Climate variability and glacial processes in eastern Iceland during the past 700 years based on varved lake sediments

JOHAN STRIBERGER, SVANTE BJÖRCK, ÓLAFUR INGÖLFSSON, KURT H. KJÆR, IAN SNOWBALL AND CINTIA B. UVO

Much of the interannual climatic variability in the North Atlantic region can be explained by fluctuations in air pressure known as the North Atlantic Oscillation (NAO). The NAO is a redistribution of atmospheric masses between the subtropical high (Azores High) and the polar low (Icelandic Low), and is the dominant cause of winter climate variability from North America to Europe (Hurrell et al. 2003). Iceland, situated in this climatically sensitive region, is to some extent influenced by changes in the NAO; for example, during positive NAO phases, when the air pressure difference between Iceland and the Azores is high, Iceland and northwest Europe typically experience milder and wetter winters with more frequent and intense storms. Iceland is also strongly influenced by the strength of the Greenland High, a variable that is not reflected directly in the NAO index (Jónsson & Valdimarsson 2005).

Ocean currents also influence the climate in Iceland (Einarsson 1984). Relatively warm and saline water originating from the North Atlantic drift is transported by the Irminger current along the western and northwestern coast of Iceland, while the East Icelandic and East Greenland currents bring cold and low-salinity polar waters from the north (e.g. Knudsen et al. 2004).

Oscillations in these oceanographic and atmospheric systems influence the mass balance of Icelandic glaciers and result in fluctuations in the amount of meltwater discharged from the glaciers (Gudmundsson 1997; Ingölffsson et al. 1997; Norddahl & Einarsson 2001; Bradwell et al. 2006). The Vatnajökull icecap is situated centrally in the Nordic Sea region, where North Atlantic deep water is formed, and its mass balance is affected by the interplay between humid-bearing cyclone systems coming from the southwest and cold polar air masses from the north. Therefore, the icecap can be regarded as a sensor for a combination of a number of regional climatic driving processes. Even though Holocene glacier advances in the North Atlantic region have been discussed in several papers (e.g. Karlén 1988; Nesje et al. 2001; Kirkbride & Dugmore 2006) and local palaeoclimatic records exist from this region (e.g. Caseldine et al. 2003; Andresen & Björck 2005; Axford et al. 2009; Geirsdóttir et al. 2009), the lack of data concerning high-resolution winter proxies and cryospheric changes in space and time is striking.

Annually laminated sediments (varves) may form in basins with seasonal sediment input and where no sediment mixing occurs, for example in glacier-fed lakes. The annual resolution provides a precise geochronology, and several studies have found relationships between varve properties and various environmental parameters. Clastic varves in proglacial lakes have previously been linked to a number of meteorological parameters (Hodder et al. 2007), for example summer temperatures (Hughen et al. 2000; Thomas & Briner 2009) and autumn precipitation (Desloges & Gilbert...
1994). However, before using varve parameters as a climate proxy it is important to fully understand what controls the formation of varves in a specific basin, as highlighted by Hodder et al. (2007). Furthermore, it is essential to identify varves correctly and to verify that the laminations are indeed seasonally/annually deposited (Jones et al. 2009).

Here we present a varve chronology from Lake Lögurinn in eastern Iceland covering the period AD 1262–2005. This lake constitutes the only sediment trap available for studying the Holocene meltwater record of one of Vatnajökull’s most conspicuous outlet glaciers, Eyjabakkajökull. Lake Lögurinn also receives sediment input from the river Grímsá, which drains an extensive catchment that is currently non-glaciated. In the following, we first show how the laminated sediments were verified as varves. Then we compare various varve properties (winter and summer lamina thickness) to meteorological records and glaciological observations. We find that different properties of varves in Lake Lögurinn can be used to infer both past climatic conditions (winter precipitation) and past glacial processes (surges).

Study site

Lake Lögurinn (latitude 65°15' N, longitude 14°25' W) is situated in the central part of the Fljótsdalur valley in eastern Iceland (Fig. 1). The lake is approximately 30 km long, up to 2.8 km wide and covers an area of 53 km². The mean water level is at 20 m a.s.l. (above sea level) and its average water depth is 51 m (Hallgrímsson 2005). The lake consists of three sedimentary sub-basins: the southern basin is the deepest (112 m) and is relatively flat-bottomed with a sediment thickness averaging 80 m. Step-like sills separate the shallower central (72 m) and northern (42 m) basins from the southern basin of Lake Lögurinn via a braided system of gravel islands. The water in Fljótsdal remains high until early September. The discharge in Jökulsá í Fljótsdal starts to increase in early May (Orkustofnun 2006) (Fig. 3).

Lake Lögurinn receives input mainly from Jökulsá í Fljótssdal and Grímsá, but also from a number of smaller streams and rivers that drain mainly heath, grassland and cultivated land (Kardjilov et al. 2006). The lake drains northwards through the Lagarfljót river into the Héraðsfloi Bay and eventually the Norwegian Sea.

Jökulsá í Fljótssdal is the main drainage channel of Eyjabakkajökull, which is a distinctive surging outlet glacier of the Vatnajökull icecap that currently terminates 55 km south of the lake in the highland north of Vatnajökull. The total area of ice that drains water to Jökulsá í Fljótssdal is about 110–127 km², and the average ice thickness is 145 m (Björnsson 1988). The equilibrium-line altitude of Eyjabakkajökull between 1995 and 2001 varied between 1040 and 1240 m a.s.l. (Björnsson et al. 2002), and its surface slope is about 2.7° (Björnsson et al. 2003). The glacier is composed of two main ice bodies: one valley glacier from the southeast (~55 km²) and one outlet glacier (~50 km²) draining the Vatnajökull icecap. The ice bodies are separated by a medial moraine.

Jökulsá í Fljótssdal originates at the Eyjabakkajökull ice front at 750 m a.s.l. About 45 km downstream the river connects with the smaller non-glacial river Keldúa, and 10 km further northwards it enters the southern basin of Lake Lögurinn via a braided system and a delta. The water in Jökulsá í Fljótssdal is whitish-opaque owing to the high concentration of light-coloured sediments delivered from the glacier. Its impact on the lake is obvious, as the lake has the same apparent whitish-opaque colour in the southern basin close to the inlet of Jökulsá í Fljótssdal. In the northern basin the water is more whitish-greyish, which reflects the mixing with water from other rivers and the gradual deposition of coarser suspended sediments during transport northwards in Lake Lögurinn. The 1963–2005 daily average discharge of Jökulsá í Fljótsdal was 37.5 m³ s⁻¹ (Orkustofnun 2006).

Grímsá, the second largest river of the Lake Lögurinn catchment, enters the lake from the north-eastern side (Fig. 1). It is a direct run-off river and its discharge increases almost instantly during heavy rains, when it flows rapidly, but it can dry out during dry conditions and frost (Einarsson 1994). It drains a nonglaciated catchment, and its 1963–2005 daily average discharge was 24 m³ s⁻¹ (Orkustofnun 2006).

The 1961–1989 mean annual temperature in the Lake Lögurinn area was 3.4°C, reaching 0.8–1.6 degrees below 0°C between December and March (Veðurstofa Íslands, http://www.vedur.is) (Fig. 2). January is the coldest month, with a mean temperature of −1.6°C. Mean 1961–1989 annual precipitation was 766 mm a⁻¹ (Veðurstofa Islands). Most of the precipitation falls during autumn and winter, both as rain and as snow (Fig. 2).

In winter, when the lake freezes over, ice starts to form over the shallower northern parts of the lake, and later spreads southwards. The lake surface is usually totally frozen over in February–March. On average, the lake ice starts to disintegrate at the beginning of April. During the past 50 years, the central areas of Lake Lögurinn have been free of ice every 3rd–5th year (Hallgrímsson 2005).

The discharge in both Jökulsá í Fljótsdal and Grímsá starts to increase in early May (Orkustofnun 2006) (Fig. 3) as a result of the spring melt. The discharge rate in Grímsá peaks in June, whereas the discharge in Jökulsá í Fljótsdal remains high until early September. The longer discharge peak in the glacial river is caused by an initial snow-melt followed by the summer melting of Eyjabakkajökull, whereas the Grímsá peak is the result mainly of snow-melt during spring/early summer. During autumn and winter the Grímsá record displays distinct flow peaks and becomes the dominant
discharge source from mid-November until the following spring. On a yearly basis the average discharge was highest in Jökulsá í Fljótstdal for every year between 1963 and 2005, except for 1972, when the average discharge in Grímsá was higher (Orkustofnun 2006).

Sediment accumulation rates in the central basin of the lake were estimated to be 9.8 mm a\(^{-1}\) between 1875 and 1993 (Gudjónsson & Desloges 1997), compared with 6.6 mm a\(^{-1}\) for the same period in the northern basin, suggesting that sediment accumulation rates decline significantly distally from the inlet of Jökulsá í Fljótstdal and northwards in the lake.

Methods

Core collection

Seismic profiles were obtained from the northern basin in June 2006 using a C-Phone hydrophone streamer connected to a C-Boom system. A number of short sediment cores (20–30 cm) were obtained for studying the sediment–water interface and to check for laminations/varve potential. Based on interpretations of these profiles and the short sections of sediments, coring was carried out in the central part of the northern basin in September 2006 at a water depth of 38 m (Fig. 1). The northern basin is protected from mass movements/gravity flows from the south by a moraine ridge (Freysnes Peninsula) forming a sill \(~10–20\) m depth below the water surface.

A cable-operated Uwitec ‘Niederreiter 60’ piston corer (ø 60 mm) was used to obtain 12 m of sediments in overlapping 3-m segments (herein referred to as L1). The overlap between core segments was approximately 1.5 m, so that the core recovery was 100% and 90% of the sediment sequence was recovered twice. The sediment–water interface was captured in a complementary 35-cm-long surface-sediment core using a LTH Kajak sediment-surface sample (ø 60 mm). In September/
October 2007, using the same equipment and coring strategy, an additional sediment sequence 14 m thick was collected about 350 m east of L1 at a water depth of 16 m (herein referred to as L2). The cores were split lengthwise into ‘D’ sections, wrapped in plastic and kept cool during transport back to Lund University.

Lamination analyses

During core splitting it was noticed that the sediments in Lake Lögurinn were distinctly laminated, especially in the upper part of the succession. The individual laminae alternate between being dark and light, and we hypothesized that they are of annual nature; that is, a dark and a light lamina are formed each year, constituting a varve. This hypothesis will be tested and henceforth we refer to either lamina(e) or varve(s), where the latter is a couplet consisting of one dark and one light lamina.

The sediments were photographed in increments of 2 cm with an AF-S Micro Nikkor 60-mm f/2.8G ED lens, and the images were mounted in Adobe Illustrator CS2 where laminated pairs, couplets, were marked and counted. The laminations were also examined with a Wild Heerbrugg M420 macroscope equipped with a polarizing filter, which was coupled to an Aniol electric tree-ring measuring system in the Laboratory for Wood Anatomy and Dendrochronology at Lund University. With this equipment it was possible to measure each lamina at micrometre precision. The measurements were taken along the central axis of the cores (within the inner 18 mm of the sediment) on three separate occasions, and the thickness of each lamina is presented as a mean of these measurements. The results of the image analyses and the measurements were finally compared with results of the X-radiography and X-ray fluorescence (XRF) analyses.

X-radiography and X-ray fluorescence analyses

XRF analyses were carried out on the sections from L1 with an ITRAX Core Scanner (Croudace et al. 2006) at the Core Processing Laboratory in the Department of Geology and Geochemistry at Stockholm University. The non-destructive ITRAX Core Scanner provides high-resolution element profiles and radiographic images that can be used to interpret laminated sediments (Guyard et al. 2007).

The XRF data, presented as peak areas of counts per second (cps), were obtained using a molybdenum anode X-ray tube. The XRF measurements were carried out on the split-core surfaces from a 4-mm-wide and 0.1-mm-thick area in 0.5-mm steps, with an exposure time of 10 s per measurement. The split cores were covered with a thin plastic film to protect the sediments from drying during the scans. Intensity variations of 22 elements were obtained from the sections.

XRF data obtained on wet sediments are, however, affected by different physical parameters, such as grain size, density and water content. In a comparison by Tjallingii et al. (2007) between a wet sediment surface and dry powder samples from the same core, light elements (Al, Si) attain strongly reduced intensities whereas heavier elements (K, Ca, Ti, Fe) are relatively unaffected by variations in any physical property. Elements displaying low counts (mean of < 40 cps), including Al and Si, were discarded for this study. The
final elements used were K, Ca, Ti, Mn, Fe, Co, Zn, Sr and Zr.

The radiographic images produced by the ITRAX Core Scanner display bulk density variations in the sediments and are ‘radiographic positives’; that is, low-density areas appear light and high-density areas appear dark. The 36-pixel-wide images (equal to a scanning width of 18 mm) are 16-bit greyscale TIFF images. The images were converted into greyscale values using ImageJ 1.38X (http://rsb.info.nih.gov/ij/index.html). The greyscale value for each 0.5-mm measurement is presented as a mean of the greyscale values obtained from the 36 pixels image width.

**Magnetic susceptibility**

Magnetic susceptibility was measured immediately after core collection, but prior to core splitting, with a Bartington Whole Core Loop Sensor every 4 cm in L1 and L2 to confirm that the 3-m segments from each core were overlapping. Magnetic susceptibility was then re-measured every 4 mm on the split-core surface in L1 using a Bartington MS2E1 Surface Scanning Sensor connected to a Bartington MS2 Magnetic Susceptibility Meter and a TAMISCAN-TS1 automatic conveyor interfaced to a PC. These methods are described by Sandgren & Snowball (2001).

**Biogenic silica**

The concentration of biogenic silica (BSi) was measured using the method described by Conley & Schelske (2001). Samples (~0.03 g) were taken at every fifth couplet at 1.2–86.0 and 333.5–380.7 cm depths. Between 86.0 and 333.5 cm, samples were collected at increments that varied between 1.1 and 9.8 cm. This variable sampling resolution was necessary to avoid tephra layers. The results of BSi are presented as weight-% of SiO₂ of total dry weight.

**137Cs analysis**

137Cs atmospheric fall-out has been detected in soils all over Iceland, including in areas close to Lake Lögríður (Sigurgeirsson et al. 2005). Because clay is the most binding agent of 137Cs in mineral soils, clay-rich lake sediments are suitable for 137Cs analysis, as opposed to organic-rich soils in which 137Cs is more mobile (Staunton et al. 2002). Significant levels of 137Cs from nuclear weapon tests were detected in the atmosphere in 1954 (Pennington et al. 1973), and fall-out in Iceland began four years later (Sigurgeirsson et al. 2005).

Fourteen samples (~200 mg, ~1.5-cm sections) were collected at various depths from the top 46.2 cm of the surface core from L1, which was assumed to cover at least the last ~60 years, including the Northern Hemi-

sphere fall-out peak in AD 1963 (Pennington et al. 1973) and the Chernobyl accident in AD 1986. Owing to the amount of sediments needed for the analyses, each sample covers about 2–3 couplets. The samples were measured by gamma spectrometry in a calibrated geometry over two days using a well-type HpGE detector.

**Tephra analysis**

Owing to the frequent volcanic eruptions in Iceland, tephra layers commonly occur in soils and sediments. Volcanic eruptions since the settlement of Iceland c. AD 870 are often mentioned in historical records, with quite accurate dates, and prehistoric tephra layers can often be dated by 14C of peat soils surrounding the tephras. Previous work has resulted in a detailed tephrochronology for Iceland throughout the Holocene (Haflidason et al. 2000; Larsen & Eiríksson 2008).

Numerous tephra layers occur in L1 and L2. Eight tephra horizons from the top 3.8 m of L1 were sampled to determine their chemical composition (these were also found in equivalent stratigraphic positions in L2). The chemical compositions of the tephras were compared with those of independently dated tephras elsewhere (Larsen 1982; Larsen et al. 1999) in order to identify their origin.

Major element geochemical analysis of the volcanic glass shards were determined by standard wavelength dispersive techniques on a JEOL JXA-8200 microprobe at the Institute for Geography and Geology, University of Copenhagen. Prior to the measurements, tephra samples were extracted by wet-sieving (>63 micrometre) and briefly treated with 10% HCl before they were impregnated and mounted on thin sections. The methods used for the tephra analysis were modified after Wastegård (2002).

**Meteorological and hydrological data**

Meteorological data were obtained from Véðurstofa Islands (Icelandic Meteorological Office, http://www.vedur.is). Temperature records used for this study were from the Hallormsstaður meteorological station (AD 1961–1989 and 1997–2005), situated east of Lake Lögríður between the inlets of Grímsá and Jökulsá í Fljótsdal, and from Egilsstaðir (AD 1990–1996), situated east of the northern basin of Lake Lögríður. The record from Hallormsstaður is based on daily observations, while the Egilsstaðir record is based on monthly means. Precipitation data were obtained from the Grímsárvirkjun meteorological station (AD 1961–2005), located in the Grímsá catchment. The record from Grímsárvirkjun is based on daily observations.

Discharge records of Grímsá and Jökulsá í Fljótsdal were obtained from Orkustofnun (2006). Both records
are based on daily observations between AD 1963 and 2005.

Statistics

Confidence intervals of Pearson product-moment correlation coefficients ($r$) were determined by means of bootstrapping (Wilks 2005), using 1000 re-samples. The 99% confidence interval for each $r$ is presented within square brackets.

Results

Core sedimentology and stratigraphy

The sediments in the top 3.81 m of L1 are minerogenic and consist of dark-coloured brownish clayey-silty laminae capped by light-coloured whitish-greyish clayey laminae (Fig. 4). These dark- and light-coloured couplets are found throughout the section, and the thickness of individual couplets ranges from 1.3 to 35 mm.

The boundaries between the dark-coloured basal lamina (herein referred to as ‘summer lamina’) and the upper light-coloured clay lamina (herein referred to as ‘winter lamina’) are distinct. No coarse particles are found in the winter laminae, and they do not exhibit any internal layering. Grain-size analysis using a Sedigraph revealed that these laminae have a clay content of ~77%. The summer laminae are more varied, coarser (~54% clay), and do in some cases consist of two or more sub-laminae. In the summer laminae where two sub-laminae are found, there is typically a coarse sub-lamina followed by an upper sub-lamina that consists of a fine-grained matrix mixed with coarser particles. In some cases the fine-grained sub-lamina is followed by an additional sub-lamina similar to the lowermost coarse sub-lamina. The boundaries between the sub-laminae within a summer lamina are in general gradual.

In addition, we find a number of atypical thick dark brown laminae that are clearly different from the summer and winter laminae. They are uniformly silty laminae with fairly sharp upper and lower boundaries, and are much darker in colour than the summer laminae.

Throughout the 3.81-m section the sediments are distinctly cyclic in colour, with alternating 5–20 cm intervals of more greyish and brownish sediments. The greyish sections are characterized by thicker greyish clay-rich laminae, whose dominance gradually fades out, until a new cycle of distinctly more greyish laminae begins.

X-radiography and X-ray fluorescence analyses

The XRF data display a distinct pattern for Fe and Ca in the top 3.81 m of L1. We note that these elements are anti-correlated. By using the ratio between Fe and Ca (Fe/Ca) this signal is enhanced, and has been shown to be a very good indicator for identifying and separating the summer and winter laminae (Fig. 5). The Fe/Ca ratio declines upwards until it reaches a minimum in the centre or in the upper part of the summer lamina. The ratio then starts to increase and reaches a maximum in the winter lamina before it starts to decrease again in the following summer lamina. This pattern is seen in most parts of the section; the capping winter laminae in most cases produce peaks in the Fe/Ca ratio, while variations can be seen in the summer laminae where sub-laminae are observed. In cases where a summer lamina consists of a fine-grained sub-lamina in between two coarse sub-laminae, the middle part of the summer lamina displays a peak in Fe/Ca.
The radiographic images of the section more or less mirror the Fe/Ca ratio. The winter laminae are shown as light areas in the images (low-density areas), whereas the summer laminae appear as dark areas (high-density areas). The radiographic images also show that the inner part of each core scanned in the ITRAX Core Scanner, namely a central core of ~18 mm diameter, was not disturbed by the coring. As expected, there are no signs of bioturbation in this varve sequence.

The cyclic and recurring greyish and brownish intervals seen throughout the sequence are best revealed by the Fe cps-values in the XRF data (Fig. 6). The greyish intervals display relatively high Fe cps-values, reflecting thicker winter laminae, while the Fe cps-values are relatively low in the brownish sections. These cycles are most distinct in the lower half of the section.

**Magnetic susceptibility**

As would be expected for Icelandic mineral-rich sediments of predominantly volcanic origin, the magnetic susceptibility is relatively high (average in the order of 300×10⁻⁵ SI). The recurring greyish and brownish intervals are also mimicked by the surface magnetic susceptibility scan data (Fig. 6). We note that in most parts where the XRF-derived Fe intensity is high, the magnetic susceptibility is low, and vice versa.

The anti-correlation between XRF-derived iron concentration and magnetic susceptibility can appear paradoxical, but there is a relatively simple explanation that invokes at least two different sources of sediment in the lake. The data tell us that at least one sediment source has higher concentrations of iron than the other(s). Similarly, at least one sediment source has higher concentrations of a strongly magnetic mineral, such as magnetite, than the other source(s) that have higher concentrations of iron. We can be confident that post-depositional iron sulphides do not contribute to the sediment magnetic properties because of the lack of organic matter and reducing conditions. Thus, only a small amount of a ferrimagnetic iron oxide, such as titanomagnetite (Fe₂₋ₓTiₓO₄), needs to be present in the source (or sources) of the brownish sediments to raise the magnetic susceptibility to values that are significantly higher than in the greyish sediment, which is derived from a source (or sources) that contain(s) more iron, but in a less magnetic form, such as paramagnetic olivine ((MgxFe1₋ₓ)2SiO₄). Our observations suggest that the greyish sediments delivered to Lake Lögurinn via Jökulsá í Fljótárdal have very different magnetic properties from the brownish sediments, which are most likely delivered via Grímsá.

**Biogenic silica**

Similar to Fe cps-values and magnetic susceptibility, BSi concentration is also reflected in the recurring greyish and brownish intervals (Fig. 6). BSi concentration is low in the clay-dominated light-coloured intervals, and high in the brown-coloured coarser intervals. Hence, changes in BSi concentration are comparable to variations in magnetic susceptibility.

**Lamination/couplet counts and their validation by independent chronologies**

A composite sequence was established using overlapping sections from L1 for counting laminations and for establishing the chronology. The top 3.81 m of L1 was chosen to test the hypothesis that the sequence consists of varves; the rather frequent occurrence of identified tephra layers can be used as time markers in the sediments, and the ¹³⁷Cs analyses can test the varve counting in the youngest sediments. This sequence also displays fairly distinct laminations, whereas further down in the sediments the laminations are less clear and
to some extent disturbed by the coring. Even though tephra layers are frequent below 3.81 m, the next tephra that has been identified is the Hekla-3 tephra (c. 3000 cal. a BP) at 10.31 m depth. No terrestrial macrofossils, which could be used for $^{14}$C dating, were found between 3.81 m and the Hekla-3 tephra.

The visual lamination count initially revealed 162 light- and dark-coloured laminae (i.e. 162 couplets) in the upper 86 cm of L1. When optical and radiographic images and XRF data (Fe/Ca) were acquired and examined, the final count revealed 131 light- and dark-coloured laminae in the same interval. This difference indicates an initial overestimation error of up to $\sim$19%. Most of this error comes from the counting of fine-grained sub-laminae found in some of the summer laminae, which initially were interpreted incorrectly as winter laminae.

The coarse dark brown layers occurring at a number of levels in the stratigraphy are interpreted as being formed during random events, such as sudden rain storms or spring floods, or even snow-melt events resulting in fluvial sediments on top of the lake ice that later fall to the bottom, and not by processes forming the normal laminations. Thus, they were excluded from the couplet count. The final count for the top 3.81 m of L1 revealed 620 couplets.

The uppermost couplet (i.e. lamination number 1–2) is assumed to correspond to AD 2005, and the result of
the $^{137}$Cs analyses from the top 46.2 cm of L1 (Table 1) supports the hypothesis that the sediments are annually laminated (i.e. varved). The distinct $^{137}$Cs peak at 30.5–31.9 and 32.3–33.9 cm depth is interpreted to represent the main fall-out peak in the Northern Hemisphere, and it comprises laminations 81–92, which according to the couplet count were deposited in AD 1960–1965 (Fig. 7). Lamination 85–86 corresponds to AD 1963, the year of peak fall-out, and is located in between the two peak samples. Furthermore, we interpret the small increase at 12.9–14.5 cm as the fall-out signal of the Chernobyl accident in AD 1986, which concurs with the observation that this was a minor fall-out in Iceland (Pállsson et al. 1994); the sample at 12.9–14.5 cm depth comprises laminations 37–40, which were deposited in AD 1986–1987 according to the couplet count. Samples at 42.6–44.1 and 44.8–46.2 cm depth do not show any measurable $^{137}$Cs activity and were deposited prior to AD 1954.

Although it is reasonable to suggest that the sediments are varved, the resolution of the samples used for measuring $^{137}$Cs activity is too low to pinpoint specific years; that is, the two peaks cannot be precisely defined as AD 1963 and 1986, respectively. When applying the tephrochronology (Table 2) to the couplet count it should become clear if varves can be identified.

The couplet count revealed 131 light- and dark-coloured pairs of laminae from the Lög-1 tephra (Table 3) to the uppermost couplet in the surface core, which corresponds to the number of years between the youngest couplet year, AD 2005, and the age of the Lög-1 tephra, AD 1875. Further down in the sediments, when our couplet counts are compared with the identified tephra layers (Fig. 8), we find that the number of couplets equals the number of years between identified tephras for the following time periods: AD 1262–1353, AD 1354–1361, AD 1362–1409, AD 1410–1476, AD 1619–1628 and AD 1629–1874. This finding supports our hypothesis that the couplets are indeed varves.

### Table 1. Results of the $^{137}$Cs analyses, indicating a distinct $^{137}$Cs peak in samples 10 and 11, and also a small increase in sample 4.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (cm)</th>
<th>$^{137}$Cs (Bq kg$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0–1.75</td>
<td>6.2±0.5</td>
</tr>
<tr>
<td>2</td>
<td>6.95–8.39</td>
<td>7.3±0.5</td>
</tr>
<tr>
<td>3</td>
<td>9.81–11.59</td>
<td>7.0±0.9</td>
</tr>
<tr>
<td>4</td>
<td>12.94–14.54</td>
<td>13±1</td>
</tr>
<tr>
<td>5</td>
<td>15.91–17.63</td>
<td>11±1</td>
</tr>
<tr>
<td>6</td>
<td>19.41–20.98</td>
<td>10±2</td>
</tr>
<tr>
<td>7</td>
<td>26.12–27.54</td>
<td>16±1</td>
</tr>
<tr>
<td>8</td>
<td>27.54–28.81</td>
<td>21±1</td>
</tr>
<tr>
<td>9</td>
<td>28.81–30.31</td>
<td>31±1</td>
</tr>
<tr>
<td>10</td>
<td>30.50–31.88</td>
<td>65±2</td>
</tr>
<tr>
<td>11</td>
<td>32.30–33.88</td>
<td>61±1</td>
</tr>
<tr>
<td>12</td>
<td>34.59–35.69</td>
<td>28±1</td>
</tr>
<tr>
<td>13</td>
<td>42.55–44.12</td>
<td>0</td>
</tr>
<tr>
<td>14</td>
<td>44.82–46.18</td>
<td>0</td>
</tr>
</tbody>
</table>

### Table 2. Depth, origin and time of deposition of the identified tephra layers from L1.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (cm)</th>
<th>Origin</th>
<th>Age (AD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lög-1</td>
<td>86.0–87.8</td>
<td>Askja</td>
<td>1875</td>
</tr>
<tr>
<td>Lög-2</td>
<td>266.0–266.5</td>
<td>Grímsvötn</td>
<td>1629</td>
</tr>
<tr>
<td>Lög-3</td>
<td>270.5–270.8</td>
<td>Grímsvötn</td>
<td>1619</td>
</tr>
<tr>
<td>Lög-4</td>
<td>282.7–285.9</td>
<td>Veidivötn</td>
<td>1477</td>
</tr>
<tr>
<td>Lög-5</td>
<td>315.4–315.7</td>
<td>Veidivötn</td>
<td>1410</td>
</tr>
<tr>
<td>Lög-6</td>
<td>333.1–333.5</td>
<td>Örafjökull</td>
<td>1362</td>
</tr>
<tr>
<td>Lög-7</td>
<td>336.9–337.7</td>
<td>Grímsvötn</td>
<td>1354</td>
</tr>
<tr>
<td>Lög-8</td>
<td>380.4–381.0</td>
<td>Katla</td>
<td>1262</td>
</tr>
</tbody>
</table>

### Table 3. Summary of the number of couplets found between the identified tephras in L1.

<table>
<thead>
<tr>
<th>Time period (AD)</th>
<th>Depth (cm)</th>
<th>No. of cal. years</th>
<th>No. of couplets</th>
<th>Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1875–2005</td>
<td>0.3–87.8</td>
<td>131</td>
<td>131</td>
<td>0</td>
</tr>
<tr>
<td>1629–1875</td>
<td>87.8–266.5</td>
<td>245</td>
<td>245</td>
<td>0</td>
</tr>
<tr>
<td>1619–1628</td>
<td>266.5–270.8</td>
<td>10</td>
<td>10</td>
<td>0</td>
</tr>
<tr>
<td>1477–1618</td>
<td>270.8–285.9</td>
<td>142</td>
<td>18</td>
<td>–126 couplets</td>
</tr>
<tr>
<td>1410–1476</td>
<td>285.9–315.7</td>
<td>67</td>
<td>67</td>
<td>0</td>
</tr>
<tr>
<td>1362–1409</td>
<td>315.7–333.5</td>
<td>48</td>
<td>48</td>
<td>0</td>
</tr>
<tr>
<td>1354–1361</td>
<td>333.5–337.7</td>
<td>8</td>
<td>8</td>
<td>0</td>
</tr>
<tr>
<td>1262–1353</td>
<td>337.7–381.0</td>
<td>92</td>
<td>92</td>
<td>0</td>
</tr>
</tbody>
</table>
In the light of the good match between tephra ages and couplet counts it is surprising that only 18 couplets could be distinguished between the Lög-3 and Lög-4 tephra layers (AD 1477–1618), that is, in 142 years of sedimentation. We find no sign of changes in the sedimentation pattern; the couplets in this interval display the same general character as couplets deposited during other time periods. Furthermore, neither of the cores indicates any sign of erosion surfaces, and couplet counting in L2 results in the same low number of couplets. Figure 9 shows the composite varve thickness record for Lake Lögurinn for the time period AD 1262–2005, including summer and winter laminae thicknesses. The time period AD 1477–1619 is excluded in the varve record because of the apparent hiatus between the Lög-3 and Lög-4 tephras.

Fig. 8. A. Chemical composition (K₂O/SiO₂) of silicic tephras (Lög-1 and Lög-6) from L1 compared with tephras independently dated elsewhere (Larsen et al. 1999). B. Chemical composition (FeO/TiO₂) of basaltic tephras (Lög-2 to Lög-5, Lög-7, Lög-8) found in L1 compared with tephras independently dated elsewhere (Larsen 1982). This figure is available in colour at http://www.boreas.dk.

Fig. 9. Composite summer laminae, winter laminae and varve thickness records for Lake Lögurinn for the time period AD 1262–2005. Data for the time period AD 1477–1619 are excluded because of the hiatus between the Lög-3 and Lög-4 tephras.
Discussion

Varve formation

Daily average discharge records from the main rivers that enter Lake Lögurinn show a distinct seasonal pattern (Fig. 3). We can assume that the bulk of suspended matter throughout a single year is transported to the lake by Grímsá and Jökulsá í Fljótstdal when their discharge rates are highest, that is, between May and September. During the spring melt the coarser sediments forming the bulk of the summer laminae are transported by Grímsá and deposited fairly soon after entering the northern basin of the lake. The small amounts of coarse sediments transported in Jökulsá í Fljótstdal are probably deposited prior to reaching the northern basin because of the sill that separates the southern basin from the central and northern basins, while most of the rock flour that forms the winter laminae stays in suspension until the flow rate in the lake is low, namely in winter when the lake usually is ice-covered and thus less wind-exposed. Consequently, the summer laminae are formed mainly by sediments transported by Grímsá, while the winter laminae are dominated by rock flour transported in Jökulsá í Fljótstdal.

Previous studies in Arctic regions have demonstrated that varve thickness can often be related to specific meteorological parameters, for example monthly or seasonal temperatures (e.g. Itkonen & Salonen 1994; Moore et al. 2001; Thomas & Briner 2009). These studies have in common that sediment transport and deposition occurs within one hydrologic year. We believe that the varve formation in Lake Lögurinn is more complex. Owing to the water depth of 38 m at the L1 coring site and the small grain size of the rock flour we suggest, according to Stoke’s Law (Stokes 1851), that there is a time-lag from the time that the rock flour enters the lake until the final settling of the clay that forms the winter laminae. If it is assumed that the mean grain size of the rock flour forming the winter laminae is 1–3 micrometres, it would take anywhere between 70 and 600 days for these particles to fall 38 m in still water. Because the water is in motion for most of the year in the lake, the sedimentation time is further delayed. In addition, the longer transport time, from the rock-flour-producing glacier, through the long river and lake system until the suspended matter reaches the northern basin, will result in a considerable time-lag. Although flocculation (Hodder & Gilbert 2007; Hodder 2009) may accelerate the settling of the rock flour, resulting in the deposition of some of the rock flour well within a hydrologic year, we hypothesize that there is a time-lag of at least one year from the time that the rock flour leaves the glacier until all of the suspended matter settles at the 38-m-deep L1 coring site.

The uppermost coarse sub-laminae that are found in some of the summer laminae are interpreted as being formed by additional sediment delivery from Grímsá associated with heavy rains and anomalous early snowmelt during autumns and winters, when most of the annual precipitation occurs.

Meteorological control of discharge rates

The mean June–September discharge rates in Jökulsá í Fljótstdal are positively correlated with mean temperature for the same period between AD 1963 and 2005 ($r = 0.72 \, [0.47 \, 0.87]$) (Fig. 10A). Correlation between discharge rates and total precipitation is low and not statistically significant ($r = 0.33 \, [-0.11 \, 0.61]$).

The mean annual discharge rates in Grímsá are positively correlated with total annual precipitation ($r = 0.82 \, [0.59 \, 0.93]$) between AD 1963 and 2005 (Fig. 10B). Correlation between mean annual discharge rate and mean annual temperature for the same time period is moderate but statistically significant ($r = 0.46 \, [0.14 \, 0.68]$).

Although the magnitude of discharge associated with spring floods is related to several factors (Sander et al. 2002) there is a clear relationship between discharge rates in Jökulsá í Fljótstdal and temperature, while discharge rates in Grímsá are closely linked to the amount of precipitation in the catchment area.

We suggest, therefore, that temperature is the most important meteorological parameter controlling the discharge rates in Jökulsá í Fljótstdal, and thus the amount of suspended matter transported in the glacial river and deposited as winter laminae. Discharge rates in Grímsá, and the amount of coarser, mostly suspended matter from the Grímsá catchment, forming the bulk of the summer laminae, are mostly related to precipitation.

Meteorological influence on laminae thickness

Consolidation of sediments can generate depth-dependent trends in measured laminae thickness if the sediment is not fully compacted close to the sediment–water interface (e.g. Hughen et al. 2000). The radiographic images do not reveal any trend of decreased sediment bulk density towards the top of the surface core, possibly as a result of the low amount of organic material, and therefore we have not accounted for changes arising from compaction in the lamination measurements further down in the sediments. Because the longest continuous time series for both temperature and precipitation based on daily observations covers AD 1961–1989 we have chosen not to use the upper 11.6 cm of the surface core for comparisons between lamina/varve thicknesses and meteorological parameters as it contains sediments deposited after AD 1989. This will
also remove the likely issue of non-fully consolidated sediments at the very top of the surface core.

Correlation coefficients between lamina/varve thickness and meteorological parameters for various periods between AD 1961 and 1989 are shown in Table 4. No statistically significant correlation is found between winter laminae thickness and any meteorological parameter. It would be expected that the correlation coefficient would be higher and statistically significant between summer temperature and winter laminae thickness, because higher temperature throughout the melt-season should enhance sediment delivery from Eyjabakkajökull to Lake Lögurinn. However, as Eyjabakkajökull is a surging glacier the amount of suspended matter transported to the lake in Jökulsá Fljótsh is heavily dependent on whether the glacier is in a surging phase. During surges, the amount of suspended matter from a glacier typically increases (Humphrey 1985), as does the discharge rate (Gilbert et al. 2002). Hence, even though the discharge in Jökulsá i Fljótsdal is climatically controlled by summer temperature (Fig. 10A), surge events will probably enhance sediment delivery to Lake Lögurinn and thus alter the ‘normal’ relationship between temperature and sediment delivery from Eyjabakkajökull. Thus, a reconstruction of any meteorological parameter based on winter laminae thicknesses is not possible without excluding periods when Eyjabakkajökull is surging.

Total autumn/winter (September–February) precipitation is positively correlated with summer laminae thickness ($r = 0.57$ [0.19 0.76]). During the autumn/winter season the correlation is highest when summer laminae thickness is compared with winter (December–February) precipitation ($r = 0.70$ [0.27 0.86]) (Fig. 11). Autumn and winter are the seasons when precipitation is highest in the area (Fig. 2); for example, nine of the top-10 days with the most precipitation recorded at Grímsvörvirkjun meteorological station since AD 1961 have occurred during autumn/winter (Veðurstofa Islands). Because the average mean temperature is close to 0°C at this time of the year, precipitation will fall as both snow and rain (Einarsson 1984). For days when the temperature is above 0°C, heavy rains will increase the discharge rapidly in Grímsá. Similar to the precipitation record, the discharge record of Grímsá shows that nine of the top-10 days with the highest discharge since AD 1963 have occurred during autumn/winter (Orkustofnun 2006). During these discharge peaks, suspended matter will be transported out into the lake, resulting in the formation of a coarser upper sub-lamina in the summer laminae. Based on these results, winter precipitation (December–February) has been reconstructed from the AD 1262–2005 summer laminae thickness record (Fig. 12).

**Table 4.** Pearson product-moment correlation coefficient for various comparisons between laminae/varve thickness and mean temperature recorded at the Hallormsstadir meteorological station and total precipitation recorded at the Grímsvörvirkjun meteorological station (Veðurstofa Islands). Spring/Summer refers to March–August, and Autumn/Winter refers to September–February. Bold values are statistically significant at $> 99%$.

<table>
<thead>
<tr>
<th>Time period</th>
<th>Mean T Total P</th>
<th>Mean T Total P</th>
<th>Mean T Total P</th>
<th>Mean T Total P</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Summer laminae</td>
<td>Winter laminae</td>
<td>Varve</td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.24 0.40</td>
<td>0.34 0.17</td>
<td>0.33 0.36</td>
<td>0.36</td>
</tr>
<tr>
<td>Mar–May</td>
<td>0.18 0.40</td>
<td>0.36 0.42</td>
<td>0.31 0.49</td>
<td></td>
</tr>
<tr>
<td>Jun–Aug</td>
<td><strong>0.45</strong> 0.13</td>
<td><strong>0.01</strong> 0.30</td>
<td><strong>0.09</strong> 0.30</td>
<td></td>
</tr>
<tr>
<td>Sept–Nov</td>
<td>–0.17 0.04</td>
<td>0.25 0.11</td>
<td>–0.11</td>
<td>0.59</td>
</tr>
<tr>
<td>Dec–Feb</td>
<td>0.21 <strong>0.70</strong></td>
<td>0.22 0.14</td>
<td><strong>0.59</strong> 0.38</td>
<td></td>
</tr>
<tr>
<td>Spring/Summer</td>
<td>0.32 0.29</td>
<td>0.31 0.36</td>
<td><strong>0.38</strong> 0.38</td>
<td></td>
</tr>
<tr>
<td>Autumn/Winter</td>
<td>0.27 <strong>0.57</strong></td>
<td>0.01 <strong>0.05</strong></td>
<td><strong>0.18</strong> 0.36</td>
<td></td>
</tr>
</tbody>
</table>

**Fig. 10.** A. Mean June–September discharge rates in Jökulsá i Fljótsdal (Orkustofnun 2006) (solid line) compared with mean June–September temperatures (Veðurstofa Islands) (dashed line) between AD 1963 and 2005, $r = 0.72$ [0.47 0.87]. B. Mean annual discharge rates in Grímsá (Orkustofnun 2006) (solid line) compared with total annual precipitation (Veðurstofa Islands) (dashed line) between AD 1963 and 2005, $r = 0.82$ [0.59 0.93]. The correlation coefficients ($r$) were calculated by removing the linear trends in (A) and (B) by computing the least-square fit of straight lines to the data, and then subtracting the results.
It can also be noted that summer temperature is moderately but significantly correlated ($r = 0.45 \pm 0.09$) with summer laminae thickness. June–August coincides with the period when discharge rates in Jökulsá Í Fljótadal are high, and hence with when most of the suspended matter from Eyjabakkajökull is transported to the lake. The moderate correlation probably indicates that the melting of Eyjabakkajökull to some extent influences the summer laminae thickness. The suspended matter from the glacier deposited in the summer laminae is seen as a finer upper sub-lamina, which is often mixed with coarser particles originating from Grimsvötn.

Varve thickness has also been tested against temperature and precipitation. We find that varve thickness is positively correlated with mean temperature during spring/summer ($r = 0.38 \pm 0.08$) and with autumn/ winter precipitation ($r = 0.36 \pm 0.01$). This is not surprising, as we know that these two climatic parameters are correlated with the thickness of winter and summer laminae, respectively. Thus, we suggest that a climatic signal in the sediments from Lake Lögurinn is best inferred from winter or summer laminae thickness alone. Nevertheless, the moderate correlations between varve thickness and mean spring/summer temperature and autumn/winter precipitation suggest that the annual amount of sediments delivered to Lake Lögurinn is controlled mainly by melt-season temperatures and late-season precipitation.

**Surges of Eyjabakkajökull and their impact on the sedimentation in Lake Lögurinn**

The remoteness of Eyjabakkajökull means that historical evidence of its fluctuations is rather sparse and...
exists only from events after AD 1890, when the glacier probably surged to its most advanced position during the Holocene (Sharp & Dugmore 1985; Sigurðsson 1996). Observations of surges after AD 1890 have been documented in AD 1931, 1938 and 1972 (Björnsson et al. 2003), as has a possible surge in 1894 (Björnsson et al. 2003). The glacier is made up of a western and an eastern branch, and it is possible that more than one surge could have occurred within a fairly short time interval if the two branches did not surge simultaneously, which might explain the close surges in 1890/1894 and 1931/1938.

The sediment load in Jökulsá í Fljótshlíð increased during the most recent surge. Observations show that the river became unusually dark in colour in late August 1972 (Sigurðsson 1996), indicating an increased sediment load. Measurements of suspended matter in Jökulsá í Fljótshlíð between AD 1966 and 1995 (Pálsson & Vigfússon 1996) confirm these observations, as the highest amount of suspended matter recorded in the river throughout the time period was during the autumn of AD 1972 (Fig. 13), that is, in the initial stage of the surge. At this time, the amount of clay in the suspended matter exceeded 90% (Fig. 13), which is unprecedented throughout the record and far above the mean amount (Fig. 14). In total, from August 1972 to October 1973 the glacier advanced 1.5–2 km (Sigurðsson 1996), which agrees with the observation of Sigurðsson (1998), who reported an advance of 1350 m during 1972 and of 620 m in 1973.

According to the sediment record from Lake Lögurinn, the thickest winter lamina for the past 100 years was deposited in AD 1972 (6.79 mm), which was

---

**Fig. 13.** Upper panel: Percentage of clay in gauged grain size distribution in Jökulsá í Fljótshlíð between AD 1966 and 1995 (Pálsson & Vigfússon 1996). Clay constituted more than 90% of the suspended matter at the time of the AD 1972 surge. Middle panel: Concentration of suspended matter (clay, silt and sand) in Jökulsá i Fljótshlíð between AD 1966 and 1995 (Pállson & Vigfússon 1996). Highest concentrations are found during the initial stage of the AD 1972 surge (Pálsson & Vigfússon 1996). Lower panel: Discharge rates in Jökulsá í Fljótshlíð recorded during the same observations. Note that the number of observations in all panels is lower after c. AD 1985. This figure is available in colour at http://www.boreas.dk.

---

**Fig. 14.** Mean grain size distribution recorded in Jökulsá í Fljótshlíð between AD 1966 and 1995 (Pálsson & Vigfússon 1996).
followed by the second thickest in AD 1973 (6.07 mm). Thus, we can conclude that the recent surge in AD 1972 resulted in significantly thicker winter laminae. This is in agreement with Sharp (1988), who suggests that sedimentation rates are likely to increase dramatically in proglacial lakes during surge events.

The amount of suspended matter transported in Jökulsá í Fljótsdal does in general compare fairly well to discharge rates at the time of observations (Fig. 13), and hence the winter laminae thicknesses could have the potential to indicate past temperature. However, during the surge in AD 1972 the amount of suspended matter increased significantly, as shown, mostly as a result of increased clay content. Surprisingly, during this surge the observed discharge rate in Jökulsá í Fljótsdal was relatively low. The mean annual discharge rate in AD 1972 was 31 m$^3$/s, which can be compared to a mean annual discharge in Jökulsá í Fljótsdal of 37.5 m$^3$/s between AD 1963 and 2005, and it is the only year during which Jökulsá í Fljótsdal had a lower annual mean discharge rate than Grímsá. The low discharge in AD 1972 can partially explain why much of the sediments delivered during the surge were deposited the same winter.

However, even though the discharge rate during the initial stage of the surge was fairly low, we cannot exclude that a surge of Eyjabakkajökull can produce jökulhlaups. Háöldulón, a lateral ice-dammed (partly supraglacial) lake, seems to develop after surges of Eyjabakkajökull. No jökulhlaup from the lake was recorded between AD 1962 and 2015, but after the recent surge, jökulhlaups were observed 15 times in Jökulsá í Fljótsdal, from December 1972 until June 1996, as a result of drainages of Háöldulón (Sigurðsson 2003). Owing to a lack of data we are not able to compare most of the jökulhlaups with the amount of suspended matter in Jökulsá í Fljótsdal, but one of the most pronounced jökulhlaups occurred during 14–15th July 1975, when the maximum discharge reached 428 m$^3$/s. The amount of suspended matter reached about 75% of the amount of suspended matter recorded during the AD 1972 surge. In the sediment record we find an increase in winter laminae thickness from 2.15 mm and 2.02 mm in AD 1974 and 1975, respectively, to 3.96 mm and 4.60 mm in AD 1976 and 1977, respectively, which we interpret to be related to the AD 1975 jökulhlaup, given the time required for suspended matter to settle. This demonstrates the complexity of sedimentation in Lake Lögurinn, especially considering the surge in AD 1972, when sediments were deposited during the same year as the surge occurred.

The winter laminae thickness record covering the known surges since AD 1890 (Fig. 15) reveals a clear increase in laminae thickness at the time of the AD 1972 surge. The thickness of the laminae then gradually declines during the following ~20 years. The surge(s) in 1931/1938 is less clear in the record, although the thicknesses are relatively high during the 1930s. The 1890/1894 surge shows an increase in winter laminae thickness similar to that for the AD 1972 surge, with declining winter laminae thicknesses for the following ~20 years.

Based on these observations, we hypothesize that the recurring cyclic pattern of more greyish fine clay-dominated laminae throughout L1 (and in L2), whose dominance gradually fades out until a new cycle begins (Fig. 6), is related to past surge events of Eyjabakkajökull; in the initial stage of a surge the amount of sediments increases significantly in the river but the discharge remains fairly low. During the surge, the glacier advances and dams up Háöldulón. When the lake reaches a certain threshold, it recurrently drains into Jökulsá í Fljótsdal, delivering significant amount of rock flour to Lake Lögurinn as long as the ice dams the lake. However, the details behind this need to be further investigated. The recurring pattern seems to be more frequent during the time period AD 1619–1900 compared with AD 1262–1477 and compared with the 1900s, suggesting that surges of Eyjabakkajökull (and thus draining of Háöldulón) occurred more frequently during the later part of the Little Ice Age. This coincides with thicker and more variable summer laminae, inferred as reflecting increased precipitation.

**Past precipitation in the Lake Lögurinn area and its relationship to NAO and glacial fluctuations**

The NAO index has previously been shown to be correlated with winter precipitation in northern and western Iceland (Hurrell 1995). The geographical
variability of precipitation varies considerably within the island, even over short distances (Einarsson 1984), and thus the correlation with the NAO index probably varies within Iceland. Observed winter precipitation at Grímsvötn meteorological station between AD 1962 and 2005 is positively although not significantly correlated ($r = 0.36 [-0.19 0.72]$) with the NAO index [National Oceanic and Atmospheric Administration (NOAA), ftp://ftp.cpc.ncep.noaa.gov/wd52dg/data/indexes/tele_index.nh]. One explanation of why the observed precipitation is not significantly correlated with the NAO could be the influence of Vatnajökull; part of the Lake Lögurinn catchment is located in a rain shadow caused by the large icecap. This result is also in agreement with Hanna et al. (2004) and Jónsdóttir & Uvo (2009), who suggest that because Iceland is situated close to one of the dipoles of the NAO there is probably a lack of a strong correlation between the climate and hydrology in Iceland and the NAO index.

In the Northern Hemisphere several atmospheric teleconnection patterns have been observed besides the NAO (Barstow & Livezey 1987). The East Atlantic (EA) pattern is the second most prominent pattern over the North Atlantic region (NOAA, http://www.cpc.noaa.gov/data/teledoc/ea.shtml). It is structurally similar to the NAO, and is often referred to as a ‘southward shifted’ NAO pattern, as the anomaly centres are displaced southwards. Similar to the NAO, oscillations in the EA pattern change the amount of precipitation that falls over Europe: when the EA pattern is in a positive phase, northern Europe and Scandinavia experience above-average precipitation while southern Europe experiences below-average precipitation. The EA pattern (December–February) between AD 1962 and 2005 is, in contrast to the NAO, significantly correlated with winter precipitation in the Lake Lögurinn area ($r = 0.56 [0.15 0.75]$). In the EA pattern, the anomaly centres are shifted southwards, which could explain why we find a significant correlation with the EA pattern and not the NAO index. However, a long-term record of the EA pattern is needed in order to verify that there actually is a robust relationship between the EA pattern and the precipitation inferred from the summer laminae thicknesses in Lake Lögurinn.

To our knowledge, there is no high-resolution record from Iceland concerning precipitation during the past 700 years. Our reconstruction (Fig. 12) suggests that precipitation has varied substantially since the mid-1200s. Most notable is the relatively high, although highly variable, precipitation between the 1600s and late 1800s. Several studies have shown that glaciers in Iceland expanded within this time period (e.g. Bradwell et al. 2006; Kirkbride & Dugmore 2006), including Eyjafjallajökull (Sharp & Dugmore 1985). With Iceland situated in the maritime climate of the North Atlantic, increased winter precipitation could have been one parameter, along with reduced summer temperatures, that enhanced glacier growth at this time. Nesje et al. (2008) conclude that the main cause of early 1700s glacier advances in western Scandinavia was mild and humid winters associated with increased precipitation and high snowfall on the glaciers. We speculate that a similar pattern could be applied to the Lake Lögurinn area and Eyjafjallajökull; increased precipitation, as snow in the highlands and on the glacier, enhanced glacier growth between the 1600s and the late 1800s.

Sedimentation between AD 1477 and 1618

The period AD 1477–1618, marked by the Lög-3 and Lög-4 tephra layers, is covered by only 11.6 cm of sediments, containing 18 varves. The same drop in sedimentation rate and decrease in varves is observed in both L1 and L2. The cores were collected from the northern basin of the lake, about 350 m apart; in other words, the marked change is most probably represented in at least the whole northern basin. As previously stated, we do not observe any indication of erosion surfaces in any of the cores. Thus, we can only speculate about what might have caused this pronounced change in sedimentation rate, expressed as the absence of most of the varves from that time period.

AD 1477–1618 is the time period in which the climatic cooling during the Little Ice Age intensified in Iceland (e.g. Eiriksson et al. 2000; Geirsdóttir et al. 2009). An extreme cold spell could have resulted in a drastic decrease in meltwater and runoff at this time, but it would probably have been restricted to autumns and winters. At least during summer, suspended matter would still be transported to the lake, resulting in sedimentation.

Owing to the large-scale morphology, it is not likely that runoff and discharge in Jökulsá í Fljótsdal and the resulting sedimentation of glacial rock flour occurred elsewhere. The inlet of Grímsá could possibly have been shifted southwards, resulting in decreased sediment transport to the northern basin. If this was the case, and if the discharge in Jökulsá í Fljótsdal was much lower as a result of a harsh climate, it is possible that the amount of sediments delivered to the northern basin was significantly lower.

What stands out in this part of the section is the 26-mm-thick Veidivötn tephra deposited in AD 1477, one of the most explosive eruptions in historical time in Iceland (Wastegård & Davies 2009). This tephra layer is more than twice as thick as the second-thickest tephra layer in the section (Lög-1, 11 mm). If this eruption caused a pronounced tephra fall-out all over the area, including the catchment of both rivers, it is possible that the discharge, and hence the transport of suspended matter, was altered.
We cannot give a fully satisfactory explanation for the reasons behind the abrupt change in sedimentation rate and the lack of varves in this interval, nor can we conclude in which years sedimentation ceased in the northern basin – it needs to be investigated further. For now, we can only conclude that this interval coincides with the transition towards the peak of the Little Ice Age.

Conclusions

We have demonstrated that various properties of varves in sediments from Lake Lögurinn can be used to infer past hydrologic and climate conditions and glacial processes.

Summer laminae thickness has been used to reconstruct winter precipitation in the Lake Lögurinn area since AD 1262 (Fig. 12). Our record shows that precipitation was low and fairly stable from AD 1262 until AD 1476. The sudden shift in sedimentation rate and the lack of varves prevent us from reconstructing precipitation between AD 1477 and 1671, a time period for which documentary sources of past climate are lacking (Ogilvie & Jónsson 2001). From AD 1619 until the late 1800s (i.e. during the peak of the Little Ice Age), precipitation was higher and more variable compared to AD 1262–1476. During the 1900s, precipitation returned to lower and less variable levels compared with those that prevailed from the 1600s until the late 1800s.

Furthermore, we have demonstrated that winter laminae thickness increased significantly during the recent surge of Eyjafjallajökull. As a consequence of the surge, the ice-dammed Lake Háöldulöon recurrently drained into Jökulsá í Fjótsdal and delivered significant amounts of rock flour to Lake Lögurinn. If our interpretation of the recurring light- and dark-coloured sections in L1 and L2 is correct, the sediment record shows that surges of Eyjafjallajökull have occurred at least during the past ~750 years. Surges seem to have occurred more frequently during the later part of the Little Ice Age, between the early 1600s and late 1800s. Interestingly, the increased surge frequency coincides with a period of increased precipitation in the area, which will be dealt with in more detail in a forthcoming study.

Acknowledgements. – This study was supported by the Swedish Research Council (Vetenskapsrådet, grant to SB), the Icelandic Research Council (Rannís) and Landsvirkjun (grants to OI), which are gratefully acknowledged. Fieldwork during the reconnaissance phase of the project carried out by Anni Madsen and participation in fieldwork by Per Sandgren are also acknowledged. Constructive comments by Kyle Hodder and an anonymous reviewer improved the presentation of this paper.

References


