# Late-Holocene glacier growth in Svalbard, documented by subglacial relict vegetation and living soil microbes

Ole Humlum,[1]* Bo Elberling,[2] Anne Hormes,[3] Kristine Fjordheim,[4] Odd Harald Hansen[5] and Jan Heinemeier[6]

(*[1]Department of Physical Geography, Institute of Geosciences, University of Oslo, Norway, and The University Center in Svalbard (UNIS), Box 156, N-9170 Longyearbyen, Norway; [2]Institute of Geography, University of Copenhagen, DK1350 Copenhagen K., Denmark; [3]Ångströmlaboratory, AMS Laboratory, Uppsala University, SE-75121 Uppsala, Sweden; [4]Department of Geography, University of Bergen, Norway; [5]Norwegian Polar Institute (NP), Tromsø, Norway; [6]AMS 14C Dating Laboratory, Institute of Physics and Astronomy, University of Aarhus, DK8000 Aarhus C, Denmark*)
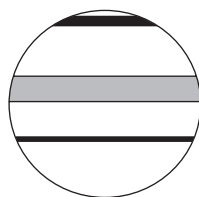
Abstract: Much renewed research interest in Arctic regions stems from the increasing concentration of atmospheric greenhouse gases and the alleged climatic sensitivity of high latitude areas. Glacier and permafrost changes are among a number of proxies used for monitoring past and present Arctic climate change. Here we present observations on frozen *in situ* soil and vegetation, found below cold-based glacier Longyearbreen (78°13′N), 2 km upstream from the present glacier terminus. Dating of the relict vegetation indicates that the glacier has increased in length from about 3 km to its present size of about 5 km during the last *c*. 1100 years. The meteorological setting of non-surging Longyearbreen suggests this example of late-Holocene glacier growth represents a widespread phenomenon in Svalbard and in adjoining Arctic regions. In addition, we use the subglacial permafrozen soil system to evaluate microbial survival capacity over considerable time periods, and we present evidence for microbes having survived more than 1100 years in a subglacial, permafrozen state.

Key words: Arctic, Svalbard, glacier variations, climate change, radiocarbon, subglacial vegetation, palaeosol, microbes, late Holocene.

Introduction

Changes in the Arctic atmosphere–ice–ocean system observed in recent years have sparked intense discussions as to whether these changes represent episodic events or long-term shifts in the Arctic environment. Early twenty-first century concerns about future climate change stem from the increasing concentration of greenhouse gases in the atmosphere, and during the last 15 years the Arctic has gained a prominent role in the scientific debate regarding global climatic change (Houghton *et al.*, 2001). Existing knowledge on Quaternary climate and Global Climate Models (GCMs) predicts that the effect of any present and future global climatic change will be amplified in

the polar regions as a result of feedbacks in which variations in the extent of glaciers, snow, sea ice and permafrost, as well as atmospheric greenhouse gases, play key roles. These are some of the reasons for renewed research interest in Arctic regions. Subcontinental-scale analysis of meteorological data obtained during the instrumental period apparently lends empirical support to the alleged high climatic sensitivity of the Arctic (Giorgi, 2002). Polyakov *et al.* (2002a,b), however, recently presented updated observational trends and variations in Arctic climate and sea-ice cover during the twentieth century, which do not support the modelled polar amplification of surface air-temperature changes observed by surface stations at lower latitudes.

There is reason, therefore, to evaluate climate dynamics and their respective impacts on high-latitude glaciers. Within the
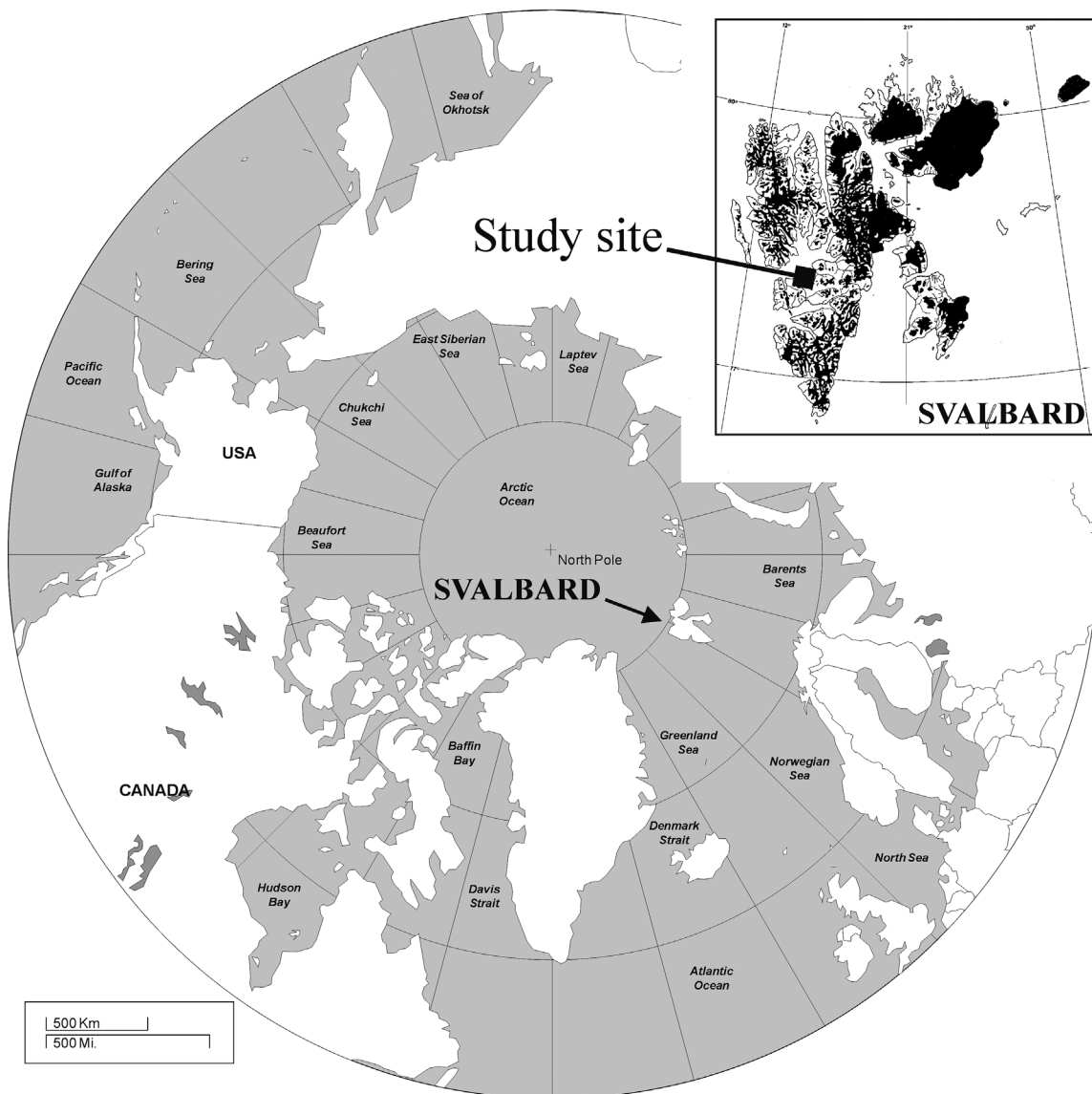
Figure 1 Map showing the Arctic regions. Insert shows the Svalbard archipelago ($63\,000$ km^2) with the study site in central Spitsbergen indicated

High Arctic, a unique climatic sensitivity applies to the Archipelago of Svalbard (Figure 1; 77–$80°$N), located on the boundary between the Norwegian Sea, the Barents Sea and the Arctic Ocean. This was recognized early by Ahlmann (1953) and Lamb (1977). Climatic variations in Svalbard during the twentieth century have been well documented by meteorological data since 1911 (Førland *et al.*, 1997). For example, a marked warming around 1920 changed the mean annual air temperature (MAAT) at sea level within only 5 years from about $-9.5°$C to $-4.0°$C. This represents the most pronounced increase in MAAT documented anywhere in the world during the instrumental period. Later, from 1957 to 1968, MAAT dropped about $4°$C, followed by a more gradual increase towards the end of the twentieth century. In the latest IPCC report (Houghton *et al.*, 2001), attention is drawn specifically to the extraordinarily high Svalbard climatic sensitivity.

The climatic sensitivity of the Svalbard region presumably derives from various forcing mechanisms. First, Svalbard is located near the confluence of ocean currents and air masses that have very different temperature characteristics. The northernmost tip of the North Atlantic Drift presently flows along the west coast of Svalbard, while cold polar water flows south along the eastern coast. Secondly, the climatic sensitivity is enhanced by rapid variations in the sea-ice extent, which are coupled with both atmospheric and oceanic circulations (Benestad *et al.*, 2002). Finally, Svalbard is located in the major transport pathway of water vapour into the Arctic Basin in association with the North Atlantic cyclone track (Dickson *et al.*, 2000). During the winter season meridional moisture transport along this pathway exhibits a positive relationship with the phase of the North Atlantic oscillation (Jones *et al.*, 2001).

These characteristics make Svalbard a key site for understanding aspects of Northern Hemisphere climatic changes; especially as most predictions of future climate suggest an amplified climatic response in the Arctic. Against this background the dynamics of Svalbard glaciers (Hagen *et al.*, 1993, 2003; Liestøl, 1993; Bruland and Hagen, 2002) represent an important source of Arctic environmental information, past as well as present. Motivated by the importance of estimating the impact of climatic change on High Arctic environments, we present here observations from a fossil soil, with intact plant cover and microbes, found in a subglacial position beneath the glacier Longyearbreen, in order to examine the effects of past climate change on Svalbard.

Longyearbreen

Glaciers presently cover about 60% of Svalbard (Hagen *et al.*, 1993, 2003; Hagen, 1996). On Spitsbergen, the main island in the Svalbard archipelago (63 000 km²), glaciation is especially extensive in areas near the eastern and western coasts, where many glaciers terminate in the sea. In contrast, glaciers in the central part of the island are smaller, mainly because of low precipitation (Humlum, 2002). The Longyearbreen glacier (Figure 2) is located in the relatively dry region of central Spitsbergen, near the main settlement of Longyearbyen. At Longyearbyen airport (45 m a.s.l.) the annual precipitation is about 180 mm (water equivalent), MAAT is about $-5°C$ and most of Svalbard is situated within the zone of continuous permafrost (Humlum *et al.*, 2003).

Longyearbreen (3.2 km²) is about 5 km long and flows from a large cirque at the head of the valley Longyeardalen (Figure 2). The valley is situated in the bioclimatic zone classified as the Inner Fjord Zone (Summerhayes and Elton, 1928; Elvebakk, 1985), the so-called Cassiope-tetragona zone of the mid-Arctic belt (Brattbakk, 1986). This represents a part of Svalbard with floristically rich vegetation, reflecting favourable climatic conditions for plant growth, mainly because of high summer insolation.

Surging glaciers are frequent on Svalbard (Dowdeswell *et al.*, 1991; 1995; Hagen *et al.*, 1993; Jiskoot *et al.*, 2000), especially in the moist coastal regions (Humlum, 2002). There is, however, no evidence for past surge behaviour of Longyearbreen. Owing to the protective debris cover on lower Longyearbreen, a typical feature of many non-surging glaciers in central Spitsbergen, this glacier has not retreated following the twentieth-century warming around 1920. Instead it has thinned 10–30 m over a considerable proportion of its length where a protective layer of supraglacial debris is lacking. It is presumably frozen to the bed except for small temperate areas and the maximum surface velocity is low, 1–3 m/yr (Tonning, 1996; Etzelmüller *et al.*, 2000).

Below the equilibrium line crevasses are narrow and few. Perennial supraglacial meandering meltwater channels erode into the glacier surface each summer. When such channels reach a depth of 8–12 m, ice deformation slowly closes the upper part of the channel. Thus, the channels are gradually transformed into englacial, meandering tunnels. A number of these meltwater conduits have now reached the glacier bed 20–50 m below the surface, exposing the glacier–bed interface for considerable distances along their course. During the winter, when water discharge has ceased, these tunnels provide a unique means of access to the glacier bed. In Svalbard, glacier deformation is usually low because of low ice temperatures. Therefore, entrances of englacial drainage systems are not completely closed and can be entered in the wintertime using alpine techniques (Vatne, 2001).

The exposed glacier bed below Longyearbreen generally takes the appearance of a normal talus or scree slope, with no visual indication of glacial erosion. Inspecting the glacier bed in one of the conduits, about 2 km upstream of the present glacier terminus, a well-preserved soil layer with *in situ* vegetation was found at the glacier–bed interface (Figure 3). The vegetation (Figure 4) was undisturbed, even though 30–40 m of active glacier ice presently covers the site. Trimlines above the modern glacier surface indicate that the ice thickness during the Little Ice Age was 20–30 m greater. The lack of evidence for glacial abrasion on boulders and bedrock exposed at the glacier bed led us to conclude that the glacier has remained cold-based at this particular place since it overran the site in the past. This is supported by the visual observation of mosses and other plants extending 1–3 cm up in the basal ice undeformed by glacial movement (Figure 4). The temperature at the glacier bed was measured at about $-4°C$ (March–May 2001, 2002, 2003).

Study samples were taken from the subglacial soil at several locations 30–35 m below the present-day glacier surface, at 450 m a.s.l. The present-day glacier terminus is at 250 m a.s.l., 2 km further downvalley. Macrofossils were investigated in the laboratory, but only a few plant species were found. This corresponds well with results from similar investigations of macrofossils from other terrestrial sediments in Svalbard (e.g. van der Knaap, 1989). *Salix polaris* (polar willow) is the single dominant species, having a high frequency in all our samples. Smaller parts of *Gramineae/Juncaceae* and moss taxa were also identified. The following moss species were identified: *Sanionia uncinata*, which typically grows on patterned ground in glacier forefields distant from any permanent water sources such as ponds and streams; *Tomentypnum nitens*, which is common in association with relatively high vegetation and moderate winter snow cover; and *Distichium* spp. and

Figure 2 Longyearbreen in late June 2000, looking south. The glacier measures about 5 km in total length. The subglacial study site is below the dot on the glacier surface, about 2 km upstream from the debris-covered glacier terminus

Figure 3 Basal part of Longyearbreen, about 35 m below the present-day glacier surface. The bed consists of a talus sheet, without indication of glacial erosion. An old soil with vegetation is found along the glacier–bed interface. Figure 4 shows the interface close to the left-hand margin of this picture. The lowermost 60–80 cm white ice presumably represents a large snowdrift overrun by the glacier, protecting soil and vegetation
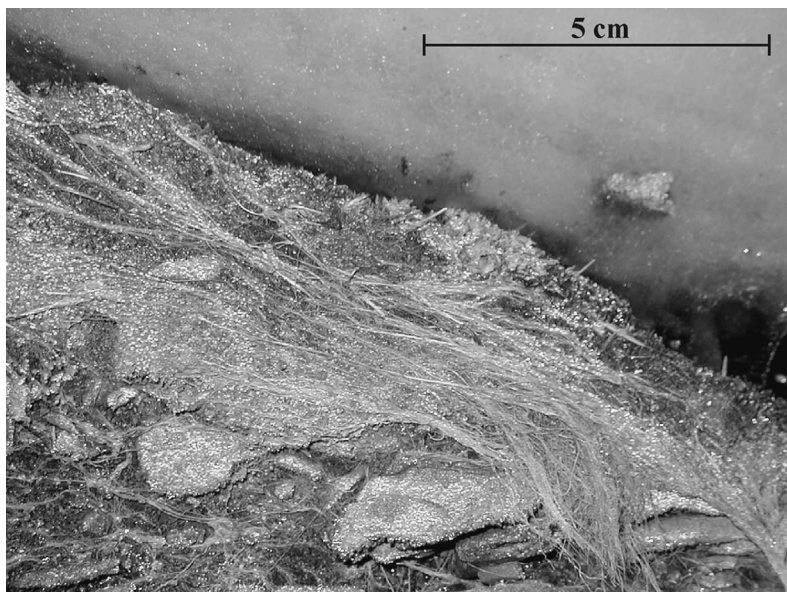
Figure 4 Close-up of *in situ* vegetation at the glacier–bed interface below Longyearbreen. The section shown measures about 10 cm across

Pogonatum urnigerum that typically live in snow-bed communities in topographic depressions. *Bryum pseudotriquetrum* is common along the margins of relatively large streams in the Arctic region. In addition *Myurella julacea* (Figure 5), *Racomitrium ericoides* and *Syntrichia ruralis* were identified, species that are typical of somewhat drier environments than the other bryophyte species mentioned here.

All species found in the subglacial soil below Longyearbreen are known from modern vegetation communities on Svalbard, where they form a widespread element of the present vegetation (e.g., Rønning, 1996). The species identified are all characterized by high adaptability to a range of environmental and climatic conditions in the Arctic. Future investigations of microfossils (pollen and spores) may result in lower taxonomic identification and identification of additional species from these soil samples.

Longyearbreen presumably overran the now-subglacial soil and vegetation during a winter advance without erosive meltwater activity at the glacier terminus. Assuming unchanged prevailing wind conditions, a thick snowdrift probably formed downvalley of the advancing glacier front, just as is the case today along the terminus of Longyearbreen during winter. Such a snow layer may have protected the vegetation to some degree during the glacier advance. We speculate that this layer of snow is represented by a white layer of bubbly ice along the glacier–bed interface (Figure 3). Oxygen isotope analysis of ice sampled from this layer yielded $\delta^{18}O$ values from -11.88 to -12.02, which is only slightly higher than modern winter snowpack values (Humlum, unpublished). Modern daily winter air temperatures at the terminus of Longyearbreen typically range from -15 to $-30°C$.

Radiocarbon dating

Methods

Frozen samples of soil and vegetation were brought to the nearby laboratory at UNIS (The University Centre in Svalbard) in polyethylene bags. The first radiocarbon dating (^{14}C; AAR-6532, Table 1) of the vegetation exposed at the glacier base yielded a calibrated age of 1700–1410 cal. yr BP. This result, and the unique preservation of the individual plants (Figure 4), encouraged us to investigate probable age variations of the different bryophyte species (mosses) when dated separately. The mosses were not subject to contamination after the glacier advance because the vegetation was perfectly preserved below, and partly in, glacier ice. This is an optimal situation for testing if radiocarbon dates of different species are mutually consistent. Mosses were selected with a pipette under magnifying glasses and binocular microscope in the clean room of the Ångströmlaboratory (University of Uppsala). Risk of contamination for small and wet macrofossil samples increases with long-term storage, as a result of uptake of modern carbon through micro-organisms (Wohlfarth *et al*., 1998). The macrofossils were therefore cleaned under binocular microscope from fungal mycelia, chemically pre-cleaned, and then identified and photographed for documentation (Figure 5).

All moss species were given acid-alkali-acid pre-treatment consisting of 1% HCl (8–10 h just below boiling point) and 0.5% NaOH (1 h at 60°C). The insoluble fraction was washed, acidified to pH 3 and converted to graphite using a Fe-catalyst reaction. This was to ensure removal of humic/fulvic acids. Two sediment samples (Lyr-1 and Lyr-6) were given acid-alkali-acid pre-treatment consisting of 1% HCl and 1% NaOH (in both cases for 8–10 h just below the boiling point). The soluble part was precipitated by adding concentrated HCl and centrifuging at 3000 rpm. The precipitate consisted mainly of
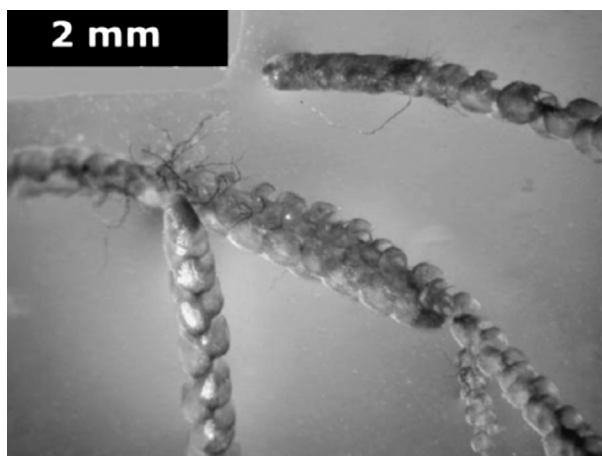
Figure 5 Microscope view of *Myurella julacea* sampled from the old vegetation layer at the glacier–bed interface below Longyearbreen

Table 1 Calibrated ages of bryophytes

Lab. No.	Material dated	^{14}C yr BP$\pm 1\sigma$	δ^{13}C ‰	cal. ^{14}C yr BP, 1σ	cal. ^{14}C yr BP, 2σ
Ua-16520	*SOL Lyr-6*	4090±70	−26.2	4810−4440	4830−4420
Ua-16380	*Tomentypnum nitens*	1875±50	−25.5	1880−1730	1930−1630
Ua-16396	*Tomentypnum nitens*	1625±40	−25.7	1570−1410	1690−1410
Ua-16377	*Sanionia uncinata*	1770±45	−24.7	1770−1600	1820−1560
Ua-16379	*Sanionia uncinata*	1515±40	−26.0	1510−1330	1520−1310
AAR-6532	*Moss* spp.	1670±45	−23.4	1690−1520	1700−1410
Ua-16375	*Myurella julacea*	1750±70	−25.1	1740−1550	1830−1520
Ua-16382	*Syntrichia ruralis*	1575±40	(−25.0)	1515−1415	1550−1350
Ua-16378	*Distichium cap.linc.*	1540±40	−22.9	1520−1350	1530−1330
Ua-16652	*Racomitrium ericoides*	1355±65	−25.2	1330−1180	1410−1130
Ua-16653	*Bryum pseudotriquetrum*	1350±45	−25.7	1310−1180	1340−1170
Ua-16651	*Pogonatum urnigerum*	1310±45	−25.5	1290−1180	1310−1140

humic acids and is labelled as SOL (soluble fraction). Before accelerator measurement the dried material (pH 4) was combusted to CO_2 and converted to graphite. The radiocarbon analyses were determined with the Uppsala EN-tandem accelerator (Possnert, 1990). A small fraction of approximately 0.05 mg carbon of the CO_2 gas was used for measurement of the natural mass fractionation, δ^{13}C, in a conventional mass spectrometer (VG OPTIMA).

A total of ten bryophyte subsamples and one SOL subsample were analysed from the subglacial samples. Usually, the bryophyte subsamples consisted of about 5–15 individual mosses in order to obtain sufficient material for graphitization. The radiocarbon ages reported in this paper are in radiocarbon years before present, expressed as yr BP. Calibrated years are given as cal. yr BP and were calculated using the OxCal v.3.8 software based on the INTCAL calibration data set.

Results and discussion

The ten calibrated ages of the bryophytes cover a time window from 1930 to 1130 cal. yr BP (95.4% probability) (Table 1). The ages of the different mosses cover a time window of about 800 years (Figure 6). The results of the calibrated ages show an explicitly continuous decrease in age with no hiatus.

Two additional samples of the bryophyte species *Sanionia uncinata* and *Tomentypnum nitens* were dated to examine the occurrence of species-specific ages. We assume that if species had a biased age, this would be observed by two different results, with one species type forming a cluster. In addition, these ages would differ from the age ranges of the other species. This is, however, not the case, since both moss species tested do not show a species-specific clustering, but distribute over the entire time window dated. In addition, the δ^{13}C values of the moss samples show apparent scattering and point to higher dependency on other parameters than to the species itself (Figure 7). Usually the δ^{13}C values are dependent on the type of plant, because of species-specific fractionation processes. The two *Tomentypnum* mosses show comparable stable isotope compositions, whereas the *Sanionia* mosses indicate varying δ^{13}C values. Different fractionation processes cannot explain the slightly different ages of *Tomentypnum*, whereas the *Sanionia* mosses indicate more complex fractionation process. The most probable explanation is long-term decomposition processes and longevity of the investigated mosses. In earlier studies it has been shown that decomposition processes led to depleted δ^{13}C values (Benner *et al.*, 1987). In addition, the apparently long time window of about 800 cal. yr BP suggests that the studied mosses decomposed over a long period of time. Using the three oldest-dated samples the probability is 95.4% for an initiation of vegetation growth at 1908 cal. yr BP. However, the site might have been covered with vegetation

several hundred years or even several thousand years before this. It is well-known that mosses are organisms with very slow growth rates (Furness and Grime, 1982). The growth rate becomes very slow in Arctic environments because of low temperatures and a low rate of decomposition (Molau, 1997; Sand-Jensen *et al.*, 1999). The mosses that were radiocarbon-dated in this study consist of ramets (physiologically independent individuals produced asexually by clonal growth) – the moss species grows at one end and decays and dies at the other. Therefore, the age of one moss species varies greatly because it consists of different ramets. A consequence of this is that clonal plants such as bryophytes might be much older than non-clonal plants, by approximately one order of magnitude (I.S. Jónsdóttir, personal communication, 2003). Clonal plants growing on Taymyr Peninsula in an Arctic environment have been estimated to have ages in excess of 3000 years (Jónsdóttir *et al.*, 2000). Taking these facts into consideration it cannot be ruled out that the mosses in the glacier forefield of Longyearbreen grew over hundreds or even thousands of years before being covered by the glacier. It is also necessary to take into account the radiocarbon date of the SOL fraction from the palaeosoil, with an age of 4830–4420 cal. yr BP; suggesting that the study site might have been ice-free since at least 4830 cal. yr BP. However, it is also well-known that clay minerals bind carbon, which can result in older humic acid ages than given by macrofossil ages of the same samples (McGechin *et al.*, 2001; Scharpenseel and Becker-Heidmann, 1992). Therefore, humic acid ages present serious problems resulting from various unknown sources. We believe that this potential
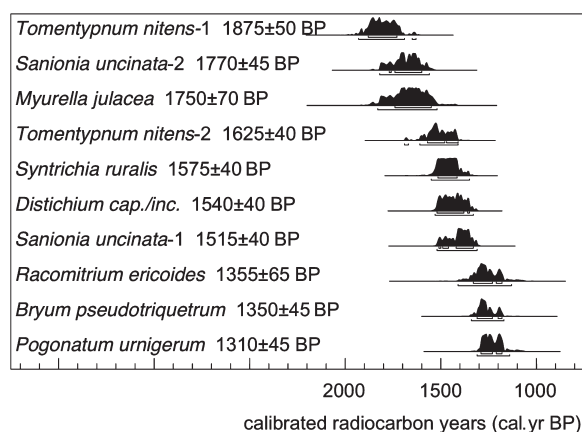
Figure 6 Radiocarbon dating results of the mosses from Longyearbreen glacier plotted on the x-axis of calibrated radiocarbon years. The histograms represent the probability distribution of the calibrated age. The calibration and the plot were done with the OxCal v 3.8 software (Bronk Ramsey, 2000) based on the radiocarbon calibration curve of Stuiver *et al.* (1998)
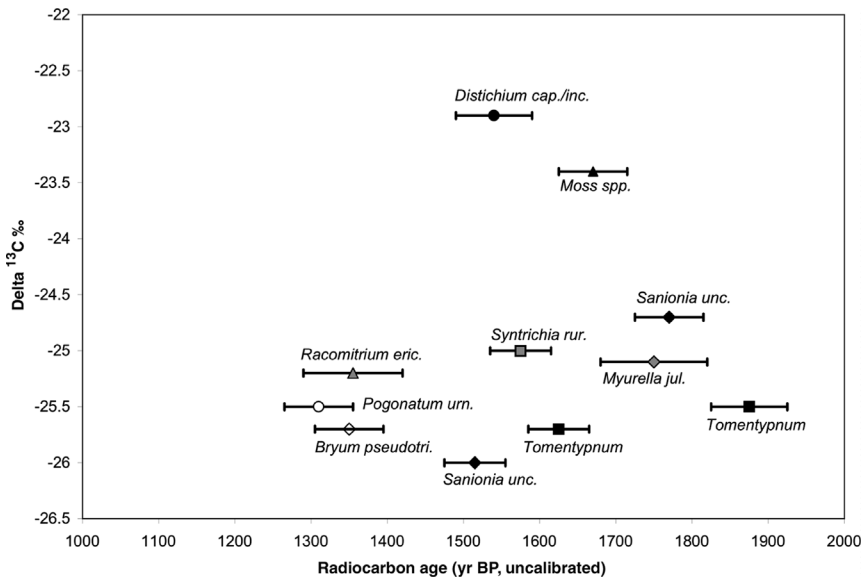
Figure 7 Stable carbon isotope composition ($\delta^{13}C$ ‰) of the moss species plotted on the *x*-axis of calibrated radiocarbon years. The age range is given with 2σ standard deviations that correspond to a probability of 95.4%. See text for discussion of longevity of moss species, which might explain the long time window of radiocarbon dates better than species-specific fractionation processes

contamination problem exists in this particular region and we are therefore presently not able to conclude that the study site was continuously ice free for the whole period between 4830 and 1104 cal. yr BP. Taking other evidence for Arctic climate development into account (see below), however, this appears very likely.

The period of vegetation growth ended at 1104 cal. yr BP (with a probability of 95.4%), taking into account the three youngest-dated samples (Figure 8). This is a minimum age for the advance of the Longyearbreen glacier over the study site, killing but not destroying the vegetation. Since then the glacier has advanced an additional 2 km downvalley, thereby lowering the altitude of the glacier terminus from about 450 m a.s.l. to 250 m a.s.l. At the time of the advance across the study site the glacier length was about 3 km. The present-day length is about 5 km.

A climatic perspective on glacier growth

The subglacial sampling site is about 2 km upstream from the present terminus of Longyearbreen. Since the advance at *c*. 1104 cal. yr BP, the glacier probably did not retreat again across the study site. Glacier retreat is a summer phenomenon, and meltwater activity along the glacier front would probably have destroyed the old vegetation and soil layer if it had been exposed by retreat.

Longyearbreen is a typical central Spitsbergen glacier with respect to topographic setting, aspect and size. There is no morphological or structural evidence suggesting past surge behaviour. In addition, based on our findings of subglacial *in situ* vegetation, a surge can be excluded for the last 1104 cal. yr BP; otherwise the mosses would have been destroyed. Most likely, the growth of Longyearbreen therefore represents a normal dynamic response to changes in air temperature, precipitation and changes in prevailing wind and the amount of drifting snow. Wind and snow drift are very important for modern glacier mass balance in the arid central part of Spitsbergen (Humlum, 2002). The late Holocene growth of Longyearbreen probably represents the local expression of a

more widespread trend for many Svalbard glaciers. In order to test this hypothesis and to place our findings into a larger spatio-temporal perspective, a brief review of palaeoclimate in Svalbard and adjoining Arctic regions is given.

Following the early twentieth-century warming mentioned above, most glaciers in Svalbard have experienced a net volume loss (Hagen and Liestøl, 1990; Liestøl 1993; Hagen *et al.*, 1993; Bruland and Hagen, 2002), adjusting to this new meteorological situation. Compared with this well-documented modern behaviour of Svalbard and other Arctic glaciers (see, e.g., Holmlund and Hagen, 1996), comparatively little was known
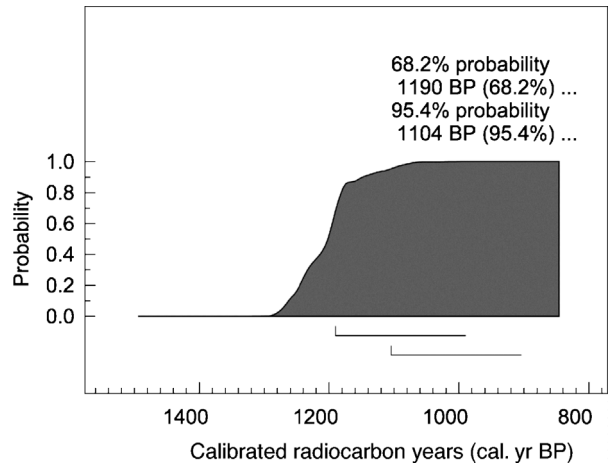
Figure 8 Probability plot (68.2% and 95.4%) of the maximum age of glacier Longyearbreen covering the study site. The calculation of the probability of being after all of the three youngest samples is given by:

$$r(t) = \prod_i \int_{t'=-\infty}^{t'=t} p_i(t')dt'$$

This is the distribution (normalized to a maximum of 1) returned by OxCal 3.8. From this a distribution $r'(t)$ can be calculated which gives a probability distribution for the last of the group of events:

$$r'(t) = dr(t)/dt$$

pre-twentieth century about Holocene glacier and climate dynamics in the High Arctic. Terrestrial micro- and macro-fossils have generally provided rather coarse data from the Svalbard region, and information about climatic change has proved difficult to interpret from observations on vegetation changes during the last 2000 cal. yr BP (e.g., Hyvärinen, 1970; 1972; Birks, 1991; Wohlfarth et al., 1995). From northwest Spitsbergen van der Knaap (1985) demonstrated a change towards lower temperatures during the eighteenth century. From around 200 cal. yr BP a somewhat warmer climate prevailed, before conditions subsequently turned colder again. There is, however, some concern that these results were based partly on data contaminated by human activity (Elvebakk, 1990).

Based on marine data, Salvigsen et al. (1992) and Salvigsen (2002) demonstrated warmer conditions in the Svalbard region during the early and mid Holocene, compared with the present-day climate. At that time, most Svalbard glaciers probably were smaller than at present or even absent. Based on lacustrine evidence from western Spitsbergen and a marine core in Billefjorden, Svendsen and Mangerud (1997) also demon-strated a warm early-Holocene climate, followed by a period of renewed glacier growth from 3000 to 2500 cal. yr BP. Furrer (1992, 1994) and Furrer et al. (1991) dated fossil organic horizons found beneath end moraines to ages between 1550 and 730 cal. yr BP (1010 ± 80 yr BP; 1315 ± 100 yr BP; 1465 ± 55 yr BP). These organic horizons were interpreted as maximum ages for glacier advances. They also dated organic horizons between 920 and 500 cal. yr BP (710 ± 60; 810 ± 65 yr BP) 640 and 290 cal. yr BP (370 ± 75; 485 ± 50 yr BP). The fact that these authors used unspecified organic material makes it difficult to estimate the validation of these dates. Contamina-tion problems of bulk sediment samples for radiocarbon dating are well-known and humic acids could account for reasonably younger ages, whereas old carbonates or coal could account for overestimated ages of the above-mentioned samples (Geyh et al., 1971; Hormes et al., 2004). Additional evidence for late-Holocene glacier growth on Svalbard was presented by Snyder et al. (2000). All the above findings are based on field evidence from moist coastal west Spitsbergen.

Marine evidence from the Nordic seas near Svalbard documents that the first half of the Holocene was the warmest period during the last 16 750–15 000 cal. yr BP (Koç et al., 1993), followed by cooling during the late Holocene (Koç and Jansen 1994). In the Barents Sea area just south of Svalbard, the early Holocene warming trend deduced from $\delta^{18}O$ records (Duplessy et al., 2001) was followed by rapid cooling during the late Holocene (Ivanova et al., 2002). The cooling was interrupted by at least two periods of warming at 2200 and 1600 cal. yr BP (Sarnthein et al., 2003). Jennings et al. (2002), working on the East Greenland Shelf, found evidence for the onset of late-Holocene cooling from c. 5500 cal. yr BP, associated with southward expansion of the Arctic sea ice and increased influence of polar water. Taking into account all marine records around Svalbard it has to be kept in mind that the marine reservoir effect and possible variations during the Holocene of coastal ocean streams is very little studied and usually a standard correction of 400 years is used for ^{14}C dating (Birkenmajer and Olsson, 1998; Feyling-Hanssen and Olsson, 1959–1960).

In Fennoscandia, south of Svalbard, oxygen isotopes of lacustrine carbonates of Lake Tibetanus in Lapland suggest a cooling trend after 3000 cal. yr BP, but interrupted by a period of increased mean July temperatures and decreased annual precipitation, peaking around 1000 cal. yr BP (Hammarlund et al., 2002). Studies of pollen assemblages from northern Finland indicate a marked late-Holocene summer (July) temperature decline, and suggest the last 2000 years to be the coolest since the early Holocene (Seppä and Birks, 2001). From diatoms, chironomids, pollen and near-infrared spectroscopy at an alpine lake in northern Sweden, Rosén et al. (2001) documented a late-Holocene lowering of the tree-limit and gradual lowering of the mean July temperature during the last 7000 cal. yr BP. In southern Norway Blikra and Selvik (1998), mapping snow avalanche deposits, demonstrated that the Holocene was characterized by highly fluctuating winter climatic conditions throughout, but that a development towards deteriorating climate has prevailed for the last 5400 cal. yr BP. In western Norway, Nesje (1992) found that valley and cirque glaciers in the Jostedalsbre region reached their greatest Neoglacial extent during the mid-eighteenth century. Elven (1978) found plant remains previously covered by the Omsbreen glacier in southern central Norway, much as is presently the situation at Longyearbreen. He concluded that Omsbreen most likely was established around 550 cal. yr BP, reached its maximum size (14 km^2) during the eighteenth century, and underwent significant retreat during the twentieth century. Heikkila and Seppä (2003) presented a temperature reconstruction based on high-resolution pollen stratigraphy from southern boreal Finland. Their results show that Holocene air temperatures reached their highest values 8–4.5 ka BP, and have gradually decreased by c. 1.5°C since, but indicate 0.5°C higher temperatures between 1200 and 1100 cal. yr BP.

In northeast Greenland, only 400 km west of Svalbard, field evidence suggests ice-free fjords during the early and mid-Holocene, followed by extensive periods with ice-covered fjords from 5700 cal. yr BP (Hjort, 1997). Kelly (1980), summarizing available glaciological evidence, concluded that the western margin of the Greenland Ice Sheet retreated to a position 60–120 km behind the present margin during the early Holocene. A late-Holocene period of growth was initiated 3000–3500 cal. yr BP, followed by major advances around 2000 cal. yr BP. This development is also reflected by ice core analysis from the Greenland ice sheet (Dansgaard et al., 1971; 1973). In addition, the establishment and growth of Hans Tausen Ice Cap (now 4208 km^2) in Pearyland, northernmost Greenland, from 3900 cal. yr BP (Hammer et al., 2001), lend support to this overall late-Holocene climatic development in the Green-land–Svalbard sector of the Arctic. Reconstructed surface temperatures from the Greenlandic Ice Sheet (Dahl-Jensen et al., 1998) directly testify to c. 3°C net cooling since 4000 cal. yr BP. Kaplan et al. (2002) reported Neoglacial cooling from southeast Greenland from 3000 cal. yr BP, and Fredskild (1973) reports botanical evidence of a sharp change in climate towards colder and drier conditions around 1550 cal. yr BP in west Greenland.

On Nordvestø in the Carey Øer group between Greenland and Ellesmere Island, an extensive deposit of peat developed between approximately 6300 and 4400 cal. yr BP at a site where this type of moss no longer grows (Brassard and Blake, 1978). From Ellesmere Island, Arctic Canada, Blake (1989) reported the observation of a lake basin that has been frozen to the bottom since about 5800 cal. yr BP, while the lake probably was completely open during the preceding warmer early Holocene. In the Canadian Arctic Archipelago, observations indicate that there was an increased frequency of driftwood between 6500 and 4500 cal. yr BP (Blake, 1972; Steward and England, 1983), suggesting diminished Arctic sea ice cover or, alternatively, that the sea ice on which the wood was carried at that time had more mobility, as was suggested from observa-tions in the Svalbard region by Häggblom (1982, 1987).

Thus, the significant net growth of Longyearbreen since 1104 cal. yr BP apparently is consistent with a widespread late-Holocene climatic development towards cooler conditions in the Arctic. This may explain why the Little Ice Age (LIA) glacier advance in Svalbard usually represents the Holocene maximum glacier extension and that little evidence for older Holocene ice-marginal morphology is found. From a glaciological point of view, the climatic sensitivity of glaciers in the central arid regions of Spitsbergen would normally be considered small compared with that of glaciers located in more maritime regions with corresponding higher mass turnover, basal sliding and higher ice-flow velocities. Therefore, we consider that the ~ 66% increase in length from 3 km to now 5 km, of Longyearbreen during the last *c.* 1100 years is quite remarkable, testifying to a climatic development towards conditions more favourable for glaciers at the gateway to the Arctic Ocean.

Cold-adapted microbes

In addition to the above analysis, the presence of the fossil soil with *in situ* plants below Longyearbreen provided us with a rare opportunity for evaluating microbial survival capacity of an entire community in a subglacial, permafrozen state within the time frame given by the ^{14}C-datings of the vegetation.

Temperature is considered one of the most important factors regulating micro-organisms, controlling their physio-chemical reactions (Herbert, 1986). Microbes, which are thought to be the earliest life-forms on this planet, occur in all types of environments including air, water and soil and have adjusted to cold environments, taking up large surface areas of the planet. Indeed, more than 70% of the planet is covered by oceans, of which about two-thirds (by volume) has a temperature of 2°C or less. Because of their ability to adjust to low temperatures, microbes can be regarded as being among the planet's most successful colonizers (Russel, 1990).

Forster (1887) was the first to isolate cold-adapted microbes in dead fish preserved by low temperatures. Since then, such microbes have been found in many types of cold environment and extensive reviews have been presented by Morita (1975), Herbert (1986), Gounot (1986), Russel (1990), Friedmann (1994) and Gilichinsky and Wagner (1995). Among other environments, viable microbes were found in permafrost as early as the early twentieth century by Russian scientists investigating Siberian mammoth localities (Omelyansky, 1911) and in buried soils in the Far East (Isachenko, 1912). In Canada, James and Sutherland (1942) found viable aerobic and anaerobic microbes in permafrost, at depths of 2–3 m at Churchill, Manitoba. Becker and Volkmann (1961) isolated different viable microbes from permafrost cores taken 6–18 m below the terrain surface near Fairbanks, Alaska, having an age of 20 ka BP and 70 ka BP, respectively. In the Antarctic, Cameron and Morelli (1974) in the Dry Valley region found viable microflora from permafrost deposits up to 1 million years of age, which have led to suggestions for addressing exobiological problems (e.g., Mars) by using terrestrial permafrost analogues. Later, Zvyagintsev *et al.* (1985), Gilichinsky *et al.* (1990; 1992) and Khlebnikova *et al.* (1990) demonstrated that large numbers of viable microbes (up to 100 million/g of soil) probably have been preserved in a viable state for several millions of years. Upon thawing, these microbes resumed their physiological activities; a remarkable feat considering the low temperature (− 12°C) and the probable great age. Apparently, microbes in permafrost regions tend to accumulate at the bottom of the active layer, as the highest number of microbes is

found in the lower part of summer thawing layer in modern tundra soils (Gilichinsky, 1995). Also at the bottom of the previous (7000–9000 cal. yr BP) thicker early Holocene active layer there is a horizon with an increased number of viable microbes in what since has been permafrost (Gilichinsky, 1995).

A fundamental problem in cryobiology is to determine under which environmental conditions microbes can survive long-term freezing (in an inactive state) and how metabolically active microbes may have been during freezing and, in particular, after thawing. This motivated us to investigate if the permafrozen subglacial soil below Longyearbreen contained viable microbes and, if so, how metabolically active (in terms of CO_2 production) such microbes would be following a major change in environmental conditions (thawing).

Analysis of the subglacial soil

Carbon dioxide (CO_2) production and its release from soil has long been used as a measure of microbial activity and has been shown to be sensitive to changes in environmental factors such as temperature and moisture (Kirschbaum, 1995; Fang and Moncrieff, 2001). Arctic soils hold large active organic carbon (C) reserves in the active layer as well as inactive reserves in permafrost layers (Post *et al.*, 1982). The release of CO_2 from such soils may therefore exert a potentially positive feedback on atmospheric CO_2 concentrations (e.g., Oechel *et al.*, 1993). This may be enhanced if buried soils become exposed as a result of increased active-layer thickness, thermocast processes or by the retreat of cold-based glaciers. In the laboratory we therefore investigated how incubated soil samples collected from the fossil soil beneath Longyearbreen responded to thawing, with a focus on the microbial production of CO_2 at temperatures between 0 and 7°C. In addition, the results were compared with similar observations from a nearby subaerial reference soil.

Frozen samples (*c.* 500 cm[3]) were collected at two sites in early May 2002. One site was the study site below Longyearbreen as described above, where soil samples were collected after removing the fresh exposed soil that might have been in contact with meltwater the previous summer. Thus, samples used for analysis have been exposed only briefly to the atmosphere as part of the sampling process and have presumably been maintained in a permafrozen state since at least 1104 cal. yr BP. The latter assumption is based on the fact that no evidence of past warm-based conditions at the glacier sole was observed. The control site has been subject to freezing and thawing on an annual basis, as has the now-subglacial soil before the glacier advance. Sampling at the control site was also undertaken in early May 2002, at a time when the soil was still frozen following the winter 2001–2002.

At both study sites (Longyearbreen and Bjørndalen) samples were collected within the upper 3 cm of the soil profile (A-horizon material only) as frozen fragments. Sampling was carried out using sterilized equipment and samples were kept frozen (below − 4°C) and stored in darkness until analysed in the laboratory. Soil pH was measured in distilled water at a soil-solution ratio of 1:1. Total organic carbon (TOC) contents were measured using a Total Organic Carbon Analyser (Dohrmann DC-190) after removal of carbonates by a weak acid (HCl). Finally, soil samples were analysed for volume, weight, porosity and solid bulk density.

Basal soil respiration (BSR) measurements were measured in subsamples that had been stored in polyethylene bags at temperatures below − 4°C. Prior to measurement, soil samples were split carefully to remove roots and stones and gently mixed. Weighed soil samples (five replicates, each equivalent to

2–3 g dry soil) were transferred to 12 ml Venoject tubes and left for pre-incubation for at least 48 h at stable temperatures equal to the experimental temperatures. Prior to measurement, tubes with soil samples were out-gassed with CO_2-free synthetic air. BSR measurements in depth-specific soil samples were subsequently taken by monitoring the linear ($R^2 > 0.95$) increase of headspace CO_2 concentrations at 7°C over two to three days by gas chromatography using a Mikrolab ML GC82-12 equipped with a Porapack Q-column kept at 30°C and a TCD detector operated at 200°C (flow rate of 25 ml He/min). The temperature dependence of BSR was investigated by incubating Venoject tubes with soil (three replicates equivalent to 2–3 g dry soil) at 0, 4 and 7 ± 0.3°C. Because of the high water content in soil samples collected below Longyearbreen, additional BSR measurements (measured at 7°C only) were made on samples from Bjørndalen after increasing the water content to the same level as observed for samples taken from the site below Longyearbreen. Sterilized samples were included after killing the microbes by chloroform addition to the soil sample (Elberling *et al.*, 2000). After 2 h the chloroform was removed by vacuum evaporation followed by BSR measurement. Substrate-induced respiration rates (SIR) were made after gently mixing soil material with a powder mixture containing glucose (1.2 mg glucose-C/g soil) and talc (Anderson and Dosch, 1978). Control incubations with empty Venoject tubes showed that CO_2 diffusion into the tubes could be neglected.

Results of soil analysis

Findings of similar plant remains in all soil samples suggest that both soil study sites have been covered by similar vegetation, mainly mosses. The soil organic carbon content was 3.2 ± 1.5% and 2.7 ± 1.9% by weight for Bjørndalen (B-site) and Longyearbreen (L-site), respectively. Soil pH values were 5–6 for all samples. The water content was 32–33% by weight at the L-site and 20–25% at the B-site.

Measurements revealed that ongoing microbial soil respiration (BSR) was measurable at both sites (Figure 9). CO_2 production rates (per weight dry soil) from soil collected at the L-site were more than ten times the rates observed from sterilized samples and about 10% of rates observed at the present-day B-site. This result strongly suggests that microbial life has survived since the soil was covered by the advancing Longyearbreen no later than 1104 cal. yr BP, and that such soil microbes can again be active after few days at above-freezing temperatures. Respiration rates observed at 0, 4 and 7°C reveal that the temperature sensitivity of CO_2 production can be fitted by a exponential function ($R^2 = 0.9$, $p < 0.005$), indicating a doubling of the rates for every 10°C increase ($Q_{10} = 2.1$). This is consistent with other studies (see review by Fang and Moncrieff, 2001). No significant differences in terms of temperature sensitivity were noted for the two Svalbard sites.

The lower rates observed for the subglacial L-site compared with the modern B-site can be related partly to higher water contents at the L-site (near-full saturation), as indicated by a 44% reduction in rates at the B-site after addition of water (Figure 9). High water content reduces the availability of oxygen and thereby results in reduced soil CO_2 respiration (Fang and Moncrieff, 2001). Taking into account the higher content of organic C in samples from the B-site relative to the L-site further reduces the differences in CO_2 production in terms of CO_2 production per unit soil-C. The CO_2 production increased in all samples after glucose addition. In samples from the modern B-site the production increased by a factor of 1.90 while samples from the subglacial L-site increased by a factor of 1.31, indicating the presence of a living biomass (microbes)
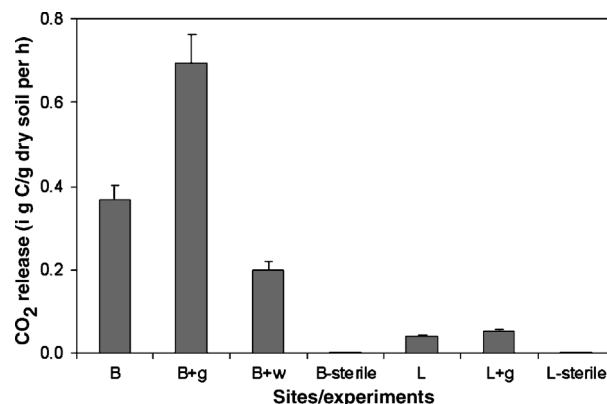
Figure 9 CO_2 release from buried soil exposed under Longyearbreen (L) and a nearby reference site Bjørndalen (B) as well as after glucose addition ($+g$), water addition ($+w$) and after sterilization. All rates are measured at 7°C. One standard deviation ($n = 5$) is shown as a vertical line

at both sites, although reduced at the L-site compared with the B-site. Results also reveal that C-substrate at both sites at least partly limits the observed CO_2 rates.

We conclude that, taking site-specific conditions such as water content into account, the average microbial respiration rate as observed in the 1104 cal. yr BP soil samples (L-site) is roughly equivalent to 20% of rates observed at the present-day B-site. Thus, microbes are not only able to survive for centuries below a cold-based glacier, but they also retain a remarkable capacity to decompose organic matter compared with present-day subaerial microbial communities.

Although the temperature sensitivity of microbial respiration below 0°C is uncertain, in particular the lower limit for activity, more recent studies from Arctic soils in Greenland suggest that Q_{10} values observed above 0°C to some extent can be extended to below-zero conditions, depending on the availability of free water (Elberling and Brandt, 2003). Applying the temperature sensitivity observed in this study ($Q_{10} = 2.1$) and assuming a rate of consumption of 0.04 µgC/g per h (at 7°C) the microbial respiration since the glacier advance can be calculated if ground temperatures experienced during the period are estimated. Presently the temperature at the glacier bed is about -4°C, and we assume that temperatures may have been somewhat lower during LIA cold periods, even though variations in glacier thickness and dynamics also represent important controls on the subglacial temperature regime. The lack of glacial abrasion at the glacier bed, however, suggests that the glacier has remained non-sliding and cold-based at the study site.

A simple mass balance calculation assuming a low average temperature of about -8°C since the site was covered by the advancing glacier, suggests a carbon loss equal to 0.19 g C/g soil, using 1100 years as an approximate time period. This is more than the expected original carbon content in the soil before the glacier advance (0.03–0.05 g/g soil). This discrepancy suggests that the microbes recovered from below Longyearbreen may have been periodically subject to temperatures below the limit for activity, at least for part of the period, but that they still were able to survive until now. Unfortunately, given the present limited state of knowledge on microbial respiration at sub-zero temperatures, our data do not allow us to differentiate between uncertainties related to the actual temperature regime experienced at the subglacial study site and uncertainties resulting from microbial activity at temperatures below 0°C.

Conclusions

- In High Arctic Svalbard, a subglacial exposure of undisturbed palaeosoil and vegetation has been found below cold-based glacier Longyearbreen. Dating demonstrates that the study site was covered by the advancing glacier no later than *c.* 1104 cal. yr BP. Before that, the site was ice-free for at least 800 years and possibly much longer.
- Cross-checking the radiocarbon dating of different moss species sampled at the glacier–bed interface showed that no species-specific ages occur and therefore, different fractionation processes can be excluded. The radiocarbon ages range over a long time window of about 800 years and point to the slow growth, clonal character and decreased decomposition of the mosses.
- Measurements revealed that microbial life has survived in a metabolically active state in the permafrozen, subglacial soil since *c.* 1104 cal. yr BP.
- Soil microbes are not only able to survive for centuries below a glacier in permafrozen state but they are also able to maintain a remarkable capacity to decompose organic matter compared with present-day microbial communities, following only few days at above-freezing temperatures.

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