ELSEVIER

Palaeogeography, Palaeoclimatology, Palaeoecology 147 (1999) 223–239

PALAEO

Dating Quaternary lacustrine sediments in the McMurdo Dry Valleys, Antarctica

Peter T. Doran [a,*], G.W. Berger [b], W.B. Lyons [c], R.A. Wharton, Jr. [b], M.L. Davisson [d], J. Southon [d], J.E. Dibb [e]

[a] *Earth and Environmental Science, University of Illinois at Chicago, Chicago, IL 60607-7059, USA*
[b] *Desert Research Institute, 7010 Dandini Blvd, Reno, NV 89512, USA*
[c] *Department of Geology, University of Alabama, Box 870338, Tuscaloosa, AL 35487, USA*
[d] *Lawrence Livermore National Laboratory, P.O. Box 808, Livermore, CA 94550, USA*
[e] *Glacier Research Group, University of New Hampshire, Morse Hall, 39 College Road, Durham, NH 03824-3525, USA*

Received 8 January 1998; revised version received 22 September 1998; accepted 23 September 1998

Abstract

Reports of erroneously old ^{14}C dates for modern Antarctic materials have thrown doubt into ^{14}C chronologies. The carbon reservoir effect purported to exist in Quaternary lacustrine sediments of the McMurdo Dry Valleys was investigated by studying ^{14}C distribution, and testing alternate dating techniques. Our results show that the carbon reservoir effect is not pervasive. Stream and near-shore microbial mats and dissolved inorganic carbon (DIC) in the surface waters of Lake Fryxell are in equilibrium with modern $^{14}CO_2$. The surface waters of Lake Hoare and Lake Bonney, however, have DIC ^{14}C ages of 1650 and 2080 yr B.P., respectively. These older age estimates are suggested to be due to the direct input of large amounts of glacial melt with relict DIC. On the other hand, Lake Fryxell receives only a minor component of its inflow directly from a glacier, while a large component must travel long distances in numerous shallow ephemeral streams after leaving local valley glaciers. This mode of melt-water input allows the water to equilibrate with modern CO_2 before entering the lake. Bottom-water ^{14}C ages for Lake Hoare closely match surface sediment ages, supporting the widely published period ~1200 yr B.P. (after a 1650 yr reservoir correction) when most dry valley lakes apparently evaporated to small brine ponds and/or disappeared completely. Lake Bonney bottom-water is ~8000 yr B.P. Carbon dating is shown to be a viable technique for lake edge deposits, and possibly lake bottom deposits where a correction to the sediment surface age can be obtained. However, we conclude that deep-water paleolake deposits can not be reliably dated using ^{14}C alone because of an inability to determine the age of the reservoir correction (i.e. accounting for the initial carbon reservoir, plus the age of the bottom water). A suite of alternative and complimentary dating techniques were tested on modern and late Holocene lacustrine deposits. These include thermoluminescence (TL), ^{210}Pb, ^{137}Cs, and paleomagnetism. Of these techniques, TL (or the more sensitive optically stimulated luminescence) holds the most promise for correcting lake sediment ^{14}C ages. TL dating of the modern Lake Hoare sediments showed an ~1000 yr relict signal. This signal is unaffected by the age of the lake water. Both ^{210}Pb and ^{137}Cs occurred in very low levels in the sediments and do not

* Corresponding author. Fax: +1-312-4137275; E-mail: pdoran@uic.edu

PII: S0031-0182(98)00159-X

appear to be viable dating techniques for the perennially ice-covered lakes of the McMurdo Dry Valleys. Paleomagnetism was not suited to the coarse-grained nature of Lake Hoare bottom sediments, but could be a useful alternative given a finer sediment type.

Keywords: Antarctica; geochronology; dating; lacustrine; lakes; sediments

1. Introduction

Quaternary dating in the Antarctic has long been a problem. In the marine and coastal environment, a variable carbon reservoir has made ^{14}C dating a challenge. The effect on dating of carbonaceous materials caused by this reservoir of relict carbon varies considerably depending upon location and sample type. For instance, in the Antarctic coastal waters, an elevated reservoir effect is common because of the upwelling of deep ocean waters (Broecker, 1963). On the other hand, recent input of bomb ^{14}C has reduced the antarctic relict signal to ~800 yr B.P. (Berkman and Forman, 1996). Reliable corrections can be obtained for the marine system by systematically collecting and dating modern carbon samples and arriving at a local offset for the older samples. In contrast, correcting for the carbon reservoir effect in lacustrine sediments is more complicated.

Dating of lacustrine deposits in coastal antarctic oases has identified a variable carbon reservoir from marine sources, and in many lakes, from old carbon supplied by glacial melt (e.g. Zale and Karlén, 1989; Bird et al., 1991; Björck et al., 1991; Melles et al., 1994; Gore, 1997). What is known about the antarctic lacustrine reservoir effect comes mainly from the surface age of lake bottom sediments. Very little systematic work has been carried out to identify the sources of carbon responsible for the lake bottom signal. In fact, to our knowledge, only two ^{14}C ages have been published for Antarctic lake water (Hendy et al., 1977; Friedman et al., 1995), and there is no available information on the age of inflowing stream water.

In this paper, we present a suite of ^{14}C dates from lakes and associated deposits in the McMurdo Dry Valleys of southern Victoria Land, Antarctica. Lake water dissolved inorganic carbon (DIC), particulate organic carbon (POC), stream and lake bottom organic material (microbial mat), soil carbonate, and paleolake deposits (carbonate and organic matter) have all been analyzed for ^{14}C content in order to define the nature and extent of the carbon reservoir in dry valley lacustrine systems. We also review the results from tests of various other dating techniques in this environment, including TL, ^{210}Pb, ^{137}Cs, and paleomagnetism.

2. Site description

The McMurdo Dry Valleys (Fig. 1) comprise the largest ice-free area in Antarctica. The McMurdo Dry Valleys are relatively ice-free largely because the Transantarctic Mountains block the flow of ice from the Polar Plateau into the region. The mean annual air temperature in the dry valleys is between −17 and −20°C (Doran et al., 1994), creating a range of permafrost in this region of 240 to 970 m thick (Decker and Bucher, 1982). Limited precipitation data suggest that the mean annual precipitation is received as snow and is on average <10 cm a^{-1}, water equivalent (Bromley, 1985). This value is well below measured ablation rates, which have ranged from 15 to 50 cm a^{-1} (Henderson et al., 1965; Clow et al., 1988). The low precipitation, low surface albedo, and dry winds descending from the Polar Plateau result in extremely arid conditions (Clow et al., 1988).

The basement metamorphic rocks of southern Victoria Land belong to the Upper Proterozoic–lower Paleozoic Ross geosynclinal complex (Lopatin, 1972). The Kukri Peneplain, a smooth erosional surface, separates the basement from the Beacon Supergroup, which contains the Middle to Upper Devonian Taylor and Permo-Carboniferous to Early Jurassic Victoria groups (Barrett et al., 1972). These strata are intruded and overlain by the Jurassic Ferrar Dolerite and Kirkpatrick Basalt (Barrett et al., 1972; Smithson et al., 1972). Coarse clastic volcanic rocks can be found in the Triassic strata and tholeiitic basalt flows which cap the Beacon Supergroup (Barrett et al., 1972).

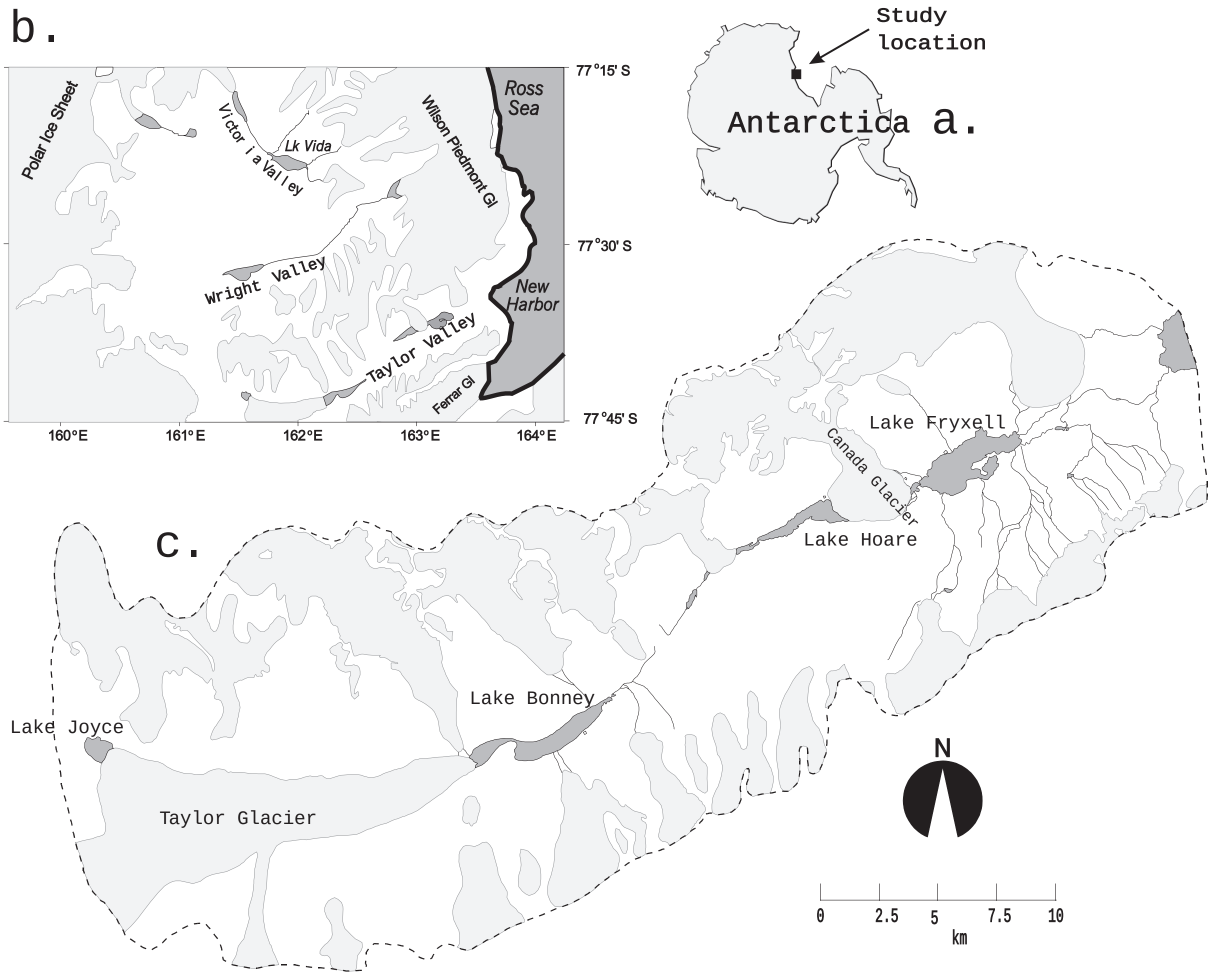

Fig. 1. Map showing (a) study location on the Antarctic Continent, (b) northern McMurdo Dry Valleys region, and (c) Taylor Valley.

Two broad drift sheets are found in the coastal areas of the dry valley region. The first drift sheet, the Ross Sea I drift (Denton et al., 1971) or Ross Sea drift (Stuiver et al., 1981), is little weathered, and the surface is kettled and marked with bands of erratics, moraines, and eskers up to 2 km long. Soils of this drift are coarse, loose, deficient in salts, weakly oxidized, and have a pH between 6.5 and 9.3. Throughout soil profiles, olive-yellow calcium carbonate coatings are common beneath coarse fragments (Stuiver et al., 1981). The second broad drift sheet is older than and distal to the Ross Sea I drift.

All three main dry valleys (Taylor, Wright, and Victoria) have a long lacustrine history (Stuiver et al., 1981; Chinn, 1993; Doran et al., 1994). In Taylor Valley, a large glacial lake, Glacial Lake Washburn, was present for much of the Late Pleistocene. In Wright Valley, an expanded Lake Vanda is marked by strand lines above present lake level. In Victoria Valley, recent field evidence has revealed repeated Pleistocene glacial lake episodes; the most recent lake episode is tentatively named Glacial Lake Victoria (B. Hall, pers. commun., 1997). Paleolake deposits are preserved in all three main valleys.

Presently, the McMurdo Dry Valleys contain more than 20 permanent lakes and ponds which vary markedly in character. Numerous authors (e.g. Armitage and House, 1962; Matsubaya et al., 1979; Priddle and Heywood, 1980; Burton, 1981; Cartwright and Harris, 1981; Chinn, 1982, 1993; Heywood, 1984; Green et al., 1988; Hammar, 1989; Wharton et al., 1993) have synthesized various aspects of antarctic limnology, including treatment of the dry valley lakes. Doran et al. (1994) provide a comprehensive review of the paleolimnology in the McMurdo Dry Valley region[1].

By far the most important feature of the dry valley lakes is the presence of a permanent ice cover (Wharton et al., 1993). Ice cover limits the amount of radiation reaching the water column (Palmisano and Simmons, 1987; Wharton et al., 1989), and greatly reduces interaction between the atmosphere and lake (Wharton et al., 1993). The ice cover also buffers the lake from the harsh Antarctic climate, especially the winter climate, thereby maintaining the lake and its ecosystem. During the summer melt all lakes develop an ice-free 'moat' to some extent. Communication between moat waters (and hence terrestrial inflow) and under-ice waters appears to happen presently in all years for the three main lakes studied in this paper.

Sediment/ice interactions have been studied on Lake Hoare in Taylor Valley (Simmons et al., 1986; Nedell et al., 1987; Wharton et al., 1989; Squyres et al., 1991). The overall conclusion from these studies was that sediment in Lake Hoare's ice cover follows a cyclical pattern of accumulation on the ice surface and fallout through the ice bottom. Fallout often occurs as localized mounds or linear ridges due to the sorting affect of the ice cover. The resulting lake bottom sediment profile is very heterogeneous.

3. Carbon-14

3.1. ^{14}C background

Since the seminal work of Libby (1955), ^{14}C dating has become the most widely used Quaternary dating technique. Based on the natural production of ^{14}C in the atmosphere by a variety of nuclear reactions, and its subsequent radioactive decay (half-life $= 5568$ yr), ^{14}C activity can be used to date many types of carbon-bearing materials. With new accelerator techniques, small sample sizes can be used to obtain high resolution results.

As mentioned above, the use of ^{14}C dating in the Antarctic is problematic, as a result of a carbon reservoir effect. Numerous authors working on both terrestrial, aquatic and coastal environments have observed surficial organic matter with very old ^{14}C ages (e.g. Domack et al., 1989; Bird et al., 1991; Björck et al., 1991; Squyres et al., 1991; DeMaster et al., 1992; Gordon and Harkness, 1992; Berkman and Forman, 1996). In Björck et al. (1996), a surface sediment carbon reservoir was implied by upward extrapolation of dates lower in the profile. Nevertheless, very little work has been done with modern lake ^{14}C dynamics in the Dry Valleys or in other Antarctic oases. The impact of the thick permanent lake ice covers and the mode of glacial input have not been tested. Hendy et al. (1977) dated the dissolved inorganic carbon (DIC) in the bottom of the west lobe of Lake Bonney at $15,450 \pm 1650$ yr B.P. Stuiver et al. (1981) collected many organic mat samples for calculating the marine reservoir effect in the region, but assumed there was no need for a lacustrine correction. Björck et al. (1991) reported algae flakes from the lake shore of Skua Lake that dated 120 ± 45 yr B.P. Friedman et al. (1995) reported a ^{14}C date of 1150 yr B.P. from a single DIC sample at 64.3 m depth in Lake Vanda.

3.2. ^{14}C methodology

In this study we used ^{14}C ages from conventional counting techniques and from Accelerator Mass Spectrometry (AMS). All dates are corrected for fractionation effects.

Conventional ^{14}C dating was performed at the Desert Research Institute's Radiocarbon Laboratory. Sample carbon is converted to CO_2 and then to benzene. A small amount of CO_2 is used to measure the stable carbon isotope ratio. ^{14}C activity is measured by beta counting in liquid scintillation counters. Four samples (two paleolake carbonates and two paleolake algae mats) from the Lake Vida basin were dated by conventional techniques.

All remaining ^{14}C dates were generated at the Lawrence Livermore National Laboratory's Center for Accelerator Mass Spectrometry (Southon et al., 1990). Samples dated using AMS included dissolved inorganic carbon (DIC) in glacier ice, stream water,

[1] Full details of the nature of the lakes discussed in this paper can be found in the above-cited publications and at http://huey.colorado.edu.

and lake water, various microbial mats (both from the modern sediment surface, and from lake cores), and soil carbonate. Liquid-water samples for DIC dating were collected by hand or using a Kemerer sampling bottle, stored in pre-ignited 100 ml serum bottles, and poisoned immediately with mercuric chloride ($HgCl_2$). Extraction of gaseous CO_2 from the DIC was performed by a combination of acid-stripping techniques described by Rao and Killey (1994) and McNichol et al. (1994). All DIC water samples were collected in the deepest oxic water possible. Glacial ice was mechanically cleaned prior to being melted in a pure nitrogen environment. Once melted, CO_2 gas trapped in the ice was collected before the meltwater was acidified as above to liberate DIC for dating. POC samples from Lake Hoare water were collected in 1988 by filtering over 400 l of lake water through glass fiber filter columns. The filter columns were stored frozen since collection, and were dried just prior to ^{14}C dating.

Microbial mats and soil carbonates were collected by hand (underwater samples were collected by divers) and kept frozen until analysis. Microbial mats were treated in 0.1 N glacial acetic acid (CH_3COOH) for 24 h to gently remove any carbonate. Separate carbonate samples were oxidized in a 50% chlorox solution to remove organics.

3.3. ^{14}C results and discussion

The abundance of ^{14}C in various materials from Taylor Valley is shown in Tables 1–3. Of particular interest is the fact that stream microbial mat and water date near modern at the point where they enter the lakes. Stuiver et al. (1981) anticipated this result when they stated " . . . we can only assume that such a (reservoir) correction is not needed because the deltas were deposited in near shore, ice-free, shallow waters by streams that have travelled as much as 1 km and whose water is therefore thoroughly mixed." (p. 345). We tested this conjecture by sampling glacier ice, and stream water DIC and microbial mat at the head and mouth of Delta Stream (an ~3 km stream originating at the base of Howard Glacier). The single glacier ice sample carried a carbon reservoir of 7430 ± 60 yr B.P. Water sampled in a pool adjacent the calved glacier ice had already 'modernized' to 570 ± 60 yr B.P. Microbial mats in this pool were older than the water at 1700 ± 60 yr B.P. The reason for the discrepancy between the pool water and mat is unclear but is likely related to the benthic mat utilizing DIC only from the very bottom of the pool. This could imply a very strong percent modern carbon (PMC) gradient in the pools or a significant amount of seepage from shallow soil flow into the pools. By the time the water has travelled to the lake edge it has completely equalized with modern $^{14}CO_2$ and microbial mats utilizing this water date modern also.

Stuiver et al. (1981) also speculated about the effect of carbonate in local rock outcrops, and conclude that it must be negligible because flow from the glaciers is over drift that is poor in carbonate clasts. Although our data support this, there is an abundance of soil carbonate in the system that has not been considered previously. Our data from Table 1 show that soil carbonate is relatively deficient in ^{14}C. We assume the reason it does not have a significant impact on the stream water is that the streams have run the same course for thousands of years (as indicated by the close association of deltas and modern streams). Any soil carbonate in the stream bed that is available to dissolve into the flowing stream water has done so long ago, and we would expect to see a decreasing carbon reservoir in the streams with age.

Lake waters display a wide range of DIC ^{14}C ages. Lake Fryxell near-surface waters are modern, while Lake Hoare and Lake Bonney carry a carbon reservoir of around 1650 and 2100 yr, respectively. Lake Bonney has an exceptionally old bottom water of ~10,000 yr, while Lake Hoare has a bottom water of ~2700 yr. Lake Fryxell bottom water is problematic. The high levels of ^{14}C in this water indicate some source of contamination, either in the water column or during sample collection. Photosynthetic studies using ^{14}C as a tracer in sealed glass bottles are common in lake studies, including in the dry valleys. One such study in 1988 reported accidentally losing 84 μCi of ^{14}C at depth in Lake Fryxell. Could this be the source of the contamination? Given a DIC content at this depth of ~15 mmol C l^{-1}, the plume would have to expand to 1.12×10^5 m^3 (~0.5% of the liquid water in the lake) in order to dilute to the present level of ^{14}C activity. We assume this plume has confined itself to a narrow density layer, but without further data we can not model its

Table 1
Summary of carbon-14 dates from various materials in the Lake Fryxell Basin

Sample location	Date	Material	% Modern C	^{14}C	±
Lake Fryxell water, 5 m depth	Nov. 1995	DIC	108	modern	
Lake Fryxell water, 5 m depth	Nov. 1996	DIC	109	modern	
Lake Fryxell water, 9 m depth	Nov. 1995	DIC	168	hot	
Delta Stream at Howard Glacier	Dec. 1996	DIC	93	570	60
Delta Stream at Howard Glacier	Dec. 1996	Mat	81	1700	60
Delta Stream at Howard Glacier	Dec. 1996	DIC [a]	40	7430	60
Delta Stream at Lake Fryxell	Dec. 1996	DIC	99	100	70
Von Guerard Stream at Lake Fryxell	Jan. 1995	Mat	111	modern	
Crescent Stream at Lake Fryxell	Jan. 1995	Mat	114	modern	
Soil on south side of Lake Fryxell basin	Jan. 1995	Carbonate	67	3230	60
Soil on south side of Lake Fryxell basin	Jan. 1995	Carbonate	27	10360	100

[a] In glacier ice.

Table 2
Summary of carbon-14 dates from various materials in the Lake Hoare Basin

Sample location	Date	Material	% Modern C	^{14}C age	±
Lake Hoare water, 4 m depth	Nov. 1995	DIC	81	1650	60
Lake Hoare water, 5 m depth	Nov. 1996	DIC	81	1660	60
Lake Hoare water, 20 m depth	Nov. 1996	DIC	73	2490	50
Lake Hoare water, 25 m depth	Nov. 1995	DIC	72	2670	50
Lake Hoare water, 6 m depth	Dec. 1987	POC	19	13530	60
Lake Hoare water, 12 m depth	Dec. 1987	POC	28	10280	70
Andersen Ck. at Lk. Hoare	Nov. 1996	DIC	97	280	70
Andersen Ck. at Lk. Hoare	Nov. 1996	Mat	102	modern	

Table 3
Summary of carbon-14 dates from Lake Bonney East Basin

Sample location	Date	Material	% Modern C	^{14}C age	±
East Lake Bonney water, 5 m depth	Nov. 1995	DIC	110	modern	
East Lake Bonney water, 5 m depth	Nov. 1996	DIC	77	2080	70
East Lake Bonney water, 35 m depth	Nov. 1995	DIC	28	10220	60
East Lake Bonney water, 37 m depth	Jan. 1995	DIC	32	9270	60

growth or movement. Assuming the plume were to spread vertically three meters, the horizontal spread could be a 110 m radius away from the deep hole. Given sampling in this lake has always been from the deepest point in the lake, and the ease of relocating this sampling site on the perennial ice cover, this plume would be an easy target for our sampling.

As stated earlier, two of the three lakes we dated have deficient ^{14}C in their surface waters. Lake Hoare receives a large amount of its inflow directly from the Canada Glacier surface, without the water ever flowing in a soil-based stream. Furthermore, thermal erosion of the glacier underwater must be contributing glacial melt directly to the lake. Lake Bonney also receives a large amount of inflow directly from the Taylor Glacier. The Taylor Glacier flows into the west lobe of Lake Bonney, but Spigel and Priscu (1998) show that significant glacial inflow from Taylor Glacier can be traced flowing into the east lobe across the sill separating the two lobes in a

Table 4
Geochemical and thermodynamic data for selected streams in Lake Fryxell Basin

Stream site [a]	Distance from head (km)	pH	DIC	p_{CO_2} [b]
Delta-u	4.9	7.4	7.1	−2.95
Delta-g	11.2	7.4	11.2	−2.75
Von Guerard-u	1.2	6.8	7.2	−2.34
Von Guerard-l	3.2	7.2	9.6	−2.62
Von Guerard-g	4.8	7.4	10.8	−2.77

[a] g = gage (within ~100 m of lake), l = lower, u = upper.
[b] Calculated with PHRQUPITZ.

single melt season. Lake Fryxell, on the other hand, has only a minor component of direct glacial input, while the vast majority of water enters the lake by stream flow. We propose that the path glacial melt takes (or mode of glacial input) is responsible for the difference in apparent age of the surface waters. Water that has been directly or quickly deposited under the lake ice from the glacier (e.g. in Lakes Hoare and Bonney) has not had the opportunity to equalize with atmospheric carbon. Although the water may run a long course on the glacier, it scours the ice surface through mechanical and thermal abrasion, collecting more glacial melt and old DIC. In Lake Fryxell, water flowing in the terrestrial streams are out of contact with the old DIC, and becomes atmospherically equilibrated as it makes its way towards the lake. This hypothesis is supported by thermodynamic data from two streams in Lake Fryxell basin that clearly show an increase in DIC with distance from the glacier (Table 4). If this increase was coming from dissolution of stream bed carbonates, we expect the ^{14}C age of the water to increase since soil carbonates have old ^{14}C signatures.

Bottom waters in the lakes are old and no obvious mechanism can be evoked to explain an increased reservoir effect with depth. There has been no indication that deep groundwater flow is occurring to any significant degree. Marine influence in these lakes is negligible. There are freeze-dried seals and penguins in the area and even on the lake ice surface, but they in no way interact with the lakes, nor are they abundant enough to account for the old ages, as in some Antarctic peninsula lakes (e.g. Björck et al., 1991). Sea birds are rare in the valleys, and therefore guano input is not a factor. The only factor we can use to explain the level of ^{14}C in the bottom of the lakes, is the actual age of the water, plus the surface reservoir. In the case of Lake Hoare this implies a bottom water date of ~1100 yr B.P., and ~8000 yr B.P. for Lake Bonney. These dates are coincident with inferred major drawdowns in the history of each lake. Lyons et al. (1999) have suggested that Lake Hoare evaporated to dryness ~1200 yr B.P. Stuiver et al. (1981) suggest that another major drying event in Taylor Valley occurred between 9800 and 10,800 yr B.P., although the state of Lake Bonney during this time is not certain. Our data certainly do not agree with the conclusions of Hendy et al. (1977) that the East Lobe of Lake Bonney was convecting and ice-free 5000 yr ago, otherwise we would expect it to have equilibrated with modern carbon at that time.

We also dated POC from two water samples collected in Lake Hoare in 1988. These samples display an exceptionally old relict signal (Table 2), and the shallow sample dates older than the deeper sample. We are not certain of the source of the old carbon, or the significance of the reversed stratigraphy. The main problem is that the components of POC are not known. There is some component of living planktonic algae, but this must be a small contribution. Possible sources for old POC could be debris that has been blown around the valleys and made its way into the lake, and material that is resuspended during lake level fluctuations over time. The deeper sample coincides with the chlorophyll-a maximum in Lake Hoare, and its younger apparent age could reflect the larger contribution by living material. Before any of these questions can be answered, the components of POC need to be more thoroughly investigated.

To test the effectiveness of ^{14}C in dating lake bottom sediments, we dated 8 microbial mat samples from a single Lake Hoare sediment core. Squyres et al. (1991) performed ^{14}C dating on Lake Hoare sediments, but their dates were obtained from several different cores and areas of the lake, and could not be used for geochronology, as they followed no coherent pattern. Our goal was to see if there is a down-core trend in organic mat dates. The results (Fig. 2) show there is a very good correlation ($r^2 = 0.97$) between core depth and ^{14}C age. Assuming that the top of the core represents modern material, the slope of the best fit line gives the sedimentation

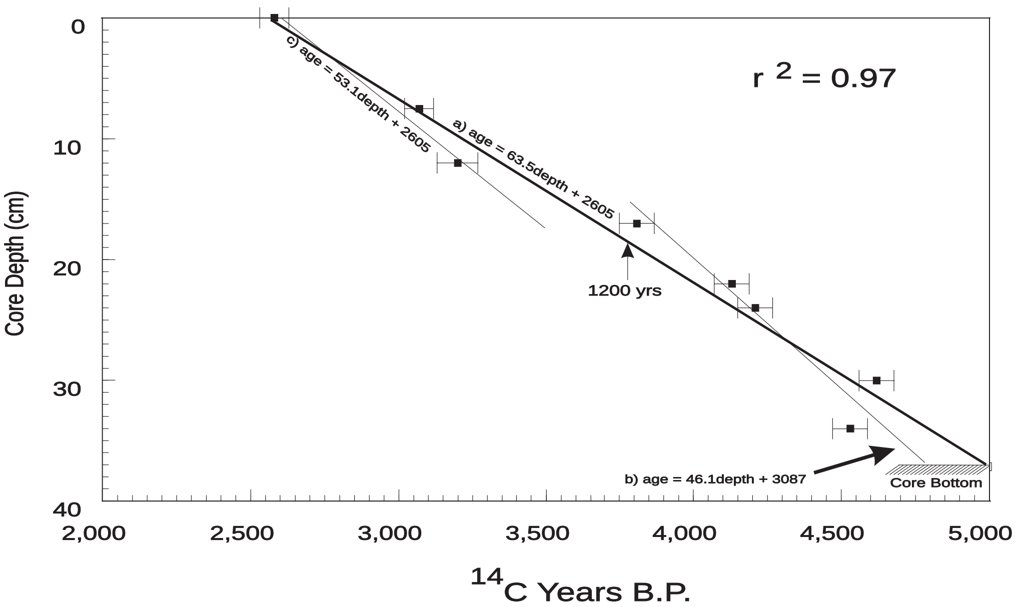

Fig. 2. AMS ^{14}C age of buried microbial mats with respect to depth in a Lake Hoare core from dive hole 2 (~11 m water depth). The calculated date of a postulated drying event in the lake is shown. Three best fit lines are calculated involving *a)* all the data ($n = 8$), and *b)* data prior to the postulated drying event.

rate of 0.015 cm yr^{-1} and suggests a basal date for this core of ~2400 yr B.P. The sedimentation rate is close to the 0.01 cm yr^{-1} suggested by Squyres et al. (1991). The basal date and continuous nature of the core profile do not support the conclusions of Lyons et al. (1999) who suggest on the basis of stable isotope data that Lake Hoare evaporated completely ~1200 yr ago. However, it is possible to create a scenario where there is an unconformity in the core representing a brief period where the sediments were exposed subaerially, and deflated a small amount. However, we did not observe any physical evidence of subaerial exposure (such as a lag deposit) in the core. Nevertheless, if lag material were not available, and these sands were somewhat sheltered by local topography, minor deflation without a lag deposit is conceivable. Interestingly, Spaulding et al. (1997) found an abrupt change in the diatom community recorded in a core taken nearby and ~4 cm lower in the profile from where we have identified the 1200 year event. This observation supports the idea that the site of our core was exposed subaerially around 1200 yr B.P.

Results of the comparison of ^{14}C ages in two Victoria Valley lacustrine sand mound deposits are shown in Table 5. In both cases carbonate ^{14}C ages are older than microbial mat ages for the same deposit, although the difference is much bigger in one case than in the other. The greater age of the carbonate deposit may be a diagenetic artifact. As discussed above, clasts in Taylor Valley soil have carbonate crusts and as shown in Table 1, the soil carbonate is variable and old. The authigenic carbonates presented in Table 5 may have minor soil carbonate crust on their surface that is contributing to the older signal. The organic mat would not show contamination from soil carbonates because they were pre-treated with acid before analysis.

Table 5
Comparison of ^{14}C dates of lacustrine sans mounds in Victoria Valley

Sample Location	^{14}C age (yr B.P.)
Victoria Valley lacustrine sand mound, 405.1 m a.s.l.	
Microbial mat	11310 ± 270
Authigenic carbonate	11990 ± 150
Victoria Valley lacustrine sand mound, 405.3 m a.s.l.	
Microbial mat	8495 ± 150
Authigenic carbonate	11380 ± 150

4. Thermoluminescence

4.1. TL background

Detrital minerals (mainly feldspars) behave as radiation dosimeters that can be reset by daylight and that emit a burial-age-dependent luminescence signal in the laboratory (Aitken, 1985; Berger, 1995). Luminescence sediment-dating methods are well suited to measuring deposition ages of deposits either lacking useful carbon for ^{14}C dating or that are older than the usual 30–40 ka limit of ^{14}C dating. Wintle and Huntley (1979, 1980) first demonstrated that TL in unheated sediment behaves as if there were a light-sensitive component (TLS) and a light-insensitive (TLIS) part. In most waterlain sediment the TLS component may not be fully zeroed at deposition (Berger, 1990), thus necessitating systematic tests of the dating method in each depositional environment. In more detail, daylight empties light-sensitive electron traps (thermally stable lattice-defect sites), and after burial, natural ionizing radiations repopulate emptied traps. In the laboratory, traps are again emptied, by heating, with electron-hole recombination producing luminescence. When the TL is related to an 'equivalent dose' (D_E) by use of calibrated laboratory radiation sources, and an effective radiation dose rate is measured in separate experiments on separate fractions, then a TL age $= D_E/D_R$ where D_R is effective dose rate. Common units are grays (1 Gy =1 joule/kg) for D_E and Gy/ka for D_R.

Because of the low solar angles and long polar darkness, we attempted first to test the effectiveness of zeroing of light-sensitive TL in a variety of depositional settings by sampling freshly deposited material, and second to apply TL dating procedures developed for unheated lower latitude sediments. The procedures we employed are those described for modern analogs of mid-latitude eolian and waterlain silts (Berger, 1990). TL dating tests on Arctic lake sediments (Berger and Anderson, 1994, 1996) have shown effective signal zeroing to be attainable in at least some high-latitude lacustrine settings.

4.2. TL methods

During the 1993–1994 austral summer in Taylor Valley, light-tight containers were used to collect suspensions from ephemeral streams, ice debris, and box cores (Table 6). Interior portions of the collected samples were removed from the light-tight containers in dim, amber laboratory illumination, for TL work. Usual steps (e.g. Wintle and Huntley, 1980) were followed to remove carbonate and organics and produce 4–11 μm polymineral grains. In estimating the D_E values, laboratory bleaching similar to protocols of Berger (1990) were employed. D_E values were determined from extrapolations of pairs of intersecting beta-dose-response curves. The back extrapolation of such pairs of curves is equivalent to extrapolation to a dose for which the light sensitive TL (or fraction thereof) is zero. This gives the deposition time (via the age equation). To remove thermally unstable and anomalous-fading TL components from the laboratory-irradiated subsamples prior to TL readout, subsamples were held at 75ºC for four days prior to signal readout. Details of our methods and results will be presented elsewhere.

4.3. TL results and discussion

The main results are listed in Table 6. Light sensitive TL has been effectively zeroed for ice-surface sample PDR-16 and suspension sample PDR-70 (see D_E values). Zeroing for suspension sample PDR-71 was less effective. This distinction is interesting because PDR-71 comes from Anderson Creek, a stream that enters Lake Hoare adjacent to the Canada Glacier, whereas the suspension PDR-70 (Huey Creek) represents an ice-distal transport. Many prior TL zeroing tests (Berger, 1988, 1995) have shown that grains transported only short distances from sub-glacial or supra-glacial meltwater flows will carry ineffectively zeroed light-sensitive TL. D_E values were not measured for sand dune sample PDR-12; rather, an optical-bleaching test (TL vs. bleaching time) was performed. This indicated effective zeroing, although the data reproducibility were poor (Doran, 1996).

For box-core sample PDR-1 (Table 6), the D_E values lie in the range 3–7 Gy, with no depth trend. The corresponding TL ages suggest that a zero-point or relict TL age of ~1.4 ka is carried to the lake bottom, by the integrated effects of available transport processes. Such a relict TL age would introduce a significant systematic error into Holocene-age lake

Table 6
Sample, dose rates and TL ages

Sample	Location (Fig. 1)	Depth [a] (cm)	D_E [b] (Gy)	D_R [c] (Gy/ka)	TL age (ka)
Suspension					
PDR-70	A — Huey Creek		0.4 ± 3.6		
PDR-71	B — Anderson Creek		5.4 ± 3.6		
Lake Hoare ice surface					
PDR-16	C — east end of lake	0–4	−1.1 ± 1.4	2.53 ± 0.18	−0.44 ± 0.56
Lake Hoare sand dune					
PDR-12	E — near Canada Glacier	36			
Lake Hoare Ekman box core					
PDR-1A	D — 333 m from Canada Glacier, Hole 1	top mat	7.1 ± 3.5	3.98 ± 0.35	1.79 ± 0.89
PDR-1B	D — 333 m from Canada Glacier, Hole 1	2nd layer (1 cm sand)	3.3 ± 1.8	3.41 ± 0.30	0.97 ± 0.54
PDR-1C	D — 333 m from Canada Glacier, Hole 1	3rd layer (thin mats)	5.27 ± 0.98	3.60 ± 0.30	1.46 ± 0.30
PDR-1D	D — 333 m from Canada Glacier, Hole 1	bottom sand and mats	6.0 ± 1.5	4.20 ± 0.32	1.43 ± 0.37
Ekman weighted mean age =				1.40 ± 0.21	1.40 ± 0.21

[a] Depth from top of deposit. For the Ekman sample 1, total length was 10–15 cm.
[b] Equivalent dose. Analytical errors are $\pm 1\sigma$.
[c] Dose rate.

bottom samples, but would be unresolvable in TL ages older than 15–20 ka. If further tests on other box-core samples reveal a consistent relict age of ~1.4 ka for Lake Hoare, then this could be used to correct TL ages lower in the sediment profile.

Independent of our work, Krause et al. (1997) tested infrared-stimulated luminescence (IR-OSL) dating assumptions in the area of the Schirmacher Oasis in east Antarctica. The potentially greater precision attainable with IR-OSL and our own limited results suggest that future tests and applications in the McMurdo Dry Valleys should also employ the IR-OSL dating methods.

5. ^{210}Pb and ^{137}Cs

5.1. Radiochemistry background

The decay series of ^{238}U includes ^{222}Rn which escapes to the atmosphere at an average rate of 42 atoms min^{-1} cm^{-2} of land surface. Once in the atmosphere ^{222}Rn decays through a series of short-lived daughters to ^{210}Pb (half-life = 22.6 yr). Rain and snow remove ^{210}Pb from the atmosphere on average over a 10-day period (Faure, 1986). When ^{210}Pb enters a lake, it is scavenged by particles and settles to the bottom. One complication occurs in lake sediments because of minerogenic ^{210}Pb formation from the decay of ^{226}Ra. This fraction of the ^{210}Pb flux is often referred to as the 'supported' fraction, and can be accounted for by measuring and subtracting the activity of ^{226}Ra in each sample. In this way, the activity of ^{210}Pb in lake sediments can be used to make inferences regarding recent sedimentation characteristics. The incoming flux of ^{210}Pb varies globally. Reliable information on the fallout of ^{210}Pb in the southern hemisphere is sparse. Appleby et al. (1995) suggest a reasonable figure for Antarctic snowfall is 567 ± 216 pCi ^{210}Pb m^{-3} water equivalent. This is very close to the value of 551 ± 50 pCi ^{210}Pb m^{-3} we have calculated from analysis of the surface layer of a snow pit on the Newall Glacier, ~1000 m above and 5 km to the north of Lake Hoare (Lyons, unpubl. data).

Atmospheric nuclear bomb testing, which began in the 1950s and peaked in 1963, generated the artificial radionuclide ^{137}C (half-life = 30.0 yr). This radionuclide is generally assumed to behave similarly to ^{210}Pb in the atmosphere (i.e. short residence time), and in the lake (i.e. scavenged by sediments). Fallout of ^{137}Cs is expected to be an order of magnitude lower than rates in mid-latitudes of the northern hemisphere (Hardy et al., 1973).

5.2. Radiochemistry methodology

Cores for this analysis were collected in 1981 by vibracorer, with the assistance of SCUBA-equipped divers (Love et al., 1982). They were frozen immediately after collection and remained that way until sectioned. In the lab, cores were split, and one half archived. The other half was cleaned with a mechanical scraper and sectioned at 0.5 cm intervals for the top 5 cm of core. ^{210}Pb and ^{137}Cs analyses were carried out by gamma spectrometry at Keck Radiochemistry Laboratory.

5.3. Radiochemistry results and discussion

The ^{210}Pb and ^{137}Cs profiles from Lake Fryxell, Lake Joyce, and Lake Hoare are shown in Table 7. These values are relatively low, and often below detection limits. ^{210}Pb ranges up to 1.12 pCi g^{-1} in Lake Joyce, and up to 1.07 pCi g^{-1} in Lake Fryxell. By way of comparison, ^{210}Pb levels in White Smoke Lake, another perennially ice-covered antarctic lake at a lower latitude, exceed 40 pCi g^{-1} near the sediment surface (Doran, 1996). It appears that there may be a ^{137}Cs peak in the top interval of Lake Joyce sediments, indicating that the sedimentation rate is $<$0.02 cm yr^{-1}. This rate is in line with the Lake Hoare sedimentation rate determined from ^{14}C above of 0.015 cm yr^{-1}. Lake Fryxell does not display any ^{137}Cs peak.

We know very little about sedimentation rates in Lake Joyce, but in Lake Fryxell, Lawrence (1982) estimated a sedimentation rate for the top of Lake Fryxell sediments of 0.007 cm yr^{-1}. Using this value along with an estimate of the ^{210}Pb flux to the area, annual precipitation, and sediment bulk density, we can calculate a range of ^{210}Pb values we would expect to see given rapid (i.e. normal) sedimentation of the radionuclide. For the bulk density we have used an average value from Lake Hoare sediments of 1.5 g cm^3 (Squyres et al., 1991). There have only been three years where annual precipitation was measured in the dry valleys (manually by winter-over researchers at Lake Vanda Camp). These three values are 7, 82, and 115 mm yr^{-1} (Bromley, 1985). If we apply this range of precipitation to the Antarctic ^{210}Pb concentration cited above (567 $\pm$ 216 pCi m^3 water equivalent), we get a range of dry valley ^{210}Pb flux of between 4.1 $\pm$ 1.6 and 499.0 $\pm$ 18.9 pCi m^{-2} yr^{-1}. In Lake Fryxell, assuming the ^{210}Pb flux and mineral sedimentation rates have not changed, and accounting for the core being stored for 15 years between collection and analysis, this should produce a ^{210}Pb activity of between 0.35 and 5.89 pCi g^{-1} in the top 0.5 cm of the sediment core. The high end of this range is comparable to levels recorded by Appleby et al. (1995) in lakes on the Antarctic Peninsula, and is much lower than levels recorded in White Smoke Lake surface sediments in Bunger Hills, east Antarctica (Doran, 1996). Given we recorded low ^{210}Pb levels throughout our core, and did not record any ^{210}Pb in the top interval, our calculations lead us to one of two conclusions: (1) the top of the core is missing (which we don't think is the case based on communication with the samplers); or (2) ^{210}Pb and presumably ^{137}Cs are not getting delivered to the water column (or the bottom of the lake). Without further data we can not answer these questions, but will hypothesize that the perennial ice cover is playing some role in retarding the deposition of ^{210}Pb and ^{137}Cs, and that inflow rates are sufficiently low that delivery of basin-deposited ^{210}Pb and ^{137}Cs (i.e. on glacier surfaces) is very low compared to other regions of the Antarctic. Analysis of the activity of the two radionuclides in the perennial ice cover and basins of these lakes would help test this hypothesis.

6. Paleomagnetism

6.1. Paleomagnetism background

Reversals of the Earth's magnetic field, and geomagnetic secular variation can be used to make temporal inferences about a lake sediment profile, although the time scales required for lake sediments generally can not make use of the reversal chronology.

Table 7
Results of ^{210}Pb and ^{137}Cs dating

Core depth (cm)	Excess ^{210}Pb [pCi (g dry wt)$^{-1}$]	1σ	^{137}Cs [pCi (g dry wt)$^{-1}$]	1σ
Lake Joyce				
0.0–0.5	0.56	0.13	0.20	0.05
0.5–1.0	1.06	0.11	B.D.	–
1.0–1.5	B.D. [a]	–	B.D.	–
1.5–2.0	B.D.	–	0.01	0.01
2.0–2.5	0.50	0.06	B.D.	–
2.5–3.0	0.27	0.07	B.D.	–
3.0–3.5	0.82	0.11	B.D.	–
3.5–4.0	0.69	0.13	0.03	0.02
4.0–4.5	1.12	0.11	B.D.	–
4.5–5.0	0.78	0.14	0.02	0.02
Lake Fryxell [b]				
0.0–0.5	B.D.	–	0.01	0.02
0.5–1.0	1.07	0.22	B.D.	–
1.0–1.5	0.26	0.06	0.03	0.01
1.5–2.0	0.24	0.05	0.03	0.01
2.0–2.5	1.00	0.12	0.05	0.02
2.5–3.0	B.D.	–	0.01	0.01
3.0–3.5	B.D.	–	0.03	0.01
3.5–4.0	B.D.	–	0.01	0.02
4.0–4.5	0.96	0.09	B.D.	–
4.5–5.0	0.39	0.06	B.D.	–
Lake Fryxell [c]				
0.0–1.0	0.35	0.12	0.03	0.03
1.0–2.0	B.D.	–	0.02	0.02
2.0–2.7	1.12	0.27	B.D.	–
2.7–3.7	0.01	0.07	0.01	0.01
3.7–4.7	B.D.	–	B.D.	–
4.7–5.7	0.09	0.07	B.D.	–
5.7–6.7	0.93	0.19	B.D.	–
Lake Hoare				
0.0–1.0	0.50	0.11	B.D.	–
1.0–2.0	0.48	0.10	B.D.	–
2.0–3.0	0.58	0.09	B.D.	–
3.0–4.0	0.47	0.09	B.D.	–
4.0–5.0	0.54	0.08	B.D.	–
5.0–6.0	0.13	0.07	B.D.	–
6.0–7.0	0.02	0.09	0.02	0.01
7.0–8.0	0.11	0.10	0.05	0.01
10.0–11.0	0.31	0.12	B.D.	–
14.0–15.0	0.22	0.12	B.D.	–
16.0–17.0	0.42	0.11	B.D.	–
19.0–20.0	0.21	0.11	B.D.	–
22.0–23.0	0.42	0.12	B.D.	–
23.5–24.5	B.D.	–	B.D.	–
27.0–28.0	0.29	0.11	B.D.	–
30.0–31.0	0.31	0.09	B.D.	–
33.0–34.0	B.D.	–	B.D.	–

[a] B.D. = below detection limits. [b] Lake Fryxell core collected in 1982. [c] Lake Fryxell core collected in 1990.

Records of short- and long-term secular behavior of the Earth's paleomagnetic vector can and have been recorded in lake sediments (e.g. Creer et al., 1983).

6.2. Paleomagnetism methods

A sediment core was collected from Lake Hoare in December 1994 using a Nesje percussion corer (Nesje, 1992) and 7.6 cm diameter downspout PVC pipe. The core was split and logged upon collection, and kept at ~5ºC. In the lab a 'u-channel' (2.5 cm square acrylic tube with one side cut off) was pushed into the sediment until refusal. A sediment cutter was used to remove the sediment-filled u-channel from the core, and the open side of the u-channel was closed with the previously removed side of the tube. The sealed u-channel was then cut into cubes, 2.5 cm on each side. Each cube was then analyzed for natural remanent magnetization on an SCI cryogenic magnetometer in the Paleomagnetism Laboratory at the University of Nevada, Reno.

6.3. Paleomagnetism results and discussion

The results of the paleomagnetic analysis are shown in Fig. 3. The samples have a strong natural remanence, but magnetic declination and inclination appear random. These results seem to indicate that the grains in Lake Hoare are too large to allow torquing by the magnetic field and the sediments do not acquire a reliable detrital magnetic remanence. Although paleomagnetism is not viable in Lake Hoare, the strong remanence is promising and we feel this technique could yield interesting results given finer sediment.

7. Overall summary and conclusions

Our data have shown that many stream, near-shore, and shallow lake waters in the McMurdo Dry Valleys are near-modern with respect to ^{14}C age. To our knowledge, we have presented among the youngest ^{14}C dates collected for any Antarctic lacustrine system. This is surprising, until it is realized that very few modern lacustrine/riverine dates exist. We propose that the mode of glacial water input has a great impact on the extent of the carbon reservoir in the surface waters of the lakes. If glacial melt travels for relatively long distances in a soil-based stream, the DIC in the stream water is allowed to 'modernize' prior to entering the lake (e.g. Lake

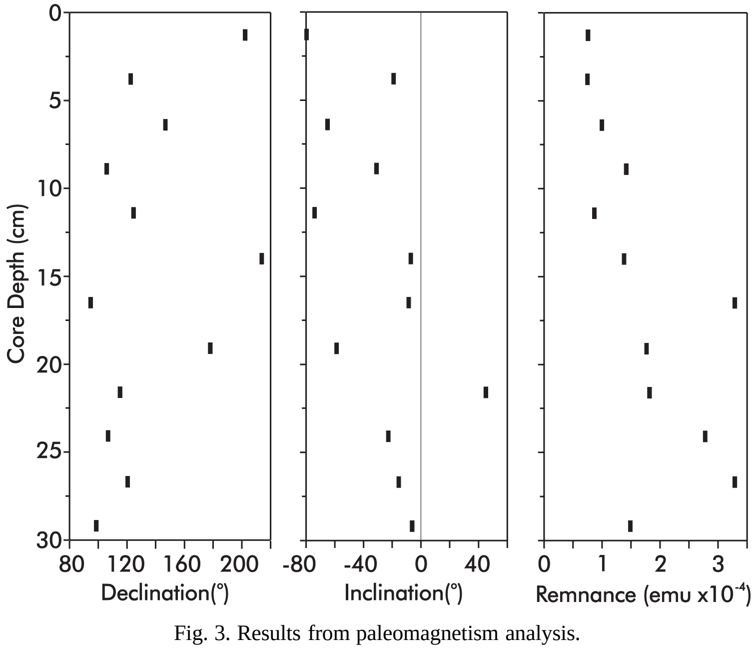

Fig. 3. Results from paleomagnetism analysis.

Fryxell). If glacial input is directly from the glacier surface (i.e. a waterfall or direct thermal abrasion beneath the lake surface), or a very short stream is used to deliver the glacial melt to the lake, then the old DIC does not have sufficient time to equilibrate with modern CO_2 before entering the lake where it is sealed from the atmosphere by the thick lake ice (e.g. Lake Hoare). Once in the lake, DIC ^{14}C decays naturally. Accepting this argument allows us to calculate for the east lobe of Lake Bonney a bottom age of ~8000 yr B.P., and for Lake Hoare a bottom age of 1100 yr B.P. The latter date is co-incident with a major lake draw-down in the history of Taylor Valley speculated on by others. Results from dating POC in Lake Hoare water reflects a very old and undefined carbon source. A younger signal at the chlorophyll-a maximum may reflect the influence of living material. What impact the settling of this old POC has on sediment surface dates is uncertain, but if the settling is constant over time, it will simply be included in the carbon reservoir.

Prior to this study, ^{14}C was not thought reliable enough in this environment to accurately date sediment profiles. However, our profile of eight microbial mat dates in a Lake Hoare core had an extremely high correlation between core depth and ^{14}C age. Furthermore, the date for the surface of the sediment core agreed with the bottom water date to within 90 yr B.P. This leads us to believe that the surface of the core carries a carbon reservoir of ~2600 yr which can be applied to correct ages lower in the profile, giving the core a basal date of ~2400 yr B.P. For this to be accurate, there must be an unconformity in the profile around 1200 yr B.P. to account for the widely documented 1200 yr drawdown of Taylor Valley lakes, and the age of the bottom water. Furthermore, if bottom water DIC is aging at the same rate as the microbial mats getting buried, we would expect uniform ages throughout the core. The fact that microbial mats get older with depth in the core points to some yet to be discovered component in the ^{14}C cycling in these systems.

It is clear from our results that ^{14}C ages from lacustrine deposits can not be interpreted without knowing something about the manner in which they were laid down. Deltaic and lake edge deposits far removed from glaciers appear to carry no carbon reservoir, and are reliable. In lake bottom deposits where it is possible to get a surface sediment date (or the age of the bottom water) to correct the profile to, reliable dates may be obtainable, but further research is required before this can be concluded. Deep-water paleolake deposits are problematic where no surface age correction is possible. Without this correction, there is no way to account for the age of the lake (and in lakes like Bonney, the error could be considerable). For samples like the Victoria Valley lacustrine sand mounds, we have no indication of the size of the carbon reservoir recorded in the deposits. In many cases such as this, utilization of a second dating technique to support ^{14}C chronologies will be necessary. A summary of the outcome of alternative dating techniques tested in this study is shown in Table 8. Of these, TL and IR-OSL hold the most promise for supporting ^{14}C chronologies.

Table 8
Summary of alternative (to ^{14}C) dating employed on Dry Valley lake sediments and results

Paleomagnetism	Lake Hoare	sediment core samples	good magnetic remanence material too coarse for rotation of grains
^{210}Pb	Lake Joyce	sediment core samples	no signal
	Lake Hoare	sediment core samples	no signal
	Lake Fryxell	sediment core samples	no signal
^{137}Cs	Lake Joyce	sediment core samples	no signal
	Lake Hoare	sediment core samples	no signal
	Lake Fryxell	sediment core samples	no signal
U/Th	Lake Hoare	sediment core carbonate	not enough authigenic material in relatively young sediment samples
	Lake Vida	paleolacustrine carbonate	reasonable date obtained
Thermoluminescence	Lake Hoare	various sediments	good potential with ~1000 yr relict signal on lake bottom

Acknowledgements

This research was funded by the National Aeronautics and Space Administration's Exobiology Program (NAGW 1947), and the National Science Foundation's Office of Polar Programs (OPP-9211773). Portions of this work were performed under the auspices of the US DOE at Lawrence Livermore National Lab, under contract W-7405-Eng-48. Desert Research Institute's Quaternary Sciences Center provided funding for the TL work. We would like to thank B. Parker and G.M. Simmons Jr. for providing us with core material from Lakes Joyce and Fryxell, Diane McKnight for the POC samples, Robert Karlin for the use of his Paleomagnetism Laboratory at the University of Nevada, and Laura Williams for her assistance in the TL lab. We would like to thank P.G. Appleby and an anonymous reviewer for comments that improved the manuscript.

References

Aitken, M.J., 1985. Thermoluminescence Dating. Academic Press, New York

Appleby, P.G., Jones, V.J., Ellis-Evans, J.C., 1995. Radiometric dating of lake sediments from Signy Island (maritime Antarctic): evidence of recent climatic change. J. Paleolimnol. 13, 179–191.

Armitage, K.B., House, H.B., 1962. A limnological reconnaissance in the area of McMurdo Sound, Antarctica. Limnol. Oceanogr. 7, 36–41.

Barrett, P.J., Grindley, P.J., Webb, P.N., 1972. The Beacon Supergroup of east Antarctica. In: Adie, R.J. (Ed.), Antarctic Geology and Geophysics. Int. Union Geol. Sci. Ser. B 1, 319–332.

Berger, G.W., 1988. Dating Quaternary events by luminescence. In: Easterbrook, D.J. (Ed.), Dating Quaternary Sediments. Geol. Soc. Am. Spec. Pap. 227, 13–50.

Berger, G.W., 1990. Effectiveness of natural zeroing of the thermoluminescence in sediments. J. Geophys. Res. 95, 12375–12397.

Berger, G.W., 1995. Progress in luminescence dating methods for Quaternary sediments. In: Rutter, N.W., Catto, N. (Eds.), Dating Methods for Quaternary Deposits. Geol. Assoc. Can. GEOtext 2, 81–103.

Berger, G.W., Anderson, P.M., 1994. Thermoluminescence dating of an Arctic lake core from Alaska. Quat. Sci. Rev. 13 (567), 497–501.

Berger, G.W., Anderson, P.A., 1996. Thermoluminescence dating tests on a core from Joe Lake, northwestern Alaska. 8th Int. Conf. Luminescence and Electron Spin Resonance Dating, 22–26 April, Australian National Univ., Canberra, Progr. Abstr.

Berkman, P.A., Forman, S.L., 1996. Pre-bomb radiocarbon and the reservoir correction for calcareous marine species in the Southern Ocean. Geophys. Res. Lett. 23 (4), 363–366.

Bird, M., Chivas, C., Radnell, C., Burton, H., 1991. Sedimentological and stable-isotope evolution of lakes in the Vestfold Hills, Antarctica. Palaeogeogr., Palaeoclimatol., Palaeoecol. 84, 109–130.

Björck, S., Hjort, C., Ingólfsson, Ó., Skog, G., 1991. Radiocarbon dates from the Antarctic Peninsula region — Problems and potential. In: Lowe, J.J. (Ed.), Radiocarbon Dating: Recent Applications and Future Potential. Quat. Res. Assoc. Quat. Proc. 1, 55–65.

Björck, S., Olsson, S., Ellis-Evans, C., Håkansson, H., Humlum, O., Manuel de Lirio, J., 1996. Late Holocene palaeoclimatic records from lake sediments on James Ross Island, Antarctica. Palaeogeogr., Palaeoclimatol., Palaeoecol. 121, 195–220.

Broecker, W.S., 1963. Radiocarbon ages of Antarctic materials. Polar Rec. 11, 472–473.

Bromley, A.M., 1985. Weather observations Wright Valley, Antarctica. N.Z. Meteorol. Serv. Inf. Publ. 11, 37 pp.

Burton, H.R., 1981. Chemistry, physics and evolution of Antarctic saline lakes: a review. Hydrobiologia 82, 339–362.

Cartwright, K., Harris, H.J.H., 1981. Hydrology of the dry valley region, Antarctica. In: McGinnis, L.D. (Ed.), Dry Valley Drilling Project. AGU Ant. Res. Ser. 33, 193–214.

Chinn, T.J.H., 1982. Hydrology and climate in the Ross Sea area. J. R. Soc. N.Z. 11 (4), 373–386.

Chinn, T.J.H., 1993. Physical hydrology of the dry valley lakes. In: Green, W., Friedmann, E.I. (Eds.), Physical and Biochemical Processes in Antarctic Lakes. AGU Ant. Res. Ser. 59, 1–52.

Clow, G.D., McKay, C.P., Simmons Jr., G.M., Wharton Jr., R.A., 1988. Climatological observations and predicted sublimation rates at Lake Hoare, Antarctica. J. Climate 1, 715–728.

Creer, K.M., Tucholka, P., Barton, C.E., 1983. Palaeomagnetism of unconsolidated sediments. In: Creer, K.M., Tucholka, P., Barton, C.E. (Eds.), Geomagnetism of Baked Clays and Recent Sediments. Elsevier, Amsterdam, pp. 197–272.

Decker, E.R., Bucher, G.L., 1982. Preliminary geothermal studies in the Ross Island–Dry Valley region. In: Craddock, C. (Ed.), Antarctic Geoscience. Univ. Wisconsin Press, Madison, WI, pp. 887–894.

DeMaster, D.J., Dunbar, R.B., Gordon, L.I., Leventer, A.R., Morrison, J.M., Nelson, D.M., Nittrouer, C.A., Smith Jr., W.O., 1992. Cycling and accumulation of biogenic silica and organic matter in high-latitude environments: the Ross Sea. Oceanography 5 (3), 146–153.

Denton, G.H., Armstrong, R.L., Stuiver, M., 1971. The late Cenozoic glacial history of Antarctica. In: Turekian, K.K. (Ed.), The Late Cenozoic Glacial Ages. Yale Univ. Press, New Haven, CT, pp. 267–306.

Domack, E.W., Jull, A.J.T., Anderson, J.B., Linick, T.W., Williams, C.R., 1989. Application of tandem accelerator mass-spectrometer dating to late Pleistocene–Holocene sediments of the East Antarctic continental shelf. Quat. Res. 31, 277–287.

Doran, P.T., 1996. Paleolimnology of Perennially Ice-covered Antarctic Oasis Lakes. Ph.D. Dissertation, Univ. Nevada, Reno.

Doran, P.T., Wharton Jr., R.A., Lyons, W.B., 1994. Paleolimnology of the McMurdo Dry Valleys, Antarctica. J. Paleolimnol. 10, 85–114.

Faure, G., 1986. Principles of Isotope Geology. Wiley, New York, NY, 589 pp.

Friedman, I., Rafter, A., Smith 1995., G.I., 1995. A thermal, isotopic, and chemical study of Lake Vanda and Don Juan Pond, Antarctica. AGU Ant. Res. Ser. 67, 47–74.

Gordon, J.E., Harkness, D.D., 1992. Magnitude and geographic variation of the radiocarbon content in antarctic marine life: implications for reservoir corrections in radiocarbon dating. Quat. Sci. Rev. 11, 697–708.

Gore, D.B., 1997. Blanketing snow and ice: constraints on radiocarbon dating deglaciation in East Antarctic oases. Ant. Sci. 9 (3), 336–346.

Green, W.J., Angle, M., Chave, K., 1988. The geochemistry of Antarctic streams and their role in the evolution of four lakes in the McMurdo Dry Valleys. Geochim. Cosmochim. Acta 52, 1265–1274.

Hammar, J., 1989. Freshwater ecosystems of polar regions: Vulnerable resources. Ambio XVIII, 6–22.

Hardy, E.P., Krey, P.W., Volchok, H.L., 1973. Global inventory and distribution of fallout plutonium. Nature 241, 444–445.

Henderson, R.A., Prebble, W.M., Hoare, R.A., Popplewell, K.B., House, D.A., Wilson, A.T., 1965. An ablation rate for Lake Fryxell, Victoria Land, Antarctica. J. Glaciol. 6, 129–133.

Hendy, C.H., Wilson, A.T., Popplewell, K.B., House, D.A., 1977. Dating of geochemical events in Lake Bonney, Antarctica, and their relation to glacial and climate changes. N.Z. J. Geol. Geophys. 20, 1103–1122.

Heywood, R.B., 1984. Antarctic inland waters. Antarct. Ecol. 1, 279–344.

Krause, W.E., Krbetschek, M.R., Stolz, W., 1997. Dating of Quaternary lake sediments from the Schirmacher Oasis (east Antarctica) by infra-red stimulated luminescence (IRSL) detected at the wavelength of 560 nm. Quat. Sci. Rev. 16, 387–392.

Lawrence, M.J.F., 1982. Origin and occurrence of Antarctic Lacustrine Carbonates, with Special Reference to Lake Fryxell, Taylor Valley. M.Sc. Thesis, Univ. Waikato, New Zealand, 246 pp.

Libby, W.F., 1955. Radiocarbon Dating. Univ. Chicago Press, Chicago, IL, 175 pp.

Lopatin, B.G., 1972. Basement complex of the McMurdo 'Oasis', south Victoria Land. In Adie, R.J. (Ed.), Antarctic Geology and Geophysics. Int. Union Geol. Sci. Ser. B 1, 287–292.

Love, F.G., Simmons Jr., G.M., Wharton Jr., R.A., Parker, B.C., 1982. Methods for melting dive holes in thick ice and vibracoring beneath ice. J. Sediment. Petrol. 52, 644–647.

Lyons, W.B., Tyler, S.W., Wharton Jr., R.A., McKnight, D.M., 1999. Recent climatic history of Lakes Fryxell and Hoare: evidence from isotopic data. Geochim. Cosmochim. Acta. (in press).

Matsubaya, O., Sakai, H., Torii, T., Burton, H., Kerry, K., 1979. Antarctic saline lakes — stable isotopic ratios, chemical compositions and evolution. Geochim. Cosmochim. Acta 43, 7–25.

McNichol, A.P., Jones, G.A., Hutton, D.L., Gagnon, A.R., 1994. The rapid preparation of seawater $\sum CO_2$ for radiocarbon analysis at the National Ocean Sciences AMS facility. Radiocarbon 36, 237–246.

Melles, M., Verkulich, S.R., Hermichen, W.-D., 1994. Radiocarbon dating of lacustrine and marine sediments from the Bunger Hills, East Antarctica. Antarct. Sci. 6 (3), 375–378.

Nedell, S.S., Andersen, D.W., Squyres, S.W., Love, F.G., 1987. Sedimentation in ice-covered Lake Hoare, Antarctica. Sedimentology 34, 1093–1106.

Nesje, A., 1992. A piston corer for lacustrine and marine sediments. Arct. Alp. Res. 24, 257–259.

Palmisano, A.C., Simmons Jr., G.M., 1987. Spectral downwelling irradiance in an antarctic lake. Polar Biol. 7, 145–151.

Priddle, J., Heywood, R.B., 1980. Evolution of antarctic lake ecosystems. Biol. J. Linn. Soc. 14 (1), 51–66.

Rao, R.R., Killey, R.W.D., 1994. Quantitative separation and determination of inorganic and organic forms of total carbon and radiocarbon in natural waters and application at a radioactive waste management site. Radiochim. Acta 65, 63–74.

Simmons Jr., G.M., Wharton Jr., R.A., McKay, C.P., Nedell, S., Clow, G., 1986. Sand/ice interactions and sediment deposition in perennially ice-covered antarctic lakes. Antarct. J. U.S. 21, 217–220.

Smithson, S.B., Fikkan, P.R., Murphy, D.J., Houston, R.S., 1972. Development of Augen-gneiss in the ice-free valley area, south Victoria Land. In: Adie, R.J. (Ed.), Antarctic Geology and Geophysics. Int. Union Geol. Sci. Ser. B 1, 293–298.

Southon, J., Caffee, M.W., Davis, J.C., Moore, T.L., Proctor, I.D., Schumacher, B., Vogel, J.S., 1990. Nucl. Instr. Meth. B 52, 301–305.

Spaulding, S.A., McKnight, D.M., Stoermer, E.F., Doran, P.T., 1997. Diatoms in sediments of perennially ice-covered Lake Hoare, and implications for interpreting lake history in the McMurdo Dry Valleys of Antarctica. J. Paleolimnol. 17 (4), 403–420.

Spigel, R.H., Priscu, J.C., 1998. Physical Limnology of the McMurdo Dry Valley Lakes. In: Priscu, J.C. (Ed.), McMurdo Dry Valleys: A Cold Desert Ecosystem. AGU Ant. Res. Ser. 72, 153–187.

Stuiver, M., Denton, G.H., Hughes, T.J., Fastook, J.L., 1981. History of the marine ice sheet in west Antarctica during the last glaciation, a working hypothesis. In: Denton, G.H., Hughes, T.H. (Eds.), The Last Great Ice Sheets. Wiley-Interscience, New York, NY, pp. 319–436.

Squyres, S.W., Andersen, D.W., Nedell, S.S., Wharton Jr., R.A., 1991. Lake Hoare, Antarctica: Sedimentation through a thick perennial ice cover. Sedimentology 38, 363–379.

Wharton Jr., R.A., Simmons Jr., G.M., McKay, C.P., 1989. Perennially ice-covered Lake Hoare, Antarctica: physical environment, biology, and sedimentation. Hydrobiologia 172, 305–320.

Wharton Jr., R.A., McKay, C.P., Clow, G.D., Andersen, D.T., 1993. Perennial ice covers and their influence on Antarctic lake ecosystems. In: Green, W., Friedmann, E.I. (Eds.), Physical and Biochemical Processes in Antarctic Lakes. AGU Ant. Res. Ser. 59, 53–72.

Wintle, A.G., Huntley, D.J., 1979. Thermoluminescence dating of a deep-sea sediment core. Nature 279, 710–712.

Wintle, A.G., Huntley, D.J., 1980. Thermoluminescence dating of ocean sediments. Can. J. Earth Sci. 17, 348–360.

Zale, R., Karlén, W., 1989. Lake sediment cores from the Antarctic Peninsula and surrounding islands. Geogr. Ann. 71A, 3–4.