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The Energy Balance of Calved Ice in Lake Jökulsarlon, Iceland

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Abstract

We describe energy fluxes involved in melting ice in the proglacial lake Jökulsarlon and the transport of thermal energy into the lake from the atmosphere and the sea. Data from earlier fieldwork and campaigns have been used to estimate the net radiation balance, the turbulent fluxes, the heat provided by inflowing seawater, and the glacial meltwater flux. From aerial photographs, DGPS measurements, and mass balance measurements, we calculated a calving flux of 260×10^6 m^3 yr^{-1} for the present. The total energy required to melt all the ice in the lake is approximately 160 W m^{-2} assuming that all the calved ice is melted during 1 yr. The most important contribution is heat from seawater. Radiation provides approximately 70 W m^{-2}. The albedo depends on the ice-covered fraction of the lake and ranges from 22% in summer to 41% in winter. The turbulent fluxes are around 10 W m^{-2}. Difficulties occurred in finding an appropriate range for the roughness parameter z_0, but the most likely values are in the range of a few centimeters. We considered different future scenarios with respect to inflow of seawater and air temperature, albedo, and even inhibition of seawater intrusion, which would have a significant impact on ice cover in the lake.

Introduction

Breidamerkurjökull, an outlet glacier of Vatnajökull, reached its Little Ice Age maximum extent in 1890. During its subsequent retreat, a proglacial lake, Jökulsarlon, first appeared in 1933 (Boulton et al., 1982; Björnsson et al., 2001). Since then the lake has grown to an extent of 15 km^2 at the end of the 20th century, and sediment from the glacier has been trapped in the lake. Sediment transport by the river Jökulsa a Breidamerkursandi has not compensated for the abrasion of the coastline by ocean waves at the mouth of the river. Therefore, during the period 1904–1989, the coastline receded inland by 700 m along an 8-km-wide strip at the river mouth at an average rate of 8.5 m yr^{-1} (Johannesson, 1994; Björnsson, 1998). This coastal erosion threatens the safety of the road and two power lines. If it continues at the same rate, the road across Breidamerkursandur will be destroyed in a few years. This threat has resulted in a call for glaciological studies of the area and the processes at work. An attempt to describe the likely future development of the lake is presented here. Figure 1 shows the outlet glacier Breidamerjökull calving into the lake Jökulsarlon, which is connected by the outlet river Jökulsa to the sea.

This study quantitatively describes the various energy fluxes involved in melting the ice in the lake and the transport of thermal energy into the lake using meteorological data from nearby weather stations to evaluate the atmospheric energy fluxes. Lake-level measurements and temperature and salinity measurements in the lake are used to describe the thermal transport from tidal currents to the melting ice front. Finally, three scenarios (increasing sea temperature, increasing air temperature, and inhibition of seawater intrusion) are discussed with respect to their impact on icemelt within the lake.

Materials and Methods

PREVIOUS INVESTIGATIONS

The terminus of Breidamerkurjökull calves into Jökulsarlon, which has been expanding since its first appearance in 1933. The lake Jökulsarlon was created during the warm conditions of the 20th century when the ice flux of Breidamerkurjökull was not able to compensate for the ice loss at the terminus by melting and calving into Jökulsarlon. Breidamerkurjökull retreated from its sill into an overdeepening that the glacier had excavated during its Little Ice Age advance (Björnsson, 1979, 1996). In 1991 the glacier thickness and the topography of the subglacial depression were mapped by radio echo sounding, revealing a trench 20 km in length and 2–5 km wide that extended 300 m below sea level (Björnsson et al., 1992; Björnsson, 1996, 1998). The lake does not occupy a rock basin but was entirely excavated from an unlithified sediment sequence at least 120 m thick during the Little Ice Age (Boulton et al., 1982). It is therefore unlikely that the area of Breidamerkursandur is a rock-floored coastal plain; it is more likely that it is a deep valley in which sediments have accumulated and filled up to above sea level. There is evidence that this could be true for the whole coastal plain south of Vatnajökull (Björnsson, 1996).

In 1975 and 1976 bathymetric surveys of Jökulsarlon were undertaken by Boulton et al. (1982). At that time the lake basin had a depth of 150 m near the ice margin and a total volume of 500×10^6 m^3. The rate of sediment transport was estimated to be on the order of 1×10^7 m^3 yr^{-1} (Björnsson, 1996) over the period 1730–1930, of which 30% was transported by the river to the sea and 70% was deposited into the lake. The accumulation of sediments in the lake over the period 1932–1975 was obtained from the interpretation of lake-sediment stratigraphy (Boulton et al., 1982: 43).

The retreat of the ice front at Breidamerkurjökull has been recorded since 1903 on maps, since 1945 in variously dated aerial photographs (Björnsson et al., 2001) taken by the Icelandic Geodetic Survey (Landmaelingar Islands), and in satellite images from 1973 and 1978 (Björnsson et al., 1999). Five maps, 12 aerial photographs, and 2 satellite photos were used to describe the growth of the lake. All aerial photos until 1990 were superimposed on the map of 1989, later photos on the map of 1996. The satellite images of 1973 and 1978 were scanned at high resolution and superimposed on the maps (Björnsson et al., 2001). Breidamerkurjökull reached its maximum extension in the 1890s and has retreated 4 km since then. Areas covered by 200 m of glacier ice in 1903 are now ice free. The retreat rate of the calving front increased suddenly from only a few m yr^{-1} to 150 m yr^{-1} in the early 1950s and remained constant until the late 1960s. For the next 20 yr it

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FIGURE 1. The lake Jökulsarlon and the outlet river Jökulsa in 1998. The 108-m-long bridge is about 300 m from the coastline (Björnsson et al., 2001).

slowed to 30 m yr^{-1} and increased again in the 1990s to an average of 200 m yr^{-1}. The calving rate into the lake has increased rapidly during the last two decades. This increase has resulted in an expansion of the lake by 0.5 km^2 yr^{-1} (Björnsson et al., 2001). The lake grew since its first appearance in 1933 to a size of approximately 15 km^2 in 1998.

Between 1997 and 1999 field measurements were carried out to estimate the components of the mass balance at the glacier terminus: downglacier flow, surface melting, and calving. With the help of aerial photographs, DGPS measurements, and mass balance measurements, the calving flux was calculated to be 260×10^6 m^3 yr^{-1} (Björnsson et al., 2001). These field observations were used to obtain an empirical relationship between water depth in m and calving rate in m yr^{-1}. (Björnsson et al., 2001). The calving rate is expected to increase rapidly when the glacier retreats 2 km inland from its 1998 terminus to where the bed slopes from 200 m to 300 m below sea level.

Jökulsarlon receives inflow of glacial meltwater from several locations along the front of the glacier; outflow of the glacial lagoon is through the river Jökulsa to the sea. At high tide the glacial lagoon receives comparably warmer water from the sea. A salt taste in the lake was first discovered in 1951, and the salinity later increased (Kjartansson, 1957). In 1974 and 1975 measurements of the lake temperatures and salinity were made by Harris (1976). The vertical and spatial distribution of salinity in Jökulsarlon showed great variations at different times of the year. During the winter months water at all depths and in all parts of the lake was isothermal at 0°C (Harris, 1976). The thermal structure showed considerable changes during summers. The maximum value of salinity, 12.5‰, was reached at the lake bottom in spring. During the summer months meltwater from the glacier flushed the saltwater out of the lake, but a bottom layer of saline water (~5‰) remained.

ENERGY BALANCE

Jökulsarlon receives thermal energy primarily from solar radiation and from inflowing water from the sea through the channel of the river Jökulsa. The energy input into the lake is the sum of total solar radiation, incoming longwave radiation minus radiation from the lake surface, heat from warm and moist air, and heat from the seawater flowing into the lake at high tide.

In quasi-stationary state the supply of energy is balanced by an equally large loss when averaged over the volume of the lake:

$$M + \Delta G = R + H_s + H_l + H_{gl} + H_i \ (\mathrm{W\ m^{-2}}) \qquad (1)$$

where M = energy available for melting, ΔG = gain of heat of a vertical column of water from the surface to the depth at which vertical heat transfer is negligible, R = net radiation, H_s = vertical eddy flux of sensible heat, H_l = vertical eddy flux of latent heat, H_{gl} = heat contribution from glacial meltwater, and H_i = heat supplied by tidal water inflow from the sea. All fluxes on the right side of this equation are positive when they supply energy to the lake.

The energy flux density required to melt all the ice in the lake during 1 yr is

$$M = L_f q_c \rho_i A^{-1} \qquad (2)$$

where L_f is the latent heat of fusion for ice (3.34×10^5 J kg^{-1}), q_c is the calving flux in m^3 s^{-1}, ρ_i is the density of glacier ice (830–917 kg m^{-3}: Paterson, 1994), and A is the lake area.

Both solar radiation, which penetrates into the water, and heat supplied by tidal inflow affect the total volume of the lake. Therefore, not only the amount of energy that enters the surface (W m^{-2}) but also the heat warming of the lake water from the surface to the depth at which vertical heat transfer is negligible must be considered.

$$\Delta G = \frac{1}{A\tau^2} \int_{Vol=0}^{Vol=V} \int_{t=0}^{t=\tau} \rho c T \, dt \, dV \approx \rho c \Delta \bar{T} \bar{Z} \tau^{-1} \tag{3}$$

where ρ is the density of water, $c = 4.2 \times 10^3$J kg^{-1} °C^{-1}, the specific heat capacity of water, $\bar{Z}$ is the mixed layer of the lake, τ is the period considered, and $\Delta\bar{T}$ is the volume mean temperature difference between beginning and end of the period τ.

Net radiation is a major component of the surface energy budget.

$$R = Q(1 - \alpha) + I_i - I_o \tag{4}$$

where R = net radiation, Q = direct and diffuse incoming solar radiation at the surface (global shortwave radiation), α = surface albedo, I_i = incoming longwave radiation at the surface, and I_o = emission of longwave radiation from the surface.

Turbulent fluxes were calculated using a one-level eddy flux model for neutral air conditions. The transfer of sensible heat in neutral conditions for a given height z above the surface can be written as

$$H_s = \rho_a c_p k_o^2 u(z) \frac{T(z) - T(z_o)}{(\ln(z) - \ln(z_o))^2} = \alpha_s(T(z) - T(z_o)) \tag{5}$$

where T(z) and u(z) are temperature and wind speed at the height z, c_p is the specific heat capacity at constant pressure with a value of 1004 J kg^{-1} K^{-1}, ρ_a is the density of air, and k_0 is a dimensionless constant, von Karman's constant, with a value of approximately 0.4. The surface-roughness parameter z_0 is the height above the mean surface at which the wind speed is zero.

The transfer of latent heat in neutral conditions for a given height z above the surface can be written as

$$H_l = L_v k_o^2 u(z) \left(0.622 \frac{\rho_a}{p}\right) \frac{e(z) - e(z_o)}{(\ln(z) - \ln(z_o))^2} \tag{6}$$

where e(z) and e(z_0) are the vapor pressures at the heights z and z_0. The vapor pressure e(z) was not measured directly but is given as e(z) = r(z) e_s/100, where r(z) is the measured relative humidity in % and e_s is the saturation vapor pressure given by

$$e_s(T(z)) = 611 \exp\left(17.5 \frac{T(z)}{T(z) + 241.2}\right) \tag{7}$$

in Pascals.

Glacial meltwater flowing down from the glacier acts as a heat sink during the melting season. This term is calculated as:

$$H_{gl} = \rho c q_{gl} \Delta T A^{-1} \tag{8}$$

where ρ is the density of water (kg m^{-3}), $c = 4.2 \times 10^3$ J kg^{-1} °C^{-1}, the specific heat capacity of water, q_{gl} is the volume flux of the glacial meltwater (m^3 s^{-1}), and ΔT is the temperature difference between the glacial meltwater and the lake water.

The heat flux density from seawater can be estimated with a similar equation:

$$H_i = \rho c_w q_s \Delta T A^{-1} \tag{9}$$

where ΔT is the temperature difference between the lake water and the inflowing seawater, $c_w = 4.2 \times 10^3$J kg^{-1} °C^{-1}, the specific heat capacity of water, ρ is the density of the seawater, and q_s is the inflow of seawater to the lake in m^3 s^{-1}.

METEOROLOGICAL AND LIMNOLOGICAL DATA

Meteorological observations are not available at Lake Jökulsarlon, but extensive data have been collected on neighboring glaciers. We use these data to make quantitative estimates as a basis for discussion of energy components at the surface of the lake. Since 1994 automatic meteorological stations have been operated on Vatnajökull ice cap. In 1996 a 2-yr glacial meteorological experiment was started on the glacier in a cooperative venture between the University of Iceland, University of Utrecht, Free University of Amsterdam, and University of Innsbruck (Oerlemans et al., 1999). The location of each station can be seen in Figure 2. Data from the lowermost stations, U3, A4, and A5, combined with data from two stations on the moraine, U2 and A1, were used to calculate the radiation components and turbulent fluxes. The stations were running from the beginning of May until 1 September. Station A4 is situated approximately 6 km northwest of the lake on the tongue of the outlet glacier Breidamerkurjökull at an elevation of 240 m and was operated continuously for several years. The stations A1 and U2 are located on the proglacial moraine at elevations of 4 and 50 m, respectively. U3 and A5 are on the glacier at elevations of 165 m and 381 m. Before and after the fieldwork, the sensors were calibrated at Cabauw, the Netherlands (Oerlemans et al., 1999). Full-yr data is available for the years 1995, 1996, 1997, and 1998. In 1998 only the individual components of the incoming solar radiation were measured.

For the present study, we use data from the summer of 1996, completed with the year-round data from station A4 in 1998. Stations used for calculations are U2 and U3 (University of Utrecht) and A1, A4, and A5 (Free University of Amsterdam), respectively. Since stations A1 and U2 are located on the moraine, we used only incoming shortwave and longwave radiation data from these two stations.

Based on the long-term record from the climate station Holar i Hornafirdi run by the Icelandic Meteorological Office, the summer of 1996 and the year 1998 were classified as representative years with respect to monthly mean temperature, maximum temperature, precipitation, and mean cloud cover.

Observations have been collected on lake temperature and salinity, the amount of ice cover on the lake, and lake-level variations (Harris, 1976; Arnason, 1998; Zophoniasson et al., 1999). The lake surface is entirely frozen from January through April (Fig. 3). During spring calving, melting begins. More than 40% of the lake is typically covered with icebergs during spring. Ice coverage is reduced as the melting rate increases during summer, and ice cover reaches a minimum in August. The lake area typically covered with icebergs was estimated from aerial photographs taken by the Icelandic Geodetic Survey. Twelve pictures are available for the period 1982 to 1998, all taken in August. An average ice cover of 40% during the summer months was estimated from these pictures. Seawater intrusion during fall is enhanced by low-pressure events. This results in a higher calving rate and correspondingly larger amounts of ice in the lake in comparison to the summer months (Fjölnir Torfason, personal communication, 2001).

Results

Energy balance components were calculated on a monthly basis. Details of the calculations are given in Landl (2002). The incoming longwave and shortwave radiation and the outgoing shortwave radiation were measured at different stations, as indicated earlier. The outgoing longwave radiation was calculated based on lake surface temperatures, which range from 273 to 276 K over the year (Harris, 1976). During spring, autumn, and winter the incoming and outgoing longwave fluxes were assumed to be approximately equal in this maritime environment (Björnsson et al., 2001).

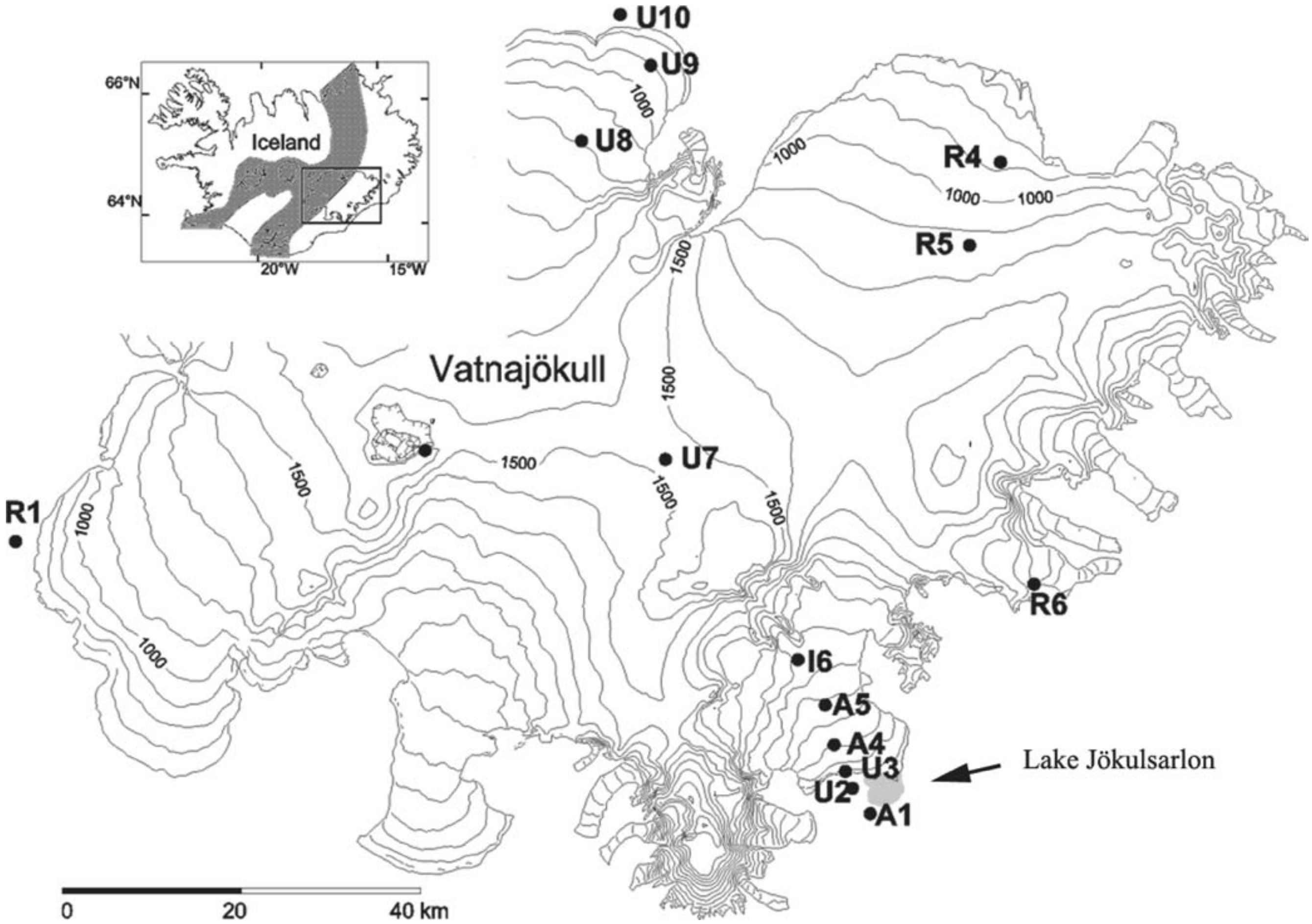

FIGURE 2. Map showing the locations of the weather stations on Vatnajökull (Obleitner and de Wolde, 1999) and the position of Lake Jökulsarlon.

Albedo values were calculated based on the area cover of ice in the lake, resulting in variation in albedo from 22% in summer to 41% in winter.

More than half of the yearly solar radiation is received during the three summer months, June, July, and August, and approximately one-third during spring (March, April, and May). The lake absorbs an average annual solar radiation of 70 W m^{-2} , which gives a total of 1050 MW over the 15 km^2 of the lake.

Sensible and latent heat fluxes were calculated using the theory described earlier, using temperature and wind speed at 2-m height. For the roughness parameter z_0, we started with z_0 values similar to those in the ablation zone of a glacier (in the range of 1 to 6 mm). The most plausible results were obtained with a surface roughness parameter z_0 of 2 mm, which was used for all calculations.

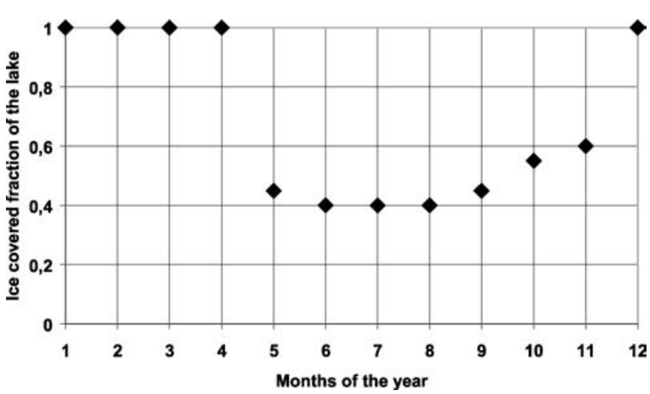

FIGURE 3. Mean ice-covered fraction of Lake Jökulsarlon.

We found that turbulent fluxes were high during the summer. This is the result of strong summer katabatic winds which flow down the steep-sloping outlet glacier Breidamerkurjökull. The contribution from turbulent fluxes to the energy balance near the snout of the glacier is much higher than at other outlet glaciers (Björnsson, 1972; Björnsson et al., 2001).

Calculated monthly mean values for net radiation and turbulent fluxes are shown in Table 1 and Figure 4.

In situ observations and photographs show that the proglacial lake has a far rougher surface than assumed in the calculations above. The implications were explored using a theory from van den Broeke (1996). Microtopographical surveys are a better basis for the determination of z_0 than wind profile measurements because of a zero-referencing problem in rough terrain (Munro, 1989). The value of z_0 is not equal to the height of the roughness elements h_0, but there is a one-to-one relation between these two parameters. Surface roughness increases with larger roughness elements. Hence, the value of z_0 of the lake was determined from the surface microtopography according to the following expression (van den Broeke, 1996):

$$z_0 = 0.5 h_0 s S^{-1} \qquad (10)$$

where h_0 is the roughness element height, s is the area cross-section of the roughness element perpendicular to the wind vector, and S is the fraction of the lake covered by roughness elements. The icebergs were approximated as pyramids. The average value of z_0 for an iceberg is in the order of magnitude of a few centimeters.

TABLE 1

Monthly means for net radiation and turbulent fluxes (W m^{-2})

Months	R	H_s	H_l	$R + H_s + H_l$
January	3	−14	−118	−129
February	16	−89	−150	−223
March	51	−161	−166	−276
April	97	−11	−68	18
May	132	72	−21	183
June	154	104	11	269
July	157	117	29	303
August	120	118	34	272
September	63	7	−1	69
October	32	−9	−68	−45
November	7	1	−77	−69
December	1	−1	−97	−97

In Table 2 sensible and latent heat fluxes were calculated for summer conditions using values for the roughness parameter z_0 ranging from 2 mm up to 100 mm. A fivefold increase in the roughness parameter results in sensible and latent heat fluxes that are 1.7 times higher.

Since the distance between the lake and the ocean has decreased, seawater is increasingly able to flow into the lake at high tides. The density of water depends on both the salinity and the temperature. In most cases of saltwater intrusion from the sea into a lake, the effect of salinity is predominant. Density differences have a major effect on the flow because they tend to cause stratification. The denser seawater sinks below the less dense lake water.

The volume of seawater entering Lake Jökulsarlon is not accurately known, nor are its tidal and seasonal fluctuations. Following a personal communication by Björnsson (2001), we assumed the ocean temperature to be constant for our calculations. From sporadic measurements by Arnason (1998), the volume rate of the inflowing water was estimated as ranging from 30 m^3 s^{-1} to 100 m^3 s^{-1} over one tidal period. Since the outflowing water has a temperature of 0°C (Zophoniasson et al., 1999) all the heat is used for melting the calving ice within the lake. Therefore, the average heat flux received by the lake from the warm seawater inflow ranges from 882 to 2940 MW per year, which is equal to 59–196 Wm^{-2}, assuming a lake surface area of 15 km^2.

During the summer months the surface temperature of Jökulsarlon reaches a maximum value of 4°C in some parts of the lake but more often does not exceed 2°C. The surface heating is transferred to an average depth of 30 m (Harris, 1976). During winter months the water is isothermal at 0°C. An average energy flux of 120 MW (8 Wm^{-2}) is required to raise the temperature in the lake during the summer months.

The present calving flux was estimated to be $q_c = 260 \times 10^6$ m^3 yr^{-1} (Björnsson, 2001). The energy required to melt all the ice in the lake during one year according to equation (2) ranges from 2300 to 2500 MW, or approximately 160 W m^{-2}.

Compared to lake surface water at 2 to 3°C (Harris, 1976), meltwater flowing down from the glacier is an additional heat sink in the lake energy balance. Nearly all of the melting on the glacier occurs in the summer months. The total area of Jökulsarlon watershed ranges from sea level to 1750 m a.s.l., and a total volume of 1790×10^6 m^3 was melted from that area in the year 1999–2000 (F. Palsson, personal communication, 2002), based on a melt rate of –2.4 m yr^{-1}. This results in an inflow of 57 m^3 s^{-1} to the lake. The average heat flux received by the lake from glacial meltwater over 1 yr is approximately –241 MW (–16 Wm^{-2}).

Assuming that all calved ice was melted, the results for the terms of equation (1) are shown in Table 3. Radiation and turbulent fluxes are

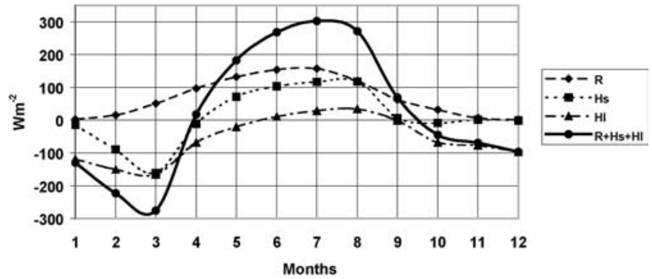

FIGURE 4. Monthly mean values for net radiation and turbulent fluxes for the year 1998.

at a maximum during summer (Table 4), while the seawater intrusion is at a minimum. Seawater intrusion is at its minimum in the summer due to a maximum in glacial meltwater flow and an enhanced melting of the ice in the lake during summer time.

By varying z_0 until calculated values and residual agreed, we accepted a value of x mm for all further calculations. Calculated values are shown in the lower part of Table 3. The differences between the calculated and residual fluxes range from 20 to 80 W m^{-2}, resulting in a volume of melted ice between 30×10^6 and 130×10^6 m^3 over 1 yr.

Discussion

In this section we discuss both climatic changes in albedo and temperature and engineering measures aimed at preventing intrusion of seawater to the lake.

The surface albedo is an important component for calculating the net shortwave radiation balance, and therefore affects the whole energy balance. The albedo of proglacial Lake Jökulsarlon depends both on impurities in the ice and on the percentage coverage of icebergs. Figure 5 shows net shortwave radiation balance for summer conditions assuming ice covers of 30%, 40%, and 50%, assuming a constant ice albedo of 0.43. Differences are largest on clear days in June, July, and August, when global radiation reached daily mean values of 13 W m^{-2}. Ice cover difference of 10% results in an average change in net shortwave radiation of 7 W m^{-2}.

Figure 6 shows the impact of changes in lake ice albedo. Three scenarios are given: clean ice, dirty ice, and a fully covered lake with clean ice. Case 1: The open squares in Figure 6 show the variation of the radiation balance for the whole year assuming ice with the same albedo as glacier ice in the ablation zone. The albedo for "clean" ice (0.43) is taken from station U2 (Fig. 2). Case 2: Since large parts of Vatnajökull are underlain by an active volcanic zone (see the gray zone in the small picture in Fig. 2), we included the effect of an ash cover. A value of 0.15 was used for the albedo of "dirty" ice because the volcanic ash is rather dark. The net shortwave radiation balance for this case is shown by the solid squares. Case 3: From January until April and in the case of inhibited seawater intrusion, the lake is entirely frozen. A mean albedo of 0.43 was taken, as in Case 1. The crosses show the shortwave radiation balance for this case. The annual mean

TABLE 2

Energy fluxes to Jökulsarlon (W m^{-2}) with different values for z_0 for summer conditions

z_0	H_s	H_l	$H_s + H_l$
2 mm	113	25	138
10 mm	186	42	228
25 mm	272	61	333
50 mm	383	86	469
100 mm	581	130	711

TABLE 3

Seasonal energy-balance components of the heat budget for Jökulsarlon (W m^{-2}) with turbulent fluxes calculated as residual: $H_s + H_l = M + \Delta G - R - H_{gl} - H_i$ (equation 1). Turbulent fluxes calculated with T (2 m), e (2 m), and $z_0 = 2$ mm

Season	M	ΔG	R	H_{gl}	H_i	Residual $H_s + H_l$
Spring	108	0	95	0	65	−52
Summer	324	8	143	−62	65	186
Autumn	164	0	34	−2	98	34
Winter	56	0	7	0	164	−115

	Calculated H_s	Calculated H_l	Calculated $H_s + H_l$	Calc.-Residual $\Delta(H_s + H_l)$
Spring	−30	−101	−131	−79
Summer	113	25	138	−48
Autumn	2	−56	−54	−20
Winter	−31	−120	−151	−36

shortwave radiation balance for dirty ice is 24% higher than for the present situation. A year-round ice-covered lake would result in a net shortwave radiation balance 20% lower than at present.

For simplicity in this investigation, we assumed that a variation in air temperature would affect only the sensible heat flux. Effects on latent heat flux and radiation balance are ignored. Keeping the latent heat flux constant, a rise in temperature means that the relative humidity is lower. An air temperature rise of 1 K results in a 25 to 30 W m^{-2} higher sensible heat flux. In these calculations the instrument height z was 2 m, the roughness parameter z_0 was taken as x mm, and the wind speed averaged over 1 yr was 5.4 m s^{-1}. This implies an average transfer coefficient α_s of 23 W K^{-1} (equation [5]).

The seawater at the southern coast of Iceland has a mean temperature of 7°C throughout the year (Stefansson, 1999). A rise in temperature of intruding seawater would provide additional energy for heating the lake water and melting the ice. Assuming an average inflow of 50 m^3 s^{-1}, a rise of seawater temperature of 1°C results in an energy increase of 14 W m^{-2}. This equals 210 MW, resulting in an extra volume of 22×10^6 m^3 yr^{-1} ice melted.

As described earlier, Lake Jökulsarlon is connected to the sea by the river Jökulsa. Since the length of the river is decreasing and the seawater is increasingly eroding the coastline, the amount of water entering the lake at high tide is increasing. Consequently, more and more warm seawater is available to melt the ice in the lake. If seawater intrusion were inhibited, an average of 98 W m^{-2} less energy would be available for heating the lake water and melting the ice. This amount of energy is equal to 1470 MW for the 15-km^2 lake, and equal to 156×10^6 m^3 yr^{-1} of ice that cannot be melted. If we assume the same ice flow as at present and take into account the fact that nowadays all the calved ice is melted during 1 yr, the amount of ice that cannot be melted comprises approximately 60% of the ice entering the lake every year. After the first spring season the volume of ice in the lake would be equal to 26×10^6 m^3. Assuming an ice cover of 100%, this would equal an average ice thickness of 1.7 m over the entire lake. In autumn of the same year the volume of ice in the lake would have a value of 91×10^6 m^3, or approximately 6-m ice thickness. Clearly interruption of seawater influx to the lake would eventually shut down calving.

TABLE 4

Shortwave radiation balance for different albedo values (W m^{-2})

Season	Varying ice cover (albedo = 43%)	Dirty ice (albedo = 15%)	100% ice cover (albedo = 43%)
Spring	90	124	83
Summer	134	153	98
Autumn	33	40	26
Winter	6	9	6
Mean	66	82	53

The intention of this work was to better define the problem of the energy balance of the proglacial lake Jökulsarlon and to construct a plan of future studies of this complex problem. Calculations were done with little data, and many assumptions had to be made. In this section we want to outline various suggestions for improving this work in the future:

- The weather stations used for calculations were situated on the moraine and on the glacier. For future investigations, weather stations closer to the lake are required to obtain accurate seasonal variations in input of energy to the lake.
- The albedo of Vatnajökull was investigated by a team at the University of Utrecht who flew over the ice cap by helicopter with radiation sensors (W. Greuell, personal communication, 2002). Their starting point was close to Lake Jökulsarlon, and therefore data for this region is available. Future studies of Lake Jökulsarlon should make more use of airborne measurements of albedo as well as surface temperature. An attempt should be made to use satellite data for this purpose.
- The turbulent fluxes above the lake are not believed to be stationary. By their very nature, katabatic winds are turbulent and fluctuating. An appropriate theory has to be developed for the aerodynamic effect of floating icebergs. Measurements of surface temperatures at fine vertical resolution might aid in calculating the turbulent fluxes more precisely.
- Since the glacier is retreating into regions of increasing bed depth, the lake is becoming increasingly deep, which further enhances the glacier calving rate. This situation has to be taken into account in calculations of iceberg influx into the lake.
- The stratification and transport of heat within the lake was last measured in 1974 (Harris, 1976). Conductivity, temperature,

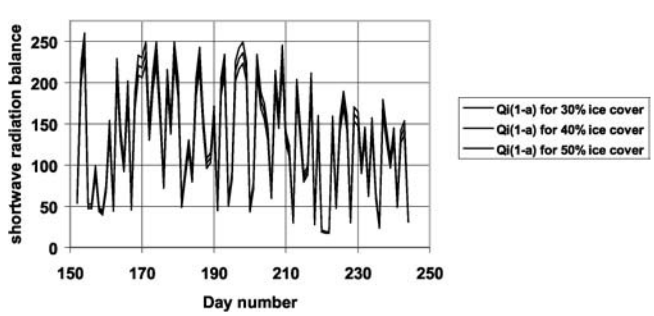

FIGURE 5. Albedo sensitivity. Daily mean values for net shortwave radiation for the summer of 1996 for ice covers of 30%, 40%, and 50% of total lake area.

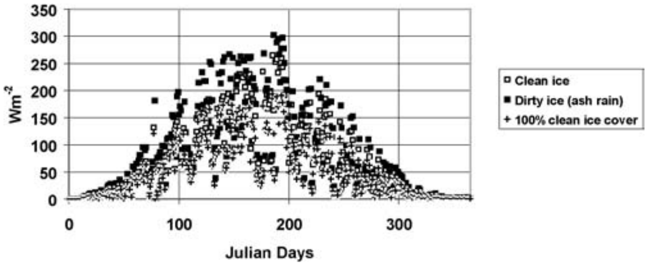

FIGURE 6. Daily mean values for net shortwave radiation for the year 1998 for three different cases.

and depth measurements would provide better information on the distribution of inflowing seawater and glacial meltwater in the lake.

- Temperature, salinity, and velocity measurements in the river channel of Jökulsa over a longer period would provide detailed information about the energy input from intruding seawater.

Conclusions

The proglacial lake Jökulsarlon is 15 km^2 in area and is expanding. Heat from different sources melts the calved ice in the lake. Intruding seawater at high tide is the most efficient component and contributes about 60% of the required energy on a yearly basis. Other sources include radiation (~43%) and turbulent fluxes (~7%). Glacial meltwater flux is an energy sink of ~−10%. The most uncertain term in the energy balance equation is the turbulent fluxes because of difficulties in determining the roughness parameter z_0 for the lake surface.

Attention must be drawn to the large energy contribution provided by intruding seawater at high tide. The study of this important component was expanded by looking at the effect of inhibiting seawater intrusion. After 1 yr the lake would be fully covered with ice at an average depth of 6 m. This would have a great impact on the calving rate and future development of the lake.

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