta^{18}O$ value of river-runoff is set to -21‰ following the average value adopted by ÖSTLUND and HUT (1984). The few available $\delta^{18}O$ values from rivers flowing into the Arctic Ocean are close to the value of -21‰ suggested by ÖSTLUND and HUT (1984) from precipitation values as the best average for Arctic river-runoff (Table 1). In this average value the measurements published by BREZGUNOV, DEBOL'SKII, NECHAEV, FERRONSKII and YAKIMOVA (1983) have been omitted, because there is no available information on their data quality. However, we will discuss possible systematic errors introduced by this omission below. The salinity of sea-ice is set to 3. We assume that the sea-ice melt water has the same $\delta^{18}O$ value as the surface water of the station for which the freshwater balance is estimated multiplied by a fractionation factor (about 1.0021, MELLING and MOORE, 1995). This approach is only a first-order approximation of the true $\delta^{18}O$ value of sea-ice because sea-ice can move independently of the surface water, and so might have formed from water with a different $\delta^{18}O$ value from that observed at the sampling site. However, without more detailed observations, this procedure seems to be reasonable, especially in view of the relatively low $\delta^{18}O$ gradients in central Arctic Ocean surface waters (compared to the $\delta^{18}O$ value of -21‰ of river-runoff).

TABLE 1. $\delta^{18}O$ values of Arctic river water

River	Location	$\delta^{18}O$(‰)	Reference
Lena	delta	-20.6	this study
Mackenzie	DELTA	-20.1	MACDONALD and CARMACK (1991)
Mackenzie	upstream	-20.3	KROUSE and MACKAY (1971)
Yukon	upstream	-22.1	COOPER and DENIRO (1990)
Ob	bay	-16.3*	BREZGUNOV *et al* (1983)
Yenisey	bay	-17.5*	BREZGUNOV *et al* (1983)

*from direct measurements and $\delta^{18}O$/salinity extrapolation to 0 salinity; no information on data quality and measurement procedure is available.

In our calculation of the river-runoff and sea-ice melt water fractions, we refer to Atlantic Water as the water of the Atlantic-derived core observed in the interior of the Arctic Ocean. The Atlantic core is modified along its flow path in the Arctic Ocean (COACHMAN and BARNES, 1963), as is apparent in the systematic changes in the temperature/salinity and $\delta^{18}O$/salinity correlations within the Atlantic core (Figs 4a,b). The highest $\delta^{18}O$, salinity and temperature values are found in the southern part of the Eurasian Basin near Spitsbergen, decreasing towards the east and north. Our data support the flow pattern of Atlantic Water suggested by COACHMAN and BARNES (1963), GORDIENKO and LAKTIONOV (1958) and RUDELS, JONES and ANDERSON (1994): one branch of Atlantic Water flows eastward along the Siberian shelfbreak while a modified branch of Atlantic-derived water flows westward towards Fram Strait in the northern part of the sections. The salinity of 34.92 and the $\delta^{18}O$ value of 0.3‰ chosen for our mass balance calculation represent the least modified Atlantic-derived water inside the Arctic Ocean.

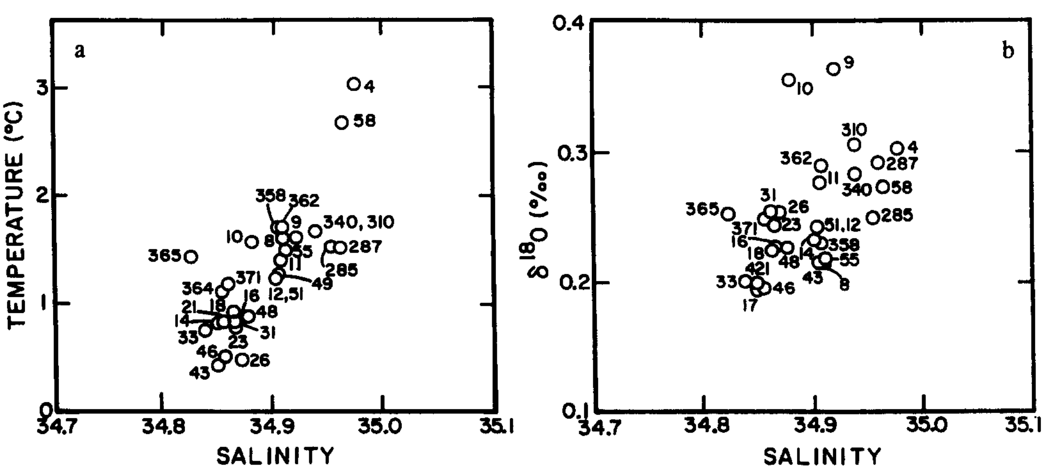

Fig.4. (a) potential temperature versus salinity plot for the Atlantic core (identified by the temperature maximum); station numbers are indicated. (b) same as Fig.4a for $\delta^{18}O$ versus salinity plot.

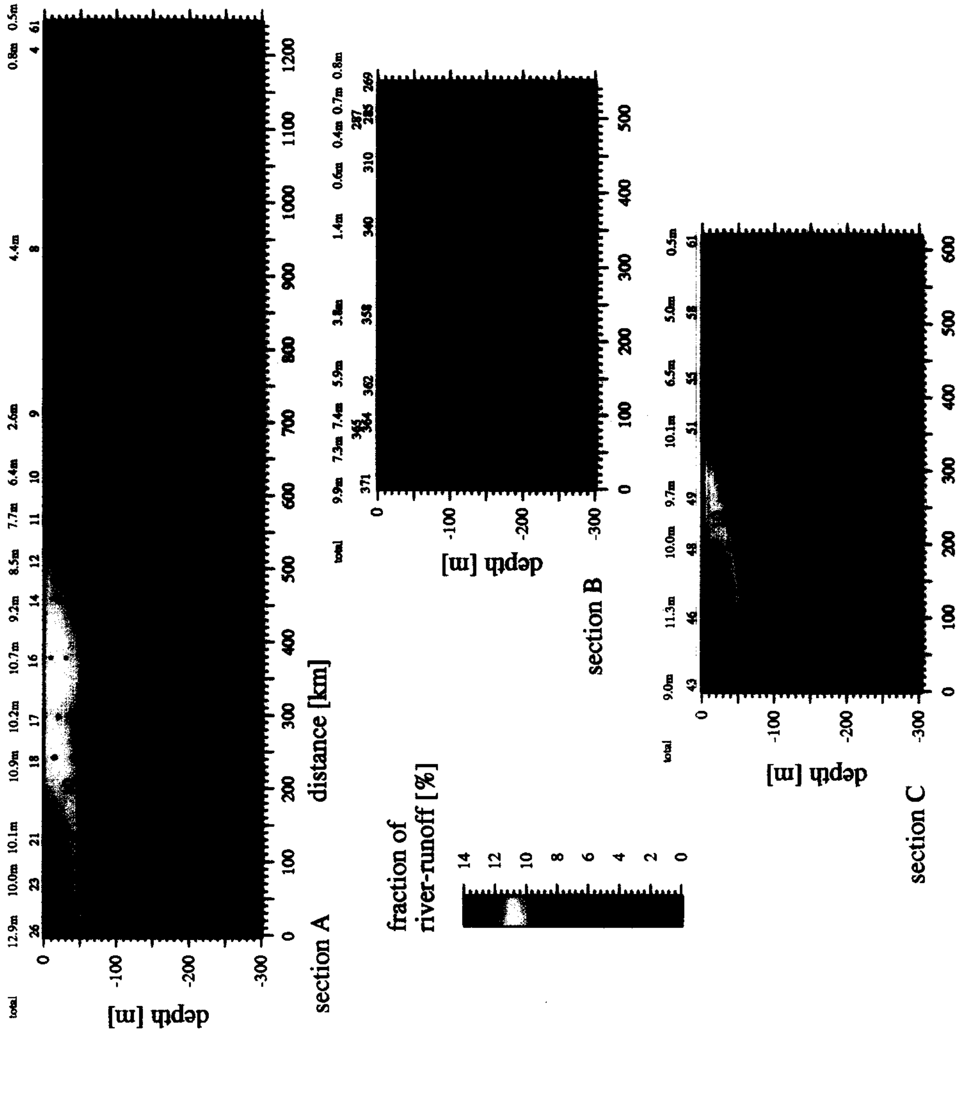

FIG.5. (a) Same as Fig.2 for the fraction of river-runoff contained in the halocline water. The calculated water column inventories of river-runoff are indicated above the station numbers.

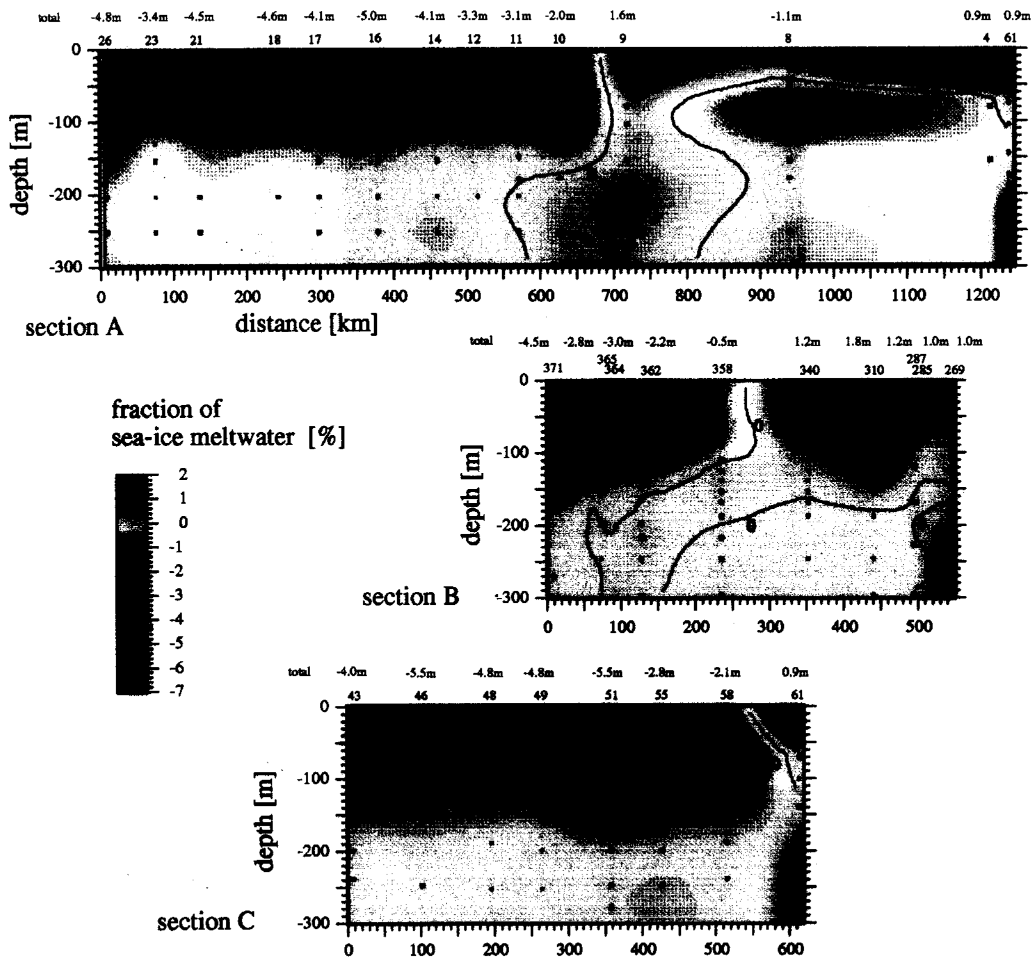

FIG..5. (b) same as Fig.2 for the fraction of sea-ice meltwater contained in the halocline waters. Positive numbers mean addition of freshwater by melting of sea-ice; negative numbers indicate formation of sea ice (removal of freshwater). The calculated water column inventories of sea-ice melt water are indicated above the station numbers.

Based on the results of the salinity/^{18}O mass balance calculations there is a trend of increasing river-runoff contribution in the freshwater component with increasing distance from the shelf. On the Barents Shelf the freshwater component is characterized by sea-ice melt water (Figs 5a,b). In the northern part of sections A and B and in the western part of section C, not only does the freshwater component consist almost completely of river-runoff, but also part of it has been extracted to form sea-ice. The fraction of river-runoff is small in the southern part of the sections; north of the sea-ice melt water boundary (i.e. the 0% sea-ice melt water fraction isopleth), the river-runoff fractions in the surface water increase from about 2% to about 12%. In contrast to the sea-ice melt water fractions, the river-runoff fractions do not level off, but increase monotonically towards the northern end of the sections.

Section C north of Fram Strait shows sea-ice melting in the southeastern part and sea-ice formation in the northwestern part. The fractions of sea-ice melt water and river-runoff reach more extreme values than in sections A and B; they amount to about -5% for sea-ice melt water and more than 12% for river-runoff. In the northwestern part of section C, we observe halocline waters with characteristics which are significantly different from those found at the northern ends of sections A and B. These waters probably originate in the Canadian Basin. This hypothesis is supported by the presence of elevated silicate values at station 43 (Fig.6a; ANDERSON, BJÖRK, HOLBY, JONES, KATTNER, KOLTERMANN, LILJEBLAD, LINDEGREN, RUDELS and SWIFT, 1994). In the $\delta^{18}O$ versus salinity correlation of station 43, this influence is also seen in a relatively high $\delta^{18}O$ value compared to adjacent stations with a salinity of about 32.5 (Fig.6b).

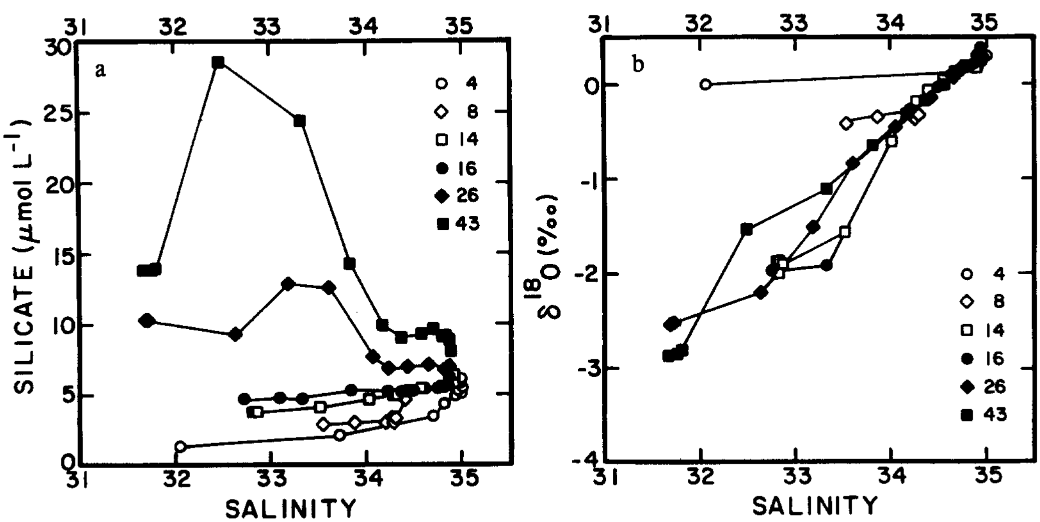

FIG.6. (a) Silicate versus salinity plot for selected stations. (b) Same as Fig.6a for $\delta^{18}O$ and salinity.

The presence of water of Pacific origin will distort the results of the mass balance calculation at station 43. The stations in the northern part of section A (stations 23, 26 and 31) also show slightly elevated silicate values at salinities of about 33 (ANDERSON *et al*, 1994) which seems to indicate a small contribution of UHW to these waters. Based on these observations, the extent of the Siberian part of the Transpolar Drift can be located at the northern end of section A and at station 43 in section C. The halocline water of the Canadian Basin flows towards Fram Strait in a separate branch and we have to assume that there are well separated regimes within the Arctic Ocean halocline (ANDERSON *et al*, 1994). The boundary between the halocline waters of the Eurasian and Canadian Basins, which roughly follows the Lomonosov Ridge, is also reflected in the surface $\delta^{18}O$ values (Fig.7). The surface waters of the Canadian Basin have lower $\delta^{18}O$ values compared to the Eurasian Basin. Interpretation of this observation requires quantification of the Pacific component contained in the halocline waters. We will attempt such a quantification below.

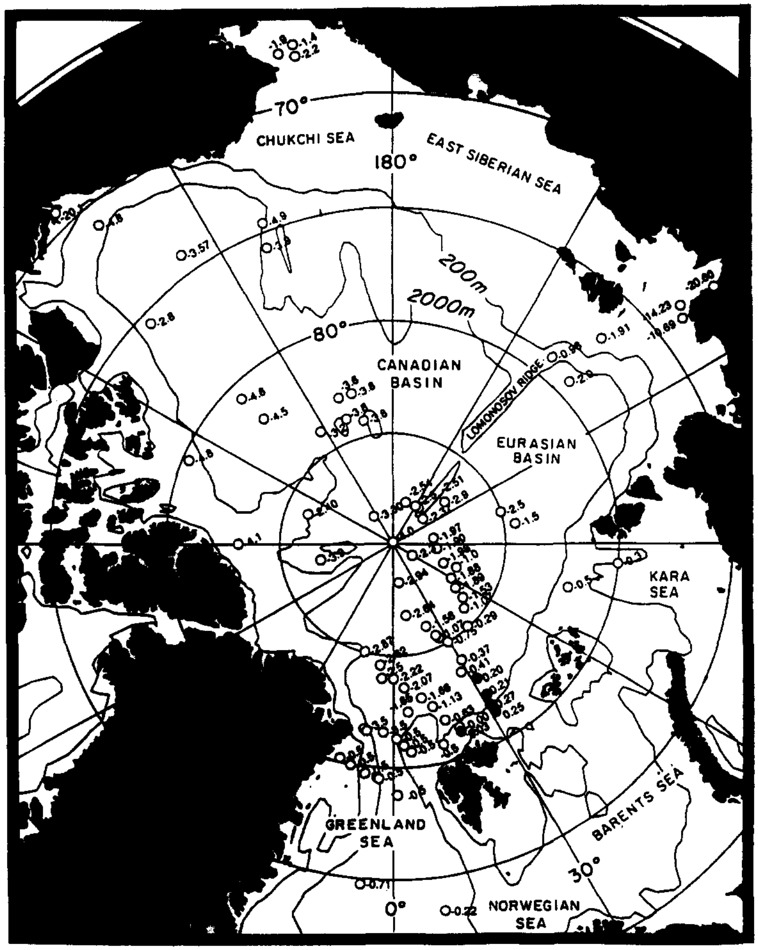

FIG.7. Geographic distribution of surface $\delta^{18}O$ values in the Arctic Ocean using our own $\delta^{18}O$ data as well as $\delta^{18}O$ data from ÖSTLUND and HUT (1984), ÖSTLUND *et al* (1987), MACDONALD and CARMACK (1991), COOPER and DENIRO (1990), DONK and MATHIEU (1969), VETSHTEYN, MALYUK and RUSSANOV (1974), FRIEDMAN, SCHOEN and HARRIS (1961) and REDFIELD and FRIEDMAN (1969). Full dots mark positive $\delta^{18}O$ values, open circles mark negative $\delta^{18}O$ values.

5.2 Water column inventories of sea-ice melt water and river-runoff

The fractions of sea-ice melt water and river-runoff can be integrated from the surface down to the Atlantic core, providing inventory values for the total amount of river-runoff and sea-ice melt water stored in the water column (values are indicated in Fig.5 above the station numbers). The depth range of the integration extends from the surface to the temperature maximum (typical depth: 250 to 300m) which is used as an indicator of the core of the Atlantic-derived water. In cases where the temperature maximum was difficult to define, salinity measurements were used as additional parameter. The uncertainty of the integration and the fraction calculation add up to an error of about ±0.5m for the total inventory of sea-ice melt water and river-runoff stored in the water column (determined by variation of the parameters used in the fraction calculation and by variation of the integration depth).

For a correct interpretation of the river-runoff and sea-ice melt water inventories it has to be kept in mind that these values represent the history of the water column. From the $\delta^{18}O$ signal we learn approximately how much river-runoff has been added to the weater column. However, this value is not the actual amount of fresh water stored in the water column. Sea-ice might melt or be formed modifying the freshwater inventory without significantly changing the ^{18}O balance. The amount of freshwater actually contained in the water column is the sum of the sea-ice melt water (H_{ice}) and river-runoff (H_{riv}) inventories ($H_{riv} + H_{ice}$).

The distribution of the inventory of sea-ice melt water shows a sharp transition between melting and formation of sea-ice over the Nansen Basin (Fig.8b). This front coincides with a strong gradient in the inventory of river-runoff (Fig.8a) indicating the common source of river-runoff and sea-ice on the shelves. Most of the initial sea-ice is formed on the shelf and it grows more slowly during its transit across the central Arctic Basins (WEEKS and ACKLEY, 1986).

Little is known about advection and mixing within the halocline. From the distribution of sea-ice melt water and river-runoff fractions we can conclude that there are well-defined lateral regimes. Although large amounts of river-runoff are flowing into the Kara and Laptev seas, the runoff does not flow directly north into the Arctic Ocean, as indicated by the relatively high surface $\delta^{18}O$ values just north of the Kara Sea (ÖSTLUND and HUT, 1984) (Fig.7). Near the shelf break of the Laptev Sea, the surface $\delta^{18}O$ values are lower (about -2‰, Fig.7), but still higher than the minimum values found in the interior of the Arctic Ocean. The river-runoff might flow off the shelf in plumes in other areas or it might be transported eastward in the Siberian Coastal Current to flow off the shelf as far east as the East Siberian or the Chukchi Seas.

5.3 Quantification of the Pacific component in the Arctic Ocean halocline

In the following approach, we use silicate as a tracer in a four-component mass balance to quantify the contribution of Bering Strait Inflow (BSI) to halocline water:

$$f_a + f_b + f_r + f_i = 1 \qquad (1)$$

$$f_a S_a + f_b S_b + f_r S_r + f_i S_i = S_m \qquad (2)$$

$$f_a O_a + f_b O_b + f_r O_r + f_i O_i = O_m \qquad (3)$$

$$f_a Si_a + f_b Si_b + f_r Si_r + f_i Si_i = Si_m \qquad (4)$$

where f_a, f_b, f_r and f_i are, respectively the fractions of Atlantic water, Bering Strait Inflow water,

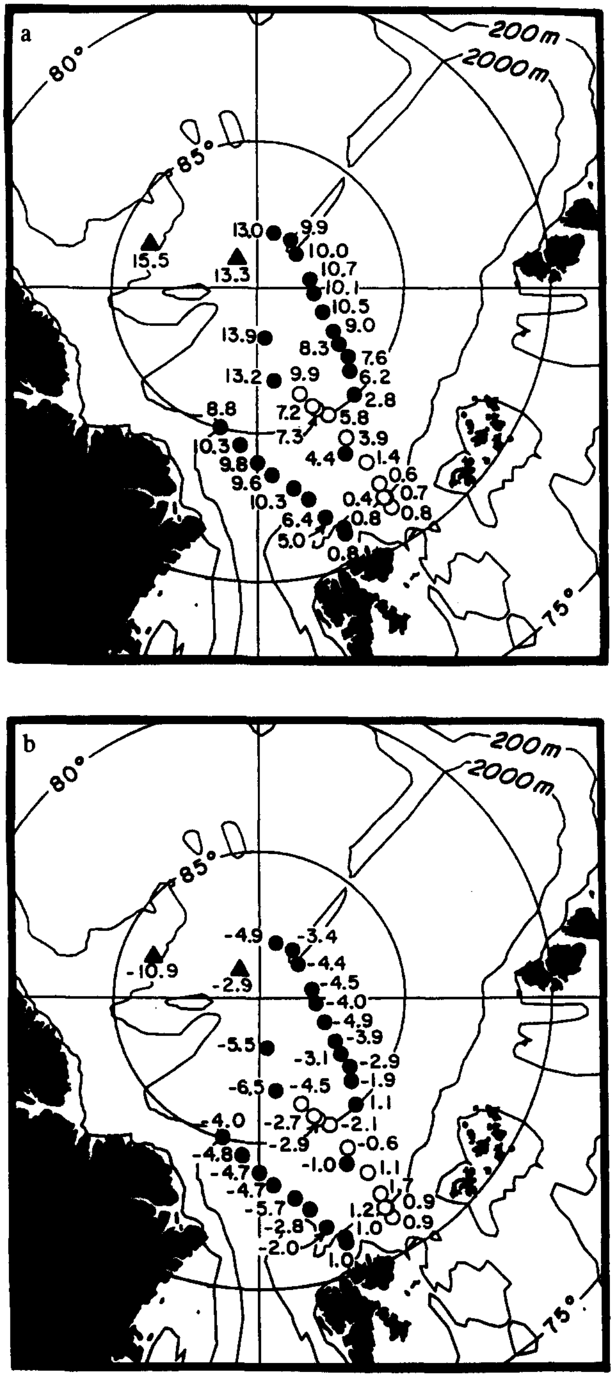

FIG. 8. (a) Geographic distribution of the water column inventory of sea-ice melt water H_{ice} contained in the upper waters of the Arctic Ocean. The values of station 43, LOREX and CESAR are obtained using the 4-component mass balance calculation, all other values using the 3-comnponent mass balance calculation. (b) Same as Fig. 8a for river-runoff H_{riv}.

river-runoff and sea-ice melt water in a halocline water parcel, and S_a, S_b, S_r, S_i, O_a, O_b, O_r, O_i and Si_a, Si_b, Si_r, Si_i are the corresponding salinities, $\delta^{18}O$ values and silicate concentrations. S_m, O_m and Si_m are the salinity, $\delta^{18}O$ value and silicate concentration measured in a specific water parcel.

In the following, the limitations of our approach to quantify the Pacific component contributing to the freshwater balance of the Arctic Ocean are discussed. Additionally, the choice of the silicate values of the end members is described (all values used in the four-component mass balance are listed in Table 2).

TABLE 2. Parameters used in the 4-component mass balance

	salinity	$\delta^{18}O$ (‰)	silicate ($\mu mol\ l^{-1}$)
Sea-ice	3	surface*	1
River-runoff	0	-21	10
Atlantic water	34.92	0.3	6
BSI	33	-1.0	40

*use $\delta^{18}O$ value of surface sample plus 2.1‰ for fractionation during sea-ice formation (MELLING and MOORE, 1994)

The water passing through Bering Strait has three components: Anadyr, Bering Shelf, and Alaskan Coastal water (COACHMAN, AAGAARD and TRIPP, 1975; WALSH, MCROY, COACHMAN, GOERING, NIHOUL, WHITLEDGE, BLACKBURN, PARKER, WIRICK, SHUERT, GREBMEIER, SPRINGER, TRIPP, HANSELL, DJENIDID, DELEERSNIJDER, HENRIKSEN, LUND, ANDERSON, MÜLLER-KARGER and DEAN, 1989). Based on salinity and nutrient distributions over the Chukchi Sea shelf (WALSH *et al*, 1989) it can be concluded that Anadyr and Bering Shelf waters are the sources of nutrient-rich water flowing off the shelf. This water mass is called UHW within the Arctic Ocean. Its salinity is about 33.1, similar to its source water masses Anadyr water with a salinity of about 33 and Bering Shelf water with a salinity ofbout 32.5 (COACHMAN *et al*, 1975). The $\delta^{18}O$ value of this water mass can be estimated to be about -1.0‰ on the basis of $\delta^{18}O$ and salinity measurements from Bering Strait (MACDONALD, CARMACK, MCLAUGHLIN, ISEKI, MACDONALD and O'BRIEN, 1989; and data cited by BJÖRK, 1990), in the northern Pacific (GEOSECS data) and from bottom salinity and ^{18}O measurements in the western Bering and Chukchi Seas (COOPER and DENIRO, 1990). This value is only accurate within about ±0.5‰.

Alaskan Coastal Water (ACW), the source water of Bering Summer Water (BSW), has a mean salinity of 31.5 (COACHMAN *et al*, 1975). Having a relatively low density, it stays near the surface and so does not come in contact with the sediments and take up silicate by dissolution. It has no silicate signature and is, therefore, not accounted for in the mass balance calculation. From current meter measurements it can be estimated that ACW contributes less than 1/3 to the water flowing through Bering Strait, since the flow through Anadyr Strait is about twice as strong as the flow through Shpanberg Strait (WALSH *et al*, 1989). MACDONALD *et al*, (1989) estimate that BSW makes up about 1/8 of the BSI found on the Beaufort shelf. Omission of about 1/3 of BSI in our mass balance calculation would lead to an error in the estimated river runoff and sea-ice melt water inventories of about 2 to 4m and 0.3m, respectively (estimated as half the difference between results of 3- versus 4-component mass balance; see discussion below). The actual error of the total river-runoff fraction might be smaller, because part of the low salinity and $\delta^{18}O$ signal of BSW is caused by freshwater of the Yukon River, which joins the Alaskan Coastal Current on the Bering Sea shelf.

For determination of the silicate concentration of the components used in our mass balance, we have to take into consideration that the silicate concentration can be changed rapidly in ice-free shallow shelf seas both by biological consumption during summer, and by dissolution from the sediments. Only off the shelves and beneath a permanent sea-ice cover these factors are limited so that silicate concentration can be treated as a more or less conservative tracer. Therefore, we specify the concentrations of each component as it flows off the shelf into the Arctic Ocean interior. We selected the following silicate values for the individual water masses contributing to the Arctic halocline.

Atlantic water: Si_a = 6µmole l^{-1} which is close to the value measured at the temperature maximum in the Eurasian Basin (KOLTERMANN and LÜTHJE, 1989).

Sea-ice: Si_i = 1µmole l^{-1} following MACDONALD *et al* (1989). The choice of this value is not critical; if we choose, for eample, 5µmole l^{-1} instead of 1µmole l^{-1}, the results change by less than 1% (absolute) in all fractions.

River-runoff: Si_r = 10µmole l^{-1} is adopted close to the average value observed in the waters of the Eurasian Basin containing river-runoff. Concentrations of 46µmole l^{-1} have been measured in the Mackenzie River (MACDONALD *et al*, 1989) and estimates for the Siberian rivers are as high as 125µmole l^{-1} (CODISPOTI and OWENS, 1975). However, no correlation is observed in our data set from the Eurasian Basin between river-runoff fraction and elevated silicate concentration (compare Figs 5a and 6b). We assume in our calculations that the high silicate concentrations carried by the rivers have been reduced to about 10µmole l^{-1}, i.e. to the level observed in the waters with the highest fraction of river-runoff.

Bering Strait Inflow: Si_b = 40µmole l^{-1} is estimated for the average silicate concentration of BSI from a section by TRESHNIKOV (1985), which shows a tongue of silicate-rich water reaching from the Chukchi Sea into the Arctic Ocean interior (Fig.9).

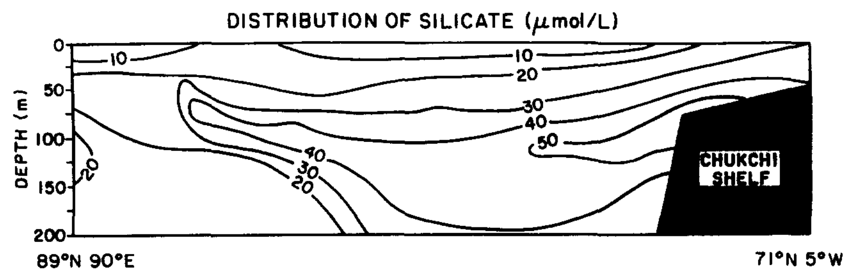

FIG. 9. Silicate concentration along a section reaching from the Chukchi Sea into the Arctic Ocean interior (from approximately 71°N 5°W to 89°N 90°E). Adapted from TRESHNIKOV (1985).

Both river-runoff and Bering Strait Inflow cross the Arctic shelves and enter the interior basins. En route silicate levels are decreased by the large seasonal growth rates of diatoms in surface waters (SAMBROTTO, GOERING and MCROY, 1984; WALSH *et al*, 1989). To obtain silicate concentrations in river-runoff as low as 10µmole l^{-1}, practically the entire amount of riverine silicate has to be deposited on the shelf as a result of biological consumption. Considering the rapid biological processes found in the Arctic Ocean during the production season (SAMBROTTO *et al*, 1984; WALSH *et al*, 1989), and a mean residence time of river-runoff on the Arctic shelves of approximately 1 to 5 years (MACDONALD *et al*, 1987, 1989; SCHLOSSER *et al*, 1994), this seems to be a reasonable assumption.

However, dissolution of sedimented silicate may be occurring over the shelves which would increase the silicate levels above 10µmole l^{-1}. For BSI much of the reduction by biological consumption may be compensated by dissolution of silicate in shelf bottom waters (WALSH *et al*, 1989).

Since at none of the stations of our data set is there a fully developed silicate maximum such as observed at locations in the central Canadian Basin, we use data of the LOREX and CESAR ice camp sites to test our balance equations ($\delta^{18}O$ and salinity data from ÖSTLUND, POSSNERT and SWIFT, 1987); silicate concentrations for LOREX from MOORE *et al* (1983), their Fig.6, and for CESAR from JONES and ANDERSON (1986), their Fig.3). At the CESAR Ice Camp station, over the Alpha Ridge in the Canadian Basin, UHW is fully developed with a silicate maximum of about 40µmole l^{-1} at 120m depth and a salinity of about 33.2. At the LOREX Ice Camp station, over the Lomonosov Ridge in the Canadian Basin, the silicate maximum of about 39µmole l^{-1} is found at 110m depth and a salinity of about 33.0. Station 43 from the ARCTIC 91 expedition has a silicate maximum of about 28.6µmole l^{-1} at 50m depth and a salinity of 32.5. Station 26 shows only a slight elevation in its silicate concentration of 13µmole l^{-1} at 70m depth and a salinity of 33.2.

At station 26, the fraction of BSI at the depth of the silicate maximum is about 15% and the fraction of Atlantic water is reduced to about 80% (Fig.10a). Below the depth of the silicate maximum we observe practically pure Atlantic water with the fraction of BSI being close to zero. At station 43 the ratio of BSI to Atlantic water at the depth of the silicate maximum is about 70:30 (Fig.10b). At the LOREX Ice Camp this ratio at the depth of the silicate maximum is about 95:5 (Fig.10c) and at the CESAR Ice Camp it becomes about 100:0 (Fig.10c). The silicate concentrations in the Atlantic layer of station 43, LOREX and CESAR are higher than at station 26 and the other stations in the Eurasian Basin of the Arctic Ocean. As a result, the fraction of BSI water at the depth of the Atlantic layer differs significantly from zero. In the Canadian Basin the Atlantic layer might have acquired a higher silicate concentration as a result of diffusion from and mixing with the overlying UHW.

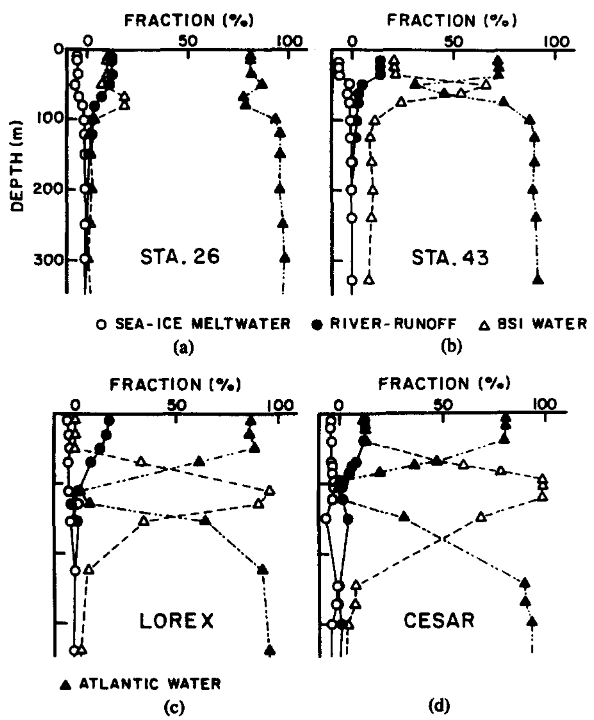

FIG. 10. Fractions of sea-ice melt water, river-runoff, Atlantic water and BSI water for stations 26 (a), 43 (b) and from LOREX (c) and CESAR (d). Calculations were done using the 4-component mass balance and the parameters listed in Table 2.

The influence of BSI water on the total river-runoff and sea-ice melt water inventories can be illustrated by comparing the inventory values calculated from 3- and 4-component mass balances (Table 3).

TABLE 3. Comparison of river-runoff and sea-ice melt water inventories using the 3- and 4-component mass balances

	Station 26		Station 43		LOREX		CESAR	
	H_{ice} (m)	H_{riv}	H_{ice} (m)	H_{riv}	H_{ice} (m)	H_{riv}	H_{ice} (m)	H_{riv}
3-comp. calc.	-4.8	12.9	-4.3	12.0	3.5	18.6	-11.7	23.7
4-comp. calc.	-4.9	12.1	-4.0	9.0	-3.0	14.3	-11.3	15.9
difference	0.1	-0.8	0.3	-3.0	0.5	-4.3	0.4	-7.8

H_{ice}, H_{riv}: inventory of sea-ice melt water and river-runoff, respectively

At station 26 the difference in the total amount of river-runoff estimated on the basis of the 4-component and the 3-component mass balances remains close to the estimated uncertainty of the 3-component mass balance. The results for the total amount of sea-ice melt water remain practically unchanged. At station 43 and at the LOREX Ice Camp, the differences in the river-runoff inventories are significant. With the result using the 4-component mass balance for station 43, the river-runoff inventories are more or less constant for the western part of Section C. The largest differences in the river-runoff inventories are observed at the CESAR Ice Camp. Additionally, there is a noticeable difference in the results for the sea-ice melt water inventory. Since the component of BSI water omitted from the calculations is about 1/3 of the total inflow, it can be estimated that the error is in first order about half of the listed differences.

The silicate values of the end members and the $\delta^{18}O$ of BSI have a relatively large uncertainty compared to the other parameters used in the calculations. Table 4 lists changes in the river-runoff and sea-ice inventories as a function of these parameters. The range of variation of each parameter reflects its estimated uncertainty.

TABLE 4. Estimate of the uncertainty in determining the sea-ice melt water and river-runoff inventories using the 4-component mass balance

	H_{ice} (m) at Station				H_{riv} (m) at Station			
	26	43	LOREX	CESAR	26	43	LOREX	CESAR
Result	-4.9	-4.0	-3.0	-11.9	12.1	9.0	14.3	15.9
ΔSi_r: ±5µmole l^{-1}	±0.0	±0.0	±0.0	±0.1	±0.2	±0.1	±0.1	±0.2
ΔSi_b: ±5µmole l^{-1}	±0.0	±0.1	±0.1	±0.2	±0.3	±0.5	±0.5	±1.5
$\Delta\delta O_b$: ±0.5‰	±0.4	±1.4	±2.4	±3.4	±0.5	±1.2	±2.2	±3.2

H_{ice}, H_{riv}: inventory of sea-ice melt water and river-runoff, respectively

It is evident that the greatest uncertainty results from the $\delta^{18}O$ value of BSI water, which needs to be determined more accurately by direct measurements from the Chukchi Sea.

5.4 Sea-ice melt water inventories: implications for sea-ice export through Fram Strait

Integration of the observed freshwater signal related to sea-ice formation along a section across Fram Strait can in principle be linked to export of sea-ice through this passage between the Arctic Ocean and the Greenland-Norwegian-Iceland Seas. Calculation of a true sea-ice export rate through Fram Strait requires:

1. Knowledge of the distribution of the river-runoff and sea-ice melt water fractions in the upper waters flowing southward through Fram Strait.
2. Quantification of the velocity profile at each station for which river-runoff and sea-ice melt water fractions are available.
3. The assumption that the sea-ice and the surface waters which carry the $\delta^{18}O$ and salinity signals imprinted during sea-ice formation are following the same pathways from the shelf seas to Fram Strait.

With this information, the export of both river-runoff and sea-ice melt water through Fram Strait could be calculated by integrating the product of the velocity and river-runoff or sea-ice melt water fraction from the surface to the core of the Atlantic layer, and along the southward flowing portion of the section.

Unfortunately, no detailed measurements of the current profile are available for the region of Fram Strait. Additionally, we do not know how closely the circulation of the upper waters carrying the sea-ice melt water signal is coupled to that of sea-ice. Therefore, we have to limit our evaluation of the sea-ice melt water inventories to fairly rough first-order approximations. For these approximations we assume the following two scenarios:

1. We use the velocities of the upper waters in Fram Strait determined by JONSSON, FOLDVIK and AAGAARD (1992) on the basis of current meter measurements. Assuming that their velocities for the upper 78 to 262m are representative for the portion of section C which contains a negative sea-ice melt water signal (i.e. assuming that the transition from positive to negative sea-ice melt water fractions marks the transition from northward flowing Atlantic derived waters to southward flowing Arctic surface waters), we can multiply the integral of the negative sea-ice melt water fractions along the section (2.45 $10^6 m^2$) by the mean velocity of 7.5cm s^{-1} (JONSSON *et al*, 1992). This yields an export rate of the freshwater deficit caused by ice formation of about 0.18Sv (6,300$km^3 y^{-1}$); this value has been corrected for the difference in densities of sea water (about 1.0) and sea-ice (about 0.9). The related flux of river-runoff through the section is about 2.47 $10^6 m^2$ at 7.5cm s^{-1}, or 0.19Sv. These fluxes represent upper limits because our section is well north of the latitude at which the current meters were deployed and the current velocities can be expected to increase from the central Arctic Ocean towards Fram Strait where a considerable fraction of the surface waters feed into the comparably narrow East Greenland Current. This hypothesis is confirmed by the fact that our calculations yield a river-runoff flux of about 0.19Sv, a value close to the sum of the total river discharge into the Arctic Ocean (about 0.1Sv) and the P-E excess (about 0.03 to 0.08Sv, AAGAARD and CARMACK, 1989; GORSHKOV, 1983).
2. We estimate the inventory of river-runoff for the portion of the ARCTIC 91 section C (Fig.1) where southward flow is assumed (H_{ice} negative). We then compare this number (4.92 $10^6 m^2$) with the river discharge data to obtain a mean velocity of the river-runoff contained in the upper waters flowing southward through Fram Strait. The result (0.13Sv/ 4.92 $10^6 m^2$) yields an average velocity of about 2.6cm s^{-1} for the river-runoff component

leaving the Arctic Ocean through Fram Strait. Assuming a constant ratio between the river-runoff fraction and the sea-ice melt water fraction allows us to estimate the export rate of the freshwater deficit related to ice-formation to be about 0.07Sv (2,280km^{3}y^{-1}).
In this calculation, we neglect outflow of river-runoff through the Canadian Archipelago. This assumption makes the calculated velocity an upper limit (diversion of a part of the river-runoff outflow through the Archipelago would decrease the velocity of river-runoff through Fram Strait). However, it seems likely that most of the sea-ice exits through Fram Strait while part of the river-runoff flows through the Canadian Archipelago. If we assume that the ratio of the sea-ice melt water fraction to the river-runoff fraction in the upper waters of the Arctic Ocean is more or less constant (first order approximation), the sea-ice export rates calculated by this method should be fairly realistic.

The analytical uncertainty of the estimate for the sea-ice export through Fram Strait is smaller than ±0.02Sv or ±650km^{3}y^{-1} (determined on the basis of an error of ±0.5m for the determination of the total river-runoff and sea-ice melt water inventories). The systematic error introduced by the assumptions used in the calculations might be considerably higher.

Table 5 summarizes the sea-ice melt water and river-runoff inventories calculated for section C and for two sections published by ÖSTLUND and HUT (1984) from YMER-80. Since the YMER-80 results do not include the influence of BSI, the results for section C are listed for both the 3-component mass balance (omitting BSI) and the 4-component mass balance (including BSI).

TABLE 5. Sea-ice melt water and river-runoff inventories calculated for three sections across Fram Strait.

Section	Average Latitude (°N)	Stations	Length (km)	I_{ice} (10^6m^2)	I_{riv} (10^6m^2)	H_{ice} (m)	H_{riv} (m)	Sea-ice export* (Sv)	Sea-ice export** (Sv)	Sea-ice export*** (Sv)
including BSI influence (4-component calculation)										
ARCTIC 91	≈83	43...58	585	-2.45	4.92	-4.2	8.4	0.065	0.090	0.18
omitting BSI influence (3-component calculation)										
ARCTIC 91	≈83	43...58	585	-2.47	5.07	-4.2	8.7	0.063	0.088	0.19
Ymer-80	≈81	164...168	≈163	-0.78	1.77	-4.8	10.9	0.057	0.079	0.06
Ymer-80	≈81	159...152	≈350	-2.49	4.47	-7.1	12.8	0.072	0.100	0.19

I_{ice}, I_{riv}: Sea-ice melt water and river-runoff inventories in the upper water layer integrated over the length of the sections
H_{ice}, H_{riv}: average thickness of sea-ice melt water and river-runoff layers
*Sea-ice export based on a P-E value of 0.03Sv (AAGAARD and CARMACK, 1989)
**Sea-ice export based on a P-E value of 0.08Sv (GORSHKOV, 1983)
***Sea-ice export using a mean velocity of 7.5cm s^{-1} (JONSSON et al, 1992).

The effect of BSI on the sea-ice export rate is not significant. There is good agreement of the sea-ice export rates between the sections located at varying latitudes between about 83 to 80°N.

Previous estimates of sea-ice export rates through Fram Strait from ice drift observations and sea-ice thickness measurements range from 0.1 to 0.18Sv for long-term averages (0.16Sv, VINJE and FINNEKASA, 1986; 0.10Sv, HIBLER, 1979; 0.18Sv, KOERNER, 1983; 0.13Sv, WADHAMS, 1983). Our estimates of the sea-ice export rate through Fram Strait have an upper limit of about 0.18Sv and a lower limit of about 0.07Sv.

5.5 Fresh water budget of Arctic Ocean halocline

Using the inventories of river-runoff and sea-ice melt water, the freshwater storage of the Arctic Ocean halocline can be calculated. With the knowledge of the annual river discharge (and the annual precipitation) we can then determine the mean residence time of the freshwater component in the Arctic Ocean. For the Canadian Basin, the thickness of the river-runoff and sea-ice melt water layer was chosen on the basis of the results of ARCTIC 91 stations 43 and 26 and the LOREX and CESAR Ice Camps including the influence of water entering through Bering Strait (BSI). For the Eurasian Basin, river-runoff and sea-ice melt water layer thicknesses were calculated using the results of the 3-component mass balance and an average value weighted by areas (Table 6).

TABLE 6. Estimates of areas, fresh water layer thickness and fresh water volume in the Arctic Ocean interior.

	Area ($10^6 km^2$)	H_{riv} (m)	H_{ice} (m)	V_{riv} ($10^3 km^3$)	V_{ice} ($10^3 km^3$)
Canadian Basin	3.2	13	-5	41.6	-16.0
Eurasian Basin	1.9	7.5	-3.5	14.3	-6.7
Arctic Ocean interior	5.1	11.0	-4.5	55.9	-22.7

H_{ice}, H_{riv}: average thickness of sea-ice melt water and river-runoff layers.
V_{ice}, V_{riv}: volume of sea-ice melt water and river-runoff

With the estimated average river-runoff inventory (11m) and the yearly river-runoff rate of 0.13Sv (equivalent to 0.8m y^{-1}, including a P-E value of 0.03Sv following AAGAARD and CARMACK, 1989), a freshwater mean residence time of 14 years is determined. Using a yearly river-runoff component of 0.18Sv including a precipitation value of 0.08Sv after GORSHKOV (1983), a mean residence time of 11 years is determined. The total amount of fresh water in the water column equals $H_{riv} + H_{ice}$, i.e. it is smaller than the amount of river-runoff initially added to the water column. Freshwater is partly removed as sea-ice and exported at a rate independent from the water column transport. Therefore, the derived mean residence time applies for the river-runoff remaining in the water column and this is equivalent to the freshwater actually found in the water column of the Arctic Ocean (ignoring the areas where sea-ice is melting). The mean residence time can also be adopted for the upper 50m to 100m of the halocline, which contain fresh water.

Mean ages derived from transient tracers point towards a faster renewal of the upper layers in the Eurasian Basin (e.g. WALLACE, MOORE and JONES, 1987; SCHLOSSER, BÖNISCH, KROMER, MÜNNICH and KOLTERMANN, 1990; SCHLOSSER, BÖNISCH, KROMER, LOOSLI, BÜHLER, BAYER, BONANI and KOLTERMANN, 1995). Based on $\delta^{18}O$ and tritium measurements ÖSTLUND and HUT (1984) derive a freshwater mean residence time of 10 years within the Arctic Ocean. These values are reasonably close considering that both of them are relatively rough estimates, but it should be pointed out that the estimate by ÖSTLUND and HUT (1984) involves qualitatively different results. ÖSTLUND and HUT (1984) do not detect a structure within the halocline and conclude that the horizontal mixing of the halocline waters occurs on a much shorter time scale than the mean residence time. In contrast, our results show a sharp boundary between the Eurasian and Canadian Basins and it can be assumed that the lateral exchange is slow compared to the mean residence time and that the mean residence times are different in each basin.

The fresh water mean residence time of 14 years is an average for the two basins. The main features of the circulation of the Arctic halocline are the Transpolar Drift in the Eurasian Basin and the Beaufort Gyre in the Canadian Basin. This circulation pattern suggests that the mean residence time of the river-runoff in the halocline of the Eurasian Basin is determined by the transfer time between the Siberian shelf areas and Fram Strait. In the Canadian Basin, on the other hand, the freshwater might be trapped in the Beaufort Gyre resulting in a longer mean residence time compared to that in the Eurasian Basin. This might be the reason that the average river water thickness in the Canadian Basin is significantly higher than in the Eurasian Basin. Since most of the river-runoff discharge of the Arctic Ocean enters the Siberian shelves, part of the Siberian river-runoff must flow eastwards and end up in the Beaufort Gyre.

5.6 Systematic errors

BREZGUNOV *et al* (1983) published $\delta^{18}O$ values for the Ob and Yenisey rivers, which are significantly different from the value we adopted for the Arctic river-runoff in the mass-balance calculation. Since there is no information on methods and measurement precision available we decided to omit these values. On the other hand, we feel that an isotopically heavier value for the Ob and Yenisey rivers could be correct for the following reasons: (1) the drainage areas of both rivers extends further south than any of the other Arctic rivers (to about 45°N compared to about 60°N), and (2) the main source of precipitation for northern Eurasia lies towards the west of the continent, so the precipitation will be less depleted in the western regions, i.e. in the drainage areas of the Ob and Yenisey rivers.

Here we estimate the systematic error introduced by using a uniform $\delta^{18}O$ value of -21‰ and -20‰ for river-runoff. Adopting a $\delta^{18}O$ value of about -17‰ for the combined runoff of Ob and Yenisey which contributes about one third of the total amount of Arctic river-runoff (about $1133 km^3 y^{-1}$ out of a total of $3303 km^3 y^{-1}$, AAGAARD and CARMACK, 1989; TRESHNIKOV, 1985), the overall value of the Arctic river-runoff is about -20‰. Using -20‰ versus -21‰ for the $\delta^{18}O$ value of river-runoff for most stations leads to a difference of about 0.5m in sea-ice melt water and in river-runoff inventory. The error for most of the ARCTIC 91 stations is about 13% for the sea-ice melt water and about 5% for the river-runoff inventories.

The sea-ice export rate through Fram Strait using -20‰ for the $\delta^{18}O$ value of river-runoff is estimated to 0.08Sv or 0.10Sv compared to 0.06Sv or 0.08Sv calculated on the basis of a $\delta^{18}O$ value of -21‰. The estimate of the mean residence time of freshwater within the Arctic Ocean stays within ±1y at about 11 to 14 years (Table 7).

TABLE 7. Comparison of results using -20‰ versus -21‰ as $\delta^{18}O$ value for the river-runoff in the mass balance calculations

	Result using -21‰	Analytical error	Result using -20‰	Difference
H_{ice}, H_{riv}	-	≈±0.5m	-	≈0.5m
sea-ice export	0.06/0.08Sv	≈±0.02Sv	0.08/0.10Sv	0.02Sv
mean residence time	11/14 yrs	-	11/14 yrs	<1yr

5.7 Sources of Arctic Ocean deep and bottom waters

The $\delta^{18}O$ profiles show only small variations below the Atlantic-derived layer. The stations in the central Arctic basins show a slight increase in $\delta^{18}O$ (≈0.03‰) below about 2600m depth (the sill depth of Fram Strait). Average values of $\delta^{18}O$ and salinity were calculated for the depth intervals between 1000 and 2600m and between 2600m and the bottom (Fig.11a; open symbols: 1000-2600m; full symbols: 2600m to bottom depth; the error bars represent the standard errors of the mean values; see also Table 8).

TABLE 8. Station averages over deep and bottom intervals

Station	Interval (m)	$\delta^{18}O$ (‰) mean	$\delta^{18}O$ (‰) error	Salinity mean	Salinity error	n
358	1000-2500	0.286	0.013	34.925	0.001	12
	2600-bottom	0.310	0.012	34.941	0.001	12
16	1000-2600	0.287	0.007	34.920	0.003	9
	2600-bottom	0.323	0.014	34.940	0.001	10
33	1000-2600	0.290	0.009	34.919	0.004	6
	2600-bottom	0.315	0.009	34.938	0.002	8
26	1000-2600	0.253	0.006	34.939	0.006	9
	2600-bottom	0.288	0.008	34.955	0.000	9
617	1000-bottom	0.241	0.009	34.920	0.003	8
79	1000-bottom	0.279	0.006	34.940	0.001	14

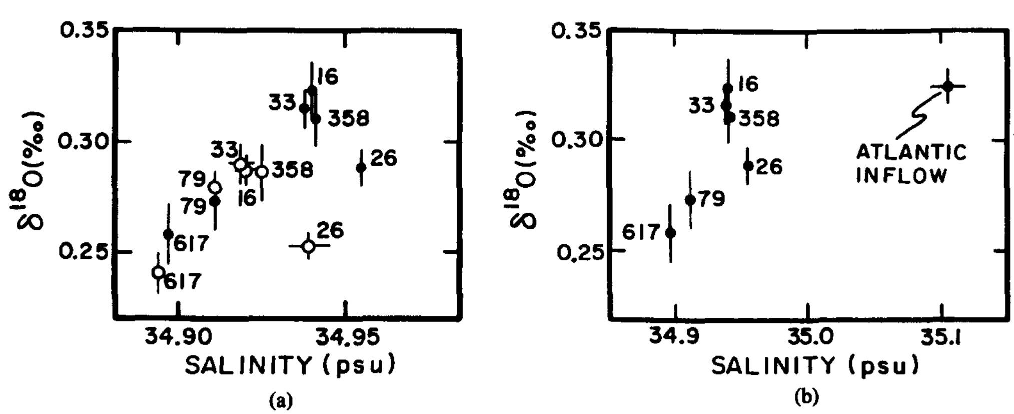

FIG.11. (a) $\delta^{18}O$ versus salinity plot for deep (1000-2600m depth; open circles) and bottom (2600m to bottom depth; full circles) waters. (b) $\delta^{18}O$ versus salinity plot for bottom waters and Atlantic Inflow.

Stations 79 and 617, located in the Norwegian Sea and in the Greenland Sea respectively, do not show any significant increase in salinity or $\delta^{18}O$ with depth (see also Tables 8 and 9). For all stations in the Arctic Ocean, there is a trend toward increasing $\delta^{18}O$ between the upper (1000 to 2600m; open symbols in Fig.11a) and the lower (2600m to bottom: full symbols in Fig.11a) depth intervals of about 0.03‰. The deep waters in the Norwegian Sea fall on a mixing line between the deep Eurasian Basin and Greenland Sea waters. However, the values of Station 26 in the Makarov Basin (Canadian Basin) do not fall on this mixing line, instead there is a shift to higher salinities and lower $\delta^{18}O$ values.

TABLE 9. Properties of deep and bottom water masses

	$\delta^{18}O$ (‰)		Salinity	
		error		error
EBDW	0.278	0.007	34.922	0.001
EBBW	0.305	0.008	34.940	0.001
GSDW	0.248	0.007	34.896	0.001
NSDW	0.276	0.007	34.911	0.002
Atlantic inflow	0.324	0.008	35.105	0.014

EBDW: average of 1000-2600m interval samples from stations 358, 16 and 33.
EBBW: average of 2600m-bottom interval samples from stations 358, 16 and 33.
GSDW: average of 1000m-bottom interval samples from station 617.
NSDW: average of 1000m-bottom interval samples from station 79.
Atlantic inflow: average of upper 500m from samples in the Norwegian Sea (8 values from station 79).

A salinity balance shows that Eurasian Basin Bottom Water (EBBW; water below 2600m depth) contains about 5‰ of freshwater, relative to the Atlantic inflow (surface to 500m depth at station 79 in the Norwegian Sea; Fig.11b). There is no decrease in $\delta^{18}O$ between inflowing Atlantic Water and EBBW, indicating that the freshwater which contributes to EBBW must be derived from melting of sea-ice. The $\delta^{18}O$ values of Atlantic inflow and EBBW are within about 2 sigma. If the contribution of river-runoff is calculated on the basis of this difference of 2 sigma, a river-runoff fraction of 1‰ is obtained. This possible small contribution of river-runoff to EBBW is still consistent with the hypothesis that the fresh water contribution to EBBW is derived from sea-ice melt water, since sea-ice formed from shelf waters containing river-runoff will indirectly carry a small river $\delta^{18}O$ signal (e.g. the mean $\delta^{18}O$ value of ice cores taken within the Barents Sea is -2.6‰ (PFIRMAN, personal communication).

The bottom waters of the Makarov Basin and the Greenland and Norwegian Seas are slightly depleted in $\delta^{18}O$ relative to Atlantic Water. The bottom water found at station 26 contains about 5‰ of freshwater (calculated from a salinity balance). About 2‰ of this freshwater might be derived from river-runoff (calculated from a $\delta^{18}O$ balance). This means that roughly one half of the freshwater is derived from river-runoff with a $\delta^{18}O$ value of -21‰.

Comparison of $\delta^{18}O$ values of the deep and bottom waters of the central Arctic Ocean with those observed in the surface shelf waters (Fig.7) indicate that the main source region of EBBW are in the Barents and Kara Seas since only on these shelf areas are the $\delta^{18}O$ values without considerable influence of river-runoff. The higher river-runoff fraction in the Canadian Basin Deep Water indicates that at least a small part of CBDW has its source on the shelves east of the Barents and Kara Seas. (For all these conclusions steady-state conditions for deep and bottom water formation are assumed.)

6. CONCLUSIONS

Balances of mass, salinity and $\delta^{18}O$ allow quantification of the fractions of the individual water masses contributing to the Arctic halocline. In the Eurasian Basin, the relative contributions of the different freshwater sources can be determined fairly precisely. In the Canadian Basin the mass balance calculation has to include BSI water in order to obtain a complete analysis of the halocline waters. Using silicate as an additional parameter in a four-component mass balance accounts for part of the Pacific component feeding into the upper halocline. The relatively low salinities and low $\delta^{18}O$ values in the Canadian Basin are caused by BSI water, as well as by larger river-runoff and sea-ice melt water fractions. For a more precise quantitative determination of the freshwater sources and content in the Canadian Basin, the parameters of the mass balance calculations have to be determined more accurately. This can be done in an oceanographic study which includes $\delta^{18}O$ and silicate measurements in the western Chukchi Sea where the UHW acquires its properties and where it flows off the shelf.

The sea-ice melt water inventories calculated for the Arctic Ocean can be converted into sea-ice export rates through Fram Strait once better current measurements are available for Fram Strait and the coupling between the circulation of sea-ice and surface waters is better known.

On the basis of a freshwater balance of the entire halocline of the Arctic Ocean, an average mean residence time of the river-runoff within the Arctic Ocean is calculated. The freshwater distribution within the Arctic Ocean reveals distinct regimes within the surface waters. These regimes can be used to outline pathways and transport mechanisms within the upper waters of the Arctic Ocean. A direct application of this knowledge is the prediction of the transport pathways of pollutants released to the Siberian shelves by riverine runoff. The $\delta^{18}O$ distribution suggests that the river-runoff from the Siberian shelves is not spreading directly into the Arctic Ocean interior, but either initially flows eastwards, or flows off the shelves in plumes.

Comparisons of the $\delta^{18}O$ values of EBBW, Atlantic Water and shelf waters allows us to locate the sources of EBBW in the Barents and Kara seas. The source region of CBDW supplies a slightly larger amount of river-runoff and cannot be located with the help of the present $\delta^{18}O$ data set.

For better understanding of oceanographic processes in the Arctic Ocean improved data coverage is needed. However, the existing observational data can already be used to validate the large-scale features of models simulating the circulation in the Arctic Ocean.

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