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Wind-driven transport pathways for Eurasian Arctic river discharge

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Abstract. Distributions of temperature, salinity, and barium in near-surface waters (depth ≤ 50 m) of the Laptev Sea and adjacent areas of the Arctic Ocean are presented for the summers of 1993, 1995, and 1996. The tracer data indicate that while fluvial discharge was largely confined to the shelf region of the Laptev Sea in the summer of 1993, surface waters containing a significant fluvial component extended beyond the shelf break and over the slope and basin areas north of the Laptev Sea in the summers of 1995 and 1996. These distributions of fluvial discharge are consistent with local winds and suggest two principal pathways by which river waters can enter the central Arctic basins from the Laptev Sea. When southerly to southeasterly wind conditions prevail, river waters are transported northward beyond the shelf break and over the slope and adjacent basin areas. These waters can then enter the interior Arctic Ocean via upper layer flow in the vicinity of the Lomonosov Ridge. Under other wind conditions, river waters are steered primarily along the inner Laptev shelf and into the East Siberian Sea as part of the predominantly eastward coastal current system. These waters then appear to cross the shelf and enter the interior Arctic Ocean via upper layer flow aligned roughly along the Mendeleyev Ridge. The extent to which either pathway is favored in a given year is largely determined by local wind patterns during the summer months, when fluvial discharge is greatest and shelf waters are at the lowest salinity of their annual cycle.

1. Introduction

It has long been appreciated that the considerable fluvial discharge to the Arctic Ocean strongly influences its properties and circulation. Fluvial sources contribute to maintaining the Arctic cold halocline layer [*Rudels et al.*, 1996; *Steele and Boyd*, 1998], which insulates the perennial sea ice cover from heat contained in the underlying Atlantic layer. The export of freshwater from the Arctic through Fram Strait and the Canadian Archipelago (of which fluvial discharge is a major component) affects water column stability in areas of deep convection in the North Atlantic and thereby influences global thermohaline circulation [*Aagaard and Carmack*, 1989]. Understanding the mechanisms by which river waters become incorporated into the circulation of the Arctic Ocean is therefore essential to investigating the links between the Arctic and global climate. Knowledge of the mixing pathways of fluvial discharge also provides information needed to assess the transport and fate of pollutants released to the Arctic Ocean [*Macdonald and Bewers*, 1996].

Exactly how and where river waters cross the broad shelves of the Arctic's marginal seas and enter the interior basins remains unclear. Based on physical and chemical tracer data, it appears that river waters flowing into the Eurasian Arctic seas are transported to the interior Arctic Ocean by surface flows roughly aligned along the Nansen-Gakkel, Lomonosov, and Mendeleyev ridges [*Östlund and Hut*, 1984; *Jones et al.*, 1991; *Anderson et al.*, 1994; *Bauch et al.*, 1995; *Guay and Falkner*, 1997; *Wheeler et al.*, 1997]. Circulation models for the Arctic Ocean also tend to show water from Eurasian Arctic rivers entering the interior basins along these ridges [*Proshutinsky and Johnson*, 1996; *Maslowski et al.*, 1998]. These results are consistent with the conceptual depiction of mean circulation in the Nansen, Amundsen, and Makarov basins as a group of basin-trapped cyclonic gyres – i.e., northward flow along the western flanks of the Nansen-Gakkel, Lomonosov and Mendeleyev ridges comprises the northward branches of the gyres occupying the Nansen, Amundsen and Makarov basins, respectively [*Rudels et al.*, 1994; *McLaughlin et al.*, 1996].

The Laptev Sea receives a large influx of fluvial discharge, both as direct runoff from rivers and as a component of river-influenced shelf waters entering from the Kara Sea through Vilkitsky Strait (Figure 1 and Table 1). Here we report temperature, salinity, and barium data collected from near-surface

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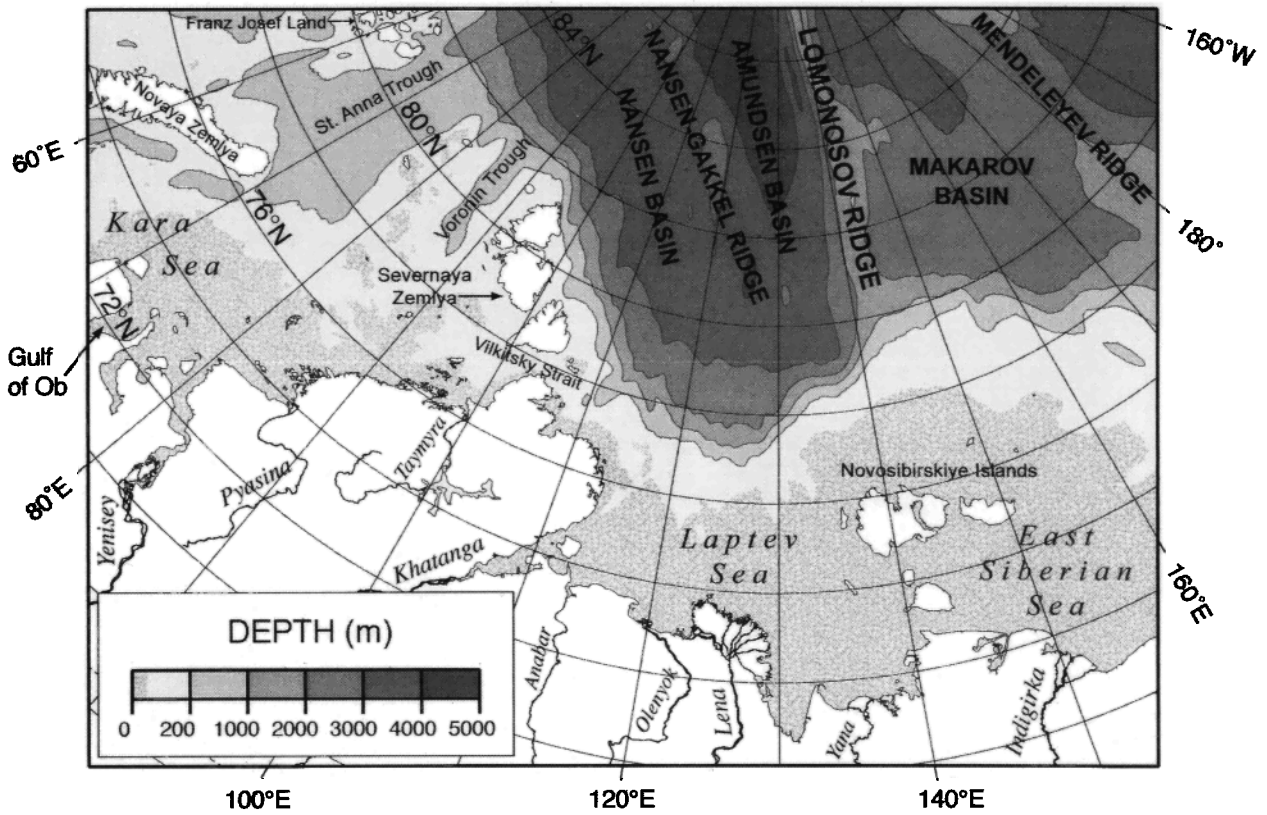

Figure 1. Map of the Laptev Sea and vicinity. Stippled area indicates bottom depth ≤ 50 m. Note that the Gulf of Ob receives discharge from the Ob, Taz, Pur, and Nadym rivers.

waters (depth ≤ 50 m) of the Laptev Sea and adjacent areas in three different years. The geographical distributions of these tracers identify regions significantly influenced by fluvial discharge and suggest pathways by which Eurasian river waters transit the shelves and enter the interior Arctic Ocean.

2. Sample Collection and Analysis

The R/V *Polarstern* occupied 236 hydrographic stations in the Laptev Sea and adjacent areas of the Nansen and Amundsen basins (collectively referred to as the Eurasian Basin) and the Makarov Basin during three cruises in the summers of 1993, 1995 and 1996 (Table 2) [*Fütterer*, 1994; *Rachor*, 1997; *Augstein*, 1997]. At all of the stations, vertical profiles of temperature and salinity were obtained using a conductivity-temperature-depth (CTD) instrument (SeaBird SBE 911 in 1995 and NBIS Mark IIIb in 1993 and 1996). The CTDs were calibrated in the laboratory before and after each cruise. In addition, the CTD salinity data were calibrated against values of salinity determined on board for discrete water samples collected during the hydrocasts. Accuracies are within 3 dbar for pressure, 0.005°C for temperature, and 0.005 for salinity. At 210 of the stations, water samples for Ba analysis were obtained from depths throughout the water column using Niskin bottles deployed on a rosette sampler. Barium concentrations ([Ba]) were determined for the samples by isotope dilution inductively coupled plasma mass spectrometry (ICP-MS) at Oregon State University on a Fisons PlasmaQuad II [*Guay and Falkner*, 1998]. The precision of the analytical procedure ranges from better than 5% at 10 nmol Ba L^{-1} to better than 3% at 100 nmol Ba L^{-1}.

3. Results

At each station, mean values of temperature and salinity were calculated from all CTD measurements made at depths above 50 m. Vertically integrated Ba inventories were determined for the upper 50 m of the water column at each station by calculating the area bounded by the [Ba] profile between 0 and 50 m. Mean values of [Ba] were calculated by dividing the

Table 1. Mean Annual Discharge of Major Rivers flowing into the Laptev and Kara seas.

River	Annual Discharge[a] $km^3\ yr^{-1}$
Laptev Sea	
Lena	525
Khatanga	85.3
Olenyok	35.8
Yana	34.3
Anabar	13.2
Kara Sea	
Yenisey	620
Ob	404
Pyasina	56.2
Taz	48.5
Taymyra	31.2
Pur	28.1
Nadym	14.8

[a]From *Meybeck and Ragu* [1995].

Table 2. Cruises to the Laptev Sea and Adjacent areas on board the R/V *Polarstern* in 1993, 1995, and 1996.

Cruise	Year	Dates	Total Number of Stations	Stations With Ba Samples
ARK IX/4	1993	Aug. 26 to Sept. 24	47	40
ARK XI	1995	Jul. 19 to Sept. 11	87	70
ARK XII	1996	Jul. 16 to Sept. 4	102	100

vertically integrated Ba inventories by 50 m (or the depth of the deepest Ba sample for stations with water depth < 50 m). This depth-weighted averaging technique was used to minimize any bias created by nonuniform vertical spacing of samples. An averaging depth of 50 m was chosen because it includes the portion of the water column most strongly influenced by fluvial discharge and allows direct comparison between shallow shelf stations (bottom depth ≈ 50 m) and deeper stations over the slope and basin areas. Performing the calculations using a greater averaging depth (down to 200 m – i.e., the maximum depth observed for the base of the halocline layer at these stations [*Schauer et al.*, 1997], below which river water would not be expected to penetrate) yields results consistent with our conclusions based on an averaging depth of 50 m.

3.1. 1993 Stations

Relatively high mean temperature (-0.42°C), low mean salinity (30.12), and highest mean [Ba] (65 nmol Ba L^{-1}) were observed at the shallowest (bottom depth = 38 m), southernmost station over the Laptev shelf (Plates 1a-1c). Warmer (mean T = -0.65° to 0.20°C), fresher (mean S = 29.15-32.09) waters slightly elevated in Ba (mean [Ba] = 45-51 nmol Ba L^{-1}) were also observed in western Vilkitsky Strait. At the rest of the stations over the northern Laptev shelf and slope and adjacent areas of the Eurasian Basin, mean temperature ranged from -1.81° to -1.24°C, mean salinity ranged from 31.04 to 33.70, and mean [Ba] ranged from 42 to 50 nmol Ba L^{-1}. Overall, temperature and [Ba] tended to decrease and salinity tended to increase moving offshore from shelf to basin.

3.2. 1995 Stations

Mean temperature ranged from -1.80° to 1.14°C, generally decreasing offshore from shelf to basin (Plate 1d). Highest temperatures were observed at stations over the northeastern Laptev shelf. Mean salinity ranged from 30.88 to 34.06 and was higher in the western part of the sampling area than in the eastern part (Plate 1e). Highest values (33.00-34.06) were observed at the stations in the vicinity of Severnaya Zemlya. Relatively low salinity waters (mean S = 30.88-33.48) covered the northern Laptev shelf and extended over adjacent areas of the Amundsen and Makarov basins and the Lomonosov Ridge. Exceptionally high values of mean [Ba] (64 and 70 nmol Ba L^{-1}) were observed at two shallow stations (bottom depths = 49 and 33 m) north of the Novosibirskiye Islands (Plate 1f). Mean [Ba] ranged from 43 to 60 nmol Ba L^{-1} at the rest of the stations, with relatively high values (50-60 nmol Ba L^{-1}) observed east of Severnaya Zemlya and over the Laptev shelf and adjacent areas of the Amundsen Basin. Mean [Ba] was lowest (43-48 nmol Ba L^{-1}) at the stations northwest of Severnaya Zemlya and at the farthest offshore stations in the Nansen Basin.

An examination of individual profiles for the 1995 stations identified 127 samples (out of 978 total samples) that had anomalously high [Ba] with respect to the other samples at that station and were not correlated with any features in temperature, salinity, or $\delta^{18}O$ profiles (Figure 2; oxygen isotope data from these cruises are presented in detail by *Frank* [1996], *Stein* [1996] and M. Mensch et al. (Water mass distribution and circulation in the Eurasian Basin of the Arctic Ocean, submitted to *Journal of Geophysical Research*, 2000)). Values of [Ba] were extremely high (between 240 and 800 nmol Ba L^{-1}) in three of these samples and ranged between 51 and 183 nmol Ba L^{-1} in the remainder. All 3 of the extreme samples and 37 of the other suspect samples were collected from the Niskin bottle in position 17 on the rosette, strongly suggesting that it was contaminated. The other anomalous samples suggest some degree of contamination from additional sources. Although we cannot be certain that these high [Ba] values do not represent contributions from some unidentified natural source, we opted for a conservative approach and excluded these suspect samples from the calculations of mean [Ba] for the 1995 stations. Note that [Ba] values for most of the 1995 samples are in agreement with values for the 1993 and 1996 samples having similar T-S properties, indicating that contamination did not affect the entire 1995 Ba data set [*Guay*, 1999]. While inferences based on the Ba data alone might still be regarded with uncertainty, our conclusions are reinforced by the combined tracer data set.

3.3. 1996 Stations

Mean temperature ranged from -1.82° to -1.34°C, with highest values occurring near the shelf break and over the Lomonosov Ridge (Plate 1g). Lowest temperatures occurred over the St. Anna and Voronin troughs and the central Eurasian Basin. Mean salinity was highest (33.76-34.43) at stations between Franz Josef Land and Severnaya Zemlya and over the central Eurasian Basin (Plate 1h). Mean salinity was lowest (32.42-33.61) along the southern (latitude < 84°N) transects across the Lomonosov Ridge and over the slope north of the Laptev Sea and Novosibirskiye Islands. Intermediate values (33.22-34.03) were observed over the St. Anna and Voronin troughs and along the northernmost stations over the Lomonosov Ridge. Mean [Ba] ranged from 43 to 53 nmol Ba L^{-1}, with highest values (47-53 nmol Ba L^{-1}) observed along the southern transects across the Lomonosov Ridge and over the slope north of the Laptev Sea and Novosibirskiye Islands (Plate 1i). Relatively high values of mean [Ba] (48-51 nmol Ba L^{-1}) were also observed at stations over the St. Anna and Voronin troughs. Mean [Ba] was lowest (43-47 nmol Ba L^{-1}) at the stations in the central Eurasian Basin.

4. Discussion

4.1. Distributions of Fluvial Discharge Inferred From Tracer Distributions

Eurasian Arctic rivers are much warmer (T > 10 °C during the summer) and lower in salinity ($S \approx 0$) than the marine waters

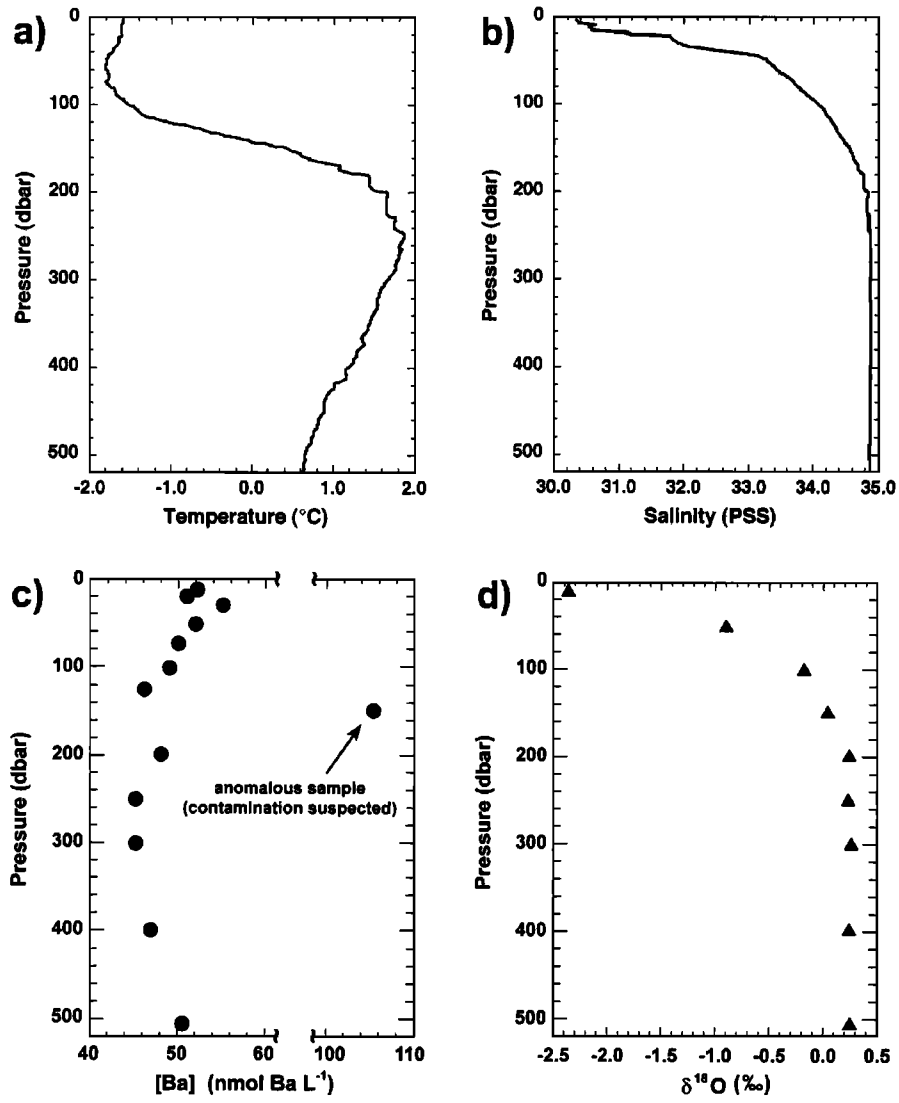

Figure 2. ARK XI station 44 (79.18°N, 135.06°E, occupied on August 17, 1995). Profiles of (a) temperature, (b) salinity, (c) [Ba], and (d) $\delta^{18}O$.

into which they flow. But temperature and salinity alone cannot be used as tracers of fluvial discharge because Arctic marine surface waters are also affected by heat exchange with the atmosphere and ice melting/formation processes. Barium can be used as an additional fluvial tracer in the Arctic due to its relative enrichment in Arctic rivers. The effective end-member Ba concentrations are 100-200 nmol Ba L^{-1} for Eurasian Arctic rivers and 42-45 nmol Ba L^{-1} for North Atlantic source waters [*Guay and Falkner*, 1998]. Thus, in Plate 1, waters significantly influenced by fluvial discharge are indicated by relatively high temperature and low salinity coincident with high [Ba]. The majority of the total annual discharge of Arctic rivers occurs from June to September, when large amounts of snow and ice melt in their drainage basins (Figure 3). We interpret the mid-to-late summer tracer distributions in Plate 1 as quasi-synoptic snapshots integrating the peak discharge period (see section 4.2 for further discussion about the residence time of Arctic river waters over the shelves).

Fluvial discharge was largely confined to the Laptev shelf during the summer of 1993. The influence of the Lena River was evident at the southernmost station near the Novosibirskiye Islands. Stations in Vilkitsky Strait reflected the eastward flow of shelf waters from the Kara Sea, which contained significant fluvial component. Fluvial influence decreased over the slope and basin areas north of the shelf break.

The most intense fluvial signals were observed over the northeast Laptev shelf in the summer of 1995. In contrast to the situation observed in 1993, surface waters bearing a significant fluvial component extended beyond the shelf break into adjacent areas over the Amundsen and Makarov basins and the Lomonosov Ridge. The influence of river waters was also apparent near the shelf break in the vicinity of Severnaya Zemlya, consistent with eastward flow from the Kara Sea and/or contributions from local sources.

Surface waters containing an appreciable fluvial component were observed over the slope north of the Laptev Sea and Novosibirskiye Islands in the summer of 1996. The influence of fluvial discharge was also evident at the stations over the Lomonosov Ridge, although it was less pronounced along the northernmost transect across the ridge than at the stations farther south. Observations over the St. Anna and Voronin troughs were consistent with an outflow of river-influenced

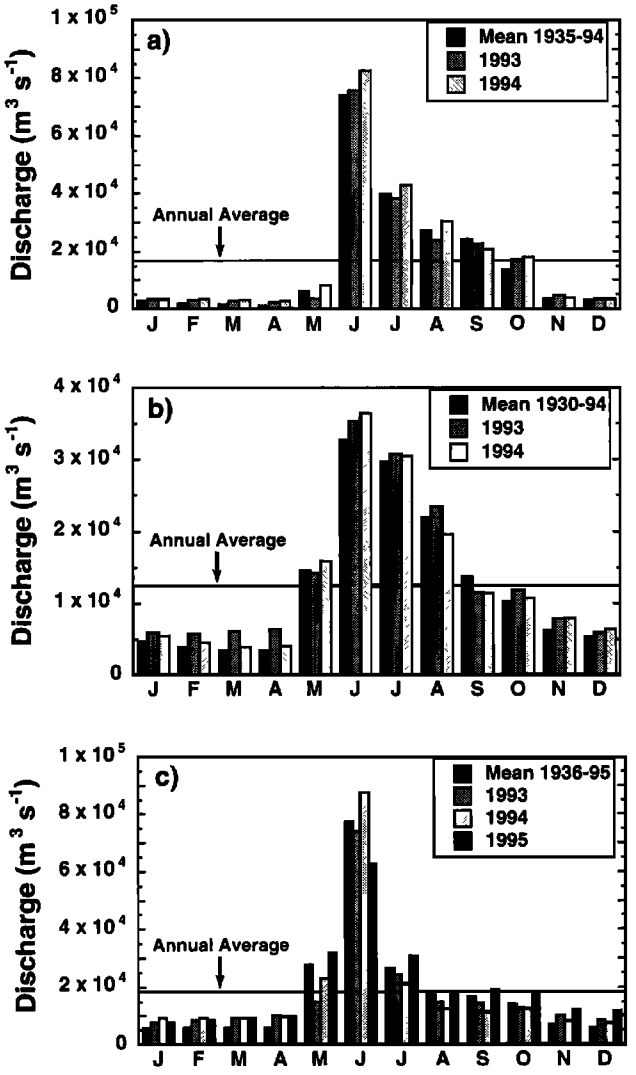

Figure 3. Mean monthly discharge for the (a) Lena River at Kyusyur, (b) Ob River at Salekhard, and (c) Yenisey River at Igarka [*Lammers and Shiklomanov*, 2001].

shelf waters from the Kara Sea. Fluvial signals were minimal at the stations in the central Eurasian Basin.

These qualitative observations are supported by statistical comparisons between the tracer data collected in 1993, 1995 and 1996 at the stations over the slope and basin north of the Laptev Sea (i.e., the geographical region of maximum overlap between sampling locations in the different years; Figure 4). At these stations, vertically integrated Ba inventories (which we interpret as proxies for river water inventories) were calculated for the upper 50 m of the water column. Mean Ba inventories were higher for both the 1995 and 1996 stations than for the 1993 stations, and the mean Ba inventory for the 1995 stations was higher than the mean Ba inventory for the 1996 stations (Table 3). River water inventories calculated independently from a salinity-$\delta^{18}O$ mass balance [*Bauch et al.*, 1995; *Frank*, 1996; *Stein*, 1996] were also higher at the 1995 stations than at the 1993 stations (Table 4). At the single station in the area at which oxygen isotope data were collected in 1996, the river water inventory was higher than the mean river water inventory for the 1993 stations and lower than the mean river water inventory for the 1995 stations.

This statistical treatment was selected as an adequately robust approach for making direct, quantitative comparisons between all 3 years given the limitations of the available data. While some degree of bias may be introduced by the nonuniform distribution of sampling locations in the different years,

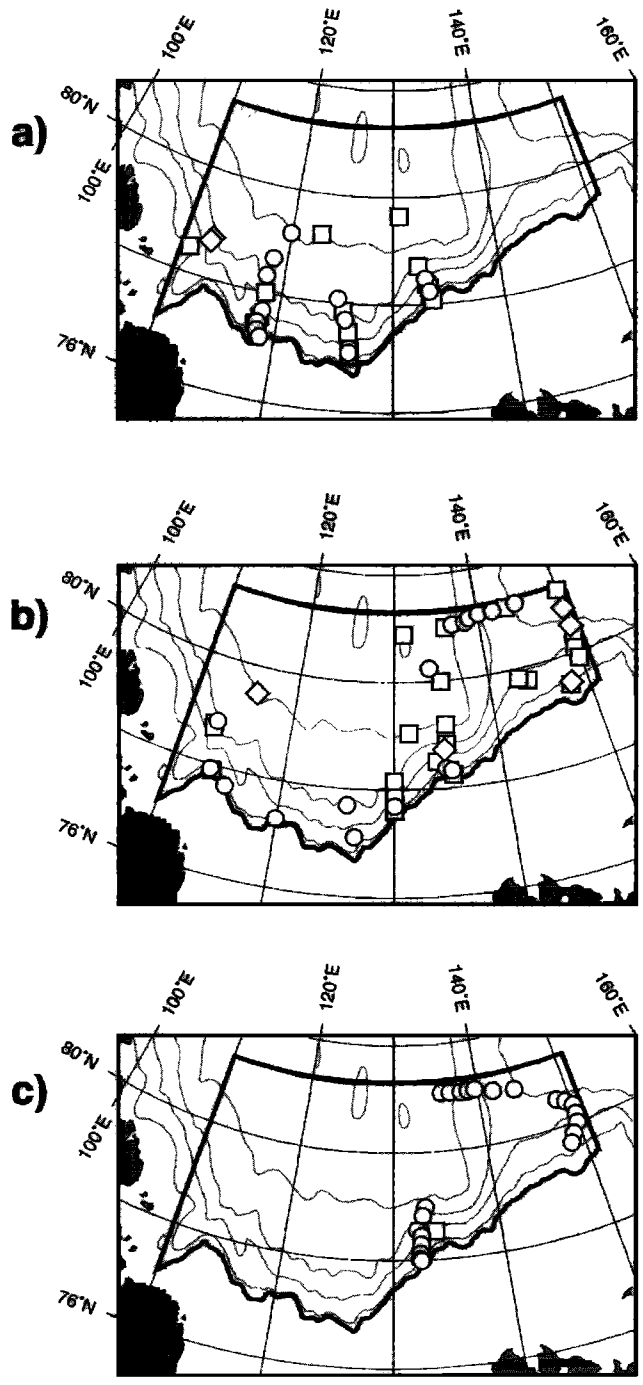

Figure 4. Stations over the slope and basin north of the Laptev Sea in (a) 1993, (b) 1995, and (c) 1996. The outlined area (100-m depth contour to 81.3°N, 110°-151°E) encloses all stations included in calculations of Ba and river water inventories (Tables 3 and 4; note that river water inventories were calculated independently based on a salinity-$\delta^{18}O$ mass balance). Circles denote stations with Ba data only, diamonds denote stations with $\delta^{18}O$ data only, and squares denote stations with both Ba and $\delta^{18}O$ data.

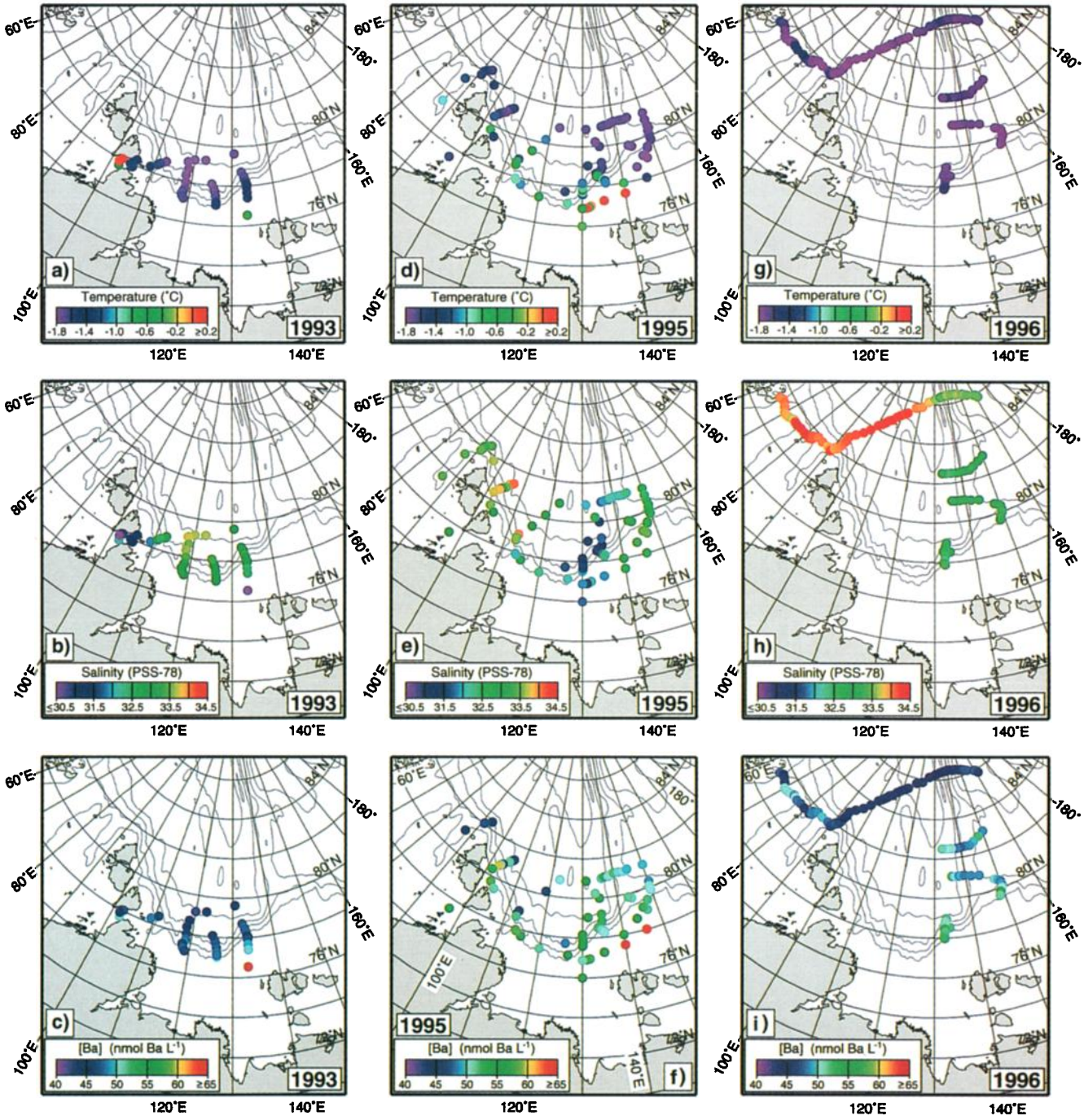

Plate 1. Distributions of mean temperature, salinity, and [Ba] in the upper 50 m of the water column in (a)-(c) 1993, (d)-(f) 1995, and (g)-(i) 1996.

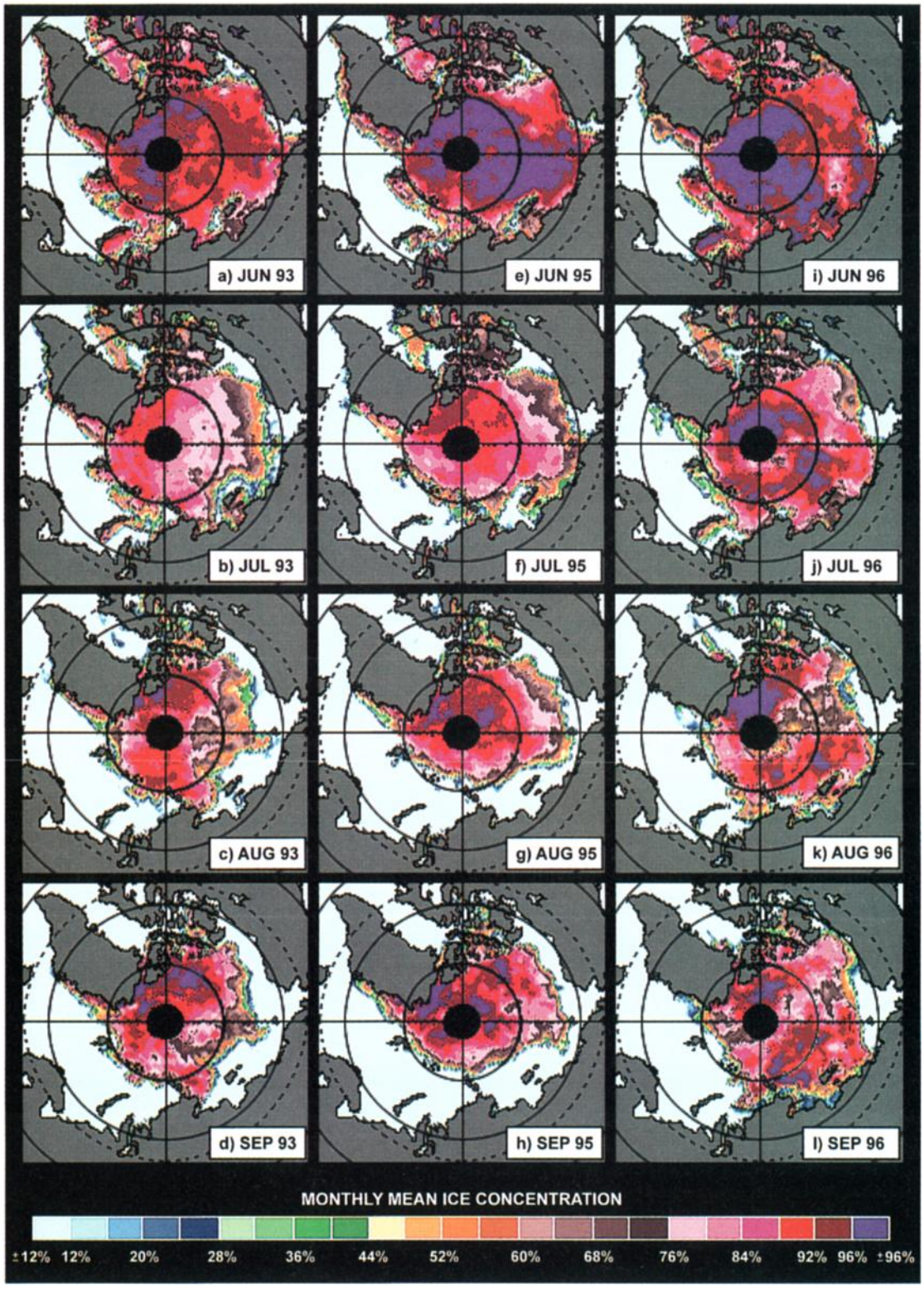

Plate 2. Monthly mean ice concentrations in the Arctic Ocean for June-September in (a)-(d) 1993, (e)-(h) 1995, and (i)-(l) 1996. Ice concentrations < 12% are considered open water. Ice concentrations were derived from passive microwave data obtained by the Special Sensor Microwave/Imager. Data from the DMSP-F11 satellite were used in 1993 and 1995, and data from the DMSP-F13 satellite were used in 1996 [*Cavalieri et al.*, 1999].

Table 3. Comparison Between Vertically Integrated Ba Inventories for the Upper 50 m of the Water Column at Stations Over the Slope and Basin North of the Laptev Sea (Within the Outlined Area in Figure 4) in 1993, 1995, and 1996.

Ba Inventory	1993 Stations	1995 Stations	1996 Stations
Number of stations with Ba samples	30	41	23
Mean Ba inventory, mmol Ba m^{-2}	2.34	2 60	2.53
Standard deviation	0 078	0 150	0.090
Two-Sample T-Test Results (H_0: $\mu_1 = \mu_2$)[a]	Degrees of Freedom	P-Value (Two-Sided)	
Comparison between 1993 and 1995 mean Ba inventories	69	< 0.0001	
Comparison between 1993 and 1996 mean Ba inventories	51	< 0.0001	
Comparison between 1995 and 1996 mean Ba inventories	62	0.096	

[a]The two-sample t-test was applied to investigate the null hypotheses that no differences exist between the mean Ba inventories in the different years.

the results clearly illustrate a large interannual variability that is consistent with the highly variable surface circulation in the Laptev Sea [*Pavlov et al.*, 1996]. The data suggest that more fluvial discharge flowed across the Laptev shelf and into the central Arctic basins during the summers of 1995 and 1996 than during the summer of 1993. The data also suggest, although less conclusively, that cross-shelf transport of river water from the Laptev Sea to the interior Arctic Ocean was somewhat greater in the summer of 1995 than in the summer of 1996.

4.2. Relationship Between Local Winds and Distributions of Fluvial Discharge

Theory predicts that the net wind-driven transport of ice and surface waters occurs to the right of the wind in the Northern Hemisphere. The northerly and westerly winds prevalent over the Laptev Sea during the summer of 1993 (Figures 5a-5d) would have tended to cause onshore transport of the ice pack and surface waters in this region. In contrast, the strong southerly to southeasterly winds present during the summer of 1995 (Figures 5e-5h) would have tended to drive the ice pack and surface waters offshore from the Laptev shelf. Winds were mixed during the summer of 1996 (Figures 5i-5l), but conditions associated with offshore transport prevailed in August and September (when all of the stations over the Laptev slope and southern Lomonosov Ridge were occupied). The distributions of river discharge inferred from the tracer data are therefore qualitatively consistent with local wind forcing during the sampling periods.

Upper ocean wind currents were estimated quantitatively based on basic wind-driven current theory. Because the Laptev shelf is broad and shallow (bottom depth = 10-50 m over much of its area), calculations were made using the equations governing pure drift currents in a homogeneous ocean of finite depth [*Neumann and Pierson*, 1966]. As an example, we considered the case of offshore (i.e., southerly) winds with an average speed of 5-10 m s^{-1}. Assuming a vertically uniform eddy viscosity of 10.0-57.5 kg m^{-1} s^{-1} [*Neumann and Pierson*, 1966, p. 195, table], an Ekman depth of 40-90 m was determined. Given these conditions, the wind-driven current at the surface would have speeds of 0.03-0.07 m s^{-1} and move in a direction between 5° and 45° to the right of the wind. At this drift rate, a parcel of water discharged from the Lena delta would reach the shelf break in the vicinity of the southern terminus of the Lomonosov Ridge (a distance of ≈ 550 km) in 3-7 months.

Table 4. Comparison Between Vertically Integrated River Water Inventories for the Upper 50 m of the Water Column at Stations Over the Slope and Basin North of the Laptev Sea (Within the Outlined Area in Figure 4) in 1993, 1995, and 1996.

River Water Inventory (from Salinity-$\delta^{18}O$ Mass Balance)	1993 Stations	1995 Stations	1996 Stations
Number of stations with $\delta^{18}O$ samples	19	30	1
Mean river water inventory, m	2.80	4.24	4.04
Standard deviation	0.457	1.002	--
Two-Sample T-Test Results (H_0: $\mu_1 = \mu_2$)[a]	Degrees of Freedom	P-Value (Two-Sided)	
Comparison between 1993 and 1995 mean river water inventories	47	< 0.0001	

[a]The two-sample t-test was applied to investigate the null hypotheses that no difference exists between the mean river water inventories in 1993 and 1995.

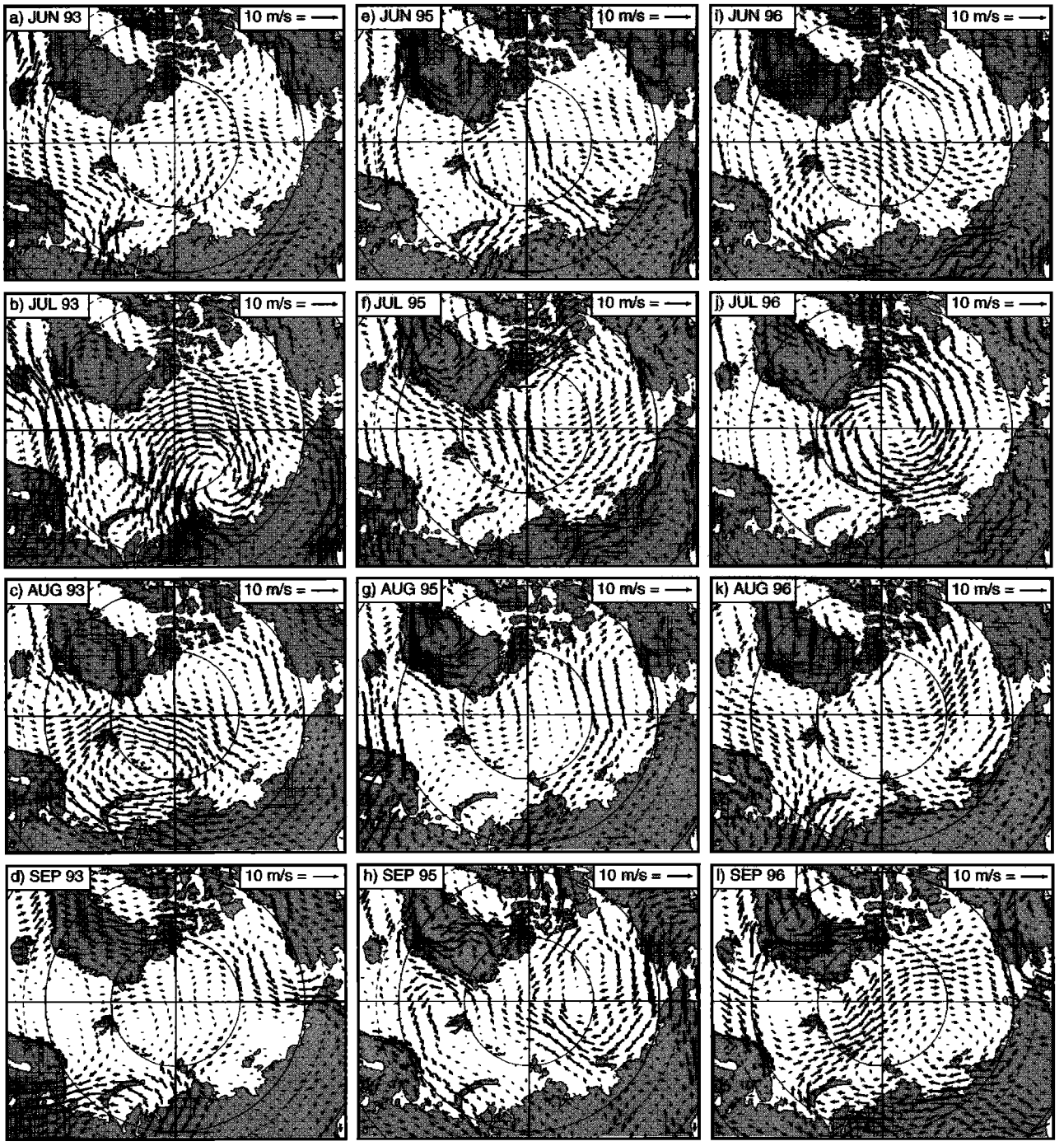

Figure 5. Monthly mean geostrophic winds in the Arctic for June-September in (a)-(d) 1993, (e)-(h) 1995, and (i)-(l) 1996. Geostrophic winds were calculated from Navy Operational Global Atmospheric Prediction System sea level pressure fields [*Hogan and Rosmond*, 1991]. (Data were provided by P. Posey and R. Preller, Naval Research Laboratory, Stennis Space Center, 1999.)

The assumption of a uniform vertical eddy viscosity is almost certainly an oversimplification given the varying degrees of water column stratification that exist during different seasons and at different locations in the Laptev Sea. In the case of a shallow surface layer overlying a sharp density gradient (a common situation over the Arctic shelves), the upper layer could "slide over" the lower layer at speeds significantly higher than our estimates while the lower layer remained nearly unaffected by the wind. Furthermore, the monthly averaged wind fields used in our calculations neglect sharp event-driven spikes in wind speed and therefore underestimate the actual wind stress (which varies as the square of the wind speed). These simplifications bias our results strongly toward the conservative and likely underestimate the actual wind-driven upper ocean current velocities. To a first approximation, our calculations show that it is physically feasible for an offshore wind field (such as the one observed during summer 1995) to drive low-salinity surface waters across the Laptev shelf on

time scales of months. A more detailed treatment is beyond the scope of this paper.

Coupled Arctic Ocean and sea ice models also indicate that some portion of fluvial discharge to the Kara and Laptev seas typically transits the shelves and enters the interior basins within a year of leaving the river mouths [*Maslowski et al.*, 2000]. This time frame is consistent with tritium/^{3}He data previously published from the ARK IX/4 cruise [*Frank et al.*, 1998], which show ages of < 2 years in the upper 50 m of the water column over the Laptev shelf and adjacent areas of the slope and basin. Further corroboration is provided by data obtained from the Kara Sea in 1994 [*Ekwurzel*, 1998], which show no tritium/^{3}He age – i.e., the presence of tritium with no accumulation of the ^{3}He daughter product – in surface waters extending to the shelf break (note, however, that near-zero tritium/^{3}He ages could also occur in waters that resided over the shelf for longer periods of time but maintained constant ventilation with the atmosphere).

These transit times are short relative to previous estimates of the residence times of Arctic shelf waters. *Hanzlick and Aagaard* [1980] determined a residence time of 2.5 years based on a mass balance for the Kara Sea. *Schlosser et al.* [1994] inferred a mean residence time of 3.5 ± 2 years from tritium/^{3}He measurements in the central Nansen Basin (an area that likely integrates runoff from the Barents, Kara and Laptev seas). In both cases (neither of which is specific to the Laptev Sea system), the calculations require the shelves to be modeled as a single well-mixed reservoir. These residence times represent volume-averaged estimates for all waters over the entire shelf and therefore do not eliminate the possibility of a low-density river-influenced surface component that transits the shelf in a much shorter time.

Ice concentrations in the Laptev Sea (Plate 2) were consistent with local winds in the summers of 1995 and 1996. In the summer of 1993, however, the ice cover over the Laptev shelf was relatively light despite the prevalence of winds associated with an onshore transport regime. This demonstrates the influence of other factors besides atmospheric forcing, such as basin-scale ice motion and local melting due to the influx of warm river waters and solar insolation during the summer. The dynamics governing the Arctic ice cover are very complex, and it is not unusual for ice motion to be decoupled from surface ocean flow. Note that ice concentrations below 50% were observed over extensive areas of the Laptev Sea in every month except June and July of 1996, indicating a significant amount of open water that wind forcing could influence directly.

Our results suggest two principal pathways by which fluvial waters can enter the interior Arctic Ocean from the Laptev Sea. When southerly to southeasterly wind conditions prevail, shelf waters containing fluvial discharge are transported offshore beyond the Laptev shelf break and over the slope and adjacent basin areas. These waters can then enter the Amundsen Basin via northward surface flow in the vicinity of the Lomonosov Ridge. Under other wind conditions, fluvial discharge is steered primarily along the inner Laptev shelf and into the East Siberian Sea as part of the predominantly eastward coastal current system [*Pavlov et al.*, 1996]. These waters then appear to cross the shelf and enter the Makarov Basin via upper layer flow aligned roughly along the Mendeleyev Ridge [*Guay and Falkner*, 1997; *Wheeler et al.*, 1997]. The prevalence of either pathway in a given year is largely determined by local wind patterns during the summer months, when fluvial discharge is greatest and shelf waters are at the lowest salinity of their annual cycle. Uncertainty remains, however, regarding the proportion of the total annual river discharge that leaves the shelf each year and the extent to which circulation over the shelves and adjacent areas integrates multiple years of discharge. Additional observations will be required to resolve these issues.

4.3. Transport Pathways of Fluvial Discharge and Cold Halocline Formation

An understanding of the temporal variability in river water mixing pathways is relevant to investigations of the recent changes occurring in the Arctic system. From the late 1980s to the mid-1990s, a decrease in sea level pressure and a strengthening of the wintertime stratospheric polar vortex have been observed in the Arctic [*Walsh et al.*, 1996; *Thompson and Wallace*, 1998]. These conditions are associated with the positive phases of the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO), i.e., principal modes of atmospheric variability that account for a significant amount of high-latitude climate variability in the Northern Hemisphere [*Hurrell.*, 1995; *Thompson and Wallace*, 1998].

Changes in Arctic Ocean circulation have also occurred over the same time period and appear to be related to changes in atmospheric forcing. The Beaufort Gyre has weakened, and waters of Atlantic origin have advanced farther into the Arctic basin [*Carmack et al.*, 1995; *Carmack et al.*, 1997; *Morison et al.*, 1998]. As a result, the boundary between Atlantic and Pacific water masses and the axis of the Transpolar Drift have shifted from the Lomonosov Ridge to a position aligned roughly over the Mendeleyev Ridge [*Carmack et al.*, 1995; *McLaughlin et al.*, 1996; *Morison et al.*, 1998]. Decreases in Arctic sea ice cover have also been observed, which may be the result of divergent flow associated with these atmospheric and oceanic conditions [*McPhee et al.*, 1998; *Macdonald et al.*, 1999]. Whether these trends reflect a naturally occurring oscillation between two dominant circulation modes of the Arctic ice-ocean-atmosphere system [*Proshutinsky and Johnson*, 1997] and/or the effects of anthropogenic climate change [*Fyfe et al.*, 1999; *Shindell et al.*, 1999] remains unclear.

Data collected from the central Eurasian Basin during two submarine cruises in August-September 1993 (SCICEX-93) and April-May 1995 (SCICEX-95) show a salinification of the upper water column and a diminishing of the cold halocline layer relative to previous years [*Steele and Boyd*, 1998]. It could be argued that the observed salinification in the Eurasian Basin simply resulted from an increase in the salinity of North Atlantic waters entering the Arctic through Fram Strait and the Barents Sea. But the atmospheric regime prevailing during the late 1980s to mid-1990s (i.e., the positive phase of the NAO) is associated with increased precipitation over the Nordic Seas [*Furevik*, 2001; *Dickson et al.*, 2000], which would have tended to reduce the salinity of Atlantic source waters to the upper Eurasian Basin. The observed changes in ocean conditions were instead attributed to changes in Arctic atmospheric forcing that presumably shifted the primary insertion point of river-influenced Kara and Laptev shelf waters from the Amundsen Basin to the Makarov Basin in the late 1980s.

Our tracer distributions and wind data for the summer of 1993 provide direct evidence supporting the shift in river water transport pathways proposed by *Steele and Boyd* [1998].

Although no tracer distributions are presented here for 1994, local winds during the season of maximum river discharge (R. Preller, personal communication, 1999) would have also tended to oppose offshore transport from the Laptev shelf to adjacent areas of the Eurasian Basin (thus forcing river waters to travel farther east before crossing the shelf and entering the interior Arctic Ocean). This scenario is consistent with tracer data from other cruises to the Arctic in 1993-1994, which show discharge from Eurasian rivers entering the Makarov Basin along the Mendeleyev Ridge pathway in these years [*Guay and Falkner*, 1997; *Wheeler et al.*, 1997].

Cross-shelf transport of river water from the Laptev Sea to the interior Arctic Ocean appears to have resumed during the summers of 1995 and 1996. Although no tracer data were available for 1997, wind conditions associated with offshore transport from the Laptev shelf prevailed during most of the summer in this year as well (R. Preller, personal communication, 1999). These observations suggest that discharge from Siberian rivers would have contributed to reestablishing the cold halocline layer in the Eurasian Basin in the years following the SCICEX-95 submarine cruise in the spring of 1995. Preliminary data from subsequent submarine cruises to the Arctic in 1996-1999 support this hypothesis (R. Muench, unpublished data, 1999).

The extent of fluvial influence and the formation of the cold halocline layer in the Eurasian Basin thus appear to be strongly dependent on wind-driven transport from the Laptev shelf to the interior Arctic Ocean along the Lomonosov Ridge pathway. The degree to which ocean circulation integrates multiple runoff seasons under varying atmospheric forcing conditions will determine how rapidly the cold halocline is diminished or restored in response to a shift in the transport regime. The limited available data suggest a timescale on the order of a few years or less.

5. Conclusions

Because the areal distribution of fluvial discharge as well as the total amount of runoff ultimately determines its effect on Arctic Ocean circulation, it is important to understand the mechanisms governing the transport of Eurasian Arctic river discharge and their temporal variability. Based on model results, observations of ice motion, and a limited number of tracer measurements, it had previously been speculated that fluvial discharge to the Kara and Laptev seas enters the interior Arctic Ocean via surface flows in the vicinity of the Lomonosov and Mendeleyev ridges. Our data provide direct evidence of these transport pathways and suggest that they are driven primarily by atmospheric forcing. While recent investigations of variability in the Arctic system have focused on basin-scale atmospheric and oceanic phenomena, our results highlight the importance of local conditions over the shelves during the summer period of maximum runoff.

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