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Abstract

A summary is presented of present opinions on the possible impact on Arctic sea ice of Soviet plans to divert a number of their northward flowing rivers, namely the Northern Dvina, Pechora, Ob and Yenisey, southwards to alleviate water shortages in Central Asia and Kazakhstan and counter falling water levels in the Aral and Caspian Seas. The background to and size of the possible diversions is outlined and discussed in the context of the magnitude of the fresh water input to the Arctic Ocean and the natural variability of riverflow. Oceanographic processes and hydrological features of the Arctic Seas, relevant to the impact of river diversions on sea ice, are summarised and the possible effects of substantial diversions, which have been suggested could become operational during the 21st century, are discussed with particular reference to recent observational and modelling studies. These indicate some sensitivity of sea ice to riverflow in the shelf regions of the Kara and Barents Seas.

Contents

Introduction

The Soviet Union has, for many years now, been acutely aware of the disparity between the areas for which its needs for freshwater are greatest and the locations of its major river systems. In terms of river flow the country has very large resources of freshwater. In all these amount to a total runoff of some 4700 $km^3\ yr^{-1}$; more than enough by far to meet demands which amount to some 309 $km^3\ yr^{-1}$ (1979 figures; Micklin, 1983). The problem is, however, that about 84% (3900 $km^3\ yr^{-1}$) of this runoff flows through sparsely utilised northern and Far Eastern regions with much of it (63%) draining into the Arctic Ocean (Micklin, 1979). Needs for freshwater are, however, greatest in the southernmost areas of the country, notably in the areas which include the drainage basins of the Azov-Black and Aral-Caspian Seas (Micklin, 1979a) and which have about 12% of the riverflow (the remaining 4% drains into the Baltic Sea). (Figure 1 shows a number of the geographical features referred to in this section). It is in the southern areas, which cover about 30% of the USSR and support 75% of the population, that the major drive towards expansion of agricultural activities is taking place and where demands have been such as to lead, over the period since the 1930s, to ecological damage to the Aral, Azov and Caspian Seas and their environs due to falling water levels and rising salinity (see e.g. Voropaev and Kosarev, 1982; Micklin, 1983). As a result the Soviet Government has long contemplated the possibility of channelling a portion of the northward flowing river waters to these southern areas. A variety of schemes have been put forward over the years involving diversions both from European Russia and from Siberian rivers. Broadly, two schemes are currently in favour for transfers from European Russia. The so-called western variant involves diversions from lakes south of the White

Sea and from the Northern Dvina/Sukhona rivers into the Volga drainage basin. The eastern variant involves diversions of water from the Pechora river for transfer southwards via the river Kama. Plans for Siberian transfers are initially for diversions from the Ob and Irtysh rivers south over the Turgay divide into South Kazakhstan and Central Asia. These, and the background to the proposals for large scale water transfer have recently been discussed in detail by Kelly, Campbell, Micklin and Tarrant (1983). Though massive in engineering terms the diversion schemes are made technically and economically feasible due to the relative proximity of the source regions for rivers flowing off the northern and southern slopes of the USSR and the relatively low divides which separate them. However, because of their scale and the complexities of the environmental issues involved, particularly internally to the USSR (Micklin, 1979), decisions to implement the schemes have been slow. Micklin (1983) presents some of the internal controversy over the diversion plans which has taken place in the Soviet Union in recent years. Present indications are for small diversions (from the Northern Dvina and Pechora rivers) of up to 20 to 25 $km^3 yr^{-1}$ by the end of the century, though there are further indications that the Soviet authorities are considering diversions of some 100 $km^3 yr^{-1}$ during the first decades of the next century and even 200 $km^3 yr^{-1}$ by 2050 (Kelly and others, 1983; Holt, Kelly and Cherry, 1984), largely by additional withdrawals from the Ob and Yenisey. As yet, however, it appears difficult to distinguish general discussion of the various schemes put forward within the Soviet Union from specific intentions to implement any of them; it is not even clear, for example, whether it is the first transfers, amounting to some 5 to 10 $km^3 yr^{-1}$ into the Volga basin, that are expected to begin by 1990 or their construction (Micklin, 1983).

The size of the withdrawals postulated, in particular by some of the more grandiose schemes that have been floated, has led over the years to speculation on the question of whether the diversions could markedly affect the structure and circulation of the Arctic Ocean, and thereby the extent of sea ice, and, as a consequence, the global climate itself. Some indication of the sensitivity of the climate to changes in the ice margins comes, for example, from the experiments of Herman and Johnson (1978) who compared two runs of the GLAS general circulation model in which the maximum and minimum winter Arctic sea ice extents observed over the period 1961-1977 were imposed. They concluded that changes in the position of the ice margin may have a significant infuence on the modelled climate on both local and hemispheric scales. More extreme experiments of this type are those in which the entire Arctic sea ice cover has been removed and replaced with water at temperatures close to freezing. Such experiments were carried out by Newson (1972) and Warshaw and Rapp (1973). The result was to lower the surface pressure and markedly reduce the vertical stability of the atmosphere over the Arctic Basin as a result of the warming of the lower layers of the troposphere. Warshaw and Rapp also found a slight increase in the strength of the circulation around the periphery of the Arctic but otherwise a general weakening of the mid-latitude westerlies, in broad agreement with Newsons' results. Whether significant changes in sea ice extent might result from the Soviet river diversions is, of course, another matter. Arguments have been presented on the one hand for a possible decrease in sea ice extent (Lamb, 1971; Aagaard and Coachman, 1975; Aagaard, Coachman and Carmack, 1981) and, stemming from a paper by Antonov (1978), for a possible overall increase on the other (Micklin, 1981). The former suggestion arises from the consequences which the withdrawals would have on the vertical structure of the upper waters of the Arctic Ocean, in particular on the role of the rivers in helping to maintain a low salinity surface layer; the latter from the possible effects the withdrawals might have on the system of Arctic Ocean currents and, though on a much smaller scale, to changes in the thermal output from the rivers.

The purpose of the present paper is to review the current position with regard to the likely effects on the Arctic ice cap of a possible reduction in the inflow from the Northern Russian rivers (notably the Northern Dvina, Pechora, Ob and Yenisey) into the Arctic Seas. Features of the hydrographic structure relevant to this problem are first outlined, followed by a brief discussion of the possible diversions in the context of the total fresh water input to the Arctic Ocean and the natural variability of the flow of the rivers likely to be involved in the schemes. Possible effects of substantial diversions of riverflow are then discussed with particular reference to recent observational studies by Holt, Cherry and Kelly (1984) and Arctic Ocean model integrations run by Semtner (1984).

The hydrographic structure of the Arctic Ocean

The characteristic structure of the Arctic Ocean reflects the influence of a number of processes and water sources. This structure is depicted in Figure 2(a) from Aagaard, Coachman and Carmack (1981). The data are for the locations shown in Figure 3 which also shows further geographical features mentioned in the text. At the surface, cold but relatively fresh water is found in a shallow well mixed layer of order tens of metres deep. Beneath this layer lies the Arctic halocline in which the salinity increases markedly but the temperature remains close to the near freezing values of the surface layer to a depth of 100 to 150m or so. Below this the temperature increases with depth with further, but slower increase in salinity to values characteristic of the water of Atlantic origin which lies below. This water largely enters the Arctic from the North Atlantic through the Fram Strait between Greenland and Spitzbergen by the West Spitzbergen current (see e.g. Coachman and Aagaard, 1974). The waters of this current lose heat by surface contact in that vicinity and submerge as they travel northwards beneath the cold but relatively fresh waters of the Arctic surface layer. The resultant temperature profile, therefore, is an unstable one overall, but is prevented from overturning by the very nature of the salinity structure which renders the density profile stable (Figure 2(b)). This restricts the effects of vertical overturning on the whole to all but the shallow surface layer. The effectiveness of the low salinity water in maintaining this stable density structure is enhanced by the greater sensitivity of changes in density to those of salinity rather than temperature for Arctic conditions. As discussed in the next section the most significant source of freshwater for this layer lies in the river input from around the peripheries of the Arctic Basin. This is partially implied by Figure 4 which shows the horizontal field of salinity in the surface waters for the summer months. The influence of saline surface North Atlantic water introduced into the Arctic from the Eastern Norwegian-Greenland Seas can clearly be seen to the north of Scandinavia. Note the low salinities found in the shelf regions along the northern coasts of European Russia and Siberia and along the northwest coast of the American continent, consistent with the addition of freshwater to the surface layers by river inflow and, in the region of the Bering Straits with the introduction of low salinity water from the Pacific. Ice melt also makes a contribution to the low salinity of these shelf waters in summer since the bulk salinity of sea ice is only of order 5‰ or less. Surface water exits from the Arctic Basin largely via the East Greenland current which also thereby removes significant amounts of freshwater in the form of sea ice.

Coachman and Aagaard (1974) suggest that the general pattern of winter salinities remains similar to that of Figure 4 with surface salinities in winter higher by 0.5 to 1‰ over the deep basins and 1 to 2‰ over the neighbouring seas. For the shelf regions, the summer to winter contrast may be much higher than this as more recent winter observations suggest. Thus Aagaard and others (1981) show an east-west section, taken using CTD equipment from a helicopter flown across the central Chukchi Sea during February and March 1977, which has salinities typically 2 to 4‰ higher than in the summer, and temperatures everywhere close to freezing. The reason for those high salinities is considered to be salt separation during freezing of sea ice, the latter being enhanced by divergence of the ice pack over that area. One would certainly expect higher salinities generally over the shelf regions in winter for at that time of year the input of fresh water by rivers is effectively cut off (see Table I) whilst mixing with higher salinity waters is continuously in progress and the salt content of the surface waters is being enhanced, as noted above, by the extrusion of salt as sea ice forms.

Returning, now, to the profiles of Figure 2(a), the extension of the low temperatures characteristic of the surface layer into the halocline has been noted and discussed by a number of authors, starting with Nansen (1902) and the question arises as to how this structure comes about.

Aagaard and others (1981) demonstrate that it cannot be a consequence of direct vertical mixing between the surface and Atlantic waters since it is not possible to produce water with these temperature and salinity characteristics by simple mixing in any proportions. Rather it must be supplied by advection; the question is from where? In fact, careful inspection of the temperature profiles in Figure 2(a) illustrates certain well known differences between the nature of the halocline structure over the Eurasian Basin (Figure 3, point 1) and that over the Canadian Basin (points 2 and 3), for in the latter there is an increase of temperature from a depth of about 50m to a slight maximum at around 75 to 100 m. Although not shown in these profiles the temperature is frequently observed to decrease again to about -1.5C before the main thermocline is reached at about 150m depth (Coachman and Aagaard, 1974). A further difference is that the salinity increase over the halocline of the Canadian Basin is generally not as rapid as it is over the Eurasian Basin. The relatively warm layer near 75m has been associated with the intrusion of summer Pacific water which passes into the Chukchi Sea through the Bering Straits. There it loses heat both from the surface to the atmosphere and by mixing with (colder) local shelf water. It is, however, initially of higher temperature and lower salinity than that which enters in winter and which has been shown to be the source for the colder water found to lie just above the main thermocline. By mixing and cooling the Bering Sea water is brought to a density intermediate between the surface waters of the Arctic and the deeper Atlantic water beneath. Its influence in the subsurface layer is found over much of the Canadian Basin, over which it spreads, in part, as a consequence of the broad scale gyral circulations. More detailed discussion of these processes will be found in Coachman and Aagaard (1974).

In fact shelf processes have been considered to play a central part in the formation of halocline water overall. Coachman and Barnes (1962) suggested the halocline water of the Eurasian Basin to be initially formed from Atlantic water forced sporadically upwards from a relatively deep position in the Arctic Basin through the submarine canyons which indent the shelf regions of the North Siberian Coast. Once on the shelves this water would not only mix but also cool to the atmosphere to develop the temperatures and salinities characteristic of the halocline. Thence it would be advected over the Arctic Basin as a source of halocline water. Such upwelling is certainly observed and some contribution to the halocline from this source is probable. However the overall importance of these processes is unknown. (Aagaard and others, 1981). A second mechanism has been proposed by these authors whereby cold and saline water is produced over the shelves by brine release during winter freezing, the effect being amplified in regions of persistent ice divergence and consequently increased ice formation. The cold upper halocline is considered then to be maintained by the eventual flow of this water out from the shelf seas. Aagaard and others (1981) point to a number of areas between Spitzbergen and the New Siberian Islands where a winter ice growth of less than about a metre would raise the salinity of the shelf water to halocline levels. This region was also proposed as a source of halocline water by Tresnikov (1959).

Freshwater sources and river discharge

Sources of fresh water for the Arctic Ocean

As implied previously the low salinity of the surface layer is maintained against export of these waters, primarily by the East Greenland current (whose additional export of fresh water in the form of sea ice is estimated, for example, by Antonov (1968) to be at the rate of 2000 $km^3 yr^{-1}$, varying by up to ± 650 km^3yr^{-1}, and by Aagaard and Coachman (1975) as 3150 km^3yr^{-1}) and against any possible mixing with saline water from below by:

i. Freshwater from the rivers of Northern Europe, Asia and Canada; estimates of river discharge into the central Arctic Basin vary. Vowinckel and Orvig (1970) quoting Antonov (1958) derive a value of 3800 km^3yr^{-1}. Coachman and Aagaard (1974) quoting Antonov (1964) for Siberian rivers and Cram (1968) for the Mackenzie give a figure of 8.5×10^4 m^3s^{-1}, equivalent to 2700 km^3yr^{-1}. Lamb (1977) gives 5000 km^3yr^{-1}, a figure consistent with estimates of Korzun and others (1978) for the total inflow into the Arctic Seas in their entirety from the American, Asian and European continents but which seems too high for the Arctic Ocean (including the Kara and Barents Seas) on its own. An estimate derived below based partly on Antonov (1958), gives 3600 km^3yr^{-1}. The rivers likely to be involved in the Russian diversion schemes contribute about 1200 km^3 yr^{-1} to the total.

ii. the excess of precipitation over evaporation; of uncertain amount, but estimated to be of order 20% of the total river flow (Lamb, 1977; Aagaard and Coachman, 1975). This is equivalent to a freshwater input into the Arctic Ocean of some 700 km^3yr^{-1}. It implies a precipitation less evaporation rate of some 70 mm yr^{-1} (taking the area of the Arctic to be of order 10^7 km^2).

iii. inflow of low salinity water from the North Pacific via the Bering Strait, though, as discussed above, much of this may go to forming halocline water rather than to making a direct contribution to the surface layer. Lamb (1977) (no references) quotes this contribution as being variously estimated as 10 to 30 10^3km^3 yr^{-1} and notes that it may well vary within this range from one period of years to another. This appears to represent the range for the mean influx of Pacific water, rather than the total freshwater contribution, since the figures are the same order of magnitude as, though smaller than, the value for the influx of Pacific water of 150 x 10^4 m^3s^{-1} (equivalent to 47 x 10^3 km^3yr^{-1}) quoted by Aagaard and Coachman (1975). Relative to the mean Arctic Ocean salinity over the upper 200m of their Figure 1 (see Figure 2(b)), namely 33.72‰ , Aagaard and Coachman calculate their value to represent a freshwater source of 5.5 x 10^4 m^3s^{-1}, equivalent to 1700 km^3 yr^{-1} (Pacific water has an average salinity of 32.5‰). Lambs figures equivalently represent a freshwater source of 400 to 1100 km^3yr^{-1}.

Variability of Arctic river drainage

Despite the discrepancies in the figures for sources of fresh water to the Arctic Basin, it is evident that river flow provides the major source for this region. As inferred previously, river flow into the Arctic is highly seasonal, the main contribution coming during the summer months as shown in Table I. Results of the International Hydrological Decade (Korzun and others, 1978) enable us to examine the year to year variability of total river flow into the central Arctic from the Asian slope, whose east-west extent is defined in Figure 3. Data for the North American and European slopes are also available, but as total runoff into all Arctic waters (including, for example, Hudson Bay) and so are not directly of use in this context. Figure 5 shows the total runoff from the mainland and islands of Asia (excluding the Pechora) for each year from 1918 to 1967. The data give a mean runoff of 2360 km^3yr^{-1} with a standard deviation of 140 km^3yr^{-1}. Following Antonov (1958) for the contributions from North America and the European USSR (also excluding the Pechora) and adding in the value for the mean flow of the Pechora derived below gives a total inflow into the Arctic Ocean of 3640 km^3yr^{-1} (Table II).

Flow of the Northern Dvina, Pechora, Yenisey and Ob rivers

Figure 1 shows the location of these rivers which are the ones likely to be involved in the diversion schemes. The Northern Dvina and Pechora provide a source of freshwater to the Arctic via the Barents Sea and the Yenisey and Ob via the Kara Sea. Data from Studies and Reports in Hydrology No 5 (UNESCO, 1971) have been extracted to show, in Figure 6, the annual runoff of these rivers over a period of some three decades. Figure 6 also shows the combined runoff and Table III summarises the corresponding means and standard deviations. These data may be expected to be of variable quality, though it is difficult to make an objective estimate of their accuracy.

As noted in the introduction, initial transfers are likely to be small, perhaps 20-25 km^3yr^{-1} by the end of the century, but may rise to 100 km^3yr^{-1} during the first decades of the 21st century and could even increase to 200 km^3yr^{-1} before its end. A reduction of 25 km^3yr^{-1} lies within the natural variability of the rivers and on the basis of the figures in Table III one would expect at least such a reduction in the total flow of these four rivers to occur by chance in approximately 10% of similar 29 year means. A reduction of this size does, of course, only represent a tiny portion of the freshwater input into the Arctic Basin as a whole, whilst there have been periods of several years when the flow of these rivers appears to have been reduced much more. It is interesting to note that the computed decreases in river runoff under the effect of economic activity as estimated by Shilklomanov (1978) already amounted in 1975 to 2% of the total runoff for the Ob (7-8 km^3yr^{-1}) and 1% for the Yenisey (5-6 km^3yr^{-1}). Interestingly the decrease was even higher for the Yenisey in the previous decade. By 1980 it was already planned to be running at 4% and 3% respectively. These are small amounts of course compared to total possible withdrawals of 100-200 km^3yr^{-1} which represent much more significant reductions, the effects of which we now consider.

Possible effects of substantial diversions

Observational, analytical and simple modelling studies

Antonov (1978) lists the direct impact of river waters on the Arctic seas to be:

(i) in the heat of river waters in facilitating the spring break up of fast ice, mainly in the estuaries and delta reaches of rivers, and in increasing the heat content of offshore waters in summer.

(ii) in their freshening effect, which promotes the earlier appearance of young ice in autumn in river mouths and offshore and creates more favourable conditions for more rapid growth of fast ice in winter.

(iii) together with excess of precipitation over evaporation, in fostering the formation of well defined gradient drainage currents (see e.g. Defant, 1961) which, in company with the corresponding atmospheric pressure field (reflecting the surface winds), determines the major features of the general system of currents and ice drift.

The main effects of a reduction in riverflow for (i) are likely to be local rather than global with a tendency to increase, rather than decrease, the ice cover in estuary and coastal regions and with consequent reduction in the period of navigability of these waters in any year. With regard to (iii) Micklin (1981) notes that a number of Soviet researchers have considered that a lower freshwater input from rivers would lead to a reduction in the gradient currents in the shelf seas and thereby in the amount of Atlantic water entering the Arctic as a whole and the shelf seas in particular. By this means the upward heat flux which results from the penetration of Atlantic water onto the shelves would be reduced so that thicker ice would result in these regions. This viewpoint is supported by a system analysis carried out by Micklin (1981) for the Kara Sea which not only bears out this hypothesis but also emphasises the complexity of the problem in the number of process linkages and feedbacks which can occur. Micklin's conclusions are qualitative because he had no basis for weighting the various linkages which he assumed to be of equal importance. The case for an increased ice cover is, however, also supported to some extent by a recent modelling study of Semtner (1984) (see below) and by a statistical analysis carried out by Holt, Kelly and Cherry (1984) who correlated annual riverflow data over the period 1966 to 1979 for the Ob, Yenisey and Lena rivers with the monthly sea ice concentration extracted from the dataset of Walsh (1979). In this context, Holt and others found areas of negative correlations (reduced riverflow associated with increased ice extent) between the flow of the Ob and Yenisey and the ice extent in the Barents Sea throughout much of the period from the mid-winter to late spring following the hydrological year (which runs from October to September) over which the riverflow was totalled and likewise between Yenisey flow and ice extent in the Kara Sea from the July to September of the following hydrological year. Correlations were considered to be significant at the 5% level. They also found a number of positive correlations, in particular between the flow of the Ob and ice extent the following autumn out to within about 200 km of the mouth of the river. They attribute this to the effect described by Antonov ((ii) above) that freshening promotes the growth of young ice in autumn. Such a correlation was not found for the other rivers, however, something which Holt and others attribute to the low spatial and temporal resolution of the data, whilst one or two substantial but more distant areas of positive correlation (e.g. between Ob flow and ice concentration along the Laptev sea coast) they attribute to chance. Holt and others signal caution in the interpretation of their results which are, nonetheless of considerable interest, tentatively indicating some sensitivity of the Arctic ice cover, at least over the shelf seas, to variations in the natural riverflow.

Because of the nature of the data available to them, the investigation of Holt and others was unable to encompass any possible effects of the changes in riverflow on ice thickness over the deeper basins of the Arctic which might arise from a change in the static stability of the upper ocean layers. Such a possibility was suggested by Aagaard and Coachman (1975) who emphasised the role of riverflow in maintaining the low values of salinity in the surface layer of the Arctic as a whole. Aagaard and others (1981) put further emphasis on the particular hydrographic structure of the upper 200m of the Arctic Ocean which, with the main pycnocline coincident with the upper halocline and a near uniform temperature profile separates the surface waters from the thermocline beneath and so largely isolates the warm Atlantic layer from the surface layer. In particular this structure forms a very effective barrier to convection brought about by cooling and freezing at the surface by preventing direct entrainment of thermocline water into the surface layers except, as we have seen, in the shelf regions where topographic effects serve to force the Atlantic water into more direct contact with the surface layers. As noted previously, the structure of the subsurface layers has been linked, in particular, by Aagaard and others (1981) to processes on the shelves and they point to the possibility that, by reducing riverflow and hence the extent to which shelf waters are freshened in summer, higher salinity levels may be achieved in these waters during autumn and winter freezing. This could have the result that the shelf waters thought to feed the halocline might descend deeper into the Arctic Basin thereby changing the structure of the upper layers towards a thinning of the pycnocline and placing the Atlantic water more in direct contact with the surface mixed layer. The mixed layer would then be more susceptible to direct heat transfer from the Atlantic water due to turbulent entrainment, with direct consequences for reducing the ice distribution and thickness.

A reduction in the flow of the rivers may, of course, also act to reduce the strength of the halocline by direct increase in the salinity of the surface and near surface waters themselves. Such a possibility was discussed by Aagaard and Coachman (1975) who concluded, on the basis of a simple budget calculation, that removal of the Arctic pycnoline as a whole within 1 decade (the typical 'residence time' of Arctic surface water) would require a totally unrealistic closure of the total inflow both through the Bering Straits and from the Arctic rivers. Likewise Stigebrandt (1981) using a two-layer diagnostic model (not strictly applicable to the more complex structure which we have seen to characterise the upper layers of the Arctic Ocean) based on conservation of water volume and salt concluded, like Aagaard and Coachman (1975), that for the Arctic to become ice free would involve cutting off at least all of the river flow. Stigebrandt's model is, however, essentially only one dimensional in nature. Aagaard and Coachman (1975) note that a particularly sensitive area for the likely impact of river diversions in directly reducing the salinity of the upper layers is the region of the southern Eurasian Basin. This region shows only a small accumulation of fresh water within the surface layers which overlie the warmest and most saline water in the Arctic, though just what reduction in riverflow could affect a significant increase in surface salinities and reduction in ice thickness is, as yet, largely unknown.

A study with an Arctic Ocean general circulation model

An attempt to assess the stability changes which might result from substantial river diversions has been made more recently in the only major, albeit preliminary, circulation modelling study to date. This has been carried out by Semtner (1984) who used a 13 level primitive equation ocean circulation model, set up to cover the Arctic Basin and Greenland/Norwegian Seas on a 100 km grid, to simulate the Arctic circulation and temperature and salinity regime (see also Semtner, 1976) and to test its sensitivity to variations in the input of those rivers likely to be involved in the Soviet diversion schemes. In the model, lateral boundary conditions used at the southernmost boundary of the Greenland/Norwegian Seas were such that some response there to changes in the temperature and salinity structure in the interior could be met via the enhanced horizontal diffusivities for heat and salt specified. A constant forcing was applied across the air/sea interface appropriate to annual mean conditions (geographically varying wind stress and specified net heat flux, dependent on diagnosed ice cover). As Semtner points out, this puts beyond the scope of the study any modelling of the seasonal effects we have seen are likely to be important. Use of annual mean forcing also precludes use of an explicit sea ice model. Instead, sea ice was presumed to exist when the surface water temperature fell below -1.5C, when surface heat exchanges were consequently reduced to very low values. No direct conclusions could be reached with regard to the influence of the river diversions on sea ice thickness or compactness, therefore, though some inferences could be made. Nevertheless, some interesting results emerged. A control integration (with the observed riverflow specified) was seen to give a reasonable simulation of the hydrographic and current structure of the Arctic Ocean, though, as Semtner notes, the halocline depth scale is greater than observed. Thus whilst the surface and Atlantic layer salinities are fairly realistically predicted, the average salinity over the upper few hundred metres of the allegedly crucial area of the Eurasian Basin is much too low. This is illustrated by Figure 7 which compares the composite salinity profile for the southern

Eurasian Basin shown by Aagaard and Coachman (1975) (see Figure 2(b)) with a profile derived from Semtner's Figure 4 for much the same region. The point here is that a much greater reduction in the input of fresh water is necessary to wipe out the halocline for Semtners profile than for Aagaard and Coachman's (1975) one. It is the shallowness of the upper mixed layer and halocline as well as the weakness of the latter which forms the basis for Aagaard and Coachman's argument. The model, may, therefore, be undersensitive to changes in the river input, a possibility which Semtner attempted to allow for by running an integration for the case in which the total flow of the Northern Dvina, Pechora, Yenisey and Ob rivers was diverted in addition to a run in which diversions from these rivers totalling some one third of their flow were specified. The total diversion case certainly showed some destabilisation north of Spitzbergen but even so this was not enough to bring about the onset of deep convection over the Eurasian basin.. We note here that careful examination of Semtner's figures shows the model temperature structure to be also untypical of that observed in that the thermocline lies immediately below the shallow upper mixed layer. This should go to increasing the sensitivity of the model somewhat in that the source of warmer water is more readily available. The reasons for the unrepresentative halocline structure may well be related to the lack of representation of seasonal effects including explicit representation of sea ice, as Semtner notes.

One important indication of the model is its adjustment of the current structure to confine most of the changes in the surface salinities to the region of the Kara and Barents Seas, into which the affected rivers flow. The salinities of these seas also increased at depth as a result of the diversions, though by a somewhat smaller amount than at the surface but with a tendency for outflow through the Fram Strait and an increase in convective activity in the ice-free Greenland Sea. Because of the anomalies in the current pattern there was a general decrease in the temperature of the thermocline and Atlantic layer waters extending up over the shelves and Semtner speculates that this might lead to a general increase, rather than decrease in ice thickness. Certainly the ice margin, as diagnosed, shows little marked change over any of the runs. Upwelling, brought about by the changes in the currents occurred over a small region of the Eurasian Basin, leading ot higher surface temperatures there and this may imply some decrease in ice thickness in this limited region. More definite answers to those questions can, as Semtner emphasises, come only from more detailed investigations including a coupled ocean - sea ice model run with seasonal forcing. Indeed, more recently Semtner (personal communication) has run his model with these effects included, but results are yet to be made available.

Concluding remarks

As yet, Soviet plans to divert their northward flowing rivers are uncertain, but present indications are that initial transfers of order 25 km^3yr^{-1} are likely before the end of the century, probably from the Pechora and Northern Dvina. Diversions of this size are unlikely to have a significant impact on the sea ice of the Arctic Basin as a whole being well within the natural variability of riverflow and insignificant compared to the total input of fresh water into the Arctic Basin overall (3600 km^3 yr^{-1} from rivers alone). There is some evidence, however, that diversions comparable in size to the magnitude of the natural variability of the riverflow (say 60 km^3yr^{-1}; cf Micklin, 1981) could have a detectable effect on the ice cover in the shelf regions of the Kara and Barents Seas. Present tentative indications are for decreased sea ice concentrations in the coastal seas adjacent to the mouths of diverted rivers but increased ice concentration (and presumably thickness) in the more central and northerly parts of these seas. The former reflects the influence of reduced freshening in discouraging the formation of young sea ice in autumn and the latter the influence of increased salinity in the shelf seas in modifying the drainage currents from them, leading to reductions in the inflow of Atlantic water entering the Arctic region. Such conclusions accrue from a number of Russian authors (see Micklin, 1981); from the system analysis carried out by Micklin (1981); from correlation studies of annual riverflow with sea ice concentration (Holt and others, 1984) and from the numerical modelling studies of Semtner (1984). It must be emphasised, however, that such results are as yet very preliminary and still lack precision. Just what the impact of larger scale diversions would be remains an open question as is the question of what reduction in riverflow would be such as to produce a climatically significant change in the ice cover. Whilst a catastrophic removal of the Arctic sea ice would seem unlikely, possible changes in the limits of the ice margin are yet to be fully assessed. Certainly Semtner's (1984) preliminary results from his numerical model integrations indicate very little change in ice extent, even with the whole of the participating rivers (Northern Dvina, Pechora, Ob and Yenisey) diverted. However, careful verification of the flow and hydrographic structure, particularly in the upper layers, together with explicit modelling of the seasonal cycle, shelf processes and sea ice is necessary before results can be considered in any way reliable. The former requires in part further observational studies, both with existing data and in the field whilst the latter requires application and development of existing ocean and sea ice models which, on modern computers, have the potential to carry out useful studies to improve our understanding of Arctic Ocean circulation and structure and, in the context of Soviet river diversions, of the role of freshwater for these.

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Table I. Distribution of flow (monthly percentage) in major arctic rivers (from Mackay and Loken (1974))

	Drainage area (10 km^2)	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Mean annual flow (km^3yr^{-1})
Canada														
Mackenzie	18,425	6	6	5	4	7	13	<u>14</u>	11	10	9	8	7	340
USSR														
Indigirka	3,050	0.22	0.11	0.06	0.04	1.6	<u>30.0</u>	<u>30.5</u>	26.4	11.5	2.6	0.70	0.43	49
Kolyma	3,610	0.42	0.28	0.23	0.20	7.9	<u>38.2</u>	19.6	15.3	12.2	3.5	1.1	0.74	70
Lena	24,300	1.3	0.97	0.70	0.57	2.3	<u>37.6</u>	20.0	13.8	12.5	7.4	1.7	1.4	514
Northern Dvina	3,480	2.5	2.0	1.8	5.7	<u>34.3</u>	17.8	7.5	5.6	6.0	7.5	6.0	3.4	106
Ob-Irtysh	24,300	3.0	2.5	2.1	2.2	9.9	<u>21.9</u>	19.9	15.0	9.3	7.0	4.1	3.4	385
Onega	557	2.8	2.4	2.1	7.6	<u>34.2</u>	14.4	7.5	5.3	6.2	7.8	6.2	3.6	15
Pechora	2,480	1.6	1.2	1.1	2.0	22.6	<u>34.0</u>	10.7	5.5	6.8	8.1	3.9	2.4	106
Yenisei	24,400	2.3	2.1	2.0	1.9	14.1	<u>35.6</u>	13.1	8.8	8.4	6.9	2.9	2.3	562

Note: Months of maximum flow are underlined.

TABLE II: Arctic River Drainage

Region	Drainage (km^3yr^{-1})	Standard Deviation (km^3yr^{-1})
Asian slope to Arctic	2360	140
European USSR	360	–
Pechora (rounded up)	110	14
North America	810	–
Total	3640	–

TABLE III: Flow of the Yenisey, Ob, Pechora and Northern Dvina. The averaging period is indicated in brackets.

River	Mean flow (km^3yr^{-1})	Standard Deviation (km^3yr^{-1})
Yenisey (29 yrs)	562	40
Ob (35 yrs)	387	55
Pechora (33 yrs)	106	14
N. Dvina (35 yrs)	100	19
Yenisey + Ob + Pechora + N. Dvina (29 yrs)	1160	85

FIG 1. Map of European Russia and part of Western Siberia showing major river systems and other significant geographical features.

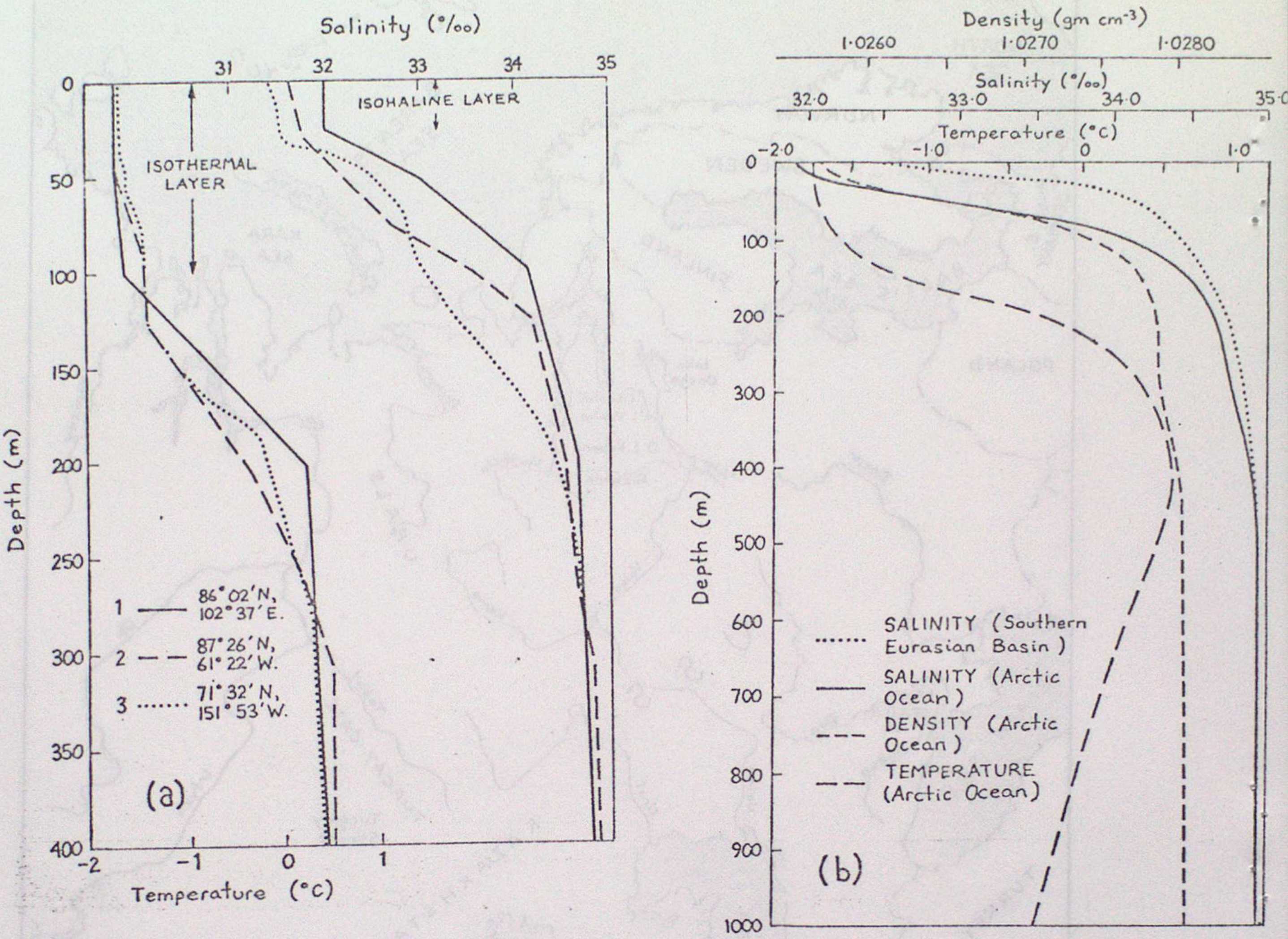

FIG 2. (a) Vertical temperature and salinity distribution in the upper 400m of the Arctic Ocean as observed during winter at the locations indicated (see also Fig 3). For details see Aagaard and others (1981).

(b) Composite vertical distributions of temperature, salinity and density (calculated for a pressure of 1 atm.) in the Arctic Ocean, and of salinity in the Southern Eurasian basin (from Aagaard and Coachman, 1975).

FIG 3. Bathymetry of the Arctic Ocean and location of geographical features. Depths are in fathoms (1 fthm = 1.829 m). The location of the stations in Figure 2(a) and the northern section of the Asian drainage slope into the Arctic Ocean (hatched) are also shown.

FIG 4. Surface (5m) salinity of the Arctic basins in summer (from AINA, 1967).

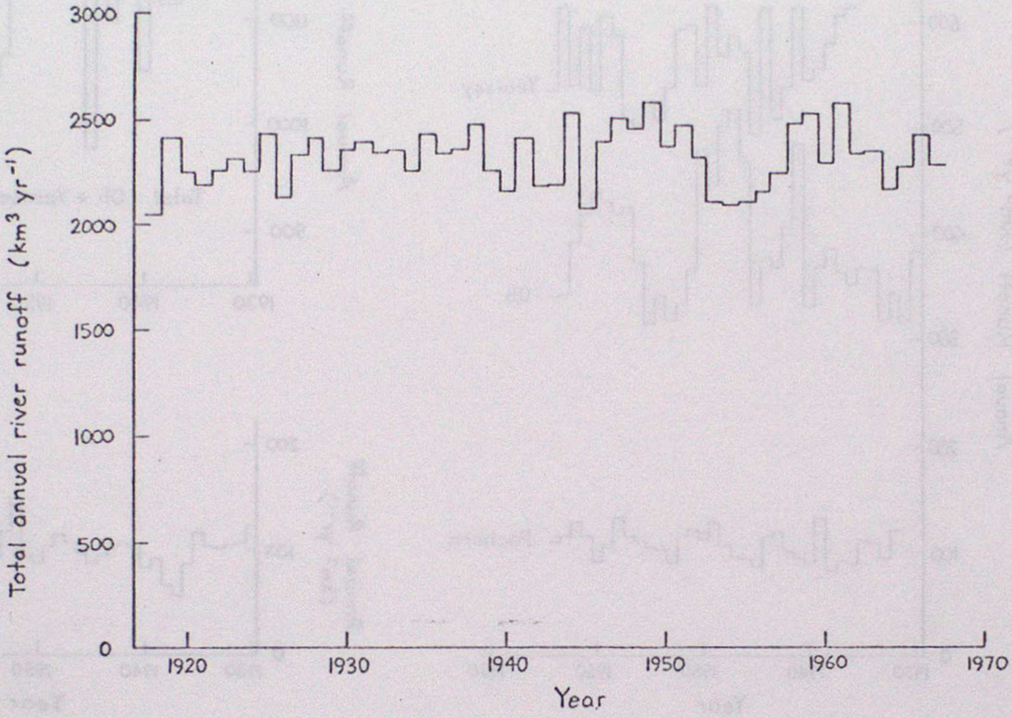

FIG 5. Total annual river runoff from the Asian drainage slope into the Arctic Ocean for the period 1918-1967. Mainland and Islands (data from Korzun and others, 1978).

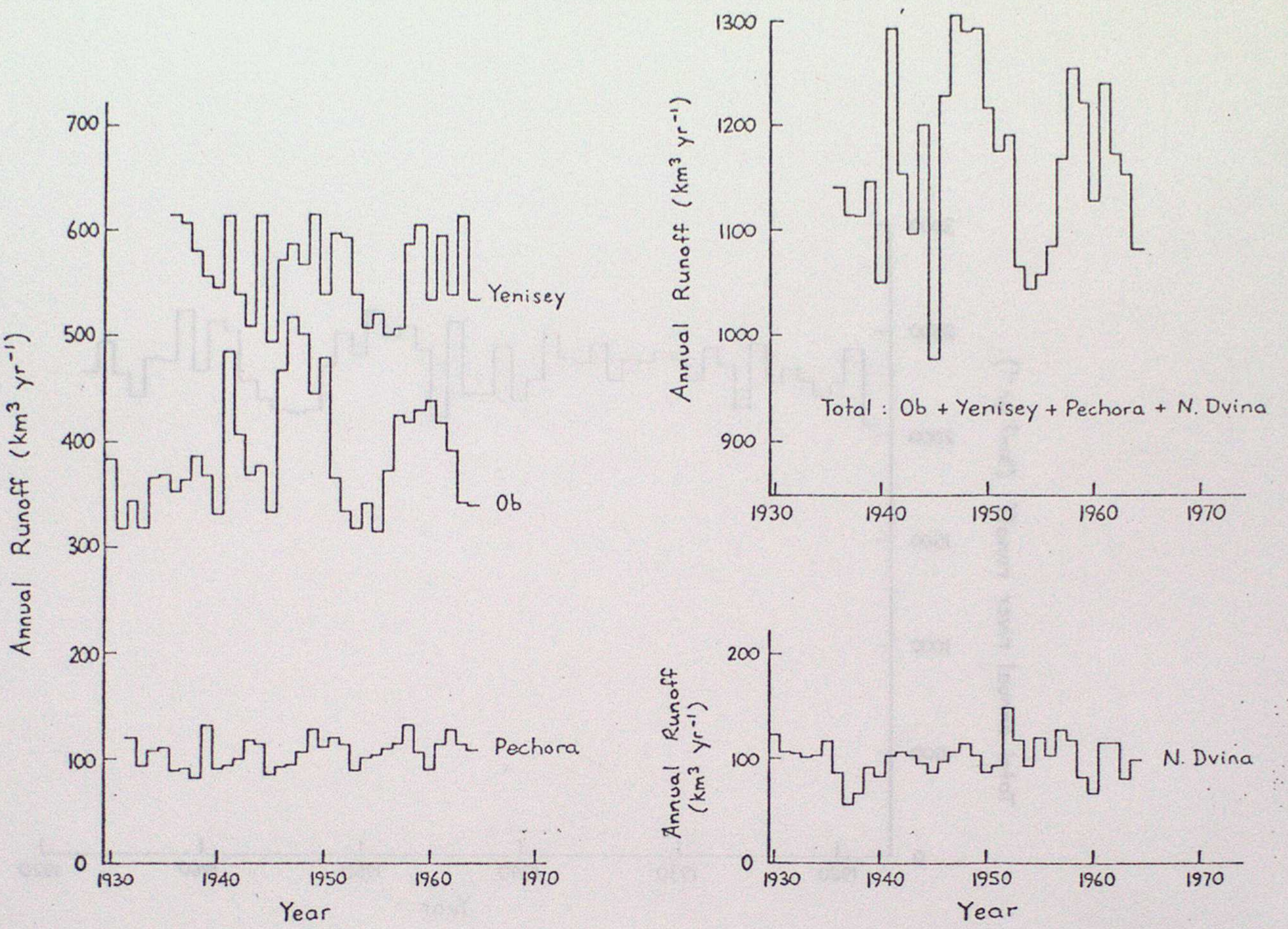

FIG 6. Annual runoff (km^3yr^{-1}) for the Yenisey, Ob and Pechora rivers (left) and Northern Dvina river (lower right) and total for all four combined (upper right) (Data from UNESCO, 1971).

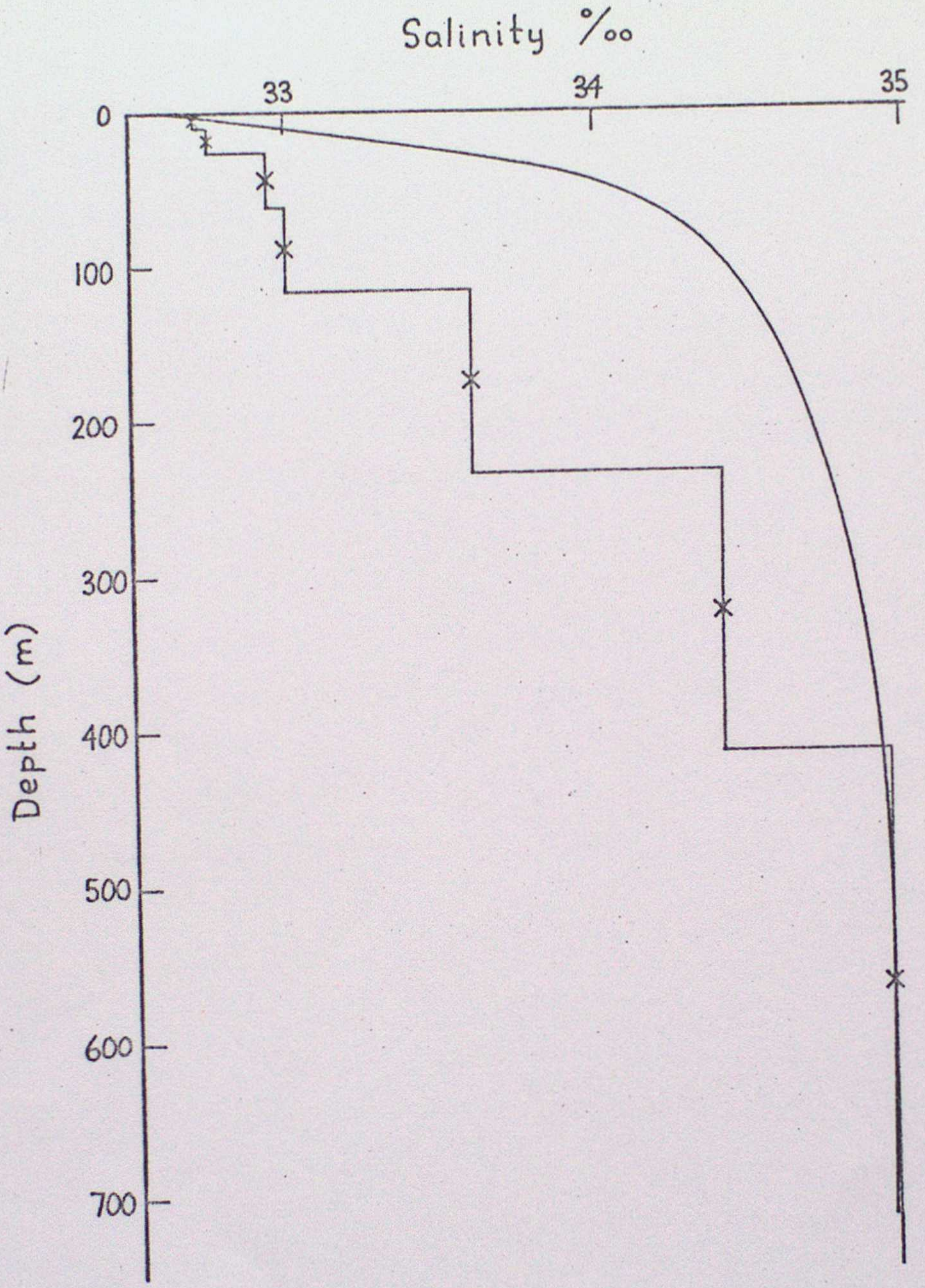

FIG 7. Comparison of the composite profile for the salinity of the southern Eurasian basin shown by Aagaard and Coachman (1975) (continuous curve) with a profile derived from Figure 4 of Semtner (1984) for much the same region. Model levels are indicated by crosses.