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Ólafsfjarðarvatn — a saline meromictic lake in North Iceland

by

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ABSTRACT

Lake Ólafsfjarðarvatn (2.25 km²), North Iceland, has an outlet to the sea through a 0.5 km long estuary. The lake has a thin fresh water surface layer (thickness: 1½–2½ m when ice-free, 0–1 m when ice-covered), separated from a saline deeper layer (maximum thickness: 7½–8½ m) by a sharp halocline. Renewal of the saline water occurs through the estuary, mainly under the conditions of strong onshore winds coinciding with spring tide. In the absence of strong winds vertical mixing is suppressed by the strong density gradient at the interface between the two layers. Therefore, the solar energy which penetrates into the lower layer in spring and summer is retained there, and as a result a temperature maximum of 14°–20°C is formed at a depth of 3–4 m. This temperature maximum can persist for several months in calm weather. Slight temperature variations in summer seem to be related to variations in wind velocity. In several instances (May-June 1980, May-June 1981 and June-July 1982) the measured heat flux into the saline layer is only slightly less than the estimated flux based on information on solar radiation and light measurements. Appreciable cooling due to vertical mixing, especially in autumn, does not occur until a wind velocity of 30–40 knots is reached. The upper part of the nutrient-rich saline layer is highly productive. As the result of the photosynthetic activity, large amounts of the oxygen formed may accumulate in the lake leading to exceptionally high super-saturation values. In winter when the lake is ice-covered the oxygen supply is largely used up for oxidizing decomposing organic matter. In the uppermost part of the saline layer, however, oxygen is not depleted, although the concentrations become quite small at depths greater than 4 m. Within the saline layer vertical eddy coefficients are found to be relatively small. It is anticipated that salinity and chemical components such as oxygen, alkalinity and silicon will differ significantly from year to year depending upon the intensity of winter storms causing inflow of sea water into the lake. This is confirmed by chemical measurements in the period 1978–1983.

INTRODUCTION

Often saline aquatic environments are in a restricted communication with the ocean — or have been in the past. In most cases such bodies of water have the character of an estuary, and the sea water within them is appreciably diluted by fresh water drainage. Frequently they are measurably renewed during each tidal cycle. This is the case of many lagoons, fjords and fjordlike regions, as typified by Nýpslón on the northeast

coast of Iceland (Stefánsson and Jóhannesson 1982). But landlocked basins are also found, where in the course of time the outlet to the sea has become partially filled with gravel, sand or clay, so that it is no longer deep enough to permit the inflow of sea water. This leads to stagnation of the saline deep water inside the basin and in time a gradual thickening of the overlying fresh water. A good example of such an environment is Lake Miklavatn on the north coast

of Iceland (Stefánsson and Jóhannesson 1978).

In exceptional cases a limited renewal of the deep water inside a basin or lake takes place under special conditions, such as high tide coinciding with abnormally low atmospheric pressure and/or storm surges. Sea water may then enter the lake along the channel bottom. Renewal under these circumstances will be slow but may nevertheless be sufficient to prevent a permanent stagnation of the deep water and maintain a thin surface layer of outflowing fresh water. Such conditions give rise to a unique phenomenon, viz. the formation in spring and summer of a temperature maximum in the upper part of the saline layer.

Basins of this last-mentioned type, both natural and man-made, have for a long time been known to exist in a few localities around the world (see e. g. Anderson 1958). Thus in ancient writings from Roman times it is reported that oyster culture was more successful in salt water ponds with fresh water drainage than in the open sea. Under the surface layers of such ponds the temperature rose rapidly and created favourable environments for spawning and growth (Gaarder and Spärck 1932). It is believed that the Romans made artificial ponds to utilize such conditions. Later on knowledge about this method was passed on to France and eventually to Scandinavia during the latter half of the last century.

The best known examples of basins of this kind, with intermediate temperature maximum formed in summer, are probably the oyster basins in western Norway. Among the first scientists to offer an explanation of the phenomenon was Professor Helland (1889). Studies of the physical characteristics of these Norwegian basins were made by Helland-Hansen (1907, 1908), whereas their biochemical properties were investigated by Gaarder and Spärck (1932).

It is now known that this intermediate temperature increase observed in a few

coastal basins and meromictic¹ lakes, is due to a "greenhouse" effect provided by the sharp pycnocline (halocline) at the interface between fresh or brackish water on top and saline water below. The solar energy which in spring and summer is absorbed by the fresh (brackish) water layer and converted to heat energy, is largely transferred to the atmosphere in the form of latent and sensible heat, while a smaller fraction is used for warming up the surface water which subsequently flows out to sea. The saline layer, on the other hand, is renewed only slowly or on rare occasions, as previously stated, and the sharp pycnocline prevents vertical mixing. The portion of the radiation energy which penetrates the pycnocline is also converted to heat energy, but since it can not escape from the saline layer, it is utilized for warming up the water. The result is the formation of a temperature maximum in the upper part of the saline layer.

Natural or artificial bodies of water of the kind here described offer unique possibilities for mariculture of warm-water species in cold-tempered or polar regions. It should also be possible to utilize these environments for accelerating the growth of species native to the area in question. A highly interesting aquatic environment of this category is Lake Ólafsfjarðarvatn in Ólafsfjörður, North Iceland. The present communication describes recent studies of its chemical and thermal properties and budgets, and an attempt is made to explain how they are affected by meteorological conditions.

PREVIOUS STUDIES

For centuries farmers in Ólafsfjörður have utilized the lake as a source of food. In former times it was considered unique because of its marine fish which were thought

¹ A lake in which some of the water remains partly or wholly unmixed with the main water mass at the circulation periods is said to be *meromictic* (Hutchinson 1957).

to be capable of living in fresh water. In his extensive treatise on the geographical and economical history of Iceland, Eggert Ólafsson (1772) wrote that the lake was abundant in fish and remarkable because of its marine species, such as cod, haddock, halibut and skate. These fish, he said, were considered good eating and they had a sweet taste, different from that of ordinary sea fish. Ólafsson believed that the sand reef separating the lake from the ocean had been formed by volcanic activity or by breaking waves, leaving marine fish behind in the lake, and he postulated that these fish had gradually acclimatized to the fresh water.

It is not unlikely that Ólafsson's description of the lake and his extraordinary hypothesis that sea fish had adjusted to fresh water, may have stimulated the French who sent an expedition to Iceland and Jan Mayen in 1891 to make a special investigation of Lake Ólafsfjarðarvatn and its strange fish. The French observations revealed that the lake was two-layered, with fresh water on top and saline water below, and the investigators concluded that there was nothing mysterious about the marine fauna in the lake: the fish could simply swim in or out through the estuary during spring tide (Rabot 1894). A further description of the lake and its estuary was given by Thoroddson (1898).

In the summer of 1900 the Icelandic fishery biologist Bjarni Sæmundsson made some observations in the lake. These included depth measurements which showed that in the outer (northern) part it generally ranged from 1 to 6 feet in depth but was much deeper in the inner (southern) part, where a sizeable area was about 5 fathoms deep, and the greatest depth measured 5½ fathoms. These results are quite similar to those found by the present authors in 1978. Sæmundsson made temperature and salinity observations at one locality where the lake was 5 fathoms deep. His results were as follows: temperature at the surface 12.5°C, at

the bottom 6.5°C; salinity at the surface 0‰, at 2 fathoms 4.06‰, at 3 fathoms 24.6‰ and at 5 fathoms 28.5‰ (Sæmundsson 1901). It is interesting to note that he did not report temperature for intermediate depths where he would no doubt have found a temperature maximum. In addition to temperature, salinity and depth measurements, Sæmundsson took some bottom samples. He remarked that he had seen no evidence of marine fish in the lake, and he did not find any small benthic animals except fresh water molluscs. Where the lake was shallow enough to be fresh at the bottom, it was covered with grayish-yellow silt with some vegetation, while in the deeper saline part he found black, foul-smelling mud. Sæmundsson believed that renewal of the saline water took place mainly by percolation of sea water through the sand bank, whereas renewal through the estuary would occur only during very high tides. He indicated that life conditions for marine fish had deteriorated a few years prior to his survey due to shallowing of the estuary, with the result that marine animals had largely disappeared from the lake as it became more fresh.

It is now well known that several species of marine fish, such as cod, pollack, plaice and even herring have been common in Lake Ólafsfjarðarvatn from the time Sæmundsson made his studies there, more than 80 years ago and up to the present time. Residents of Ólafsfjörður maintain that such fish may enter the lake, probably every year, under exceptional meteorological conditions, as will be discussed in a later section of this paper.

MATERIAL AND METHODS

1. Depth measurements

When the present investigations were started in July 1978, there was no depth chart available of Lake Ólafsfjarðarvatn. The first task, therefore, was to make depth

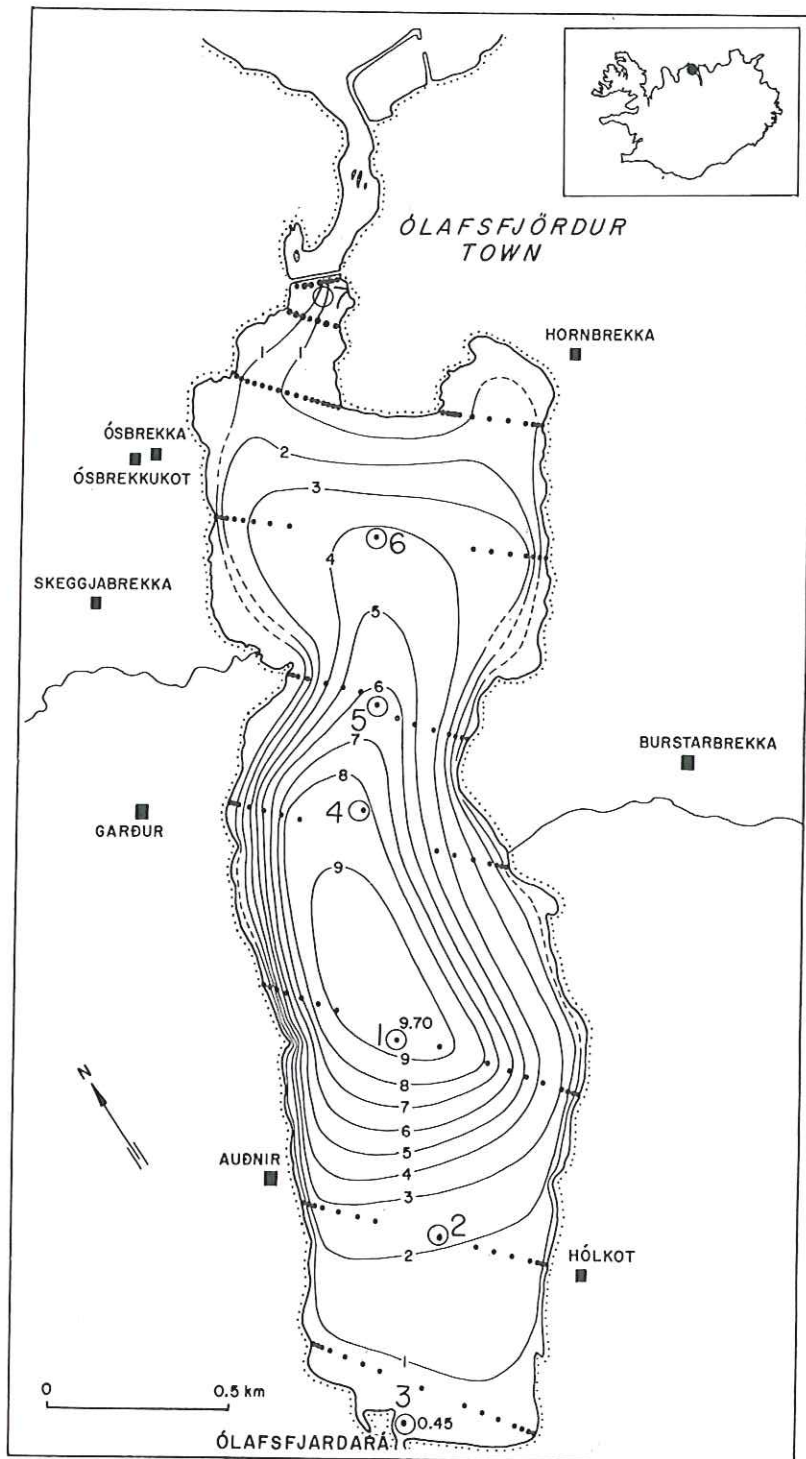


Fig. 1. Bathymetrical chart of Lake Ólafsfjardarvatn. Depths in meters. Small circles denote stations occupied.

measurements and draw up a bathymetric chart. Depths were determined using a fine sounding line with a small weight attached to it. Exact positions were obtained by erecting flag poles on the shore and measuring distances along straight transects using a graduated 200 m steel tape. In the deepest part, outside the reach of the tape, positions were determined by taking bearings by means of a prism. The results were plotted on an aerial map of the lake (Fig. 1).

There are seasonal variations in the water level of the lake, but their magnitude has not been determined. It is estimated, however, that they rarely exceed 0.5 m. At the time of the depth measurements in July 1978, the height of the water level was considered to be about average.

2. Surveys and station data

In the period July 1978 – July 1983 a total of 14 excursions were made to Lake Ólafsfjarðarvatn for collecting samples and carrying out various observations. These included determinations of salinity, dissolved oxygen, pH, alkalinity, inorganic and total phosphorus, nitrate, nitrite, ammonia, chlorophyll a, primary production and suspended matter in addition to measurements of temperature and transparency of visible light. Analytical results and other relevant data are presented in Table I in the Appendix and in Table 4 (productivity data).

Stations occupied during the first survey are shown in Fig. 1. Since the results from these first measurements indicated that there was no significant difference between properties of samples taken at the same depth but from different parts of the lake, the subsequent surveys were limited to one locality (St. 1) in the deepest part. On a few occasions additional samples were taken in the estuary at the northern end of the lake and from the river Ólafsfjarðará discharging into the southern end.

3. Physical and chemical methods

Water samples were taken from different depths by means of an NIO plastic water sampler with attached deep sea thermometers (precision: $\pm 0.01^\circ - 0.02^\circ\text{C}$).

Dissolved oxygen concentration, alkalinity and pH were determined in Ólafsfjörður immediately after the samples had been collected. Other analyses were made in the chemical laboratory of the Marine Research Institute, Reykjavík. Determinations of oxygen, alkalinity, pH, salinity and nutrients were made by standard methods (see e. g. Einarsson and Stefánsson 1983). Samples for determination of chlorophyll a were filtered through a glass fiber filter (GF/C 1.2 μ), dried, extracted with 90% acetone and analyzed spectrophotometrically (Anon. 1966). Phaeo-pigments were determined by a modification of the method of Marker et al. (1980). Suspended matter was determined by filtering 0.5–0.1 litre of water through a preweighed glass fiber filter (GF/C 1.2 μ). Following washing and drying the filters were reweighed (Stefánsson 1981).

Primary production was measured according to the well known method of Steemann-Nielsen (1952): samples were taken from different depths, transferred to glass bottles, a known amount of $\text{NaH}^{14}\text{CO}_3$ added and the bottles subsequently suspended in the lake at the depths from which the samples had been taken. The *in situ* incubation time was 4–6 hours between 10:00 a. m. and 4:00 p. m. The samples were then filtered through a membrane filter, washed and dried, and the radioactivity determined with a Geiger flow counter with a Mylar window (Theódórsson 1975).

Transparency was estimated with a Secchi disk and on two occasions (June and July 1982) also measured directly using a Li-COR 192S underwater quantum meter (Lambda Instrument Corporation) coupled with a Li 190S sensor for reference measurements.

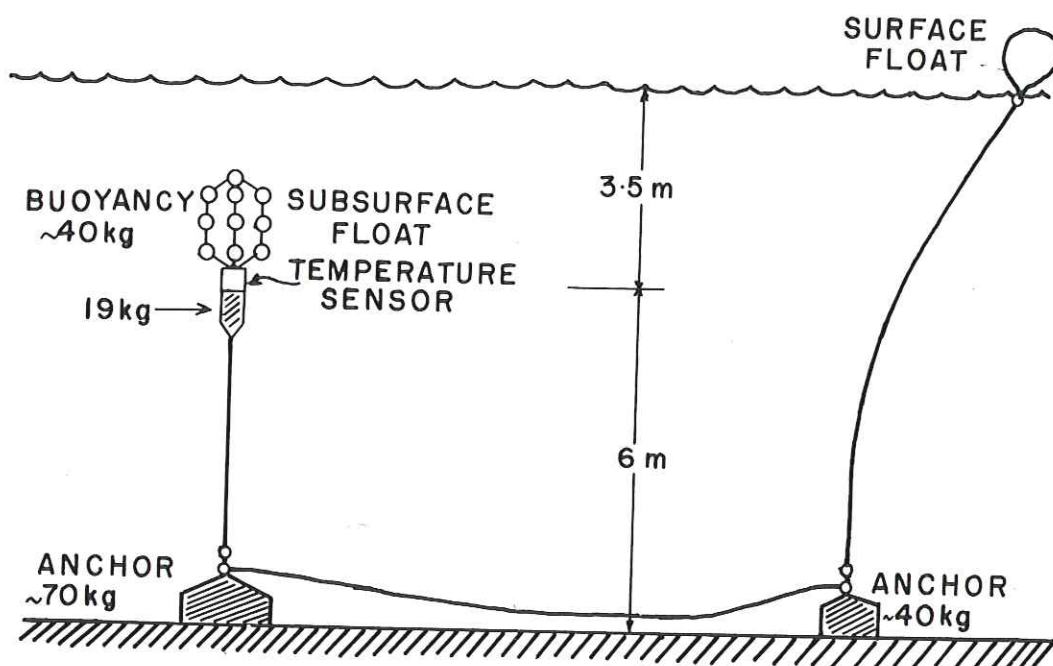


Fig. 2. Installation of Andraea data logger for temperature measurements at $3\frac{1}{2}$ m at St. 1.

In the one year period May 21, 1980 – May 25, 1981, temperature was measured every hour by means of a temperature sensor which was anchored at St. 1 (see Fig. 1) in the deepest part of the lake at a depth of $3\frac{1}{2}$ m. The mooring arrangement is shown in Fig. 2. The surface float was removed just before ice formed on the lake in the autumn and the instrument recovered by dredging in May 1981. The sensor, a thermistor inside the pressure housing of an Andraea current meter, recorded temperature every hour and stored the signals on a magnetic tape.

Current measurements were carried out on June 26, 1981, using a simple device: a wooden cross on a vertical pole was ballasted with a lead weight at the lower end and an adjustable piece of styrofoam on top. By moving the styrofoam up and down the pole, the level of the wooden cross could be varied for measuring currents at different depths. A light nylon line, attached to the top of the wooden pole, was paid out as the

float drifted with the current. The current velocity was then determined from the time it took the float to drift a measured distance. A total of 11 measurements were made in calm weather on a 150 m long section across the lake just south of the estuary.

RESULTS

1. Bathymetry and size

The lake which is located just south of the town Ólafsfjörður occupies the head of the fjord and is now separated from the sea by a low-reef barrier, 500–700 m wide. The barrier is breached by the lake outlet which runs along the northwest side of the town. The lake is about 3.3 km long from the estuary to the river Ólafsfjarðará which discharges into the lake at its southern end (Fig. 1). The width is generally between 0.5 and 1.0 km. The lake is appreciably deeper in the southern part, between the farms Garður and Auðnir, where there is a de-

pression with depths from 7 to 9 m. The greatest depth measured was about 10 m. In the southern part the bottom is considerably steeper on the west side where depths increase uniformly down to 8 m about 150 m offshore. In the northern (outer) part most of the lake is between 1 and 5 m deep. In the outlet the greatest depths are between 1 and 2 meters.

Areas and volumes inside different isobaths are given in Table 1. The total area of the lake is 2.25 km² and the volume 8.3 × 10⁶ m³. Its mean depth is thus 3.7 m.

0.15 m³s⁻¹. These figures correspond to a mean total discharge into Lake Ólafsfjarðarvatn of about 12.8 m³s⁻¹ and a minimum total discharge of about 1.5 m³s⁻¹.

Fairly comprehensive data are available on the discharge into the neighbouring lake Miklavatn in Fljót (Stefánsson and Jóhannesson 1978). If we assume the runoff per unit area per year to be similar for these two drainage areas, the annual average discharge into Lake Ólafsfjarðarvatn will be 8.1 m³s⁻¹, the maximum mean monthly discharge will be 20 m³s⁻¹ (in June) and the

TABLE 1

The area and volume inside different isobaths in Lake Ólafsfjarðarvatn as determined planimetrically from Fig. 1.

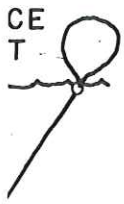
Depth interval (m)	Area (km ²)	Mean Depth (m)	Volume (10 ⁶ m ³)
0-1	0.370	0.5	0.185
1-2	0.418	1.5	0.627
2-3	0.291	2.5	0.727
3-4	0.322	3.5	1.127
4-5	0.232	4.5	1.044
5-6	0.128	5.5	0.704
6-7	0.118	6.5	0.767
7-8	0.097	7.5	0.727
8-9	0.146	8.5	1.241
>9	0.124	9.35	1.159
Total	2.246	3.7	8.308

2. Water budget

In addition to the river Ólafsfjarðará, which is by far the largest, a few small ones enter the lake. No direct measurements have been made of the fresh water discharge into the lake except for the river Garðsá (Rist 1956) on the west side, between the farms Skeggjabrekka and Garður. The drainage area of this small river is estimated to be 17 km², which is about 1/10 of the total drainage area for Lake Ólafsfjarðarvatn, 168 km². The mean runoff from Garðsá was found to be 1.30 m³s⁻¹ and the minimum runoff during extended frost periods about

minimum mean monthly discharge about 3 m³s⁻¹ (in January). These values are somewhat lower than those obtained on the basis of the discharge of the river Garðsá. Therefore, the true mean discharge is probably in the range 8.1–12.8 m³s⁻¹, the highest mean monthly discharge 20–30 m³s⁻¹ and the lowest mean monthly discharge 3–5 m³s⁻¹.

The results of the current measurements are given in Table 2. The difference between the 11 values is hardly significant. From the mean value (26 cm s⁻¹) and the cross-sectional area (94.6 m²) the mean discharge is found to be 24.6 m³s⁻¹. Since the



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TABLE 2

Current velocities (cm/s) on the bathymetrical section across Lake Ólafsfjarðarvatn some 100 meters south of the bridge (see Fig. 1).

Depth (cm)	Distance (m) from land on the west side							
	6	17	30	43	70	82	107	130
25	29.4	23.3	22.6	25.2	26.8	30.7	24.6	22.2
50	—	—	—	—	27.6	27.1	—	—
75	—	—	—	—	—	27.3	—	—

Mean velocity: 26.07 ± 2.75 cm/s

measurements were made near the time of the year when the runoff is normally maximal, this result is in good agreement with the estimated one.

As we have indicated, the renewal of the saline layer of the lake is quite slow and the exchange of water between the two layers very limited. The flushing time of fresh water in the lake is therefore determined by the rate of discharge and the volume of the upper layer, which on the average will be about 2 m in thickness when the lake is ice-free. From Table 1 we find that the fresh water volume of the lake will be $3.73 \times 10^6 \text{ m}^3$. Using the value of $10.5 \text{ m}^3 \text{ s}^{-1}$ for the mean discharge, we arrive at a mean renewal time of $(3.73 \times 10^6 \text{ m}^3) / (10.5 \text{ m}^3 \text{ s}^{-1} \times 60^2 \times 24 \text{ s day}^{-1}) = 4.1$ day. The maximum renewal time would be about 11 days and the minimum about 2 days. Thus the flushing time of fresh water in Lake Ólafsfjarðarvatn is quite short.

3. Temperature and salinity

When the lake is ice-free the uppermost 2 meters normally consist of fresh water only. Below there is a sea water layer which reaches a thickness of about 8 meters in the deepest part of the lake. The boundary between these two layers is extremely sharp, the salinity increasing by as much as 22–23 units per meter at the interface.

The fresh water thickness changes season-

ally. During spring and summer, when the lake is ice-free, the layer is $1\frac{1}{2}$ – $2\frac{1}{2}$ m thick (Fig. 3a, Table I), but when the lake is frozen and fresh water afflux at a minimum, the fresh water layer under the ice may be very thin (Fig. 3b). Thus the salinity will be highly variable between 0 and 2 m in winter but between 2 and 3 m when the lake is ice-free.

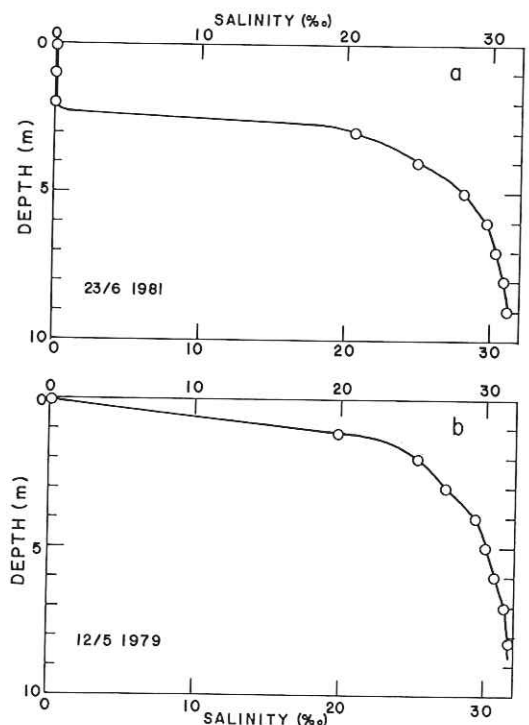


Fig. 3. Vertical distribution of salinity in Lake Ólafsfjarðarvatn. a: 23/6 1981; b: 12/5 1979.

TABLE 3

Temperature of the river Ólafsfjarðará during the winter of 1965–1966, near the farm Reykir. Beyond this locality (farther away from the lake) the river was flowing under snow cover.

Water temperature in river Ólafsfjarðará		Mean monthly air temperature Siglunes (about 18 km away)	
Date and year	Temperature °C	Month and year	Temperature °C
16/9 1965	2.9	October 1965	5.4
15/10	2.4	November	0.1
20/12	1.9	December	- 1.4
10/1 1966	2.4	January 1966	- 1.9
8/2	2.4	February	- 2.4
3/4	2.4	March	- 2.1
10/5	2.4	April	0.8
26/5	2.4	May	2.5
24/6	3.4	June	7.7

In the deeper part of the saline layer the salinity ranges between 31 and 32 which is only 2–3 salinity units lower than in the adjacent ocean area.

Since density depends largely on salinity, there is a very sharp pycnocline at the boundary between the fresh water on top and the saline water below. Mixing between the two layers will therefore be very limited.

The winter temperature of the river Ólafsfjarðará has been found to be relatively high (Table 3). It is also of interest to note that on May 12, 1979, at a time when there was a thick snow cover in Ólafsfjörður, the river temperature was as high as 3.4°C. This can probably be attributed to a large proportion of spring-fed water in winter, whereas the melting of ice and/or snow would be expected to produce water relatively low in dissolved solids. Consequently, the composition of the fresh water drainage into the lake appears to be variable with respect to some of the dissolved components. This is in a minor way reflected in the alkalinity but to a much greater extent in the dissolved silicon (Table I, Fig. 11). Thus in the spring and summer of 1980 and in the spring of 1981 the silicon concentration was

much smaller than during the other surveys with silicon observations.

When the first survey was carried out in July 1979, the authors were surprised to find a temperature maximum of about 14.5°C at a depth of 3½ m, i. e. near the top of the saline layer, just below the fresh water layer where the temperature was only 7°–8°C. Similar temperature distribution was observed during the subsequent summers, but the temperature maximum is only found in the uppermost part of the saline layer when the lake is ice-free. This will be clear from Fig. 4 which illustrates the vertical temperature distribution in different seasons. It will be seen that on May 21, 1980, shortly after the lake became ice-free, a temperature maximum had begun to develop at a depth of 3–3½ m, where the temperature had reached 8.5°C. About the middle of June it had increased to 15°C at a depth of 3½ m, and on August 19, following a spell of warm and calm weather, it was almost 20°C at a depth of 4.25 m. This value is comparable to summer temperatures in the Bay of Biscay. At the end of November 1980, when the lake was ice-covered, a marked cooling had taken place in the uppermost layers. The

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Lake Ólafsfjarðará

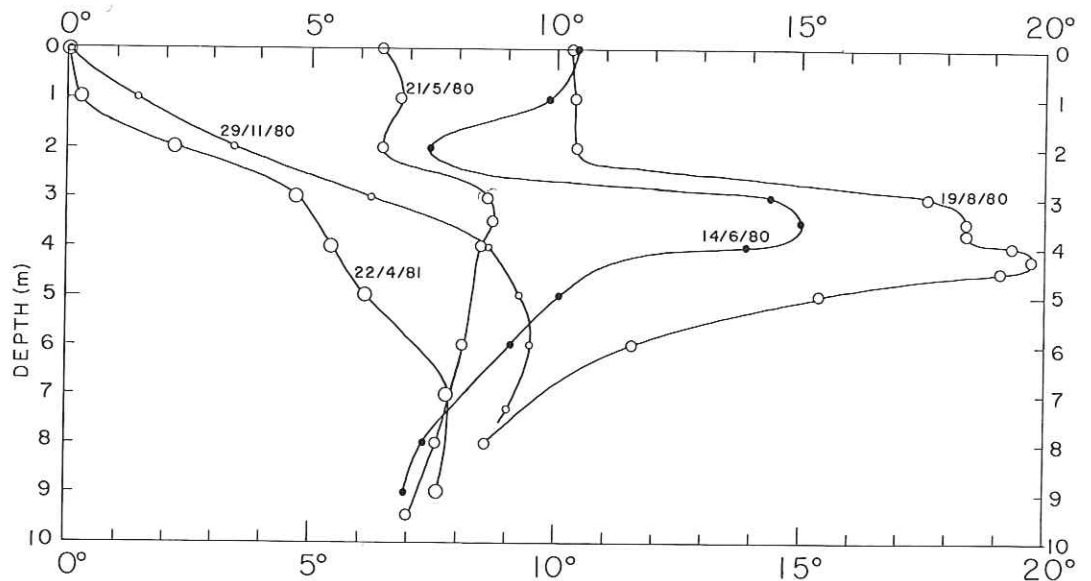


Fig. 4. Vertical distribution of temperature, 21/5 1980, 14/6 1980, 19/8 1980, 29/11 1980 and 22/4 1981.

maximum was now located at about 6 m depth with a temperature of 9.5°C. Finally, in late April 1981, a few weeks before the ice melted, the lake had cooled considerably from November, and a temperature maximum of about 8°C was now found at roughly 7 m depth. The temperature maximum is thus gradually displaced downwards during the one year period following the melting of the ice. While the lake is ice-free, this displacement is no doubt caused by vertical mixing from above, primarily due to the action of the wind. In the course of time this mixing gradually eats its way farther and farther downwards.

A detailed record of temperature changes in the uppermost part of the saline layer was obtained with a temperature sensor which, as stated before, was placed at a depth of 3½ m (Fig. 2). Daily temperature means for the whole period 21/5 1980 – 25/5 1981 are shown in Fig. 5. It will be noted that the temperature rose rapidly from the time the sensor was installed until June 20, when the temperature dropped by about 2°C. It then increased again to more than 18°C in the

beginning of July. During the following 8 weeks it fluctuated slightly about a mean value of just below 18°C. In early September it dropped by about 2°C and then remained 15°–16°C for another week. At noon September 16, the temperature dropped very suddenly from 14.7°C to 6.5°C at 13:00 hours, i. e. by about 8°C during one hour. For the rest of the day and the following day the temperature fluctuated up and down, but then rose rather rapidly to 12°C. During the 6th and 7th of October the temperature fell very abruptly to about 3°C. Again it rose and reached 7°C under the ice cover at the end of the year 1980. During the next 5 months the temperature remained quite uniform at this depth, decreasing very gradually down to 5.5° at the beginning of May 1981.

4. Oxygen

In the fresh water layer, saturation values were at all times near 100%, and only a slight supersaturation occurred during the photosynthetic period. The relatively low

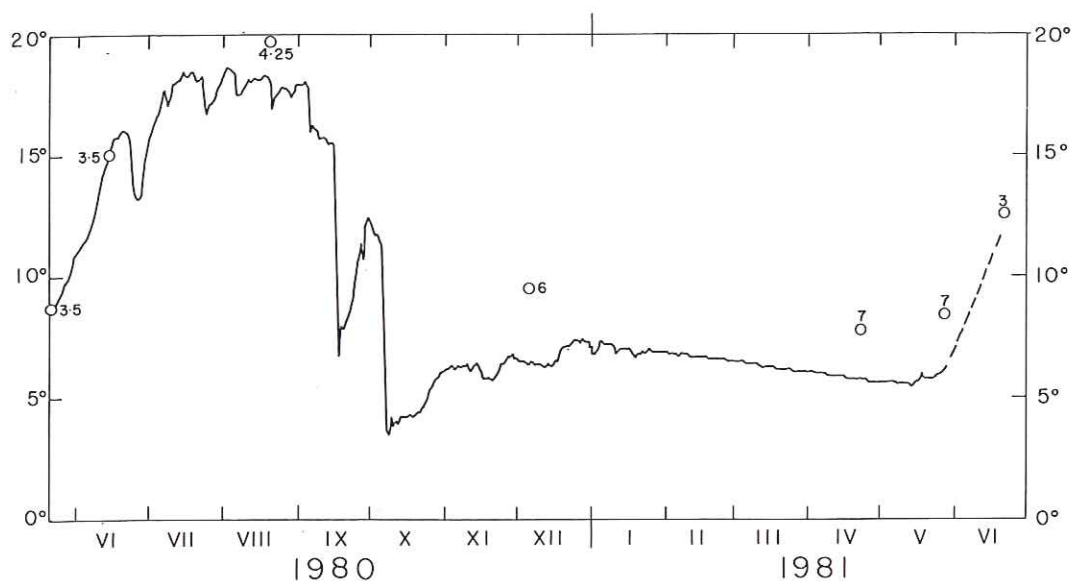


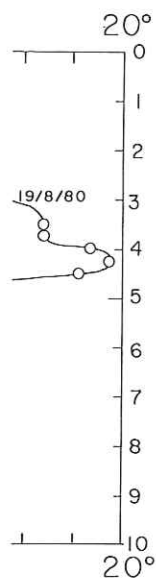
Fig. 5. Temperature variations at a depth of 3½ m at St. 1 in the period 20/5 1980–21/5 1981. Circles denote temperature maxima, when samples were taken at St. 1 and numbers indicate depth of these maxima.

supersaturation values of the fresh water were no doubt primarily due to exchange with the atmosphere as normally found in the surface layers of aquatic systems. Besides, oxygen production in the fresh water was probably at all times limited, as indicated by the low values of primary production in this water.

During the photosynthesis of plants in the uppermost part of the saline layer in spring and summer a considerable production of organic matter takes place. At the same time oxygen is formed and it will largely accumulate in the saline layer, as seen in Fig. 6. In June 1980 the oxygen saturation had reached more than 200% at a depth of 3–4 m and in August 1980 it approached 400%. This is as far as we know the highest supersaturation which has been reported anywhere for aquatic environments. Gaarder and Spärck (1932), who investigated similar meromictic environments in Western Norway, found about 250% saturation, which is about the highest saturation value we have seen reported. It should be noted, however, that equilibrium solubility of gases decrea-

ses with increasing temperature. It is also well known that it will take the water some time to reach equilibrium again, when the solubility decreases or increases due to seasonal temperature or salinity changes. Therefore, in temperate regions in spring a slight supersaturation is always found. In Icelandic coastal or oceanic waters this amounts to about 105% on the average (Stefánsson, unpublished data), and a corresponding undersaturation (about 95%) is normally found in early winter. The exceptional temperature increase in the upper part of the saline layer in Lake Ólafsfjarðarvatn in spring and summer could as a maximum lead to an apparent supersaturation of 30% (saturation value of 130%). Thus the observed saturation is mostly due to oxygen production. This is also clearly seen by the actual concentration (Table I), e. g. in June and August 1980.

In winter when the lake is ice-covered, organic matter will be decomposed and oxidized. This of course leads to reduction of the oxygen concentration rendering the deepest part of the saline layer oxygen



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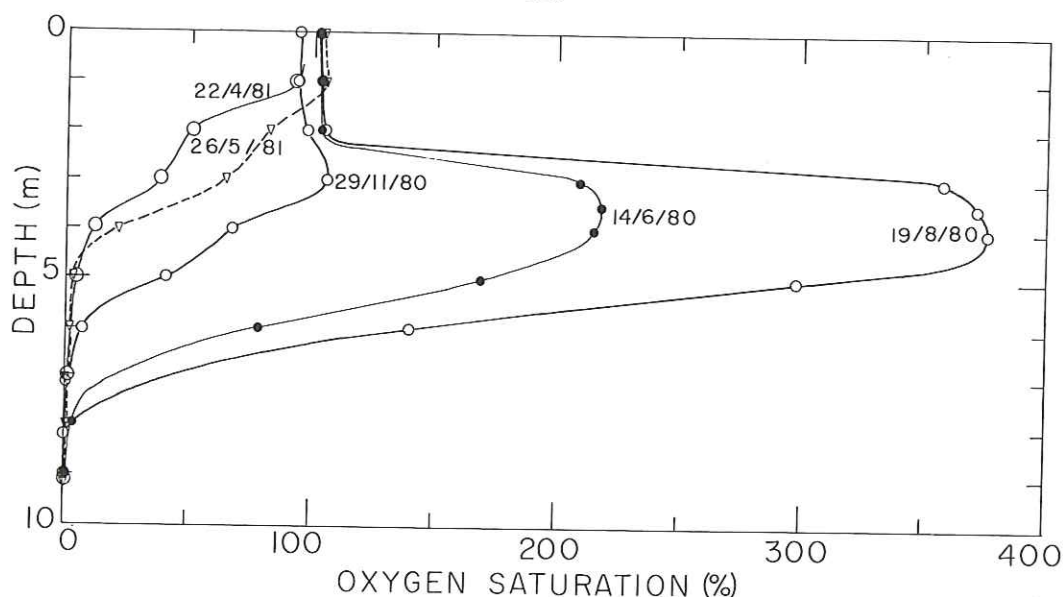


Fig. 6. Oxygen saturation (%) at various depths 14/6 1980, 19/8 1980, 29/11 1980, 22/4 1981 and 26/5 1981.

deficient. In the uppermost part of the saline layer, however, oxygen is not depleted, although saturation values become quite low at depths greater than 4 m at the end of winter (Fig. 6).

Sulfide was not measured in Lake Ólafsfjarðarvatn. It is assumed that its concentration must at any rate have been small, since H_2S smell was never detected in any of the samples except in April 1981. At that time a distinct H_2S smell was found of the sample from 9 m depth, and consequently the sulfide concentration at that depth may then have been considerable.

5. Productivity and chlorophyll *a*

Results of the primary production measurements are shown in Table 4. It will be noted that the *in situ* exposure time was 6 hours (10:00–16:00) in 1980, but 4 hours (12:00–16:00) in 1981, with additional samples exposed for 24 hours. According to Jónasson and Adalsteinsson (1979) the proportion of evening and night production (18:00–06:00) to morning and afternoon

production (06:00–18:00) in Lake Mývatn was on the average 34.6% in the period May 15 – June 20, 1974. This corresponds to an average of $(4 / (12 + 12 \cdot 0.346)) \cdot 100 = 24.8\%$ for the ratio $(P_4/P_{24}) \cdot 100$, which is in fair agreement with our results (26.5%) for Lake Ólafsfjarðarvatn.

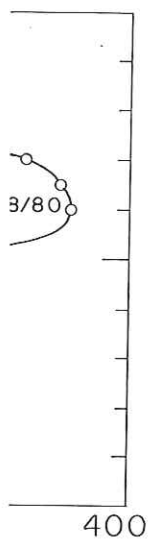
A common feature for both years, 1980 and 1981, was a conspicuous productivity in the upper part of the saline layer, whereas practically no production was observed in the fresh water layer on top (Table 4). This can probably be largely attributed to much more favourable nutrient conditions in the saline layer than in the fresh water (cf. Fig. 9), as will be discussed later. But also, due to the short residence time of the fresh water, phytoplankton growth may not have time to develop in the fresh water layer. At greater depths than 5 m the productivity decreased markedly and was quite small at 8 m.

The maximum values of productivity found in Lake Ólafsfjarðarvatn are similar to or higher than those observed in Lake Mývatn in May–June 1972–1974 but some-

TABLE 4
Results of productivity measurements in Lake Ólafsfjarðarvatn

14/6 1980				23/6 1981					
Bottles	Depth (m)	Exposure time: 6 hours Production (mgC m ⁻³ h ⁻¹)	Chlorophyll a (mg m ⁻³)	Bottles	Depth (m)	Exposure time: 4 hours Production (mgC m ⁻³ h ⁻¹)	Chlorophyll a (mg m ⁻³)	Exposure time: 24 hours Production (mgC m ⁻³ day ⁻¹)	P ₄ /P ₂₄ ¹ (%)
Light	1/2	—	—	Light	1/2	0.33	0.25	5.37	24.6
	1	0.05	—		1	—	—	—	—
	2	0.42	0.34		2	0.52	—	—	—
	2 1/2	5.09	—		2 1/2	—	—	—	—
	3	8.33	—		3	18.93	4.74	252.74	29.9
	3 1/2	9.73	5.99		3 1/2	—	—	—	—
	4	10.10	—		4	8.21	—	—	—
	5	9.74	8.04		5	—	5.10	—	—
	6	—	—		6	2.71	—	—	—
	8	1.60	1.67		8	0.27	6.03	4.30	25.1
Dark	1 1/2	0.04	—	Dark	1/2	0.03	—	—	—
	—	—	—		4	1.08	—	—	—
								Mean:	26.5

¹ P₄/P₂₄ denotes the ratio between the production measured in 4 hours and that measured in 24 hours.



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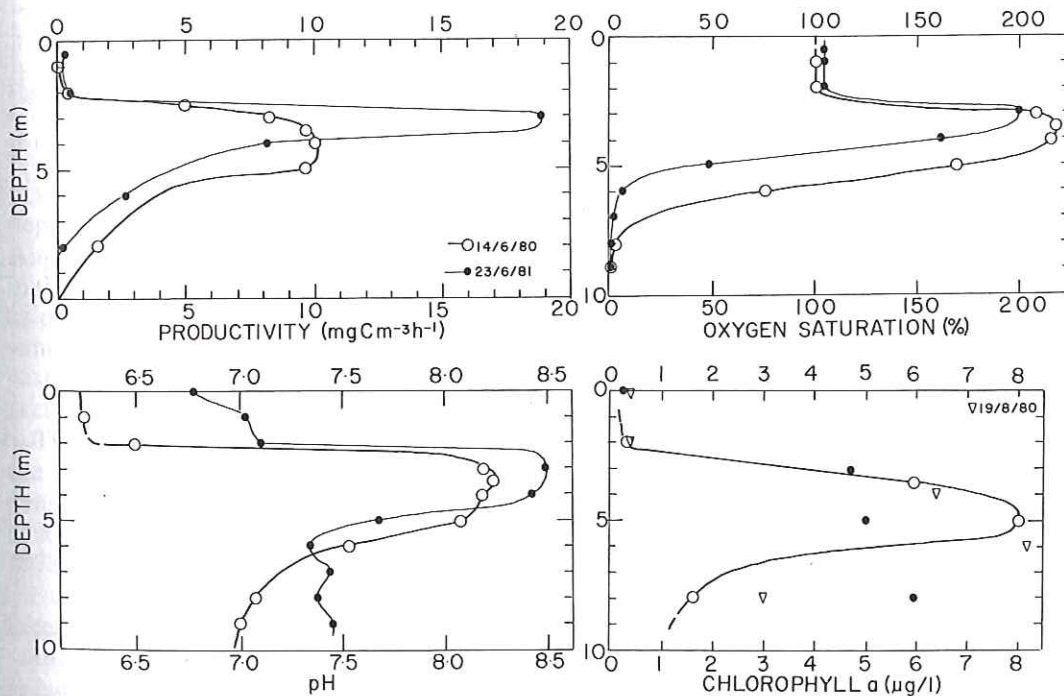


Fig. 7. Vertical distribution of productivity, oxygen saturation, pH and chlorophyll a, 14/6 1980 and 23/6 1981.

what lower than measured there in July-August (Jónasson and Adalsteinsson 1979).

In Fig. 7 a comparison is made of the vertical distribution of productivity and of some chemical properties which would be presumed to reveal similar features. Thus, as expected, the oxygen saturation formed a maximum which coincided with that of productivity and the same applied to pH.

In general there is a moderate correlation between chlorophyll a, which is a measure of the biomass, and phytoplankton productivity (e. g. Bougis 1974) and this is indicated in Fig. 7. The chlorophyll maximum, however, appears to be at a somewhat greater depth than the productivity maximum, and as seen in Fig. 8, relatively high chlorophyll concentrations were found in the deepest part of the lake, even at a depth of 8 m. Since productivity was very small at this depth, where oxygen is practically de-

pleted and light very limited, the high chlorophyll a values are difficult to explain, unless it is presumed that plant particles including chlorophyll a must have sunk down below the main production zone and accumulated at greater depths. This presumption is supported by the phaeo-pigment data (Table I) which reveal that with increasing depth there was an increased percentage of this pigment which is a decomposition product of chlorophyll a.

It is interesting to note that in the period June 22, 1982 - August 20, 1982 (Fig 8) there was a three- to fourfold increase in the total chlorophyll a concentration at depths below 5 m, whereas in the fresh water layer and the uppermost part of the saline layer the concentrations remained similar throughout the summer. A chlorophyll maximum was indicated in June and August 1982, but it did not appear in July. It may

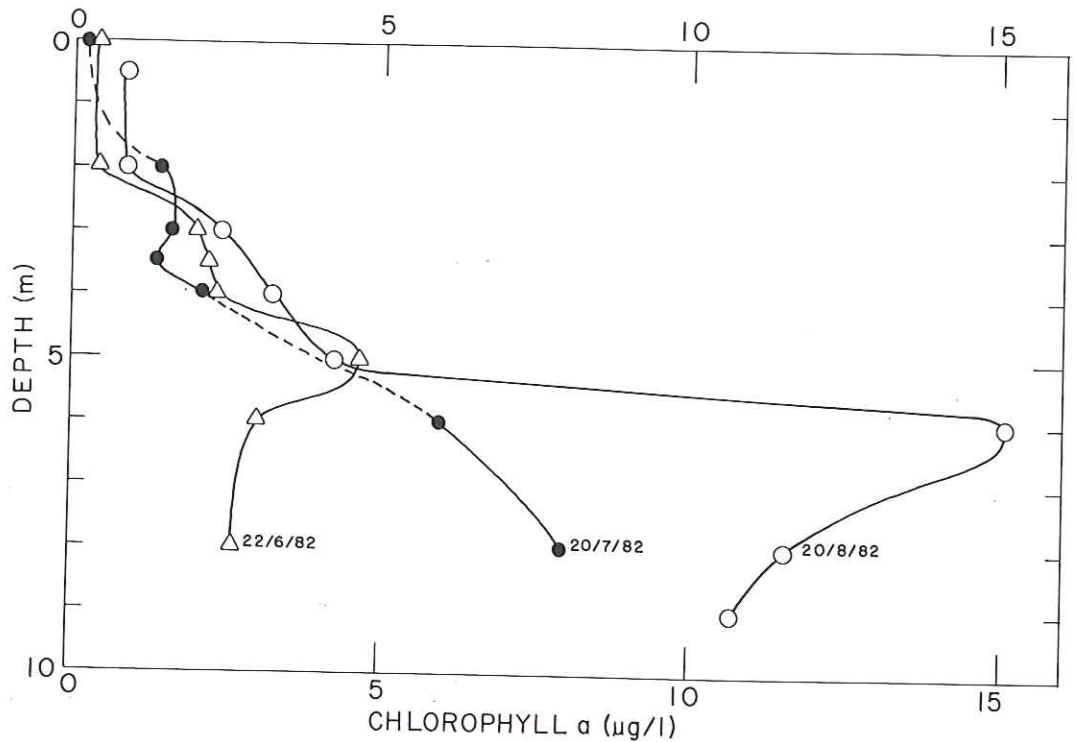


Fig. 8. Vertical distribution of chlorophyll a, June 22, July 20 and August 20 1982.

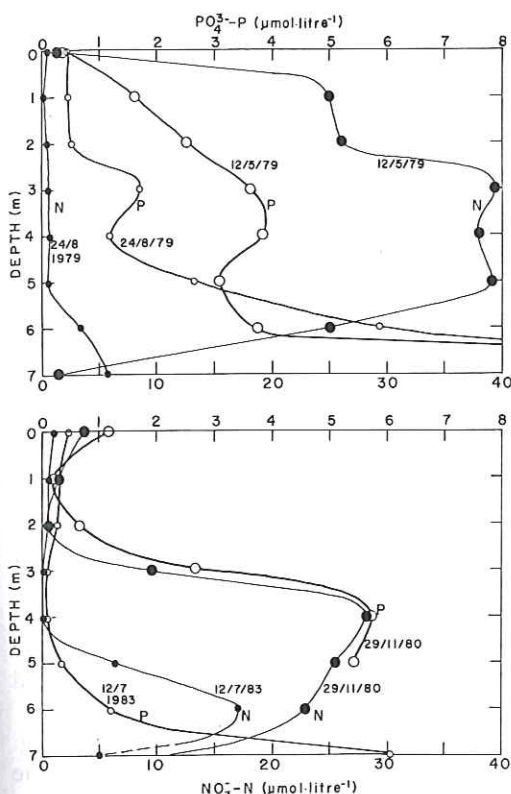


Fig. 9. Vertical distribution of inorganic phosphate and nitrate down to a depth of 7 m. a: 12/5 and 24/8 1979 and b: 29/11 1980 and 12/7 1983.

nevertheless have been present, e. g. at the depth of 5 m which was sampled in June and August but not in July. The marked increase in the chlorophyll concentration in the deeper part of the lake from June to July and a similar increase from July to August suggests the effect of a gradual and fairly even accumulation.

The chlorophyll concentrations found in the saline part of the lake were in general considerably higher than found for Lake Miklavatn (Stefánsson and Jóhannesson 1978) and also somewhat higher than measured in Nýpslón, Northeast Iceland (Stefánsson and Jóhannesson 1982). In Mývatn, however, much higher chlorophyll values were measured (Jónasson and Adalsteinsson 1979), but these were mostly due

to *Anabena* bloom in the lake. Before the *Anabena* bloom started, the chlorophyll concentration in Lake Mývatn was somewhat smaller than that measured in Lake Ólafsfjarðarvatn.

A comparison between the productivity distribution in June 1980 and 1981 reveals two main dissimilarities: a) The maximum was much more conspicuous in 1981 than in 1980 and b) the main production layer was thicker and located slightly deeper in 1980 than in 1981. A corresponding difference was also apparent in pH, but less so in oxygen saturation which proved to be slightly greater at 3 m depth in 1980 than in 1981 in spite of much greater productivity at that depth in the latter year than in the former. A possible explanation might be that as 1980 was a relatively warm year, the lake then became ice-free at an earlier date than in 1981 and therefore the plant production had started earlier and thus lasted longer when measured in 1980 than in 1981. Consequently, biologically produced oxygen may have accumulated to a greater extent in 1980 than in 1981 because of an earlier advent of spring.

6. Nutrients

Inorganic phosphate and nitrate concentrations showed marked variations both with depth and seasons (Fig. 9a,b). In the river water which feeds the fresh water layer of the lake, the concentrations were relatively low, both in summer and winter (Table I). In the saline layer, however, they were much greater. Within this layer the inorganic phosphate concentrations increased sharply with depth, while nitrate increased with depth only in the uppermost part of the saline layer, but then decreased to very low values, as expected, in the reducing environment of the oxygen deficient deeper water.

It is well known that in oxygen deficient

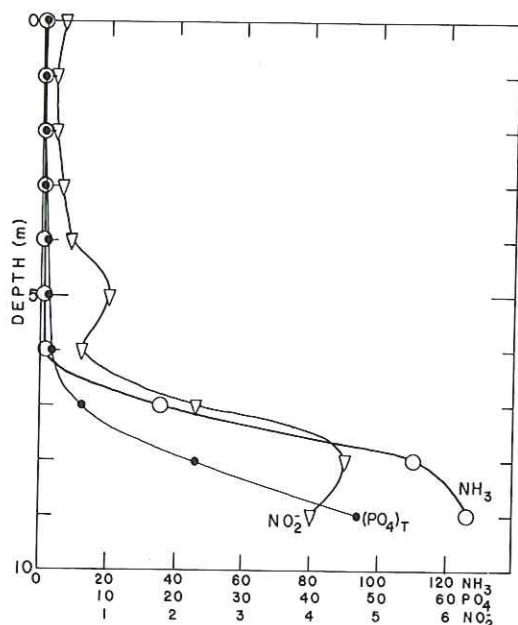


Fig. 10. Vertical distribution of nitrite, ammonia and total phosphorus 12/7 1983. Units: $\mu\text{mol l}^{-1}$.

environments nitrate reduction leads initially to the formation of nitrite and subsequently to the formation of ammonia and probably free nitrogen (see e. g. Deuser 1975). Similar denitrification processes clearly apply to Lake Ólafsfjarðarvatn. Thus nitrite concentrations were quite low in the fresh water and in the oxygenated part of the saline water (Table I, Fig. 10). At 7 m and deeper, where the oxygen was reduced to less than 0.25 ml l^{-1} , nitrite increased significantly, reaching a maximum of $4.5 \mu\text{mol l}^{-1}$ at a depth of 8 m. Very low and insignificant ammonia concentrations were measured in the surface layer and in the upper part of the saline layer down to a depth of 6 m but increased from there on quite sharply all the way to the bottom, where ammonia constituted by far the most important inorganic nitrogen component measured (Table I, Fig. 10). Other denitrification products besides nitrite and ammonia, such as free nitrogen or nitrous oxide, were not determined.

Concentrations of total phosphorus were somewhat (10–50%) higher than those of inorganic phosphorus in the fresh water, whereas in the upper part of the saline layer they were 3–5 times higher. It thus seems that in this productive part of the water column phosphorus is mainly in the organic and/or the particulate form in spring and summer. At greater depths both inorganic and total phosphorus increased dramatically, reaching a value of $50 \mu\text{mol l}^{-1}$ in the deepest part of the oxygen deficient layer.

It should be remarked that at 7 m and deeper, the concentration of the inorganic fraction as measured was slightly greater than the total phosphorus. This must be due to interference in the method used for inorganic phosphate (Murphy and Riley 1962), possibly the presence of sulfide (cf. Koroleff 1976; see also Jonge and Villerius 1980). The values reported for inorganic phosphorus at 7–9½ m in Table I should therefore be considered doubtful.

The silicate concentration (Table I, Fig. 11) proved to be highly variable both in the surface layers and within the saline deeper water. Usually, surface and near-surface concentrations were from 120 to $140 \mu\text{mol l}^{-1}$ which probably is the concentration range of the inflowing river water. Occasionally, however, considerable "dilution" appeared to have taken place in the fresh water, presumably as the result of melting of snow and ice, as we have stated before. This effect was particularly conspicuous in June 1980 (Fig. 11) but appeared also to a lesser extent in August 1980 and May 1981. In most cases an intermediate silicate minimum was found with concentrations ranging from $50\text{--}100 \mu\text{mol l}^{-1}$ near the depth of 3–4 m. This minimum must be caused by admixture of river water with sea water which in the adjacent sea area can be assumed to have a winter concentration of only $7\text{--}10 \mu\text{mol l}^{-1}$. At greater depths the silicate concentration increases markedly, no doubt due to dissolution from diatom

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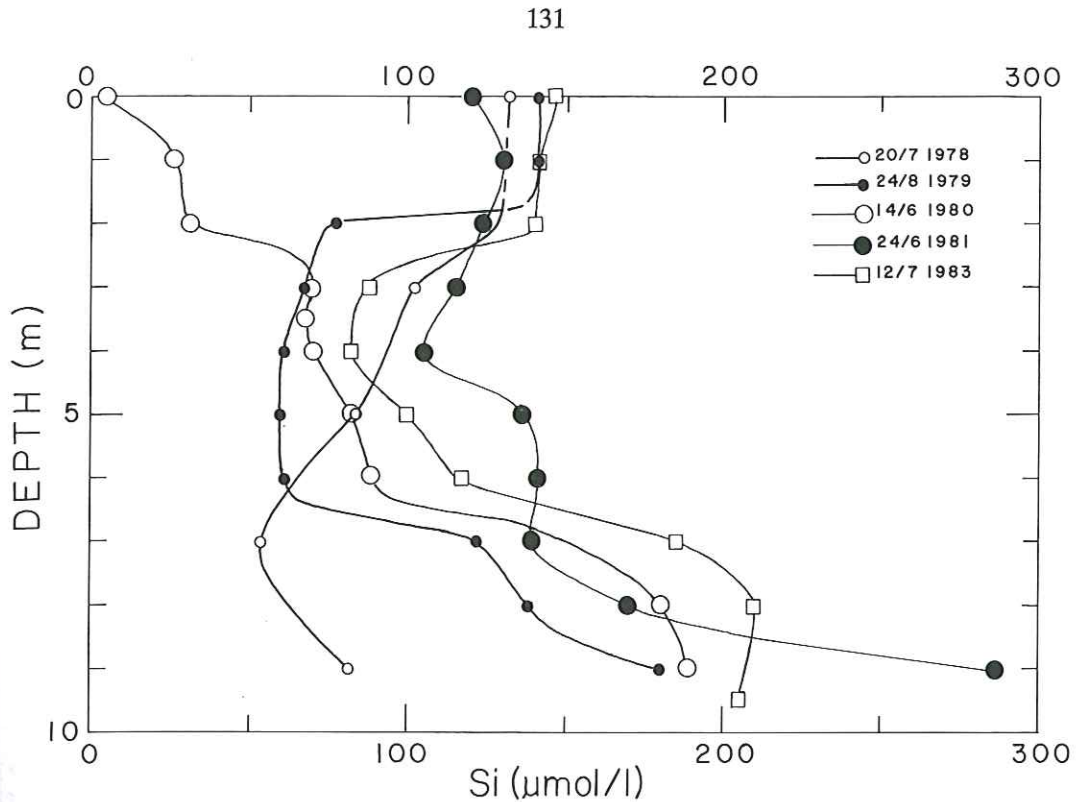


Fig. 11. Vertical distribution of silicate 20/7 1978, 24/8 1979, 14/6 1980, 24/6 1981 and 12/7 1983.

tests as organic matter is oxidized. Clearly, there are great year to year variations in the near-bottom silicate concentrations.

7. Alkalinity and pH

Alkalinity in aquatic environments is a measure of the concentration of anions of weak acids and is determined by the number of equivalents of strong acid needed to convert these anions in a unit volume or unit weight to the weak unionized acids. In normal sea water only the anions of the carbonate system, HCO₃⁻ and CO₃²⁻, and the borate, B(OH)₄⁻, will be quantitatively important. When river water is mixed with

normal sea water and there are no special processes acting in the water which provide additional buffering capacity, the alkalinity will be a simple linear function of salinity and is referred to as conservative alkalinity (A_c).

The mean alkalinity of the river Ólafsfjarðará, based on all observations from the period 1978–1983, is 0.23 meq l⁻¹, and the specific alkalinity of normal open sea water is assumed to be 0.126 meq l⁻¹ (Cl)⁻¹ (Skirrow 1975). If the salinity of the sea water entering Lake Ólafsfjarðarvatn is taken to be 35 and the salinity of the river to be zero, it follows that the conservative alkalinity of the water in the lake will be:

$$A_c = (0.126 \cdot 35/1.80655 - 0.23) \cdot S/35 + 0.23 = 0.063 \cdot S + 0.23 \quad (1)$$

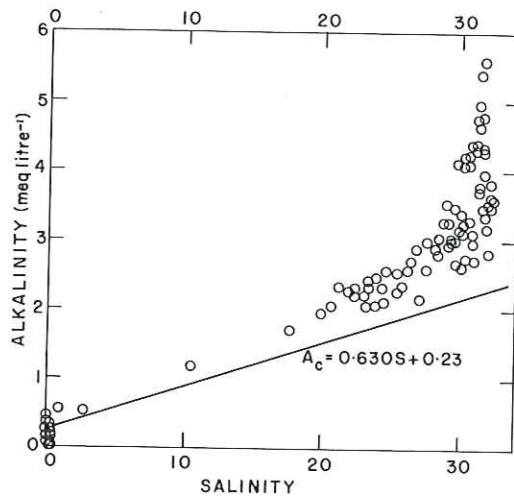


Fig. 12. Titration alkalinity as a function of salinity. The straight line denotes the conservative alkalinity.

In estuaries or lakes where mixing is intense and the deeper layers are rapidly renewed, such as in Nýpslón in Vopnafjörður (Stefánsson and Jóhannesson 1982), the total (titration) alkalinity consists of conservative alkalinity only. In oxygen-poor or anoxic environments, on the other hand, additional components give rise to excess alkalinity. Thus in the anoxic Lake Miklavatn, where high concentrations of sulfide are found in the underlying saline water, the conservative alkalinity constituted only about 30% of the total alkalinity (Einarsson and Stefánsson 1983). In Lake Ólafsfjarðarvatn (Fig. 12) the titration alkalinity was found to be slightly higher than the computed conservative alkalinity for salinities below 26. In the salinity range 26–30, excess alkalinity rose considerably, and in the deepest part, where salinity ranges between 30 and 32, the titration alkalinity reached values of 4–5 meq l⁻¹, and thus exceeded conservative alkalinity by a factor of about 2.

In the fresh near-surface layer pH was around 7.0 or somewhat less but increased in the saline layer (Table 1). We have seen (Fig. 7) that the pH distribution in June

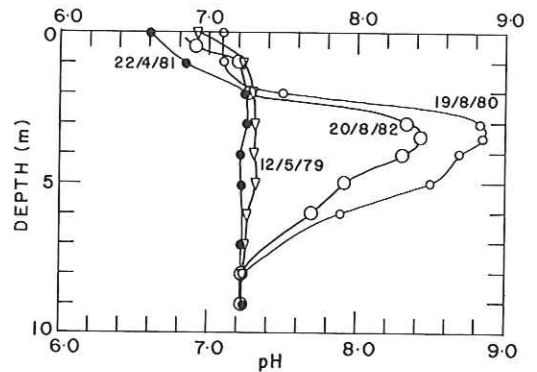


Fig. 13. Vertical distribution of pH 12/5 1979, 19/8 1980, 22/4 1981 and 20/8 1982.

1980 and 1981 revealed an intermediate maximum which practically coincided with the productivity maximum. In general, such a maximum in the upper part of the saline layer characterized the pH variations with depth during the summer season. In the main production layer the pH in summer ranged from 8.2–8.8. In winter (or while the lake is ice-covered) no intermediate maximum is observed. In the deepest part pH generally varied between 7.2 and 7.4. Examples of typical summer and winter distributions are illustrated in Fig. 13.

8. Particulate matter

Suspended matter was measured on one occasion, in June 1981. The lowest concentrations were found in the fresh water layer (Fig. 14). In the upper part of the saline layer the values were 2–3 times greater than at the surface and about 7 times greater at the depth of 8 m. These concentrations are comparable to those reported for the Ytri flói of Lake Mývatn (Ólafsson 1979).

Chemical analysis of the composition of the particulate matter (seston) was not carried out. In the surface layer, where primary production was negligible and the biomass quite small, the seston may have consisted largely of inorganic debris carried with the river. The increase in suspended matter at

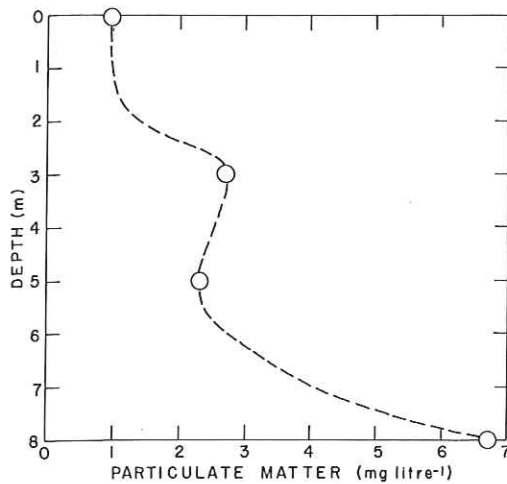


Fig. 14. Vertical distribution of suspended matter 24/6 1981.

intermediate depth was, however, probably due to organic particles formed through biological processes. In this context it is of interest to note that somewhat higher concentration of particulate matter was found at 3 m, where the productivity maximum occurred, than at 4 m depth. The high concentration found at 8 m most likely reflects the effects of accumulation of particles sinking from above, although stirring up from the bottom can not be excluded.

9. Light measurements

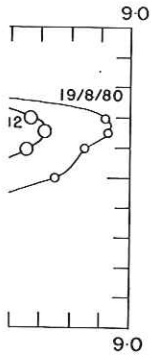
Transparency of the lake was measured with Secchi disk in spring and summer during the years 1980–1982. The results are given in Table 5. The values obtained ranged from 2.8 to 5.0 meters. They are much higher than found for Lake Mývatn (0.5–1.5), where wind stirring causes resuspension of sediments (Jónasson and Adalsteinsson 1979), but similar to the results found in Lake Miklavatn (3.0–3.7) in June 1977 (Stefánsson and Jóhannesson 1978). They are, however, considerably smaller than found in Nýpslón (6.75 m) in July 1981 (Stefánsson and Jóhannesson 1982) and much smaller than typically found in the ocean area around Iceland.

By using the simple formula $\kappa = 1.7/D$ (Poole and Atkins 1929), where D denotes the Secchi disk depth, extinction coefficients were estimated. In addition, they were determined for the Secchi disk depth by direct light measurements in June and July 1982 using a Quantum meter (400–700 nm). The values so obtained (Table 5, column 4) agreed very well with those estimated from the Secchi disk measurements. Thus the rather crude technique of Secchi disk measurements for estimating extinction

TABLE 5

Results of transparency measurements with a Secchi disk and light measurements at Secchi disk depth.

Date	Secchi disk depth (m)	Extinction Coefficient (κ)	
		From Secchi disk depth (m^{-1})	From direct light measurements at Secchi disk depth (m^{-1})
25/5/80	3.1–3.3	0.53	—
14–15/6/80	3.8–4.0	0.44	—
19/8/80	4.4	0.39	—
26/5/81	2.8	0.61	—
23–24/6/81	2.8–3.1	0.58	—
22/6/82	3.6–3.7	0.47	0.46
20/7/82	5.0	0.34	0.34
20/8/82	3.3–3.7	0.49	—



2/5 1979, 19/8

intermediate depth was, however, probably due to organic particles formed through biological processes. In this context it is of interest to note that somewhat higher concentration of particulate matter was found at 3 m, where the productivity maximum occurred, than at 4 m depth. The high concentration found at 8 m most likely reflects the effects of accumulation of particles sinking from above, although stirring up from the bottom can not be excluded.

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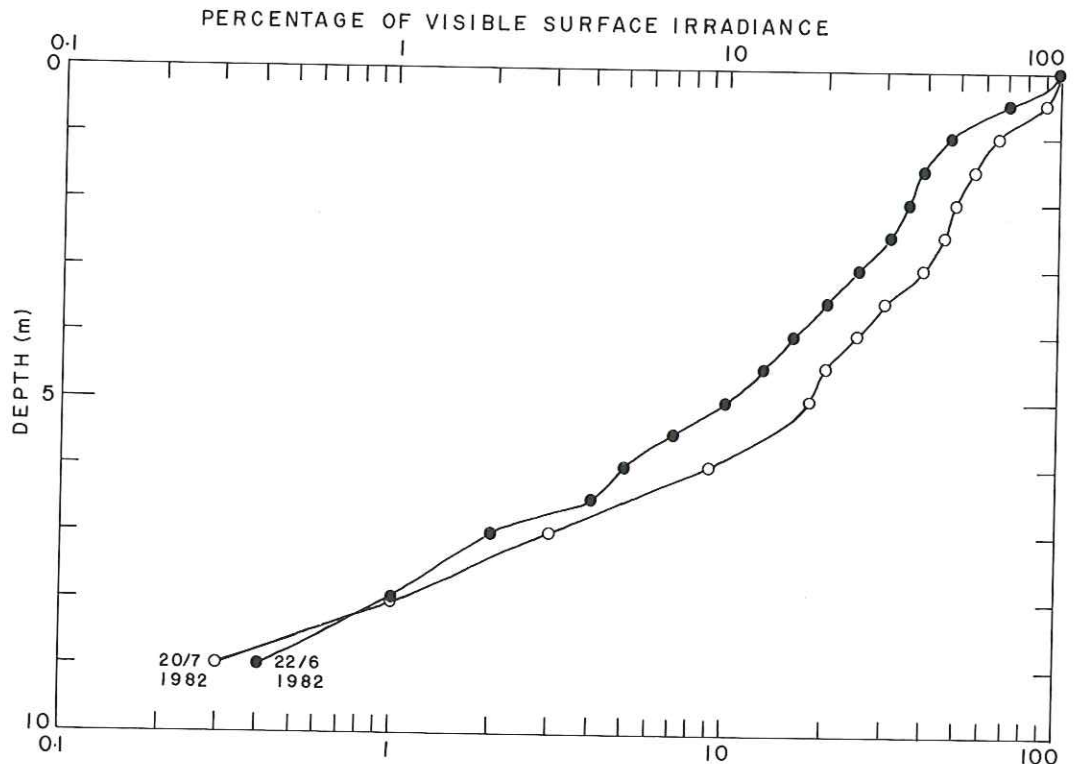


Fig. 15. Relative irradiance (%) at different depths 22/6 and 20/7 1982.

coefficients seem to yield satisfactory results for Lake Ólafsfjarðarvatn.

From the values of the relative irradiance of visible light at different depths (Fig. 15) the transparency can be calculated for different depth intervals (Table 6).

Comparison between the results of the two determinations, in June and July 1982, reveals that very marked light attenuation took place within the top 1 meter in June. Clearly, this was the main cause of smaller Secchi disk depth in June than in July. Presumably, suspended matter carried by the rivers at the time of maximum runoff was the main reason for this reduced transparency in the uppermost meter. On both occasions the maximum transparency was found near the interface between the two layers, i. e. at the lowest part of the fresh

water layer and in the uppermost part of the saline layer. With increasing depth the transparency was reduced, and near the bottom it amounted to only about one half of the value found near the top of the saline layer. It is interesting to note that slightly lower transparency values were observed for the 2–3 m interval in June than in July. This probably reflects greater biomass in this interval at the former observation time, as indicated by the chlorophyll *a* values. In July, however, transparency was significantly smaller in the deepest part of the lake than it was in June. This may have been caused by the accumulation of particles sinking from above. At 8 meters depth the irradiance of visible light was reduced to 1%.

TABLE 6

Transparency for different depth intervals in Lake Ólafsfjarðarvatn, computed from the ratio I_{z+1}/I_z . I_z is % of the surface illumination at depth z .

Depth Interval (m)	Mean transmission per meter (%)	
	22/6 1982	20/7 1982
0-1	47	65
1-2	74	75
2-3	71	80
3-4	64	64
4-5	63	72
5-6	50	50
6-7	40	33
7-8	50	33
8-9	40	30

DISCUSSION

1. Chemical Components

Information on the renewal of the saline deeper water of Lake Ólafsfjarðarvatn is scanty. It has been suggested that seepage may take place through the sand and gravel ridge between the lake and the sea, but in view of the considerable width of the ridge (0.5–0.7 km), we believe this source of sea water exchange to be insignificant. Therefore, the renewal of the saline water will presumably be very limited except for the occasions when sea water enters the lake through the estuary. Under normal conditions this does not take place, not even during spring tide. According to people with good knowledge this occurs when a spring tide coincides with a heavy swell during strong N or NE winds, especially in winter. Therefore, it would be expected that salinity and other chemical components of the deep water inside the lake would differ appreciably from year to year depending upon the frequency, duration and intensity of winter storms.

At the time this paper is written a conductivity meter, moored at the bottom of the estuary under the main bridge over the

seaside estuary, has been operating over a period of a few months. It is connected to a registering device. With this instrument which was installed by the staff of the Department of Physics, Science Institute, University of Iceland, it is hoped that information will be secured on the frequency of sea water intrusions into the lake and under which conditions they occur. However, on the basis of chemical measurements alone it is possible to confirm that significant year to year variations exist in the deepest part of the lake.

Near-bottom alkalinity (Fig. 16) increased markedly from the time these investigations started in 1978 until 1981 but decreased somewhat again in 1982. The salinity variations from the same period turned out to be nearly a mirror image of the alkalinity variations, i. e. a general decrease in salinity until 1981 and then an increase. This can be explained by the fact that with increasing age of the deep water its salinity will decrease due to a slight fresh water admixture, and at the same time alkalinity will increase as the water becomes more and more oxygen-deficient, leading to high ammonia concentrations in the deepest layers and possibly reduction of sulfate. Corresponding changes took place in the oxygen saturation values in the deeper layers of the lake. At

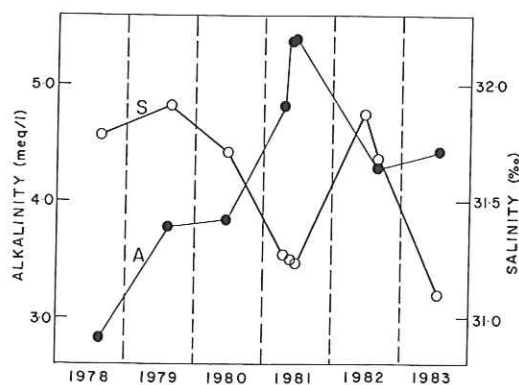


Fig. 16. Alkalinity and salinity at 9 m depth in different years.

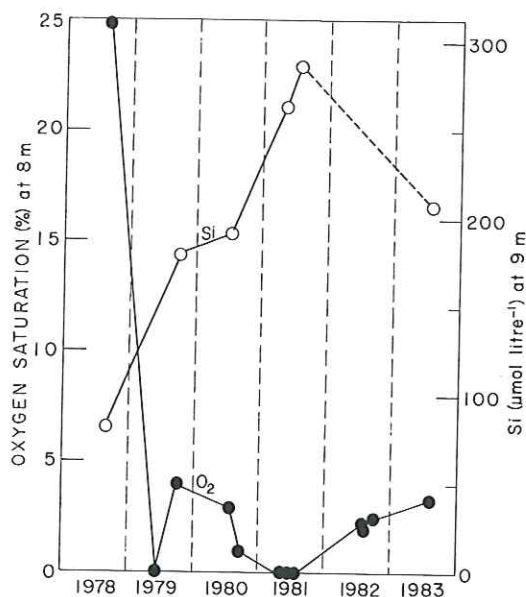


Fig. 17. Oxygen saturation (%) and concentration of silicate at the depth of 8 m in different years.

the depth of 8 m (Fig. 17) it dropped from 25% in 1978 to zero in 1979, then increased slightly (~ 2–4%)¹ in 1980 but dropped again to zero in 1981. Finally, a slight but measurable increase was found in 1982 and 1983. At 9 m depth the saturation was 13% in 1978 but undetectable at this depth in the following years. Parallel variations were found in the silicate concentration (Fig. 11), i. e. increasing silicate with decreasing oxygen, which can be attributed to dissolution of silica from diatom tests.

The high concentrations of excess alkalinity (1–2½ meq l⁻¹, see Fig. 12) for salinities exceeding 30, are somewhat puzzling. One source is certainly ammonia which was found in relatively high concentrations in the reducing environment near the bottom (Fig. 10), but it accounts for only 5–10% of

¹ These values may represent an upper limit, since observations have indicated that the true oxygen concentrations may be slightly lower than determined by the Winkler method at very low concentrations (Broenkow and Cline 1969; Cline and Richards 1972).

the excess alkalinity. Other sources may include:

- Reduction of sulfate in the anoxic environment below 8–9 m. This reduction might initially lead to the formation of sulfide, which on mixing with oxygenated water at higher levels and with sea water entering the lake through the channel, might form other sulphur compounds, such as sulfite, thiosulfate or even free sulphur (Stefánsson and Jóhannesson 1978; Einarsson and Stefánsson 1983). Sulfide might also be lost by precipitation as iron sulfide (Berner 1974; Goldhaber and Kaplan 1974).
- Dissolution of carbonate from decomposing organisms may have increased the concentration of carbonate forms in the lake.

The silicon of biogenic origin was estimated using the mean value of 135 µmol l⁻¹ for the silicate concentration of the river water and assigning a value of 7 µmol l⁻¹ to the originally inflowing sea water. Salinity of "pure" sea water was taken as 35.0 and the salinity of the river water was assumed to be zero. Using these data the following equation is derived (cf. Einarsson and Stefánsson 1983) relating the original silica concentration to salinity:

$$Si_{orig.} = -3.66 \cdot S + 135 \quad (2)$$

and from this relationship the silica of biogenic origin was computed as:

$$Si_{biog.} = Si_{total} - Si_{orig.} \quad (3)$$

Excess alkalinity was plotted against biogenic silica so found (Fig. 18). The computed ratio of change was $\Delta A : \Delta Si = 122 : 11.1$ which is practically identical to that found for Lake Miklavatn (Einarsson and Stefánsson 1983), suggesting similar sources of excess alkalinity for both lakes.

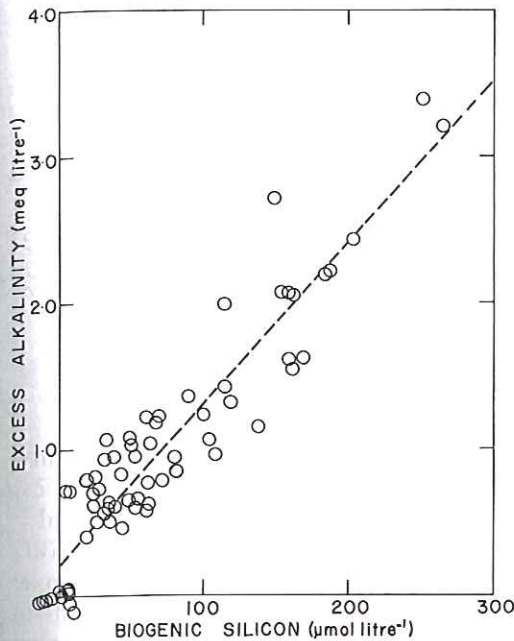
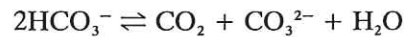


Fig. 18. Excess alkalinity as a function of biogenic silicon. Computed slope of line is $\Delta A: \Delta Si = 122:11.1$.

As was shown earlier (Table 4, Fig. 4) the productivity profiles were characterized by very low values in the fresh water layer, a conspicuous maximum in the upper part of the saline layer and low values in the deepest part. These features of the productivity distribution can be attributed to chemical and physical conditions at various depths of the lake. Thus in 1979 nutrient levels were quite low in the surface layers, even at the end of winter (Fig. 9a), while both nitrate and phosphate concentrations were high in the upper part of the saline layer. Nutrient data are not available for late winter 1980, but phosphate concentrations measured in both late winter and spring of 1981 revealed similar distribution as in 1979. Furthermore, light attenuation seems to influence the depth and the thickness of the main productivity layer. Thus the transparency was considerably greater in June 1980 than in June 1981, and this may explain why the productivity maximum appeared at a

smaller depth and was narrower in 1981 than in 1980. It is, however, remarkable that according to Table 4 the rate of assimilation ($\text{mg C m}^{-3}\text{h}^{-1} / (\text{mg chlorophyll a m}^{-3}) = 18.93/4.74 = 4.0$) was considerably greater at 3 m in 1981 than at 3½ m in 1980 ($9.73/5.99 = 1.6$). Conceivably, this was due to greater incident illumination on the day of observation in 1981 than in 1980. It is also possible that the percentage of phaeo-pigment was greater at 3½ m in 1980 than at 3 m in 1981. In the deeper layers the productivity is suppressed by reduced light ($< 1\%$ for depths exceeding 8 m) and very low oxygen concentration.

Since the plant production involves an uptake of CO_2 from the water, the equilibrium



will be displaced to the right as the result of the photosynthesis. This will cause the concentration of carbonate to increase and consequently also pH. Therefore, pH changes in the saline layer should provide a useful although rough indication of the biomass produced. This is clearly suggested by Fig. 19 where pH is plotted as a function of the apparent oxygen production.

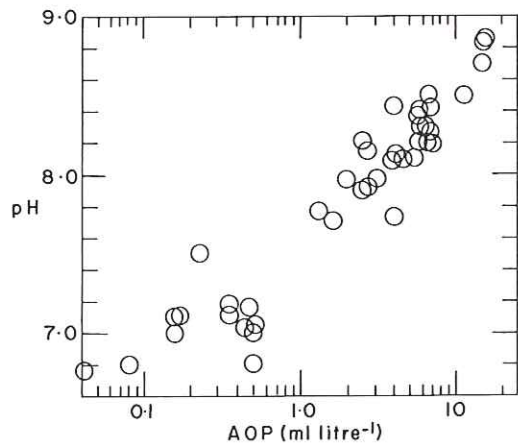


Fig. 19. pH as a function of the apparent oxygen production (AOP), for $\text{pH} \geq 6.7$ and $\text{AOP} \geq 0.04 \text{ ml l}^{-1}$.

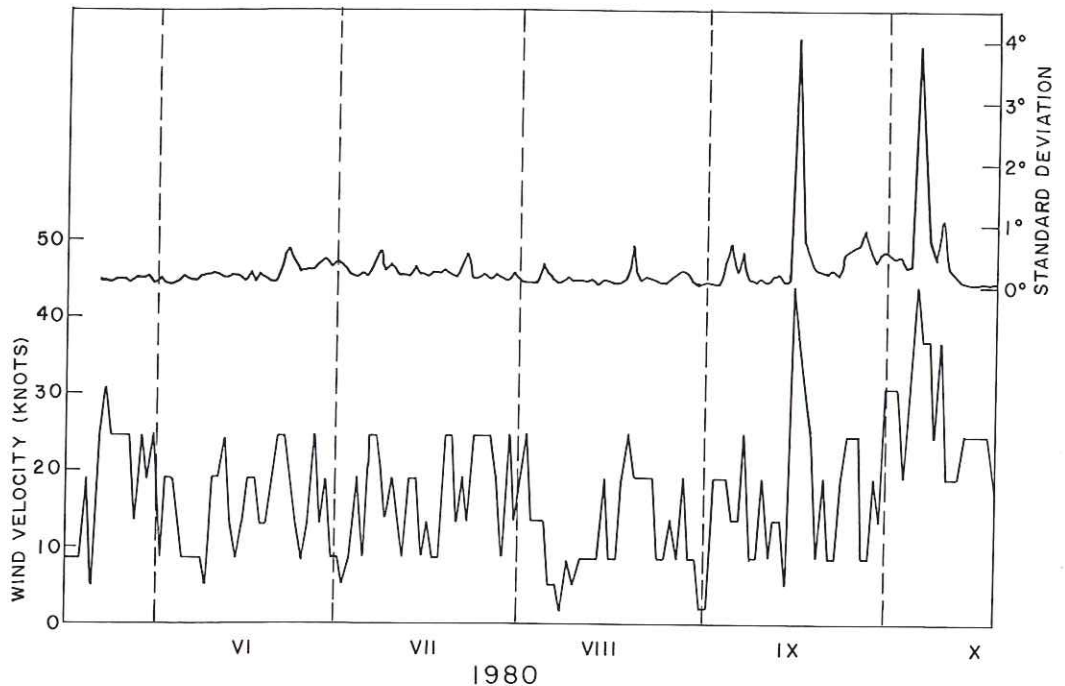


Fig. 20. Daily variations in maximum wind velocity and standard deviations of mean daily temperature at a depth of $3\frac{1}{2}$ m during the one year period 20/5 1980–21/5 1981.

2. Heat budget

The results of the short-periodic temperature measurements at a depth of $3\frac{1}{2}$ m which we have described, showed clearly that cooling in the uppermost part of the saline layer does not by any means proceed uniformly and gradually. On the contrary, it takes place intermittently and suddenly. This suggests that the cooling is mainly brought about by wind mixing. However, in view of the fact that the intermediate temperature maximum can persist for months in the upper part of the saline layer, it was postulated that appreciable vertical mixing due to wind action does not penetrate the halocline until a considerable wind velocity is attained.

To test this hypothesis variations in wind speed were plotted against the standard deviation of temperature at the depth of the sensor, i. e. at $3\frac{1}{2}$ m. It will be seen (Fig 20)

that in general temperature changes correspond rather closely to variations in maximum wind speeds. This relationship applies in particular at great wind speeds. Thus the abrupt temperature drop on September 16, and again on October 6–7, 1980, coincided with the largest wind speeds recorded. This result strongly supports the postulate that the autumn cooling in the upper part of the saline layer is largely wind-induced. However, it is also seen that wind mixing down to the level of $3\frac{1}{2}$ m does not become appreciable until the wind has reached a speed of 30–40 knots. In winter, when the lake is frozen, the wind will no longer cause stirring up of the water. During that season the temperature decrease will therefore depend largely on thermal diffusion through the ice, a process which will be slow. This is indicated in Fig. 4 which reveals that during the 4 months period January–May 1981 the

TABLE 7

Measured net heat flux to saline layer as compared to estimated heat flux.

Period	Mean total radiation day ⁻¹ - 7% (cal./cm ²)	% total radiation at 2 1/2 m	Estimated heat flux over period (cal./cm ²)	Measured net heat flux	
				(cal./cm ²)	As % of estimated
21/5-14/6 1980	458	12.3	1408	1310	93
15/6-19/8 1980	372	15.1	3654	2004	55
26/5-24/6 1981	389	10.1	1165	1200	103
22/6-20/7 1982	407	15.8	1800	1660	92
21/7-20/8 1982	311	15.4	1487	-110	-7

temperature decreased gradually by only 1.5°C.

Investigations by Smith and Sinclair (1972) have indicated that in homogeneous lakes the mixed depth equals about one half of the wavelength. From their data on relationships between wind speed, fetch and depth of water disturbed by wave action (loc. cit. Table 1), it can be inferred that for a fetch of about 2 km which is roughly the maximum fetch for St. 1 in Lake Ólafsfjarðarvatn (see Fig. 1), a wave-mixed depth of roughly 3 m will result from wind speed of about 18-19 knots and a depth of roughly 3 1/2 m from a wind velocity of 22 knots. These are considerably smaller wind speeds than seem needed to cause an appreciable mixing and cooling of the upper part of the saline layer (Fig. 20). However, it is clear that the density discontinuity at the interface between the two layers will strongly suppress vertical mixing, and the upper layer will tend to slide over the lower (see e. g. Neumann and Pierson, p. 197). Consequently, eddy viscosity coefficients below the pycnocline will be greatly reduced and much stronger winds needed to produce wind mixing reaching down into the saline layer.

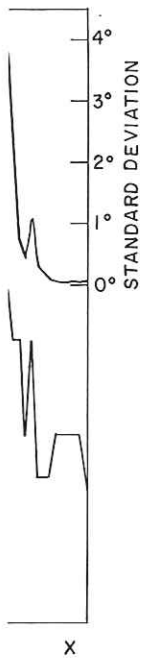
In view of the observed correlation between winds and temperature variations, it was attempted to estimate how insulating the halocline is under different conditions. For this purpose a comparison was made

between the estimated heat flux into the saline layer and the measured heat flux from temperature measurements at various depths. Information on solar radiation measured at Akureyri, some 50 km from Ólafsfjörður, was obtained from the Icelandic Meteorological Office. Visible light was assumed to amount to 42% of the total radiation (Ivanoff 1975, p. 24) and albedo 7%. From direct light measurements as well as Secchi disk depths the fraction of the radiation reaching a depth of 2 1/2 m was computed, and from this the heat flux into the saline layer in different periods (Table 7, column 4). These values can be compared to the measured net heat increase between two observation periods, ΔQ , based on actual temperature measurements at different depths (Table 7, column 5), using the formula

$$\Delta Q = \sum_i \frac{1}{2} [(T_{j+1,z_i} - T_{j,z_i}) + (T_{j+1,z_{i+1}} - T_{j,z_{i+1}})] c_p (z_{i+1} - z_i) \quad (4)$$

where T_{j,z_i} denotes temperature measured at depth z_i during observation j , and c_p stands for specific heat of the water.

It should be emphasized that the estimated values for the heat flux are, due to various uncertainties, at best rough approximations. Nevertheless, it will be clear (Table 7,



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columns 4 and 5) that for the shorter periods the two sets of values show a fair agreement which suggests that the halocline must be a good insulator. However, it is also seen that the halocline is far from preventing heat loss at all times from the saline layer. Thus it is indicated that during the relatively long period between June 15 and August 19, 1980, only about 50% of the heat received was retained within the saline layer, and in the period July 20 to August 20, 1982, there was a net heat loss, contrary to what was found for the other periods.

These findings are, at least partly, in agreement with meteorological data. A

comparison of wind speeds (both means and maxima) for the months June, July and August during the years 1980, 1981 and 1982 (Table 8) reveals that in these three months the weather was definitely calmer in 1980 than in the other two years. This is consistent with higher temperatures of the saline layer that year than in 1981 and 1982. On the other hand, winds were considerably stronger in July-August 1982 than in either 1980 or 1981. According to information received from the Meteorological Office very strong winds were observed on August 18-19 at Siglunes, about 18 km away, where the maximum wind speeds were up to 55 knots.

TABLE 8
*Comparison between wind velocities at Siglunes
(18 km from Ólafsfjörður), in the years 1980, 1981 and 1982.*

JUNE				
Year	Wind velocity		Number of days with	
	Mean (knots)	Maximum (knots)	≥ 24 knots	≥ 30 knots
1980	9.7	24	4	0
1981	9.8	30	9	3
1982	10.3	41	9	2

JULY				
Year	Wind velocity		Number of days with	
	Mean (knots)	Maximum (knots)	≥ 24 knots	≥ 30 knots
1980	8.4	24	8	0
1981	9.8	30	9	4
1982	12.0	37	8	6

AUGUST				
Year	Wind velocity		Number of days with	
	Mean (knots)	Maximum (knots)	≥ 24 knots	≥ 30 knots
1980	5.6	24	2	0
1981	7.1	37	8	3
1982	8.8	(55)	7	3

Residents of the Ólafsfjörður town did not recall gales of this magnitude at the time in question, but they confirmed nevertheless strong northerly winds, exceeding 30 knots in the period August 15–18. These winds may have caused the observed heat loss during the period July–August 1982.

3. Oxygen budget

An estimation of the oxygen budget of the saline layer can be made in a similar way as was done for the heat flux, by computing changes in the oxygen content for different periods using data from Table 1. The results so obtained are shown in Table 9. It will be seen that the oxygen production in spring and summer is on the average 0.4–0.7 litres $m^{-2} day^{-1}$. In winter the process is reversed: oxygen is consumed due to decomposition and oxidation of organic matter, and possibly some loss of oxygen to the atmosphere takes place due to wind mixing in the autumn. In the period August 19 to November 29, 1980 the oxygen loss amounted to 0.5–0.6 litres $m^{-2} day^{-1}$. During the latter half of winter, i. e. in the period November 29 to April 22, 1981, the oxygen consumption was quite slow, or at the rate of only 0.08 litres $m^{-2} day^{-1}$. In the last column of Table 9 the estimated oxygen production is shown, based on the primary production measurements of June 14, 1980 and June 23, 1981, using the simple formula

$$O_2(\text{ml l}^{-1}) = (\text{mg C l}^{-1}) \cdot (11.2/12) \cdot (276/106) \quad (5)$$

These values are 2–3 times higher than corresponding figures in column 4 of Table 9. A possible interpretation might be that the primary production and consequently the oxygen produced was greater in June when measured than later in the summer. An alternative interpretation would be that 50–70% of the oxygen produced was lost from

the saline layer, presumably in the form of gas bubbles. Until more data are available it can not be decided whether either of these explanations apply.

Another way of estimating the oxygen budget within the productive part of the saline layer, is to compare the measured rate of change of oxygen with the rate of change of nutrients. We will consider changes in the upper part of the saline layer (2–4 m) from May 12 to August 24, 1979. During this period the phosphate concentration decreased by about 2.2 $\mu\text{mol l}^{-1}$, and the nitrate concentration by about 34 $\mu\text{mol l}^{-1}$, which gives a $\Delta N : \Delta P$ ratio of 15.5 : 1. Using the model of Richards et al. (1965) this change in nutrient concentrations corresponds to a change of 303 $\mu\text{mol l}^{-1}$ of dissolved oxygen. The actual observed net oxygen increase during the period in question was 420 $\mu\text{mol l}^{-1}$ or almost 40% higher. A possible explanation of this discrepancy might be that the observation in May 1979 was made under ice, preventing equilibrium between the lake surface and the overlying air. At the time when the ice melted and the plant production began, oxygen may have diffused from above down into the oxygen deficient layer. This would add to the oxygen produced by photosynthesis. On the other hand, it is by no means certain that the stoichiometric relationship proposed by Richards et al. applies exactly to Lake Ólafsfjarðarvatn. To obtain a reliable measure of the oxygen budget by this approach, observations should be made just before and immediately after the lake becomes ice-free.

The conclusion we can draw from these rough estimations is that at any rate a large portion of the oxygen produced by photosynthesis in the lake may be preserved for a considerable length of time. In view of this, oxygen measurements in the lake may yield a rough indication of the amount of organic matter produced, provided that the wind velocity remains below a critical level.

TABLE 9
The oxygen budget

Period	No. of days	Net oxygen increase		Estimated oxygen production	
		during period litres/m ²	mean per day litres/m ²	Date	litres/m ² per day
15/6-19/8 '80	65	25.8	0.40	14/6 '80	1.30
20/8-29/11 '80	102	-54.8	-0.54		
30/11'80-22/4 '81	144	-11.3	-0.08		
23/4-26/5 '81	34	2.8	0.08		
27/5-24/6 '81	29	20.0	0.69	23/6 '81	1.30
22/6-20/7 '82	29	17.3	0.60		
21/7-20/8 '82	31	3.3	0.11		

4. Thermal diffusion within the saline layer

From the temperature profiles a rough calculation can be made of the vertical eddy coefficients, by simply computing the rate of increase in heat content of a column of water of unit area below a certain depth over a given period and dividing by the value of the mean vertical thermal gradient at the depth in question.

As an example we will consider the two temperature profiles from June 14 and August 19, 1980 (Fig. 4). To estimate the vertical heat flux at a depth of 5 m we extrapolate the temperature curves in Fig. 4 to the bottom (10 m). Then from the diffusion equation:

$$\partial T/\partial t = \partial/\partial z (K_z \partial T/\partial z) \quad (6)$$

it follows that

$$\frac{\partial}{\partial t} \int_{\text{bottom}}^{-5} T dz \approx \frac{1}{\Delta t} \int_{-10}^{-5} (T_2 - T_1) dz = K_z \overline{(\partial T/\partial z)}_{5m} \quad (7)$$

where T denotes temperature, t time, z depth, K_z the vertical eddy diffusion coefficient and $\overline{(\partial T/\partial z)}_{5m}$ the mean vertical temperature gradient at 5 m depth. The sub-

scripts 1 and 2 refer to the observation dates 14/6 and 19/8, 1980 respectively.

In this particular example the integral

$$\int_{-10}^{-5} (T_2 - T_1) dz$$

was evaluated from the area between the two temperature curves from 5 m depth down to the bottom (16.3 m), $\overline{(\partial T/\partial z)}$ was found to be about 3.0°m^{-1} and $t \approx 66$ days or $5.7 \cdot 10^6$ seconds.

Then:

$$(K_z)_{5m} = \frac{\int_{-10}^{-5} (T_2 - T_1) dz}{\Delta t \cdot \overline{(\partial T/\partial z)}_{5m}} = \frac{16.3\text{m}}{5.7 \cdot 10^6\text{s} \cdot 3.0^\circ\text{m}^{-1}} = 0.95 \cdot 10^{-6}\text{m}^2\text{s}^{-1}$$

This value for K_z is 1-2 orders of magnitude smaller than generally estimated for the ocean (Garrett 1979) but is comparable to that computed for small and shallow lakes (Hutchinson 1957, p. 472-473).

CONCLUDING REMARKS

Lake Ólafsfjarðarvatn is unique among Icelandic lakes, both with regard to thermal conditions and chemistry. Due to suitably thin surface layer, appreciable light energy penetrates down into the upper part of the saline layer which normally is well insulated from the fresh water on top by a strong halocline. This energy causes a temperature maximum to develop at 3–4 m depth in summer, providing a submarine "greenhouse" with temperatures of up to 20°C. The renewal of the saline deeper water is limited but nevertheless sufficient to prevent oxygen depletion in the upper part of the saline layer in winter. These conditions might, as stated in the introduction to this paper, offer unique possibilities for mariculture of warm water species, or alternatively for accelerating the growth of species native to this area, such as salmon. Experiments are already under way to examine these possibilities. Higher saturation values of oxygen have been measured in the saline layer of the lake than reported in the literature. A future task of great interest would be to study the chemical and biological effects of these exceptional oxygen concentrations.

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APPENDIX

TABLE I
Distribution of temperature and chemical properties in Lake Ólafsfjörðarvatn 1978-1983.

Date	St.	Depth (m)	Temp (°C)	Salinity	Oxygen		Titration alk (meq/l)	pH	Reactive phosphate		Si (µmol/l)	NO ₃ ⁻	NO ₂ ⁻	Chlorophyll a (µg/l)
					(ml/l)	(% sat.)			phosphate	phosphate				
20/7 1978	1	0	7.90	0.1	8.82	106	0.22	6.80	0.42	132	0.2	0.0	—	
		3	12.12	11.16	12.78	180	—	8.37	1.47	102	—	0.5	—	
	2	5	10.60	27.39	9.56	146	3.03	7.93	5.64	84	—	0.7	—	
		7	7.24	30.68	3.86	56	2.74	7.65	7.24	54	7.6	0.8	—	
	3	9	6.20	31.78	0.92	13	2.85	7.37	—	82	7.2	1.1	—	
		0	5.11	0.0	8.94	100	0.24	6.76	0.63	139	0.4	0.1	—	
	4	2	4.80	0.1	9.55	107	0.19	6.72	0.66	139	0.4	0.1	—	
		0	7.40	0.0	9.26	110	0.25	6.67	0.52	135	0.3	0.1	—	
	5	2	8.50	—	—	—	—	—	—	—	—	—	—	—
		3	13.66	—	—	—	—	—	—	—	—	—	—	—
		3	12.45	—	—	—	—	—	—	—	—	—	—	—
		4	13.77	—	—	—	—	—	—	—	—	—	—	—
		5	11.22	—	—	—	—	—	—	—	—	—	—	—
		0	8.19	0.1	8.83	107	—	—	—	0.60	134	0.4	0.1	—
2		8.41	—	—	—	—	—	—	—	—	—	—	—	
3		11.46	14.01	12.90	185	—	—	8.40	1.57	103	0.5	0.5	—	
4		13.35	—	—	—	—	—	—	—	—	—	—	—	
5 1/2		10.76	29.29	10.63	165	3.15	—	8.12	7.91	61	0.6	0.8	—	
6	3 1/2	14.36	23.81	—	—	2.55	8.20	2.25	72	0.6	0.6	—		
	0	7.86	0.1	8.85	106	0.22	6.99	0.53	127	0.5	0.1	—		
12/5 1979	1	1.2	7.83	0.1	8.32	100	0.23	7.16	0.52	133	0.6	0.2	—	
		0	0.00	0.3	8.26	81	0.47	6.92	0.37	102	1.3	—	—	
	2	1.1	1.51	19.90	2.62	31	2.00	7.24	1.63	98	25.0	—	—	
		3	3.80	25.27	2.73	35	2.29	7.30	2.51	88	26.0	—	—	
	3	5.58	27.30	1.84	25	2.60	7.31	7.31	3.63	84	39.5	—	—	
		4	6.21	29.39	1.14	16	2.66	7.31	3.83	87	38.0	—	—	
	4	5	6.64	30.06	1.01	14	2.76	7.32	3.08	79	39.5	—	—	
		6	6.68	30.66	0.14	2	3.01	7.27	3.78	105	25.0	—	—	
	7	6.55	31.40	0.08	1	3.36	7.25	7.25	18.1	159	1.5	—	—	
		8.2	6.28	31.64	0.00	0	3.19	7.24	20.3	—	—	—	—	

24/8	7	0	1.10	0.9	—	—	0.59	6.91	0.72	151	4.7	—	—
1979	R ¹	0	3.40	0.1	9.62	103	0.47	7.02	0.25	132	1.1	—	—
	1	0	8.25	0.2	—	—	0.25	6.72	0.44	141	0.5	0.2	0.39
		1/2	8.45	—	8.73	107	—	—	—	—	—	—	—
		1	8.43	0.2	8.68	106	0.25	—	0.46	141	0.1	0.2	—
		2	14.00	22.24	11.65	186	2.33	8.10	0.54	78	0.5	0.5	—
		3	14.64	25.98	12.47	206	2.60	8.30	1.73	68	0.5	0.4	—
		4	13.78	28.69	9.96	164	2.98	8.08	1.17	61	0.7	0.5	4.56
		5	11.15	29.93	8.35	131	2.72	7.97	2.63	60	0.6	0.5	—
		6	9.62	30.65	4.10	63	3.12	7.59	5.87	61	3.5	0.6	—
		7	8.50	31.27	0.50	8	3.44	7.38	10.1	122	5.7	0.7	5.85
		8	7.70	31.71	0.25	4	3.55	7.32	—	138	1.1	0.8	—
		9	7.14	31.91	0.10	1	3.80	7.28	19.7	180	0.8	0.8	—
	7 ^a	1/2	8.65	0.2	—	—	—	—	0.43	140	0.1	0.3	—
		1 1/2	8.40	0.2	—	—	—	—	0.40	140	0.0	0.2	—
	7 ^b	1/2	8.60	—	—	—	—	—	0.57	142	0.2	0.4	—
	R ¹	0	7.60	—	—	—	—	—	0.43	151	3.1	0.3	—
	H ^c	—	—	—	—	—	—	—	0.08	1134	0.6	0.2	—

l R: River Ólafsfjörðará.

a At high tide.

b At low tide.

c H: Sample of hot spring water, taken from a faucet in the Ólafsfjörður town.

TABLE I (continued)

Date	St.	Depth (m)	Temp (°C)	Salinity	Oxygen		Tita- tion alk (meq/l)	pH	Reactive		Chloro- phyll <i>a</i> ($\mu\text{g/l}$)
					(ml/l)	(% sat.)			phos- phate	Si ($\mu\text{mol/l}$)	
21/5 1980	1	0	6.45	0.1	—	—	—	—	—	—	—
		1	6.80	0.1	—	—	—	—	—	—	—
		2	6.45	0.1	—	—	—	—	—	—	—
		3	8.60	19.49	—	—	—	—	—	—	—
		3½	8.70	25.17	—	—	—	—	—	—	—
		4	8.50	26.69	—	—	—	—	—	—	—
		6	8.10	29.56	—	—	—	—	—	—	—
		8	7.58	31.18	—	—	—	—	—	—	—
		9.4	7.00	31.60	—	—	—	—	—	—	—
14/6 1980	1	0	10.40	0.1	—	—	0.07	—	0.71	4	—
		1	9.85	0.1	8.15	103	0.02	6.25	0.58	26	—
		2	7.40	0.1	8.69	103	0.16	6.50	0.69	31	0.34
		2½	10.60	—	—	—	—	—	—	—	—
		3	14.38	22.97	12.98	209	2.08	8.20	1.05	70	—
		3½	15.00	25.63	13.12	218	2.36	8.25	1.06	67	5.99
		4	13.90	27.45	13.11	216	2.60	8.20	0.65	70	—
		5	10.10	29.28	11.08	170	3.02	8.10	0.63	82	8.04
		6	8.60	30.05	5.24	78	3.17	7.55	0.86	88	—
8	7.77	31.25	0.23	3	3.82	7.09	18.8	180	1.67		
9	6.94	31.71	0.00	0	3.85	7.02	(96.9)	189	—		
19/8 1980	1	0	10.33	0.2	8.01	102	0.26	7.10	0.39	69	0.44
		1	10.40	0.3	7.97	102	0.27	7.10	0.39	59	—
		2	10.42	0.2	8.02	103	0.30	7.50	0.39	59	0.47
		3	17.60	21.79	20.93	358	2.34	8.83	—	58	—
		3½	18.40	23.25	21.23	372	2.43	8.85	—	57	—
		3¾	18.40	—	—	—	—	—	—	—	—
		4	19.35	28.11	20.47	376	2.82	8.70	0.69	52	6.38
		4¼	19.75	—	—	—	—	—	—	—	—
		4½	19.60	—	—	—	—	—	—	—	—
5	15.40	29.82	17.38	298	3.15	8.50	0.61	76	—		
6	11.60	30.43	8.80	140	3.34	7.90	0.69	91	8.23		
8	8.60	31.45	0.04	1	4.30	7.25	(63.4)	178	3.10		

29/11	1	0	0.05	0.1	9.58	94	0.43	—	1.14	—	3.7	—
1980		1	1.42	2.97	9.05	94	0.58	—	0.23	—	1.5	—
		2	3.42	10.56	8.45	97	1.22	—	0.68	—	0.5	—
		3	6.22	24.21	7.82	106	2.28	—	2.67	—	9.6	—
		4	8.65	28.00	4.60	68	2.93	—	5.70	—	28.0	—
		5	9.25	29.24	2.71	41	3.04	—	5.41	—	25.5	—
		6	9.50	30.19	0.47	7	3.28	—	—	—	23.0	—
		7.3	9.05	31.20	0.00	0	4.71	—	—	—	2.0	—
22/4	1	0	0.00	0.1	—	—	0.23	6.60	0.30	—	—	—
1981		1	0.26	0.1	9.36	92	0.26	6.85	0.23	—	—	—
		2	2.22	17.63	4.32	51	1.77	7.25	1.48	—	—	—
		3	4.69	23.04	2.91	38	2.23	7.27	2.45	—	—	—
		4	5.41	25.91	0.90	12	2.67	7.22	2.82	—	—	—
		5	6.12	28.83	0.33	5	2.92	7.23	3.42	—	—	—
		7	7.80	30.92	0.05	1	4.32	7.24	(45.8)	—	—	—
		9	7.60	31.27	(0.12) ^a	(2)	4.83	7.25	(97.2)	—	—	—

^a A distinct H₂S smell of sample.

TABLE I (continued)

Date	St.	Depth (m)	Temp (°C)	Salinity	Oxygen		Titrati- on alk (meq/l)	pH	Reactive phos- phate		Si (μ mol/l)	Chlorophyll a		Suspended matter (mg/l)
					(ml/l)	(% sat.)			(μ mol/l)	(μ g/l)		(% phaeo- pigment)		
26/5 1981	1	0	5.50	0.0	9.12	103	0.22	—	—	—	58	—	—	—
	1	1	5.52	0.0	9.21	105	0.26	—	—	—	69	—	—	—
	2	2	6.15	0.87	7.98	83	1.04	—	—	—	117	—	—	—
	3	3	6.19	22.67	4.88	65	2.26	—	—	—	105	—	—	—
	4	4	6.95	26.06	1.52	21	2.67	—	—	—	112	—	—	—
	5	5	7.72	28.11	0.21	3	2.95	—	—	—	111	—	—	—
	6	6	7.73	29.81	0.11	2	3.23	—	—	—	88	—	—	—
	7	7	8.30	30.41	0.09	1	4.23	—	—	—	179	—	—	—
	8	8	8.21	31.07	0.00	0	4.63	—	—	—	272	—	—	—
9.2	9.2	7.90	31.26	0.00	0	5.60	—	—	—	272	—	—	—	
24/6 1981	1	0	5.30	—	—	—	0.20	6.78	0.44	—	120	0.25	—	0.983
	1	1	5.50	0.0	9.23	105	0.21	7.03	—	—	130	—	—	—
	2	2	5.76	0.0	9.13	104	0.20	7.11	0.28	—	124	—	—	—
	3	3	12.60	20.61	13.10	200	2.18	8.50	0.39	—	115	4.74	—	2.727
	4	4	11.23	24.97	10.53	161	2.58	8.43	0.36	—	105	—	—	—
	5	5	9.36	28.15	3.24	48	3.07	7.68	1.01	—	136	5.01	—	2.306
	6	6	9.00	29.80	0.60	9	3.54	7.35	10.2	—	141	—	—	—
	7	7	8.65	30.35	0.15	2	4.15	7.45	16.3	—	139	—	—	—
	8	8	8.50	31.05	0.03	0	4.91	7.39	26.9	—	170	6.03	—	6.717
9	9	8.16	31.23	0.00	0	5.41	7.46	(45.6)	—	286	—	—	—	
22/6 1982	1	0	6.80	0.0	8.76	103	—	—	—	—	—	0.42	—	—
	2	2	6.84	0.1	8.79	103	—	—	—	—	—	0.43	—	—
	3	3	10.77	24.84	9.42	142	—	—	—	—	—	2.01	—	—
	3 1/2	3 1/2	10.97	25.89	9.75	149	—	—	—	—	—	2.19	—	—
	4	4	9.94	27.69	8.25	124	—	—	—	—	—	2.34	—	—
	5	5	8.70	29.17	5.53	82	—	—	—	—	—	4.64	—	—
	6	6	8.07	30.27	0.55	8	—	—	—	—	—	3.03	—	—
	7	7	7.60	30.86	0.17	3	—	—	—	—	—	—	—	—
	8	8	6.97	31.02	0.16	2	—	—	—	—	—	—	—	—
9.3	9.3	6.50	31.89	0.02	0	—	—	—	—	—	—	—	—	
20/7 1982	1	0	8.34	—	—	—	—	—	—	—	—	—	—	—
	1/2	1/2	8.45	0.3	8.27	101	0.13	6.80	—	—	—	—	—	—

TABLE I (continued)

Date	St.	Depth (m)	Temp (°C)	Salinity	Oxygen		Titr- ation alk (meq/l)	pH	Reactive phos- phate	Σ P	Si				
					(ml/l)	(% sat.)					NO ₃ ⁻	NO ₂ ⁻	NH ₃		
12/7	1	0	5.60	0.0	—	—	0.12	6.50	0.50	0.83	146	1.0	0.3	0.2	
1983	1	1	5.83	0.0	9.09	104	0.18	7.18	0.35	0.48	142	0.8	0.0	0.1	
	2	2	5.61	0.0	9.36	106	0.22	7.05	0.32	0.34	140	0.8	0.1	0.1	
	3	3	9.30	23.41	9.63	139	2.32	8.15	0.15	0.50	88	0.2	0.3	0.6	
	3 1/2	3 1/2	8.65	—	—	—	—	—	—	—	—	—	—	—	—
	4	4	8.03	26.35	9.50	136	2.72	8.22	0.18	1.14	82	0.3	0.4	0.7	
	5	5	7.62	28.74	4.84	70	3.28	7.97	0.34	1.20	100	6.4	1.0	1.1	
	6	6	7.48	29.53	1.06	15	3.47	7.52	1.20	1.56	117	17.0	0.6	1.0	
	7	7	7.28	30.24	0.24	4	4.21	7.60	(6.60)	6.05	185	—	2.3	35.8	
	8	8	7.10	30.84	0.23	3	4.41	7.45	(23.2)	23.1	210	—	4.5	110	
9 1/2	9 1/2	7.06	31.12	0.00	0	4.40	7.40	(52.5)	47.0	205	—	4.0	126		