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Mild Little Ice Age and unprecedented recent warmth in an 1800 year lake sediment record from Svalbard

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ABSTRACT

The Arctic region is subject to a great amplitude of climate variability and is currently undergoing large-scale changes due in part to anthropogenic global warming. Accurate projections of future change depend on anticipating the response of the Arctic climate system to forcing, and understanding how the response to human forcing will interact with natural climate variations. The Svalbard Archipelago occupies an important location for studying patterns and causes of Arctic climate variability; however, available paleoclimate records from Svalbard are of restricted use due to limitations of existing climate proxies. Here we present a sub-decadal- to multidecadal-scale record of summer temperature for the past 1800 yr from lake sediments of Kongressvatnet on West Spitsbergen, Svalbard, based on the first instrumental calibration of the alkenone paleothermometer. The age model for the High Arctic lake sediments is based on ^{210}Pb, plutonium activity, and the first application of tephrochronology to lake sediments in this region. We find that the summer warmth of the past 50 yr recorded in both the instrumental and alkenone records was unmatched in West Spitsbergen in the course of the past 1800 yr, including during the Medieval Climate Anomaly, and that summers during the Little Ice Age (LIA) of the 18th and 19th centuries on Svalbard were not particularly cold, even though glaciers occupied their maximum Holocene extent. Our results suggest that increased wintertime precipitation, rather than cold temperatures, was responsible for LIA glaciations on Svalbard and that increased heat transport into the Arctic via the West Spitsbergen Current began ca. A.D. 1600.

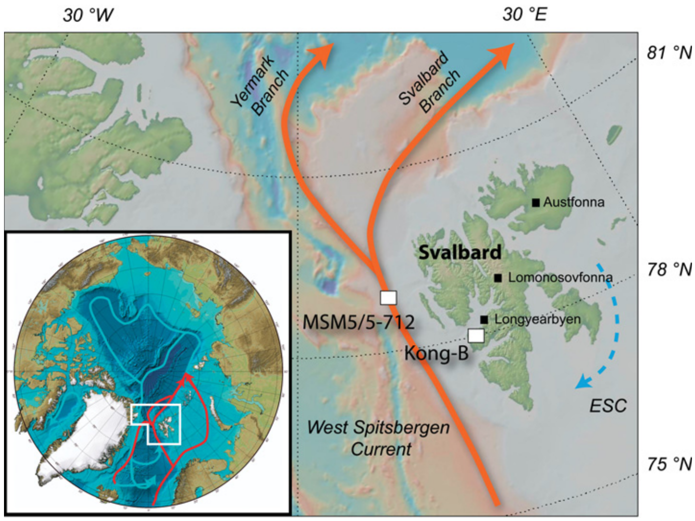

Figure 1. Site map with location of lake core Kong-B from Kongressvatnet and marine core MSM5/5–712 (Spielhagen et al., 2011). Longyearbyen, Lomonosovfonna, and Austfonna ice caps, West Spitsbergen Current, and East Spitsbergen Current (ESC) are marked. Inset map shows Arctic Ocean surface currents.

INTRODUCTION

The Arctic exerts a large influence on global climate and is currently undergoing profound change (Intergovernmental Panel on Climate Change; Solomon et al., 2007). In order to put current changes into greater context and to anticipate future climate changes we must understand the degree to which, and the reasons why, Arctic climate has varied in the past. Instrumental records of Arctic temperature rarely extend more than 60–70 yr into the past, so such records provide us with only a very limited perspective on climate variability. An extension of climate records beyond the brief instrumental period requires examination of geologic archives with quantifiable climate sensitivity. The Svalbard Archipelago (77°–80°N; Fig. 1) is situated between physically distinct ocean currents and air masses and along major conduits of oceanic and atmospheric heat transfer to the Arctic. At present, the northernmost extension of the Gulf Stream flows along Svalbard's western coast as the West Spitsbergen Current (WSC) (Fig. 1), and cold polar water flows along the eastern coast. Svalbard is located along the primary atmospheric moisture pathway between the Atlantic and the Arctic Basin, and precipitation over Svalbard is positively correlated with the mode of the North Atlantic Oscillation (NAO) (Dickson et al., 2000). In addition, changes in sea ice extent can result in an amplified response of Svalbard climate to changes in oceanographic and atmospheric conditions. For these reasons, small perturbations in the ocean and atmosphere have the potential to trigger large climatic responses over Svalbard. Climate modeling studies project that Svalbard will warm more than any other landmass by the end of the twenty-first century (Solomon et al., 2007) and its geographic location makes Svalbard important for understanding Northern Hemisphere climate dynamics.

A homogenized instrumental air-temperature record for Longyearbyen Airport (Fig. 1) begins in A.D. 1911, and documents multidecadal temperature variability and a significant warming trend during the past 100 yr (mean annual air temperature increased by 2.4 °C; Nordli, 2010). This air temperature record is long for the High Arctic, although not sufficiently long to examine the full range of natural climate variability over human (multidecadal to centennial) time scales during the recent past. Existing paleoclimate reconstructions from Svalbard are based on biological (Birks et al., 2004; Guilizzoni et al., 2006; Holmgren et al., 2010; Jiang et al., 2011; Velle et al., 2010), sedimentological, and bulk geochemical proxies (Svendsen and Mangerud, 1997) from lake sediments, and a variety of analyses from ice cores of the Lomonosovfonna and Austfonna ice caps (Divine et al., 2011; Grinsted et al., 2006; Isaksson et al., 2005; Nagornov et al., 2006). Biological proxies have provided limited paleoclimate information due to coarse temporal resolution, and sedimentological measurements reflect glacier variation rather than climate directly. Ice core proxies are sensitive to moisture source, air temperature, sea ice extent, ice melt, meltwater percolation, seasonality of precipitation, and accumulation rate, thereby complicating paleotemperature interpretation (Grinsted et al., 2006; Nagornov et al., 2006). There is weak correlation

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($r^2 = 0.15$) between ice core $\delta^{18}O$ and Svalbard mean annual air temperature during the instrumental period (Isaksson et al., 2005), and it has been proposed that Svalbard ice core $\delta^{18}O$ is a better proxy for winter temperature (Divine et al., 2011); however, the complications outlined above remain and additional paleotemperature records are needed to evaluate temperature changes outside of the winter season.

We report the first continuous, high-resolution, quantitative summer air temperature reconstruction for coastal West Spitsbergen, extending the record of air temperature for this location to ca. A.D. 200. Our record is from analysis of lake-sediment gravity core Kong-B, recovered in summer 2009 from 47 m water depth in Kongressvatnet (Fig. 1). Kongressvatnet (94 m above sea level, surface area 0.82 km^2) receives inflow from three mineral springs, leading to elevated ionic concentrations, water column meromixis, and hypolimnetic anoxia (Holm et al., 2011), and preservation of undisturbed, finely laminated sediments enabling sampling at high (4–30 yr) temporal resolution.

TEPHROCHRONOLOGY

The lack of terrestrial macrofossils and the presence of coal seams in the watershed that contribute ^{14}C-dead carbon to the lake (Guilizzoni et al., 2006) preclude radiocarbon-based age modeling of Kongressvatnet sediments; instead, we used ^{210}Pb and $^{239+240}Pu$ dating and tephrochronology to construct an age model for core Kong-B (Fig. 2; Table 1; see the GSA Data Repository[1]). Three distinct peaks in tephra concentration occurred within the core from 19–21, 27–29, and 39–44 cm; tephra geochemistry corresponds with that of glass shards attributed to the Icelandic eruptions of Öræfajökull, A.D. 1362 (Hall and Pilcher, 2002; Pilcher et al., 2005, 1995), Hekla, A.D. 1104 (Chambers et al., 2004; Hall and Pilcher, 2002; Pilcher et al., 2005, 1995, 1996; Vorren et al., 2007), and Snæfellsjökull, A.D. 170 (Kristjánsdóttir et al., 2007; Larsen et al., 2002), respectively (Fig. 2; Table 1; see the Data Repository).

Tephra horizons produced by Icelandic volcanic eruptions throughout the late Quaternary have been identified in distal sediments as cryptotephra at sites around the North Atlantic, providing key chronostratigraphic marker horizons (Turney et al., 2006; Wastegård and Davies, 2009). However, there is little information on these deposits at higher latitudes due to difficulty of isolating grains present in very low concentration. We were able to isolate and analyze tephra from dominantly minerogenic sediments and compile a robust geochemical data set to extend the known geographic distributions of three Holocene tephra horizons. Our results represent the northernmost lacustrine application of tephrochronology and highlight the potential of this approach for geochronology in the High Arctic.

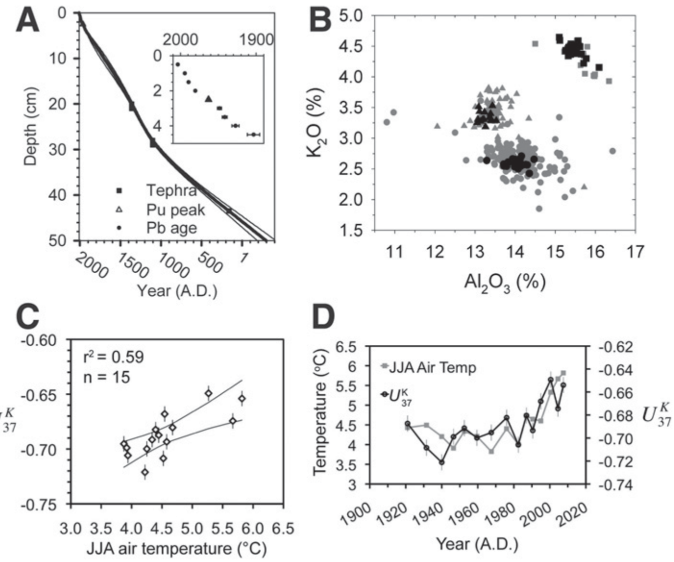

Figure 2. Chronology and U^K_{37} calibration. A: Age model for core Kong-B from tephrochronology, ^{210}Pb, and $^{239+240}$Pu dating (inset). B: Biplot of tephra shard geochemistry (Table 1) (see the Data Repository [see footnote 1]) from Kong-B (black) and reference samples (gray). C: U^K_{37} calibration from this study, analytical precision (±1σ), and 95% confidence interval of estimation (see the Data Repository). JJA—June, July, August. D: U^K_{37} time-series overlap with instrumental JJA air temperature (Nordli, 2010).

TEMPERATURE RECONSTRUCTION

To reconstruct summer (June, July, August—JJA) air temperature (T), we used the alkenone unsaturation index (U^K_{37}), a well-established marine paleothermometer (Brassell et al., 1986) that is also applicable in lakes (D'Andrea et al., 2011, 2006). We developed the first calibration in time for the alkenone paleothermometer ($U^K_{37} = 0.0255\ (T) - 0.804$; $r^2 = 0.59$; n = 15;

TABLE 1. MAJOR OXIDE AND Cl CONCENTRATIONS OF TEPHRA SHARDS FROM CORE KONG-B COMPARED TO REFERENCE TEPHRA

Sample (reference)	n	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Cl	Total
Kongress 19–21 cm	15	71.92	0.22	13.26	3.24	0.08	0.00	0.93	4.66	3.34	0.22	97.88
		(0.93)	(0.02)	(0.15)	(0.08)	(0.02)	(0.01)	(0.02)	(0.30)	(0.10)	(0.02)	(0.99)
Öræfajökull, A.D. 1362	71	71.86	0.27	13.47	3.20	–	0.04	1.04	5.08	3.39	–	98.36
(1, 3, 6)		(1.66)	(0.05)	(0.42)	(0.40)	–	(0.07)	(0.28)	(0.37)	(0.25)	–	(1.79)
Kongress 27–29 cm	21	71.24	0.20	14.03	3.27	0.11	0.10	1.92	4.35	2.59	0.06	97.87
		(0.95)	(0.03)	(0.26)	(0.14)	(0.03)	(0.03)	(0.11)	(0.38)	(0.07)	(0.01)	(0.91)
Hekla, A.D. 1104	148	71.51	0.25	14.10	2.98	0.11	0.15	1.96	4.60	2.66	–	98.32
(1–3, 5, 6, 8)		(1.36)	(0.11)	(0.66)	(0.45)	(0.12)	(0.13)	(0.28)	(0.36)	(0.25)	–	(1.53)
Kongress 39–44 cm	28	68.21	0.25	15.46	3.24	0.11	0.11	1.06	4.03	4.43	0.25	97.16
		(0.79)	(0.04)	(0.20)	(0.22)	(0.03)	(0.05)	(0.13)	(0.51)	(0.11)	(0.02)	(0.66)
Snæfellsjökull, A.D. 170	8	71.1	0.25	14.5	3.12	–	0.09	1.16	5.14	4.54	–	100.06
(7)		(1.20)	(0.03)	(0.36)	(0.17)	–	(0.01)	(0.05)	(1.13)	(0.21)	–	–
Snæfellsjökull, A.D. 170	6	67.27	0.38	15.69	4.08	–	0.33	1.72	5.07	4.36	–	98.96
(4)		(1.21)	(0.04)	(0.37)	(0.40)	–	(0.12)	(0.35)	(0.19)	(0.23)	–	(0.67)

Note: Sample references: 1—Pilcher et al., 1995; 2—Pilcher et al., 1996; 3—Hall and Pilcher, 2002; 4—Larsen et al., 2002; 5—Chambers et al., 2004; 6—Pilcher et al., 2005; 7—Kristjansdottir et al., 2007; 8—Vorren et al., 2007. Analytical precision (1σ) reported in parentheses. Major oxide and Cl concentrations are reported as percent of total. Dashes indicate values that were not reported in the original references.

[1]GSA Data Repository item 2012283, Figure DR1 (SEM image of a tephra shard from core Kong-B), Table DR1 (calibration data), Table DR2 (geochemical data for all analyzed tephra), Table DR3 (chronological markers used in the age model), and Equation DR1 (standard error calculation used in the temperature reconstruction), is available online at www.geosociety.org/pubs/ft2012.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

p < 0.001) by comparing U^K_{37} values in the radiometrically dated section of core Kong-B with instrumental JJA temperatures from Longyearbyen, ~40 km away (Nordli, 2010) (Fig. 2; see the Data Repository). The correlation is strong and statistically significant, notwithstanding chronological uncertainty and calibration to air temperature rather than directly to lake water temperature. Temperature estimates below 3.3 °C required extrapolation of the U^K_{37}-temperature relationship beyond the calibration range; therefore, the standard error of estimation below U^K_{37} of –0.72 is undetermined.

The U^K_{37} reconstruction reveals JJA temperatures in Longyearbyen, Svalbard, ranging from ~2 °C to 6 °C between A.D. 230 and A.D. 2009 (Fig. 3). From A.D. 230 to A.D. 1600, summer air temperatures were 2–4 °C, exhibiting multidecadal- to centennial-scale variability. Summer temperatures reached the lowest values of the past 1800 yr (2 °C) during the 10th century (A.D. 910–990) and from ca. A.D. 1100–1300. The latter interval can be separated into two distinct cold periods (A.D. 1100–1160 and A.D. 1250–1300) interrupted by 90 yr during which air temperatures warmed by ~0.5 °C. From A.D. 1600 to present, temperature increased from 4 °C to 6 °C, with multidecadal variability superimposed upon a long-term warming trend of ~0.4 °C per century. Two intervals of particularly warm summers stand out during this period: the first between ca. A.D. 1750 and 1780, and the second between A.D. 1987 and 2009. The A.D. 1987–2009 warming is documented by instrumental air temperature measurements (Nordli, 2010) and is apparently unprecedented in West Spitsbergen for at least the past 1800 yr, including during the Medieval Climate Anomaly (MCA; A.D. 950–1250). Two episodes of relative summer warmth (A.D. 1010–1060 and A.D. 1160–1250) during the MCA were 2 °C and 2.5 °C colder, respectively, than mean JJA temperature from A.D. 1987–2009.

Our summer temperature reconstruction can help distinguish the roles of summer ablation and winter accumulation in determining glacier activity on Svalbard (Baranowski and Karlén, 1976; Humlum et al., 2005; Svendsen and Mangerud, 1997; Werner, 1993) during the past 1800 yr, including during the Little Ice Age (LIA). Cold summers ca. A.D. 1250–1300 (Fig. 3) correspond to an episode of extensive alpine glaciation on Svalbard (Baranowski and Karlén, 1976; Werner, 1993), suggesting that reduced ablation contributed to glacial advance at this time. Moraine stabilization in the early 14th century (Svendsen and Mangerud, 1997; Werner, 1993) suggests a climatic shift that limited glacier advance and allowed moraine development. Warming ca. A.D. 1300–1330 documented

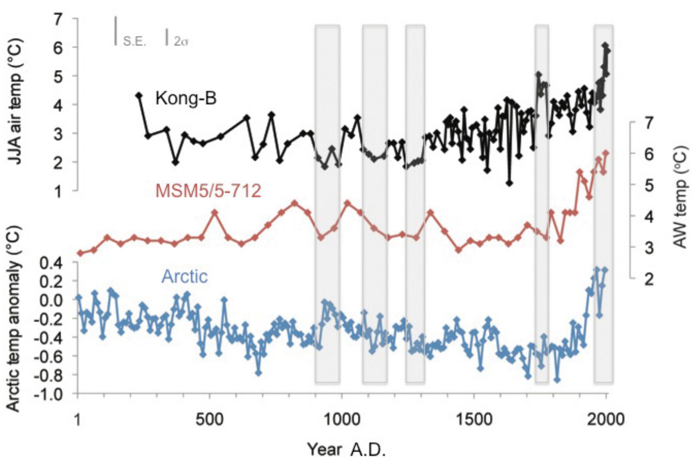

Figure 3. Temperature reconstructions. Upper plot: U^K_{37}-based June, July, August (JJA) air temperature (temp) from Kongressvatnet. Gray shading denotes intervals of coldest and warmest summer temperature during past 1800 yr. S.E.—standard error. Middle plot: Atlantic Water (AW) temperature in eastern Fram Strait based on foraminiferal assemblages from core MSM5/5–712 (Spielhagen et al., 2011). Bottom plot: Arctic temperature anomalies from A.D. 980–1800 mean (Kaufman et al., 2009).

in our record suggests that increased summer temperature may have limited glacier advance at the time. An earlier episode of moraine stabilization ca. A.D. 1000 (Werner, 1993) also corresponds to a warming event (~1 °C) in our record that interrupted a century-long cold period, consistent with the idea that summer ablation changes modified glacier activity.

However, the most recent period of glacier advance on Svalbard, during which time glaciers occupied nearly the same positions as ca. A.D. 1300 (Werner, 1993), occurred within the past 200 yr (Salvigsen and Høgvard, 2005; Werner, 1993), when summer temperatures (3–4.5 °C) were 1–2.5 °C higher than during previous periods of glacier advance on Svalbard. The 18th to 19th century LIA glaciation on Svalbard has generally been considered a response to low temperatures (Humlum et al., 2005; Svendsen and Mangerud, 1997). However, our findings do not support this explanation and instead suggest that increased wintertime precipitation led to the most recent glacier advance on Svalbard. Indeed, there is evidence that glacier expansion in western Scandinavia during the 18th century was not due to low summer temperatures, but to increased winter precipitation associated with changes in the NAO (Nesje et al., 2008, and references therein), which is positively correlated with winter precipitation in both Scandinavia and Svalbard (Dickson et al., 2000). Tree ring studies from Scandinavia and northern Europe do not reveal cold summer temperatures during 18th century glaciations (Nesje et al., 2008), while proxy-based NAO studies suggest positive NAO mode dominance during the 19th and early 20th centuries (Glueck and Stockton, 2001; Luterbacher et al., 2004; Trouet et al., 2009), supporting our interpretation that winter precipitation exerted primary control on LIA glaciations on Svalbard.

A synthesis of paleotemperature proxy records (Kaufman et al., 2009) (Fig. 3) indicates that the Arctic has undergone a cooling trend from A.D. 0 to 1900 that was interrupted by recent warming. Our temperature reconstruction does not show this trend for West Spitsbergen, and we note that no records from Svalbard were included in the synthesis. In addition, the Kaufman et al. (2009) reconstruction shows A.D. 1600–1900 as the coldest interval of the past 2000 yr in the Arctic, while our record indicates gradual warming in West Spitsbergen from A.D. 1600 to 1900. We propose that the proximity of West Spitsbergen to the WSC distinguishes its temperature history from that of the Arctic as a whole, and that the slow increase in air temperature over West Spitsbergen beginning ca. A.D. 1600 is related to warming of water in eastern Fram Strait at the time (Fig. 3; Spielhagen et al., 2011). Our U^K_{37} record and the foraminiferal assemblage record of Spielhagen et al. (2011) both reveal gradually increasing temperature between A.D. 1600 and 1890 (0.2–0.3 °C per century), followed by accelerated warming during the past ~120 yr. An initial increase in the volume of warm Atlantic Water transported northward by the WSC from A.D. 1600 to 1890, or increased temperature of this water, could have resulted in warmer summers in proximal West Spitsbergen, while an Arctic-wide temperature response to increased heat transfer and associated feedbacks did not commence until ca. A.D. 1900.

Sea surface temperature (SST) reconstructions from lower latitude sites sensitive to the North Atlantic Current also indicate warming within the period A.D. 1600–1900 (Richter et al., 2009; Sejrup et al., 2011; Sicre et al., 2011). However, these records indicate greater warmth from ca. A.D. 1000 to 1300 followed by cooling SSTs during the 13th century, features that have been attributed to an MCA-LIA transition (Sejrup et al., 2011; Sicre et al., 2011) but that are not apparent in the higher latitude records (this study; Spielhagen et al., 2011). These differences reflect the complexity of North Atlantic oceanography and climate dynamics and highlight the need for a greater number of regional paleoclimate reconstructions to better constrain past patterns of variability.

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