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Glacial and postglacial sedimentation in the Fryxell basin, Taylor Valley, southern Victoria Land, Antarctica

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Abstract

A 9.14 m long sediment sequence was recovered from Lake Fryxell, Taylor Valley, southern Victoria Land, Antarctica, and investigated for its chronology and sedimentological, mineralogical, and biogeochemical changes. The basal part of the sequence is dominated by coarse clastic matter, i.e., mainly sand. The sediment composition suggests that a lake existed in Fryxell basin during the Middle Weichselian by ca. 48,000 cal. year BP. After a short period of lake-level lowstand ca. 43,000 cal. year BP, lower Taylor Valley became occupied by the proglacial Lake Washburn, which was at least partly supplied by meltwater and sediments from the Ross Ice Sheet that was advanced to the mouth of Taylor Valley. Evaporation of Lake Washburn to lower levels started during the Last Glacial Maximum at ca. 22,000 cal. year BP, long before the Ross Ice Sheet retreated significantly. Lake-level lowering was discontinuous with a series of high and low stands. From ca. 4000 cal. year BP environmental conditions were similar to those of today and lower Fryxell basin was occupied by a small lake. This lake evaporated to a saline or hypersaline pond between ca. 2500 and 1000 cal. year BP and refilled subsequently.

Keywords: Lake Fryxell; Lake Washburn; Late Quaternary; Paleolimnology

1. Introduction

Taylor Valley in southern Victoria Land, Antarctica, has a number of closed drainage basins, of which the easternmost contains Lake Fryxell. This lake is supposed to be a remnant of a much larger proglacial lake; Lake Washburn (Stuiver et al., 1981; Hendy, 2000a). Evidence for the existence of this proglacial lake comes from relict deltas and shorelines, and from lacustrine or glaciolacustrine sediments in the catchment of the modern lakes. The lacustrine conditions are documented by the occurrence of non-marine diatom assemblages in these deposits (Kellogg et al., 1980; Brady, 1982).

The relict shorelines and deltas along the valley slopes indicate the morphology of Lake Washburn. On the walls of central and eastern Taylor Valley, several strandlines extend up to 336 m above sea level (a.s.l.) (Hendy, 2000a) and indicate that Lake Washburn was

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more than 300 m deep at its greatest extent. Even higher deposits from proglacial lakes have been reported from adjacent Wright and Victoria Valleys (Hall et al., 2001, 2002).

Radiocarbon dating of preserved algae and uranium–thorium dating of lacustrine carbonates indicate that Lake Washburn was present during the Last Glacial Maximum and into the Holocene (Hall and Denton, 2000; Hall, 2003). Large lake level fluctuations have been common, but their detailed reconstructions are sparse (Stuiver et al., 1981). The meltwater that has fed Lake Washburn during the Late Weichselian is supposed to originate from an advanced grounded ice sheet that blocked the outlet of Taylor Valley (Hendy, 2000a). Distinct lowering of the water level of Lake Washburn after the Last Glacial Maximum has been attributed to the retreat of the grounded ice sheet and to climatic changes (Hall and Denton, 2000; Whittaker et al., 2003). Both climatic and glacial oscillations apparently affected the present-day lakes in Taylor Valley in their younger history and led, for example, to a desiccation of Lakes Hoare and Fryxell prior to ca. 1200 ^{14}C year BP (Lyons et al., 1998).

The reconstruction of the climatic and environmental history of southern Victoria Land is based on several records, but they do not provide a consistent history. The isotopic record from the nearby ice sheet at Taylor Dome correlates well with the Byrd and Vostok ice core records during the Middle and Late Weichselian (Grootes et al., 2001), but indicates regional peculiarities at the Pleistocene/Holocene transition (Steig et al., 1998). A comparison between temperature changes deduced from isotopic changes in the Taylor Dome ice core record (Masson et al., 2000), glacier advances and retreats in the Dry Valleys inferred from geomorphic features (e.g., Higgins et al., 2000), ice shelf advances and retreats reconstructed from marine sediments (Leventer et al., 1993; Cunningham et al., 1999), and the occurrence of penguin rookeries in the region (Baroni and Orombelli, 1994) shows a complex picture of Holocene changes, which suggests that local alpine-type glacier advances and retreats are controlled by different reasons than those of the ice shelf (Doran et al., 1994a). For example, local alpine-type glaciers are controlled predominantly by precipitation (e.g., Higgins et al., 2000), whilst the advance of the Ross Sea Ice Sheet is apparently controlled more by global sea-level lowering (B. Hall, personal communication). This has resulted in the out of phase relationships seen in Taylor Valley (e.g., Doran et al., 1994a).

The modern ecology of the lakes in Taylor Valley, their physical and biogeochemical processes, and the links between climate changes and their impact on the lakes and their environments have extensively been studied (e.g., Wharton et al., 1983; Priscu, 1998; Doran et al., 1999; Lyons et al., 1999; Neumann et al., 2001; Lawson et al., 2004). Present bottom water quality in the lakes varies from entirely freshwater (Lake Hoare) to brackish (Lake Fryxell) to hypersaline (Lake Bonney). These geochemical differences can be traced back to lake-level changes and variations in lake ice cover in response to climate change (Wilson, 1964; Chinn, 1993; Lyons et al., 1998; Fountain et al., 1999; Doran et al., 2002).

Despite the excellent knowledge about the recent conditions in the lakes and their interactions with the environment, relatively little has been published about their histories using their sedimentary records. Most of the existing investigations focus on the chronology (Doran et al., 1999) and/or are restricted to the near-surface sediments to about 1 m depth (Doran et al., 1994a; Hendy, 2000a). Longer sedimentary records have recently been recovered (Whittaker et al., 2003), but detailed information extending deep into the deposits of proglacial Lake Washburn or earlier stages is so far lacking. In order to obtain continuous, high-resolution information about the environmental conditions prior to or during the existence of proglacial Lake Washburn, its subsequent shrinking, and the transition to the modern lakes, a long sediment sequence was recovered from Lake Fryxell in 2002 and subsequently investigated using chronological, sedimentological, biogeochemial, and mineralogical methods. The information obtained complements the paleoclimatic and paleolimnologic information existing so far and shed further light on the environmental history of the region.

2. Study area

The McMurdo Dry Valleys in southern Victoria Land (Fig. 1), with a combined area of approximately 4800 km^2, is the largest ice-free area in Antarctica. Drainage basins in Taylor Valley contain several lakes with permanently unfrozen water beneath a perennial ice cover. Lake Fryxell is located at an altitude of ca. 15 m a.s.l. (Hendy, 2000a; Lawson et al., 2004) in the easternmost drainage basin (Fig. 1). Canada Glacier separates Lake Fryxell from Lake Hoare located further upvalley. Branches of the Asgard Range and the Kukri Hills, elevated up to several hundred meters above sea level and covered by alpine glaciers, delimit the catchment of Lake Fryxell towards the north and south. Towards the east, where Taylor Valley meets

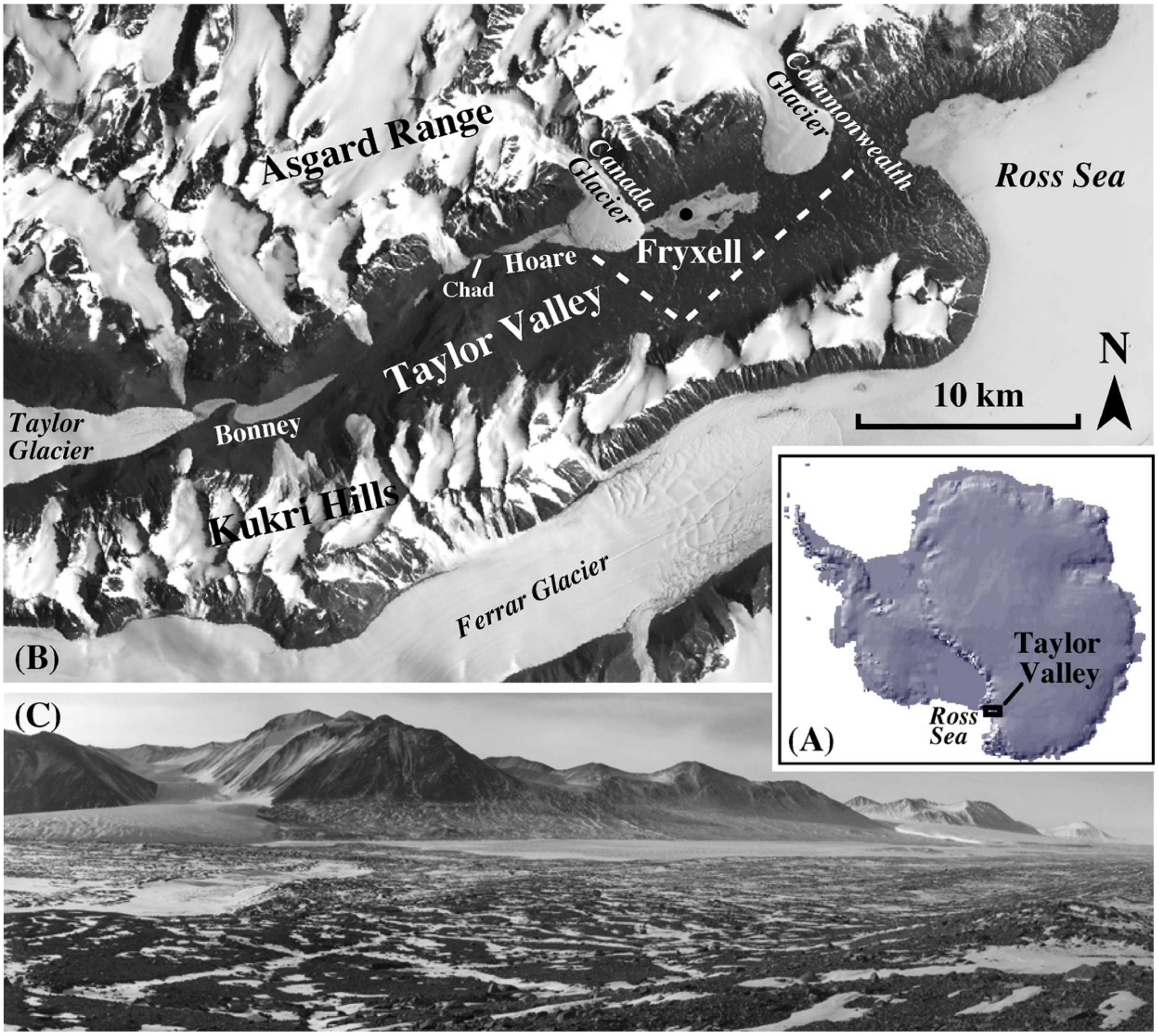

Fig. 1. (A) Location of Taylor Valley in Antarctica. (B) Satellite image of Taylor Valley, southern Victoria Land, Antarctica, indicating the location of most important lakes and glaciers. The black dot in Lake Fryxell indicates the coring location Lz1021. Note that (B) is rotated by 180° in comparison with (A). The dashed lines indicate the perspective of this figure (C). (C) Photography of Fryxell basin with view from the south. Canada Glacier is located at the left side, Commonwealth Glacier and the outlet of Taylor Valley at the right side, and Lake Fryxell in the center.

the Ross Sea, the Commonwealth Glacier and a saddle, 78 m a.s.l., border the catchment.

Lake Fryxell has a surface area of 7.1 km^2. The perennial lake-ice cover floats almost exclusively eolian sediments today and has an average thickness of 3.3–4.5 m. Ice-free moats form around the edge of all Dry Valley lakes during summer. The bathymetry of Lake Fryxell is characterized by very gentle slopes with a maximum depth of only 20 m (Lawson et al., 2004). Water temperatures are between just above 0 °C at the surface and ca. 2.5 °C in the bottom waters. A circumneutral pH exists throughout the water column. Waters are relatively fresh and super-saturated with respect to dissolved oxygen near the surface, and a brackish and anoxic monimolimnion is formed below ca. 9.5 m depth (Lawson et al., 2004). Lake Fryxell is fed by meltwater from local glaciers both directly (Canada Glacier) and via glacial-fed streams (89% during the 2001/2002 summer season, Ebnet et al., 2005) during the short austral summer. An outlet does not exist today, but a relict stream channel draining both ways across the saddle to the east of Lake Fryxell implies that at times the lake overflowed towards the Ross Sea, and, at other times, the lake was fed by an inlet from the east (Hendy, 2000a). The bedrock in the catchment is dominated by early Palaeozoic granitoid plutons, gabbros, and lamprophyre dykes, which are referred to as Granite Harbour Intrusive Complex, and metasediments of the Skelton Group (Laird and Bradshaw, 1982; Smillie, 1992). Alkali volcanic complexes assigned to the McMurdo Volcanic Group are only sporadically exposed upvalley, but are dominant in the southern and eastern parts of the McMurdo Sound, including Ross Island (LeMasurier and Thomson, 1990). Quaternary sediments overly the bedrock in the lower parts of Taylor Valley (Stuiver et al., 1981).

The recent climate conditions in the Dry Valleys are characterized by mean air temperatures of −15 to −30 °C, with summer temperatures rising above 0 °C and winter temperature as low as −60 °C (Thompson et al., 1971; Doran et al., 2002). The precipitation is almost exclusively snow and amounts to <100 mm/year water equivalent (Bromley, 1985). This value is well below the ablation rates, which have been determined to 150–600 mm/year (Lyon, 1979; Clow et al., 1988). Main wind directions are easterly from the Ross Sea during summer and westerly from the ice sheet during the rest of the year. Changes in the wind regime have a distinct impact on the temperatures and precipitation (Doran et al., 2002). The low precipitation in combination with dry katabatic winds descending from the ice sheet result in extremely dry conditions (Clow et al., 1988). Sensitivity of the lake levels to climate change has been seen in lake level records over the past several decades (Wharton et al., 1992).

3. Material and methods

Sediment coring was carried out in November 2002 through two holes in the lake ice cover, ca. 3 m separated from each other, and close to the deepest part of Lake Fryxell at 18.3 m water depth (77°36.63′S, 163°08.39′E; Fig. 1). A continuous sediment sequence (Lz1021) was recovered by the deployment of a gravity corer for undisturbed near-surface sediments and a 3 m long piston corer (both UWITEC Corp.) for deeper sediments. In order to prevent disturbance of sediment structures by flushing water, the cores were frozen in vertical position immediately after their recovery. Then, they were cut into pieces of up to 150 cm length and transported to McMurdo Station, where the magnetic susceptibility (MS) was measured in 1 cm steps on the frozen cores using a Bartington MS2C sensor with an inner diameter of 72 mm.

In the laboratory, the cores were cut lengthwise into halves at frozen conditions using a stainless steel band saw and carbide-tipped blade. After cleaning the sediment surface, the cores were described, photographically documented, and optically correlated. One of the halves was then cut again into two quarters, one of which was used for subsampling in 2 cm intervals throughout the cores. The subsamples were freeze-dried and their water contents determined by the loss of weight.

For grain-size analyses aliquots of the subsamples were pre-treated with 10% acetic acid (CH_3COOH) and 3% H_2O_2 in order to remove carbonate and organic matter. Gravel and sand contents were determined by sieving. Silt and clay contents were quantified after separation in settling tubes.

Another set of aliquots was ground to <63 μm and homogenized. These samples were used for the measurement of grain density and for biogeochemical and mineralogical analyses. The grain density was determined with a Micromeritics Accupyc 1330 pycnometer. Total carbon (TC), total nitrogen (TN), and total sulphur (TS) contents were measured with an Elementar III (VARIO Corp.) analyser. Total organic carbon (TOC) contents were measured with a Metalyt CS 1000S (ELTRA Corp.) analyser, after pretreating the sediment with 10% HCl at 80 °C in order to remove the carbonate. Total inorganic carbon (TIC) was calculated from the difference in TC and TOC. Mineralogical analyses of the bulk sediment were performed at a MiniFlex X-ray diffractometer (Rigaku Corp.) with CoKα radiation (30 kV, 15 mA). After adding an internal standard of Corundum (Al_2O_3) at a sample/standard ratio of 5:1, random powder mounts were X-rayed from 3 to 40° 2Θ with a step size of 0.02° 2Θ and a measuring time of 2 s/step. Evaluation of the diffractograms followed a method described in detail in Neumann and Ehrmann (2001).

For enumeration of volcanic glass and brownish and greenish hornblende of selected samples, the heavy minerals in the fine sand fraction (63–125 μm) were separated from the light minerals in a centrifuge with a sodium-metatungstate solution. The heavy minerals were fixed with Meltmount on glass slides. In this fraction, volcanic glass concentrations, i.e., pure glass or glass with heavy mineral intrusions, and brownish and greenish hornblende were identified under a polarising microscope, counting at least 400 grains. The glass concentrations are presented as grain percentages in the total heavy mineral fraction. Grain percentages were also used to calculate the ratio between brownish and greenish hornblende.

Radiocarbon dating was conducted by accelerator mass spectrometry (AMS) at the Leibniz Laboratory for Radiometric Dating and Isotope Research in Kiel, Germany. Since macrofossil remains were lacking throughout the core, 24 sediment samples were selected for bulk TOC dating. These samples had elevated amounts of TOC, however, seven measurements failed, either because TOC contents were too low to produce sufficient gas amounts or TS amounts were too high thus hampering complete reduction of the samples. The ages obtained by dating the humic acid free fraction (HAF) on most of the samples were constrained by dating the humic acid fraction (HA) on only a few samples. In order to compare the sedimentological and biogeochemical data from core Lz1021 with information provided from other records nearby, radiocarbon ages were calibrated into calendar years before present (cal. year BP; Table 1). For this purpose the recent reservoir effect in Lake Fryxell was estimated by dating the near-surface sediment and subtracted from the radiocarbon ages to obtain reservoir corrected ages (corr. year BP). The corrected ages were then calibrated using CALIB rev 5.0 and the SHCal04 and INTCal04 data sets (Stuiver and Reimer, 1993; McCormac et al., 2004; Reimer et al., 2004). Means and uncertainties of the calendar ages were calculated from the upper and lower boundaries of the probability distribution at the 2σ level, and a polynomial function was used to describe the age–depth correlation. In addition, published radiocarbon dates from the region given in ^{14}C year BP were calibrated for better comparison. Calendar ages are used in order to be able to compare our data to the Taylor Dome ice core record (Steig et al., 2000). We recognize that variability in the lacustrine carbon reservoir over time may cause the calendar ages to

Table 1
Radiocarbon dates of the humic acid free fraction (HAF) and the humic acid fraction (HA) from bulk sediment samples of core Lz1021, Fryxell basin

Sample	Core	Corr. depth (cm)	Weight (g)	Material	C (mg)	δ^{13}C (‰)	^{14}C age (^{14}C year BP)	Corr. age (corr. year BP)	Cal. age (cal. year BP)
KIA24525	Lz1021-1	0–2	1.0	HAF	8.5	−30.87	2475±40	0±0	65±70
				HA	2.5	−28.53	2670±30	195±30	145±145
KIA24526	Lz1021-1	4–6	1.0	HAF	5.2	−29.89	3020±35	545±35	525±30
KIA26079	Lz1021-1	6–8	1.0	[a]					
KIA24527	Lz1021-1	10–12	3.0	HAF	3.1	−27.15	5230±50	2755±50	2820±80
				HA	2.1	−27.39	4800±35	2325±35	2260±105
KIA24533	Lz1021-1	14–16	3.0	HAF	1.7	−24.16	4880±35	2405±35	2400±90
				HA	1.5	−25.86	5115±40	2640±40	2640±150
KIA24528	Lz1021-1	22–24	3.0	HAF	13.7	−30.30	4885±40	2410±40	2500±190
KIA24532	Lz1021-1	26–28	3.0	HAF	7.9	−27.43	5435±25	2960±25	3060±110
				HA	3.4	−28.59	5510±35	3035±35	3140±140
KIA24529	Lz1021-1	30–32	3.0	HAF	3.5	−25.40	6165±35	3690±35	3960±120
				HA	2.1	−27.82	6170±40	3695±40	3960±120
KIA26418	Lz1021-2	48–50	3.0	[a]					
KIA26080	Lz1021-2	50–52	3.0	[a]					
KIA25876	Lz1021-2	56–58	3.0	HAF	0.8	−26.44	14,630±110	12,155±110	14,090±340
KIA24531	Lz1021-2	68–70	3.0	[a]					
KIA26081	Lz1021-2	72–74	3.0	[a]					
KIA24530	Lz1021-2	84–86	3.0	HAF	0.6	−21.03	14,660±220	12,185±220	14,290±660
KIA24534	Lz1021-3	115–117	3.0	HA	0.7	−27.23	11,250±90	8775±90	9840±310
KIA25127	Lz1021-3	117–119	3.0	HAF	0.6	−29.08	14,070±180	11,595±180	13,460±340
KIA26419	Lz1021-3	147–149	3.0	HAF	0.5	−25.62	14,150±230	11,675±230	13,560±440
KIA26420	Lz1021-4	205–207	3.0	HAF	0.8	−18.80	18,420±260	15,945±260	19,150±420
KIA26421	Lz1021-4	265–267	3.0	HAF	0.4	−23.85	19,380±510	16,905±510	20,130±1110
KIA26082	Lz1021-4	267–269	3.0	[a]					
KIA25654	Lz1021-5	459–461	3.0	HAF	1.7	−20.40	26,420±240	23,945±240	~28,500[b]±0
KIA25655	Lz1021-6	689–691	3.0	HAF	9.2	−25.29	34,650±250	32,175±250	~38,500[b]±0
KIA26083	Lz1021-7	826–838	3.0	[a]					
KIA26422	Lz1021-7	836–840	4.8	HAF	0.6	−15.36	18,600±320	16,125±320	19,350±550

All dates were calibrated into calendar years before present (cal. year BP) using the calibration program CALIB 5.0 and the SHCal04 and INTCal04 data sets (Stuiver and Reimer, 1993; McCormac et al., 2004; Reimer et al., 2004), after subtracting a reservoir effect of 2475 year (corr. year BP), corresponding with the radiocarbon age of the surface sediment, from the radiocarbon ages (^{14}C year BP). Means and uncertainties were calculated from the lowest and highest dates at the 2σ probability distribution.

[a] Failed.

[b] Estimated.

be inaccurate, as also the chronology of the Taylor Dome record may provide some uncertainties.

Correlation of the overlapping core segments was first guided by the good control we had on core collection depth through rigorous cable measurement of the depth of piston release (field depths). Then alignment of the MS signal allowed a more precise match of the overlapping segments. Finally, the cores were correlated in the laboratory on the basis of macroscopic core description, grain densities, and water contents. Measurements of biogeochemical and sedimentological data were additionally included at a later stage.

4. Results and discussion

As seen in the MS values (Fig. 2), the correlation of the overlapping core segments was not unambiguous throughout the sequence. A comparison with published data of sediment cores from Lake Fryxell (Lawrence and Hendy, 1989) revealed a relatively good correspondence in the upper ca. 35 cm, but also some distinct differences below that depth, particularly in the occurrence of aragonite horizons. This can best be explained by small-scale variations in the sediment deposition, as described from Lake Hoare nearby (Squyres et al., 1991), and is confirmed by ongoing studies in Lake Fryxell (B. Hall; personal communication). After including all information available, the compiled core composite led to a continuous sediment sequence of 914 cm length (Fig. 2). The sediment composition is dominated by clastic matter, except for the uppermost 34 cm, where coarse sandy horizons of up to several centimetres thickness are interspersed by organic layers with submillimeter laminations.

4.1. Physical properties and grain-size distribution

The physical properties MS, water content, and grain density correlate relatively well and reflect the general lithology of core Lz1021 (Fig. 2). Horizons with lower MS and grain density, as well as higher water content, document relatively fine grain sizes and/or increased amounts of organic matter. MS may indicate changes in the sediment supply from the catchment or diagenetic processes after sedimentation (Nowaczyk et al., 2001). High grain density values between 2.6 and 2.8 g/cm^3 and water contents between ca. 10% and 75% reflect the overall high proportion of clastic matter in core Lz1021.

The grain-size distribution is characterized by a clear dominance of sand and silt throughout most parts of the core (Fig. 2). This indicates that fine material has not been provided from the catchment or that high transport energies prevailed during times of deposition on the lake ice and/or the sediment surface. Nowadays, eolian-derived sand and silt migrates through the lake ice or falls through seasonal cracks and forms mounds and ridges at the lake bottom (Squyres et al., 1991; Hendy et al., 2000). The horizons with increased amounts of gravel or clay likely can be related to changes in the lake ice thickness, the transport regime, the lake level, or the occurrence and proximity of a glacier front. Lake level changes would affect the distance to inlets and thus lead to changing influence of fluvial sediment supply. The occurrence of a glacier front can induce a temperature driven lake-ice conveyor, as been demonstrated in modern proglacial lakes of the Dry Valleys (Hendy et al., 2000). The proglacial lake-ice conveyor transports icebergs and their clastic sediment load, as well as sediments in the lake ice to the distal parts of the lake and thus has a distinct impact on the grain-size distribution of the sediments deposited. Deposition of sediment at the lake bottom is complex and related to, for example, the thickness of the lake ice cover, the amount of eolian sediment in the ice, the occurrence of seasonal cracks in the lake ice, the grain-size distribution of the subaerially exposed sediments, or the formation of a moat at the lake shore (Hendy, 2000b; Hendy et al., 2000). Sediments deposited in these proglacial lakes are dominated by sand and silt with a decrease in grain-sizes towards the glacier front. Coarse grains, which accumulate by melting or ablation of glacial ice on the lake ice, prefer to remain on the lake surface and are commonly deposited in the moat zone (Hall et al., 2000; Hendy, 2000b). However, rotating or subaquatic melting of icebergs can deposit dropstones also directly at the glacier front. The grain-size distribution in core Lz1021 suggests that iceberg or lake-ice related transport processes, or lake-ice thickness have significantly affected sedimentation in Fryxell basin in the past.

4.2. Biogeochemistry

The biogeochemical proxies exhibit significant changes throughout core Lz1021. TOC and TN contents correlate well (Fig. 2) and both reflect the amount of organic matter in the sediment. In the lower part of core Lz1021, the sporadic occurrence of organic matter is related to horizons with finer grain-sizes. At the top of the sequence, alternating horizons of sand and laminated microbial mats reflect the modern sedimentation conditions. These alternating horizons are also reported from other surface sediment cores from Lake Fryxell (Hendy,

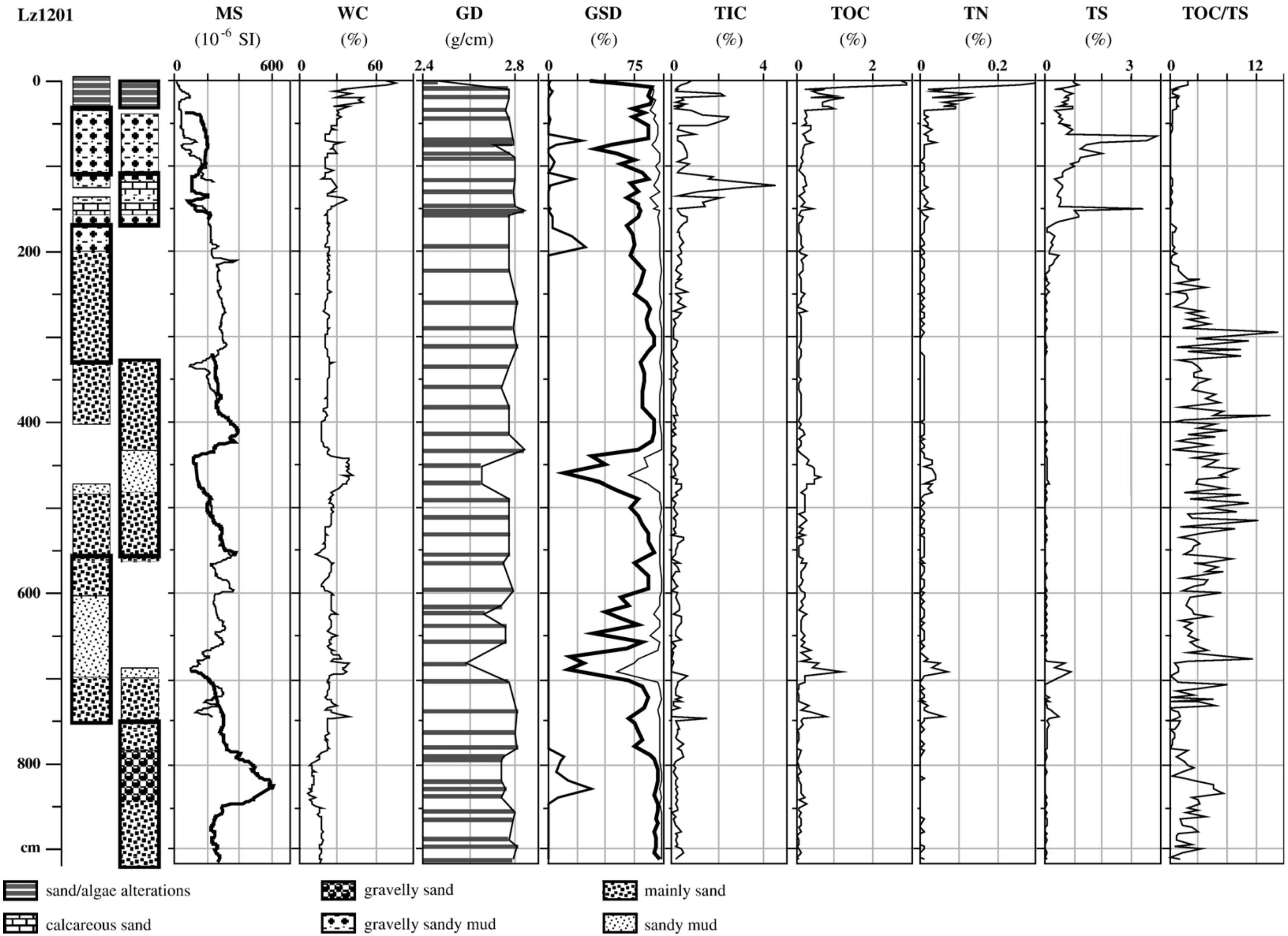

Fig. 2. Lithology, magnetic susceptibility (MS), water content (WC), grain density (GD), grain-size distribution (GSD; with cumulative gravel, sand, and silt fraction from left to right), total inorganic carbon (TIC), total organic carbon (TOC), total nitrogen (TN), total sulphur (TS), and the TOC/TS ration of core Lz1021 from Fryxell basin. The core consists of 7 overlapping segments. The MS data are completely shown in order to demonstrate the difficulties in correlating the overlapping parts and to indicate small-scale variations in sediment deposition. Analyzed sequences are framed in black.

2000a). Whilst the fine laminated microbial mats have been shown to be biogenic varves (Ian Hawes, personal communication), the interspersed horizons of coarse sand are probably due to the formation of mounds and ridges (Squyres et al., 1991; Hendy et al., 2000). However, the formation of mounds and ridges would probably have caused deposition of the overlying microbial mats in an angle, which was not observed in the core investigated. Hence, a more likely explanation is that the sand horizons represent periods with no or reduced ice cover, thus promoting deposition of eolian transported material.

Significant changes are also observed in the amount of TIC, particularly in the top 150 cm of the core. TIC likely represents calcite or aragonite in the sediment. Hendy (2000a) suggested two processes for the formation of calcite and aragonite layers in Lake Fryxell. Very thin laminae of calcite and aragonite between the microbial mats likely have been generated by carbonate precipitation due to photosynthetic CO_2 depletion in the water column. Thicker layers of aragonite were suggested to be precipitated during lake drawdown. Fine calcite or aragonite laminae have also been observed in core Lz1021 (Fig. 3). However, because TIC and TOC do not correlate, we propose that the distinct variations in TIC document lake level fluctuations rather than changes in biogenic accumulation.

The co-occurrence of TS and TOC in the lower part of the core suggests that the amount of TS is mainly formed by organic sulphur. At the top of core Lz1021, in contrast, a distinct increase of TS up to 4% is inconsistent with the TOC increase. Such high amounts of TS can be formed during strongly reducing conditions (Håkanson and Jansson, 1983; Müller, 2001) due to the formation of pyrite, a mineral observed in core Lz1021 (Fig. 3). Because TOC and TS do not correlate throughout the core, the TOC/TS ratio varies significantly. This ratio is known to reflect the salinity in the water column (Berner and Raiswell, 1984; Müller, 2001; Cohen, 2003). It can be used to reconstruct changes in the duration of lake ice cover or in the ion concentration of the water column in response to variations in meltwater and moisture supply or evaporation rate (Wagner et al., 2004).

4.3. Mineralogy

Mineralogy is characterized by slightly higher proportions of quartz and feldspars at the base and at the top of core Lz1021, and a broad maximum of volcanic glass in between (Fig. 4). Both quartz and feldspars could be supplied from the bedrock in the vicinity of Lake Fryxell, mainly from the acidic Granite Harbour Intrusive Complex. The occurrence of glass, in contrast, has been attributed to the McMurdo Volcanic Group (Ehrmann and Polozek, 1999), which is sporadically exposed in the upper Taylor Valley but widespread in the Ross Island region (LeMasurier and Thomson, 1990).

Hornblende has a distinct maximum at the core base. High amounts also occur in the upper core between ca. 420 and 70 cm sediment depth. There, brownish hornblende dominates and seems to correlate positively with glass (Fig. 4). Brownish hornblende has also been

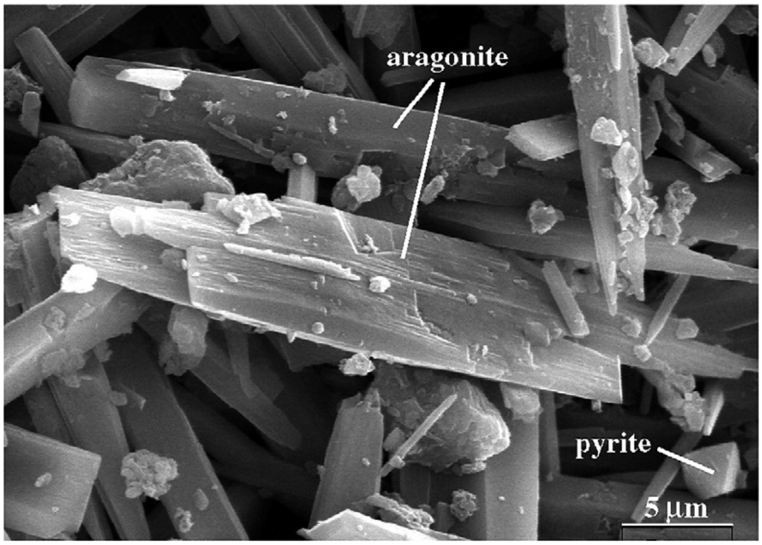

Fig. 3. SEM photography of bulk sediment from 140 cm depth in core Lz1021, showing aragonite and pyrite crystals.

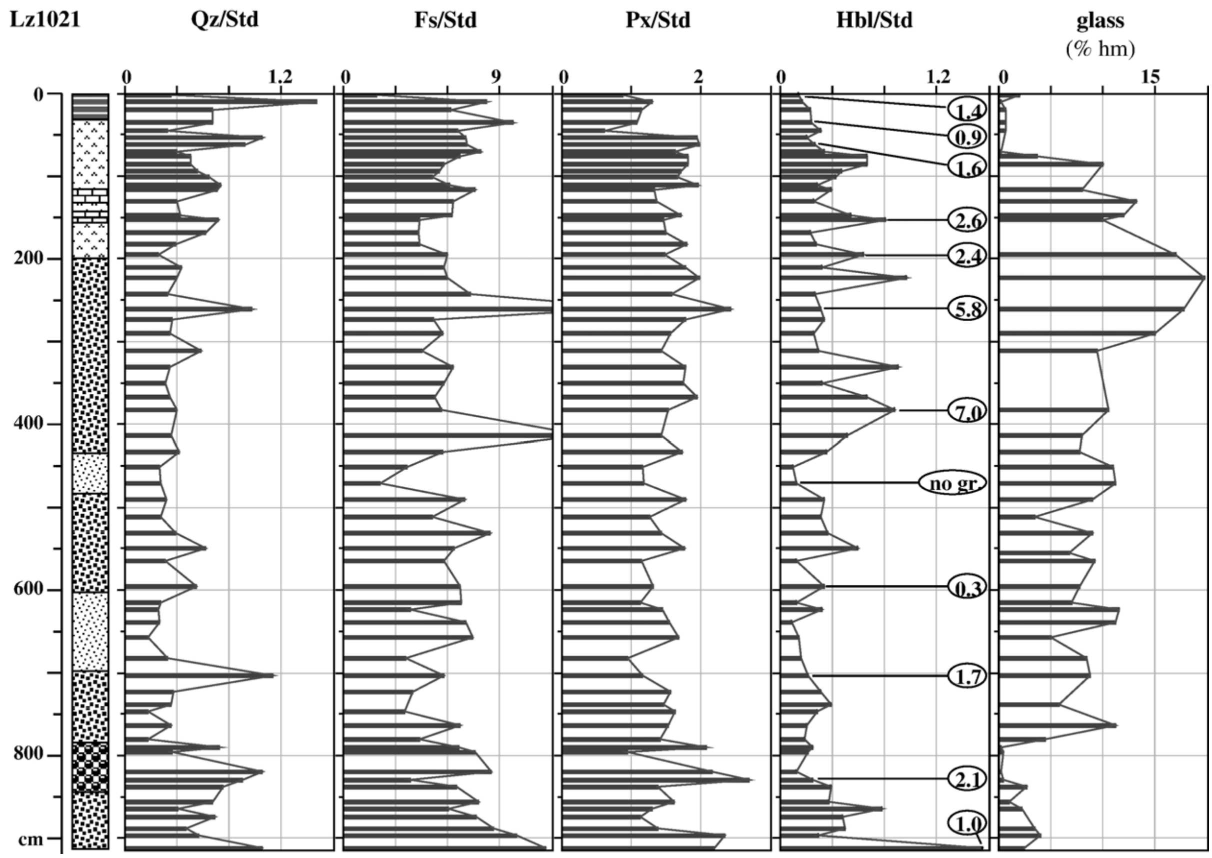

Fig. 4. Amounts of quartz (Qz), feldspars (Fs), pyroxene (Px), hornblende (Hbl) versus standard (Std) in core Lz1021. Encircled numbers indicate the ratio between brownish and greenish hornblende. Volcanic glass concentration is given as percentage of the heavy mineral fraction. The position of the horizontal grey bars represents the horizons investigated.

attributed to the McMurdo Volcanic Group (Ehrmann and Polozek, 1999), which may provide a common source of hornblende and glass for this part of the core. Pyroxene exhibits no significant changes throughout the core. Because pyroxene comprises minerals from various sources, its origin cannot be assigned to a specific geological formation in the vicinity of Lake Fryxell. In summary, the distinct long-term alternations between higher occurrence of quartz and feldspars on the one hand and volcanic glass on the other hand suggest that considerable changes in the sediment supply from the catchment have occurred.

4.4. Chronology

Radiocarbon dating is the most common method to establish a reliable chronology for Antarctic lake sediment sequences, although there exist numerous sources of error (e.g., Melles et al., 1997; Doran et al., 1999). Old carbon can be delivered from coal, carbonate, or glacier ice in the catchment. Whilst the supply of old carbon from coal and carbonate in the catchment of Lake Fryxell can widely be excluded by the geology in the catchment, the supply of old carbon from glacial meltwater likely occurs or has occurred in the past. Erroneously old ^{14}C ages can also be due to permanent ice cover and salts creating a density stratification of the water column, which leads to the lack of gas exchange between the modern atmosphere and the bottom waters.

Dating the sediment surface in Lake Fryxell yielded an age of 2475 ^{14}C year BP for the humic acid free (HAF) fraction (sample KIA24525; Table 1; Fig. 5). This age likely indicates the modern reservoir effect on the lake bottom. The surface sediment ^{14}C age of Lake Fryxell is in close agreement with respective dates from Lake Hoare nearby, where the surface sediments have been dated to ca. 2600 ^{14}C year BP and match very well the age of the bottom waters (Doran et al., 1999). Additionally, Lawson (2005) measured an age of ca. 2800 ^{14}C year BP for total benthic organic carbon at the deepest part of Lake Fryxell. The reservoir effect in

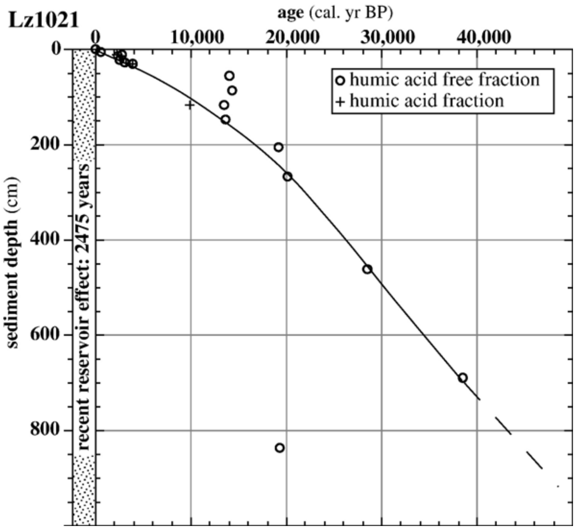

Fig. 5. Age–depth correlation of core Lz1021. All ages were corrected by a reservoir effect of 2475 years that have been dated at the sediment surface. The curve is based on a polynomial function 5th order.

proglacial lakes is only to a minor part affected by surface melt of glaciers, because of rapid gas exchange with the atmosphere (Doran et al. 1999; Hendy and Hall, 2006). This is confirmed at Lake Fryxell, where stream and nearshore microbial mats and dissolved inorganic carbon in the surface waters are in equilibrium with modern $^{14}CO_2$ (Doran et al., 1999). Subsurface melt of glaciers can contribute a significant amount of old carbon, particularly in lakes with a large glacier cross-sectional area at the grounding line. In these lakes, reservoir ages can be as much as 20,000 years immediately at the grounding line (Hendy and Hall, 2006). Today, Canada Glacier is expected to contribute only to a minor part to the reservoir age of Lake Fryxell, because its proportion of subsurface meltwater supply is relatively small. Subsurface meltwater supply from glaciers or ice sheets however may have been more significant in the past, particularly during times of the existence of the proglacial Lake Washburn. Another contribution to the reservoir ages of lakes comes from the in situ aging of lake water, the so-called residence age, which is due to the lack of gas exchange with the atmosphere as consequence of a permanent ice cover and strong salinity stratification of the water column. This contribution can play a major role and forms residence ages as much as 10,000 years, which was shown, for example, in the bottom waters in hypersaline Lake Bonney (Fig. 1; Doran et al., 1999). Lake Fryxell has today a brackish monimolimnion and a permanent ice cover, which indicate that the bottom waters do not exchange gas with the atmosphere. These conditions will have significantly changed in the past, in dependance of lake level fluctuations, meltwater supply and temperature changes. Since terrestrial macrofossils do not exist in core Lz1021 and comparable dating techniques, such as OSL dating, provided no reliable ages so far, a constant reservoir effect for the period recovered is assumed as best estimate. Hence, the surface age of 2475 ^{14}C year BP years has been subtracted from all ^{14}C ages prior to calibration into calendar years (Table 1), although changes in the reservoir effect in a range of a few thousand years have likely occurred in the past due to changes in the glacial meltwater supply, and particularly the mixing conditions of the water column and the intensity of the lake ice cover. Therefore, we are aware of the potential error of up to several thousand years involved in our reservoir correction. However, Lake Washburn was much larger, deeper and fresher, and not likely to have strong salinity stratification. Additionally, the large amount of meltwater in the basin indicates a higher melt potential than today, and so the ice cover was likely thin and potentially seasonal. Finally, the depth of the Lake Washburn (300 m) would restrict benthic photosynthesis to a narrow band on the valley walls. Organic carbon in the core is thus likely derived from fallout of pelagic organisms fixing carbon at the top of the lake (i.e., regions closer to equilibrium with the atmosphere). Thus the values we are currently applying for a reservoir correction, are likely near maxima for Lake Fryxell.

Results obtained from deeper sediments in core Lz1021 from Lake Fryxell indicate a general trend towards higher ages with increasing depth (Fig. 5). Nevertheless, a few reversals occur, particularly in the upper 160 cm (Table 1). Reasons for erroneous ages could be the low content of carbon in some of the samples and significant changes in the reservoir effect of the lake.

Ages obtained from samples KIA25654 and KIA25655 (460 and 690 cm depth) were too old for calibration into calendar years using standard calibration data sets (Table 1). Their calendar ages, therefore, have been estimated by the assumption that a linear extrapolation of the least squares line between ^{14}C ages and calendar ages is appropriate. The lowermost sample KIA26422 has an age much younger than the overlying samples and obviously was contaminated during the sampling procedure. Therefore, this sample was not taken into account calculating the age–depth correlation, based on the ages obtained by dating the less mobile HAF fraction. Samples KIA25876 (57 cm), KIA24530 (85 cm), KIA25127 (118 cm), and KIA26420 (206 cm) have also been excluded from the calculation, because they indicate apparently too old ages as a result of sediment redeposition, changing reservoir effects, or low contents of organic carbon.

The age model presented here is supported by published data. An aragonite horizon at ca. 60 cm sediment depth in another core from Lake Fryxell has been dated to 10,410 ^{14}C year BP (Hendy, 2000a), and basaltic sands at 419 cm depth have been dated to ca. 23,000 ^{14}C year BP (Whittaker et al., 2003), both matching relatively well our chronology. Nevertheless, taking the obvious dating problems and likely shifts in the reservoir effect, the age–depth correlation presented here should be regarded as a first attempt to provide data comparable with those from other records.

5. Interpretation

5.1. Stage I (ca. 48,000–45,000 cal. year BP)

Given that the changes of reservoir ages are within a range of a few thousand years throughout core Lz1021, the well sorted sand at the core base has been deposited during the Middle Weichselian, probably at ca. 48,000 cal. year BP (Figs. 5 and 6). The sorting could be the result of fluvial transport, but can also be explained by other processes. For example, sediments of similar composition in the Dry Valleys have been interpreted as windblown deposits (Higgins et al., 2000), probably trapped on the uneven surface of ice-covered lakes. Sorting takes place today in the ice covers because finer grains melt into the ice and coarser grains stay on top. Sorting of sediments can also occur in proglacial lakes, where a lake-ice conveyor transports glacial and eolian debris on the lake ice from the glacier front to the lake shore (Hendy et al., 2000). Because finer material is deflated during the transport and coarser grains are either deposited right in front of the glacier or in the moat zone, unimodal silts and sands can accumulate at locations proximal to the lake shore. Hence, the grain-size composition alone cannot be used to define the origin of the sediments and the environmental conditions in Fryxell basin at that time.

The primary source for volcanic glass in Quaternary Taylor Valley sediments most likely is the Ross Sea drift, because exposures of the McMurdo Volcanic Group occur only sporadically in the central and upper Taylor Valley (Ehrmann and Polozek, 1999). The Ross Ice Shelf incorporated volcanic material from Ross Island and terminated in front of Taylor Valley. This setting led to the formation of a proglacial lake in lower Taylor Valley, in which well-sorted sand and volcanic glass were deposited. According to Steig et al. (2000) the Ross Ice Shelf started to expand during the late marine isotope stage 5 or early stage 4, and was still expanded during the relatively warm stage 3. Although there is no further evidence that an advanced Ross Ice Shelf reached the mouth of Taylor Valley at that time, such a scenario cannot be excluded. Alternatively, volcanic material, which was deposited at the Taylor Valley slopes during earlier Ross Ice Shelf advances, could have been eroded from the valley slopes and transported by streams during times of low lake level.

In any case, the existence of a lake in Fryxell basin in the Middle Weichselian is very likely (within the constraints of our chronology). Relatively warm temperatures, which according to the Taylor Dome δ^{18}O record have been comparable to those of today (Steig et al., 2000; Fig. 6), may have led to significant meltwater supply thus supporting the existence of a lake. The trend towards higher TOC/TS ratios until ca. 45,000 cal. year BP implies a gradual increase of the freshwater conditions in the lake presumed.

5.2. Stage II (ca. 45,000–42,000 cal. year BP)

Between ca. 45,000 and 42,000 cal. year BP a significant accumulation of poorly rounded gravel, a distinct peak in the MS, a decrease in the TOC/TS ratio, and a minimum in volcanic glass characterize the sedimentation in the Fryxell basin. A mass movement event as reason for deposition of these sediments can be excluded because of a gradual transition from the underlying sediments. More likely, the significant changes in sediment composition are related to long-term environmental changes. For instance, a lake-level lowering could be reflected by the decrease of the TOC/TS ratio, indicating a negative water balance with successively more brackish conditions. A diminished inflow of meltwater could have been due to colder temperatures, as indicated in the Taylor Dome δ^{18}O ice core record (Steig et al., 2000; Fig. 6). Lake-level lowering also might explain the deposition of gravel, either because the distance of the coring location to the inlets has decreased or because the coring site has become located in the moat zone of a proglacial lake. More likely is, however, that higher salinity during lake-level lowering led to thinning or loss of the ice cover, which released coarser material from the ice surface. This is supported by maxima in quartz and feldspars and a minimum in volcanic glass, indicating that sediment was supplied from the close vicinity of the lake rather than by an advanced Ross Ice Shelf. Since the grain-size distribution and the poor roundness of the gravel are similar to those in the surface sediments deposited in

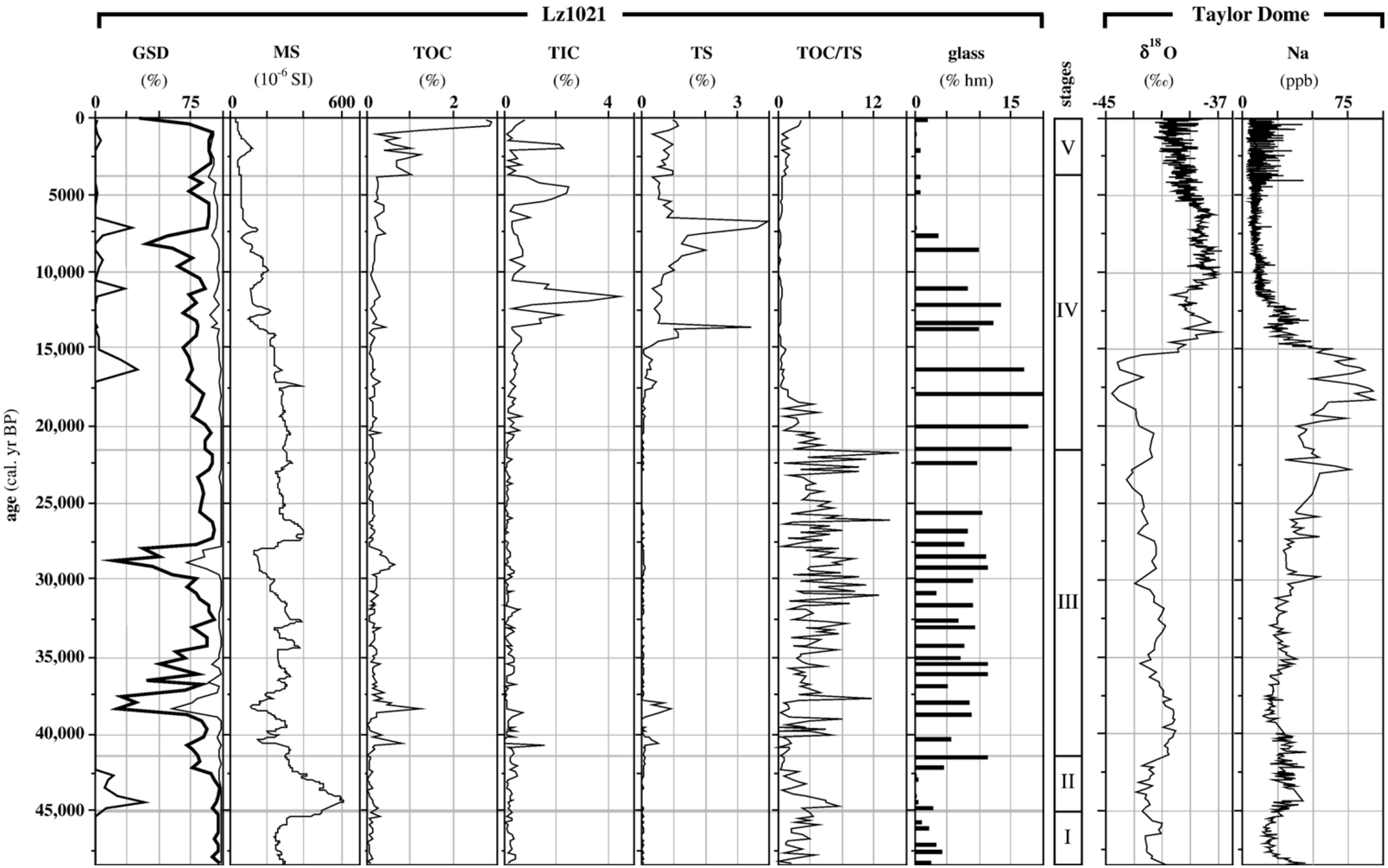

Fig. 6. Comparison between the magnetic susceptibility, the grain-size distribution, most important biogeochemical proxies, and volcanic glass in core Lz1021 from Fryxell basin with the Taylor Dome ice core $\delta^{18}O$ isotopic ratio and Na concentration. Stages (I–V) refer to significant periods of lake evolution in Fryxell basin. Note that the age–depth correlation for core Lz1021 from Lake Fryxell is based on a constant reservoir correction of 2475 years, which likely has shifted by up to several thousand years during the period recovered. Therefore, the comparison with the data from Taylor Dome should be regarded critically.

Lake Hoare today (Wagner, unpublished data), Fryxell basin at that time may have been occupied by a relatively small lake with similar environmental conditions.

5.3. Stage III (ca. 42,000–22,000 cal. year BP)

The sediments deposited between ca. 42,000 and 22,000 cal. year BP are characterized by significant fluctuations in MS, grain-size, organic matter, and TOC/TS ratios (Fig. 6). Overall higher TOC/TS ratios suggest more freshwater conditions in the lake due to higher meltwater supply and the fluctuating TOC/TS ratio might reflect a lake system responding to dynamic climate regimes, inducing changes in lake level and salinity. The relatively high amount of volcanic glass during this period implies that a significant amount of the sediments supplied to the lake originated from the McMurdo Volcanic Group on Ross Island. If our chronology is correct, this suggests that an advanced Ross Ice Shelf may have reached the mouth of Taylor Valley 10,000–14,000 years earlier than so far assumed. Our results contradict with ages of reworked marine shells from the outlet of Taylor Valley, which suggest that McMurdo Sound was free of grounded ice as late as ca. 30,000 ^{14}C year BP (Hall and Denton, 2000). However, these ages are considered to be minimum ages, given the trouble with dating old shells, and we need further confirmation for our chronology.

The geomorphological evidence for the earliest existence of proglacial Lake Washburn comes from a delta exposed well above the extant lakes in Taylor Valley. This delta has been dated to 23,800 ^{14}C year BP (Denton et al., 1985), corresponding to ca. 28,500 cal. year BP (Fig. 7). The altitude of the delta at 222 m a.s.l. (Hall and Denton, 2000) indicates that Lake Washburn must have been more than 200 m deep at this time. It also provides a minimum elevation for the Ross Ice Shelf at the Taylor Valley outlet. Meltwater from the ice shelf and from Taylor Glacier and, with minor contributions, from local alpine glaciers is supposed to have fed Lake Washburn (Hendy, 2000a). Because the lake during this period occupied large parts of the eastern Taylor Valley, the coring location Lz1021 must have been located somewhere in the central part of a proglacial lake (Fig. 8), where particularly sands and silts are deposited. This is confirmed by the grain-size distribution in core Lz1021, dominated by sand and silt in the sediments of this period.

The proglacial lake was likely too deep to produce and to preserve significant amounts of organic matter. A few periods of slightly increased organic matter accumulation, finer grain sizes, and partly coincident fluctuations in the TOC/TS ratio, suggest fluctuating lake-levels or short-term climate changes. A relict delta of Lake Washburn at 314 m a.s.l. has been dated to ca. 18,600 ^{14}C year BP (Denton et al., 1989), corresponding with ca. 22,000 cal year BP (Fig. 7). Our corresponding maximum in the TOC/TS (Fig. 6) at this time apparently confirms the high lake level. Lake-level fluctuations could have been caused by the position of the Ross Ice Shelf controlling outflow to the Ross Sea. However, the relatively constant and high amount of volcanic glass between 42,000 and 22,000 cal. year BP implies that Ross Ice Shelf was in front of Taylor Valley and supplied sediments into Fryxell basin during more or less the complete period. Hence, a more likely explanation for the presumed lake-level fluctuations is climate-driven variations in meltwater supply and evaporation.

5.4. Stage IV (ca. 22,000–4000 cal. year BP)

Between ca. 22,000 and 4000 cal. year BP distinct changes are recorded predominantly in the biogeochemistry in core Lz1021. The gradual decrease of the TOC/TS ratio between ca. 22,000 and 18,000 cal. year BP implies increasingly salinity, which was likely accompanied by increased density stratification and the formation of an anoxic monimolimnion. The anoxic bottom waters promoted the formation of insoluble sulphides and sulphurization of organic matter, which are documented by an increase of the

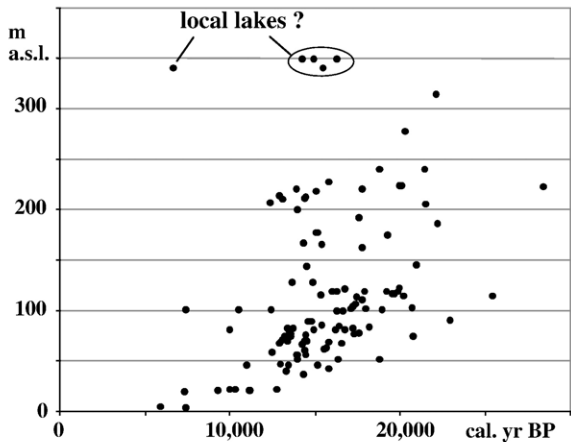

Fig. 7. Radiocarbon ages of organic remains from ancient deltas versus their altitudes in Taylor Valley, southern Victoria Land. Data are from Hall and Denton (2000). Ages have been calibrated into calendar years (cal. year BP) using the calibration program CALIB 5.0 and the SHCal04 and INTCal04 data sets (Stuiver and Reimer, 1993; McCormac et al., 2004; Reimer et al., 2004).

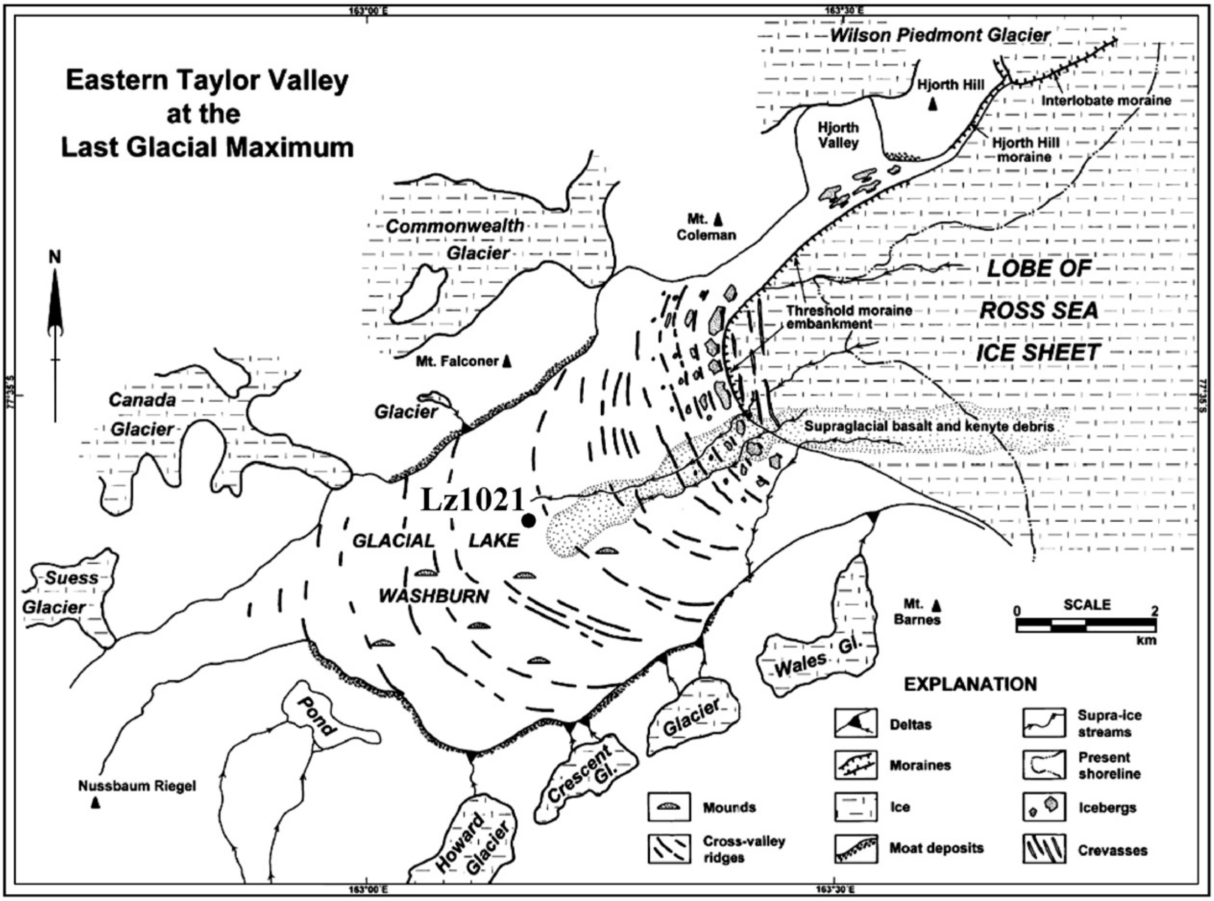

Fig. 8. Sketch of Taylor Valley during the Last Glacial Maximum, when proglacial Lake Washburn was close to its maximum level (from Hall et al., 2000).

TS amount in the sediments (Fig. 6). A significant peak in TS at ca. 13,500 cal. year BP is followed by a distinct maximum of TIC between 13,000 and 11,000 cal. year BP suggesting that further desiccation led to the precipitation of carbonate from the water column (cf. Doran et al., 1994a). Similar observations concerning a succession of sulphur and carbonate accumulation have been made, for example, in the Baltic Sea (Alvi and Winterhalter, 2001). Aragonite horizons have already been found in sediments from the Fryxell basin, and explained by evaporation of Lake Washburn to a lower level (Lawrence and Hendy 1989; Hendy, 2000a). Lake-level lowering during this period is confirmed by the relict deltas, which successively are formed at lower altitudes (Fig. 7). A few relict deltas at altitudes of more than 300 m a.s.l. with ages less than ca. 6700 cal. year BP do not challenge this interpretation, because they were likely formed by local, smaller lakes on the valley slopes (Hall and Denton, 2000).

Lake Washburn evaporation was likely induced by cold conditions during the Last Glacial Maximum, which are indicated in the oxygen isotope ratios in the ice core from Taylor Dome (Steig et al., 2000; Fig. 6) and led to reduced formation of meltwater. Low accumulation rates at that time, documented in the ice core from Taylor Dome (Steig et al., 2000) may have supported lowering of Lake Washburn. Sea-level lowering during the Last Glacial Maximum allowed the Ross Ice Shelf to ground, expand and thicken into an ice sheet, i.e., the Ross Ice Sheet (RIS) (Steig et al., 2000). Simultaneously, Taylor Glacier and alpine glaciers in Taylor Valley retreated, likely because the widespread grounded ice sheet in the Ross Sea hampered moisture supply to the Dry Valleys (Higgins et al., 2000). Additional reduction of moisture supply could be due to more extensive sea-ice cover at that time. This is indicated in elevated Na concentrations in the Taylor Dome record (Fig. 6), taken that sea salts in ice cores originate from the surface of the sea ice

(Rankin et al., 2004). However, diminished moisture transport is likely of much less importance for the negative water balance in Lake Washburn than cold temperatures, since the water balance is mainly controlled by glacial meltwater influx.

Our data support the hypothesis that lowering of Lake Washburn was initiated by evaporation rather than by drainage. The maximum position of RIS at the Taylor Valley outlet is recorded by the formation of a moraine at Hjorth Hill (Fig. 8). This moraine at ca. 350 m a.s.l. has been dated to 12,700–14,600 ^{14}C year BP (Hall and Denton, 2000), corresponding with ca. 15,000–18,000 cal. year BP. Despite at least one slight retreat, prominent moraines of a RIS were formed at Hjorth Hill as late as ca. 10,500 ^{14}C year BP (ca. 12,000 cal. year BP) close to the maximum position. The existence of a flow line of the Ross Sea ice around northern Ross Island to the eastern Taylor Valley from at least 17,500 until at least 9000 ^{14}C year BP, corresponding with ca. 21,000 and 10,000 cal. year BP, respectively, is indicated by the deposition of kenyte erratics in the eastern Taylor Valley (Stuiver et al., 1981). This is confirmed by a maximum in volcanic glass deposition in core Lz1021 between 20,000 and 15,000 cal. year BP (Fig. 6). Hence, evaporation started simultaneously with the beginning of maximum RIS expansion. The results obtained here confirm the suggestion of Hall and Denton (2000) that a maximum of lacustrine deposits between ca. 16,500 and 11,000 ^{14}C year BP, corresponding with ca. 19,500 and 13,000 cal. year BP, was either due to a very active lake-ice conveyor or to a fast lake-level lowering. Lake-level lowering and increasing salinity may have led to lake ice thinning, thus promoting higher sedimentation of gravel and clay (Fig. 6). The increasing salinity likely also caused reduced mixing of the water column, which likely resulted in a distinct increase of the reservoir effect. This could at least partly explain the relatively high occurrence of erroneous ages around 14,000 cal. year BP (Fig. 5).

Last significant supply of volcanic glass is documented in core Lz1021 at ca. 9000–8000 cal. year BP, suggesting that the RIS retreated from the mouth of Taylor Valley afterwards. The low concentrations of volcanic glass after 7500 cal. year BP imply only limited erosion of volcanic material that was deposited at the valley slopes during the past RIS advances. A mid-Holocene recession of the RIS is confirmed by marine shells, of which one from a location adjacent to Explorers Cove was dated to 5370 ^{14}C year BP (Hall and Denton, 2000), corresponding with ca. 6200 cal. year BP. The RIS retreat probably caused higher moisture supply to Taylor Valley during the mid to late Holocene (Steig et al., 2000), although increased Na concentrations in Taylor Dome ice core record also might document increased sea-ice cover (Rankin et al. 2004; Fig. 6). Increased moisture supply may have caused readvance of alpine glaciers in Taylor Valley during the so-called Alpine I drift from ca. 6000 ^{14}C year BP (ca. 6800 cal. year BP). A coeval lake-level high-stand inferred from the occurrence of relict deltas (Hall and Denton, 2000), however, is not documented in core Lz1021. In contrast, a second period of inferred salinity increase, with a first increase of TS and a subsequent peak of TIC, is indicated between ca. 9000 and 4000 cal. year BP (Fig. 6). These discrepancies could be explained by dating uncertainties, or the relict deltas being from local lakes.

5.5. Stage V (<ca. 4000 cal. year BP)

The sediments deposited at site Lz1021 from ca. 4000 cal. year BP to present are comprised of fine laminated microbial mats with interspersed thick horizons of coarse sand, similar to those found at the sediment surface in Lake Fryxell and Lake Hoare (Squyres et al., 1991; Doran et al., 1994a,b). They thus indicate environmental conditions and water depths comparable to those of today. The deposition of the microbial mats is documented in significantly elevated TOC contents, and low MS (except for a peak at ca. 2000 cal. year BP) and TIC (Fig. 6). Freshening of the water towards the present is indicated in a slight increase of the TOC/TS ratio.

The TOC increase and the occurrence of microbial mats indicate increased photosynthetic activity as a consequence of distinctly lower lake level and volume. Microbial mat laminations have been shown to be annual (Ian Hawes, personal communication). In the modern environment, 1.2% to 2.7% of incident light penetrates the ice cover, which is sufficient for photosynthesis by organisms adapted to these low light levels. Photosynthesis occurs through most of the Lake Fryxell water column, but the benthic microbial mats appear to be the major primary producers in this lake and provide the majority of organic carbon in the sediments (Lawson et al., 2004). The boundary between microbial mats and interspersed sand horizons apparently does not show an angle and is rather horizontal. Such an angle likely would have been observed, if the microbial mats were deposited on mounds and ridges of sand at the lake floor, which would have been formed by sand melting through the lake ice (Squyres et al., 1991). Therefore, other formation processes for the sand horizons cannot be excluded. For example, periods with ice-cover absence may have promoted the immediate deposition of eolian sand. Alternatively, reduced eolian accumulation may have occurred, when Canada Glacier separating Lakes Fryxell and Hoare in Taylor Valley advanced and thus formed a natural barrier for eolian transport down valley.

At the end of the Alpine I drift between 3000 and 2000 ^{14}C year BP (Hall and Denton, 2000), Lake Vanda ca. 40 km to the northwest of Lake Fryxell was extensively fed by meltwater and had a period of lake-level high stand (Hendy 2000a). Apparently, relict deltas document a lake-level high stand during the same period also for Lake Fryxell (Hall, 2003). This lake-level high stand was followed by a desiccation, during which Lake Fryxell evaporated to an ice-free, hypersaline pond (Wilson, 1964; Lyons et al., 1998). Evaporation and hypersaline conditions could explain the relatively constant ages around ca. 2500 cal. year BP between ca. 24 and 10 cm depth in the surface core Lz1021-1 from Lake Fryxell (Table 1). Upwards increasing hypersaline conditions in Lake Fryxell may have hampered the gas exchange of the bottom waters with the atmosphere, thus increasing the reservoir ages. Alternatively, they may have promoted enhanced deposition of eolian sands due to ice cover loss, or substantial thinning, which is supported by a spike in gravel at this level. Probably, also the TIC peak at ca. 2000 cal. year BP in core Lz1021 corresponds with the period of evaporation. Starting at ca. 1000 cal. year BP, the high amount of TOC and the distinctly reduced amount of TIC indicate that Lake Fryxell filled again with meltwater. This is also suggested by the increase of the TOC/TS ratio upwards. The increased supply of meltwater in recent history has been attributed to higher summer temperatures or the number of clear, calm and snowless midsummer days (Lyons et al., 1998; Hendy, 2000a).

6. Conclusions

Detailed investigation of a sediment sequence from Fryxell basin, southern Victoria Land, Antarctica, using chronological, sedimentological, biogeochemical, and mineralogical methods, confirms, complements, and provides new information about the environmental history of the region. However, the palaeoenvironmental information provided by the sediment sequence should be handled carefully, because small-scale variations in the sediment deposition in the Fryxell basin are common and the chronology of the sediment sequence shows some uncertainties. Large shifts in the reservoir effect as consequence of significant changes in glacial meltwater supply and, particularly, in hydrological conditions in the lake have likely occurred over time, but cannot be determined at this stage. Therefore, the comparison of our results with those from other records should be regarded as a first attempt, and the results need validation in the future.

Our data confirm that eastern Taylor Valley was occupied by a proglacial lake, so-called Lake Washburn, during the Last Glacial Maximum and that this lake was dammed by an advanced RIS blocking the mouth of Taylor Valley. Between 9000 and 8000 cal. year BP, the RIS retreated from the McMurdo Sound. Lake-level lowering of Lake Washburn was discontinuous. Last significant evaporation of the remnant of Lake Washburn, the present Lake Fryxell, occurred between ca. 2500 and 1000 cal. year BP.

We can complement information that lowering of Lake Washburn started long before RIS retreated significantly from the Taylor Valley outlet. Carbonate precipitation observed during evaporation periods of Lake Washburn after the Last Glacial Maximum was preceded by sulphide accumulation in the sediments. Environmental conditions similar to those of today were reached around 4000 cal. year BP.

New aspects are that Fryxell basin was occupied by a small lake already during the mid Weichselian. After a short period of lake-level low stand, an advanced RIS, as indicated in the mineralogical sediment composition, likely led to the formation of proglacial Lake Washburn as early as ca. 42,000 cal. year BP, i.e., 10,000–14,000 years earlier than previously assumed. Lake Washburn commenced significant lake-level fluctuations. As indicated in the increasing salinity of the lake water, final lowering of Lake Washburn from the Last Glacial Maximum was induced by evaporation rather than drainage, likely as a result of cold temperatures.

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