WATER, SALT AND HEAT BALANCE OF THE NORTH POLAR SEA AND OF THE NORWEGIAN SEA

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Summary. An attempt has been made to establish estimates of the average values of the different factors, in claiming that water, salt and heat are simultaneously in balance. In many cases the different sources diverge considerably; this may be due to true variations of the factors or to unreliable methods of determination.

For the *North Polar Sea* it appears that the major part of the heat loss (88%) is due to conduction through the ice, while the heat gain is mainly due to currents. Moderate factors are evaporation and radiation within the ice-free areas and the export of ice. Both for water and for salt balance the outflow in the East Greenland Current must be approximately equal to the sum of inflow of Atlantic and bottom water (eq. 4, 5, 6). For heat balance a relation (eq. 7) results between the different factors, of which some are first introduced as unknown terms. The values finally adopted as most probable, are given in Table 3.

For the *Norwegian Sea and the Barents Sea* it is found that the major part of the heat loss must be due to evaporation (82%), while radiation balance and the Atlantic current account for most of the heat gain. Water and salt balance give the same result, that Atlantic inflow and East Greenland outflow must be approximately equal (eq. 9, 10, 11). For heat balance (eq. 12, 13) these amounts will depend upon the evaporation, and an approximate relation is established between the volume transport of Atlantic water and the average evaporation (eq. 15). The values adopted as most probable are given in Table 4.

Introduction. The climate of Northern Europe is known to be abnormally mild; according to Schott (1942, PL. X, XX) the mean annual air-temperatures are from 5 up to 12.5°, and the mean annual sea surface temperatures up to 5° higher than the averages for the corresponding latitudes. Within the Norwegian Sea[1] these high anomalies are generally ascribed to the transport of warm Atlantic water through the Faeroe-Shetland Channel.

[1] This name will here be used according to the definition given by Helland-Hansen and Nansen (1909) for the whole of the sea between Greenland and Scandinavia.

The purpose of the present paper is to review available estimates on the exchange of water, salt and heat, for the possible establishment of complete balance sheets. This has been attempted first for the Polar Sea and then for the Norwegian Sea and the Barents Sea. A brief treatment of the problem was given by SVERDRUP for the Arctic Mediterranean as a whole (SVERDRUP, JOHNSON and FLEMING 1946, pp. 655—656). Only after publishing some preliminary results (MOSBY 1960—1961) and nearly finishing the present study I became aware of the papers by TIMOFEEV (1956, 1958, 1960), by ZAITSEV (1959, 1960) and by VOWINCKEL and ORVIG (1960).

Many of the values introduced below appear to be poorly known, and so do some of the processes involved. It is believed, however, that the results obtained may at least illustrate the orders of magnitude; and it is hoped that they may perhaps also encourage further efforts to establish more reliable data.

In later years several field investigations are known to have been carried out, from which final results have not yet been published, or the publications have not been available to the present author. Cordial thanks are due to colleagues who have kindly placed their preliminary results at my disposal.

Formulation of the problem. Consider within the sea an arbitrary volume with clearly defined limits. If the hydrographical conditions, *i.e.* the stratification of the water layers and thus also the average specific volume remain constant, the principle of conservation of mass leads to the conclusion that the net outflow of water per unit time must be zero. If the surface of the sea forms part of the limiting surface of the volume considered, the same conclusion holds good in case of a constant sea-level. In the following an estimate of mean annual values of transport shall be attempted, and our first equation may be written

$$\Sigma V = 0 \tag{1}$$

where V denotes any single average volume of water transported into or out of the area under consideration. V will here be given in $10^6\ m^3 sec^{-1}$.

If the average salinity of each volume of water transported is known, the conservation of salt my be expressed by

$$\Sigma V \cdot S = 0 \tag{2}$$

where S denotes the content of salt of the corresponding volume of water, expressed in $^0/_{00}$. The transport of salt $V \cdot S$ will then be given in 10^9 g. sec^{-1}.

Unfortunately further conservative characteristics of the waters of the ocean are rarely sufficiently well known for the establishment of other similar equations. But on the assumption that the amount of heat dissipated within the volume considered, may be neglected, the principle of conservation of energy yields a third equation, expressing that the net amount of heat transported out of the volume under consideration must be zero. Transport in this case must, of course, include any process by which heat is received or lost, such as incoming and outgoing radiation, effects of evaporation and of freezing or melting of ice. The third equation may therefore be written

$$\Sigma V \cdot t + \Sigma Q = 0 \tag{3}$$

where t denotes temperature in °C of any volume of water transported and Q denotes the amounts of heat received or lost. $V \cdot t$ will then be given approximately in 10^9 kcal. $\sec^{-1}$. It should, however, be noted that equation (3) holds good only if equation (1) is satisfied. In this way the heat *balance* may be established, but each single term $V \cdot t$ has in itself no clear significance, dependent as it will be f. inst. on the zero point of the temperature scale used.

Nevertheless, it may appear desirable to express in figures the amount of heat brought into an ocean basin by one single current. In order to do this one may think of expressing the amount of heat active to increase the total content of heat of the ocean basin, *i.e.* its average temperature. The latter being denoted by t_0, the said amount of heat will be expressed by $V(t - t_0)$. For the heat balance this definition will, of course, be of no influence, because $\Sigma V t_0 = t_0 \Sigma V = 0$. But most currents are flowing in the upper layers of the oceans, and in certain cases the deep water may be nearly isolated from the surface currents by a strong transitional layer. Slow processes may conduct an influence of the surface waters to the deeper layers, and relatively rapid fluctuations may have no perceptible effect. Against defining the heat transport of a current as $V(t - t_0)$ it may also be remarked that the amount of heat transported out of one basin will then be different from the amount transported into another basin by one and the same current. This paradox should be borne in mind when heat transport by currents is discussed.

In view of the poor knowledge of many of the values needed for a practical application, the simultaneous use of the three equations provides an aid for the evaluation and comparison of the single terms.

THE NORTH POLAR SEA. The water masses of the Polar Sea may communicate directly with those of the adjacent seas in different ways, of which the following are believed to be most important.

1. Inflow of water through the Bering Strait.
2. Outflow of water to the Barents Sea.
3. Inflow of Atlantic water from the Norwegian Sea.
4. Inflow of bottom water from the Norwegian Sea.
5. Outflow of Arctic water east of Greenland.
6. Outflow of Arctic water through the Canadian Archipelago.
7. Outflow of ice.
8. Outflow of melting water.

Further factors to be taken into consideration are

9. Fresh water received as precipitation.
10. Fresh water received as run-off from land.
11. Fresh water lost by evaporation.

and

12. Heat lost by evaporation as heat of vaporization.
13. Heat lost by evaporation as sensible heat.
14. Heat gained by radiation.
15. Heat lost by conduction through the ice.
16. Heat gained by conduction through the bottom.
17. Heat gained by the freezing of ice.
18. Heat lost by discharge of ice from North Greenland.

Each of these factors is estimated below, and the resulting values are compiled in Table 3, in the order indicated by the numbers in column 2. In view of the poor accuracy the specific heat of the water has been chosen as 1 gramme-calory per gramme and °C and the density as 1 gramme per cubic centimeter.

An estimate of the radiation balance of the Polar Sea is difficult, perhaps first of all due to the variable albedo of the ice-cover. As most of the surface is covered by ice, this difficulty has, however, been avoided by introducing the heat lost by conduction through the ice.

The poor knowledge of the Polar water and its transport, especially between Spitsbergen and Greenland, makes a reliable estimate of its temperature and salinity difficult. However, it is known that the extreme values are always very near to — 1.8° and 34.0‰, corresponding to a water produced by the freezing process. When the ice melts, a brackish or nearly fresh water of about 5‰ salinity is produced, and this is mixed with the extreme Polar water, here called Arctic water. Below the Arctic water, mixing goes on with the Atlantic water. Where the East Greenland Current passes the border between the Polar Sea and the Norwegian Sea, *i.e.* between the 80th and the 81st parallels, we may assume its waters to contain a certain, but unknown, percentage of Atlantic water as well as of melting water. We may below neglect the Atlantic water contained, as its amount must be assumed to be small as compared to the amount of Atlantic water probably flowing into the Polar Sea. And by making a rough estimate of the amounts of melting water and of ice transported, we may thus introduce a virtual transport of Arctic water, the magnitude of which will be determined later, but which can hardly be very different from the total transport of Polar water.

The values of volume transport found in literature are usually determined by dynamical computations, and their validity therefore depends on certain theoretical conditions, which do not always seem to be satisfied in nature. Here may be referred to some cases from the Sognefjord Section crossing the Atlantic Current off the Norwegian Current (SÆLEN 1959). In occupying this section twice or even four times in sequence, changes were found amounting to 100 *per cent* in the total transport and to 80 *per cent* in the transport of Atlantic water. It must be admitted that the values can hardly be considered as trustworthy and representative, when such great changes are found within a few days. In the following we shall, therefore, as far as possible avoid to rely upon values found by this method only.

1. Inflow of water through the Bering Strait. Mr. GENE L. BLOOM (1961) of the U.S. Navy Electronics Laboratory, San Diego, California, has kindly put at my disposal a summary of transport and temperature measurements in the Eastern Bering Strait in the years 1953 to 1958, based on not yet published data. The transport values were determined from electromagnetic silver-silver chloride system records. The electrode systems were calibrated by direct oceanographic current profile measurements at simultaneous stations out to 20 nautical miles by an acoustic system. The transport values are considered characteristic of the flow in the Eastern Bering Strait out to 25 miles or approximately of 55 *per cent* of the strait. They show that in winter and spring the transport is irregular and changes greatly from year to year. Thus in May 1954 a southward transport of $3.8 \cdot 10^6\,m^3sec^{-1}$ was found, while in May 1957 there was a northward flow of $3.2 \cdot 10^6\,m^3sec^{-1}$. More regular conditions prevail from August to November, and from these months the most complete sets of measurements exist. It is believed that the average monthly values given by BLOOM may be fairly representative; they are given in the second column of Table 1. The average values for the four months of August to November in the years 1953 to 1958 were: 1.70, 1.82, 2.68, 0.73, 0.67 and 1.05 respectively. For the whole year the values of Table 1 give an average transport of $0.66 \cdot 10^6\,m^3sec^{-1}$.

Table 1. *Eastern Bering Strait* (BLOOM *1953—1958 and 1953—1960*).

Month	$10^6\,m^3sec^{-1}$	°C
January	0.01	−1.61
February	0.18	−1.72
March	0.44	−1.78
April	1.07	−1.78
May	−0.23	−1.50
June	−0.23	0.39
July	0.49	4.94
August	1.83	7.18
September	1.64	5.80
October	1.22	2.94
November	1.21	−0.31
December	0.26	−1.24
Year	0.66	(3.18)

The temperature measurements were made from 1953 to 1960 by resistance thermal units at different distances from the bottom and at up to 5 miles offshore. They show a regular seasonal variation. The mean monthly values, given in the third column of Table 1, may be considered as fairly representative of the average conditions within the section. For the heat transport it is of importance that the highest temperatures occur in July to October, or mainly within the months of great northward transport.

Taking the transport values into consideration we find from Table 1 an average temperature of 3.18°C. On the assumption that these measurements are representative of the whole of the section of the Bering Strait, we should have a volume transport of $1.2 \cdot 10^6 m^3 sec^{-1}$ and a temperature of 3.2°C.

TIMOFEEV (1960, Table 24) refers the monthly values of water transport and of heat transport for 1941—1943 computed by G. A. BASKALOV. From these data the mean monthly temperatures can be found; they agree well with those of BLOOM (see Table 1 above) except for the summer months. The total transport is $1.15 \cdot 10^6 m^3 sec^{-1}$, and by taking the transport values into consideration, we find an average temperature of only 0.93°. As these values are probably based on measurements in the western part of the strait, we may adopt for the whole cross section a transport of $1.2 \cdot 10^6 m^3 sec^{-1}$ and the average temperature 2.1°. As salinities are not available, we shall adopt the value estimated by SVERDRUP (SVERDRUP, JOHNSON and FLEMING 1946, p. 655), or 32.0‰.

2. Outflow of water to the Barents Sea. From the results of the "Quest" expedition (MOSBY 1938) it is seen that the exchange of water between the Polar Sea and the Barents Sea is very moderate, but that in July 1931 there was a net transport of Atlantic water from the Polar Sea. In this case the agreement with direct current measurements was not very good, but the value derived is so small that it shall here be adopted for the sake of completeness. We may then use the values for the "extreme" Atlantic water (*loc.cit.* pp. 36—37), or $0.05 \cdot 10^6 m^3 sec^{-1}$ of 2.7° and 34.92‰.

3. Inflow of Atlantic water from the Norwegian Sea. By dynamical computations within a section from the north-west corner of Spitsbergen towards the north, occupied by the "Quest" in 1931 (MOSBY 1938, p. 34), a total transport towards the east of $1.36 \cdot 10^6 m^3 sec^{-1}$ was found in the southern part of the section. Direct current measurements were carried out for 30 hours, and the results agreed within $\pm$ 10 *per cent* of the calculated currents. Through the northern part of the section $0.58 \cdot 10^6 m^3 sec^{-1}$ was found to move towards the west. This agrees with the image of the current pattern in this area established by NANSEN (1915) and with the conception of SVERDRUP (1933, a) that the Atlantic water spreads out towards the north and gradually turns west.

When integrating within the east-going part of the current the values of speed multiplied by the corresponding temperatures and salinities, the mean values are found to be 3.25° and 35.10‰, and these values will be adopted for the water masses as they enter the Polar Sea. And from the above measurements we may expect the volume transport to be at least $1.4 \cdot 10^6 m^3 sec^{-1}$.

The latter value is considerably lower than the values reported by Russian authors. Thus TIMOFEEV (1958) gives 94'082 $km^3 year^{-1}$ or $3.0 \cdot 10^6 m^3 sec^{-1}$, referring to an earlier paper, not available to the present author. LAKTIONOV (1959) gives 102'500 $km^3 year^{-1}$ or $3.2 \cdot 10^6 m^3 sec^{-1}$. In a communication to the Special IGY Meeting of the ICES in 1959 (ZAITSEV, FEDOSOV, ILJINA and ERMACHENKO 1959, ZAITSEV 1960) the total transport between Spitsbergen and Greenland is reported to be 16.25 km^3hour^{-1}, or $4.5 \cdot 10^6 m^3sec^{-1}$. And TRESHNIKOV (1960) gives for the Atlantic water 128'500 km^3year^{-1}, or $4.1 \cdot 10^6 m^3sec^{-1}$. Without having at hand the complete material of data on which these results are based, it is not possible to explain the considerable divergencies. However, if the Russian results are based on dynamical computations, sections in different latitudes may be expected to have led to rather different results. This is understood from our knowledge of the general circulation in the area: from the Atlantic Current running northwards off the western coast of Spitsbergen, one part branches off towards the east, while farther south off Spitsbergen considerable volumes of water are known to branch off or spread out towards the north-west, west, and south and thus to remain within or to return to the Norwegian Sea.

In view of the great discrepancies, we may as a first step denote the transport of Atlantic water by A.

4. Inflow of bottom water from the Norwegian Sea. According to NANSEN (1909, 1912) bottom water is formed in the Norwegian Sea and flows across a ridge, later called the Nansen Ridge, between Greenland and Spitsbergen at an approximate saddle depth of some 1500 m. Numerous soundings from Russian expeditions have revealed a narrow channel through the Nansen Ridge, reaching more than 3000 m of depth (BALAKSHIN 1959). In spite of this the deep water of the Polar Sea nowhere shows temperatures below $-0.9°$, or values usually found within the Norwegian Sea at depths of about 1500 m. The explanation of this will now probably have to be sought in the dynamical equilibrium of the water masses within the northern part of the Norwegian Sea, by which the coldest bottom water is kept away from the areas immediately south of the Nansen Ridge.

For our present study the discovery by Russian expeditions of a ridge across the Polar Sea from Greenland in about 50°W towards the shelf in about 145°E is of importance. The saddle depth seems to be nowhere greater than 1500 m (TRECHNIKOV 1960), and this ridge therefore represents a barrier to the inflowing Norwegian Sea deep water. On the Spitsbergen side of the ridge the temperature of the deep water may be about $-0.85°$ or perhaps even slightly lower, while on the Beaufort side the lowest temperature is hardly below $-0.45°$. The general features are clearly illustrated on PL. I, Fig. 2 of TRESHNIKOV's paper (1960). A rough evaluation indicates a horizontal extension of the Spitsbergen basin of the Polar Sea at 1500 m depth of nearly $1.5 \cdot 10^6 km^2$, or nearly equal to the extension of the Norwegian Sea at the same level.

Whenever bottom water is formed, the original water layers are lifted into higher levels. Recent investigations (MOSBY 1959) indicate that in the Norwegian Sea the deep water is in this way moved vertically upwards at an average rate of some 25 m per year. According to what has been said above a similar vertical motion must be expected also within the Spitsbergen basin. The average production of bottom water may, therefore, amount to a volume of the order of $2 \cdot 1.5 \cdot 10^{12} \cdot 25 m^3year^{-1}$ or about $2.4 \cdot 10^6 m^3sec^{-1}$, of which one-half must flow into the Polar Sea across the Nansen

Ridge. A value of the inflow of bottom water into the Polar Sea of $1.2 \cdot 10^6\ m^3sec^{-1}$ may thus be expected; but as it is a computed, not an observed value, its influence on our final conclusion must be considered separately later. In Table 2 we may introduce the characteristics $-0.9°$ and $34.92‰$, but as a first step the transport will be denoted by B.

5. Outflow of Arctic water east of Greenland. This part of the East Greenland Polar Current is poorly known as yet. To form a first, rough idea, let us remember that the distance between Spitsbergen and Greenland in 80—81°N is 500—600 km. In the east the Atlantic Current flows towards the south, and in the middle one part of the Atlantic water branches off towards the west to return to the south. A reasonable guess would be that the Polar Current may in these latitudes hardly be wider than one-third of the 500—600 km, or about 200 km. The outflowing water, due to its relatively low salinity, is light and forms a surface current. Judging from measurements farther south, as well as within the Polar Sea, the average vertical extension of this current must probably be less than 100 m. If in its eastern part the current has a speed of 20 $cm \cdot sec^{-1}$ and if the speed decreases gradually towards the coast, one may perhaps expect an average speed of 10—15 $cm \cdot sec^{-1}$. This would give a total transport of from 2 to $3 \cdot 10^6\ m^3sec^{-1}$. As a first step we may introduce a value $G = 0.2 \cdot c \cdot 10^6\ m^3sec^{-1}$, where c denotes the average speed in $cm \cdot sec^{-1}$.

Russian measurements are known to have been carried out in the region, but the results have not yet been available to the present author. However, TRESHNIKOV in his paper (1960) reports that the East Greenland Current transports a total amount of 161'500 km^3year^{-1} or $5.1 \cdot 10^6\ m^3sec^{-1}$. ZAITSEV (1960) reports for the total transport between Spitsbergen and Greenland 16.25 km^3hour^{-1}, or $4.5 \cdot 10^6\ m^3sec^{-1}$.

Again the discrepancies are great, and as this current will appear below to be one of the most important factors, it shall, as a first step, be denoted by G.

As explained above, the water transported by the East Greenland Polar Current consists of a mixture of extreme Polar water of about $-1.8°$ and about $34‰$, and the brackish water originating near the surface through the melting of ice. The content of salt in the Polar ice is usually about $5‰$ (MALMGREN 1927), to which would correspond a melting point of about $-0.3°$. Observations farther south have shown the frequent occurrence of Polar water of $-1.8°$ and $34.0‰$ (JAKHELLN 1936), covered by a surface water of salinities down to 31 or even $30‰$, the temperature of which may be variable in summer.

Information on the conditions in the Polar Sea is obtained from PL. XV in NANSEN's paper (1902), from which the average depths of the isohalines 34 and $32‰$ have been found to be 75 and 30 m respectively. When neglecting a few extraordinarily low salinities observed at the surface, the average surface salinity is found to be about or slightly above $30‰$, indicating a roughly linear increase of the salinity with depth from the surface to about 75 m. The average salinity of the layer is about $32.2‰$, which would result if a layer of 71 m of a salinity of $34‰$ be diluted by the addition of 4 m of fresh water.

During the last part of the drift of the "Fram", in the Spitsbergen area, the average surface salinity was above $32^0/_{00}$, while $34^0/_{00}$ was found at 60 m depth. The surface layer might here result from a 60 m layer of $34^0/_{00}$ by addition of less than 2 m of fresh water.

For our present purpose we may, therefore, distinguish between the transport of ice, of fresh water (both estimated below), and of Arctic water of a salinity of $34^0/_{00}$ and of a temperature of $-1.8°$.

6. Outflow of Arctic water through the Canadian Archipelago. Extensive investigations have been carried out in later years, and some of the preliminary results have been issued. Some of them are based on dynamical computations, others also on direct current measurements. According to COLLIN (1960) "a summary of recent information indicates that the total eastward transport of water from the Arctic Ocean through the Canadian Arctic Archipelago is of the order of 40'000 $km^3 year^{-1}$ ($1.26 \cdot 10^6\ m^3 sec^{-1}$). Of this total approximately 16'000 $km^3 year^{-1}$ enters Baffin Bay through Smith Sound. The remainder leaves the Arctic Ocean by way of Jones Sound and Hudson Strait." This is somewhat less than the average southward transport through the Davis Strait of $1.5 \cdot 10^6\ m^3 sec^{-1}$ computed by SMITH, SOULE and MOSBY (1937) on the basis of five different sections.

Russian authors give different values, thus TIMOFEEV (1956) $1.0 \cdot 10^6\ m^3 sec^{-1}$, TRESHNIKOV (1959) $0.4 \cdot 10^6\ m^3 sec^{-1}$, ZAITSEV (1960) 7.1 to $7.5 \cdot 10^6\ m^3 sec^{-1}$ and ZAITSEV, FEDOSOV, ILJINA and ERMACHENKO (1959) 7.0 to $7.5 \cdot 10^6\ m^3 sec^{-1}$.

The best established value seems to be that reported by COLLIN, and for the present study we may adopt his value, approximated to $1.2 \cdot 10^6\ m^3 sec^{-1}$. As explained above for the East Greenland Polar Current, also this outflow shall here be considered as mainly consisting of Arctic water of $-1.8°$ and $34.0^0/_{00}$ to an amount of $1.1 \cdot 10^6\ m^3 sec^{-1}$, while the rest is supposed to be ice and melting water.

7. Outflow of ice. This may be expressed as $g = 0.002 \cdot cd\ 10^6\ m^3 sec^{-1}$, if the horizontal extension is assumed to be the same as for the East Greenland Current or 200 km, c is the average speed in $cm \cdot sec^{-1}$ and d is the average thickness of the ice in metres.

According to MALMGREN (1927) the mean annual temperature of the ice investigated on the "Maud" expedition increased from $-15.7°$ at the surface to $-4.82°$ at 2 m depth. For ice of 2 m thickness we may thus adopt an average temperature of $-10°C$.

According to TIMOFEEV (1958) this transport of ice has been studied by P. A. GORDIENKO and D. B. KARELIN (1945, not available to the present author): "For the period 1933 through 1944 on an average 1'036'000 km^2 of ice were removed annually". Assuming the ice to be 3 m thick on the average, TIMOFEEV arrives at a transport of 3'108 km^3 of ice per year, or nearly $0.1 \cdot 10^6\ m^3 sec^{-1}$.

The outflow of ice through the Canadian Archipelago must be expected to be relatively much smaller than that carried by the East Greenland Current, because the

Canadian Sounds are so narrow in comparison with the sea east of Greenland. In our calculations we shall, therefore, consider this ice to be included in the above expression for the East Greenland Current.

8. Outflow of melting water. This is of minor importance and may be estimated for the East Greenland Current and for the currents through the Canadian Sounds at about $0.1 \cdot 10^6\, m^3sec^{-1}$ of a temperature of some $-0.3°C$.

9. Fresh water received as precipitation. Measurements from the "Fram" expedition (MOHN 1907) gave an average of 74.7 mm per year, those from Gjøahavn (GRAARUD 1932) 109.8 mm per year, and those from the "Maud" expedition (SVERDRUP 1933, b) 102.5 mm per year. A probable average value may therefore be about 100 mm per year, and the amount received for the total of nearly $10 \cdot 10^6\, km^2$ of the Polar Sea will then be about $0.03 \cdot 10^6\, m^3sec^{-1}$. This is somewhat more than the value given by TRESHNIKOV (1959) of 700 km^3 $year^{-1}$, or $0.022 \cdot 10^6\, m^3sec^{-1}$.

Most of the precipitation must reach the surface as snow, and will, in the end, add to the water balance. Although its temperature is certainly low, it cannot influence our heat balance, as this is to be based on the conduction of heat through the ice.

10. Fresh water received as run-off from land. This may be roughly estimated from available climatological maps. For the whole of the drainage area surrounding the Polar Sea, the Oxford Advanced Atlas (BARTHOLOMEW 1936) gives by rough computations about $0.1 \cdot 10^6\, m^3sec^{-1}$, no loss by evaporation from land being subtracted. ZUBOV (1940) reported the run-off to be $0.16 \cdot 10^6\, m^3sec^{-1}$. According to VOWINCKEL and ORVIG (1961) a reasonable value, based on more recent investigations, would be 3'767 km^3 $year^{-1}$, or $0.12 \cdot 10^6\, m^3sec^{-1}$, of an average temperature of nearly 5°C.

11. Fresh water lost by evaporation. When computing as explained below, using the surface temperature of the ice instead of that of the water, we find a total "evaporation" of about 4.2 cm $year^{-1}$, corresponding to a total loss of water of $0.013 \cdot 10^6\, m^3sec^{-1}$. When in summer the surface temperature of the ice becomes zero, water will certainly evaporate from the wet ice and from the ponds of water accumulating on the ice-floes. For our study of the heat balance this need not be taken into consideration, since our study is based upon the heat loss by conduction through the ice-cover. And from the values of evaporation from the ice-free areas computed below in Table 2 it may be estimated that for the water balance the total amount of water evaporating from ice and open water is hardly more than some $0.02 \cdot 10^6\, m^3sec^{-1}$.

For the heat balance the evaporation from the ice-free areas may, however, be of considerable importance. For these areas the heat balance of the sea-surface has been studied in Table 2, in which radiation and evaporation have been evaluated for the months April to September, on the basis of measurements from the "Maud" expedition. The total radiation income values in column 1 are taken from MOSBY (1932, Table 20, p. 33). Due to multiple reflection between the clouds and the snow-cover the ice cannot be assumed to have received the whole of this energy; but for the open lanes of water this will probably be true. For the sea-surface temperatures above − 1.8° the highest values of the outgoing, "nocturnal" radiation measured on the "Maud" expedition agree well with the formula deducted by ÅNGSTRØM from measurements in more southern latitudes (MOSBY 1932, p. 59). When taking the cloudiness (*loc.cit.* Table 19, p. 32) into consideration, we find the values of the effective outgoing radiation in column 2. After reduction by 8% due to reflection, the resulting radiation balance is found to be as given in column 3.

Table 2. *Heat balance of the ice-free areas in kcal. cm⁻²month⁻¹.*

Month	Radiation			Evaporation				Heat balance
	In	Out	Balance	cm	Vapour	Sens	Total	
April	11.4	6.4	4.4	6.9	4.1	11.6	15.7	−11.3
May	15.8	3.3	11.5	5.8	3.4	7.7	11.1	0.4
June	16.7	2.5	12.9	0.8	0.5	0.4	0.9	12.0
July	15.6	2.1	12.4	0.8	0.5	0.3	0.8	11.6
August	9.3	1.9	6.8	0.8	0.5	0.3	0.8	6.0
September	4.7	2.2	2.2	4.3	2.6	4.9	7.5	−5.3

Evaporation has been computed from the conventional formula $E = 0.142\,(e_s - e_a)\,W$, where E denotes evaporation in mm · day^{-1}, e_s the water vapour pressure in saturated air at the temperature of the sea surface, e_a the observed water vapour pressure of the air, both in mb, and W the wind speed in m · sec^{-1} (see SVERDRUP 1951). The sea-surface temperatures used are those for 1923 (SVERDRUP 1929, p. 65), the air-temperatures also those for 1923 (SVERDRUP 1933, b, p. 80), the water vapour pressures those for 1922—24 (SVERDRUP 1933, b, Fig. 100, p. 253) and the wind speeds those for 1923 —24 (SVERDRUP 1933, b, Table 88, p. 214). The heat of vaporization was taken as 560 gcal · cm^{-3} and the ratio between the heat conveyed to the atmosphere by convection (the "sensible" heat) and the heat of vaporization was taken to be the Bowen ratio, or $0.65(t_s - t_a)/(e_s - e_a)$, where t_s and t_a denote the temperatures of the sea-surface and of the air respectively.

It is seen from Table 2, last column, that in April and September the net radiation income is not sufficient for the evaporation. This must mean that in April, when the water is cold and nearly unstable, there will be a tendency towards the formation of ice, while in September the uppermost surface layer of about 3 m thickness will be cooled. When being cooled from 0.8° in July to − 1.6° in winter this layer gives off 0.7 kcal · cm^{-2}, and this amount will therefore have to be subtracted on the heat balance sheet of the water.

As explained above there will, in winter, be a tendency towards ice-formation on the lanes of open water, and for the water balance we may estimate the amount evaporated for the months May to August in Table 2, column 4 to be valid over 5 *per cent* of the Polar Sea to be 8.2 cm, or about $0.004 \cdot 10^6 \ m^3 sec^{-1}$. This amount can be neglected for the water balance.

12. The heat lost by evaporation as heat of vaporization is found from Table 2, column 5, for the months May to August to be $4.9 \ kcal \cdot cm^{-2}$. The areas of open water, however, are difficult to estimate; they may be expected to cover some 5 to 10 *per cent* of the Polar Sea in summer. When denoting by p the percentage of the total area that is ice-free in summer, the heat of vaporization may be expressed as $0.16 \cdot p \ 10^9 \ kcal \cdot sec^{-1}$.

13. The heat lost in connection with the evaporation by convection to the atmosphere as "sensible heat" will then according to Table 2, column 6, amount to $8.7 \ kcal \cdot cm^{-2}$ or $0.27 \cdot p \cdot 10^9 \ kcal \cdot sec^{-1}$.

14. The heat gained by radiation is found from Table 2, column 3 to be, for May to August, $43.6 \ kcal \cdot cm^{-2}$. From this amount must be subtracted the $0.7 \ kcal \cdot cm^{-2}$ stored in the surface water in summer (see above). For p *per cent* of the Polar Sea the resulting $42.9 \ kcal \cdot cm^{-2}$ will correspond to $1.36 \cdot p \cdot 10^9 \ kcal \cdot sec^{-1}$.

15. The heat lost by conduction through the ice was determined on the "Maud" expedition (MALMGREN 1927) to be $7.67 \ kcal \cdot cm^{-2}$ for the months of September to April. In May and June heat was conducted downwards into the ice and in July and August heat received from above was used for the melting of ice. Thus the heat balance of the water is affected only by the loss of heat by conduction in September to April, to an amount of $24.3 \cdot 10^9 \ kcal \cdot sec^{-1}$, when neglecting the $p \approx 5$ *per cent* of the area which is free of ice, mainly in the summer months.

Russian expeditions farther north, at about 85°N, have found lower values. LAIKHTMAN (1959) gives $3.46 \ kcal \cdot cm^{-2}$ for September to April and $3.62 \ kcal \cdot cm^{-2}$ for September to May, but concludes that "it is quite probable that there exists a constant heat flux from the ocean depth to the surface, equal to $3.0 - 5.0 \ kcal \cdot cm^{-2} \ year^{-1}$." BESPALOV (1959) found a decrease of the heat flux through the ice at station North Pole 5 from 1.34 to $1.01 \ gcal \cdot cm^{-2} hour^{-1}$ during the period from 5. November 1955 to 1. March 1956, and he found the heat flow to the snow surface to vary between 0.54 and $1.61 \ gcal \cdot cm^{-2} hour^{-1}$, the average for the mentioned period being $1.25 \ gcal \cdot cm^{-2} hour^{-1}$. Thus the highest annual value possible also according to BESPALOV seems to be about $5.0 \ kcal \cdot cm^{-2}$, or for the total area of 10 mill. km^2 of the Polar Sea $15.9 \cdot 10^9 \ kcal \cdot sec^{-1}$, *i.e.* hardly two-thirds of the "Maud" value. Due to its importance, the heat loss by conduction through the ice will first be considered as unknown and be denoted by C.

16. Heat gained by conduction through the bottom. According to measurements by REVELLE and MAXWELL (1952) and by von HERZEN (1959) this may be estimated to be about $1 \cdot 10^{-6}$ gcal $\cdot$ cm^{-2}sec^{-1}. When applied to the Polar Sea this gives a heat gain of the order of $0.1 \cdot 10^9$ kcal $\cdot$ sec^{-1}.

17. Heat gained by the freezing of ice. The transport of the East Greenland Current was estimated above to be $G = 0.2 \cdot c \cdot 10^6$ m^{3}sec^{-1}, c denoting the average speed of the current in cm $\cdot$ sec^{-1}. Denoting the average thickness of the ice by d metres, the transport of ice may then be estimated as $g = 0.002 \cdot c \cdot d \cdot 10^6$ m^{3}sec^{-1}. Due to the importance for the heat balance, our first step shall be to introduce this expression, which leads to a heat gain of $0.16 \cdot c \cdot d \cdot 10^9$ kcal $\cdot$ sec^{-1}, when the heat of fusion is taken as 80 gcal $\cdot$ cm^{-3}.

18. Heat lost by discharge of ice from North Greenland. The amount of this ice has been estimated by BAUER (1954) as 10 km^{3}year^{-1} or $0.0003 \cdot 10^6$ m^{3}sec^{-1}, which is insignificant as far as the water balance is concerned. Its heat of fusion will amount to $0.024 \cdot 10^9$ kcal $\cdot$ sec^{-1}. As its temperature is certainly low, we may adopt for this heat loss a total of $0.03 \cdot 10^9$ kcal $\cdot$ sec^{-1}.

A compiltation of the estimates established above is given in Table 3. When introducing these values into equations (1), (2) and (3), we obtain for water balance

(4) $$G = 0.98 \cdot (A + B) + 0.08$$

and for salt balance

(5) $$G = 1.03 \cdot (A + B) - 0.02$$

These equations are approximately identical and may be replaced by

(6) $$G \approx A + B$$

For heat balance we must have

(7) $$3.25 \cdot A - 0.9 \cdot B + (1.8 + 0.9 \cdot d) \cdot G = C - 5.06 - 0.93 \cdot p$$

or by elimination of G

(8) $$(5.05 + 0.9 \cdot d) \cdot A + 0.9 \cdot (1 + d) \cdot B \approx C - 5.06 - 0.93 \cdot p$$

Reasonable estimates of the average thickness of the ice that is carried out of the Polar Sea, may vary from $d = 2$ to $d = 3$ m. When introducing $A = B = 0$, we find for $C = 24.3$ and 15.9 the values $p = 21$ and $p = 12$ respectively, indicating that with such ice-free areas no transport of Atlantic nor of East Greenland water would be needed. The true, alternative values of p must, therefore, be expected to be lower. But even with $p = 0$ we find: for $d = 2$ that $C_2 = 5.06 + 6.85 \cdot A + 2.7 \cdot B$, and for $d = 3$ that $C_3 = 5.06 + 7.75 \cdot A + 3.6 \cdot B$. For $A = 1.4$ this gives $C_2 = 14.65 + 2.7 \cdot B$ and $C_3 = 15.91 + 3.6 \cdot B$, indicating that even without any transport of bottom water, *i.e.* for $B = 0$, C cannot be below 14.

But we know that some ice-free areas exist; as a reasonable value we may adopt

$p = 5$, the average value reported by TIMOFEEV (1958) "according to the data of ice charts for the central part of the Arctic Basin". As a possible minimum of the ice-thickness we may adopt $d = 2$. As long as bottom water is hardly known to be formed within the Polar Sea itself, an inflow of bottom water from the Norwegian Sea must be considered as necessary to maintain the vertical thermal stratification. The value $B = 1.2$, estimated above on the basis of computations only, is certainly not very reliable, and may be too high. Adopting the lowest estimate of the transport of Atlantic water, $A = 1.4$, we have for $p = 5$ and $d = 2$: $C = 19.3 + 2.7 \cdot B$, from which it is seen that on these assumptions $C = 19.3$, or near to the average of the values found for 75°N (24.3) and the highest probable value found for 85°N (15.9). As a compromise we may introduce $B = 0.6$, and accordingly $C = 20.9$.

As seen from the above discussion, a reliable balance sheet for the Polar Sea cannot be established on the basis of the available estimates. But it appears that a consistent picture may be obtained in adopting $p = 5$, $d = 2$, $A = 1.4$, $B = 0.6$, $G = A + B = 2.0$ and $C = 20.9$, and these values have been introduced in Table 3.

Table 3. *Water, salt and heat balance of the North Polar Sea.*

		$10^6 m^3 sec^{-1}$	‰	°C	10^9 kcal.sec^{-1}	
Inflow.						
Bering Strait	1	1.2	32.0	2.1	[2.52]	
Norwegian Sea Atlantic water	3	A → [1.4]	35.10	3.25	[4.55]	
Norwegian Sea bottom water	4	B → [0.6]	34.92	−0.9		[−0.54]
Precipitation	9	0.03	0.0	0.0	[0.00]	
Run-off	10	0.12	0.0	5.0	[0.60]	
Outflow						
East Greenland Current	5	− G → [−2.0]	34.0	−1.8	[3.60]	
Canadian Archipelago	6	−1.1	34.0	−1.8	[1.98]	
Ice	7	− g → −0.04	5.0	−10.0	[0.40]	
Melting water	8	−0.1	0.0	−0.3		[−0.03]
Atlantic water to Barents Sea	2	−0.05	34.92	2.7		[−0.14]
Evaporation.						
Fresh water loss	11	−0.02	0.0			
Heat loss, vaporization	12					−0.80
Heat loss, "sensible" heat	13					−1.35
Radiation, net gain	14				6.80	
Conduction through ice	15					− C → −20.9
Conduction through bottom	16				0.1	
Freezing of ice	17				3.20	
Glacier ice	18					−0.03

G = 0.2 . c, g = 0.002 . c . d and freezing of ice = 0.16 . c . d for c = 10 cm . sec^{-1} and d = 2 m. Vaporization = 0.16 . p, sensible heat = 0.27 . p and radiation = 1.36 . p for p = 5%. Heating by currents in [] based on average Polar Sea water temperature = 0.00°C.

No attempt shall be made here to determine the average temperature of the waters of the Polar Sea; a rough guess might be about $-0.5°C$. As the true value can hardly be too far from zero, the amounts of heat transported by currents have, however, been computed on the arbitrary, but apparently often used assumption of $t_0 = 0.0°C$; they are given in Table 3 in parentheses.

As $G = 0.2 \cdot c$, it is seen that the values adopted above correspond to an average speed of the East Greenland Current of $c = 10$ cm $\cdot$ sec^{-1}, in good agreement with the estimates quoted by KOCH (1945). And the heat transport by Atlantic water of $4.55 \cdot 10^9$ kcal $\cdot$ sec^{-1} agrees well with TIMOFEEV (1958), who reports that $214'357 \cdot 10^{12}$ kcal $\cdot$ year^{-1} is brought into the Polar Sea and that $72'881 \cdot 10^{12}$ kcal $\cdot$ year^{-1} is brought out again, both by Atlantic water; the difference is seen to be equivalent to $4.48 \cdot 10^9$ kcal $\cdot$ sec^{-1}.

Discussion. Although many of the values given in Table 3 are questionable, it is hoped that the general picture brought out is not too wrong. It is then seen that of the total loss of heat from the Polar Sea of $23.8 \cdot 10^9$ kcal $\cdot$ sec^{-1} nearly 90% is conducted through the ice cover. The loss is covered in part (nearly 30%) by net radiation income to the estimated 5% of the surface which is ice-free in summer. Part of this gain is lost by evaporation, making the net gain 20%. The rest of the heat gain is mainly due to currents and the melting of ice outside the Polar Sea. It appears that a considerable amount of heat is brought through the Bering Strait, and somewhat more by the Atlantic Current from the Norwegian Sea. But also the outflow of cold water and ice east of Greenland and through the Canadian Archipelago means a considerable net gain of heat.

THE NORWEGIAN SEA AND THE BARENTS SEA. The water masses of the Norwegian Sea and the Barents Sea may communicate directly with those of the adjacent seas in different ways, of which the following are believed to be most important.

19. Inflow of water from the Polar Sea (into the Barents Sea).
20. Outflow of Atlantic water to the Polar Sea.
21. Outflow of bottom water to the Polar Sea.
22. Inflow of Arctic water east of Greenland.
23. Inflow of ice east of Greenland.
24. Inflow of melting water east of Greenland.
25. Inflow of water from the Atlantic Ocean.
26. Outflow of Arctic water between Iceland and Greenland.
27. Outflow of ice between Iceland and Greenland.
28. Outflow of melting water to the Atlantic Ocean.
29. Overflow of deep water into the Atlantic Ocean.
30. Outflow of Atlantic water to the North Sea.
31. Inflow of Baltic water from the North Sea.

Further factors to be taken into consideration are

32. Fresh water received as precipitation.
33. Fresh water received as run-off from land.
34. Fresh water lost by evaporation.

and

35. Heat lost by evaporation as heat of vaporization.
36. Heat lost by evaporation as sensible heat.
37. Heat gained by radiation.
38. Heat lost by conduction through the ice.
39. Heat gained by conduction through the bottom.
40. Heat lost by the melting of ice.
41. Heat lost by glacier ice being received from East Greenland.

Each of these factors is treated below, and the resulting values are compiled in Table 4, in the order indicated by the numbers in column 2.

19—22. Exchange of water with the Polar Sea. This has been treated above under items 2—5.

23. Inflow of ice east of Greenland. This was estimated above to be $0.04 \cdot 10^6\,m^3sec^{-1}$ when including the ice transported into the Canadian Archipelago. In view of the poor accuracy we may adopt for the East Greenland Current a transport of $0.03 \cdot 10^6\,m^3sec^{-1}$ of ice of 10° and $5^0/_{00}$.

24. Inflow of melting water east of Greenland. This was estimated above to be $0.1 \cdot 10^6\,m^3sec^{-1}$ when including the melting water transported into the Canadian Archipelago. For the East Greenland Current we may adopt $0.07 \cdot 10^6\,m^3sec^{-1}$ and a temperature of — 0.3°.

25. The inflow of Atlantic water from the Atlantic Ocean has been studied by TAIT (1957) by dynamical computations on the basis of 46 sections across the Faeroe-Shetland Channel. Great variations were found, the extreme values being 0.4 and $6.5 \cdot 10^6\,m^3sec^{-1}$; the average of all values reported (*loc.cit.* Table 3, p. 52) is $2.3 \cdot 10^6\,m^3sec^{-1}$. Studies farther east, in the Sognefjord Section by SÆLEN (1959) have given (27 cases) variations of the total transport from 2.0 to 7.0, and of the transport of Atlantic water from 1.8 to $6.4 \cdot 10^6\,m^3sec^{-1}$; the average transport of Atlantic water is (*loc.cit.* Table 1, p. 6) $3.8 \cdot 10^6 m^3sec^{-1}$. In cases where observations were made within a few days under way towards the west-north-west and immediately afterwards on the return, divergencies of up to 80 per cent were encountered; thus in July 1955 the transport of Atlantic water was computed to be 2.1 and $3.8 \cdot 10^6\,m^3sec^{-1}$. As such strong and rapid fluctuations did not appear reasonable, several parallel sections were occupied, whereby semipermanent eddies were disclosed. Only preliminary reports on these investigations have as yet been published (MOSBY 1955, 1961); by thorough dynamical computations the different sections appeared to give greatly diverging values of transport. The problem shall not be discussed here, but apparently the applicability of the method of dynamical computations will have to be investigated very closely, before the results of such computations can be considered as valid. Even when considering the fluctuations found by TAIT (1957) as representative, it is difficult to obtain a reasonable estimate of the average or "normal" transport — the value desired for our present purpose, and this shall, therefore, be denoted by At. The simple mean of all values mentioned above (46 from TAIT's paper and 27 from SÆLEN's paper) is $2.8 \cdot 10^6$ m^3sec^{-1}, a value which may be expected to indicate the magnitude of the average transport.

As salinity and temperature of the Atlantic water in the Faroe-Shetland Channel we may adopt the simple averages of all of the narly 8000 salinities not below $35.00^0/_{00}$ given in TAIT's paper for the years 1927 to 1952 and of the corresponding temperatures, *i.e.* $35.3^0/_{00}$ and 8.9°C.

26. Outflow of Arctic water to the Atlantic Ocean between Iceland and Greenland. From data collected by the "Anton Dohrn" expedition in 1955 (DIETRICH 1957) it is seen that the temperature in summer is hardly higher than — 1.8 or — 1.7°, except within a thin surface layer. The corresponding salinity is usually about $34.3^0/_{00}$, decreasing to perhaps $32.4^0/_{00}$ near the surface. It therefore appears justified to consider the Arctic water, the ice and the melting water separately, as done above for the East Greenland Current between Greenland and Spitsbergen. As no well established estimate of the transport is known to have been made, we shall denote this volume transport by Ar.

27. Outflow of ice between Iceland and Greenland. The transport of ice from the Polar Sea into the Norwegian Sea was estimated above (Table 3) at about $0.04 \cdot 10^6$ m^3sec^{-1} including ice transported into the Canadian Archipelago. The transport between Iceland and Greenland will here be estimated as somewhat lower, $0.02 \cdot 10^6$ m^3sec^{-1} of ice of $5^0/_{00}$ salinity and of a temperature about the melting point or — 0.3°. For the water balance this somewhat arbitrary assumption is of no importance, but also for the heat balance it is of minor importance, as will be seen below.

28. Outflow of melting water to the Atlantic Ocean. From what was said above of the Polar water, its average salinity may be estimated at about $33.5^0/_{00}$, corresponding to a dilution of the Arctic water of $34^0/_{00}$ by melting water of $5^0/_{00}$ to an amount of hardly 2 *per cent.* We may therefore adopt for the transport of melting water a value of $0.02 \cdot Ar$.

29. Overflow of deep water into Atlantic Ocean. The overflow across the Iceland-Faeroe Ridge was studied from nine vessels in co-operation in 1960. The analysis of the data collected has not yet been finished, but preliminary investigations indicate that this transport may here be neglected.

30. Outflow of Atlantic water to the North Sea. Also this transport will here be neglected, the only consequence being that the transport of Atlantic water into the Norwegian Sea may come out slightly too low.

31. Inflow of Baltic water from the North Sea. According to JACOBSEN (1925) the transport of Baltic water into the Kattegat amounts to about $0.016 \cdot 10^6 m^3 sec^{-1}$ of water of a salinity of $18^0/_{00}$, while in return about $0.032 \cdot 10^6 m^3 sec^{-1}$ of water of a salinity of about $8^0/_{00}$ is transported from the Kattegat into the Baltic. From these figures it appears justified to conclude that the inflow of Baltic water through the North Sea (Norwegian coastal current) into the Norwegian Sea may here be neglected.

32. Fresh water received as precipitation. Specially careful attempts of measuring precipitation at sea have been carried out onboard the "Polarfront I" at Weather Station M (SKAAR 1955). From these an annual precipitation of $333 \text{ mm} \cdot \text{year}^{-1}$ was obtained. SPINNANGR (1958) from a comparison with measurements at certain stations ashore, for which reasonable estimates on the orographic effect can be made, concluded that $333 \text{ mm} \cdot \text{year}^{-1}$ must be too low and that the true value must probably be at least twice as great. This agrees with the map prepared by BIRKELAND and FØYN (1932), in which the position of station M falls somewhat nearer to the isohyet for 775 than to that for $500 \text{ mm} \cdot \text{year}^{-1}$. This map also illustrates the regular decrease of annual precipitation towards the north. For the Polar Sea we have above estimated the precipitation at about $100 \text{ mm} \cdot \text{year}^{-1}$, and a reasonable assumption might be about $300 \text{ mm} \cdot \text{year}^{-1}$ for the Barents Sea and about $600 \text{ mm} \cdot \text{year}^{-1}$ for the Norwegian Sea. This would correspond to a total amount of fresh water received as precipitation of nearly $0.07 \cdot 10^6 m^3 sec^{-1}$, a value which must at least be of the right order of magnitude, and which shall be adopted here.

33. Fresh water received as run-off from land. From detailed mesaurements (Hydrol. unders. i Norge 1958) it is found that the run-off along the Norwegian coast from the Skagerak to the Russian border amounts to about $0.011 \cdot 10^6 m^3 rec^{-1}$. The total amount of run-off from all coasts must, accordingly, be of little importance for the water balance; it may be estimated at about $0.02 \cdot 10^6 m^3 sec^{-1}$.

34. The fresh water lost by evaporation has been computed by conventional methods (see SVERDRUP 1951) from observations at station M to be $1222 \text{ mm} \cdot \text{year}^{-1}$. This is considerably more than illustrated by the map of ZAITZEV (1960), in which the highest values, above $800 \text{ mm} \cdot \text{year}^{-1}$, are found along the "tongue" of Atlantic water outside the coast of Norway. Without knowing the method of computation and the observational basis of ZAITSEV's map it is not possible to explain this great discrepancy. But the position of station M is certainly not representative of the entire area, and for our present purpose the fresh water lost by evaporation has been deducted from the values of heat lost by vaporization to be about $0.04 \cdot 10^6\,m^3sec^{-1}$ (see below).

35. Heat lost by evaporation as heat of vaporization. BUDYKO (1955) in his Atlas has drawn on his map for the annual heat loss by evaporation the isoline for 40 kcal. $cm^{-2}year^{-1}$ from Greenland through Eastern Iceland to the north-east corner of Scotland, and has covered most of the Norwegian Sea up to about 70°N by the colour corresponding to values between 40 and 20 kcal $\cdot\, cm^{-2}year^{-1}$. As his monthly maps do not cover this area, one may think that the annual map is perhaps in this case based on a relatively poor material of data. An approximate estimate of the average heat loss for the whole of the area on this basis would lead to about 30 kcal.cm^{-2} $year^{-1}$. The ice-covered areas in the west and north may in this connection be neglected, as the heat loss from the sea is estimated below from the amount of heat conducted through the ice. From different atlases somewhat diverging values are found for the areas covered by ice in different seasons. When taking into consideration only areas which are always, or nearly always, covered by ice, the probable extremes seem to be about $1.0 \cdot 10^6\,km^2$ in April and about $0.2 \cdot 10^6\,km^2$ in August. But considerable areas may be partially covered by drift-ice. The extremes seem to be about $2.2 \cdot 10^6\,km^2$ in April and about $0.8 \cdot 10^6\,km^2$ in August. In these areas the surface temperature must be very nearly uniform, and the evaporation must therefore be low or maybe even negligible. When applying the above approximate average value to the remaining ice-free areas of $2.7 \cdot 10^6\,km^2$ only, we arrive at a total heat of vaporization of $25.7 \cdot 10^9$ kcal $\cdot\, sec^{-1}$.

ZAITSEV (1960) has published a more detailed map of the heat loss by evaporation from the Norwegian Sea. By extrapolation for the Barents Sea a total amount of $25.3 \cdot 10^9$ kcal $\cdot\, sec^{-1}$ has been found for the always ice-free areas of $2.7 \cdot 10^6\,km^2$, while for the $3.6 \cdot 10^6\,km^2$ of no permanent ice a value of $27.3 \cdot 10^9$ kcal $\cdot\, sec^{-1}$ was derived. The total heat of vaporization H may thus tentatively be put at $26 \cdot 10^9$ kcal $\cdot\, sec^{-1}$, and this corresponds to an amount of water evaporating of $0.043 \cdot 10^6\,m^3sec^{-1}$.

36. Heat lost by evaporation as „sensible" heat. The heat lost to the atmosphere by convection in connection with the evaporation, the heat sometimes called "sensible heat", is given by ZAITSEV (1960) for the Norwegian Sea alone to be 80 *per cent* of the heat of vaporization of $24.4 \cdot 10^9$ kcal $\cdot\, sec^{-1}$, or $19.5 \cdot 10^9$ kcal $\cdot\, sec^{-1}$. We shall adopt for the heat lost to the atmosphere $r = 80$ *per cent* of the above established heat of vaporization, or $21 \cdot 10^9$ kcal $\cdot\, sec^{-1}$.

37. The heat gained by radiation may be evaluated for the ice-free areas as follows (see MOSBY 1936). The total radiation income is approximately equal to $0.024 \cdot [0.29 + 0.71(1 - \bar{C})] \cdot \bar{h}$ gcal $\cdot\, cm^{-2}min^{-1}$, where $\bar{C}$ denotes the part of sky covered by clouds and $\bar{h}$ the average solar altitude. The outgoing radiation is approximately $0.94 \cdot \AA$ $(1 - 0.83 \cdot \bar{C})$, where $\AA$ is taken from the table of ÅNGSTÖRM (1920). For Weather station M the radiation balance 1948—1958 has in this way been found to be 17.9 kcal · cm^{-2}year^{-1}, or somewhat lower than demonstrated by the map of ZAITSEV (1960, Fig. 1), which gives for the position 66°N, 2°E about 22 kcal · cm^{-2}year^{-1}. The average value of the solar altitude throughout the whole year (including night and day) is 10.5° at 60°N, decreasing northwards to 7.4° at 80°N. According to BIRKELAND and FØYN (1932, Fig. 20, 21) the average cloudiness within the ice-free areas should be a little lower than at station M. For these reasons it seems difficult to understand the great decrease of the radiation balance from 25 kcal · cm^{-2}year^{-1} at 63°N to 2.5 kcal · cm^{-2}year^{-1} at 80°N demonstrated by the map of ZAITSEV.

Supported by data on cloudiness and relative humidity of the air at Hopen Island, and taking only the ice-free areas in different latitudes into account, an independent estimate has been made; this led to an average radiation balance of 16.3 kcal · cm^{-2} year^{-1}. Using the values from the map of ZAITSEV (1960, Fig. 1) for the same areas we arrive at 15.1 kcal · cm^{-2}year^{-1}. We may therefore adopt for the $3.6 \cdot 10^6$ km^2 in question the average of these two values, or for the total heat gain approximately $R = 18.0 \cdot 10^9$ kcal · sec^{-1}.

38. Heat lost by conduction through the ice. The measurements from the "Maud" expedition (MALMGREN 1927) show that, with a certain lag, the heat conduction through the ice follows the air temperature, so that for the annual mean we may roughly adopt the heat loss to be between $-0.04 \cdot t_a$ and $-0.035 \cdot t_a$ kcal · cm^{-2} month^{-1}, where t_a denotes air temperature in °C. According to BIRKELAND and FÖYN (1932) the air temperature in January over the ice east of Greenland may be expected to be from − 2° near Iceland to − 17° in 80°N latitude, while it will be about − 20° between Spitsbergen and Novaja Zemlja. In July the corresponding figures are: from 7° to 2° in the west and about 2° in the Barents Sea. On this basis we may estimate the average heat loss through the ice at about 0.4 kcal · cm^{-2}month^{-1}, or at nearly 5 kcal · cm^{-2}year^{-1}. As the completely ice-covered area was estimated above at about $0.6 \cdot 10^6$ km^2, we thus arrive at a total loss of nearly $1.0 \cdot 10^9$ kcal · sec^{-1}. In view of the small importance of this value for our total balance, this rough estimate may be considered as satisfactory.

39. Heat gained by conduction through the bottom. As for the Polar Sea we may use the value $1 \cdot 10^{-6}$gcal · cm^{-2}sec^{-1}. For the $4.2 \cdot 10^6$ km^2 of the Norwegian Sea and the Barents Sea this corresponds to $0.04 \cdot 10^9$ kcal.sec^{-1}.

40. Heat lost by the melting of ice. On the above assumptions that $0.03 \cdot 10^6$ m^{3}sec^{-1} of ice is received from the Polar Sea, while $0.02 \cdot 10^6$ m^{3}sec^{-1} is transported into the Atlantic Ocean, the heat of fusion of $0.01 \cdot 10^6$ m^{3}sec^{-1} or $0.8 \cdot 10^9$ kcal · sec^{-1} must be the heat loss of the Norwegian Sea for the melting of ice.

41. Heat lost by glacier ice being received from East Greenland. For the whole of the east coast of Greenland BAUER (1954) estimates a discharge of 120 km^3 $year^{-1}$. A rough estimate might then be that about two-thirds of this quantity, or some 80 $km^3 year^{-1}$ is melting in the Norwegian Sea. For the water balance this will be insignificant, while the heat of fusion will amount to some $0.2 \cdot 10^9$ kcal $\cdot$ sec^{-1}.

Table 4. *Water, salt and heat balance of the Norwegian Sea and the Barents Sea.*

		10^6 m^3sec^{-1}	‰	°C	10^9kcal. sec^{-1}	
Inflow						
Polar Sea to Barents Sea	19	0.05	34.92	2.7	[0.14]	
East Greenland Current	22	2.0	34.0	−1.8		[−3.6]
Ice east of Greenland	23	0.03	5.0	−10.0		[−0.3]
Melting water east of Greenland	24	0.06	0.0	−0.3		[−0.02]
Atlantic water from Atlantic Ocean	25	At → [3.6]	35.3	8.9	[32.0]	
Baltic water from North Sea	31	0.0				
Precipitation	32	0.07	0.0			
Run-off	33	0.02	0.0			
Outflow.						
Atlantic water to Polar Sea	20	−1.4	35.10	3.25		[−4.55]
Bottom water to Polar Sea	21	−0.6	34.92	−0.9	[0.54]	
Arctic water to Atlantic Ocean ..	26	−Ar → [−3.6]	34.0	−1.8	[6.5]	
Ice to Atlantic Ocean	27	−0.02	5.0	−0.3	[0.01]	
Melting water to Atlantic Ocean.	28	−0.07	0.0	−0.3	[0.02]	
Deep water to Atlantic Ocean ..	29	0.0				
Atlantic water to North Sea	30	0.0				
Evaporation.						
Fresh water loss	34	−0.04				
Heat loss, vaporization	35					−H → −26.0
Heat loss, "sensible" heat	36					−rH → −21.0
Radiation, net gain	37				R → 18.0	
Conduction through ice	38					−1.0
Conduction through bottom	39				0.04	
Melting of sea-ice	40					−0.8
Melting of inland ice	41					−0.2

A compilation of the estimates established above is given in Table 4. When introducing these values into equations (1), (2) and (3), we obtain for water balance

$$Ar = 0.98 \cdot At + 0.17 \tag{9}$$

and for salt balance

$$Ar = 1.04 \cdot At - 0.01 \tag{10}$$

or approximately

$$(11) \qquad Ar \approx 1.04 \cdot At \approx At$$

For heat balance we find

$$(12) \qquad 1.8\,Ar + 8.9\,At = 9{,}7 + (1 + r)H - R$$

whereby

$$(13) \qquad Ar \approx At \approx 0.9 + 0.093[H(1 + r) - R]$$

and for $H = 26.0$, $r = 0.8$ and $R = 18.0$

$$(14) \qquad Ar \approx At \approx 3.6.$$

By a rough estimate also the average temperature of the waters of the Norwegian Sea and the Barents Sea has been taken as 0.00°, leading to the figures given in parentheses in the last two columns of Table 4.

Discussion. In so far as the figures of Table 4 may be relied upon it is seen that the main loss of heat is due to evaporation, amounting to 82 *per cent* of the total loss. The net gain of heat by radiation is 31 *per cent*. On the assumption of an average water temperature of 0°, the total effect of the East Greenland Current including ice, melting water and the melting of ice is nearly negligible, but the heat brought by the Atlantic current through the Faeroe-Shetland Channel makes 56 *per cent* of the total.

An approximate relationship between the transport At in $10^6\,m^3sec^{-1}$ of the Atlantic Current and the average evaporation E in mm $\cdot$ day^{-1} over the ice-free areas of $2.7 \cdot 10^6$ km^2 may be established when all other factors are accepted as given in Table 4. We find

$$(15) \qquad At \approx 1.6 \cdot (1 + r) \cdot E - 0.8$$

where r is the Bowen ratio, *i.e.* the ratio between the amount of heat conducted to the air and the heat of vaporization. On the above assumption of ZAITSEV (1960) that $r = 0.8$ we find $E = 1.5$ mm $\cdot$ day^{-1} if $A = 3.6 \cdot 10^6\,m^3sec^{-1}$. This agrees well with the average value 1.4 mm $\cdot$ day^{-1} for the Norwegian Sea alone, computed from the map of ZAITSEV. It is interesting to note that the values of At determined by TAIT (1957, Fig. 7) seem to indicate an annual variation of between $\pm$ 1 and $\pm\, 2 \cdot 10^6\,m^3sec^{-1}$, the highest values occurring in autumn-winter and the lower ones in spring-summer. This may agree with the annual variation of the evaporation at Weather station M, which shows a maximum of above 5 mm $\cdot$ day^{-1} in January-February and a minimum below 2 mm $\cdot$ day^{-1} in June-July.

According to Table 4 about $21.0 \cdot 10^9$ kcal $\cdot$ sec^{-1} may be expected to be transferred to the atmosphere from an ice-free area of $2.7 \cdot 10^6$ km^2, or 67.2 gcal $\cdot$ cm^{-2}day^{-1}. This would suffice for an increase of the temperature of the atmosphere above the same area by about 0.28°C per day.

SYMBOLS AND UNITS

A	$10^6\, m^3sec^{-1}$	Atlantic water into Polar Sea
Ar	$10^6\, m^3sec^{-1}$	Arctic water (— 1.8°, 34.0°/$_{00}$) out of Norwegian Sea
At	$10^6\, m^3sec^{-1}$	Atlantic water into Norwegian Sea
B	$10^6\, m^3sec^{-1}$	bottom water into Polar Sea
C	10^9 kcal · sec^{-1}	conduction through ice-cover of Polar Sea
c	cm · sec^{-1}	average speed of East Greenland Current, 80°N
d	m	average ice-thickness East Greenland Current, 80°N
E	mm · day^{-1}	evaporation
e_a	mb	water vapour pressure in air
e_s	mb	water vapour pressure at sea-surface
G	$10^6\, m^3sec^{-1}$	Arctic water (— 1.8°, 34.0°/$_{00}$) by East Greenland Current, 80° N
g	$10^6\, m^3sec^{-1}$	ice by East Greenland Current, 80°N
H	10^9 kcal · sec^{-1}	heat of vaporization by evaporation from Norwegian Sea and Barents Sea
p	%	average ice-free areas of the Polar Sea
Q	10^9 kcal · sec^{-1}	heat transport
R	10^9 kcal · sec^{-1}	radiation balance of Norwegian Sea and Barents Sea
r	dim. less	ratio sensible heat to heat of vaporization (equiv. Bowen ratio)
S	°/$_{00}$	salinity
t	°C	temperature
t_0	°C	average temperature of all waters in a basin
t_a	°C	air temperature
t_s	°C	sea-surface temperature
V	$10^6\, m^3sec^{-1}$	volume transport
W	m · sec^{-1}	wind speed
$Å$	gcal · $cm^{-2}min^{-1}$	outgoing radiation
$0.16 \cdot p$	10^9 kcal · sec^{-1}	heat of vaporization, Polar Sea
$0.27 \cdot p$	10^9 kcal · sec^{-1}	sensible heat, Polar Sea
$1.36 \cdot p$	10^9 kcal · sec^{-1}	radiation balance, Polar Sea
$0.16 \cdot c \cdot d$	10^9 kcal · sec^{-1}	heat gain by melting of ice outside the Polar Sea
$0.02 \cdot Ar$	$10^6\, m^3sec^{-1}$	transport of melting water into Atlantic Ocean
$G = 0.2 \cdot c$		water transport of East Greenland Current, 80°N
$g = 0.002 \cdot c \cdot d$		ice transport of East Greenland Current, 80°N

Polar Sea taken as	$10.0 \cdot 10^6 km^2$
Norwegian Sea total	$2.6 \cdot 10^6$ -
Barents Sea total	$1.6 \cdot 10^6$ -
Polar ice	$0.6 \cdot 10^6$ -
Drift ice	$1.5 \cdot 10^6$ -

Specific heat of water taken as 1 gcal per gramme and °C

Density of water	—	1 g · cm^{-3}
Heat of fusion of ice	—	80 gcal · cm^{-3}
Heat of vaporization	—	560 gcal · cm^{-3}

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