Lacustrine evidence of Holocene environmental change from three Faroese lakes: a multiproxy XRF and stable isotope study

Jesper Olsen a,*, Svante Björck b, Melanie J. Leng c, Esther Ruth Gudmundsdóttir d, Bent V. Odgaard a, Christina M. Lutz e, Chris P. Kendrick c, Thorbjørn J. Andersen f, Marit-Solveig Seidenkrantz a

a Centre for Past Climate Studies, Department of Earth Sciences, Aarhus University, DK-8000 Aarhus, Denmark
b Department of Geology, Lund University, Sölveg. 12, SE-223 62 Lund, Sweden
c NERC Isotope Geosciences Laboratory, British Geological Survey, Nottingham, UK
d Earth Science Institute, University of Iceland, Iceland
e Department of Physics and Astronomy, Aarhus University, DK-8000 Aarhus, Denmark
f Department of Geography and Geology, University of Copenhagen, DK-1310 Copenhagen, Denmark

A R T I C L E   I N F O

Article history:
Received 10 February 2010
Received in revised form 8 June 2010
Accepted 16 June 2010

Abstract

The vegetation history of the Faroe Islands has been investigated in numerous studies all broadly showing that the early-Holocene vegetation of the islands largely consisted of fellfield with gravely and rocky soils formed under a continental climate which shifted to an oceanic climate around 10,000 cal yr BP when grasses, sedges and finally shrubs began to dominate the islands. Here we present data from three lake sediment cores and show a much more detailed history from geochemical and isotope data. These data show that the Faroe Islands were deglaciated by the end of Younger Dryas (11,700 – 10,300 cal yr BP), at this time relatively high sedimentation rates with high δ13C imply poor soil development. δ13C, Ti and γ data reveal a much more stable and warm mid-Holocene until 7410 cal yr BP characterised by increasing vegetation cover and build up of organic soils towards the Holocene thermal maximum around 7400 cal yr BP. The final meltdown of the Laurentide ice sheet around 7000 cal yr BP appears to have impacted both ocean and atmospheric circulation towards colder conditions on the Faroe Islands. This is inferred by enhanced weathering and increased deposition of surplus sulphur (sea spray) and erosion in the highland lakes from about 7400 cal yr BP. From 4190 cal yr BP further cooling is believed to have occurred as a consequence for increased soil erosion due to freeze/thaw sequences related to oceanic and atmospheric variability. This cooling trend appears to have advanced further from 3000 cal yr BP. A short period around 1800 cal yr BP appears as a short warm and wet phase in between a general cooling characterised by significant soil erosion lasting until 725 cal yr BP. Interestingly, increased soil erosion seems to have begun at 1360 cal yr BP, thus significantly before the arrival of the first settlers on the Faroe Island around 1150 cal yr BP, although additional erosion took place around 1200 cal yr BP possibly as a consequence of human activities. Hence it appears that if humans caused a change in the Faroe landscape in terms of erosion they in fact accelerated a process that had already started. Soil erosion was a dominant landscape factor during the Little Ice Age, but climate related triggers can hardly be distinguished from human activities.

© 2010 Elsevier Ltd. All rights reserved.

1. Introduction

Various terrestrial and marine records have shown synchronous global climate change through the course of the Holocene (Mayewski et al., 2004; Allen et al., 2007; Wanner et al., 2008; Debret et al., 2009). These (patterns of climatic change are often related to coupled atmospheric–oceanic circulation modes such as ENSO and AO/NAO; Rimbu et al., 2004). Studies of modern oceanography have linked atmospheric NAO with Atlantic Meridional Overturning Circulation (AMOC) and oceanic sea-surface temperature (SST)
The vegetation history of the Faroe Islands has been investigated by numerous palynological and plant macrofossil studies (Hannon et al., 2001, 2003, 2005; Lawson et al., 2005, 2007, 2008; Jessen et al., 2008) all broadly confirming the pioneering works by Persson (1968) and Jóhansen (1985). Jóhansen (1985) described the early-Holocene (> 10,000 cal yr BP) vegetation of the islands as extremely open while the landscape was largely covered by fellfield with gravely and rocky soils formed under a continental climate. According to Jóhansen (1985) climate became more oceanic by around 10,000 cal yr BP. The mid-Holocene (c 8000 cal yr BP) vegetation history was studied by Hannon et al. (2003, 2010) in much greater detail. Although warmer than today, the early-Holocene climate on the Faroe Islands was dynamic and characterised by rapid changes as revealed by variations in organic carbon and nitrogen, biogenic silica, grain size and magnetic susceptibility (Jessen et al., 2008). Grasses and sedges became dominant from c 10,000 cal yr BP and developed into shrub-dominated vegetation indicating a warmer climate (Jóhansen, 1985). Between c 7000 and 1200 cal yr BP increasing concentrations of Calluna pollen and decreasing Juniperus pollen suggest lower summer temperatures and soil degradation (Jóhansen, 1985). In agreement with pollen data an early-Holocene climatic optimum ending around 8400 cal yr BP succeeded by a gradual change to colder climatic conditions is inferred from organic carbon, biogenic silica, grain size and magnetic susceptibility in lake sediments and from foraminifera variations in fjord sediments (Roncaglia, 2004; Andresen et al., 2006). The dominance of grasses and herbs subsequent to the early-Holocene fellfields suggests that soils remained base-rich for several millennia. Gradually soils became more acidic and waterlogged with the development of blanket peats. According to magnetic susceptibility, organic carbon and palynological information substantial erosion began sometime after 5000 cal yr BP alongside generally cooler conditions (Andresen et al., 2006; Lawson et al., 2008).

After c 1150 cal yr BP the first cereal pollen grains indicate the introduction of agriculture. This period has been extensively studied with focus on the first settlers and their possible influence on the Faroese landscape (Jóhansen, 1985; Hannon and Bradshaw, 2000; Hannon et al., 1998, 2001, 2005; Lawson et al., 2005, 2007, 2008). The connection between the arrival of the first settlers to the Faroe Islands and landscape disturbance has been the subject of intense debate. Archaeological and palaeoecological data suggest that substantial settlement (landnám: meaning taking land) by the Norse on the Faroe Islands did not occur until the mid 9th century (Hannon et al., 2001; Arge et al., 2005; Dugmore et al., 2005). Although claims have been made that the first settlers arrived as early as 1450–1250 cal yr BP (500–700 AD, Hannon et al., 2001) the best consensus age for arrival is c 1150 cal yr BP (800 AD) according to Arge et al. (2005) and Dugmore et al. (2005).

A typical response of landscapes to human settlement and land use is an increase in erosion due to destruction of vegetation that binds the topsoil, whether by cultivation or grazing animals (Edwards and Whittington, 2001; Rasmussen and Olsen, 2009; Olsen et al., 2010). The Faroese landnám was in some places accompanied by significant structural changes to the landscape, such as building of ditches (Hannon et al., 1998, 2001; Hannon and Bradshaw, 2000), while at other sites the anthropogenic change to vegetation were much more subtle (Lawson et al., 2008). There seems to have been some geographical variation in pre-settlement vegetation. Lawson et al. (2008) conclude that the landscape on the island of Sandoy today is probably similar to the natural pre-Norse settlement landscape whereas Jóhansen (1985) conclude that pre-settlement landscapes on Streymoy and Eysturoy had a vegetation with more tall herbs and shrubs such as Juniperus and Salix than today. It seems a fact that no extensive woodlands existed on the Faroe Islands although smaller groves of stunned and crooked tree branches have been documented at protected places (e.g. Mahler, 2007). Blanket peat spread across large areas driven purely by natural processes, and grass–sedge communities dominated on the mineral soils (Lawson et al., 2008). Hence the landnám on the Faroe Islands is different than elsewhere principally because there was little deforestation and, given the sparse vegetation, erosional processes were already very important in the landscape. Landscape changes were most acute in protected places where tall herb and low shrub communities were strongly affected by domestic grazing while heath-type and sedge-dominated vegetation was probably less disturbed. During occupation by the Norse, intensive land use was spatially limited by geology, topography and climate conditions, and cultivation seemed not to have made any profound changes to the vegetation and soils except in a few places around settlements on fertile sloping areas (Lawson et al., 2008). As elsewhere, the primary landscape effect of Norse settlements was through animal husbandry.

To supplement the main palynological mid- to late-Holocene studies we are using sediment cores from three Faroese lakes to present XRF data (e.g. Ti, Al, K, Ca, Sr) and organic matter (OM) data (total organic carbon (TOC), total nitrogen (TN), from which we derive C/N, and carbon isotope ratios (δ13C) along with other geochemical (total sulphur (TS), total carbon (TC)) and geophysical proxies to investigate environmental change on the Faroe Islands through the Holocene. Further we compare our observations with the general climate trends of the Northern Hemisphere, in particular those of the North Atlantic Ocean. The investigated lakes are situated fairly close to each other, but are different both hydrologically and limnologically allowing us to qualify the impact and magnitude of local environmental change and thus to distinguish it from regional/hemispheric climate forcing. Overall this investigation has two aims: to investigate environmental change on the Faroe Islands and to study the linkage between the North Atlantic Ocean dynamical behaviour and atmospheric circulation throughout the Holocene.

2. Regional setting

The Faroe Islands are located at approximately latitude 62°N, longitude 7°W and consist of 18 small, hilly islands. The islands have a total area of 1399 km², and extend 113 km from north to south and 75 km from east to west (Fig. 1). The highest elevations, reaching nearly 890 m a.s.l. are found in the northern islands. The islands are remnants of a large low relief plateau of Tertiary flood basals deriving from volcanic eruptions almost 60 Myr ago prior to the opening the North Atlantic ocean basin (Rasmussen and Noe-Nygard, 1970).

The climate on the Faroe Islands is greatly influenced by the warm North Atlantic Drift (NAD) and by the passage of frequent cyclones. Accordingly, the climate is humid, unsettled and windy with mild winters and cool summers (Cappelen and Laursen, 1998).
Precipitation reflects the local topography and is smallest near coastal regions (c 1000 mm/a) and peaks (c 3000 mm/a) in the most elevated inland regions (Cappelen and Laursen, 1998). The Faroe Islands are situated in the main path of the NAD. The part of the NAD flowing west and north of the islands is the Faroe Current. Polar water of the East Icelandic Current flows towards the Faroe Islands from the northwest and is separated from the Faroe Current by the Iceland–Faroe front (Fig. 1).

Lake Mjáuvötn is situated in an elongated basin and consists of two parts: Eystara (eastern) Mjáavatn and Vestara (western) Mjáavatn (Fig. 1). The lake is located at an altitude of 76 m a.s.l. in a high relief alpine landscape with 52% of its catchments area above 300 m a.s.l. (Mortensen, 2002). The lake has approximately the same depth, around 6–7 m, in both parts (Fig. 1). Mjáuvötn is surrounded by flat shores consisting mostly of sand and gravel, and the catchment consists of extensive grassy fields growing on thin pebbly soils. Input of nutrients and other particulate matter to the lake originates from natural erosion of the surrounding soils and bedrock. Currently, Mjáuvötn is an oligotrophic lake with a low phytoplankton biomass (Brettum, 2002; Jensen et al., 2002), but holds a wide range of phytoplankton species (c 20, Brettum, 2002) found in nutrient poor and alpine lakes. Cryptomonads comprise c 80% of total biomass volume; the most abundant species belong to the genera Chroomonas and Cryptomonas. Mjáuvötn contains sparse submerged vegetation dominated by charophytes and isoetids, but includes also elodeids. Vascular plants dominate the littoral zone in 0–5 m water depth and charophytes from 5 to 10 m (Schierup et al., 2002).

Brúnavatn and Stóravatn are mountain plateau lakes situated at c 260 m a.s.l. (Fig. 1). The highland plateau is a relatively flat area consisting of extensive grassland and wetlands with rolling hills rising up to about 300 m a.s.l. towards the north—northwest where it reaches the mountains with elevations up to 585 m a.s.l. The river Tyggará drains the main part of the area, but a smaller proportion is drained by the brook running into Brúnavatn (Fig. 1). Brúnavatn and Stóravatn are adjacent basins and separated by a narrow bedrock ridge with an elevation difference of c 10 m, with Stóravatn situated at the higher elevation.

Brúnavatn is surrounded by relatively steep sided slopes (up to a few metres high) towards the northeast, southeast and southwest with stony shores. Towards the northwest the area is relatively flat and consists of a swampy peatland adjacent to Brúnavatn. The northwest shore consists of sand and gravel. The major inlet to Brúnavatn runs through the peatland area, draining the area to the north—northwest towards the mountains, and colouring the water of the lake (Brún = brown). Stóravatn is surrounded by relatively steep slopes and has a stony and sandy shore.
Mjáuvötn receives high amounts of surface run-off water compared to Stórvatn and Brúnavatn (Mortensen, 2002) and all lakes have an open hydrology dominated by precipitation according to monitoring 2007–2009 of lake water δD and δ18O (Olsen et al., submitted for publication).

3. Materials and methods

The sediments were collected using a 100 cm long Russian corer with a diameter of 8 cm and with 20 cm overlap between drives. The sediment successions at Mjáuvötn, Brúnavatn, and Stórvatn were cored in water depths of 558, 118 and 458 cm, respectively, and the length of each sequence is 910, 259 and 426 cm, respectively. Top sediments for 210Pb dating were retrieved using a kajak corer. All depths are expressed as cm below the water surface. The overlap between the 100 cm core sections has been determined by cross-correlating high resolution XRF data.

Samples from the top 30 cm of Mjáuvötn and Stórvatn sediments were analysed for the activity of 210Pb, 226Ra and 137Cs via gamma spectrometry at the Gamma Dating Centre, Department of Geography and Geology, University of Copenhagen. The measurements were carried out on a Canberra ultralow-background Germanium detector. 210Pb was measured via its gamma-peak at 46.5 keV, 226Ra via the granddaughter 214Pb (peaks at 295 and 950 keV) contaminants, such as carbonates and inorganic fraction) (Abbott et al., 1996). After chemical pre-treatment graphite for AMS14C measurements via reduction with H2 using a Carlo Erba 1500 online to a VG TripleTrap and Optima dual-inlet mass spectrometer at the NERC Isotope Geosciences Laboratory (UK), δ13C values were calculated to the VPDB scale using within-run laboratory standards calibrated against NBS-19 and NBS-22. Replicate analysis of well-mixed samples indicated a precision of ±0.1‰ (1 SD). TOC and TN percentages were also measured (from which C/N were derived and presented in atomic units) and calibrated against an Acetanilide standard. Replicate analysis of well-mixed samples for TOC and TN indicate a precision of ±<0.1. Principal component analysis (PCA) was conducted using MatLab 7.8 on a normalised dataset. Normalisation was done by dividing each parameter with its standard deviation and by log transforming all percentage values.

4. Results

4.1. Lithology

The lowest sediments in the lakes mostly consist of homogenous gray sandy clays and do not contain macrofossils (Fig. 2), in Brúnavatn these sediments are underlain by coarse gravel. In Stórvatn the sediments contain small gravel clasts at the base of the core. From 1446, 296 and 840 cm in Mjáuvötn, Brúnavatn and Stórvatn, respectively, the sediments change to a light brown clay gyttja with scattered and numerous terrestrial macrofossils remains of Racotritium, Hylocomium and Salix herbacea (Fig. 2). From 1340 and 775 cm the Mjáuvötn and Stórvatn sediments show a gradual change to a brown dark-brown detritus gyttja with an upwards increasing content of organic matter. A similar change is observed in the Brúnavatn sediments at 280 cm, but here as a sharp boundary between the two sediment facies indicating a sediment hiatus (Fig. 2). The Mjáuvötn sediments contain scattered plant remains and small clasts throughout the sequence, and with the lowest organic content of the three sequences, whereas the Brúnavatn sediments display highest organic content (Fig. 2). Numerous visible tephra layers occur in all three sediment sequences, from millimetre to centimetre thick.

4.2. Chronology

The age models for the three Faroese lakes are based on 40Ar/39Ar dated samples and on 11 tephra layers (see online Supplementary Material S1). Of the 40Ar/39Ar samples, 31 dates are from plant macrofossils, the remaining 9 are from paired humic acid (HA) and humic acid residues (HAR) (Table 1, Fig. 3). Additionally, the top sections of Mjáuvötn and Stórvatn are dated using 210Pb and depth-to-age relationships constructed using a modified CRS modelling approach (Appleby, 2001), in which the activity in the deepest part of the cores was estimated on the basis of a regression of ln (210Pbxs) vs. accumulated mass depth.
A uniform age probability distribution with a span of 1 year has been used in the age model for the historic Hekla 1104. The Mjáuvötn Tephra A (MTA) is dated to 6596 ± 66 cal yr BP (Olsen et al., in press; Wastegård et al., 2001) and the age for the Saksunarvatn tephra (SAK) is from the NGRIP ice core, 10,347 ± 45 kyr BP (GICC05) (Rasmussen et al., 2007), i.e. 10297 ± 45 cal yr BP. Multiple dates exist for Hekla 3, Hekla S and Hekla 4 tephras and these ages were combined to a single age and outliers were removed (Table 2). The ash layer at 1358.5 cm in Mjáuvötn is assumed to be the SAK tephra (Table 1).

In particular, the mid-Holocene sediments have insufficient amounts of macrofossils for 14C dating. Therefore other substances like humid acids were 14C dated to obtain satisfactory chronologies. However, chronologies based on 14C dating of different bulk sedimentary organic fractions have their problems; these include low...
Table 1

Chronological information. $^\delta^{13}$C values in parentheses denote estimated $^\delta^{13}$C values used for fractionation correction to conventional $^14$C dates by the AMS Dating Centre, University of Aarhus. $^\delta^{13}$C dates in parentheses are excluded from age models. HA – humic acid (alkali soluble fraction) and HAr – humic acid residue (alkali insoluble fraction). Botanical nomenclature follows Tutin et al. (1964–1980) and Smith (2004).

<table>
<thead>
<tr>
<th>Lab. ID/Tephra</th>
<th>Material</th>
<th>Depth, cm</th>
<th>$^\delta^{13}$C, VPDB</th>
<th>Age, $^14$C BP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brunavatn</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12241</td>
<td>Stem fragment of Racomitrium (terrestrial moss) and leaf of juniperus</td>
<td>140.5</td>
<td>(−27)</td>
<td>945 ± 55</td>
</tr>
<tr>
<td>AAR-12267</td>
<td></td>
<td>141.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12240</td>
<td>Racomitrium sp. (moss)</td>
<td>152.5</td>
<td>−27.53</td>
<td>1054 ± 35</td>
</tr>
<tr>
<td>AAR-12239</td>
<td>Little twig or root, indet.</td>
<td>158.5</td>
<td>−24.42</td>
<td>1383 ± 20</td>
</tr>
<tr>
<td>Hekla 3°</td>
<td></td>
<td>159.5</td>
<td>−25.20</td>
<td>2463 ± 40</td>
</tr>
<tr>
<td>Hekla 5°</td>
<td></td>
<td>175.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12242</td>
<td>Characeae (macro-algae) spores</td>
<td>201.5</td>
<td>−24.13</td>
<td>3120 ± 55</td>
</tr>
<tr>
<td>Hekla 4°</td>
<td></td>
<td>216.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12453</td>
<td>HA</td>
<td>248.0</td>
<td>−25.62</td>
<td>(5670 ± 60)</td>
</tr>
<tr>
<td>AAR-12451</td>
<td>HAr</td>
<td>248.0</td>
<td>−22.24</td>
<td>(7097 ± 48)</td>
</tr>
<tr>
<td>AAR-12448</td>
<td>Unidentified</td>
<td>240.5</td>
<td>−24.31</td>
<td>5320 ± 55</td>
</tr>
<tr>
<td>AAR-12444</td>
<td>cf. Vaccinium myrtillus</td>
<td>260.5</td>
<td>(−27)</td>
<td>5635 ± 65</td>
</tr>
<tr>
<td>Mjaavotn Tephra A</td>
<td></td>
<td>262.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12274</td>
<td></td>
<td>277.5</td>
<td>(−26)</td>
<td>6230 ± 75</td>
</tr>
<tr>
<td>AAR-12816.1</td>
<td>HA</td>
<td>279.0</td>
<td>−25.48</td>
<td>(6098 ± 48)</td>
</tr>
<tr>
<td>AAR-12816.2</td>
<td>HAr</td>
<td>279.0</td>
<td>−22.67</td>
<td>(6095 ± 65)</td>
</tr>
<tr>
<td>Combined AAR-12816.1, -12816.2</td>
<td></td>
<td>279.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12466</td>
<td>Racomitrium aquaticum (moss)</td>
<td>290.5</td>
<td>−21.71</td>
<td>9910 ± 70</td>
</tr>
<tr>
<td>Mjaavotn</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12615</td>
<td>Polystichum (moss)</td>
<td>564.5</td>
<td>−24.19</td>
<td>−2031 ± 28b</td>
</tr>
<tr>
<td>AAR-12616</td>
<td>Rhytidiothecium lutescens (2 moss stem fragments), Hylocomium splendens (2 moss stem fragments), Dicranella subulata (7 moss stems), Racomitrium sp. (4 moss stems)</td>
<td>589.5</td>
<td>−24.09</td>
<td>199 ± 36</td>
</tr>
<tr>
<td>AAR-12617</td>
<td>Unidentified little twig, Anisotrichum palustre (moss), Dicranella subulata (moss)</td>
<td>620.5</td>
<td>(−27)</td>
<td>264 ± 46</td>
</tr>
<tr>
<td>AAR-12618</td>
<td>Pleurozium schreberi (2 moss stem fragments), unidentified twigs</td>
<td>719.5</td>
<td>−26.64</td>
<td>816 ± 43</td>
</tr>
<tr>
<td>AAR-12619</td>
<td>Rhytidiothecium squarrosum (2 moss stems), Hylocomium splendens (moss), Anisotrichum palustre (moss), Bryum sp. (moss)</td>
<td>745.5</td>
<td>−24.87</td>
<td>889 ± 31</td>
</tr>
<tr>
<td>AAR-12621</td>
<td>Racomitrium fasiculare, Hylocomium splendens (2 moss stem fragments), Dicranella subulata (moss)</td>
<td>837.5</td>
<td>−24.96</td>
<td>1326 ± 33</td>
</tr>
<tr>
<td>AAR-12620</td>
<td>Unidentified terrestrial leaf fragments</td>
<td>919.5</td>
<td>(−27)</td>
<td>1790 ± 90</td>
</tr>
<tr>
<td>AAR-12625</td>
<td>cf. Dicranella cerviculata (moss)</td>
<td>935.0</td>
<td>−24.68</td>
<td>2070 ± 120</td>
</tr>
<tr>
<td>AAR-12622</td>
<td>Hylocomium splendens (moss)</td>
<td>1048.5</td>
<td>(−27)</td>
<td>2785 ± 50</td>
</tr>
<tr>
<td>Hekla 5°</td>
<td></td>
<td>1137.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hekla 4°</td>
<td></td>
<td>1174.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12623</td>
<td>Salix herbacea (1 leaf), Hylocomium splendens (moss)</td>
<td>1177.5</td>
<td>(−27)</td>
<td>4890 ± 80</td>
</tr>
<tr>
<td>AAR-13269</td>
<td>Racomitrium canescens (Hedw.) Brid. (2 moss stems), Dicranacea indet. (16 moss stems), Calliergonella cuspidata (1 moss stem), Calluna vulgaris (2 twigs without leaves)</td>
<td>1248.5</td>
<td>−23.20</td>
<td>5330 ± 36</td>
</tr>
<tr>
<td>AAR-13268</td>
<td>Dicranacea indet. (34 moss stems), Philonotis cf. fontana (2 moss stems), Plagiothecium sp. (1 moss stem), leaf fragments (7)</td>
<td>1250.5</td>
<td>−25.17</td>
<td>5450 ± 70</td>
</tr>
<tr>
<td>AAR-13272.2</td>
<td>HA</td>
<td>1260.0</td>
<td>−26.98</td>
<td>(5831 ± 35)</td>
</tr>
<tr>
<td>AAR-13272.1</td>
<td>HAr</td>
<td>1260.0</td>
<td>−25.53</td>
<td>(5934 ± 35)</td>
</tr>
<tr>
<td>AAR-13271.2</td>
<td>HA</td>
<td>1268.0</td>
<td>−27.03</td>
<td>(6103 ± 37)</td>
</tr>
<tr>
<td>AAR-13271.1</td>
<td>HAr</td>
<td>1268.0</td>
<td>−25.69</td>
<td>(6223 ± 47)</td>
</tr>
<tr>
<td>Mjaavotn Tephra A</td>
<td></td>
<td>1276.5</td>
<td></td>
<td>5500 ± 100</td>
</tr>
<tr>
<td>Mjaavotn</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-13270</td>
<td>Dicranum scoparium (1 moss stem), Racomitrium canescens (Hedw.) Brid. (1 moss stem), Dicranacea indet. (34 moss stems), Philonotis sp. (2 moss stems), Campylium stellatum (1 moss stem), Bryum sp. (1 moss stem), Hylocomium splendens (3 moss stems), Deyryhynchium vawarzi (1 moss stem)</td>
<td>1286.5</td>
<td>−21.30</td>
<td>5964 ± 48</td>
</tr>
<tr>
<td>AAR-12624</td>
<td>Bryum sp. (moss), Hylocomium splendens (moss), Dicranella subulata (moss)</td>
<td>1292.5</td>
<td>(−27)</td>
<td>6275 ± 65</td>
</tr>
<tr>
<td>SAK</td>
<td></td>
<td>1358.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12820</td>
<td>Hylocomium pyrenaicum (moss)</td>
<td>1384.5</td>
<td>−26.61</td>
<td>9480 ± 75</td>
</tr>
<tr>
<td>AAR-12821</td>
<td>Salix herbacea, leaves</td>
<td>1396.5</td>
<td>−27.66</td>
<td>9420 ± 75</td>
</tr>
<tr>
<td>AAR-12819</td>
<td>Resting eggs (ephippiae) of Cladocera, torn off leaves of several mosses</td>
<td>1423.5</td>
<td>−25.41</td>
<td>9880 ± 75</td>
</tr>
<tr>
<td>Storavatn</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12449</td>
<td>2 twigs, terrestrial</td>
<td>576.0</td>
<td>−26.15</td>
<td>920 ± 45</td>
</tr>
<tr>
<td>AAR-12462</td>
<td>HA</td>
<td>629.0</td>
<td>−26.66</td>
<td>(2243 ± 31)</td>
</tr>
<tr>
<td>AAR-12460</td>
<td>HAr</td>
<td>629.0</td>
<td>−25.01</td>
<td>2046 ± 31</td>
</tr>
<tr>
<td>AAR-12459</td>
<td>HA</td>
<td>680.0</td>
<td>−26.06</td>
<td>(2631 ± 37)</td>
</tr>
<tr>
<td>AAR-12457</td>
<td>HAr</td>
<td>680.0</td>
<td>−23.56</td>
<td>2550 ± 37</td>
</tr>
<tr>
<td>AAR-12246</td>
<td>juniperus</td>
<td>704.5</td>
<td>−26.92</td>
<td>2952 ± 37</td>
</tr>
<tr>
<td>Hekla 5°</td>
<td></td>
<td>738.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hekla 4°</td>
<td></td>
<td>749.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AAR-12817.1</td>
<td>HA</td>
<td>760.0</td>
<td>−26.18</td>
<td>(4695 ± 60)</td>
</tr>
<tr>
<td>AAR-12817.2</td>
<td>HAr</td>
<td>760.0</td>
<td>−23.45</td>
<td>(4750 ± 75)</td>
</tr>
<tr>
<td>Combined AAR-12817.1, -12817.2</td>
<td></td>
<td>760.0</td>
<td></td>
<td>4716 ± 48</td>
</tr>
</tbody>
</table>

(continued on next page)
OM content, often with uncertain origin and slow decomposition rates. Slow terrestrial carbon decomposition allows OM to be stored for prolonged periods prior to being transported into the lake, resulting in apparently too old \(^{14}C\) ages. Even macrofossils may yield apparent too old \(^{14}C\) ages because of reworking from pre-existing sediments and soils. The decomposition of ancient carbon may generate a significant reservoir effect even for living aquatic plants (Abbott et al., 1996; Wolfe et al., 2004; Geirsdóttir et al., 2009).

However, dating of bulk sediment from the softwater lakes of the Faroe Islands has often shown to give reliable ages (Björck and Wohlfarth, 2001). In general the HA and HAr \(^{14}C\) dates yielded identical ages and a combined age of the fractions has accordingly been used in the age model. Where discrepancies exist the youngest \(^{14}C\) date is chosen as this is believed to best reflect the true sediment age (Table 1). Only in a few cases where both the HA and HAr fraction are significantly older than adjacent \(^{14}C\) dates, these results have not been considered in the age model (Table 1) (Olsen et al., in press).

The age model was constructed using depositional models in OxCal 4.1 (Ramsey, 2008, 2009). Age-to-depths relationships for Mjáuvötn, Brúnavatn and Stóravatn were constructed in a combined age model by cross-linking identical tephra layers in between the cores (Fig. 3B). Initial model runs where used to estimate the model parameter \(k\) for each core and to identify possible outliers. Based on the initial age model for each site and applying the measured geochemical and geophysical parameters, further age models included additional cross links between sites such that all zone boundaries from zone 5 to 13 are cross correlated (Fig. 3). The final depositional model yielded an agreement index of 79.2\% with a \(k\) value of 50. The base date of each core was determined by interpolation and is therefore uncertain (Fig. 4).

A sharp sedimentary boundary as well as the \(^{14}C\) dates indicates that there is a hiatus in Brúnavatn at 280 cm (Figs. 2 and 3). Because only one \(^{14}C\) date exists below the hiatus the dating of the early-Holocene sequence of Brúnavatn is based on cross correlation to the absence of the SAK tephra in the Brúnavatn sequence implies that the hiatus spans the age of this commonly found tephra on the islands close to the sea. In the case of the Faroese lakes we assume that Br can be a proxy for marine aerosols, i.e. sea spray, which was transported from the coast to the site. Higher Br values should then indicate higher wind activity. The positive correlation between Br and TOC suggests that Br is adsorbed to organic matter (Boyle, 2001; Sai et al., 2007). The modern plants generally fall into the global aquatic plants (moss) 816.5 (moss) 836.5 (Fig. 4). The XRF counts of Br from the three lakes show different absolute values. The early-Holocene \(\delta^{13}C\) and \(\delta^{13}C/N\) values are positively correlated with Al, Ti, K and Sr (Fig. 4).

Br is often used as a geochemical proxy for hypersaline lakes and playa salt deposits (Ullman, 1995; Risacher and Fritz, 2000) and is abundant as bromide salts in sea water, but it occurs in only extremely small concentrations in fresh water systems (Song and Müller, 1993). Br is therefore a useful indicator for saline waters. Its dominant aqueous species, \(Br^-\), shares many chemical characteristics with the abundant salt constituent chloride, Cl\(^-\) (Ullman, 1995). Br can thus be used as an indicator of direct influence of marine water on lakes that are located close to the sea. In the case of the Faroese lakes we assume that Br can be a proxy for marine aerosols, i.e. sea spray, which was transported from the coast to the site. Higher Br values should then indicate higher wind activity. The positive correlation between Br and TOC suggests that Br is adsorbed to organic matter (Boyle, 2001; Sai and King, 2004; Schmidt et al., 2006).

The \(\delta^{13}C\) and \(\delta^{13}C/N\) values from modern plants collected in and around the three lakes (Olsen et al., submitted for publication) show significant differences between aquatic and terrestrial plant values (Fig. 5). Mean \(\delta^{13}C\) values and \(\delta^{13}C/N\) ratios are for terrestrial plants \(-12.7\) at 280 cm (Fig. 3). Because only one \(^{14}C\) date exists below the hiatus the dating of the early-Holocene sequence of Brúnavatn is based on cross correlation to variations in \(\delta^{13}C\) and \(\delta^{13}C/N\) values in Störavatn and Mjáuvötn. This part of the age model may therefore, to some extent, be dubious. Further, the absence of the SAK tephra in the Brúnavatn sequence implies that the hiatus spans the age of this commonly found tephra on the Faroe Islands, and that the age of the sediments below the hiatus are older than 10,300 cal yr BP. A \(k\) value of 200 for the early-Holocene Brúnavatn sequence has been applied. The length of the hiatus is determined to be 4000 ± 100 yrs (Fig. 3).

4.3. Data description

PCA on parameters from each core suggest that generally two major groups (an organic and a minerogenic) explain the major variability of the three lake sediment sequences and can be accounted for by the first principal component (PCA 1, Fig. 4). The organic group elements are TOC, TN, TS and Br and the minerogenic group elements are Ti, Al, K, Ca, Sr and \(\chi\) (Fig. 4). The XRF counts of Ti, Al, K, Ca and Sr in Mjáuvötn, Brúnavatn, and Stóravatn are strongly correlated (\(r > 0.90\), Fig. 4) and roughly reflect the amount of minerogenic material, including weathered minerals, transported to the lake from the catchments. Furthermore, these element counts correlate with \(\chi\) (\(r > 0.60\), Fig. 4). Interestingly, Si is correlated to Ti in two of the lakes but only weakly in Brúnavatn (\(r = 0.39\)). This may indicate that while the Si in Mjáuvötn and Stóravatn may mainly have been derived from weathering, the Si content in Brúnavatn is more strongly influenced by biogenic silica such as diatoms and phytooliths. The OM data (TOC, TN, TS and Br) are all either positively or negatively correlated with \(\delta^{13}C\) values are positively correlated with Al, Ti, K and Sr (Fig. 4).

\(\begin{array}{|c|c|c|c|}
\hline
\text{Lab. ID/Tephra} & \text{Material} & \text{Depth, cm} & \delta^{13}C_{\text{vc}} \text{ VPDB} & \text{Age, } \text{^{14}C}\text{BP} \\
\hline
\text{AAR-12818.1} & \text{HA} & 770.0 & -26.00 & (6199 ± 50) \\
\text{AAR-12818.2} & \text{HAr} & 770.0 & -23.34 & (6190 ± 50) \\
\text{Combined AAR-12818.1, -12818.2} & & 770.0 & 6195 ± 35 & \\
\text{AAR-12465} & \text{HA} & 779.0 & -24.51 & 8493 ± 46 \\
\text{AAR-12463} & \text{HAr} & 779.0 & -21.36 & 8660 ± 80 \\
\text{SAK} & & 790.5 & 8991 ± 49 & \\
\text{AAR-12279} & \text{Racomitrium (moss)} & 804.5 & -25.34 & 9391 ± 49 \\
\text{AAR-12280} & \text{Racomitrium (moss)} & 816.5 & -25.45 & 9449 ± 44 \\
\text{AAR-12247} & \text{Racomitrium (moss)} & 836.5 & -23.46 & 9940 ± 55 \\
\hline
\end{array}\)

\(\text{See Table 2 for age details.}\)

\(\text{bomb4NH1 calibration curve applied (Hua and Barbetti, 2004).}\)
Fig. 3. (A) Calculated accumulation rate based on age models. Zones are based on proxy parameter variations (see further details in main text). Calibrated dates of each zone are shown at the top. (B) Combined age models for Brúnavatn, Stóravatn and Mjáuvötn. Dashed lines are depth correlations based on either ash layers or proxy parameters according to zones (see further details in main text).
The temporal variability is presented in Figs. 6–8 and is divided into 13 zones. Zones are made according to variations in proxy values in each lake separately and may therefore not reflect common changes across all lakes at a time. The presented temporal variability is described in the context of the early-, mid- and late-Holocene time periods below. Additionally, the first principal component (PCA 1) is presented in Fig. 9. PCA 1 explains 68.8%, 62.0% and 65.6% of the observed variability of Mjáuvötn, Brúnavatn and Stóravatn respectively.

4.3.1. Early-Holocene (c 11,700–10,300 cal yr BP)

Data from the lowermost part (zone 1, c 11,430 cal yr BP) is limited and only exist from the Mjáuvötn and Brúnavatn cores. Zone 1 is characterised by high χ and Ti values (Fig. 6). Mjáuvötn TOC (0.16 ± 0.07%) and Br (84 ± 23 counts) values are very low, which is also reflected as low Br values in Brúnavatn (Fig. 6). The δ13C values in both Mjáuvötn and Brúnavatn are around −26‰ (Fig. 6).

Zone 2 (11,430–10,105 cal yr BP) is marked by decreased χ and Ti and slowly increasing to constant TOC in Mjáuvötn. In Brúnavatn the Ti and χ are decreasing whereas Stóravatn shows constant values (Fig. 6). TOC values in Brúnavatn and Stóravatn display an increasing trend. δ13C shows elevated values in all three lakes (Fig. 6). Zone 3 (11,015–10,935 cal yr BP) in Brúnavatn and Stóravatn shows a decrease in χ and Ti whereas zone 3 starts in Mjáuvötn with a change to higher χ and Ti followed by decreasing values. TOC increases further in Brúnavatn and Stóravatn towards 5% and 2%, respectively, while Mjáuvötn still shows fairly low values around 0.35% (Fig. 6). The δ13C values are around −25.5‰ and −23‰ for Mjáuvötn and Stóravatn, respectively (Fig. 6), while there was no deposition in Brúnavatn between c 11,130 and 7125 cal yr BP. Zone 4 (10,935–10,705 cal yr BP) shows decreasing χ and Ti (Fig. 6). Stóravatn shows higher TOC (between 4 and 7%) and C/N as high as 22 (Fig. 6). TOC in Mjáuvötn is around 0.7%. Zone 5 (10,705–10,300 cal yr BP) is characterised by lower TOC (c. 4%) in both Mjáuvötn and Brúnavatn, c 26‰ to −25‰ and −23‰ to −24‰ respectively (Fig. 6). Ti and χ decrease in both lakes (Fig. 6).

4.3.2. Early- to Mid-Holocene (10,300–3200 cal yr BP)

A rapid increase in χ and Ti marks the transition from zone 5 to 6, concurrent with increasing δ13C (Fig. 7). Thereafter the Stóravatn χ and Ti decrease, a trend also seen in the Mjáuvötn χ and Ti, albeit to a lesser extent (Fig. 7). TOC increases while δ13C decreases in both Mjáuvötn and Stóravatn (Fig. 7). In both lakes C/N ratios are around 12 and gradually increase.

Zone 7 is characterised by almost constant χ and Ti in Mjáuvötn and Stóravatn (Fig. 7). This is also the case (from 7000 cal yr BP) for

![Fig. 4](image-url)

**Fig. 4.** Biplots of PCA components 1 and 2 for Mjáuvötn, Brúnavatn and Stóravatn based on the combined set of variables measured. PCA axes 1 and 2 explain 72.2%, 75.9% and 77.2% of the total observed variance of Mjáuvötn, Brúnavatn and Stóravatn, respectively.
Brúnavatn until 6320 cal yr BP where a significant peak in C appears which might correspond to the MTA tephra (Fig. 7). TOC and $\delta^{13}$C continue to increase and decrease, respectively, though the rise in TOC is much slower in Brúnavatn and Stóravatn (Fig. 7). In zone 8 the Brúnavatn and Stóravatn $\chi$ and Ti decrease whereas the Mjáuvötn $\chi$ and Ti show no change (Fig. 7). TOC stabilises around c. 4, 8 and 12% in Mjávötn, Brúnavatn and Stóravatn, respectively, and with C/N ratios of 15–18. $\delta^{13}$C continues its decreasing trend with the exception of Stóravatn where $\delta^{13}$C increases (Fig. 7).

4.3.3. Late-Holocene (3200 cal yr BP–present)

Zone 9 is defined by the significant increase in Brúnavatn of the C/N and C/S ratios and a decrease in $\delta^{13}$C to around $-27.5\%_{o}$ (Fig. 8). Both Mjáuvötn and Stóravatn have gradually decreasing $\chi$ and Ti, increasing TOC while $\delta^{13}$C decreases to around $-27.5\%_{o}$ and $-26\%_{o}$, respectively (Fig. 8). Zone 10 shows an increase in Mjáuvötn and Stóravatn in $\chi$ and Ti succeeded by decreasing values and increasing TOC (Fig. 8). Mjáuvötn $\delta^{13}$C values are constant, whereas Stóravatn $\delta^{13}$C drops to c. $-27.5\%_{o}$. The high Brúnavatn TOC and C/N begin to decline while $\delta^{13}$C remains low around $-27.5\%_{o}$ (Fig. 8). Zone 11 is defined by increasing $\chi$, Ti and C/N and decreasing TOC in Mjáuvötn and Brúnavatn (Fig. 8). Stóravatn shows decreasing $\chi$ and Ti values and increasing C/N ratios and TOC values (Fig. 8). $\delta^{13}$C values in Brúnavatn and Stóravatn remain low with a decreasing and increasing trend, respectively, while they increase in Mjáuvötn (Fig. 8). Zone 12 in Mjáuvötn is marked by decreasing $\chi$ and Ti and increasing C/N and TOC, whereas the Brúnavatn and Stóravatn records show opposite trends (Fig. 8). Zone 13 displays decreasing $\chi$ and Ti in Mjáuvötn and Brúnavatn and increasing $\chi$ and Ti in Stóravatn (Fig. 8). C/N and TOC decreases in all three lakes (Fig. 8). Mjáuvötn $\delta^{13}$C increases value, while the opposite is found in Brúnavatn and Stóravatn (Fig. 8).

5. Discussion

5.1. Early-Holocene (c. 11,800–10,300 cal yr BP)

The onset of sedimentation is estimated by extrapolation of the age models to 11,630 ± 160 cal yr BP, 12,105 ± 295 cal yr BP and
12,365 ± 330 cal yr BP for Mjáuvötn, Brúnavatn and Stóravatn, respectively (Fig. 3). Brúnavatn and Stóravatn were cored as deep as possible and at the Brúnavatn site the last drive reached gravel sediments indicating that sediments from the entire postglacial lake history likely were retrieved. Radiocarbon dating of bottom sediments in c 10 different basins on the Faroe Islands (Andresen et al., 2006; S. Björck, unpublished; Jessen et al., 2008) has shown that it is difficult to retrieve pre-Holocene sediments. The oldest 14C ages are in the range of 10,200–10,400 14C yrs BP (c 12,100 cal yr BP, all on bulk sediments). This suggests that the Faroe archipelago was covered by an ice cap at least until the later part of Younger Dryas. Previous 14C measurements (S. Björck, unpublished) indicate that the older ages are often found in lakes/bogs outside valleys and at higher altitudes, which is also the case for Stóravatn and Brúnavatn. This indicates that the Faroe ice cap experienced a rapid down-wasting in late Younger Dryas, and that glacial ice survived longest in the deep and narrow valleys but completely disappeared during the Preboreal warming (cf. Humlum et al., 1996; Humlum and Christiansen, 1998).

The sparse vegetation cover during the early-Holocene is likely to have been accompanied by high erosion rates which is reflected in the high accumulation rates (c 8 mm/yr for Mjáuvötn and c. 4 mm/yr for Brúnavatn and Stóravatn) as well as the mainly minerogenic sediments in all three lakes during the early-Holocene (Figs. 3A, 4 and 5–9).

The mechanisms responsible for the hiatus in Brúnavatn are not clear, but extensive coring campaigns on the islands by Björck et al. have shown that hiatus are fairly common in many of the shallower lakes. The occurrence of hiatus in shallow lakes may be related to sediment resuspension and erosion caused by high wind speeds. High wind speeds from variable directions are often experienced on the Faroe Islands between and at the foot of steep mountain slopes. Such conditions may create strong counter currents to the prevailing wind directions in near-by lakes, which can easily mobilize surface sediments and cause erosion and sediment refocusing. The fact that sedimentation covering c 4000 years is missing, corresponding to perhaps 1 m of sediments and including the fairly heavy tephra particles of the SAK tephra, implies a long and efficient erosion period. The onset of sedimentation by c 7100 cal yr BP indicates that wind conditions/directions may have changed considerably by then.

The sedimentary early-Holocene record reported here is remarkably similar to the highly variable record from Sandoy, Faroe Islands (Jessen et al., 2008). The very high χ and Ti and limited OM of the homogenous gray sandy clays of zone 1 are consistent with a landscape with little or no vegetation cover and poorly developed soils (Figs. 6 and 9). A sedimentary change in zone 2 (11,460 cal yr BP) to light brown gyttja with lower χ and Ti values probably reflects the establishment of terrestrial vegetation (Jessen et al., 2008). The enhanced TS values and decreased C/S in zones 3–5 in Stóravatn (11,255–10,300 cal yr BP, Figs. 6 and 9) may indicate deposition of sea spray (NaSO4) as also observed in Lykkjuvötn (Jessen et al., 2008).

The early-Holocene OM generally displays higher δ13C values than the modern terrestrial plants from the catchment, but values do lie within the global field of terrestrial plants. However, samples from zones 3 and 4 have low δ13C values, with a trend towards the field of terrestrial plants (Fig. 5). High δ13C (especially zones 2 and 5) combined with low C/N and TOC values suggest that aquatic plants contributed significantly to the OM (e.g. Hammarlund, 1993; Turney,
1999; Nunez et al., 2002; Mackie et al., 2007). High $\delta^{13}C$ values in aquatic plants are common in lakes with longer water residence times leaving aqueous C enough time to equilibrate with the $^{13}C$ enriched atmospheric CO2 (Heaton, 1999; Turney, 1999; Mackie et al., 2007). Long residence time possibly suggests a drier climate.

5.2. Early- to mid-Holocene (10,300 – 3000 cal yr BP)

5.2.1. Zones 6–7 (10,300 and 4190 cal yr BP)

Between 10,300 and 7400 cal yr BP the accumulation rates are significantly lower (2 mm/year for Mjáuvötn and 0.4 mm/yr for Stóravatn) which is likely reflecting the increasing vegetation cover and a build up of organic soils (Fig. 3A, Jóhansen, 1985; Lawson et al., 2007). These processes are also implied by the lower Ti and $\delta^{13}C$ and rising C/N ratios (Figs. 3, 7 and 9). Overall, slowly decreasing Ti and $\chi$ in the lowland lake Mjáuvötn indicates stable conditions in zones 6 and 7 (Figs. 7 and 9). In contrast, the rapidly increasing $\chi$ and Ti in Stóravatn at the beginning of zone 6 (c 10,300 cal yr BP) suggest increased catchment erosion in the highlands (Figs. 7 and 9), either caused by the so-called 10.3 ka climate cooling found on the Faroe Islands (Björck et al., 2001) or by the introduction of particles from the SAK tephra. The succeeding slow decrease of $\chi$ and Ti appears as a gradual return to low catchment erosion rates throughout zone 6. Zone 7 shows significant variations in Brúnavatn and Stóravatn $\chi$ and Ti records from 7410 to 4190 cal yr BP and further demonstrate the differences between low- and highland environments with a re-appearance of increased highland erosion at c 6300 cal yr BP (Fig. 7).

Despite the differences in $\chi$ and Ti variability all three records display decreasing $\delta^{13}C$ values and increasing OM and C/N ratios indicating either an increased terrestrial OM fraction or increasing lacustrine organic productivity (which would increase $\delta^{13}C$ and OM, but not C/N unless there was a change in plant type) during this time period (zones 6–7, Fig. 7). When compared to the global ranges of aquatic and terrestrial plants, the mid-Holocene Mjáuvötn $\delta^{13}C$ values display a mixed OM in contrast to a more aquatic origin indicated by high $\delta^{13}C$ values and low C/N ratios at Brúnavatn and Stóravatn (Fig. 5). The parallel increase in mid-Holocene C/N ratios and TOC in these oligotrophic lakes may reflect low nitrate concentration. Low nitrate is presently observed in numerous Faroese lakes (Jensen et al., 2002; McGowan et al., 2008).

Because the increasing trend in C/N values is subtle and their values...
are around 12–15 the $\delta^{13}C$ trend towards lighter values may be interpreted as weakened dominance of atmospheric CO$_2$ and enhanced development of soils providing $^{13}C$ depleted DIC to the lakes in concord with increased vegetation cover (Jóhannsen, 1985).

Altogether these findings point towards a milder and drier climate with high terrestrial and aquatic organic productivity, a situation which may be associated with a fairly early Holocene Climatic Optimum (Mayewski et al., 2004; Roncaglia, 2004; Andresen et al., 2006; Wanner et al., 2008; Langdon et al., 2010). A warm phase characterised by greater seasonality is also recorded in the ocean off the Faroe Island (Rasmussen and Thomsen, in press). This is broadly consistent with recent GCM model experiments showing a time transgressive Holocene Climatic Optimum. The transgressive optimum was associated with atmospheric and oceanic circulation change coupled to the final melting of the Laurentide ice sheet at 7000 cal yr BP and indicates that not before 7000 cal yr BP did summer temperatures closely follow summer insolation (Debrez et al., 2009; Renssen et al., 2009).

The higher variability of weathering products ($\chi$ and Ti, in particular in zone 7) observed in the highland lakes Brúnavatn and Stóravatn possibly indicates a much slower and gradual development of the vegetation cover at higher elevations relative to the valley lakes. In this the stabilised TOC values of Brúnavatn and Stóravatn at c 7 and 15%, respectively, in zone 7 (c 7410), suggesting the onset of colder climate conditions affecting elevated areas more strongly (Fig. 7). This may point towards colder winter conditions resulting in prolonged snow cover and increased erosion at higher elevations. In contrast with this are increased terrestrial influence observed in sediments from Skálafjörður (Roncaglia, 2004). As stated above it may also be noteworthy that the sedimentation pattern changed in Brúnavatn at the onset of zone 7, indicating a changing wind climate towards stronger winds as supported by the high TS values and low C/S in Stóravatn (zone 7, Fig. 7), i.e. surplus of sulphur deposition due to increased sea spray. The onset of a mid-Holocene cooling is also observed in other archives. In Skálafjörður lower temperatures are recorded by faunal proxies from c 7500 cal yr BP in concord with reports of cold arctic Norwegian water masses off the Faroe Islands and in the ocean east of Iceland (Roncaglia, 2004; Witon et al., 2006; Thorvald et al., 2009; Rasmussen and Thomsen, in press).

These trends may be due to a decrease in North Atlantic Deep-Water (NADW) as suggested by frequent cold events observed in Irish speleothems between c 8000 and 4200 cal yr BP (McDermott et al., 2001). Another link could be to a southeastward movement of the Iceland–Faroe front with increased influence of polar water as indicated by a distinct decrease in salinity between 8200 and 6800 cal yr BP off Faroe Islands (Rasmussen and Thomsen, in press).

5.2.2. A response to the 8.2 kyr event?

Clear evidence of the 8.2 kyr event (Alley et al., 1997) has not previously been found on the Faroe Islands (Andresen and Björck, 2005; Andresen et al., 2006). In Mjáuvót a pronounced decrease in Ti counts occurs at 8330 ± 400 cal yr BP (Figs. 7 and 9). This could imply that decreased erosion rates characterise this event on the Faroe Islands, and indicates that drier conditions may have prevailed (Fig. 7). This is consistent with a southward displacement of the polar front as indicated in Greenland ice cores (O’Brian et al., 1995; Thomas et al., 2007), drier conditions and prolonged winters north of 50°N due to reduced Atlantic Meridional Overturning Circulation (Magny et al., 2003; Seppa et al., 2007; Ljung et al., 2008; Prasad et al., 2008), but in contrast to increased precipitation and soil erosion as reported from Southern Scandinavia (Hammarlund et al., 2005; Hede et al., in press). One reason that this climate event has not clearly shown up elsewhere on the Faroe Islands may be due to the often low deposition rates of that time period (Fig. 3A), but the lack of evidence may also be related to the fact that the Faroe landscape was less sensitive to such a brief cooling event within an already fairly cool region.

5.2.3. Zone 8 (4190–3000 cal yr BP), onset of the Neoglacialation

A pronounced climate change around 4000 cal yr BP has been reported from numerous proxy types around the world (Edwards et al., 1996; Hammarlund et al., 2002; Olafsdottir and Guomundsson, 2002; Came et al., 2007; Olsen et al., 2010; Solignac et al., in press). On the Faroe Islands, increased mid-Holocene erosion rates have been reported, possibly related to gully erosion of peats, which focussed the erosion and led to higher sedimentation (Humlum and Christiansen, 1998; Edwards, 2005; Lawson et al., 2007). Furthermore, foraminiferal data from the eastern Faroese shelf indicate colder sea-surface temperatures from c 4000 cal yr BP and onwards (Rasmussen and Thomsen, in press). The pollen data presented by Hamon et al. (2001) show that a major change took place above the Hekla 4 ash layer, indicating encroachment of shrub woodland and heath communities, and Lawson et al. (2008) report that the Gróthúsvatn pollen record indicates similar changes. In the lakes studied here, a return at 4190 cal yr BP to higher accumulation rates of 3.4, 0.3 and 1.7 mm/yr for Mjáuvót, Brúnavatn and Stóravatn, respectively, is also observed (Fig. 3A). These higher accumulation rates may possibly be interpreted as a significant increase in catchments erosion and thus as a climate deterioration (Andresen et al., 2006; Lawson et al., 2008).

In contrast, Zone 8 (4190 cal yr BP) shows a decreasing trend of $\chi$ and Ti in all three lakes indicating decreasing catchment erosion and likely warmer or drier climate conditions (Fig. 7). Similarly, the $\delta^{13}C$ values continue to decrease in all three cores with increasing C/N ratios suggesting an increased fraction of terrestrial derived OM in the sediments. This could point towards increased soil instability due to increased influence of e.g. freeze/thaw cycles and thus colder climate (Fig. 7). This is particularly valid for Brúnavatn where the strongly decreasing $\delta^{13}C$ values and increasing C/N ratios at the top of zone 8 clearly suggest an increased influx of terrestrial OM to the sediments (Fig. 7).

5.3. Late-Holocene (3000 cal yr BP—present)

5.3.1. Zones 9–10 (3000–1360 cal yr BP)

The period from c 3000 cal yr BP marks the onset of a further cooling and increased climate instability on the northern hemisphere presumably related to a weakening of AMOC (Köc et al., 1993; Eiriksson et al., 2000; Jiang et al., 2002; Andresen et al., 2004; Kaplan and Wolfe, 2006; Seidenkrantz et al., 2008; Ren et al., 2009; Olsen et al., 2010).

The further decreasing trends of $\chi$ and Ti in all three records indicate decreasing catchment erosion and likely warmer or drier climate conditions (Figs. 8 and 9). The high C/N ratios and low $\delta^{13}C$ values of the late-Holocene show that terrestrial OM dominates the sediments when compared to global values, in particular zone 10 with high C/N ratios (>20) and $\delta^{13}C$ values around ~27‰ (Figs. 5 and 8). Only Stóravatn zone 9 indicates a mixed OM origin (Figs. 5 and 8). Combined with the increasing TOC trend of Mjáuvót and Stóravatn this suggests an increased fraction of soil-derived OM in the sediments and thus continued soil instability as observed in the late mid-Holocene (Figs. 7 and 8). This may have been caused by wetter conditions and/or colder climates resulting in soil instability due to freeze/thawing processes. Overall, taking the Ti and $\chi$ in account the climate was likely drier and colder. Alternatively, increased wind activity may have increased the input of terrestrial OM to the lakes (Andresen et al., 2004; Solignac et al., in press).

The increased Ti counts and $\chi$ values between 1850 and 1750 cal yr BP (zone 10, Fig. 8) in Mjáuvót and Stóravatn suggest
an increase in catchment erosion, perhaps a short period with increased precipitation, in concert with increased soil erosion recorded in Lake Hvannavatn (Suduroy), Heygsvatn and Litlavatn (Sandoy) (Graeuert, 2005; Lawson et al., 2007, 2008). Taking into account the warmer SST in Skálafjord (Roncaglia, 2004), this suggests a short warm and wet period around 1800 cal yr BP.

The TOC of Brúnavatn rapidly increase at the onset of zone 9 (c 3000 cal yr BP) towards 40% combined with C/N ratios above 100, which together with constant δ13C values of c −27‰, suggest that the sedimentation of OM is more or less limited to a terrestrial origin (Fig. 8). This development is unique to Brúnavatn and likely reflects an ontogenetic change. The likely mechanism behind this distinct change is the onset of peat overgrowth in the north–eastern part of the lake around the main inlet as seen today and is thus attributed to a local lake specific change in Brúnavatn (Fig. 1). This may explain the very constant and low δ13C values as well as the very high C/N ratios, both due to direct sedimentation of in-washed humic matter from the surrounding peat land and also by modifying the DIC carbon isotope composition (Fig. 8). According to Lawson et al. (2007) the initiation of blanket peat cannot be related to climate or anthropogenic impacts on the Faroe Islands explaining the uniqueness of the Brúnavatn development when compared with Mjáuvötn and Stóravatn.

5.3.2. Zones 11–13 (1360 cal yr BP–present), anthropogenic impacts?

The period 1360–725 cal yr BP (zone 11) is characterised by increased Ti counts and δ values suggesting enhanced weathering in the Mjáuvötn catchment (Figs. 8 and 9). The OM content decreases (c 8–4%) and the decreasing C/N ratios and δ13C values of around −26.5‰ indicate a reduced input of terrestrial OM to the Mjáuvötn sediments (Fig. 8). A similar general pattern is found in Brúnavatn, while the opposite is recorded by the Stóravatn sediments. At that site decreased minerogenic content and a significant increase in OM content, together with increasing C/N ratios and δ13C values approaching −28‰, suggest a very high terrestrial OM fraction (Fig. 8). Overall, this leaves the impression of a period with significant mineral erosion in a wetter or colder climate in contrast to the warmer climate generally found during the Medieval Warm Period (Jiang et al., 2002; Geirsdóttir et al., 2009; Mann et al., 2009). Wetter conditions may have been linked to an enhanced NAD strength of the AMOC (cf Jessen et al., 2008). From the basal ages of lake sediments it appears that the Faroe Islands had been deglaciated by the end of Younger Dryas (cf Humlum et al., 1996, Humlum and Christiansen, 1998). The early-Holocene sparse vegetation cover accompanied relatively high sedimentation rates with high bulk OM δ13C indicating poor soil development. The inferred highly fertile climate was perhaps a consequence of recurring meltwater outbursts around the North Atlantic resulting in repeated disturbances of the AMOC (cf Jessen et al., 2008).

δ13C, Ti and δ values data reveal a much more stable and warm mid-Holocene characterised by an inferred increasing vegetation cover and build up of organic soils towards the Holocene thermal maximum on the Faroe Islands around 7400 cal yr BP. The final meltdown of the Laurentide ice sheet around 7000 cal yr BP appears to have impacted both ocean and atmospheric circulation towards colder conditions on the Faroe Islands as suggested by enhanced weathering and increased deposition of surplus sulphur (sea spray) and erosion in the highland lakes from about 7400 cal yr BP. As a consequence we see indications of more severe and prolonged winter conditions at higher elevations.

Clear evidence of the 8.2 kyr event (Alley et al., 1997) is not found in our data although Mjáuvötn shows a pronounced decrease in Ti counts at 8330 ± 400 cal yr BP. This could imply that decreased erosion rates characterise this event on the Faroe Islands in which case drier conditions may have prevailed. This is consistent with a southward displacement of the polar front as indicated in Greenland ice cores (O’Brien et al., 1995; Thomas et al., 2007) and drier conditions and prolonged winters north of 50°N due to reduced Atlantic Meridional Overturning Circulation (Magny et al., 2003; Seppa et al., 2007; Ljung et al., 2008; Prasad et al., 2009).
From 4190 cal yr BP further cooling is believed to have occurred as judged by increased soil erosion. This change may have been caused by an increase in freeze/thaw sequences related to oceanic and atmospheric variability. A cooling trend appears to further advance from 3000 cal yr BP. A short period around 1800 emerges as a warm and wet phase in between a general cooling characterised by significant soil erosion lasting until 725 cal yr BP. Interestingly, increased soil erosion seems to begin already by 1360 cal yr BP, significantly before the arrival of the first settlers on the Faroe Island around 1150 cal yr BP. There are indeed indications of additional erosion taking place around 1200 cal yr BP which may have been a direct consequence of human activities. Hence it may be hypothesized that human impact on the landscape following the landnam merely accelerated an erosional process that had already started before this event. Alternatively, the increased soil erosion could be interpreted as an effect of anthropogenic activity, implying that the Faroe Islands were inhabited already by 1360 cal yr BP as suggested by some studies. Soil erosion was a dominant landscape factor during the Little Ice Age, but climate related triggers cannot easily be distinguished from human activities.

The three investigated lakes (Mjávøtn, Stórarvatn and Brúnavatn) are situated fairly close to each other (within 20 km), but are different hydrologically and limnologically (catchments area, water depth and elevation). Apart from the fact that general trends are fairly similar throughout the Holocene they show different variations. In particular, Brúnavatn shows a unique development after c 3000 cal yr BP. This may presumably be attributed to local entangling climate variability from lake sediments and multiple catchment processes not determined by external factors, i.e. climate. This emphasises the care that must be taken when distinguishing climate variability from lake sediments and multiple sites are always advisable.

Acknowledgements

We are grateful to Rineke Gieles and Thomas Richter (Royal Netherlands Institute for Sea Research, Texel) who assisted with XRF-scanning. We would also like to thank Inge Juul (Department of Geography and Geology, University of Copenhagen) for TC, TS and CaCO3 measurements. We are also very grateful to the COWI financing the fieldwork on the Faroe Islands. Jeppe Joel Larsen (Department of Geography and Geology, University of Copenhagen) and Palle Lindegård (Department of Earth Sciences, Aarhus) are thanked for their hard work and assistance with coring the three lakes. Jesper Olsen would like to thank the Carlsberg Foundation for financial support.

Appendix. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.quascirev.2010.06.029.

References


