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A 3000-year varved record of glacier activity and climate change from the proglacial lake Hvítárvatn, Iceland

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ABSTRACT

A suite of environmental proxies in annually laminated sediments from Hvítárvatn, a proglacial lake in the central highlands of Iceland, are used to reconstruct regional climate variability and glacial activity for the past 3000 years. Sedimentological analysis is supported by tephrostratigraphy to confirm the continuous, annual nature of the laminae, and a master varve chronology places proxies from multiple lake cores in a secure geochronology. Varve thickness is controlled by the rate of glacial erosion and efficiency of subglacial discharge from the adjacent Langjökull ice cap. The continuous presence of glacially derived clastic varves in the sediment fill confirms that the ice cap has occupied the lake catchment for the duration of the record. Varve thickness, varve thickness variance, ice-rafted debris, total organic carbon (mass flux and bulk concentration), and C:N of sedimentary organic matter, reveal a dynamic late Holocene climate with abrupt and large-scale changes in ice-cap size and landscape stability. A first-order trend toward cooler summers and ice-cap expansion is punctuated by notable periods of rapid ice cap growth and/or landscape instability at ca 1000 BC, 600 BC, 550 AD and 1250 AD. The largest perturbation began ca 1250 AD, signaling the onset of the Little Ice Age and the termination of three centuries of relative warmth during Medieval times. Consistent deposition of ice-rafted debris in Hvítárvatn is restricted to the last 250 years, demonstrating that Langjökull only advanced into Hvítárvatn during the coldest centuries of the Little Ice Age, beginning in the mid eighteenth century. This advance represents the glacial maximum for at least the last 3 ka, and likely since regional deglaciation 10 ka. The multi-centennial response of biological proxies to the Hekla 3 tephra deposition illustrates the significant impact of large explosive eruptions on local environments, and catchment sensitivity to perturbations.

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1. Introduction

An important goal of paleoclimate research is to understand the causes and scale of past climate variability, including the associated feedbacks that resulted in distinct system changes on local and regional scales (Jansen et al., 2007). Recently, attempts to synthesize empirical data, multiproxy records, and computer simulations, have produced global and hemispheric temperature reconstructions placing twentieth century warmth within the context of the last few millennia (Mann and Jones, 2003; Luterbacher et al., 2004; Moberg et al., 2005; Jones et al., 2009; Kaufman et al., 2009). The results from these efforts depict a highly variable late Holocene

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climate, characterized by nonlinear expressions of change that have been associated with internal modes of variability (i.e. El-Niño and the North Atlantic Oscillation) and external forcing agents, including changes in solar irradiance, aerosol concentrations, and explosive volcanism (Stuiver et al., 1997; Shindell et al., 2001; Ammann et al., 2007; Gao et al., 2008; Jones et al., 2009; Mann et al., 2009; Trouet et al., 2009). Such findings are compatible with increasing evidence for abrupt (Mayewski et al., 2004) and periodic (Denton and Karlén, 1973; Bond et al., 2001; Hodell et al., 2001) climate change occurring throughout the Holocene, illustrating the dynamic nature of the climate system and the continued need for well-dated, high-resolution paleoclimate records.

Iceland is the largest landmass in the central North Atlantic, with a total area above sea level of ca 103,000 km². Paleoceanographic investigations performed around the Iceland shelf describe a variable late Holocene, with fluctuating sea-surface and bottom-water temperatures (Jiang et al., 2002, 2005; Bendle and Rosell-Melé,



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2007; Sicre et al., 2008; Patterson et al., 2010), biotic assemblages (Andrews and Giraudeau, 2003; Ran et al., 2008), drift ice (Massé et al., 2008; Andrews et al., 2009), and ocean currents (Eiríksson et al., 2004; Justwan et al., 2008; Ólafsdóttir et al., 2010). Although significant variability exists regarding the timing and magnitude of climatic events, there is a pervasive summer cooling trend after 3 ka. following the earlier mid Holocene warmth.¹ For example, during the past 2 ka. summer sea-surface temperatures (SST) fluctuated with amplitudes of 3 °C (Jiang et al., 2005), and some of the coldest temperatures of the last 10 ka were recorded during the last millennium (Andersen et al., 2004). However, despite the abundance of marine paleoclimate data and the thermally sensitive position of Iceland, comparatively few continuous Holocene terrestrial records exist, and uncertainties remain regarding the relationship between land and marine environments (e.g. Geirsdóttir et al., 2009a).

The few records with annual to decadal resolution for late Holocene terrestrial environments in Iceland are restricted to coastal lowland sites (e.g. Axford et al., 2009; Geirsdóttir et al., 2009b). Information on past glacier activity is fragmentary and is primarily concerned with the timing of maximum extension or rates of twentieth century retreat. Temperate glaciers presently cover 11% of Iceland and are sensitive to climate fluctuations (Björnsson and Pálsson, 2008), notably temperature changes (Flowers et al., 2005), yet little is known about their evolution through the late Holocene. Most reconstructions rely on dated moraine complexes in front of select outlet glaciers and small mountain glaciers (e.g. Gudmundsson, 1997; Stötter et al., 1999; Kirkbride and Dugmore, 2001, 2006; Geirsdóttir et al., 2009b), and the results are inherently limited and temporarily discontinuous.

Glacially derived lake sediments, deposited proximal to the ice limits, yet outside the region of disturbance, are best suited for obtaining constructive archives of the glacial activity and corresponding environmental change. This study capitalizes on this fact by targeting the proglacial lake Hvítárvatn, adjacent to Langjökull, second largest of Iceland's ice caps. We provide a 3000-year continuous, varved, multiproxy record directly related to late Holocene icecap activity in Iceland, that also provides important information on climate and environmental change in the central highlands.

2. Geographic and geologic setting

Hvítárvatn is a large, proglacial lake adjacent to the eastern margin of Langjökull in central Iceland (925 km²; Fig. 1a). The lake is situated 422 m a.s.l., has an area of 28.9 km² (ca 10×3 km), and a maximum depth of 83 m. Approximately one-third its 820 km² watershed is occupied by the ice cap which, through erosion and meltwater transport, governs the sediment flux. Hvítárvatn drains to the southeast over a bedrock sill that is overlain by large boulder residuals, and has an average monthly discharge of 45 $m^3 s^{-1}$ (range: $23-94 \text{ m}^3 \text{ s}^{-1}$; reference period: 1961-1990 AD), with an average suspended sediment load of 35 mg L^{-1} . The outlet stream Hvítá removes roughly half of the sediment delivered to the lake before it settles (Black et al., 2004). Major sediment sources are from two outlet glaciers, Suðurjökull and Norðurjökull, and two meltwater streams that emerge from the glacier farther north. Suðurjökull and Norðurjökull are warm-based outlet glaciers that have catchments of 58 and 61 km² respectively. Both outlet glaciers advanced into Hvítárvatn during the Little Ice Age (LIA; Black, 2008; Geirsdóttir et al., 2008), but have receded from their LIA maxima, which are clearly defined by lateral moraines, trimlines, and multibeam bathymetric data (Fig. 1; Geirsdóttir et al., 2008).

Suðurjökull receded from the lake ca 60 years ago, whereas Norðurjökull only exited the lake in 2009.

A seismic reflection survey of the lake bottom reveals over 65 m of stratified post-glacial sediment in the main depositional basin, distal to the LIA terminal moraines (Black et al., 2004). The presence of the Saksunarvatn tephra in the sediment fill confirms that the Iceland Ice Sheet had retreated from Hvítárvatn prior to 10.2 ka (Jóhannsdóttir, 2007). Bedrock surrounding the lake consists primarily of early Holocene basalt lava flows and late Pleistocene subglacial volcanics (i.e. pillow lavas intercalated with hyaloclastite tuffs, lapilli tuffs and breccias), overlain by alluvium and drift (Sinton et al., 2005). The sparse soils in the region are typically Andosols and are restricted to small, vegetated "islands" or areas near stream channels. Soil profiles commonly contain a coherent record of mid-to-late Holocene tephra layers, suggesting a nearcontinuous soil development. However, in recent time (post 1300 AD), the soil properties changed abruptly, becoming coarsergrained and indicating that enhanced wind erosion has significantly reduced the extent of soil and vegetation in the region (e.g. Geirsdóttir et al., 2009b). A notable exception to this degradation is the prograding Hvítárnes delta, an expansive wetland area along the northeastern shore that is well vegetated by Cyperaceae and Ericaceae (sedges and heath). Seismic profiles reveal thickening lacustrine sediment sequence towards this feature, suggesting active progradation of the meltwater fed delta (Black et al., 2004).

Field measurements of pH, specific conductance, dissolved oxygen, and water temperature confirm that the lake is well mixed. with little or no thermal stratification in summer (ΔT of water $column = 2.5 \circ C$). The Iceland Meteorological Office has maintained a weather station at Hveravellir (641 m a.s.l.), located approximately 30 km north of Hvítárvatn. This station provides a continuous record of temperature, precipitation and wind speed from 1966 to present. Mean annual temperature and precipitation for 1966–2003 are -0.9 °C and 730 mm, respectively. In most years, the lake is ice covered from November to April, although early winter thaws and/or late season ice cover may occur. Recent work on the distribution of mountain permafrost in Iceland indicates the presence of permafrost in regions with a thin or absent snow cover at elevations above 800-1000 m a.s.l., and in regions with mean annual air temperatures below -3 °C (Etzelmüller et al., 2007; Farbrot et al., 2007; Christiansen et al., 2010). Combining available data and field observations with the geographic and climatic conditions of the watershed, it can be concluded that persistent permafrost has not been present in the Hvítárvatn catchment to any significant extent during the late Holocene.

3. Materials and methods

3.1. Sediment cores and physical proxies

Ten sediment cores were recovered from four sites in Hvítárvatn using the DOSECC GLAD-200 core rig (http://www.dosecc.org/). This study focuses on three core sites: HVT03-1 in the central northern flats, and HVT03-2 and HVT03-3 located on a hyaloclastite ridge fronting the Norðurjökull terminal moraine (Fig. 1b and Table 1). Offset core pairs (and triplets), drilled from the same sites, ensure continuous sediment recovery, and surface cores were taken at each site, following Glew (1991), to capture the sediment—water interface and undisturbed upper sediments. The recovery of the spliced cores exceeded 95% at each individual coring site. Three-meter-long core sections were cut into 1.5 m lengths before being shipped to the Limnological Research Center (LRC) at the University of Minnesota for initial core processing, description, and sub-sampling. Continuous, whole-core density (gamma-ray) measurements were taken in 1 cm intervals using the gamma-ray attenuation porosity evaluator

¹ All ka ages used in this paper refer to 10³ calendrical years before 1950 AD.



Fig. 1. (a) Topographic map of field site showing position of Hvítárvatn adjacent to the Langjökull ice cap (gridlines on map are 1 km² for scale). Inset map highlights the position of the lake within Iceland. Current ice margins of the two outlet glaciers, Norðurjökull and Suðurjökull, are shown along with their approximate LIA maximum limits (dotted lines). The multibeam bathymetry has been superimposed to illustrate lake bottom morphology. Image modified from Geirsdóttir et al. (2008). (b) Bathymetric map of Hvítárvatn showing core sites in the northern basin. Cores HVT03-2 and HVT03-3 are located on a bedrock ridge in front of the Norðurjökull moraine, and HVT03-1 is situated in the northern flats proximal to the Hvítárnes delta.

(GRAPE) logging system. Core segments were then split, and core halves were photographed using a DMT CoreScan Color flatbed scanner. Medical X-ray images of core halves were taken at the Wardenburg Health Center, University of Colorado, and digitally scanned (at 300 ppi resolution) to better document sedimentary structures, including ice-rafted debris (IRD). In this study, IRD is defined as clasts with a grain diameter ≥ 2 mm and was counted in 20-year intervals for two size fractions (≥ 2 mm and ≥ 5 mm) according to the age model developed for each core. The counts were performed in Adobe Illustrator on scanned X-ray core images placed alongside respective Corescan images.

The upper third of each core consists of well-defined, olive black to light yellow laminations, composed of siliciclastic fine sand, silt and clay (Fig. 2a–c). Grain size analyses were conducted, using a Malvern laser diffraction particle size analyzer, on a selected suite

Table 1	
Hvítárvatn Sediment o	core information

Core ID	Latitude	Longitude	Water depth (m)	Spliced core length (m)
HVT03-1	64° 38.52 N	19° 50.42 W	36.5	18.2
HVT03-2	64° 38.60 N	19° 50.98 W	23.1	13.7
HVT03-3	64° 38.29 N	19° 51.57 W	30.4	12.1

of samples to examine the physical properties of individual laminae and to evaluate whether the laminated packages are clastic varves. Continuous, overlapping epoxy-impregnated (18×2 cm) sediment slabs were created following Francus and Asikainen (2001), from which thin sections were prepared at the University of Colorado. The thin sections were digitally scanned and microstratigraphic analysis was achieved with the aid of image enlarging and color/ contrast adjustments.

3.2. Biological proxies

The Hvítárvatn cores were sub-sampled for total carbon (TC) and total nitrogen (TN) to produce semi-quantitative proxies for paleoenvironmental changes taking place in the catchment. These concentrations are expressed as weight ratios and can be influenced by independent factors, such as sediment accumulation rates. Total carbon is derived exclusively from organic matter and represents total organic carbon (TOC) due to the lack of inorganic carbon in the lake sediment. Because the sedimentation rates in Hvítárvatn were found to be highly variable, TOC concentrations were converted to fluxes using: $TOC_Q = (TOC_{*})\rho_B Z$, with the yearly flux of carbon (TOC) equal to the product of TOC concentration (TOC_{*}), bulk density (ρ_B) and varve thickness (Z). In this manner, the



Fig. 2. Images of laminated sediments obtained by different techniques (a)–(c) and pie charts of grain size distribution from seasonal laminae (d). X-ray (a) and flatbed scan (b) are taken from the same section of a split core half and are displayed at the same scale (ruler in centimeters). Note the presence of IRD seen in the X-ray toward the bottom of the image. Black rectangle in (b) marks the thin section sampling interval seen in (c). Enlarged image of thin section (c) reveals clastic varve structure composed of three seasonal packages labeled by their varve age (year AD). Varve years were identified by the sharper contact at the top of the winter clay layer signifying the end of seasonal ice cover and the transition from winter to spring. Grain size analysis demonstrates the correlation between laminae color and sediment size with lighter colors containing a higher proportion of clay-sized particles (d). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

mass of carbon deposited annually at each core location could be evaluated through time, irrespective of changes in the flux of clastic material. Samples for TOC and TN were taken at 1–10 cm intervals from cores HVT03-1A/1B and HVT03-2A/2B/2C, and at 2.5 cm intervals from the surface core HVT03-1C.

To distinguish between the relative proportion of autochthonous and allochthonous organic matter (OM) in the lake sediment, the elemental mass ratio (C:N) of each sample was calculated. In most lacustrine systems, sedimentary OM is derived from a combination of aquatic and terrestrial sources, each with characteristic C:N signatures that are independent of sedimentation rate. C:N of aquatic algae, notably diatoms, is typically between 4 and 10, whereas vascular plants have values ≥ 10 (Meyers and Teranes, 2001). To calibrate the average terrestrial component of C:N, soil samples from terrestrial stratigraphic sections around Hvítárvatn were analyzed at the same time as the lake sediment. These samples include sediment deposition spanning the last \sim 7 ka, following the regional tephrostratigraphy (Kirkbride and Dugmore, 2006 and references therein). Freeze-dried samples were combusted at 1020 °C and measured at the University of California, Davis, using a PDZ Europa ANCA-GSL elemental analyzer.

3.3. Chronology

Tephrochronology and annual layer counting are the primary techniques used to constrain the ages of the sediment cores. Radiocarbon proved unreliable for dating Hvítárvatn sediment due to variable fluxes of aged terrestrial carbon reworked by wind and ice (Black, 2008).

3.3.1. Tephrochronology

Conspicuous basaltic and rhyolitic tephra of Icelandic origin offer a means for correlating between cores and for absolute dating. Written accounts (Íslendingabók) date the onset of permanent settlement of Iceland to 874 AD, and this event is demarcated in Holocene sediment archives by the so-called Settlement Layer (Vö): a widespread tephra layer that has been identified and dated in Greenland ice cores to 871 ± 2 AD (Grönvold et al., 1995). The tephrochronology of post-Settlement (historical) eruptions has been derived from written records dating back to the twelfth century, supplemented by dates from soil sections, lake sediments, and glacial ice (e.g. Thordarson and Larsen, 2007; Larsen and Eiríksson, 2008 and references therein).

Prominent tephra horizons in the Hvítárvatn cores were detected visually, and sub-sampled for chemical composition. Each tephra sample was sieved to obtain the optimum size fractions $(100-500 \ \mu m)$, and examined under a binocular microscope to verify the purity and to document the range of grain types before mounting into a probe plug. Sample surfaces were polished with 6 µm and 1 µm diamond paste and then cleaned in an ultrasonic bath for 40 min with deionized water before carbon coating. Major element glass analyses were carried out on a Cameca SX100 energy dispersive electron microprobe at the NERC Tephra Analysis Unit at the University of Edinburgh and University of Hawaii at Manoa using a 5-8 µm beam size, 15 kV acceleration voltage, 2 nA beamcurrent at Edinburgh, and 15 kV acceleration voltage, 15 nA beamcurrent and 8 \times 12 µm rastered and defocused beam at Hawaii (Jóhannsdóttir, 2007). The uncertainty on the major element analysis is <1%, and the set up used at Edinburgh circumvents the problem of Na-loss in silicic glasses during analysis. The international standards a99, VG2, BHVO2G and Lipari 1 were analyzed as unknowns at regular intervals (after a batch of 30-40 analyses) as a quality check and to monitor for instrument drift.

The geochemical composition of multiple size fractions from each tephra horizon was compared to the University of Edinburgh tephra glass chemistry database, and combined with stratigraphic information to identify source volcano and eruption, according to the procedures of Jóhannsdóttir (2007). One sample of disseminated organic matter found beneath a thick, well-preserved Hekla 3 (H3) tephra deposit exposed in a stream-cut proximal to Hvítárvatn was radiocarbon dated.

3.3.2. Lamination analysis

Laminated packages display a clastic varve structure (thin, finegrained winter layers separating thicker, coarser-grained summer deposits) and the strong seasonal variation in deposition resembles that of other glacially dominated lake systems (e.g. Desloges, 1994). In order to demonstrate that the laminated packages are produced annually and continuous in nature, we performed sedimentological analysis to identify seasonal components before developing a time series for each core. The preliminary layer counts were correlated to each other by identifying common marker beds, and then cross-dated using the positions of identified, historically dated tephra layers.

Annual sediment accumulation was measured using discernible differences between seasonal laminae color and grain size distribution. A typical year comprises a three component (seasonal) sequence of brownish olive fine silt (spring) that coarsens upward to olive black silt and fine sand (summer), before transitioning to a light yellow clay-rich winter layer. Because of the tendency for an individual sequence to be interrupted by intra-seasonal laminations, such as sand lenses interstitial to silt and clay deposits, the sharp transition at the top of each clay unit (signifying the end of seasonal ice cover) was used as the primary marker when identifying and measuring a single annual layer (Fig. 2c). Each seasonal component is physically differentiable based on sediment characteristics (i.e. grain size and color), and the annual stratigraphy is clearly visible in flatbed scanner, X-ray, and thin section images (Fig. 2a–c). Grain size analysis illustrates the seasonal depositional cycle and also demonstrates the relationship between sediment color and grain size, where lighter colors reflect an increased clay concentration and darker colors represent greater amounts of sand (Fig. 2d).

To test the continuous nature of the laminae, annual couplets were measured and counted manually from digital images of thin sections. The counts were performed independently on cores HVT03-1 and HVT03-2 down to the Vö tephra layer, before the two series were cross-correlated using distinct layer patterns and cleared of inconsistencies. After the two chronologies were annually correlated, laminae were counted on HVT03-3 for the past ~ 1 ka and then compared to the others. This layer count was initiated at the core top and continued down to a prominent rhyolitic tephra layer, at 4.01 m depth, that was common to all cores, and subsequently identified as the H1104 layer. The addition of the third chronology proved helpful in resolving conflicting layer counts in core segments that contained vague structures, poorly defined clay units, or disturbed sediments in one or both of the original two cores. Discrepancies between the preliminary chronologies were resolved by matching the records from the three core sites and correlating individual varves and varve sequences (e.g. Lamoureux and Bradley, 1996). Lastly, the stratigraphic positions and documented ages of seven historical tephra horizons were used to implement independent age verification and increase the accuracy of the three chronologies. In this manner, a tephra-constrained, varve chronology was created for the upper 3000 years of the Hvítárvatn sediment record.

4. Results

4.1. Tephrostratigraphy

The tephrostratigraphy around Hvítárvatn is well established from soil and lacustrine sequences (Kirkbride and Dugmore, 2006; Jóhannsdóttir, 2007; Jagan, 2010). Prominent tephra deposits in the region are derived from the Hekla, Katla and Veiðivötn volcanic systems. Major element composition of seven conspicuous tephra layers identified in the lake cores (Table 2) confirm their correlation with the historical eruptions of Hekla 1766, Katla 1721, Veiðivötn 1477, Hekla 1300, Hekla 1104, Eldgjá (934–942 AD), and Vatnaöldur ~870 (Settlement Layer (Vö); 871 \pm 2 AD), and the pre-historic Hekla 3 layer (Thorarinsson, 1967; Larsen and Thorarinsson, 1977; Larsen, 1984, 2000; Thordarson et al., 2001; Jagan, 2010). The composition of Hekla 3 (1080 \pm 180 BC) is presented as four separate groups due to the wide range in major element chemistry; complete chemical analyses for this and all other tephra are available as Supplementary Table 1.

4.2. Varve chronology

The initial varve chronologies for HVT03-1 and HVT03-2 were within 10% of the Vö tephra age and showed strong similarities in overall structure and varve thickness patterns. The close agreement of the two records to each other and with the tephra age, support the continuous varved nature of the sediment, and indicate that lake-wide changes in sediment accumulation rates have occurred in a coherent manner during the late Holocene. Both counts fell short of the 871 AD Vö tephra, with counting errors in HVT03-1 and HVT03-2 of 41 and 117 years (~4 and ~10%), respectively. The greater "under-counting" error in HVT03-2 reflects the lower sedimentation rate at this site and relates to the increased chances of missing varves when counting thinner laminae (e.g. Ojala and Tiljander, 2003).

Table 2

Major element chemistry of the seven historical and Hekla 3 tephra layers identified in the Hvftarvatn sediment cores. The standard deviation for each major element is shown in the second row of all tephra layers.

Tephra	Eruption	Varve	# samples	# of	SiO ₂	TiO ₂	Al_2O_3	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	$P_{2}O_{5}$	Total
	date (year AD)	age (year AD)	analyzed	analyses											
H1766	1766	1766	1	7	59.93	1.17	15.05	9.39	0.27	1.63	5.19	4.37	1.45	0.48	98.93
					0.88	0.13	0.08	0.51	0.02	0.23	0.27	0.08	0.07	0.09	0.46
K1721	1721	1721	2	42	47.57	4.54	12.98	14.63	0.23	4.82	9.51	3.08	0.80	0.59	98.73
					0.58	0.12	0.44	0.31	0.02	0.12	0.16	0.08	0.08	0.09	0.63
V1477	1477	1483	2	40	49.89	2.02	13.55	12.87	0.22	6.38	11.21	2.51	0.26	0.19	99.10
					0.68	0.43	0.32	0.82	0.02	0.47	0.63	0.14	0.08	0.05	0.67
H1300	1300	1300	2	27	59.68	1.38	14.59	9.28	0.25	2.00	5.35	4.23	1.72	0.47	98.95
					6.19	0.55	0.99	2.03	0.07	2.21	2.39	0.71	0.61	0.15	0.59
H1104	1104	1104	3	98	72.24	0.21	13.92	3.22	0.11	0.11	1.93	4.59	2.69	0.02	99.04
					0.67	0.02	0.38	0.21	0.01	0.02	0.13	0.23	0.14	0.01	0.74
Eldgj	938 ± 4	940	2	38	47.98	4.33	12.77	14.84	0.21	5.06	10.05	2.90	0.75	0.44	99.33
					4.71	1.08	1.13	2.53	0.04	1.23	1.64	0.34	0.38	0.14	0.99
Settlement	871 ± 2	871	2	95	49.67	1.95	13.54	12.93	0.22	6.39	11.24	2.46	0.26	0.19	98.84
Layer (Vo)	1000 1 1003	000 BC		100	0.60	0.42	0.37	0.63	0.02	0.54	0.83	0.28	0.20	0.12	0.87
Hekla 3	1080 ± 180^{a}	982 BC	4	120	64.78	1.00	13.92	6.82	0.19	1.18	4.02	4.21	1.95	0.26	98.32
					8.19	1.25	0.89	4.03	0.08	1.50	2.42	0.56	0.63	0.28	0.83
11-1-1-2-	1000 + 1003	002 PC		10	46.50	4 40	10.74	15.01	0.22	5.00	0.00	2.02	0.70	0.40	07.00
Hekla 3a	1080 ± 180^{-1}	982 BC	1	12	46.59	4.40	12.74	15.01	0.23	5.00	9.80	2.92	0.73	0.48	97.90
Ushis 2h	1000 + 1003	092 86	2	21	0.57	0.16	14.00	10.05	0.01	0.24	0.31	0.18	0.07	0.02	0.57
Hekia 3D	1080 ± 180^{-1}	982 BC	3	31	29.19	1.44	14.00	10.05	0.28	1.94	0.44	3.93	1.50	0.00	98.32
Uable 2a	1000 1 1003	092 86	4	25	1.32	0.19	1477	0.78	0.03	0.48	0.44	0.34	0.15	0.12	0.88
Hekia 30	$1080 \pm 180^{\circ}$	982 BC	4	30	00.42	0.48	14.77	0.10	0.20	0.51	3.55	4.51	1.96	0.13	98.62
Holdo 2d	1000 1 1002	092 86	4	40	2.13	0.15	12.40	1.11	0.03	0.19	1 70	0.23	0.20	0.07	0.74
пекіа 30	$1000 \pm 180^{\circ}$	902 DC	4	42	1 20	0.15	15.48	2.71	0.10	0.08	1./8	4.52	2.02	0.01	96.20
					1.30	0.06	0.50	0.58	0.02	0.06	0.30	0.23	0.19	0.01	0.85

^a (year BC; Dugmore et al., 1995).

Following the preliminary varve counts, the bottom of the two chronologies were age-constrained by the date of the Vö tephra, and then correlated to each other using distinct laminae, tephra and stratigraphic sequences until every varve in the series could be linked between cores. Although the addition of the HVT03-3 chronology and the position of the H1104 tephra increased the accuracy of the basin-wide varve chronology for the last ~ 1 ka, further chronologic control was necessary to reduce the level of counting error in the three records, and to increase the overall age accuracy. This was accomplished by identifying the historic tephra (H1766, K1721, H1300 and Eldgjá) that were visible in thin section and already fingerprinted geochemically. The calendar ages of each tephra were inserted into the varve series as time-marker horizons, and the three chronologies were divided into six shorter sequences, each bounded by dated marker beds. The shorter varve sequences were inspected to identify missed or incorrectly identified varves (e.g. Lamoureux and Bradley, 1996), and this process was repeated until the varve counts correlated with the marker-bed ages or until further improvements to the varve sequences were impractical. A final, multi-core varve chronology was constructed for the historical portion of the Hvítárvatn sediment by combining the shorter sequences into three individual, correlated records.

Age uncertainty for the final varve chronology was evaluated using the position of the Veiðivötn 1477 (V1477) tephra horizon. This thin, relatively fine-grained ash layer was not initially detected in thin section or incorporated in the construction of the varve chronologies, but was targeted on the basis of its stratigraphic position and geochemically identified in the sediment cores by Jagan (2010). The varve age produced by the stratigraphic sample depth of the tephra is 1483 AD, indicating a counting error of six years. This tephra is located near the middle of the longest time interval between control points in the historical portion of the cores (421 years between K1721 and H1300), when counting error is presumed to be highest. Using the position of V1477 and the length of time between the nearest control points, we assign a maximum age uncertainty for the varve chronology of 3 years for every 210 years between identified tephra layers.

The laver count of HVT03-2 was continued down core to the H3 tephra layer and produced a varve age of 982 BC for this eruption. Given the greater age uncertainties associated with varve counts made on a single core, the increased counting error associated with periods of low overall sedimentation rates (thinner, less-defined varves), and the length of time (\sim 1800 years) from the nearest control point, we assign a conservative maximum counting error of \pm 100 years for the pre-historic portion of the varve chronology. The magnitude of this uncertainty is low at the Vö tephra, and increases in a relatively consistent manner down to H3. The sample taken directly below H3 in the exposed soil section yielded a calibrated radiocarbon age of 2956 \pm 86 cal yr BP (2 σ ; Calib 6.0; Stuiver and Reimer, 1993). This date is statistically indistinguishable from previously published results (3030 \pm 180 cal yr BP, 2 σ , Dugmore et al., 1995, and 2997 \pm 41 cal yr BP, 2σ , van den Bogaard et al., 2002). To facilitate the evaluation of all proxies using the varve chronology, we assign the varve age of 982 BC (2932 yr BP) to H3, acknowledging that the radiocarbon-inferred ages are slightly older.

Age-depth models for the three sediment cores, constructed using the varve counts and the positions of the constraining tephra layers, are in close agreement, with relatively constant differences in sedimentation rate between cores (Fig. 3). The age model for HVT03-1 was continued down to H3 using four distinctive prehistoric tephra layers common to this core and HVT03-2 as tie points. All physical and biological proxies are evaluated using the age model inherent to the core from which they were sampled. The few core sections (a total of 12 sections from all three cores) that contained short segments (<8 varves) of highly deformed laminae were omitted from the individual varve chronologies. Two notable exceptions occur in core HVT03-01 from 1912 AD to 1947 AD and in HVT03-2 from 1968 AD to 1975 AD. These gaps were filled by transposing the average annual sedimentation rate from the other two cores, accounting for differences in sedimentation rate between core sites. One ~8 cm section in HVT03-3 was highly deformed and precluded the ability to measure varve thickness for the years 1805 AD to 1817 AD (Fig. 3).



Fig. 3. Age models for the three sediment cores constructed from annual layer counts and tephra layers (shaded circles). The six historical tephra horizons that were geochemically identified (Table 1) and used to improve the accuracy of the varve chronology are labeled. Note that the age model for core HVT03-3 ends at H1104. The age model for core HVT03-1 was continued from the Settlement tephra to Hekla 3 using the age model from HVT03-2 and the depths to four unidentified tephra layers common to both cores.

4.3. Physical proxies

Individual core varve thickness records show similar patterns (Fig. 4 and Fig. S1). Low sedimentation rates persist into the second millennium AD, and then increase significantly in two steps, beginning at ca 1250 AD and again at ca 1710 AD. Despite a roughly 150-year decline at the start of the record, varve thickness remains relatively constant from 982 BC to 500 AD with an average sedimentation rate in core HVT03-2 of 0.25 cm a^{-1} . The brief period of minimum varve thickness centered around 550 AD is followed by increasing and variable sedimentation rates after 600 AD that lead up to the Vö tephra at 871 AD. Immediately following this tephra layer, there is an abrupt increase in varve thickness before a more gradual return to pre-Settlement levels following a pronounced peak at 900 AD (Fig. 4).

Consistently thin varyes characterize the period from 980 to 1180 AD when average thickness in HVT03-1 and HVT03-2 is 0.26 cm and 0.23 cm respectively. Beginning at ca 1250 AD, varve thickness steadily increases and attains values greater than twice the average of the previous 2000 years early in the fourteenth century. The local maximum between 1400 and 1600 AD is represented in each core, with a common peak in varve thickness at ca 1500 AD. Following this interval, sedimentation rates decline in a stepped manner toward relative minima in the late seventeenth century. Sediment deposition increases rapidly after ca 1700 AD, with three defined peaks at approximately 1845, 1900, and 1935 AD. The time of thickest varves for the last 3 ka occurs in all three cores during the 1935 AD peak, when average sedimentation in HVT03 core sites 1, 2 and 3 is 2.50, 1.13, and 1.25 cm a^{-1} respectively. Subsequent to this peak, varve thickness decreases to pre-1800 AD levels by 1980 AD with a short interruption around 1966 AD. The last 20 years of the record show a trend of increasing varve thickness.

Significant IRD concentrations are confined to the period of maximum sedimentation rate (Fig. 4 and Fig. S1). Cores HVT03-1 and HVT03-2 contain only a limited number of clasts deposited every few centuries for the first 2.7 ka of the record. The time series in Core 3 begins at H1104 and shows significantly more IRD in both size fractions ($\geq 2 \text{ mm}$ and $\geq 5 \text{ mm}$) during all time periods. This is most notable beginning after 1550 AD with HVT03-3 showing consistent levels of about 7 clasts/20 years. By 1700 AD small numbers of IRD are also present in Cores 1 and 2. There is a striking increase in the flux of IRD to the lake beginning in 1760 AD, which is bracketed by the K1721 and H1766 tephra layers. Two peaks around 1840 and 1890, separated by a relative minimum around 1860 AD, define the period of maximum IRD deposition. Shortly after 1940 AD, there is a sudden decrease in IRD that is coincident with the documented retreat of Suðurjökull from the lake. Each core contains low numbers of clasts for the last 60 years.

Sediment bulk density is often used to evaluate changes in sediment character related to changes in mineral composition, sediment compaction, and the relative abundance of sediment organic matter. The sediment fill deposited in Hvítárvatn during the late Holocene has an average density of 2.0 ± 0.1 g cm⁻³ (n = 1934; Fig. 4). Notable deviations of low density occur following the H3 tephra layer and around 600 AD, while peak values occur in the tenth and fifteenth centuries AD. The sediment accumulation in the last 500 years has a slightly lower bulk density with more periodic fluctuations, including minima at 1920 AD and at the core top. There is no long-term trend in sediment bulk density, nor is there a clear relationship between density and varve thickness. The bulk density data confirms our assertion that varve thickness measurements faithfully represent changes in the flux of clastic erosion products to the lake.



Fig. 4. Combined records of varve thickness, IRD (≥ 2 mm and ≥ 5 mm size fractions shown as the average from all three core sites), and bulk density for the last 3ka. The bold lines in the varve thickness panel are 20-year running averages superimposed on the annual data shown behind in gray. IRD was counted in 20-year intervals and the gray bars in both IRD panels represent the number of clasts per 20 years, while the bold line reflects a cubic spline smooth. Density data for cores HVT03-1 and HVT03-2 is at 5–10 year resolution. The bold line in the density panel represents an approximately 20-year smooth on the combined data from both cores. The position of all tephra layers used in the creation of the varve chronology are shown: a) H1766, b) H1721, c) V1477, d) H1300, e) H1104, f) Eldgjá, g) Vö, h) H3. Individual core varve thickness and IRD data for the historical period is presented in Supplementary information (S1 a–C).

4.4. Biological proxies

Sedimentary organic matter is characterized by TOC_{*} values between 0.1 and 0.6% (Fig. 5). The majority of soil samples contain higher concentrations of organic material, with TOC_{*} ranging from 0.5 to 5.5%. The strong linear correlation between TOC and TN in the soil samples is defined by a C:N of ~ 14 (Fig. 5). This ratio reflects the purely terrestrial OM contained in the soils surrounding Hvítárvatn, and represents the upper boundary for C:N of the lake sediment. Currently, an average purely aquatic C:N value cannot be defined, but few lake sediment samples are <5, which we assume is close to the purely aquatic carbon value (e.g. Meyers, 1994). The C:N for >95% of the lake sediment samples lie between 5 and 14.

Both cores show a trend of increasing C:N and TOC₀ over the 3 ka record (Fig. 6), indicating that the delivery of terrestrial OM to Hvítárvatn has increased through the late Holocene. Notable positive excursions occur at ca 1000 BC, 600 AD, and after 1300 AD (Fig. 6). $TOC_{\%}$ broadly tracks TOC_{O} until the thirteenth century, when increasing varve thickness dilutes the signal. The cores contain an average TOC flux of roughly 2 mg cm⁻² a⁻¹, excluding the disturbance associated with the H3 tephra layer. Because of this large disturbance at the start of the varve chronology, the age model was continued past H3 by approximately 200 years using an average sedimentation rate appropriate to the respective cores. This extension provides the opportunity of evaluating the response of sedimentary OM to the deposition of the H3 tephra (Fig. 6). TOC_Q and C:N are fairly constant leading up to H3, with TOC_Q between 1 and $2 \text{ mg cm}^{-2} \text{ a}^{-1}$, and C:N around 6. Immediately following the tephra layer however, TOC₀ and C:N increase sharply for approximately 50 years, before decreasing more-gradually toward pre-eruptive levels. At the height of the perturbation, TOC_Q is more than an order of magnitude higher than pre-eruption levels, $TOC_{\%}$ is >2.0% and C:N approaches 12. In both cores, C:N peaks slightly after TOC and is slower to recover, maintaining elevated values for >200 years.

After recovering from the H3 disruption, the amount and source of sedimentary OM is stable until the sixth century AD (Fig. 6). At ca 550 AD there is a large shift in C:N toward higher values that is followed by an increase in TOC₀ and TOC_%. All three proxies attain a distinct peak at around 600 AD. Following this peak, TOC₀ and C:N decrease swiftly. However, while TOC₀ falls back to minimum values, C:N drops only slightly, and remains elevated for the duration of the record. The brief local minima seen in both proxies during the tenth century is followed by steadily increasing TOC_0 and variable, yet increasing C:N through the nineteenth century. Maximum C:N occurs at ca 1480 AD in HVT03-2 and ca 1910 AD in HVT03-1, and reflects the greatest contributions of terrestrial OM to overall TOC at each site. The total flux of TOC increases considerably after ca 1700 AD and throughout the eighteenth century, reaching consistently high values of nearly 10 mg cm⁻² a⁻¹ during the nineteenth century. The last 60 years of the record shows decreasing TOC_0 and C:N, as both proxies are approaching pre-thirteenth century values.

5. Interpretation

5.1. Varve thickness and IRD

Sedimentation rates in Hvítárvatn are determined by the processes responsible for the production and delivery of sediment



Fig. 5. Total organic carbon (TOC) and total nitrogen (TN) from cores HVT03-1 and HVT03-2 along with soil samples removed from soil pits dug around the lake. The 19 soil samples, taken from various depths and spanning the last 6 ka, show a strong linear correlation between TOC and TN, and are characterized by a C:N of roughly 14. This ratio represents the terrestrial end member for sedimentary OM preserved in the lake sediments. The aquatic end member was inferred from the distribution of samples and given a C:N ratio of 5. Greater than 95% of the lake sediment samples have C:N that fall between these two values (see inset), illustrating the two end member mixing model that is used to evaluate source contribution of OM to the lake sediments. The dashed box marks the area of the plot that is covered in the inset.

to the lake. Previous studies have demonstrated that in proglacial environments, lacustrine sedimentation rates are largely controlled by the activity of upstream glacier-fluvial systems (Leonard, 1986, 1997; Desloges, 1994; Loso et al., 2006; Hodder et al., 2007). Although Langjökull presently occupies ~35% of the Hvítárvatn watershed, a comparison of lake discharge to simulated glacierderived recharge (for the 1961–1990 AD reference period), suggests that the ice cap dominates inflow to the lake, contributing 72% of the total (Flowers et al., 2007). From both the empirical observations and modeling studies we conclude that the sediment flux to Hvítárvatn is dominantly controlled by the integrated rate of sediment production by erosion beneath Langjökull, modulated on annual to decadal timescales by the efficiency of the subglacial fluvial sediment delivery system. Strong similarities in the varve thickness records of the three cores suggests that relative sedimentation rates change uniformly throughout the basin, and that such changes represent catchment-wide variations in sediment yield. The relatively constant density profile indicates that the observed increases in varve thickness during the last millennium are unrelated to changes in sediment compaction (Fig. 4).

The existence of clastic varves in the sediment fill confirms that Langjökull has been continuously present in the lake catchment for the last 3 ka, and that it has been contributing to suspended sediment deposition through erosion and subglacial discharge. We interpret changes in sedimentation rate to reflect multi-decadal changes in Langjökull's aerial extent. Increased (decreased) erosion rates due to the steady growth (decay) of Langjökull over these timescales will result in on average thicker (thinner) varves. This relationship is evident in the first-order trend of increasing sedimentation rates in Hvítárvatn through the late Holocene, and is consistent with the pattern of Neoglacial cooling in Iceland (Geirsdóttir et al., 2009a). Similarly, variance in varve thickness is a measure of the magnitude of variability in sediment deposition, which we interpret to be a function of changes in the total amount of erosion products available for transport to the lake. Because the size of Langjökull is the dominant control on lake sedimentation rates, high (low) variance in varve thickness can only be achieved when there is an abundance (shortage) of glacially derived erosional products, which in turn is a function of Langjökull's aerial extent.

On shorter timescales, the direct relationship between varve thickness and glacier activity becomes more complex. Absolute peaks in sedimentation rates, including the one at ca 1935 AD, may be associated with initial periods of rapid ice retreat following a highstand rather than the maximum glacier stand itself (e.g. Leonard, 1997). In addition, high frequency variations in varve thickness exist and are superimposed on the longer-term trends. These variations have been explored by Ólafsdóttir K. B (2010), and likely reflect changes in the efficiency of the subglacial fluvial system to deliver erosional products to the lake, in response to inter-annual and decadal changes in local climate, such as variations in the North Atlantic Oscillation (NAO). However, because high frequency signals must be transmitted to the lake through Langjökull, it is the overall size of the ice cap that determines multidecadal patterns in varve thickness. In this study we restrict our discussion to the longer-term trends, which record the broader evolution of climate and ice cap activity.



Fig. 6. Total organic carbon bulk concentration (TOC₈), the mass flux of total organic carbon (TOC₀), and C:N ratios from Hvítárvatn cores HVT03-1 and HVT03-2 over the last 3.2 ka. The age model for both cores was extended ca 200 years below Hekla 3 using an average sedimentation rate appropriate to each core to evaluate pre-eruption conditions. Thick lines show the 10-point moving average through the data and represent a smoothing of 20–40 years.

The principle mechanism for the transport and deposition of IRD to the core locations is through iceberg rafting, although it is possible that strong winds or spring flooding could carry small amounts of sediment over a frozen lake surface. Therefore, the presence of IRD in the sediment fill is generally restricted to intervals when either of the two outlet glaciers advanced into Hvítárvatn and maintained an active calving front. During such periods, sediment-laden icebergs produced at the ice margin were transported around the lake by wind, releasing any clasts they contained as they melted. The shallow water depths in the southern portion of Hvítárvatn, and particularly its outlet (<5 m deep and <150 m wide), would have limited the evacuation of icebergs and confined the majority of icebergs to the northern basin (Fig. 1a; Geirsdóttir et al., 2008). Although the release of IRD is highly stochastic, the replication of the IRD signal at all three core sites, and in both size fractions, suggests that the flux of IRD accurately reflects the volume (number) of icebergs in the lake at any given time (Fig. S1). The pattern in average IRD flux can thus be used to infer changes in iceberg production (Fig. 4).

Consistent IRD delivery to all cores began shortly after 1760 AD, but remained relatively low until 1820 AD. This time window represents the advance of one or both outlet glaciers into Hvítárvatn in response to continued summer cooling to sustained levels not experienced in at least the previous 2 ka. The period of maximum IRD lasted from 1820 to 1940 AD, and reflects the continuous presence of a calving margin. In all three cores IRD peaks around 1840 and 1890 AD, separated by a distinct minimum at approximately 1860 AD and followed by a decline to lower values after 1900 AD (Fig. 4). The IRD abundance structure has two possible explanations. The period of low IRD around 1860 may reflect a period of intense cold summers during which the lake was filled with icebergs that failed to melt completely each summer, thereby stabilizing outlet glacier margins (Geirsdóttir et al., 2008). Alternatively, the IRD minimum may reflect a period of relative warmth within the nineteenth century when the outlet glaciers had smaller calving margins. Both IRD peaks are contemporaneous with varve thickness maxima. The decline in IRD after 1890 AD marks the retreat of the outlet glaciers from Hvítárvatn. Although Norðurjökull maintained a calving front until ca 2009 AD, aerial photographs show that Suðurjökull retreated from the lakeshore before 1945 AD (Landmaelingar Íslands: the National Land Survey of Iceland). The small flux of IRD during the last 60 years is thus credited solely to Norðurjökull.

Prior to the onset of IRD delivery in the eighteenth century, small numbers of IRD are present in individual cores every few centuries, with limited inter-core commonality. We interpret this background IRD signal to reflect occasional delivery of IRD by winter wind or



Fig. 7. Summary of Hvítárvatn proxy data for the past 3 ka including (from top to bottom): varve thickness variance, varve thickness, IRD, and the Organic Matter proxies C:N and TOC_Q . Prominent climate stages are shown along with the three conventional periods of climate change noted in *stage 3*. DACP = Dark Ages Cold Period, MWP = Medieval Warm Period, LIA = Little Ice Age. The number of sediment cores represented for each proxy is shown in gray in the upper left corner of respective panels. Varve thickness variance was calculated from 20-year running bins. Annual varve thickness data (shown in gray) is overlain by a 20-year smoothing (bold line). IRD was counted in 20-year intervals and gray bars represent the average number of clasts per 20 years, while the bold lines reflect a cubic spline smooth. The bold lines in the lower panel represent an approximately 20–30-year average of C:N and TOC_Q data.

meltwater flowing over a frozen lake surface, rather than indicative of calving glaciers in the lake. The relatively reproducible signal of low IRD flux in the decades leading up to 1760 AD may also be explained by icefall on the Norðurjökull terminus (i.e. the tumbling of seracs down the steep proglacial slope into the lake) or the advance of the terminus just to the lakeshore.

5.2. Sedimentary OM

The relative abundance of sedimentary organic carbon and nitrogen is a function of OM source and production from vegetation growing in and around the lake. TOC concentrations (TOC_%) are often used as indicators of within-lake primary productivity and thus past environmental conditions. However, additional factors, including sedimentation rate, preservation potential, and the flux of terrestrial OM, can influence TOC_% (Meyers and Teranes, 2001). In Hvítárvatn, when high sedimentation rates are sufficient to minimize remineralization and primary productivity is low (due to the long seasonal ice cover, shallow photic zone, and low summer temperatures), TOC_% is strongly influenced by the flux of terrestrial OM and sedimentation rate. The influence of allochthonous OM on lake sediment TOC_% is illustrated by the strong positive correlation between TOC $_{\%}$ and C:N in HVT03-1 and HVT03-2 for the first 600 years of the record ($r^2 = 0.71$). This relationship begins to break down after the fifth century AD, and the two proxies are only marginally correlated for the entirety of the record ($r^2 = 0.20$; Fig. 6). The decoupling of TOC_% and C:N reflects decreasing aquatic primary productivity, as the lake becomes increasingly turbid and cold, and greater dilution by minerogenic material, both due to increasing glacial activity. Elevated C:N implies increased soil erosion (low temperatures and wind) across the catchment, whereas the low TOC_% suggests increased delivery of glacially eroded minerogenic material that is largely organic-free. The diluting effect of high sedimentation rates on OM concentrations is most evident through the last millennium when TOC_% no longer follows TOC₀ (Fig. 6). During this period, and particularly after 1250 AD, the flux of OM to the lake is higher than the preceding 2 millennia, despite low TOC_%. Because of the substantial impact of varying sedimentation rates on TOC_% at Hvítárvatn, we emphasize changes in TOC_0 rather than $TOC_{\%}$ to quantitatively assess the evolution of sedimentary OM during the late Holocene.

The two large positive excursions in TOC_Q that occur immediately following H3 and in the early sixth century are indicative of abrupt environmental disturbances (Fig. 6). Both peaks are associated with a decrease in sediment bulk density and an increase in C:N, representing episodes of intense soil erosion within the catchment. Although the implications of these events are explored in greater detail below, we refer to them here to illustrate the sensitivity of the proxies to environmental perturbations. Variations in TOC_0 are suggestive of changes in the delivery of erosional products (soils or clastic material) to the lake, and hence a function of landscape stability and glacier activity. Similarly, C:N tracks the relative contribution of OM from terrestrial sources and, like TOC₀, is independent of sedimentation rate. The primary mechanisms for the mobilization of terrestrial OM in the lake catchment are wind erosion of soils and glacial advance over soil-covered terrain. Both mechanisms increase the delivery of soil OM during colder summers.

Differences in the timing of maximum C:N (Fig. 6) between cores HVT03-1 (ca 1910 AD) and HVT03-2 (ca 1480 AD and 1910 AD) may be related to the locations of the two core sites within the basin. The northern flats core site (HVT03-1) is likely dominated by sediment delivered across the meltwater fed Hvítárnes delta, whereas the ridge-crest core (HVT03-2) is dominated by suspension settling an may be more representative of a lake-wide average (Fig. 1). While the steady increase of terrestrially derived OM in HVT03-1 during the past millennium suggests a general growth of Langjökull and/or the progradation of the delta, the two peaks seen in HVT03-2 are coincident with peaks in varve thickness and support multiple glacial advances.

6. Discussion

6.1. Climate and ice cap evolution

The combined, multiproxy record of late Holocene glacial activity and environmental conditions at Hvítárvatn can be subdivided into three stages discussed below (Fig. 7).

6.1.1. Stage 1 (H3 to ca 600 BC)

Proxy data reflect mild summer temperatures and limited glacial activity in the lake catchment following the abrupt and prolonged perturbation associated with the H3 tephra layer (Fig. 7). Hekla 3 (varve age: 982 BC) produced approximately 12 km³ of tephra and is regarded as the most severe eruption of Hekla during the Holocene (Larsen and Thorarinsson, 1977). Although the timing of this event has been associated with widespread climatic,

environmental and archeological events (Eiríksson et al., 2000a; Grattan and Gilbertson, 2000 and references therein; Knudsen and Eiríksson, 2002), there are no continuous terrestrial records documenting its environmental impact in Iceland.

The most probable mechanism responsible for the sudden increase in TOC_% and C:N following H3 (Fig. 8) is vegetation damage due to abrasion and increased soil erosion within the catchment. The H3 tephra horizon in the area is characterized as a coarse ash to fine lapili, ranging in thickness from 7 cm in the lake cores to over 30 cm in nearby soil pits (e.g. Larsen and Thorarinsson, 1977). Subsequent to the eruption, the tephra layer likely killed a large portion of vascular plants in the watershed through root system suffocation, impaired photosynthesis, abrasion and acidification (e.g. Mack, 1981; Grattan and Gilbertson, 2000). With the vegetation cover depleted, underlying soils would have become vulnerable to deflation and alluviation, resulting in pronounced landscape destabilization. This is supported by increased varve thickness, and suggests that catchment instability was contributing to lake sedimentation. The protracted C:N recession limb, relative to TOC_%, indicates that although the total amount of terrestrial OM delivered to the lake was relatively small, it still comprised the majority of OM flux (Fig. 8). Persistent elevated C:N values also suggest suppressed primary production related to less-favorable growing conditions. Both scenarios point toward a cooler and windier environment lasting for >100 years after H3 that is temporally consistent with reduced North Atlantic Deep Water formation (Oppo et al., 2003), increased IRD (Bond et al., 2001), shifting ocean currents (Knudsen and Eiríksson, 2002) and low SST (Eiríksson et al., 2000b; Jiang et al., 2002; Ran et al., 2008). On land, evidence for cooler conditions includes declining birch forests and increased soil erosion (Hallsdóttir, 1995). Although the timing of this cooling episode in relation to H3 is debated (e.g. Knudsen and Eiríksson, 2002), proxy evidence from this study suggests that prior to the eruption, conditions in the highlands were stable, and it was only after the tephra deposition that regional environmental changes occurred.



Fig. 8. C:N and TOC_x from cores HVT03-1 and HVT03-2 through the interval spanning the Hekla 3 tephra layer (dashed vertical line). During the ca 200 years prior to the eruption, TOC and C:N are low and steady, reflecting stable environmental conditions with limited delivery of terrestrial OM to Hvítárvatn. Immediately following the tephra deposition both proxies increase dramatically and remain elevated for >100 years. The more gradual decline in C:N relative to TOC indicates that even as the amount of OM washing into the lake decreases rapidly following peak soil erosion, the landscape remains unstable for many decades and is contributing a large proportion of OM to the lake sediments relative to aquatic algae.

After recovering from H3, varve thickness maintains a 250-year (850 BC to 600 BC) average of 0.24 cm a^{-1} and contains low variance (Fig. 7). Combined with the return of TOC_Q and C:N to minimal, pre-eruptive levels, these data reflect stable conditions around Hvítárvatn; Langjökull dominated the sediment flux, but exhibited little variability.

6.1.2. Stage 2 (ca 600 BC to ca 550 AD)

The second stage lasts for nearly 1200 years, beginning ca 600 BC (~2550 years BP), when all proxies have fully recovered from the H3 disturbance, and terminating in the sixth century AD. Broadly contemporaneous with a period of widespread glacier advances in Iceland (e.g. Stötter et al., 1999; Kirkbride and Dugmore, 2006), the onset of this stage is marked by a small increase in varve thickness and varve thickness variance, reflecting increased glacial activity. However, the lack of detectable long-term trends in sedimentation rate suggests Langjökull's aerial extent remained relatively constant. In addition, steady and generally low TOC_Q levels, combined with low C:N ratios, indicate that Neoglacial conditions in the region were relatively stable, and that sedimentary OM was dominated by within-lake primary productivity (Fig. 7).

6.1.3. Stage 3 (ca 550 AD to present)

Subsequent to H3, the most notable features of the Hvítárvatn record are the pronounced intervals of abrupt change in all proxies and irregular high variability in varve thickness, beginning in the sixth and increasing after the thirteenth centuries AD (Fig. 7). All proxy data reflect a shift toward increased glacial erosion and landscape destabilization from ca 550 AD to ca 900 AD and from ca 1250 AD to ca 1950 AD, separated by an interval of relatively mild conditions. The timing of these intervals coincides with the well-documented periods of climate change commonly known as the Dark Ages Cold Period, the Medieval Warm Period, and the Little Ice Age, and is associated with changes in solar activity (Stuiver et al., 1998) and/or explosive volcanism (Gao et al., 2008).

6.1.3.1. The dark ages cold period (DACP). An abrupt cooling event at ca 550 AD is expressed as spikes in TOC_Q and C:N, accompanied by a decrease in bulk density and subsequent broad increase in varve thickness (Figs. 4 and 7). Both OM proxies increase sharply for approximately 50 years before reaching relative maxima at ca 600 AD, as annual laminae begin a period of increased thickness and greater inter-annual variability (Fig. 7). Stratigraphically, the depth to these OM peaks is directly below and above the first in a series of three prominent basaltic tephra horizons. TOC_Q and C:N increase prior to the tephra deposition, and reflect an abrupt and severe cooling event that stimulated landscape instability and increased soil erosion in the lake catchment. Although the main flux of terrestrial OM occurs between 550 and 650 AD, the persistence of high C:N until present suggests that the abrupt perturbation of the catchment may reflect a threshold change that promoted subsequent instabilities.

A sudden shift toward cooler conditions is consistent with an increase in soil erosion in western Iceland (Fig. 9; Geirsdóttir et al., 2009b) and historical accounts of crop failures, famines, and dry cold fogs in Europe and Asia in the 530's AD (e.g. Stothers and Rampino, 1983; Rampino et al., 1988). Evidence for a prolonged temperature decline includes the advection of cold surface waters on the north Iceland shelf (Fig. 9; Jiang et al., 2002), and the disappearance of birch forests in northwest Iceland (Andrews et al., 2001). These records, combined with the multiple proxies at Hvítárvatn, reflect both rapid-onset and persistent cooling, and suggest that volcanism, combined with weaker solar activity, may have triggered the hemispheric cooling in the sixth to ninth centuries AD.

The prominent peak in varve thickness that begins directly after the Vö tephra at ca 871 AD, and that reaches a maximum at ca 900 AD (Fig. 7), is not attributed entirely to glacial activity or landscape instability. There is no evidence for increased soil erosion, and although the timing coincides with a short cooling event seen in multiple Northern Hemisphere proxy records (Jiang et al., 2005 and references therein; Geirsdóttir et al., 2009b), we are hesitant to ascribe all of the change in varve thickness to glacial erosion. A visual inspection of the cores indicates that this change is most likely due to secondary tephra deposition from the surrounding watershed. The Vö tephra is the thickest tephra horizon in the Hvítárvatn cores after H3. Consequently, it is possible that the tephra fallout on the lake catchment, including over Langjökull, directly contributed to the increase in sediment accumulation for multiple decades.

6.1.3.2. The medieval warm period (MWP). The MWP is subtly expressed as a period of minimal glacial activity and reduced soil erosion in the catchment. Varve thickness decreases after 950 AD and remains consistently low through Medieval time with slightly thinner annual laminations (0.23 cm a⁻¹) than for any other multicentennial period in the past 3 ka (Fig. 7). Both TOC_Q and C:N decrease through the tenth century AD. The transition out of the MWP occurs as a step-wise increase in varve thickness, beginning with an initial discrete pulse and subsequent decline at ca 1200 AD. Although this transition is not smooth, we place the termination of the MWP at ca 1250 AD, when sedimentation rate exceeds the prior 2200 year average and varve thickness never returns to the levels of previous centuries.

The period of Norse exploration and settlement in the tenth century, including the colonization of South Greenland, suggests that mild conditions persisted in the North Atlantic during this time. Recent proxy-based temperature reconstructions demonstrate that the Northern Hemisphere experienced peak medieval warmth between ca 900 and 1100 AD (Moberg et al., 2005; Mann et al., 2008). Although the combination of such records defines a broadly synchronous period of warming in the North Atlantic, not all regions experienced increased temperatures (e.g. Seidenkrantz et al., 2008), and individual localities in the North Atlantic seem to have been influenced by local atmospheric and oceanographic conditions (Eiríksson et al., 2004). At Hvítárvatn, the combined proxy data suggest a period of relative warmth from the tenth to late twelfth centuries AD, compared to the DACP and subsequent LIA cooling, but similar to conditions early in the first millennium AD.

6.1.3.3. The little ice age (LIA). The onset of the LIA is marked by the multi-centennial increase in varve thickness, varve thickness variance, and C:N at ca 1250 AD, reaching a broad maximum between 1450 and 1550 AD (Fig. 7). This represents the first of two major advances of Langjökull during the past millennium. After 1500 AD, although sedimentation rates remain relatively high, a rapid drop in C:N and a decrease in varve thickness suggest a glacier standstill or retreat. The start of the second major advance of Langjökull began shortly after 1700 AD and coincides with the prominent increase in TOC_0 . This advance is not accompanied by an increase in C:N, suggesting that the glacier advanced over previously glaciated terrain containing little soil or vegetation. All proxy evidence indicate that Langjökull continued to grow through the eighteenth century, with outlet glaciers reaching the lakeshore by ca 1760 AD. The maximum ice stand was achieved in the nineteenth century with inferred peaks occurring around 1840 and 1890 AD (Fig. 7). The absolute peak in varve thickness in the 1930's coincides with rapid declines in IRD and C:N, and occurred only 15 years before aerial photography documents the retreat of Suðurjökull from



Fig. 9. Comparison of Hvítárvatn varve thickness and C:N (lower curves) with other northern North Atlantic climate records and 60°N summer insolation. From the top: June 60°N insolation from Berger and Loutre (1991), North Iceland shelf diatom-inferred SST from Jiang et al. (2002, 2005), periods of inferred Hrútfell glacier advances (Kirkbride and Dugmore, 2006), Lake Haukadalsvatn TOC (Geirsdóttir et al., 2009a), paleomagnetic secular variation (PSV) determined average sediment accumulation rates from Lakes Haukadalsvatn (black) and Hvítárvatn (gray; from Ólafsdóttir S., 2010). Approximate age resolution of Hvítárvatn varve thickness, Hvítárvatn C:N, Haukadalsvatn TOC, and diatom-inferred SST is 20 years. PSV determined sedimentation rates have an age resolution of 50 years.

Hvítárvatn. This peak also occurred during a period of pronounced warming recorded at the Stykkisholmur weather station in western Iceland. Thus the period of maximum sedimentation rates in Hvítárvatn reflects the retreat of Langjökull in response to early twentieth century warmth. Both varve thickness and C:N increase briefly in the 1960's and again after 1980 suggesting small readvances within the overall retreat or the tapping of stored erosion products by meltwater streams.

The LIA was the most severe multi-centennial cold interval of the late Holocene and numerous Northern Hemisphere temperature reconstructions illustrate a pronounced cold climate between approximately 1250 and 1850 AD, following the relative warmth of medieval times (Mann and Jones, 2003; Moberg et al., 2005; Mann et al., 2008). This cooling has been attributed to decreased solar irradiance and high atmospheric sulfate loading associated with explosive volcanism (Stuiver et al., 1997; Overpeck et al., 1997; Mann et al., 1998; Bond et al., 2001; Ammann et al., 2007; Gao et al., 2008). The North Iceland shelf is thought to have experienced exceptionally cold but variable LIA conditions due to the advection of cold-water masses and the delivery of sea ice (Eiríksson et al., 2000a; Massé et al., 2008; Sicre et al., 2008). In Iceland, LIA cooling is manifested in widespread glacier advances, with the majority of Late Holocene maxima occurring in the midto-late nineteenth century (Geirsdóttir et al., 2009a and references therein).

Based on multiple proxy data, we place the start of the LIA in the highlands in the mid-thirteenth century, when Langjökull began a series of two long periods of expansion, with high stands at ca 1500 AD and in the nineteenth century. This onset date correlates well with a shift toward cooler SST off the North Iceland shelf and landscape destabilization in western Iceland (Fig. 9; Jiang et al., 2002; Geirsdóttir et al., 2009b). The prominent two-step event seen in Langjökull's advances may reflect a "bimodal" character of the LIA in Iceland that is composed of two cold periods separated by less severe summer cold during the sixteenth and seventeenth centuries. Langjökull's late Holocene glacial maximum occurs during the second LIA advance and significantly exceeded any previous advance for at least the last 3 ka. The outlet glaciers Norðurjökull and Suðurjökull entered the lake around 1760 AD and reached their respective maximum extensions of 1.5 km and 3 km into the basin (Fig. 1a) between 1820 and 1900 AD, with two distinct pulses at ca 1840 and 1890 AD. Using an ice cap response time of roughly 100 years, we infer minimum LIA temperatures in the highlands occurred during the mid eighteenth to early nine-teenth centuries.

6.2. Late holocene climate in iceland

The pattern of Neoglacial climate reconstructed here stands in broad agreement with terrestrial and marine climate records from Iceland and the North Atlantic. The Hvítárvatn sediment archive reflects a 3 ka trend toward cooler summers and increasing ice coverage in the highlands that broadly follows the summer insolation forcing at 60°N (Berger and Loutre, 1991; Fig. 9). This trend is complemented by chironomid-derived summer temperature reconstructions from three coastal lakes in northern Iceland that reflect colder summers after 3 ka (Axford et al., 2007). Multiple climate proxies from lake Haukadalsvatn, including sediment accumulation rate and TOC, demonstrate that cold summers and windy conditions became more prevalent through the late Holocene (Fig. 9; Geirsdóttir et al., 2009b; Ólafsdóttir S., 2010). In the marine realm, a decline in warm water diatom and foraminifera species is detected after 3 ka in sediment cores extracted from the north Iceland shelf (Eiríksson et al., 2000b; Jiang et al., 2002, 2007), and alkenone-derived SST's indicate a \sim 5 °C decline in water temperature (Sicre et al., 2008). Evidence for glacier fluctuations during the late Holocene is generally limited to periods of maximum extension (e.g. Nesje, 2009; Geirsdóttir et al., 2009a), and rarely predates the LIA. However, one study of small mountain glaciers proximal to Hvítárvatn suggests three individual glacier advances during the last 3 ka (Fig. 9; Kirkbride and Dugmore, 2006).

7. Conclusions

The sediment fill in glacial lake Hvítárvatn is dominated by the erosion products of the adjacent Langjökull ice cap and changes in the stability of soils in the unglaciated portions of its catchment. Grain size analysis and tephrostratigraphy demonstrate the continuous presence of clastic varves, and variations in varve thickness record changes in the annual production and delivery of sediment from Langjökull to the lake and serve as a proxy for ice cap size. Varve thickness is used in conjunction with additional proxies including: IRD, varve thickness variance, sediment bulk density, TOC (mass flux and bulk concentration), and C:N to evaluate glacier activity and environmental change.

Large increases in C:N and TOC immediately following the H3 tephra layer suggest that H3 tephra induced significant soil erosion that lasted for over 100 years. Subsequent to the recovery from this disturbance, all proxies indicate relatively stable conditions from ca 850 BC to ca 550 AD. Landscape instability and ice expansion was initiated abruptly at ca 550 AD, with cold and variable conditions lasting until ca 900 AD, corresponding to the DACP. An interval of variable yet relatively mild conditions existed from ca 950 AD to 1250 AD. During this time, sedimentation rate decreased to the lowest levels on record, reflecting subdued ice cap activity prior to the LIA. Inferred MWP warmth is strongest from 950 AD to 1150 AD, when proxy data reflect particularly low glacier activity and soil erosion in the catchment. The century between 1150 AD and 1250

AD is characterized by more variable conditions, including a brief pulse of glacial activity centered around 1200 AD, and subsequent return to the earlier warm state.

The onset of LIA conditions is marked by an unprecedented increase in varve thickness beginning in the mid-thirteenth century, accompanied by an increasing flux of TOC and a sharp rise in C:N. Maximum ice cap extent occurred between 1760 and 1940 AD, with distinct high stands at ca 1840 and 1890 AD. There is no evidence for outlet glaciers terminating in Hvítárvatn prior to the LIA for at least the last 3 ka, indicating that this advance likely represents Langjökull's Holocene glacier maximum. Shortly after 1900 AD, varve thickness decreases, indicating Langjökull began a steady phase of recession signifying the end of the LIA and rising twentieth century temperatures.

The Hvítárvatn record compares favorably to paleoenvironmental records from the North Atlantic and helps to define the late Holocene climate in Iceland. Strong increases in glacier activity and landscape instability seen at ca 3 ka, and beginning at ca 550 and 1250 AD, suggest colder summers and prolonged landscape destabilization associated with the changes in solar insolation and volcanism. In particular, the step-wise transition into the LIA, including the multi-centennial scale shift at ca 1250 AD, along with the bimodal nature and timing of peak cooling describe the nature of the LIA with unprecedented age resolution.

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Appendix. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.quascirev.2011.05.02.

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