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Paleohydrology of Kangerlussuaq (Søndre Strømfjord), West Greenland during the last ~8,000 years

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Abstract

Major fluctuations of hydroclimate in West Greenland are recorded in paleoshoreline terraces that encircle several lakes near Kangerlussuaq, Greenland (67°01’N, 50°40’W). Geomorphic and stratigraphic analyses were used to construct a lake-level curve for Hunde Sø, a large closed-basin lake in this region. Changes in lake volume associated with lake-level fluctuations were calculated, and a water-balance model was used to determine the primary factors influencing lake volume and the changes in those factors necessary to affect reconstructed lake-level change. Sensitivity tests suggest that mean annual precipitation and the relative proportion of summer versus winter precipitation are the primary climate drivers of lake-level change. Temperature effects are less important unless coupled with changes in one of the other variables. Two pluvial periods occurred centered around 4600 and 2000 cal. yr BP, with mean annual precipitation estimated at 130 mm and 70 mm higher than modern, respectively. In contrast, a low-stand prior to 6070 cal. yr BP with water levels as much as 18 m below modern was a result of summer temperatures ~2–3°C above modern, coupled with a 14 day increase in the length of the ice-off period and a reduction in mean annual precipitation of ~80 mm relative to modern.

Keywords: paleohydrology, Greenland, Holocene, pluvial, paleoclimate, geomorphology, lake shorelines, lake-level variations, water balance

Introduction

Hydrology in Greenland is strongly linked to atmospheric and oceanic processes. Runoff and ice-discharge from the Arctic’s largest island impact the temperature and salinity of the North Atlantic Ocean through interactions with the West Greenland Current, the East Greenland Current and the Irminger Current (Moros et al., 2006). In turn, sea surface temperatures and ice cover influence both local and hemispheric atmospheric processes, including circulation patterns, storm tracks and modes of the North Atlantic Oscillation (Cassou et al., 2004; Deser et al., 2004; Magnusdottir et al., 2004). Because of these significant atmosphere–ocean feedback mechanisms, the hydrological system of Greenland is both a result and a source of variability in large-scale atmospheric processes that affect the climate throughout the Northern Hemisphere. The historic record of climate and weather in the North Atlantic region has been studied extensively by both the modeling and observation-based scientific community. However, because of the relatively short period of record, our knowledge of centennial- to millennial-scale variability has come from investigations of climate proxy records, the most well-known being the ice cores. Oxygen isotope and ice accumulation records from the ice cores form an important record of temperature and precipitation for the North Atlantic region but are subject to local errors (i.e., drifting) and don’t necessarily reflect conditions along the margins of the ice sheet. Further, although the ice accumulation records display high decadal to centennial variability in the mid to late Holocene, large longer-term trends are not obvious, in contrast with the record of closed-basin lakes at the margins of the ice sheet, which indicate significant hydrologic change.

The ice-free region near Kangerlussuaq (Figure 1) contains hundreds of thousands of lakes. Many of these are close-basin lakes, making them ideal for the retrieval and study of paleohydrologic proxies from sediment cores. Studies of regional lacustrine proxy records indicate a highly variable mid- to late Holocene climate, but there are discrepancies among the records and proxies. For example, McGowan et al. (2003) used diatom-inferred (DI) conductivity to reconstruct effective moisture from two lakes (Braya Sø and SS6) near Kangerlussuaq (Figure 2), based on the premise that conductivity in a closed-basin lake is a proxy for change in precipitation minus evaporation (P − E). Dissolution of the diatom record from 6,800 to 5,300 cal. yr BP suggests high conductivity and low effective moisture, followed by a continuous decline in DI conductivity to minimum values at ~4,700 cal. yr BP, with low (<400 µS/cm) values persisting un-
Anderson and Leng (2004) measured stable isotopes of carbonates in these same lakes (Figure 2). The interpretation of the δ¹⁸O record for SS6 largely agrees with the diatom records prior to ~4,000 cal. yr BP, in that higher δ¹⁸O values indicate evaporative enrichment between ~7,000 and 5600 cal. yr BP, followed by a wetter period inferred from depleted values between ~5,600 and 4,000 cal. yr BP. The δ¹⁸O for Braya Sø, however, only shows a slight trend towards enrichment with two large excursions. After 4,000 cal. yr BP, the diatom and δ¹⁸O records are difficult to reconcile. The DI conductivity curve for SS6 displays large variations on the decadal to millennial scale, ending with a trend of decreased conductivity from ~1,500 years to present. In contrast, the DI conductivity reconstructions for Braya Sø show much smaller-scale fluctuations and remain fairly constant over the last ~1,500 years. The δ¹⁸O for both lakes shows very little variability and a slight trend towards δ¹⁸O enrichment.

These studies indicate that moisture balance in this area was highly variable throughout the Holocene, but the magnitude of variation, the pattern of late-Holocene change, and the climatic parameters responsible for inferred variability are unclear. Furthermore, lacking a direct transfer function to translate DI conductivity change into lake volume change or δ¹⁸O to P-E change, the source(s) of the discrepancies among the records cannot be easily constrained. Quantitative changes in lake elevation and volume can be reconstructed from geomorphic evidence along with detailed mapping of the catchment and lake; these data can then be used in water-balance models to estimate changes in precipitation and to investigate the factors influencing evaporation (Hostetler and Benson, 1990).

The "salt lakes" Hunde Sø, Braya Sø, and Limneæa Sø near the head of Søndre Strømfjord are well known (e.g., Böcher, 1949; Williams, 1991). The paleoshorelines that surround these lakes have been cited as evidence for changes in regional effective moisture (Anderson et al., 2001; Anderson and Leng, 2004) but have never been investigated in detail. This study documents the timing and magnitude of lake-level fluctuations recorded by these shorelines and attempts to quantify the climatic parameters responsible for hydrologic change. A lake-level curve is developed for Hunde Sø, and a water balance model is used to test the sensitivity of the lake to changes in regional climatic parameters. The model provides a more quantitative understanding of the parameters that likely resulted in the lake-level fluctuations, as well as the magnitude of those changes.

Figure 1. Study location of the Kellyville lake basins showing the major features mentioned in the text. Triangles mark locations of trenches and exposures (NSE, North Shore Exposure; HT Cove, Hunde Sø Terrace Cove; B, Braya Sø; WB, West Braya Sø; LB, Land Bridge: CI = 25 m).
Site location and climate

The study basins are located approximately 15 km west-northwest of Kangerlussuaq, Greenland (67°01′N, 50°40′W), just northwest of Kellyville (Figure 1). The three larger lakes of the Kellyville basin are Hunde Sø (SS3), Braya Sø (SS4), and Limnæa Sø (SS5); however, SS6, Upper Braya (SS85) and two small unnamed lakes to the northeast (NE Lake) and east (East Lake) of Hunde Sø are of interest, as they are all hydrologically connected or have been in the past. With the exception of Upper Braya and NE Lake, these lakes are oligosaline, with conductivities that range from 3,250 to 4,910 µS/cm. Upper Braya and NE Lake seasonally drain into the larger lakes, flushing salts and maintaining lower conductivities (380–600 µS/cm). All the lakes are above the marine limit (~ 100 m above modern sea level), and thus the origin of the salinity is not from the capture of seawater. The modern elevations (above modern sea level) of Hunde Sø, Braya Sø, and Limnæa Sø are 170, 173, and 173.5 m, respectively, with a modern bench between 2 m and 3 m deep. The sill for the system of lakes is located on the west side of Limnæa Sø at an elevation of 186 m, where overflow passes through an elongate lake into Søndre Strømfjord. The elevation of the land-bridges separating Hunde Sø from Limnæa Sø and Hunde Sø from Braya Sø are 174.5 m and 173.9 m above sea level, respectively.

The modern climate of Kangerlussuaq is summarized in Figure 3. Mean annual precipitation (MAP) between 1949 and 2003 is 158.8 mm, and calculated rates of evapotranspiration are ~300 mm/yr, most of which occurs between June and August (Hasholt and Sogaard, 1978). Monthly precipitation data show that the majority of annual precipitation is received between May and September. Mean annual temperature is -5.1°C, and mean temperature for June through August (JJA) is 9.4°C. Wind direction is dominantly from the east as katabatic winds are funneled through the fjord, but strong westerly winds also are produced by the onshore movement of storms from the Atlantic Ocean. The mean wind speed between 1985 and 1999 at Kangerlussuaq is 3.6 m/s.

Shoreline stratigraphy and lake-level reconstruction

To reconstruct the magnitude and timing of lake-level fluctuations, several paleoshorelines around Hunde Sø, Braya Sø, and Limnæa Sø (Figure 1) were surveyed and excavated. Detailed topographic mapping, stratigraphy, and ground penetrating radar (GPR) profiling of the shorelines were acquired during the summers of 2002–2004. Seismic reflection data from Braya Sø was also acquired. These have been interpreted within the context of samples collected from depositional environments of the modern littoral zones, three- dimensional terrace geometry, and well-established principles of transgressive-regressive cycles and their associated stratigraphic relationships (see Posamentier and James, 1993). Chronologic control was obtained by radiocarbon dating of terrestrial wood, where available, and bulk organic material (Table 1). A complete description of the stratigraphy, GPR, and seismic data can be found in Aebly (2007); the stratigraphy of NSE and HT Cove is summarized here as it pertains to lake-level changes.

When reconstructing lake levels using paleoshoreline stratigraphy, the elevation of the deposit represents a minimum elevation depending on the depositional depth, as well as lake geometry (e.g., fetch and bathymetry) and meteorological conditions. Although it is impossible to know the meteorological conditions at the time of deposition, the depth can be constrained by analysis of modern sediments in a similar location. For the current study, modern sediments were collected at several depths just to the east of NSE and just offshore from HT Cove. The modern beach along the north shore is a mixed sand and gravel beach with a coarse swash-zone and foreshore, a finer-grained shoreface, and an offshore transition (Figure 4). Medium to coarse sands and gravels with sporadic irregular, asymmetrical ripples dominate the foreshore area. Beyond a depth of ~ 0.5 m, ripples become more symmetrical and continuous in the shore-parallel direction, and percent gravel drops sharply to less than 1% and is seen only in ripple troughs. Although preservation of bedforms is expected to be very low in this environment owing to low sediment supply and accommodation, the modern beach conforms to models of wave-dominated, low to medium energy, mixed sand and gravel coasts (e.g., Elliot, 1986; Hart and Plint, 1995). Therefore, these models can be applied to the north shore.
Figure 3. Modern climate parameters at Kangerlussuaq, Greenland: (a) monthly mean precipitation (1948–2003); (b) monthly mean temperature (1948–2003); (c) distribution of summer precipitation (proportion of MAP that occurs from July through August; \( P_s \)); and (d) distribution of annual precipitation. (Sources: 1976–1999 Danish Meteorological Institute (DMI; Cappelen et al., 2001), 1949–2003 Global Historical Climatology Network, Version 2 (GHCN V2, http://www.ncdc.noaa.gov)

Table 1. Radiocarbon AMS dates and lab-reported date error (National Ocean Sciences Accelerator Mass Spectrometry; NOSAMS) for samples used in the lake-level reconstruction

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>NOSAMS accession no.</th>
<th>Elevation of sample (m a.s.l.)</th>
<th>Age (^{14} \text{C yr BP} )</th>
<th>Age (cal. Yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HT12</td>
<td>OS-38397</td>
<td>171.29</td>
<td>1150 ± 50</td>
<td>981 (1065) 1167</td>
</tr>
<tr>
<td>HT17</td>
<td>OS-38398</td>
<td>171.5</td>
<td>1730 ± 45</td>
<td>1570 (1642) 1699</td>
</tr>
<tr>
<td>HT19</td>
<td>OS-38399</td>
<td>171.55</td>
<td>1620 ± 40</td>
<td>1418 (1504) 1555</td>
</tr>
<tr>
<td>HT19W*</td>
<td>OS-38399</td>
<td>171.55</td>
<td>1210 ± 30</td>
<td>1081 (1134) 1174</td>
</tr>
<tr>
<td>HT111</td>
<td>OS-38401</td>
<td>171.6</td>
<td>1080 ± 30</td>
<td>939 (1051) 987</td>
</tr>
<tr>
<td>HT23</td>
<td>OS-38425</td>
<td>173.6</td>
<td>2470 ± 35</td>
<td>2468 (2561) 2617</td>
</tr>
<tr>
<td>HT31*</td>
<td>OS-38402</td>
<td>174.67</td>
<td>2260 ± 35</td>
<td>2183 (2237) 2339</td>
</tr>
<tr>
<td>HT33</td>
<td>OS-38403</td>
<td>174.81</td>
<td>1950 ± 35</td>
<td>1868 (1901) 1945</td>
</tr>
<tr>
<td>HT36B*</td>
<td>OS-38404</td>
<td>174.96</td>
<td>810 ± 25</td>
<td>694 (717) 732</td>
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<tr>
<td>HT37</td>
<td>OS-38410</td>
<td>174.99</td>
<td>635 ± 82</td>
<td>553 (608) 665</td>
</tr>
<tr>
<td>NSE14*</td>
<td>OS-42034</td>
<td>170.77</td>
<td>5300 ± 35</td>
<td>5999 (6070) 6177</td>
</tr>
<tr>
<td>NSE12*</td>
<td>OS-42033</td>
<td>171.26</td>
<td>2220 ± 35</td>
<td>2156 (2231) 2310</td>
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<tr>
<td>NSE13*</td>
<td>OS-42032</td>
<td>170.98</td>
<td>4980 ± 35</td>
<td>5658 (5700) 5735</td>
</tr>
<tr>
<td>LE6</td>
<td>OS-42075</td>
<td>175.71</td>
<td>3220 ± 35</td>
<td>3396 (3429) 3468</td>
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<tr>
<td>LE3*</td>
<td>OS-42074</td>
<td>174.04</td>
<td>1160 ± 35</td>
<td>992 (1069) 1168</td>
</tr>
<tr>
<td>LE10*</td>
<td>OS-42073</td>
<td>175.07</td>
<td>5750 ± 60</td>
<td>6454 (6547) 6638</td>
</tr>
<tr>
<td>LE8*</td>
<td>OS-42072</td>
<td>176.61</td>
<td>3350 ± 35</td>
<td>3487 (3581) 3636</td>
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<tr>
<td>WB6</td>
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<td>174.26</td>
<td>4180 ± 50</td>
<td>4672 (4706) 4829</td>
</tr>
<tr>
<td>WB9</td>
<td>OS-42070</td>
<td>174.86</td>
<td>2330 ± 35</td>
<td>2214 (2342) 2358</td>
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<tr>
<td>WB14*</td>
<td>OS-42069</td>
<td>176.2</td>
<td>840 ± 25</td>
<td>707 (738) 786</td>
</tr>
<tr>
<td>WB1*</td>
<td>OS-42037</td>
<td>173.45</td>
<td>2200 ± 25</td>
<td>2152 (2236) 2305</td>
</tr>
<tr>
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<td>OS-42036</td>
<td>173.67</td>
<td>1250 ± 30</td>
<td>1147 (1195) 1260</td>
</tr>
<tr>
<td>B1*</td>
<td>OS-42035</td>
<td>174.69</td>
<td>2220 ± 30</td>
<td>2157 (2230) 2309</td>
</tr>
</tbody>
</table>

Calibrated dates were obtained using Calib 5.0.2 (Stuiver and Reimer, 1993). The range of calibrated dates is given with the median probability date in parentheses. Dates quoted in the text and used to reconstruct the lake-level curve are the median probability dates rounded to the nearest decade. Sample IDs for the HT Cove and West Braya Se (HT and WB, respectively) are numerically numbered following the area and terrace prefix (i.e., HT23 is the third sample from HT2). An asterisk * denotes wood, all others are bulk organic material.
PALEOHYDROLOGY OF WEST GREENLAND DURING THE LAST ~8,000 YEARS

Area. Observations of the lower shoreface and offshore transition were not possible, but models predict parallel laminated sands and silts with storm-induced bedding, such as hummocky cross-stratification (HCS) or swaley cross-stratification (SCS), also possible because of the frequent occurrence of high-wind storms that affect the area (Elliot, 1986; Walker and Plint, 1992).

In HT Cove, modern grab samples were dominated by organic material with little silts and clays. Minor occurrences of fine sand are observed at the very edge of the shoreline at depths less than 0.5 m. The geometry of the cove results in a fetch that is essentially zero from the west, and a peninsula ~300 m to the east shelters the cove from easterly winds. Because of the heavy cover of vegetation, modern runoff to the cove is extremely limited and clastic input is dominated by aeolian deposition.

Stratigraphy of NSE

The North Shore of Hunde Sø is currently eroding from high-energy wave action. Erosion has exposed ~1 m of sediment along the shoreline, which was excavated back and down about ~0.5 m to expose the sediments and structures (Figure 5). Only one section is described here, but all facies and depositional environments can be traced laterally and correlated through three sections spanning the north shore of Hunde Sø (Aebly, 2007). At the base of the section, coarse sands and gravels with trough cross-stratification and some scour and fill structures indicate a fluvial to fluvial deltaic environment. This facies is laterally extensive across the north shore until a buried moraine is reached to the east, which separates the NSE sites from a small beach in the northeast corner of the lake. A valley immediately northwest of the NSE site shows evidence of significant fluvial activity, including a dry waterfall with plunge pool. This valley was likely the source of the fluvial sediments. The absence of this facies east of the moraine and the calibrated age of 6,070 yr BP both indicate that the fluvial system was postglacial and likely a result of climatically forced changes in lake level and runoff. Above this facies, foreshore (FS) to upper shoreface (USF) delta front facies overlie by offshore (OS) silts indicates a rise in lake level. As the foreshore migrated landward, wave action eroded the underlying sediments producing a gravel lag, and ultimately the OS silts and clays were deposited, representing the maximum flooding surface. The source of the gravel within the lag deposit was most likely the underlying FS/USF delta front sediments. This is comparable with the type-1 lag deposit described for marine sediments by Van Wagoner et al. (1990) and represents a transgressive lag. The OS facies are overlie by offshore transition (OST) facies, which are then truncated by FS/USF facies of gravel and sand, representing declining lake levels. A second OST facies overlying the FS/USF facies is dated at 2,230 cal. yr BP and denotes a second rise in lake level. A shoaling upwards through middle to lower shoreface (M-LSF) deposits completes the section. At the top of NSE, facies Sp is capped by a thin (~8 cm) veneer of massive sand and gravel. Bioturbation, aeolian reworking, and compaction have destroyed any structures that may have been present, but this is likely a swash zone deposit.

The lack of channel incision or fluvial deposits younger than the basal sediments on the north shore is evidence that the modern lakes are the lowest they have been since the lowstand prior to ~6,070 cal. yr BP. Although the elevation of this lowstand cannot be directly determined for Hunde Sø, acoustic subsurface profiles of Braya Sø show seismic facies indicative of drowned beach ridges approximately 11 m below the modern surface (Figure 6). The beach ridges are stratigraphically located at the base of the Holocene seismic section indicating an age similar to the fluvial sediments at NSE. The lake experienced two major periods of lake-level rise; the highest was recorded at 5,700 cal. yr BP. Between these two transgressions, a forced regression is evident in a shoaling of the sediment until the truncation of the OST facies by USF to FS facies.
As defined by VanWagoner et al. (1990: figure 5, example 1), three parasequences are recognized: a lower parasequence (PS I), bounded on the bottom by deltaic sediments and on top by a flooding surface; a middle parasequence (PS II), bounded on the bottom and top by flooding surfaces; and an upper parasequence (PS III), bounded on the bottom by a flooding surface and on the top by the modern soil. These parasequences form the basis of the climate zones discussed later.

**Facies Codes**

- **Gh** - Massive to faintly laminated clast-supported gravels
- **Gp** - Planar cross-beded gravels and sands.
- **St** - Coarse sands and gravels; trough cross stratification; scour and fill
- **Sm** - Massive coarse sands and gravels
- **She** - Horizontal to wavy laminated fine sands and coarse silts; interbeds of organic detritus; some cross bedding.
- **Snr** - Normally graded fine sands and coarse silts with minor occurrences ripple cross-lamination.
- **Si** - Inversely graded fine sands and coarse silts
- **Sp** - Planar cross bedded sands with some gravel
- **Sr** - Rippled sands with some gravel; ripple-scale hummocky cross-stratification in HT3.
- **Fl** - Massive to faintly laminated fine sands and silts
- **Fm** - Massive to faintly laminated fine silts and clays; dominated by organic material in some places

**Figure 5.** Photograph and graphic log of North Shore Exposure (NSE) with facies codes and depositional environments. Dashed lines indicate flooding surface separating PS1, PS2, and PS3. Dates shown (bold) are in cal. yr BP, with * indicating dates based on wood. USF, upper shoreface; M-LSF, middle to lower; OST, offshore transition; FS, foreshore (FS).

**Stratigraphy of HT Cove**

Three obvious terraces are located in the HT Cove area (Figure 7). Trenches were excavated to the depth of permafrost (~60 cm) in all three. Readily apparent is the lack of any dates prior to 2,560 cal. yr BP. However, analysis of GPR radar facies indicates that an older erosion surface is present beneath the permafrost in HT3 (Aebly, 2007), which may be related to the earlier
Paleohydrology of West Greenland during the last ~8,000 years

transgressive lag described for NSE. In addition to the facies described for NSE, two additional facies are found at HT Cove. Sr is present only at the base of the top terrace (HT3) of the HT Cove site. It is composed of fine-grained sands with ripple cross-lamination and ripple-scale hummocky cross-stratification (HCS) similar in scale to those discussed by Yang et al. (2006). Laboratory experiments by Dumas et al. (2005) indicate that ripple-scale HCS is a transitional feature between symmetrical and asymmetrical ripples with increasing unidirectional flow velocities. The appearance of the ripple-scale HCS in HT3 alongside asymmetrical ripple cross-lamination may be analogous and suggests an increase in the longshore current.

Fl forms the base of HT2 and consists of massive to faintly laminated fine sands and silts. The base of the unit lies within the permafrost and could not be excavated. The sediment dynamics in this cove are highly variable, with changes in fetch occurring as water levels fluctuate above and below the landbridge to the west and the peninsula to the east. As a result, the determination of the depositional environment is difficult, but the similarity in dates between the top of Fl in HT2 (2,560 cal. yr BP) and Sr at the base of HT3 (2,240 cal. yr BP) suggests coeval deposition.

Lake-level reconstruction

Using the stratigraphy of the shorelines and the calibrated 14C dates, a lake-level curve was developed for Hunde Sø (Figure 8). Basal dates from sediment cores of Braya Sø and SS6 (McGowan et al., 2003) and the glacial history of the area (van Tatenhove et al., 1996) show that glacial drainage into these lakes stopped ~8,000 cal. yr BP. While receiving glacial runoff, all of the lakes presumably were connected, and water levels were at the sill elevation of 186 m. Field measurements with a Global Positioning System (GPS) confirmed old lake sediments at an elevation of 185 m, and the sill shows evidence of past erosion and deposition.

Prior to 6,070 cal. yr BP, lake levels had dropped to below modern, as evidenced from the fluvial/deltaic parasequence boundary. The transgressive lag and overlying OS sediments are dated at 5,700 cal. yr BP. Using a depositional depth of at least 3 m for OS silts and clays, the lake-elevation is estimated

Figure 6. Portions of three subsurface acoustic profiles from Braya Sø showing drowned beach ridges at a depth of approximately 11 m.

Figure 7. Photograph of Hunde Sø Cove showing the locations of HT1–HT3 and graphic logs of the Hunde Sø Cove trenches. Shoreline profile shows vertical and lateral spatial relationships. Dates shown (bold) are in cal. yr BP, with * indicating dates based on wood. Solid line on profile shows elevation at which Hunde Sø and Limnaea Sø are connected.
to be at least 174 m. However, facies Fl at the base of a trench in WB dated at 4,700 cal. yr BP at an elevation of 174.3 m, and planar laminated sands from Limnæa Sø (LE), dated at 3,580 cal. yr BP at an elevation of 176.6 m, indicate that Hunde Sø continued to rise until the lakes were connected at an elevation of approximately 177 m (see Table 1).

A lowering of lake level between 3,580 and 2,560 cal. yr BP is interpreted from the FS/USF sediments at the top of PS II. Depositional depth is estimated from modern sediments to be about 1 m resulting in a lake elevation of ~172 m. This is also supported by GPR data from this site that contain radar facies characteristic of foreshore deposits at an elevation of approximately 172 m, which are stratigraphically correlated with the top of PS II (Aebly, 2007). The timing of the lowering is constrained between 3,580 and 2,560 cal. yr BP using the highstand dates from LE, the flooding surface that forms the upper boundary of PSII dated at 2,230 cal. yr BP, and the dates at the bases of HT2 and HT3, which places the lake at an elevation above 175.4 m by 2,560 cal. yr BP. This is the elevation at which Hunde Sø connects with Limnæa Sø; this connection, and the resulting increase in fetch, is also supported by GPR data from this site that contain radar facies characteristic of foreshore deposits at an elevation of approximately 172 m, which are stratigraphically correlated with the top of PSII (Aebly, 2007).

The group of dates between 990 and 1,640 cal. yr BP at elevations around 171.5 m indicate a lowering of the lake surface to near modern levels, however, there is no evidence in the stratigraphy or the GPR profiles to support any subaerial exposure (i.e., soil development) or erosion of the shorelines prior to the higher levels recorded between 610 and 720 cal. yr BP. It should also be noted that these dates are all from HT1, which contains a chronological reversal and could have been contaminated by reworking of sediments from higher up the shoreline.

**Model development**

To ascertain what parameters contributed to the reconstructed lake-level fluctuations, a sensitivity analysis was performed with a water balance model. The water balance for any lake can be described by the equation

$$\Delta V = I_s + I_l + P_s - Q_s - Q_u - E \cdot A_l$$

where $\Delta V$ is change in lake volume; $I_s$ is subsurface inflow; $I_l$ is surface inflow; $P_s$ is precipitation onto the lake; $A_l$ is area of the lake; $Q_s$ is surface discharge out of the lake; $Q_u$ is subsurface outflow; and $E$ is evaporation.

The conditions present in the Kellyville basins allow for some assumptions and modifications to Equation (1). The hard bedrock and deep permafrost make it unlikely that groundwater is a factor. The hydraulic sill of the basin is at an elevation of 186 m, well above the highest shorelines present, so there are no outflows. There are no active major drainages leading into the lakes, and runoff is restricted to the spring melt season and precipitation events that occur while the ground is still saturated from the melt. Precipitation falling in the drainage areas during July, August, and September is likely to be quickly removed via high rates of evapotranspiration, so that only precipitation falling directly on the lake contributes to volume change. Winter precipitation, on the other hand, is likely to experience minor evaporation and ablation and will runoff quickly to the lakes during the melt season. These conditions have been observed by a number of workers in the area. Hasholt (personal communication, 2003) derived a continuity equation specific to these lake basins, which is re-written here as:

$$\Delta V = P_w \cdot MAP \cdot (A_l + A_s) + (P_s \cdot MAP \cdot A_l) - (E \cdot A_l)$$

where $P_w$ is the proportion of winter precipitation, $A_s$ is the area of the drainage basin, and $P_s$ is the proportion of summer precipitation. Summer precipitation is defined here as the precipitation that falls during the months of July, August, and September; winter precipitation falls during the remainder of the year. Although precipitation in May could be liquid, it would likely run off directly to the lakes as a result of either saturated or frozen ground. In addition, field observations have shown that the drainage basins are still providing direct-runoff during June precipitation events because of saturated soils. In the present model, Equation (2) is used, as it accounts for all the major gains and

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**Figure 8.** Lake-level curve for Hunde Sø and DI conductivity for Braya Sø along with depositional depth of dated samples. Depositional depth estimated using GPR and modern depositional environments. Zones I, II, III, and subzones a–c correspond to those discussed in text and in Table 2. Dashed and dotted lines indicate the elevation at which Braya Sø and Hunde Sø and all the lakes become connected, respectively.
losses of water in these lakes. Ignored in this equation is sublimation of snow and ice during the winter months.

Evaporation

Equations to estimate evaporation from an open water body are numerous, and each has assumptions and estimates that are required when using climatological data. Many of the equations for evaporation are based on the principal of Dalton’s Law, which states that all other factors being constant, evaporation is proportional to the wind speed and the vapor pressure deficit (Dalton, 1802 as cited in Ward, 1967). Shuttleworth (1993) provides several methods for estimating evaporation under different conditions of data availability. Of the methods presented, the following variant of the Penman equation was chosen, as it accounts for wind speed, vapor pressure deficit, and the energy available for evaporation, while requiring the smallest number of assumptions.

\[ F_p = F_p^1 * A + F_p^2 * \tilde{D} \]

(3)

where \( E_p \) is potential evaporation (mm/day), \( A \) is the energy available for evaporation (which in this case is equivalent to the net radiation, \( R_n \), since advected energy is negligible to the absence of any inflowing rivers), \( \tilde{D} \) is the average vapor pressure deficit, \( F_p^1 \) and \( F_p^2 \) are coefficients defined as

\[ F_p^1 = \Delta / (\Delta + \gamma) \]

(4)

\[ F_p^2 = [\gamma / (\Delta + \gamma)] * [6.43(1 + 0.536U_2^2)/1] \]

(5)

where \( \Delta \) is the slope of the saturated vapor pressure curve, \( \gamma \) is the psychrometric constant, \( \lambda \) is the latent heat of vaporization, and \( U_2 \) is the wind speed measured 2 m above the surface. The methods used to estimate the various parameters in Equations (3) to (5) are discussed in Shuttleworth (1993). The vapor pressure deficit can be calculated using the air temperature in combination with dew-point temperature, relative humidity, or the wet-bulb temperature. Apart from air temperature, these data do not exist, so minimum air temperature was substituted for dew-point temperatures. Minimum air temperature and water surface temperature were estimated using the modern empirical relation between these variables and mean air temperature. A third-order polynomial was fit to the output of Berger and Loutre’s (1991) radiation calculations for mid July at 65°N and used to calculate the radiation for each year modeled.

Although the average wind speed recorded at Kangerlussuaq is 3.6 m/s, hilly terrain that alters the surface winds from those measured in the fjord at Kangerlussuaq surrounds the study lake basins. The modal summer wind speed for June 2001 to August 2002 recorded by an Automatic Weather Station (AWS) at Lake E was 3.9 m/s. Lake E (see Figure 1) is also surrounded by hills and located at approximately the same distance from the ice sheet and the coast as the study basins; therefore this value is used in the model. Average monthly cloud cover was available from the Danish Meteorological Institute (Cappelen et al., 2001) for the years 1973–1999 and averaged ~60% for the summer months. The solar zenith angle at Kangerlussuaq varies from approximately 60° to 75° during the summer months and, with a 60% cloud cover, results in a water surface albedo of 0.1 (Hartmann, 1994).

Basin geometric properties

A digital elevation model (DEM) was created in ArcGIS using a combination of published topographic maps, bathymetry data, and field surveys. Detailed surveying of Hunde Sø and Braya Sø shores up to an elevation of ~180 m was performed using a Trimble GeoExplorer Global Positioning System (GPS) in the summer of 2003. Local ephemeris data collected at the Kellyville Scat-
Figure 9. Sensitivity results for the lowstand condition. The plotted lake elevations are those at the end of the 2000 yr period, but models reached equilibrium within 200 years: (a) ΔJJA_m and P_s; (b) ΔJJA_m and ΔMAP; (c) ΔMAP and P_s; and (d) ΔJJA_m and ΔE_d. The large steps seen in the response between 170 and 168 are a result of the switching on and off of the contribution from NE Lake, while the plateaus are a result of the large land bridges between the three lakes and the paleoshorelines surrounding them. An asterisk denotes values from the modern climatology (JJA_m = 9.4°C; P_s = 0.43; MAP = 148.0 mm; and E_d = 91 days).

Figure 10. As in Figure 5, but for the highstand condition. Only those experiments that returned an elevation above 175 m are plotted.
dromologic parameters. In all cases this occurred within 30 time steps (~300 years), after which lake elevation would start rising slightly but at a rate of less than 0.5 m over ~1,700 years.

**Lowstand condition**

Immediately obvious in the results of the sensitivity analysis for the lowstand condition is the presence of systematic jumps and plateaus in calculated lake elevation (Figure 9). The jumps are a result of switching NE Lake drainage off and on based on the modeled elevation. The plateaus result from the large flat areas of land around Hunde Sø that are part of the ancient shorelines, including the land bridges between the other lakes, and the modern shelf below 170 m. The modeled lake elevation appears to be more sensitive to changes in precipitation parameters (P<sub>s</sub> and MAP) than those affecting evaporation (JJA<sub>d</sub> and E<sub>d</sub>). A ΔJJA<sub>m</sub> of +4°C combined with an increase in E<sub>d</sub> of 14 days is insufficient to lower the lake enough to turn off drainage from NE Lake (<168 m). In contrast, the water drops below the 168 m threshold with a ΔMAP of -40 mm with all other parameters held to modern values. The standard deviations (SD) of JJA<sub>m</sub>, MAP, and P<sub>s</sub> for the period of record are 0.86, 49.97, and 0.15, respectively. Therefore, to reduce lake levels below the 168 m threshold, a change of more than 1×SD in JJA<sub>m</sub> is required, whereas less than 1×SD change in MAP results in the same reduction of water level. Similarly, a -4°C change in JJA<sub>m</sub> combined with a 14 day reduction in E<sub>d</sub> is required to increase modeled water levels to 175 m, whereas only a modest increase in MAP of about 1×SD is required. The combined effect of changes in P<sub>s</sub> and MAP (Figure 9c) result in the widest ranges of modeled lake elevation.

**Highstand condition**

As with the lowstand condition, the modeled lake elevations above 175 m appear to be more sensitive to precipitation parameters than evaporation parameters. However, the importance of evaporative processes has increased with the larger lake area (Figure 10). In particular, for the modeled lake elevation to reach the sll elevation of 186 m, a combination of -4°C change in JJA<sub>m</sub> and an increase in MAP above 50 mm is required. Comparing Figure 9b with Figure 10b, it can be seen that the change in lake level per degree change in JJA<sub>m</sub> becomes much greater for the highstand condition. However, a large change in JJA<sub>m</sub> is still required to raise lake elevation to the height recorded in the field evidence, and the model still shows a higher sensitivity to the combined effect of P<sub>s</sub> and MAP (Figure 10e).

**Climate reconstruction**

The sensitivity tests provide broad envelopes of climate parameters that are required to maintain either the highstand or lowstand condition and highlight the importance of precipitation variabil-

Table 2. Estimated parameters for Zones I to IIIc and the corresponding time periods

<table>
<thead>
<tr>
<th>Zone</th>
<th>Sub-zone</th>
<th>Time period (cal. yr BP)</th>
<th>Panel in</th>
<th>ΔJJA&lt;sub&gt;m&lt;/sub&gt; (°C)</th>
<th>ΔE&lt;sub&gt;d&lt;/sub&gt; (days)</th>
<th>ΔMAP range (mm)</th>
<th>Estimated P&lt;sub&gt;s&lt;/sub&gt;</th>
<th>Estimated ΔMAP (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td></td>
<td>-8000–6070</td>
<td></td>
<td></td>
<td></td>
<td>-10 to -80</td>
<td>0.30</td>
<td>-80</td>
</tr>
<tr>
<td>II</td>
<td>a</td>
<td>6070–4600</td>
<td>a, d</td>
<td>+2.5</td>
<td>+13</td>
<td>+60 to +130</td>
<td>0.60</td>
<td>+130</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>4600–3600</td>
<td>e</td>
<td>+2.0</td>
<td>+10</td>
<td>+60 to +130</td>
<td>0.40</td>
<td>+80</td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>3600–3000</td>
<td>b</td>
<td>+1.0</td>
<td>+5</td>
<td>0 to -70</td>
<td>0.40</td>
<td>+20</td>
</tr>
<tr>
<td>III</td>
<td>a</td>
<td>3000–2600</td>
<td>c</td>
<td>0.0</td>
<td>0</td>
<td>+30 to +90</td>
<td>0.50</td>
<td>+70</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>2600–700</td>
<td>f</td>
<td>+1.0</td>
<td>+5</td>
<td>+40 to +100</td>
<td>0.30</td>
<td>+40</td>
</tr>
<tr>
<td></td>
<td>c</td>
<td>700–0</td>
<td>e</td>
<td>0.0</td>
<td>0</td>
<td>-30 to +20</td>
<td>0.40</td>
<td>-20</td>
</tr>
</tbody>
</table>

Values of ΔJJA<sub>m</sub> and ΔE<sub>d</sub> are those used in the model to produce Figure 11. ΔMAP range results from using P<sub>s</sub> from 0.30 to 0.60. Estimated P<sub>s</sub> was determined from GISF2 Na<sup>+</sup> flux records, and Estimated ΔMAP was calculated using the P<sub>s</sub> determined from Na<sup>+</sup> records (see text).

**Zone I (8,000–6,070 cal. yr BP)**

To estimate both the climate parameters and the magnitude of the Zone I lowstand in Hunde Sø, the modeling methodology described above was applied to the Braya Sø bathymetry and catchment. Using a ΔJJA<sub>m</sub> of +2.5°C and ΔE<sub>d</sub> of +13 days, a ΔMAP range of -80 to -40 mm was necessary to reduce Braya Sø by 11 m, the depth of the drowned beach ridges. For any combination of P<sub>s</sub> and ΔMAP required to attain a stable Braya Sø lake elevation 11 m below modern, the resulting Hunde Sø elevation is 151.4 and 152.3 m. This equates to a decrease of 18 to 19 m below modern and represents a significant dry period in the Holocene record for this area.

**Zone II (6,070–3,000 cal. yr BP)**

Zone IIA (6,070–4,600 cal. yr BP) begins with a rapid rise in lake levels resulting first in the connection between Braya Sø and Hunde Sø, and then with Limnaea Sø. The ΔJJA<sub>m</sub> was +2.5°C and ΔE<sub>d</sub> was calculated to be +13 days. With the increased temperature and duration of summer evaporation, a ΔMAP of +60 to +130 mm was needed to raise Hunde Sø lake elevation above 175 m. All the lakes except Upper Braya remained connected through Zone IIB (4,600–3,600 cal. yr BP). Once connected a ΔMAP between +30 and +100 is sufficient to maintain the connection, but in order for the lakes to continue to rise above 176 m, ΔMAP still needed to be between +60 and +130 mm. The high temperature and precipitation of Zones IIB are consistent with the peak in local vegetation density found by Eisner et al. (1995) between 4,990 and 3,650 cal. yr BP, and with the minimal DI conductivity reported for SS6 and Braya Sø. However, the data presented here indicate that the increase in precipita-
tion began several hundred years earlier. Zone IIc (3,600–3,000 cal. yr BP) shows a rapid decline in Hunde Sø lake levels to an elevation of 172 m by about 3,000 cal. yr BP. A $\Delta JJA_m$ of +1°C and a $\Delta E_d$ of +5 days resulted in an estimated $\Delta MAP$ between +10 and +60 mm.

Zone III (3,000–0 cal. yr BP)

The lowering that began in Zone IIc was short-lived through Zone IIIa (3,000–2,600 cal. yr BP) and coincided with a period of temperature close to modern. Although Zone IIIa was modeled using modern $JJA_m$ and $E_d$, a $\Delta MAP$ between +50 and +100 mm was required to raise lake levels above the 175 m sill. The lakes again remained connected through Zone IIb (2,600–700 cal. yr BP), but because of the lower temperatures and shorter evaporation period than Zone IIb, a $\Delta MAP$ range of +40 to +100 mm was sufficient to maintain the connection between the lakes and raise the lake elevation to 176 m. Although the lakes may have experienced a brief decline ~1,000 cal. yr BP coincident with the “Medieval Warm Period,” no definitive evidence of this was found in the sedimentary record. Zone IIIc (700 cal. yr BP–modern) was modeled with modern $JJA_m$ and $E_d$ and the resulting $\Delta MAP$ range was between -30 and +20 mm.

Precipitation seasonality and the NAO

The broad ranges of $\Delta MAP$ provided when given a range of $P_s$ make it apparent that some constraint on the change in the seasonality of precipitation would be beneficial in further constraining the estimates of $\Delta MAP$. In West Greenland, the North Atlantic Oscillation (NAO) has a significant effect on the seasonality of precipitation, particularly winter precipitation (Chen et al., 1997; Appenzeller et al., 1998; Bromwich et al., 1999). During positive modes of the NAO, a strong Icelandic low coupled with the Azores high results in enhanced zonal flow across the North Atlantic and cold and dry winters in West Greenland. During negative modes of the NAO, the pressure gradient is reversed and a blocking ridge is set up over the East Atlantic, resulting in increased meridional flow and more storms tracking north to Greenland. West Greenland experiences mild and wet winters during negative NAO years. These two modes are known as Greenland Below (GB) and Greenland Above (GA), respectively.
Following the arid climate of Zone I, our results indicate Zone II to be characterized by warm and very moist conditions. The beginning of Zone II is estimated to have a ΔMAP of +130 mm but initially the DI conductivity remains high. However, it is reasonable to expect a lag between the significant lowstand in Zone I and the time when the lake would be dilute enough to preserve diatoms. It would also be reasonable to expect selective preservation of salt-tolerant diatoms during the initial refilling of the lake, which may have been prolonged by runoff through salt-enriched sediments deposited during Zone I. Modern runoff has been measured in the field to be an order of magnitude higher in conductivity as it flows through ancient lake sediments.

From Zone IIc through Zone III the records are difficult to reconcile. Although the DI conductivity increases slightly just prior to the Zone IIc decline in lake elevation, the conductivity remains higher through Zones IIIa and IIib even though the lake level has rebounded to above 175 m. However a slight trend towards decreased DI salinity is seen in the Braya So data during the first half of Zone IIib before increasing sharply to near modern levels (Figure 8). The DI conductivity curve for SS6 displays high millennial variability through this period with a trend towards decreasing salinity in the last ~1500 years. The reason for this is unknown, but the trend does not match either the climate reconstruction here or the other proxy curves. The overall trend towards lower ΔMAP values, agrees well with the trend towards enriched δ18O of both Braya So and SS6.

Conclusions

A history of lake-level fluctuations for the Kellyville basins was reconstructed using geomorphic, stratigraphic, and geophysical evidence from these basins. These lakes have experienced a series of fluctuations and have not merely evaporated to modern levels since glacial times. The lake-level curve agrees well with previous reconstructions using diatoms and isotopes from neighboring lakes during the mid Holocene, but discrepancies exist during the late Holocene. The reason for these discrepancies is not clear but may be a result of within-lake processes affecting lake chemistry (McGowan et al., 2003) or lags between changing climate and lake response. Additionally, re-working of salt-encrusted sediments from old shorelines and inflow from neighboring lakes may be causing noise in the biological and geochemical proxy records. Nonetheless, the overall trend towards increased aridity since ~4,500 cal. yr BP seen in the proxy records does agree with the modeled ΔMAP estimates reported here, which indicate a general decrease in MAP since ~4,600 cal. yr BP.

A model was developed to constrain the climatic changes necessary to cause the lake-level fluctuations seen in the sedimentary record. Lake levels were very sensitive to changes in precipitation and the seasonal distribution of precipitation, and less sensitive to temperature changes. The sensitivity of the model to precipitation seasonality creates uncertainty in estimating MAP from the model results. However, using evidence of long-term shifts in the atmospheric patterns of the North Atlantic region to infer changes in precipitation seasonality allowed quantitative estimates of MAP to be calculated.

Results indicate that this region of West Greenland has had a much warmer and moister climate compared with modern conditions through most of the mid to late Holocene. Estimates suggest that MAP was 80 to 130 mm higher than modern from 5,700 to 3,600 cal. yr BP and from 40 to 70 mm higher from 3,000 to 700 cal. yr BP. In contrast, MAP was ~80 mm less than modern during the postglacial lowstand prior to 6,070 cal. yr BP. The last

**Figure 12.** Five-year moving average of the station-based December through March NAO (dashed line) and \( P_s \) (solid line) from 1950 to 2004 (\( P_s \) data are missing from 1971–1975). Correlation between the two time series is insignificant but they are shown here to demonstrate the overall trend towards higher \( P_s \) during negative (positive) NAO modes. Horizontal black line denotes the average \( P_s \) from 1950 to 2004. NAO data from [http://www.jisao.washington.edu/data_sets/nao/](http://www.jisao.washington.edu/data_sets/nao/).

O’Brien et al. (1995) have shown distinct periods of increased or decreased marine Na+ in the GISP2 ice core through the Holocene, and Dawson et al. (2003) demonstrated that above average Na+ deposition at the GISP2 site occurs during GB years, while below average Na+ deposition occurs during GA years. Although \( P_s \) is highly variable from year to year and has no significant correlation with the NAO on an annual basis, it does appear to respond to major changes of state in the NAO (Figure 12); averaging 0.38 during negative NAO years (GA) and 0.48 during positive NAO years (GB). Therefore, direction of change in \( P_s \) can be inferred to be toward smaller (larger) values for times when Na+ flux is lower (higher), because GB winters are typically drier and GA winters are typically wetter.

O’Brien et al. (1995) consider the time periods 610 to 0 and 6,100 to 5,000 yr BP to be times of large increases in Na+, with lesser increases occurring from 3,100 to 2,400, and 8,800 to 7,800 yr BP; periods of decreased Na+ occur from 960 to 610, 2,700 to 1,500, 7,900 to 6,300 yr BP. Using O’Brien et al.’s periods of increased or decreased sodium flux and the SD of the modern distribution of \( P_s \) (rounded to the nearest 0.10), values of \( P_s \) were assigned to each zone (Table 2) in the following manner for the time periods of concern: + ~1 SD (0.60) from 6,070 to 4,600 cal. yr BP, + ~0.5 SD (0.50) from 3,000 to 2,600 cal. yr BP, + ~1 SD (0.30) prior to 6,070 and from 2,600 to 700 cal. yr BP, and 0.40 from 4,600 to 3,000 cal. yr BP when Na+ fluxes were about average. Using these values for \( P_s \), a single estimate of ΔMAP can be selected from the range of values given previously (Table 2).

**Comparison with proxy records**

The lake-level curve and climate reconstruction agree reasonably well with the diatom and δ18O record through most of Zones I and II. The drought conditions reconstructed for Zone I are reflected in McGowan et al.’s (2003) diatom record. An absence of diatoms in Braya So just prior to 5,784 14C yr ago (~6,630 cal. yr BP) and SS6 just prior to 5,280 14C yr ago (~6,170 cal. yr BP) was attributed to extreme salinity and dissolution resulting from evaporative enrichment during low lake levels (Figures 2 and 8). Anderson and Leng (2004) found that δ18O isotope from SS6 were significantly enriched during this time, also indicating high aridity (Figure 2). Immediately following this arid period, a depletion of δ13C in their record was attributed to increased runoff. The fluvial sediments in NSE, in an area where fluvial transport and deposition occurs today as only minor overland flow, support this.
700 cal. yr BP also appear to be much drier than the last ~6,000 years, with the lakes falling to modern levels, which are the lowest since the initial lowstand following deglaciation.

Geomorphic and sedimentary evidence provides only slices in time for interpretation of environmental change rather than continuous records, such as diatoms and isotopes. However, coupled with numerical modeling, these kinds of studies can provide valuable constraints within which the proxy record should be interpreted. Further modeling of these lake basins to couple hydrology and lake chemistry should be considered, as it may provide a better understanding of the complex relationships between hydrology, biology, and limnology.

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