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Exobiology Site Selection for Future Mars Missions: Martian Paleolake Sediments and Terrestrial Analogs

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Exobiology Site Selection for Future Mars Missions: Martian Paleolake Sediments and Terrestrial Analogs

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This research was conducted to establish the scientific framework for the exobiological study of sediments on Mars and to encourage the selection of these sedimentary deposits as sampling sites for future Mars missions. A study was completed on the Antarctic Dry Valley lakes (terrestrial analogs of the purported martian paleolakes) and their sediments that allowed the development of quantitative models relating environmental factors to the nature of the biological community and sediment forming processes. Publications resulting from this research include:

Baublis, J.A., R.A. Wharton, Jr., and P.A. Volz. 1991. Diversity of micro-fungi isolated in an Antarctic dry valley. *J. Basic Microbiol.* 31: in press.

Squyres, S.W., D.W. Andersen, S.S. Nedell, and R.A. Wharton, Jr. 1991. Lake Hoare, Antarctica: sedimentation through a thick perennial ice cover. *Sedimentology* , in press.

McKay, C.P., R.L. Mancinelli, C.R. Stoker, and R.A. Wharton, Jr. 1990. The possibility of life on Mars during a water-rich past. Mathews, M.S. (ed.), University of Arizona Press, in press.

Andersen, D.T., C.P. McKay, R.A. Wharton, Jr., J.D. Rummel. 1990. An Antarctic research outpost as a model for planetary exploration. *J. Brit. Interplanetary Soc.*, 43:507-512.

Wharton, R.A., Jr., C.P. McKay, R.L. Mancinelli, and G.M. Simmons, Jr. 1989. Early martian environments: the Antarctic and other terrestrial analogs. *Adv. Space Res.* 6:147-153.

Palmisano, A.C., R.A. Wharton, Jr., S.E. Cronin, and D.J. Des Marais. 1989. Lipophiliic pigments from the benthos of a perennially ice-covered Antarctic lake. *Hydrobiologia* 178:73-80.

Wharton, R.A., Jr., G.M. Simmons, Jr., and C.P. McKay. 1989. Perennially ice-covered Lake Hoare, Antarctica: physical environment, biology, and sedimentation. *Hydrobiologia* 172:306-320.

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Perennially ice-covered Lake Hoare, Antarctica: physical environment, biology and sedimentation

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Key words: Antarctica, climatic changes, gas solubility, ice-covered lakes, Lake Hoare

Abstract

Lake Hoare (77° 38′ S, 162° 53′ E) is a perennially ice-covered lake at the eastern end of Taylor Valley in southern Victoria Land, Antarctica. The environment of this lake is controlled by the relatively thick ice cover (3–5 m) which eliminates wind generated currents, restricts gas exchange and sediment deposition, and reduces light penetration. The ice cover is in turn largely controlled by the extreme seasonality of Antarctica and local climate. Lake Hoare and other dry valley lakes may be sensitive indicators of short term (< 100 yr) climatic and/or anthropogenic changes in the dry valleys since the onset of intensive exploration over 30 years ago. The time constants for turnover of the water column and lake ice are 50 and 10 years, respectively. The turnover time for atmospheric gases in the lake is 30–60 years. Therefore, the lake environment responds to changes on a 10–100 year timescale. Because the ice cover has a controlling influence on the lake (*e.g.* light penetration, gas content of water, and sediment deposition), it is probable that small changes in ice ablation, sediment loading on the ice cover, or glacial meltwater (or groundwater) inflow will affect ice cover dynamics and will have a major impact on the lake environment and biota.

Introduction

The largest relatively glacier-free region on the Antarctic continent is the southern Victoria Land 'dry valleys' (~4000 km²) near McMurdo Sound (Heywood, 1972). The dry valleys, which have also been called the southern Victoria Land 'oasis' (Parker *et al.*, 1982a) or 'Ross desert' (Friedmann & Weed, 1987), exhibit glacial and periglacial features, temperatures usually below freezing, low precipitation, cyclonic storms, high velocity winds, and four months each of continuous sunlight, twilight, and darkness (Solopov, 1967). The southern Victoria Land dry valleys contain several closed basins in which perennially ice-covered lakes are found. The ice covers on these lakes overlie liquid water which contains plankton and benthic microbial communities.

The lakes of southern Victoria Land were first discovered during the Expeditions of R. F. Scott during the early 1900's (Scott, 1905; Huxley, 1913). For half a century after their discovery, the limnology, geology, geomorphology, and climate of the southern Victoria Land dry valleys and lakes remained unknown (Parker *et al.*, 1982a). It was not until the International Geophysical Year

(1957–58) and the establishment of United States and New Zealand scientific research bases on Ross Island that studies in the dry valleys were resumed.

From the late 1950's until 1978, limnological data from these lakes were limited to samples and measurements taken through 10–23 cm drill holes in the ice. These studies focused primarily on the characterization of the water column and have been reviewed by Heywood (1984), Hobbie (1984), and Vincent & Howard-Williams (1985). Beginning in 1978, we developed a method for melting holes through the 3–5 m thick perennial ice covers for the purpose of allowing research divers to work under the ice (Simmons *et al.*, 1979; Love *et al.*, 1982). This development and the data generated by these activities opened a new dimension to limnological studies of the dry valley lakes.

Some of the more important results of this research have led to a better understanding of gas exchange mechanisms between the atmosphere, ice, and water column (Wharton *et al.*, 1986, 1987), the toxicity effects of high oxygen concentrations (Mikell *et al.*, 1984), the biological adaptations to low levels of light (Palmisano & Simmons, 1987) and temperature (Seaburg *et al.*, 1982), the relationship between sediment accumulation on the ice cover and resulting ice cover dynamics (McKay *et al.*, 1985; Nedell *et al.*, 1987b; Simmons *et al.*, 1986), and the species composition, distribution, morphology, and ecology of benthic microbial mats (Allnutt *et al.*, 1981; Love *et al.*, 1983; Parker *et al.*, 1981, 1982b; Simmons *et al.*, 1983; Wharton *et al.*, 1982, 1983).

The majority of the studies listed above lead to an important conclusion; namely, environment in a dry valley lake is to a large extent controlled by the presence of a relatively thick perennial ice cover. The ice cover eliminates wind generated currents within a lake (Ragotzkie & Likens, 1964; Hawes, 1983b) and greatly restricts exchange of gases between the water column and atmosphere (Wharton *et al.*, 1986, 1987). The ice cover also greatly reduces light penetration (Palmisano & Simmons, 1987) and restricts sediment deposition (Simmons *et al.*, 1986; Nedell *et al.*, 1987b) into the water column below.

The ice cover is in turn largely controlled by the extreme seasonality of Antarctica and local climate. Therefore, it is possible that the perennially ice-covered lakes could represent a sensitive indicator of short term (< 100 yr) climatic and/or anthropogenic changes in the dry valleys resulting from intensive exploration over the past 30 years. As shown in a subsequent section, the time constants for turnover of the water column and lake ice are ~50 and ~10 years, respectively. The turnover time for atmospheric gases in the lake is 30–60 years (Wharton *et al.*, 1987). Thus, the lake environment responds to changes on a 10–100 year timescale. Because the ice cover has a controlling influence on the lake (*e.g.* light penetration, gas content of water, and sediment deposition, etc.), it is probable that small changes in ablation, sediment loading, and glacial meltwater (or ground water) inflow will affect ice cover dynamics and will have a major impact on the lake environment and biota.

In this paper, we synthesize our recent research findings for Lake Hoare, Taylor Valley, Antarctica. Specifically, we will discuss the climate of Taylor Valley, the physics of Lake Hoare's ice cover, and the gas balance within the lake. We will also consider the effects of the ice cover on the water column and benthic environments, as well as sediment deposition. We will conclude with a discussion of future research objectives, which we believe necessary to complete our understanding of the lake environment and its potential as an indicator of short term climatic and/or anthropogenic changes in the dry valley region.

Study site

Lake Hoare (77°38′ S, 162°53′E) is at the eastern end of Taylor Valley in southern Victoria Land, Antarctica. The lake is 58 m above sea level, 4.1 km long, 1.0 km wide, with a surface of 1.8 km^2, a maximum depth of 34 m, and a mean depth of 14.2 m (Fig. 1). The perennial ice cover of Lake Hoare overlies water at a temperature of

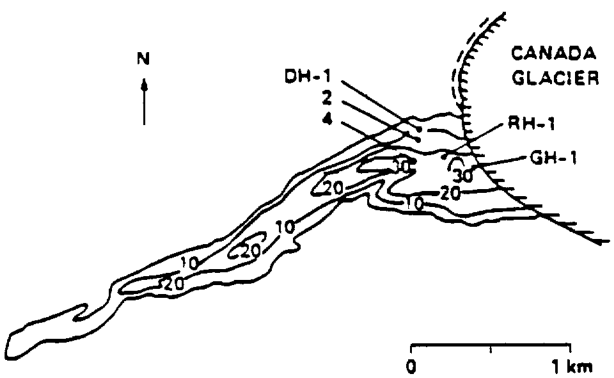

Fig. 1. Map of Lake Hoare, Taylor Valley southern Victoria land Antarctica. DH, GH, RH indicate location of research dive holes. Depth contours are in meters.

0 °C. The lake receives both water and sediment from glacial meltstreams and from nearby Lake Chad during the austral summer; lacking outflow streams, it loses water only by ablation and sublimation at the surface of the ice and by evaporation from the moat.

Results and discussion

Taylor Valley climate

The climate of the dry valleys is strongly seasonal as a result of the southern latitude. There is ~4 months of sunlight in summer and 4 months during which the sun does not cross above the horizon. The light, temperature and wind regimes follow this basic polar cycle (Clow *et al.*, in press).

In order to develop a in-depth understanding of any ecosystem it is necessary to have year-round data as to the climatic conditions. To achieve this objective without the cost or environmental impact of an over-winter team we have deployed automatic sensing and recording systems at Lake Hoare. A meteorological station was deployed in December 1985 and is situated on a small peninsular kame in the northeastern (down valley) end of Lake Hoare (about 100 m up valley from the Canada Glacier meltstream). The station consists of a cup anemometer, wind direction indicator, relative humidity measurement device, and a shielded T-type thermocouple with a electronic reference thermistor. Identical wind and temperature instruments were placed at two elevations, 1 and 3 m above ground level. A Li-Cor PAR quantum sensor, which measures photosynthetically active radiation (PAR, 400–700 nm), was attached to the surface of a flat rock approximately 10 cm above the ground, near the station. The data acquisition and storage system is based on a Campbell Scientific Data Recorder.

Two environmental variables of prime interest in understanding the ecology of Lake Hoare are temperature and light. These variables also are characterized by unique cycles associated with the polar regions. For these reasons, we focus on these variables in this paper, while a more complete description of the meteorology and instrumentation is in Clow *et al.* (in press).

Figure 2 shows the six-hour averages of temperature at 3 m height above the ground surface. Minimum temperatures during the austral winter were slightly less than −40 °C. However, during the winter foehn winds the temperature rises sharply often to as high as −10 °C. Summer temperatures were often above freezing and in mid January 1987 there were days in which the temperature never fell below freezing. The mean annual temperature for 1986 was −17.3 °C, which is very close to the value reported for Lake Vanda of -20.0 °C (Thompson *et al.*, 1971).

Figure 3 shows the six-hour averages of the PAR incident on the lake surface. A diurnal cycle, even on a clear day, is evident in the summer. This is due to the ~26° difference between the zenith angle of the sun at noon and at midnight. As can be seen by comparing Figs. 2 and 3, the air temperature closely follows the incident sunlight. The summer climate is dominated by the presence of sunlight (Clow *et al.*, in press). This is true both on the scale of seasonal variations and on day by day variations in the summer. Hence, the warmest day of the year will typically be in late December or early January, close to the summer solstice. This is quite unlike lower latitudes in which there is a much longer delay between the summer light maximum and the warmest temperatures. During the summer, days with cloud cover are associated

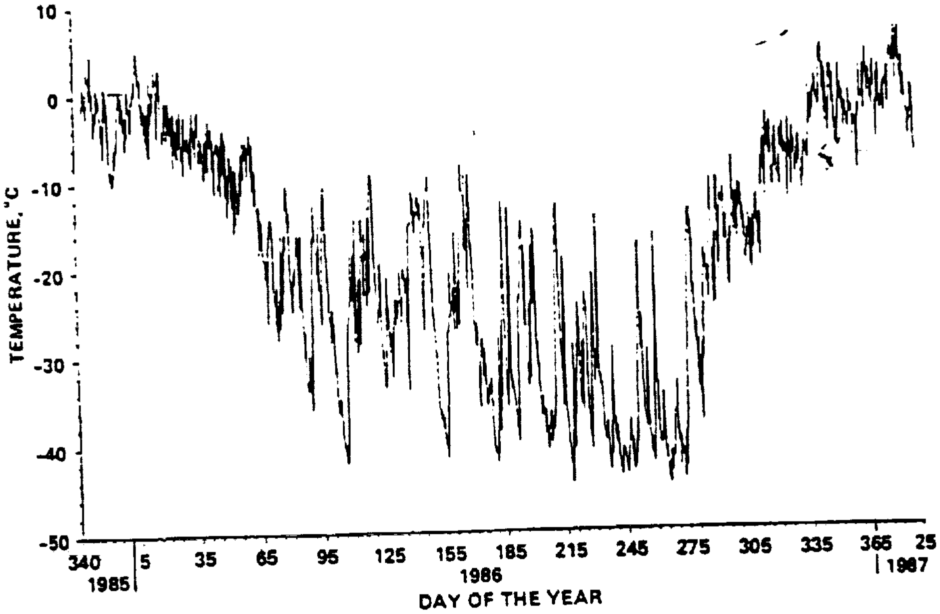

Fig. 2. Temperature at 3 meter height using shielded (against direct sunlight) T-type thermocouple with an electronic reference temperature (thermistor). Each data represents the average of data points taken every 30 seconds over six hour intervals. The data from the meteorology station is plotted versus the day of the year; zero on the x-axis corresponds to Jan. 1, 1986. The data for Dec. 1985 are plotted as well. A more detailed discussion of these results is in Clow *et al.* (in press). Errors are about ± 0.5 °C. The 1986 yearly average value is − 17.3 °C.

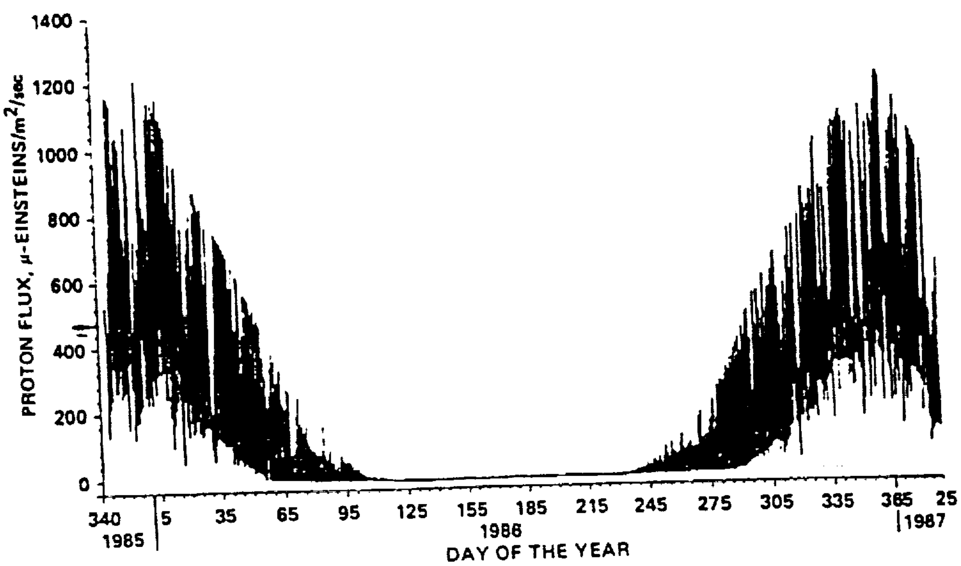

Fig. 3. Photosynthetic quantum flux at Lake Hoare (meteorology station) obtained with a Li-Cor PAR quantum sensor sensitive to light between 400–700 nm. The yearly average PAR is 188 $\mu E\ m^{-2}\ s^{-1}$ with a maximum error of 4.6%. The x-axis is as defined in Fig. 2.

with significant drops in temperature, as can be seen in Fig. 2 for the interval between days 358 and 361 of 1985 (corresponding to 25–28 Dec. 1985). The total amount of photosynthetically active sunlight incident on the lake can be determined by integrating the curve in Figure 3 which gives $5.93 \times 10^9\ \mu E\ m^{-2}$. This corresponds to a yearly average PAR of $188\ \mu E\ m^{-2}\ s^{-1}$. The maximum error on this value based upon the temperature sensitivity of the photodiode (0.15% per degree from 25 °C) is 4.6%.

Figure 4 shows the cumulative PAR as a function of time of year beginning on the first of September. To compile this figure, the data in Fig. 3 were continued by assuming that the latter portion of January and all of February and March 1987 were identical to the corresponding months in 1986. As can be seen in the figure only a small percentage of the total light (~5%) is incident before the start of typical field operations in mid October.

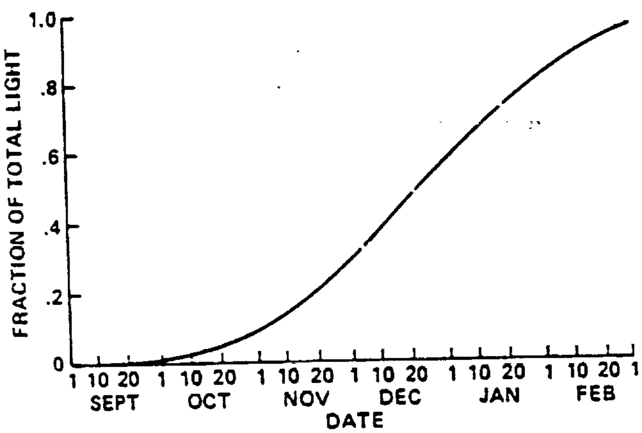

Fig. 4. Cumulative sunlight as a function of time of year beginning on the first day of sunrise. This is the fraction of the total summer sunlight that has reached the ground by the date specified on the x-axis. To compile this figure, the data in Fig. 3 for the summer season 1986–1987 were continued by assuming that the latter portion of January and all of February and March 1987 were identical to the corresponding months in 1986.

Physics of the ice cover

The dry valley lakes are perennially ice-covered because the mean annual temperatures are so low (−20 °C). The presence of liquid water beneath the ice is primarily due to the fact that for a few days in the summer the air temperatures are above freezing. Consequently, the presence of a year-round ice-water interface in a lake, unique to Antarctica, is due to this combination of very cold mean temperatures and comparatively warm summer maximums. A smaller seasonal temperature distribution would result in a lake that is either frozen completely or one that melts fully in the summer.

McKay *et al.* (1985) have developed a simple model which relates the thickness of ice on a perennially ice-covered lake to climatological variables. In this annually averaged, steady state model, the thickness of ice is determined primarily by the balance between the conduction of energy from the ice and the inputs of energy via sunlight and the transport of latent and sensible heat by the summer meltstream. The latent heat released upon freezing at the ice-water interface is the largest term in this equation. Because steady state conditions are assumed, the freezing rate of water at the ice bottom must be offset by ablation from the ice surface. Through this indirect relationship, ablation is the key variable that predicts the ice thickness. Other factors constant, higher ablation rates correspond to thinner steady state ice cover thicknesses. The ice thickness is given by (McKay *et al.*, 1985):

$$Z = \frac{b \ln(T_o/T_s) + c(T_s - T_o) - S_o(1-a)(1-r)h(1-e^{-Z/h})}{v\rho l + F_g}$$

where T_o is the temperature of the ice-water interface, T_s is the yearly averaged temperature of the surface, both in Kelvin, Z is the equilibrium thickness of the ice cover, v is the rate of formation of new ice averaged over the entire year (equal to the ablation rate), ρ is the density of the ice, l is the heat of fusion of water, a is the albedo of the lake, r is the fraction of the lake that is covered by dark absorbing material such as sand and silt, S_0 is the annual average solar radiation incident on the lake surface, h is the mean e-folding extinction path length multiplied by the cosine of the mean solar zenith angle, b and c are constants that define the thermal conductivity of the ice, $k = b/T - c$, taken to be $780\ \mathrm{W\,m^{-1}}$ and $0.615\ \mathrm{W\,m^{-1}\,K^{-1}}$ respectively and F_g is the geothermal heat flux.

We can use the difference in climate between Wright and Taylor Valleys to calculate the difference in steady state ice cover thickness on Lake Vanda and Lake Hoare. The mean annual temperature for 1969–70 at Lake Vanda (Wright Valley) was $-20.0\ ^\circ\mathrm{C}$ (Thompson *et al.*, 1971), while that for Lake Hoare over 1986 was $-17.3\ ^\circ\mathrm{C}$. There is also a significant difference in the yearly average light reaching the surface. A value of $104\ \mathrm{W\,m^{-2}}$ has been reported for Lake Vanda (Thompson *et al.*, 1971). A value for Lake Hoare can be determined by converting the average PAR ($188\ \mu\mathrm{E\,m^{-2}\,s^{-1}}$) into total radiation (Clow *et al.* in press), which gives a value of $92\ \mathrm{W\,m^{-2}}$. Using Eq. (1) and the nominal values of all parameters other than light (the nominal values are those from curve '1' from Fig. 2 of McKay *et al.*, 1985, with a 30 cm/yr ablation), we have computed that the ice thickness corresponding to $104\ \mathrm{W\,m^{-2}}$ is 3.36 m, and the ice thickness corresponding to $92\ \mathrm{W\,m^{-2}}$ is 4.43 m. Hence, we can predict that the ice cover on Lake Hoare would be about a metre thicker than the ice cover on Lake Vanda.

The model of McKay *et al.* (1985) assumes that the ice cover is in steady state, however, there is evidence that the thickness of the ice on Lake Hoare has undergone a significant change over the last ten years. Fig. 5 shows the thickness of the ice on Lake Hoare as determined by drill and melt holes over the past ten years. The 1980–81 point labeled with a solid circle and a set of error bars represents the results of an extensive survey of the ice cover that season. Thirty-five holes were drilled along the length and breath of the lake. The average, with standard deviation, of the ice-thickness values obtained is 4.77 ± 0.34 metres, $N = 35$. It is also significant that 88% of the values were within one standard deviation of the mean and no values were more than two standard deviations from the mean. The distribution is significantly tighter than a gaussian. These statistics imply that the ice cover is very uniform throughout the lake and that single measurements can be used with a fair degree of confidence. In general, any single measurement will be within ± 0.34 m of the mean ice thickness. The error associated with the mean of three measurements will be significantly less. For these reasons, we conclude that the trend shown in Fig. 5, a decrease of ice thickness on Lake Hoare, is not a measurement error but represents a pronounced, and as yet unexplained, change in the lake.

The data show a clear trend; the ice thinned at a roughly uniform rate of about 28 cm/yr from 1977 to 1986. This value is comparable to the ablation rates reported for Lake Fryxell (30 cm/yr, Henderson *et al.*, 1965) and predicted for Lake Hoare from meteorological data (35 cm/yr) by Clow *et al.* (in press). This indicates

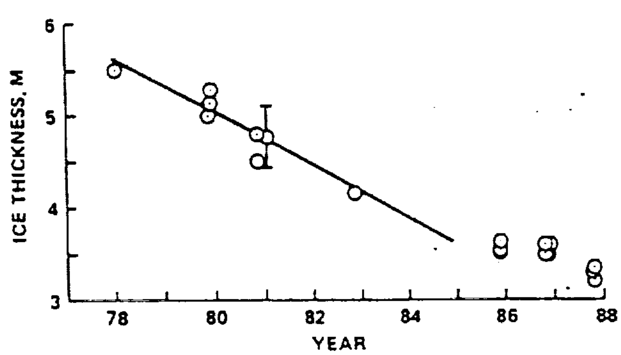

Fig. 5. Measured values of ice thickness on Lake Hoare, Taylor Valley, Antarctica over the last ten years as a function of the date. The data point plotted on 20 Jan 1981 represents the mean of 35 data points. The range indicated is ± one standard deviation. The linear fit to the data from 1977–1986 has a slope of −28.3 cm/year.

that there has been either a change in the local climate in Taylor Valley (*e.g.* mean annual temperature) or a change in the physical properties of the ice cover on Lake Hoare (*e.g.* sediment loading). To model these changes in the ice cover requires extending the results of McKay *et al.* (1985) to include time dependent effects such as seasonal changes in sunlight and ablation.

One of the significant biological implications of the ice cover is on the quantity and spectral distribution of radiation reaching the liquid phase beneath the cover. The reduction in light transmission due to the thick ice cover also affects the heat budget (Hoare, 1966; Ragotzkie & Likens, 1964; Bydder & Holdsworth, 1977; Adams & Lasenby, 1978); primary production, and plankton distribution and composition (Rodhe, 1956; Tominaga, 1977; Rigler, 1978; Priddle, 1980; Light *et al.*, 1981; Vincent, 1981; Vincent & Vincent, 1982; Cathey *et al.*, 1982; Parker *et al.*, 1982a; Hawes, 1983a,b, 1985). From an ecological perspective, this reduction in the quantity and alteration of the spectral distribution of light will exert selective pressures on the photoautotrophs (Seaburg *et al.*, 1983; Priscu *et al.*, 1987).

Palmisano & Simmons (1987) recently discussed the spectral downwelling PAR irradiance (400–700 nm) in Lake Hoare. For measurements taken near noon at summer solstice, the full waveband PAR beneath the ice was $< 44\ \mu E\ m^{-2} s^{-1}$ or $\leqslant 3\%$ of surface downwelling irradiance. The ice cover absorbed longer wavelengths and maximum light transmission was in the blue region between 400–550 nm. The bulk attenuation coefficient of the water column ranged between 0.45 and 1.33 for five depths measured beneath the ice at two diver holes. Light attenuation by phytoplankton was greatest in the 400–550 nm and 656–671 nm regions. The spectral distribution of sunlight penetrating the lake ice depends upon seasonal factors including solar zenith angle, day length, cloud cover, and ice characteristics.

Goldman *et al.* (1967) first reported data on the seasonal variations in ice cover optical properties. To further document these seasonal effects, we have extended the transmission measurements of Palmisano & Simmons (1987). In Fig. 6, we show the ice cover transmission from early in the season (11 Nov. 1986), at summer solstice (23 Dec. 1982, from Palmisano & Simmons, 1987), and late in the season (11 Jan 1987). The ice cover thickness was roughly comparable in the 1982 data to the 1986–87 data (see Fig. 5). The results show an interesting pattern. Early in the season when the ice is fairly clear, the transmission of blue light greatly exceeds that of red light. This is because the transmission of blue light is dominated by scattering, which is minimal in the clear ice. However, in the red, transmission is dominated by the absorptive properties of ice which is fairly independent of whether the ice is clear or not. As the season progresses, the surface of the ice clouds due to freeze-thaw processes on the surface, and as a result, the scattering optical depth greatly increases, attenuating the blue light (Fig. 6). This scattering has little effect on the red light. The result is that later in the season the spectral

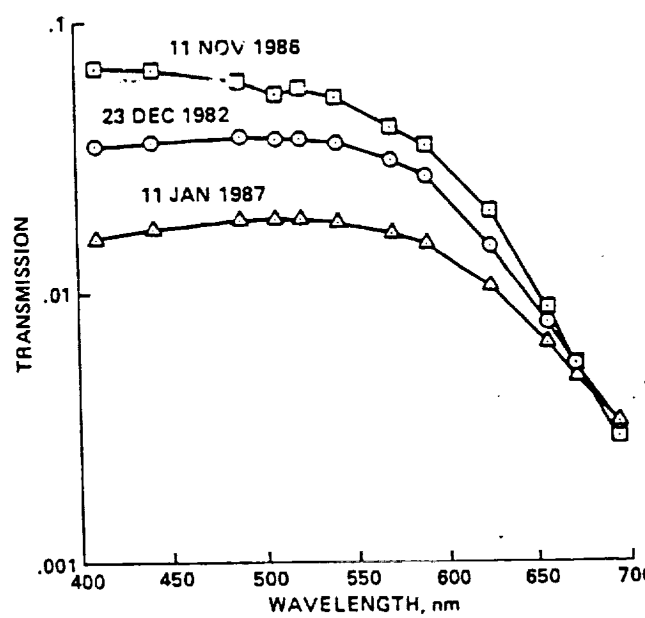

Fig. 6. Light transmission through a 3.7 m thick ice cover at various times during the season. Uppermost curve is for early in the season (11 Nov. 1986), middle curve is at summer solstice (23 Dec. 1982) from Palmisano & Simmons (1987) and lower curve is for late the season (11 Jan. 1987). The ice cover thickness was similar in the 1982 data to the 1986–87 data (see Fig. 5).

properties of the ice become more uniform with wavelength.

Roulet & Adams (1984) found spatial heterogeneity of light penetration in an ice-covered Canadian lake. They emphasized that a single point measurement could overestimate the quantity of light by 700% compared to an integrated areal measurement. During the 1986 austral summer, we measured the spatial heterogeneity of light penetration over a 10 m diameter circle immediately beneath the ice at DH4 (Fig. 1) and found the total quanta penetrating the ice surface to range between 15–51 $\mu E\, m^{-2}\, s^{-1}$ ($\bar{x} = 28.5 \pm 12.7$, N = 6) and the percent transmission to range between 0.43–1.45 ($\bar{x} = 0.81 \pm 0.36$, N = 6). The variation between our highest and lowest reading was 58.8% and most of the readings (5) were $\leqslant \pm 30\, \mu E\, m^{-2}\, s^{-1}$. This indicates that the lake is relatively uniform with respect to optical properties, which is in agreement with ice thickness measurements.

Commonly encountered conditions indicate that light penetration is ~1% of surface irradiance values. The phytoplankton and water mass then combine to absorb much of the remaining light, so that the quantity of light reaching the bottom of the lake is $\ll 1\%$. This aspect of light penetration and attenuation is particularly interesting given the abundance of benthic photoautotrophs that thrive on the lake's bottom (Wharton *et al.*, 1983).

Gas balance in the lake

One of the most unusual features of these lakes is the occurrence of supersaturated O_2 and N_2 in the water column ranging from slightly over saturation to over 400% for O_2 and 160% for N_2. To quantitatively explain the high O_2 concentrations, we developed a bulk O_2 budget. There are two primary net sources of O_2: a physical source resulting from gases carried into the lake by the meltstreams and forced into the water column when water freezes onto the bottom of the ice cover, and a biological source resulting from photosynthesis and from the burial of reduced carbon in the lake sediments.

While it has been known for many years that the dry valley lakes are supersaturated with O_2, quantitative budgets were not available. We have recently developed quantitative models of both oxygen and nitrogen in Lake Hoare (Wharton *et al.*, 1986, 1987). In our model of gas flow into and out of the lake, we considered both biological and nonbiological sources and sinks. Our model (Wharton *et al.*, 1986) predicted, and direct measurements have subsequently shown that there is a supersaturation of N_2 in the lake water, as well as oxygen. Dissolved N_2 levels of 145% and 163% were determined from samples taken just below the ice cover and at a depth of 12 m, respectively (Wharton *et al.*, 1987).

The two principal atmospheric gases (N_2, O_2) are both influenced by the non-biological processes affecting the gas balance of the lake water (such as freezing of water and bubble formation). Non-biological processes will act on N_2 and O_2 equally, maintaining the N_2/O_2 ratio at that value characteristic of water in equilibrium with the atmosphere (~1.8). However, biological processes (primarily photosynthesis and respiration) will affect the O_2 concentrations to a much larger extent than they will affect N_2, thus altering the N_2/O_2 ratio (Wharton *et al.*, 1987). Therefore, this ratio can be a useful 'signature' of biological and non-biological gas production. In Lake Hoare, this ratio was 1.20 at the ice/water interface and 1.05 at 12m; considerably different from the ratio in equilibrium with air (~1.8). Based on these results, we have determined that about half of the net O_2 production in the lake is the result of biological processes.

The approach discussed above by which we can infer net biological productivity was validated by the independent publication of a study of the open ocean by Craig & Hayward (1987), in which ratios of supersaturated gases were used in a similar way.

By considering the total reservoir of gas in the lake and the inflows and outflows of O_2 and N_2, we have estimated the residence time of these gases in Lake Hoare. We obtain a value of ~30 years for oxygen and ~60 years for nitrogen (Wharton *et al.*, 1987). Thus, gases are cycled relatively rapidly through the lake.

The importance of the gas concentrating mechanisms found in perennially ice-covered Antarctic lakes includes the development of lift-off benthic mat in shallow waters, the natural selection for organisms that can adapt to perennial supersaturated oxygen levels (Mikell *et al.*, 1984), and the production of gas channels in the ice, which may serve conduits for sediment to penetrate the ice cover.

Lake biology

Because of the combination of snow cover, sediment and gas bubbles, light transmission through the ice cover is much less than would be expected with an equivalent column of pure ice or water. This reduction in light limits the plankton density, and the lack of internal currents that keep free floaters suspended (*e.g.* Langmuir spirals) also restricts the plankton population to mainly swimming forms (Parker *et al.*, 1982a). Over 200 measurements of chlorophyll *a* (Chl *a*) were made during the 1985–86 austral summer at four different locations (DH1, DH2, DH4, RH1) in Lake Hoare (Fig. 1). Chl *a* was measured fluorometrically using the technique described in Parker *et al.* (1982a). The integrated Chl *a* concentration for the season was 16.92 mg m^{-2}. However, if a hypsographic curve is used to accommodate changes in volume with depth, the value then becomes 22.26 mg m^{-2} at the lake surface. When these data are compared against 5 integrated profile measurements made during the 1979–80 season, the 1985–86 values decrease by 43.2% (22.6 vs. 39.19). The reasons for the decrease measured during the 1985–86 season may have been due to the melting of ice which would have diluted the concentrations of Chl *a* in the upper water column. Also, during the 1979–80 season the first Chl *a* collections were made nearly three weeks before we began the 1985–86 collections. In fact, the highest Chl *a* measurement of the 1979–80 season was obtained on the first profile measurement. When the 1985–86 data were examined on a seasonal basis, they showed that the maximum Chl *a* level was at 13 m piezometric depth (maximum O_2 concentration was at 10 m).

The fate of the phytoplankton community during the Antarctic winter remains to be determined. The future of the phytoplankton community under a dynamically changing ice cover is important to our understanding of this lake ecosystem. If the ice cover continues to thin and sand continues to be dumped from the ice cover, we would expect light transmission to increase, and therefore phytoplankton density, as measured by Chl *a*, to also increase.

Microbial mats composed primarily of the cyanobacteria, eukaryotic algae, and heterotrophic bacteria occur abundantly throughout the benthic region of Lake Hoare (Wharton *et al.*, 1983). These microbial mats are precipitating calcite, iron, and sulfur, and trapping and binding sediment forming alternating laminae of organic and inorganic material. An unique feature of many of these benthic mats is their development into modern, cold water stromatolites (defined by Awramik *et al.*, 1976 as organosedimentary structures produced by sediment trapping, binding, and/or precipitation as a result of the growth and metabolism of microorganisms). Wharton *et al.* (1983) have described microbial mats resulting in four types of modern Antarctic stromatolites, including lift-off, pinnacle, aerobic and anaerobic prostrate mats.

One interesting effect of the elevated gas levels is on the formation of stromatolites. Beneath the perennial ice in the shallower, more brightly lit areas of the lake is a sizeable biomass of columnar lift-off mat. In these mats, an excess of dissolved gas (primarily N_2 and O_2) accumulates as bubbles causing the mats to lift off the substrate (Parker *et al.*, 1982b; Wharton *et al.*, 1983). These pieces of mat often tear loose from the substrate and float up to the underside of the ice, thus disturbing the integrity, or prohibiting the development of a laminated structure. Therefore the formation of bubbles, which is controlled by the environment, can directly influence mat morphology and stromatolite formation.

Wharton *et al.* (1976) have developed a formula for relating O_2 and N_2 dissolved in the water column to bubble formation with depth:

$$z = 10.3[0.21(S_o - 1) + 0.78(S_n - 1) + p_w] \quad (2)$$

where z is the maximum depth of bubble formation, S_o is the supersaturation of O_2 with respect to equilibrium with the atmosphere, S_n is the corresponding number for N_2, and p_w is the vapor pressure of water. For example, in Lake Hoare the high concentration of dissolved gases in the water column results in bubble formation to a depth of ~10 m. Below 10 m bubbles cannot form in the water as a result of the increased hydrostatic pressure. Consequently, we expect bubble formation and mat lift-off to disrupt mat morphology at depths above 10 m, while at depths below 10 m laminated, prostrate mats would be expected. Indeed, the former and latter cases are observed *in situ* (Wharton *et al.*, 1983).

Another interesting aspect of the elevated gas levels in the water column is on the oxygen content of the sediment. During the 1986–87 field austral summer, oxygen profiles of the sediment were obtained with a Diamond Electro-Tech dissolved-O_2 microprobe calibrated to microwinkler determinations of water from the sediment-water interface. This method was used to determine the oxygen level of sediment in cores taken from different locations at the bottom of the lake. Contrary to what is typically observed in lake sediment, sediment collected in 10 m water depth and above remained oxic to a depth of 30 cm (maximum core depth). Sediment collected from 23 m depth and below were anoxic several mm below the sediment water interface. The sediment which remained oxic to 30 cm are overlaid with water supersaturated with O_2 (and other gases) while the anoxic sediment is overlaid by anoxic water or water with saturated levels of O_2 similar to what would be observed in a temperate lake. These results are another indication of the important role that the supersaturation of atmospheric gases play in regulating the lake environment.

Sediment/ice interactions and sedimentation

The physical environment for sedimentation in Lake Hoare is unusual because of the presence of the perennial ice cover. The ice catches and traps wind blown sediment and provides a surface for the movement (by saltation, rolling and drift on the ice) of larger sediment particles into the middle of the lake. Although the ice cover does contain large boulders right in the middle of the lake, most of the mass of the ice cover burden is in the form of sand-sized and finer particles. Based on numerous melt holes we estimate that the average sediment loading is ~0.2–2 g cm^{-2}. This range of values corresponds to less than 0.6% of the mass of a 4 m ice cover.

The sediment in the ice will have a major influence on the optical properties of the ice cover. Sediment opacity causes heating of the ice cover, which would tend to thin the ice. At the same time, sediment opacity also prevents the transmission of light through the ice, which would tend to thicken the ice (McKay *et al.*, 1985). Consequently, the amount of sediment in the ice cover is an important variable in regulating ice thickness.

Small particles of sediment will not melt their way directly through the ice cover (Simmons *et al.*, 1986). This can be shown theoretically by considering the energy balance of a radiatively heated particle. A small, dark sand-sized particle on the surface of the ice or embedded in the ice cover absorbs sunlight. If the heating rate is sufficient to raise the surface temperature of the particle above the melting point, the particle will sink into the ice cover. Because the particles are very small compared to the thickness of the ice cover, the particle surface temperature can be determined by the spherically symmetric heat equation:

$$F(1 - \omega)\pi r^2 = 4\pi r k \Delta T \quad (3)$$

where F is the radiation field in the ice cover averaged over the upward and downward directions (including scattered light), ω is the single scattering albedo of the particle ($\omega \simeq 0.2$), r is the radius of the particle, k is the thermal conductivity of the ice (at −1 °C, k ~2.3 W K^{-1} m^{-1}), and ΔT is the difference between the temperature of the particle surface and the temperature of the ice at the depth of the particle. Using this equation, Simmons *et al.* (1986) have shown that in order to melt through ice that is only 1° below freezing requires a particle of 1.5 cm radius at the surface, 3.8 cm at a depth of 1 m, and 9.3 cm at a depth of 2 m. Melting through colder ice requires even larger particles. Hence, sand-sized particles that have radii much less than 1 cm, will not melt through the ice cover and are carried into the ice cover by surface meltwater percolation during the austral summer. Aggregates of individual particles are effectively like a particle with a larger size and are therefore able to more effectively melt the ice, but only in approximately the first meter. From this we conclude that sediment will not melt through a several meter-thick ice cover and must be carried by water through cracks or gas-bubble channels in the ice cover.

The results (Table 1) from four sets of sediment traps placed in the northeastern end of the lake have helped us unravel the unusual sedimentation processes (Simmons *et al.*, 1986; Nedell *et al.*, in prep.). Traps from DH1 (~8 m depth) showed a sedimentation rate of 4.11 mg cm^{-2} yr^{-1}. Sediment traps from DH2 (~11 m depth) and DH4 (~27 m depth) averaged 3.76 and 2.87 mg cm^{-2} yr^{-1}, respectively. It is interesting that one trap (DH2 A and DH4 C) from each of these latter two sites contained significantly more sand-sized sediment than the other traps from the same site. Traps from GH1 (~23 m depth) contained a substantial quantity of sediment and had a mean sedimentation rate of 142 mg cm^{-2} yr^{-1}. The sediment traps at GH1 were predominantly coarse sand, while farther away from the glacier at DH4, both coarse sand and finer, silty material were collected. In DH1 and DH2, which are closer to shore, the traps collected silt and clay-sized particles. The observation of different amounts of sediment from the same area supports the hypothesis that sediment enters the lake through the ice cover at distinct locations via cracks in the ice and/or gas bubble channels. Also, small mounds of sand 0.5–1.0 m high and 1.0–3.0 m wide were observed at the sediment/water surface near DH2 and GH1, further suggesting point sources for sediment discharge into the lake through the ice cover (Nedell *et al.*, in prep.). Therefore, instead of

Table 1. Total dry mass and composition of sediment trap material collected from dive holes (DH) 1, 2, 4 and glacier dive hole (GH) 1 in Lake Hoare, southern Victorian Land, Antarctica (modified from Simmons *et al.*, 1986).[1]

	Total dry mass (g)	Organic matter (g)	Carbonate (g)	Gravel (g)	Sand (g)	Mud (g)	Other[2] (g)
DH 1-A	17.52	0.81	6.61	0.00	0.25	9.85	0
DH 1-B	24.38	0.63	13.79	0.00	0.43	9.53	0
DH 1-C	16.95	1.10	6.04	0.00	0.34	9.47	0
DH 2-A	44.25	3.11	2.46	3.88	32.65	2.15	0
DH 2-B	4.58	0.21	1.52	0.00	0.00	2.85	0
DH 2-C	4.98	0.48	1.41	0.64	0.10	2.35	0
DH 4-A	2.58	0.10	0.22	0.00	2.13	0.13	0
DH 4-B	2.08	0.13	0.18	0.00	1.67	0.10	0
DH 4-C	36.35	0.32	0.89	2.58	32.39	0.17	0
GH 1-A	633.00	39.69	18.38	0.00	572.42	1.51	1.00
GH 1-B	544.00	4.30	15.30	0.00	522.74	2.24	0.02
GH 1-C	856.20	7.53	30.74	0.00	813.97	2.94	1.02

[1] Traps were deployed in Dec/Jan 1982 and retrieved in Nov/Dec 1985; methods for determining composition of trap material in Simmons *et al.* (1986) and Nedell *et al.* (in prep). Traps were composed of an aluminium funnel with a 45 cm diameter opening, 47 cm tall, attached to a plastic container 10 cm in diameter and 20 cm tall.

[2] Other = aluminium oxide material.

receiving most of its sediment from the lake margin and inflowing streams, the majority of sediment at the bottom of Lake Hoare is transported downward through the ice cover. The similarity in grain-size distribution and mineralogy between samples from the ice cover and lake bottom supports this conclusion (Simmons *et al.*, 1986; Nedell *et al.*, 1987b). However, the fine-grained sediment population that appears in the lake bottom samples but not in the samples from the ice cover, is probably brought in via meltstreams. This is corroborated by the observation of an under ice plume of glacial flour 250–300 m from the entrance of the Canada Glacier meltstream into the lake.

It is not clear what controls the total sediment burden on the ice cover. We suggest at least two general possibilities. One possibility is that the ice-sediment system is in steady state; the rate of sediment percolation through the ice cover is sufficient to balance the rate at which sediment is added to the ice cover. In this case, the current sediment burden is the steady state value and therefore its low amount suggests that there must be a fairly efficient mechanism for getting sediment through the ice cover. A second possibility is that the ice-sediment system is cyclic. Sediment builds up on an initially clean ice cover until the level of sediment causes a change in the ice cover sufficient to dump the sediment. Clearly, an extreme case would be when sufficient sediment is loaded on the ice cover which then becomes negatively buoyant. Simmons *et al.* (1986) have discussed a possible ice-sediment interaction cycle. In their cycle,increased sediment loading results in increased surface topography on the ice cover and increased ablation. The increased ablation results in a thinner ice cover until eventually a point is reached at which the sediment, which accumulates at the topographic lows (ponds), can pass through the ice cover. The clean ice cover reseals itself and the cycle begins anew. The changes in ice cover thickness shown in Fig. 5 could be evidence that the ice-sediment system is cycling and the ice cover of Lake Hoare just underwent a transition as part of such a cycle (Simmons *et al.*, 1986).

Another important aspect of the benthic sediment is their involvement with possible groundwater influx. We have obtained the first actual measurements of groundwater flow into a dry valley lake. Seepage meters were placed in two locations in the lake for a pilot study to measure ground-water flow. One location was in approximately 8 m of water beneath the ice near the shore by the meteorological station, and the other was in approximately 27 m of water at the base of the Canada glacier near GH1 (Fig. 1). The seepage meters were constructed of polyvinyl chloride (PVG) and 1 L Nalgene collection bags were used for the collection of the seepage water (see Simmons & Netherton, 1987). Seepage meter water was collected from two areas in the lake during the 1985–86 and 1986–87 austral summers over a six day period. During 1985–86, $\sim 625\ \mathrm{ml\ m^{-2}\ day^{-1}}$ was collected from the site at the glacier's base, and $\sim 416\ \mathrm{ml\ m^{-2}\ day^{-1}}$ from the shallow water site. During the 1986–87 austral summer, $\sim 31\ \mathrm{ml\ m^{-2}\ day^{-1}}$ was collected at both sites also over a six day period. The difference between years may be due to the time period during which the samples were collected. In 1985–86, we collected the samples in early December; whereas, in 1986–87, the samples were collected in mid-January. The importance of this pilot study is that earlier suggestions (Wilson, 1979; Chinn, 1982) of groundwater movement into lakes have been corroborated and deserve further study because groundwater seepage into these lakes could play an important role in nutrient cycling to the benthic community at the sediment water interface.

Conclusions

The environment of the Antarctic lakes could represent a sensitive indicator of short term (< 100 yr) climatic and human-induced changes in the dry valleys. In fact, they may be the only sensitive indicator for changes on these times scales. The other class of physical systems that are in dynamic equilibrium with the climate are the glaciers. Chinn (1985) has discussed this aspect of the structure and equilibrium of the dry valley glaciers and concludes that they respond on many thousand year timescales. Some dry valley glaciers are retreating and some are advancing (Chinn, 1985). The other biological system in the dry valleys is the cryptoendolithic microbial communities described by Friedmann (1982). These communities grow very slowly and again the timescale of response to changes is measured in thousands of years (Friedmann & Weed, 1987).

The time constants in the lake are much shorter. We can estimate the turnover time for the water by dividing the mean depth of Lake Hoare (14.2 m) by the inflow rate (30 cm/yr, assumed to equal the ablation rate). This gives a timescale of ~50 yr. A similar calculation for a 4 m ice cover gives a timescale for lake ice turnover of ~13 years. We have determined the turnover time for atmospheric gases in the lake (O_2 and N_2) by considering the total amount in the lake divided by the sources. This gives a value of ~30 years for O_2 and ~60 years for N_2 (Wharton *et al.*, 1986, 1987). A time constant for biological turnover is more difficult to estimate but is probably also between 10 and 100 years.

We now think that the ice cover has a controlling influence on the lake (*e.g.* light penetration, gas content of water, and sediment deposition). Therefore, small changes in ablation, sand loading, and glacial meltwater (or groundwater) inflow will affect ice cover dynamics and will have a major impact on the lake environment and the biota. These records of changes should be most obvious in the sedimentary record. For example, we know that sediment is deposited through the ice cover, perhaps via cracks which develop when the ice cover thins to ca. 3 m thickness. This sediment buries portions of benthic microbial mat, which will ultimately recolonize the sediment surface after a period of years. The resulting sediment lens in the benthic profile indicates a period of sediment deposition through the ice cover. Thinning of the ice cover (and resultant sediment deposition) probably results from changes in local climate which result in ice cover changes. Consequently, it may be feasible to understand past climate regimes as well as to be able to predict the effect of future modifications of climate (either natural or human-induced changes) on the dry valley lake ecosystem.

There is evidence of a recent change in climatic conditions in the dry valleys. Chinn (1982) has documented a water level rise in virtually all of the dry valley lakes over the period of 1972 to 1982. He suggests that this is related to climate but no causal mechanism is proposed. Our data for ice cover thickness on Lake Hoare also suggests changes in the lake environment, and again we have no causal mechanism to explain it. There is not sufficient data to show whether ice cover thicknesses have changed on the other lakes also.

If the dry valley lakes are to be useful as indicators of climate change in that region of Antarctica, then it is necessary to develop the baseline data that documents the present condition of the lakes. Furthermore, detailed predictive models that relate the biogeochemical cycles and biological processes in the lake to the external environment must be developed.

We feel the following future research objectives are necessary to complete our understanding of the major physical, chemical, and biological interactions that regulate the lake's ecology:

1. More refined models of the climatological controls on the thickness of the ice cover.
2. A re-examination of phytoplankton density changes in relation to changes in lake ice thickness.
3. Development of a carbon cycle model, in particular, quantification of the carbon dioxide sources and sinks and their relationship to carbonate formation particularly in benthic microbial mats.
4. Quantification of sediment loading on the ice cover and dumping.
5. Quantification of benthic microbial biomass accumulation and decomposition rates.
6. Resolution of abiotic/biotic contributions to sediment composition.
7. A sediment profile study of oxygen and oxidation-reduction potentials.
8. Quantification of ground water nutrient and mineral re-cycling processes.

In addition to helping us unravel the climatic and environmental history of the dry valley region, the ice-covered lakes serve as useful models for increasing our understanding of early life on this planet and possibly Mars. The first 2.5 billion years of life on Earth was microbial. These microbes left behind a fossil record of their Precambrian existence in the form of stromatolites. It is a common misconception that stromatolites form *only* in warm and/or saline environments. As discussed previously, microbial mats are forming stromatolites in the Antarctic dry valley lakes. Other studies have shown that several periods of glaciation occurred during the Precambrian (Anderson, 1983; Walter & Bauld, 1983). We suggest that studies of Antarctic lakes may play an important role in the re-interpretation of stromatolite formation during the Precambrian and specifically to their occurrence in Precambrian polar environments.

Another intriguing aspect of research in the dry valley lakes is the connection to extraterrestrial habitats. The Antarctic dry valley lakes have been suggested as analogs of paleolakes on Mars which may have sheltered early Martian life. Geological and climatological studies suggest that conditions on early Mars (> 3 b.y.a.) were very different from what they are today, and were similar to early earth (McKay, 1986). Because life on earth is known to have originated during this early period on earth, the Martian environment could have also been conducive to the origin of life. The record of the origin and early evolution of life on Earth has been obscured by extensive erosional and tectonic activity. However, on Mars much of the ancient heavily cratered terrain, dating back to this early period, remains in pristine condition and may hold a record of events that led up to the origin and early evolution of life.

Recent studies by Nedell *et al.* (1987a), have described an area of ancient (> 3 b.y.a.) lake sediments in the Valles Marineris canyon system on Mars. As the Martian atmosphere thinned and the surface grew cold, these putative Martian paleolakes, like the Antarctic lakes, would have contained liquid water beneath a layer of ice, as opposed to being frozen solid (McKay *et al.*, 1985). In addition to providing a relatively warm, liquid water environment, the process of concentrating atmospheric gases beneath the ice cover could have significantly affected the gas budget in those lakes, possibly enhancing the levels of biologically important gases from the thin Martian atmosphere (Wharton *et al.*, 1987). Also, it is possible that the sediments observed in the Valles Marineris canyons could have been deposited by sediment passing through ice in much the same fashion as observed in Lake Hoare, Antarctica (Nedell *et al.*, 1987a,b; Simmons *et al.*, 1986).

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Lipophilic pigments from the benthos of a perennially ice-covered Antarctic Lake

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Abstract

The benthos of a perennially ice-covered Antarctic lake, Lake Hoare, contained three distinct 'signatures' of lipophilic pigments. Cyanobacterial mats found in the moat at the periphery of the lake were dominated by the carotenoid myxoxanthophyll; carotenoids: chlorophyll *a* ratios in this high light environment ranged from 3 to 6.8. Chlorophyll *c* and fucoxanthin, pigments typical of golden-brown algae, were found at 10 to 20 m depths where the benthos is aerobic. Anaerobic benthic sediments at 20 to 30 m depths were characterized by a third pigment signature dominated by a carotenoid, tentatively identified as alloxanthin from planktonic cryptomonads, and by phaeophytin *b* from senescent green algae. Pigments were not found associated with alternating organic and sediment layers. As microzooplankton grazers are absent from this closed system and transformation rates are reduced at low temperatures, the benthos beneath the lake ice appears to contain a record of past phytoplankton blooms undergoing decay.

Introduction

Lake Hoare (77° 38′ S, 162° 53′ E) is one of seven perennially ice-covered oasis lakes in Southern Victorialand, Antarctica, which have received considerable attention in recent years (Canfield & Green, 1985; McKay *et al.*, 1985; Parker & Wharton, 1985; Wharton *et al.*, 1986, 1987, in press; Priscu *et al.*, 1987). This freshwater lake is 58 m above sea level with a surface area of 1.8 km, a maximum depth of 34 m, and a mean depth of 14.2 m. The perennial ice cover of Lake Hoare, which has thinned from 5.5 to 3.5 m between 1978 and 1987 (Wharton *et al.*, in press), overlies water at a temperature of 0 °C. The lake receives both water and sediment from glacial meltstreams and from nearby Lake Chad during the austral summer. Lacking outflow streams, it loses water only by ablation at the surface of the ice and by evaporation from the surrounding moat.

During the austral summer, photosynthetically available radiation (PAR; 400–700 nm) between 1100-1500 h on December 23 ranged from about 44 $\mu E \cdot m^{-2} \cdot s^{-1}$ just beneath the ice cover to 0.06 $\mu E \cdot m^{-2} \cdot s^{-1}$ below the chlorophyll *a* maximum 12 m depth (Palmisano & Simmons, 1987). A phytoplankton bloom of small flagellates belonging to the Cryptophyceae, Chrysophyceae, and Chlorophyceae occurs during the austral summer (Parker *et al.*, 1982; Seaburg *et al.*, 1983). In addition, four types of benthic communities have been described in Lake Hoare which include cyanobacteria, diatoms, and green algae: 1. A surface community in the ice-free moat surrounding the lake; 2. columnar lift-off mat below the ice in shallow, brightly lit areas; 3. aerobic prostrate mats in deeper, dimly lit areas; and 4. anaerobic prostrate mats below 26 m. The alternating bands of organic and sediment layers (AOSL) observed in all cores of aerobic and anaerobic prostrate mat by Wharton *et al.* (1983) suggested that these were modern stromatolitic algal-bacterial mats.

Pigment analysis of marine and freshwater sediments has proven useful for identifying source organisms, that is, organisms contributing organic matter to sediments (Repeta & Gagosian, 1982; Edmunds & Eglinton, 1984). Carotenoids, in particular, are promising as chemotaxonomic indicators, because many are specific to one algal taxon such as myxoxanthophyll from cyanobacteria (Liaaen-Jensen, 1979). Our objective was to examine both the benthos at the sediment-water interface and the AOSL for lipophilic pigments which may serve as biomarkers for microalgae in Lake Hoare.

Methods

Sediment cores were obtained from six locations in Lake Hoare (Fig. 1) during the austral summers of 1985–86 (moat, DH1, DH2, RH1, GH1 and DH4) and 1986–87 (moat, DH1, and DH2) by SCUBA divers using 10 cm diameter Plexiglas tubes. The cores were stoppered underwater, brought to the surface where they were protected from direct sunlight, and returned to a field laboratory for processing. In the laboratory, water above the sediment material was siphoned off, and the cores were extruded on aluminum foil. Discrete layers were sectioned from the cores using a clean blade and forceps. These layers were placed in clean glass vials, covered with black tape, and returned frozen to NASA-Ames Research Center. In addition, samples of surface mat were also collected from the ice-free moat areas at the periphery of the lake during 1985–86 and 1986–87.

Lyophilized mats were homogenized in 0 °C 90% acetone, extracted overnight, decanted, then re-extracted for 1–2 h in fresh solvent. Pigments

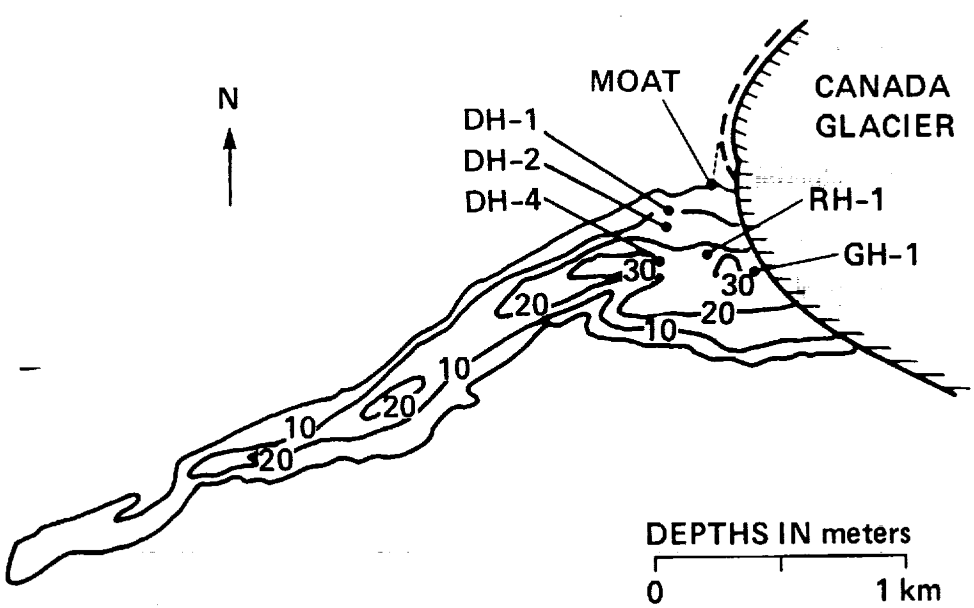

Fig. 1. Bathymetric map of Lake Hoare showing the location of six sites where benthic samples were collected.

were analyzed by high performance liquid chromatography (HPLC) on a reverse phase ODS Hypersil column. The details of these analyses will be described elsewhere (Palmisano *et al.*, 1988). Briefly, a non-linear gradient of 82% methanol, 13% acetonitrile, and 5% deionized water was run against 100% acetone; deionized water contained tetrabutyl ammonium acetate as an ion pairing agent (Mantoura & Llewellyn, 1983). Pigments were identified by absorption maxima using ultraviolet-visible spectroscopy (Foppen, 1974) and co-chromatography with standards. β-carotene and chlorophylls *a* and *b* were obtained from Sigma Chemical Co.; phaeophytins *a* and *b* were made by acidification of corresponding chlorophylls. Lutein, zeaxanthin, fucoxanthin, canthaxanthin, and echinenone were generously donated by Hoffmann-LaRoche Co. Myxoxanthophyll and chlorophyll *c* were isolated from cyanobacteria and diatoms, respectively, using thin layer chromatography (Jeffrey, 1981). Samples were preserved for microscopic examination in Lugol's iodine or 2.5% glutaraldehyde.

Results

The dominant microflora in the benthic samples based on light microscopy and the depths at which they were found are summarized in Table 1. The benthos was dominated by pigments from either cyanobacteria, diatoms, cryptomonads or green algae. Microscopic examination of freshly collected as well as preserved samples revealed that detrital material was abundant in the under-ice sites; microalgae from these sites had shrunken chromatophores and appeared to be senescent.

Table 1. Dominant photosynthetic microbiota in the benthos of Lake Hoare, Antarctica.

Site	Date	Depth	Microflora
Moat	11/85	0.3 m	cyanobacteria
	1/87		
DH2	11/85	10.5–11.5 m	diatoms, frustules
	1/87	10.5–11.5 m	golden-brown monads
RH1	11/85	18–20 m	diatoms
GH1	11/85	21–26 m	coccoid green algae
DH4	11/85	28–30 m	coccoid green algae

Three different lipophilic pigment 'signatures' were evident in the chromatograms of lake sediments (Fig. 2). Altogether, a total of 23 different pigments were isolated of which 17 have been tentatively identified.

A pigment signature characteristic of cyanobacteria was found in the surface 0.5 cm of the moat mats (Fig. 2; Table 2); these mats were primarily composed of *Nostoc* sp., *Oscillatoria* sp., and *Phormidium* sp. Myxoxanthophyll, a carotenoid specific to cyanobacteria, was the dominant pigment in both the 1985–86 and 1986–87 moat samples; it represented >60% of the total carotenoids by weight. Carotenoids: chlorophyll *a* ratios in the moat mat were very high, ranging from 3 to 6.8.

A second, distinct pigment signature was associated with the aerobic benthos from a depth of 10 to 20 m (DH2, RH1) beneath the perennial lake ice (Fig. 2; Table 3). The top few mm of the RH1 sample was divided into an upper and lower section. Samples from two sites at this depth in botn 1985–86 and 1986–87 austral summer contained pigments characteristic of golden-brown algae such as diatoms and chrysophytes. Fucoxanthin accounted for >90% of the total carotenoids, and chlorophyll *c*, diatoxanthin, and diadinoxanthin were also present. Microscopic examination of these samples revealed diatoms,

Table 2. Lipophilic pigments from the cyanobacterial mat in the moat of Lake Hoare ($\mu g \cdot g^{-1}$ dry weight).

Pigment	1985–86	1986–87
myxoxanthophyll	9.9	4.7
lutein/zeaxanthin	0.8	0.3
chlorophyll *a*	5.2	1.0
echinenone	3.1	1.0
β-carotene	1.6	0.8
phaeophytin *a*	trace	trace
carotenoids:chlorophyll *a*	3.0	6.8

Trace amounts of 2 unknown carotenoids were also present.

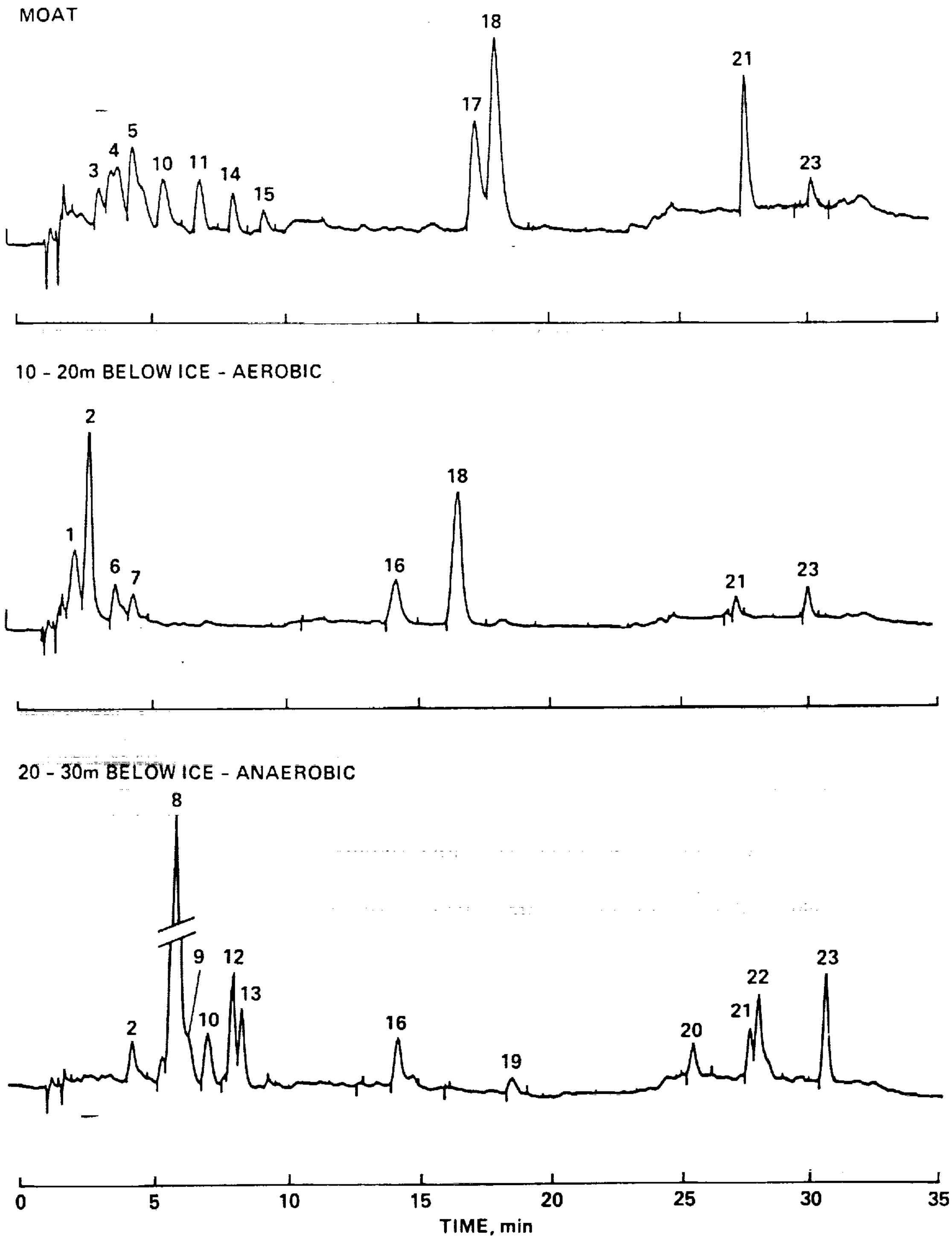

Fig. 2. High performance liquid chromatograms showing the relative absorbance at 440 nm of samples collected from the lake moat, aerobic, and anaerobic sediments in Lake Hoare. Peaks are numbered as follows: 1, chlorophyll *c*; 2, fucoxanthin; 3–5 myxoxanthophyll; 6, diadinoxanthin; 7, diatoxanthin; 8, alloxanthin; 9, phaeophorbide *b*; 10, lutein/zeaxanthin; 11–15, unknown carotenoids; 16, chlorophyll *a'*; 17, echinenone; 18, chlorophyll *a*; 19, unknown; 20, phaeophytin *b'*; 21, β-carotene; 22, phaeophytin *b*; 23, phaeophytin *a*.

Table 3. Lipophilic pigments from aerobic benthos beneath perennial lake ice cover ($\mu g \cdot g^{-1}$ dry weight).

Pigments	RH1 top layer 1985–86	RH1 lower layer 1985–86	DH2 1985–86	DH2 1986–87
chlorophyll *c*	38.8	27.0	62.0	1.2
fucoxanthin	246.9	117.7	449.2	6.5
chlorophyll *a'*	55.0	32.0	53.0	2.1
chlorophyll *a*	110.2	59.8	242.2	5.0
β-carotene	10.5	5.0	25.0	0.2
phaeophytin *a*	2.7	0.7	7.4	trace
carotenoids: chlorophyll *a*	1.5	1.3	1.5	0.9

Trace amounts of diadinoxanthin, diatoxanthin, and three unknown carotenoids were also found.

empty diatom frustules, and, in samples from site DH2 1986–87, golden-brown monads. Carotenoids: chlorophyll *a* ratios ranged from 0.9 to 1.5.

A third pigment signature was found at depths of 20 to 30 m (GH1, DH4) where the benthos is anaerobic (Fig. 2; Table 4). The pigment profile was dominated by a single carotenoid which has been tentatively identified as alloxanthin based on its spectral characteristics in chloroform (λ max = 436, 460, 489 nm) and in ethanol (λ max = 427, 451, 480 nm; Foppen, 1974), and by its relative migration on thin layer chromatographic plates. This carotenoid is found in cryptomonads such as *Chroomonas lacustris*, a prominent member of the lake phytoplankton during the austral summer (Parker *et al.*, 1982). Alloxanthin accounted for > 80% of the total carotenoids in these samples. Phaeophytins *b* and *b'* and lutein/zeaxanthin were probably derived from senescent green algae.

Layers from the AOSL below the top 1 mm of cores from DH2, GH1, and DH4 did not contain any detectable lipophilic pigments in sections down to a 15 cm depth. Moreover, an entire core from DH1 in both 1985–86 and 1986–87 contained almost exclusively empty diatom frustules. Pigments were not detected in this core, nor were cells stained with acridine orange to detect DNA or Nile red to detect neutral lipids using epifluorescent microscopy, suggesting an absence of viable cells.

Table 4. Lipophilic pigments from anaerobic benthos beneath perennial ice in Lake Hoare ($\mu g \cdot g^{-1}$ dry weight).

Pigments	GH1-Core1	GH1-Core2	DH4
fucoxanthin	11.5	6.8	59.4
alloxanthin	95.5	108.9	701.1
lutein/zeaxanthin	6.5	5.1	42.2
chlorophyll *a'*	11.8	12.5	72.1
phaeophytin *b*	-	-	trace
β-carotene	3.8	6.3	31.8
phaeophytin *a*	1.6	1.4	18.9

Three unknown carotenoids in trace amounts phaeophorbide b and phaeophytin b' were present.

Discussion

Three sites in Lake Hoare – moat, aerobic benthos (10–20 m) and anaerobic (20–30 m) benthos – had distinct lipophilic pigment profiles. These pigments probably reflected differences in 1) *in vivo* production by microalgae, 2) allochthonous inputs to the benthos from plankton, and 3) differential degradation of pigments.

Cyanobacterial moat communities resembled those found in nearby glacier-fed, ephemeral streams (Vincent & Howard-Williams, 1986). The epilithic stream communities are only metabolically active during brief periods in the austral summer when glacial meltwater allows rehydration of the community. During the remainder of the year, communities exist in a freeze-dried, but viable, state. Upon rehydration, photosynthesis is initiated within 20 min, suggesting that photosynthetic pigments are preserved during the winter (Howard-Williams & Vincent, in press). Moat communities had carotenoids: chlorophyll *a* ratios of 3 to 6.8. Carotenoids, such as β-carotene, undoubtedly help to protect cyanobacteria in moat and stream communities from surface downwelling irradiance which can reach 1680 $\mu E \cdot m^{-2} \cdot s^{-1}$ during mid-day in the austral summer (Palmisano & Simmons, 1987). While *in vivo* production contributes to pigment synthesis, allochthonous inputs are minimal. Pigment degradation in surface communities can result from photooxidation, however, temperatures below freezing would help to preserve pigments (Vincent & Howard-Williams, in press).

Golden-brown pigments in the aerobic benthos beneath perennial ice may be partly the result of a small amount of *in vivo* production by benthic diatoms. Most benthic diatoms, however, were senescent and failed to stain with vital stains. Thus, at 10–20 m, allochthonous inputs from sinking phytoplankton are probably the most important contributors to the pigment signature in our samples. A record of the brief plankton bloom during the austral summer appeared to be preserved in the top few mm of sediment.

Lipophilic pigments in the senescent or detrital-based benthic microbial communities were subject to degradation. Repeta and Gagosian (1984) suggested three processes for the transformation of fucoxanthin in marine systems: ester hydrolysis via heterotrophic metabolism primarily by zooplanktonic herbivores, dehydration via bacterial metabolism, and epoxide opening via slow chemical reaction. Because microzooplankton grazers are virtually absent from the perennially ice-covered lake, ester hydrolysis by this route would be very limited. Chlorophyll degradation to phaeophorbide by the actions of microzooplankton grazers (Shuman & Lorenzen, 1975) also would be severely reduced. Bacterial metabolism is slower at ambient temperatures of 0 °C (Pomeroy & Diebel, 1986), as are abiotic chemical reactions. The combination of low light, low temperatures, and the absence of microzooplankton grazers might lead to a reduction in rates of pigment degradation despite the supersaturation of lake water with oxygen (Wharton *et al.*, 1986).

It is unlikely that *in vivo* production in the anaerobic benthos, which contained apparently senescent green algae, contributed significantly to the pigments. Chlorophyll *b*, the primary light harvesting pigment in green algae, was absent; however, the presence of small amounts of phaeophytin *b'* (an epimer of phaeophytin *b*) probably resulted from decaying green algae. The pigment profile was dominated by alloxanthin, the principal carotenoid of cryptomonads; alloxanthin accounts for $>70\%$ of the total pigments in pure cultures of *Rhodomonas* sp. and *Cryptomonas ovata* (Pennington *et al.* 1985). The source of this pigment might be a planktonic bloom of cryptomonads. The structure of alloxanthin is similar to diatoxanthin in that both lack a 5,6 epoxide; diatoxanthin degrades more slowly than carotenoids such as fucoxanthin (Repeta & Gagosian, 1984). Thus, the relative abundance of alloxanthin in the anaerobic benthos in Lake Hoare may result from inputs from phytoplankton coupled with restricted degradation under anaerobic conditions.

Although laminations were apparent during macroscopic examination of cores (AOSL), lipophilic pigments were not found within these layers with one exception. Fucoxanthin was found at several mm depth in one core from DH2 on 11/85. While we did not find any evidence of cyanobacterial mats at depth in our pigment analyses, this may be due to a patchy distribution of mats beneath the ice cover, or to a PAR below the compensation intensity for cyanobacteria. Some redistribution of lipophilic pigments in the benthos may have occurred by sediment focussing by sliding of slopes (Hilton, 1985). However, our study of pigment biomarkers clearly showed that the lake benthos contains planktonic-derived detrital material in addition to the cyanobacterial mats previously reported (Wharton *et al.*, 1983).

Lipophilic pigments, as well as other lipid biomarkers such as stenols, stanols, and polar lipid fatty acids, serve as a chemotaxonomic record of microbial components. Orcutt *et al.* (1986) found that lipid biomarkers in an anaerobic prostrate mat from Lake Hoare (31 m depth) strongly implicated diatoms as well as cyanobacteria as biogenic sources. However, their argument for cyanobacterial contribution was largely based on the presence of C_{29} stenols which are also found in unialgal diatom communities from Antarctic sea ice (P. D. Nichols, personal communication). The phylogenetic diversity reflected in the lipophilic pigments from the benthos of Lake Hoare is, nevertheless, remarkable, considering the selective pressures of low, episodic irradiance, low temperature and low nutrients. With its virtual absence of grazers, ice-covered Lake Hoare provides a simplified system for studying the sources of pigment biomarkers and pathways of pigment degradation.

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Sedimentology.

LAKE HOARE, ANTARCTICA: SEDIMENTATION THROUGH A THICK PERENNIAL ICE COVER

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Abstract

Lake Hoare in the Dry Valleys of Antarctica is covered with a perennial ice cover more than 3 meters thick, yet there is a complex record of sedimentation and of growth of microbial mat on the lake bottom. Rough topography on the ice covering the lake surface traps sand that is transported by the wind. In late summer, vertical conduits form by melting and fracturing, making the ice permeable to both liquid water and gases. Cross sections of the ice cover show that sand is able to penetrate into and apparently through it by descending through these conduits. This is the primary sedimentation mechanism in the lake. Sediment traps retrieved from the lake bottom indicate that rates of deposition can vary by large amounts over lateral scales as small as a meter. This conclusion is supported by cores taken in a 3×3 grid with a spacing of 1.5 meters. Despite the close spacing of the cores, the poor stratigraphic correlation that is observed indicates substantial lateral variability in sedimentation rate. Apparently, sand descends into the lake from discrete, highly localized sources in the ice that may in some cases deposit a large amount of sand into the lake in a very short time. In some locations on the lake bottom, distinctive sand mounds have been formed by this process. They are primary sedimentary structures and appear unique to the perennially ice-covered lacustrine environment. In some locations they are tens of cm high and gently rounded with stable slopes; in others they reach $\sim$1 m in height and have a conical shape with slopes at angle of repose. A simple formation model suggests that these differences can be explained by local variations in water depth and sedimentation rate. Rapid colonization and stabilization of fresh sand surfaces by microbial mat composed of cyanobacteria, eukaryotic algae, and heterotrophic bacteria produces a complex intercalation of organic and sandy layers that are a distinctive form of modern stromatolite.

Introduction

Lake Hoare, Antarctica is one of several perennially ice-covered lakes in the Dry Valleys of South Victoria Land. It has been a site for biological research since the late 1970's (*e.g.* Allnutt, 1979; Allnutt *et al.*, 1982; Parker *et al.*, 1981; Wharton *et al.*, 1983). More recently, the lake has been intensively examined, including the local climatology (Clow *et al.*, 1988), the effect of climate on the ice cover (Wharton *et al.*, 1989), the water chemistry (Wharton *et al.*, 1986, 1987), and the sedimentary geology (Nedell *et al.*, 1987). One motivation for this work has been the use of Lake Hoare as an analog for what may have been a similar environment on Mars early in its history (*e.g.*, Squyres, 1989). Another has been that ice-covered lakes may be sensitive indicators of climatic change (Simmons *et al.*, 1987).

Modern lakes that formed as a result of glacial action are numerous (Smith and Ashley, 1985; Drewry, 1986), and glaciolacustrine settings are well known. The extensive literature on processes in these lakes and on modern and ancient glaciolacustrine deposits has been reviewed thoroughly (Gravenor *et al.*, 1984; Smith and Ashley, 1985; Drewry, 1986; Edwards, 1986; Brodzikowski and van Loon, 1987). Perennially ice-covered lakes are known in the antarctic (Wilson, 1982), the arctic (Barnes, 1960), and probably in high-altitude settings at lower latitudes (Johan Reinhard, pers. comm.). Drewry (1986) noted that perennially ice-covered lakes are distinct from other lakes in glacial settings because there is little direct coupling between wind and water. Little has been known until recently, however, about sedimentary processes in such lakes. In this paper, we describe processes operating in Lake Hoare, and report the occurrence of unusual sedimentary structures that may be useful in recognition of deposits of perennially ice-covered lakes elsewhere.

In a previous study, we suggested that the main source of deposition in Lake Hoare is downward transport of sediment through the ice cover (Nedell *et al.*, 1987). Although there were no direct observations of this process, sedimentation through the ice cover was inferred by the similar grain sizes and textures of the pebbly sand that is trapped on the surface of the ice and also comprises most of the sediment at the lake bottom. We suggested that sediment may penetrate the ice cover by migrating downward through porous ice at the surface, through water-filled vertical bubble columns that penetrate partially through the ice cover, and possibly through cracks in the ice that act as conduits.

In this paper, we present conclusive evidence that sediment migrates through the ice cover, using new data from the 1986/87 and 1987/88 field seasons. We describe unusual sedimentary structures, discovered on the lake bottom during 1986, whose origin can be explained by vertical settling of grains from sources in the ice cover. We also attempt to correlate the stratigraphy of cores taken from a grid on the lake bottom by comparing the texture, $CaCO_3$ and organic matter content, and biostratigraphy of the cores. All of these results are used to infer the nature of sedimentation in Lake Hoare.

Background

Lake Hoare (77°38′S, 162°53′E) is at the eastern end of Taylor Valley in the Transantarctic Mountains. The lake is 58 m above sea level, 4.1 km long, and 1.0 km wide at its widest point. It has a surface area of 1.8 km^2, a maximum depth of 34 m, and a mean depth of 14.2 m (Fig. 1). The perennial ice cover, which exceeds 3 m in thickness, overlies water at a temperature of 0°C. Less than 1% of the photosynthetically active radiation (0.4–0.7μm wavelength) striking the surface penetrates the ice cover (Parker *et al.,* 1982a). The lake is a terminoglacial lake in the usage of Brodzikowski and van Loon (1987), and is dammed by the Canada Glacier, which separates it from Lake Fryxell to the east. It receives water from glacial meltstreams, from nearby Lake Chad, and from groundwater inflow near the Canada glacier (Wharton *et al.,* 1989) during the austral summer. Lacking outflow streams, Lake Hoare loses water only by ablation at the surface of the ice, and by evaporation from a zone of open water 3–5 m wide (called a "moat") that forms at the lake margins during most summers. The material along the shoreline (except near the inflow streams) is coarse and poorly sorted, with boulders up to 50 cm in diameter (Nedell *et al.,* 1987).

The climate of the Dry Valleys is strongly seasonal as a result of the extreme southern latitude. There are about four months of continuous sunlight in the summer, and four months of darkness in the winter. The valleys are free of ice primarily because glacial flow from the polar plateau is obstructed by the Transantarctic Mountains. In addition, local ablation rates of snow and ice greatly exceed the annual snowfall. Recently, the first year-round meteorological observations were obtained for Taylor Valley by Clow *et al.* (1988). They established that the mean air temperature during 1986 at Lake Hoare was −17.3°C, the mean annual solar flux was ~ 92 W m^{-2}, the mean annual relative humidity was 55%, and the mean wind speed was 3.3 m s^{-1} (though highly variable).

Lake Hoare, like many of the other lakes in the Dry Valleys, supports numerous benthic and planktonic microorganisms that have adapted themselves to near-freezing temperatures, low light levels, and high oxygen concentrations (*e.g.* Parker *et al.*, 1982a; Wharton *et al.*, 1983). Microbial mats composed primarily of cyanobacteria, eukaryotic algae, and heterotrophic bacteria cover most of the lake bottom. An unusual feature of the lake is that the water column is supersaturated with O_2 and N_2, ranging up to 400% of saturation for O_2 and 160% of saturation for N_2 (Wharton *et al.*, 1986, 1987). Wharton *et al.* (1987) proposed a model in which gases are supplied to the lake by inflow of aerated water, with O_2 supplemented by biological activity. The presence of a perennial ice cover prevents free atmospheric exchange and apparently leads to the high gas concentrations. Only about half of the O_2 present is a result of biological processes; the rest is supplied by aerated inflow.

The data presented in this paper were collected in the 1985/86, 1986/87, and 1987/88 field seasons. During 1985/86, twelve sediment traps were retrieved from four dive holes, the locations of which are shown in Fig. 1. There were three traps in each dive hole. The traps were deployed in December 1982, and retrieved in November 1985. During the 1986/87 and 1987/88 field seasons, core samples were collected from the bottom of Lake Hoare, and the bottom topography and ice cover were observed by SCUBA divers working through five dive holes. Throughout the entire 1986/87 field season (October 1986 through January 1987) a video-equipped remotely-operated vehicle (Phantom 500 made by Deep Ocean Engineering, San Leandro, CA) was used to visually survey the lake bottom at distances of up to $\sim$100 m from the dive holes.

The Ice Cover

The environment in a Dry Valley lake is controlled to a large extent by the presence of the thick perennial ice cover (Wharton *et al.*, 1989). The ice cover eliminates wind-generated currents within the lake (Ragotskie and Likens, 1964; Hawes, 1983). It greatly limits exchange of gases between the water column and atmosphere (Wharton *et al.*, 1986, 1987) and light penetration into the water column below (Palmisano and Simmons, 1987). It also appears to have a profound effect on the nature and rate of sedimentation that takes place (Nedell *et al.*, 1987).

Lake Hoare has a perennial ice cover because the mean annual air temperature is so low. Water exists beneath the ice cover due to the fact that, for a limited period in the summer, air temperatures are above freezing and there is inflow of runoff and groundwater. The presence of a year-round ice/water interface in the lake therefore is due to the combination of very cold mean temperatures and comparatively warm summer maxima (Wilson, 1982). A smaller seasonal temperature range would result in a lake that is either frozen completely or one that melts fully in the summer.

The topography of the ice cover on Lake Hoare is extremely rugged. Ridges of friable ice that are 10–30 m wide and are oriented roughly parallel to the long axis of the lake (called "ablation tables") stand ~1 m above troughs, which are 1–10 m wide. At the beginning of the summer season, the ice in the troughs is hard and clear. Sand that has become trapped there may be visible in a layer a few tens of cm down. Later in the season after there has been some melting, this ice becomes porous and friable. In many instances, melting begins below the surface near a trapped sand layer, as solar energy is preferentially absorbed by dark sand grains. As melting progresses, small pools of water form on the ice surface in local topographic lows. Because the ice surface is extremely rough, wind-blown sediment is easily trapped on the ice, much of it in these pools. Sediment settles in the pools of water, from which it can migrate into the porous ice below.

Opposing any migration of material down through the ice cover is the upward movement of the ice. New ice freezes at the ice/water interface, while ice is removed from the surface by ablation. Parker *et al.* (1982b) have shown that microbial mat from the lake bottom migrates upward with the ice. In shallower regions of the lake, which are locally supersaturated with dissolved gases, pieces of mat that contain gas bubbles tear loose from the lake bottom and float up to the underside of the ice cover where they become frozen into newly-forming ice. Parker *et al.* (1982b) estimated that it takes 5–10 years for the mat to reach the surface, dry, and be dispersed by the wind. The cycling time for the ice cover was independently estimated by Clow *et al.* (1988) to be ~10 years, based on their calculated sublimation rate (35 cm yr^{-1}) and the measured ice thickness (3.3 m in 1987/88).

In order to gain a better understanding of the physical nature of the ice cover and its interaction with sediment, we made observations of the distribution of sand and bubbles in the ice during the melting of Dive Hole 2 (DH2). These observations were made during the last week of November 1986. Dive holes are created by a melting process, using a coil through which heated ethylene glycol is pumped (Love *et al.*, 1982). Melted water is pumped immediately from the hole. In order to avoid any alteration of the ice that might have been caused by the hole melting process itself, slabs of ice were chain-sawed from the edge of the hole, cutting into undisturbed ice. The hole was melted in this fashion down to a depth of 2.1 m. Once the hole reached this depth, it filled rapidly with water and could no longer be pumped dry. Apparently, the ice at and/or below 2.1 m was sufficiently permeable at that time that lake water was able to penetrate into the ice. All observations between 2.1 m and the bottom of the ice layer at 3.5 m were made by divers in the hole.

Figure 2 shows the features observed in the ice at DH2. A number of types of void spaces are present. These include: *(a)* large open void spaces floored by a thin layer of sand (only near the surface), *(b)* contiguous vertical bubble columns typically $\sim$ 10 cm in length, *(c)* vertical bubble columns composed of small, separated quasi-spherical bubbles, *(d)* columns of small quasi-spherical bubbles arranged to form distinctive pinnate structures, and *(e)* tabular bubbles $\sim$ 1 cm high and several cm in diameter. No obvious fractures were observed at this location.

Within 2.25 m of the ice surface, some void spaces of all types except the pinnate structures were observed to contain sand (Fig. 3). During the 1986-87 field season, we found no large tabular sand layers of the sort reported previously (Nedell *et al.*, 1987). Below 2.25 m, no sand was observed in the ice. It is interesting that the depth below which no more sand was observed coincided fairly closely with the depth below which the hole filled with water.

In October and November 1987, large lenticular sand bodies were observed in dive holes at GH1 and RH1. These sand bodies underlie ice with fewer, smaller bubbles than that described above. The ice and sand are inferred to represent shallow ponds that formed on the ice surface during the previous summer. The sand body at GH1 had a maximum thickness of about 3 cm. The sand at RH1 was in a flat layer 2 cm thick at a depth of about 1 m. The coarsest clast observed was a granitic boulder 36 cm in length.

Beneath each of these sand layers was ice with a bubble stratigraphy similar to that described in Fig. 2. The ice in each of the dive holes also contained a vertical fracture. The fractures were both about 5 mm wide at the widest point and were oriented in an east-west direction, parallel to the length of the lake. The fracture in GH1 extended to a depth of 2.7 m below the ice surface, and no sand was observed in it. The fracture at RH1 extended to a depth of 2.6 m and contained coarse-grained, pebbly sand to a depth of 1.8 m. It is possible that each fracture formerly extended to the base of the ice cover, and that the unfractured ice below has been added since the fractures formed.

There is direct evidence that the ice cover becomes porous through its entire thickness during the late summer season. During the dives in January 1987, air expelled by SCUBA divers was observed at the ice surface near three dive holes, where it produced bubbles in small pools of water on the surface of the ice.) During a diving operation at DH1, bubbles were observed surfacing through a fracture ~8 m away from the dive hole. The fracture itself was ~5 cm wide at the top and extended ~20 m up the lake towards the west. The fracture was located on the edge of one of the ablation tables paralleling the long axis of the lake. When bubbles were observed surfacing through this fracture, no bubbles were observed surfacing through the dive hole itself. Also, no bubbles were observed surfacing through the fracture except when a diver was under the ice in the vicinity of the crack.

At DH2 bubbles were observed to surface through small (< 5 cm diameter) holes at two locations in the ice cover ~12 and 15 m from the dive hole towards the Canada Glacier. It was not clear if these holes were associated with a fracture. As at DH1, when bubbles were observed surfacing through these holes, no bubbles were observed surfacing through the dive hole. A day after the dive, bubbles were still surfacing through some of the same small holes near the dive hole. Finally, bubbles were observed surfacing through small holes in the ice ~5 m west of GH1. All of these observations indicate that the ice cover is locally permeable to both gases and liquids.

One other observation regarding the ice cover may be noteworthy. Boulders are observed scattered over much of the ice surface. Typical sizes are ~ 1 m. Near the Canada Glacier, boulders may have been deposited directly onto the ice by the glacier. However, the most plausible source of the boulders toward the other end of the lake is that they rolled from the steep slopes bordering the lake (especially on the north side) out onto the ice surface. Interestingly, some boulders lie several hundred m from the lake shore. When boulders are intentionally rolled out onto the lake surfaces from heights of up to 1000 m above the lake shore on the steep north side during the summer, they do not progress more than ~ 10 m onto the ice surface. One possible explanation for this discrepancy is that boulders might roll a great deal further over smoother ice during the winter. Another is that they may roll out onto the margin of the lake, and then somehow become rafted toward the center of the lake by lateral movement of the ice. We understand neither the details of any possible lateral ice motion nor a mechanism that could cause it. However, the possibility of lateral ice motion must be kept in mind when the formation of sedimentary structures on the lake bottom is considered, since we show below that the primary mechanism by which sediments enter the lake is downward transport through the ice cover.

Lake Bottom Materials

Our previous work (Nedell *et al.,* 1987) showed that the lake bottom sediment is moderately sorted, medium-grained sand. The composition ranges from lithic arkosic to feldspathic lithic sand (using the classification scheme of Folk, 1980). Materials from the ice cover and lake bottom are mineralogically indistinguishable.

Microbial mats composed primarily of cyanobacteria, eukaryotic algae, and heterotrophic bacteria cover much of the bottom of Lake Hoare (Wharton *et al.,* 1983; Parker and Wharton, 1985). The filamentous cyanobacterium *Phormidium frigidum* forms the matrix of the mat, and diatoms comprise the largest number of algal species. The actively growing surficial mat is usually less than 5 mm thick. Mats form a variety of distinct morphological structures including small (< 10 cm) columns, knobs, and pinnacles (Fig. 4).

The benthic microbial mats of Lake Hoare may be considered modern stromatolites. Stromatolites are defined as organosedimentary structures produced by sediment trapping, binding, and/or precipitation as a result of interaction with microorganisms (Awramik *et al.*, 1976; Walter, 1977). The antarctic microbial mats are currently trapping and binding sediments, precipitating various minerals, and forming alternating laminae of organic and inorganic material (Parker *et al.*, 1981; Parker and Wharton, 1985; Wharton, 1982;

Wharton *et al.*, 1983). The type of mat and resultant stromatolitic structure depend on environmental factors including the amount of light penetrating the ice cover, the alkalinity of the water, dissolved gas levels, and sedimentation rates and processes. Because there is no bioturbation of the sediment or microbial mats that would result in a disturbance of the organosedimentary structures, a coherent record of alternating organic and inorganic-rich layers is preserved in the benthic sediment.

An additional important input of organic matter to the sediment is phytoplankton deposition. Palmisano *et al.* (1989) have shown that pigments characteristic of phytoplankton found in Lake Hoare contribute to the photosynthetic pigment content of lake-bottom samples. It is probable that phytoplankton are deposited annually on both sediment surfaces and the surfaces of microbial mats. Also, organic matter and calcium carbonate may be deposited through the ice surface.

Sediment Trap Data

Twelve sediment traps were deployed at the bottom of Lake Hoare in December 1982 and January 1983. They collected sediment settling out of the water column for approximately three years, and were retrieved in late November 1985. Three traps were arranged near each of four dive holes: DH1, DH2, DH4, and GH1. Each trap consists of an aluminum funnel that is 45 cm in diameter at the top, 47 cm deep, and narrows to 10 cm at the bottom. It is attached at the bottom to a 4 l plastic Nalgene container. The traps were placed in metal stands that place the top of the trap about 2 m above the lake bottom. At each dive hole, the traps were placed in a straight line with a spacing of 1-2 m, ~10 m away from the hole. The material retrieved from the traps was analyzed for total dry mass, and for amounts of organic matter, carbonate, gravel, sand, and mud. The data are shown in Table 1.

The average sedimentation rate for the three years that the traps were deployed was 3-4 mg cm^{-2} yr^{-1} near dive holes 1, 2, and 4. Near the glacier (GH1), the average sedimentation rate was substantially higher, 142 mg cm^{-2} yr^{-1}. Taking an average density of the sediment of $\sim$ 1.5 g cm^{-3}, the average thickness of sediment that was deposited in a year was between 0.002 and 0.003 cm yr^{-1} near dive holes 1, 2, and 4, and $\sim$ 0.09 cm yr^{-1} near the glacier (Table 1).

From the data on the total mass of material recovered from each trap, it is evident that the amount of sediment deposited from one dive hole to the next is extremely variable. A more interesting observation is that at dive holes 2 and 4, the amount of sediment collected in each trap was also highly variable. Even among traps separated by 1-2 m, the sedimentation over a three-year period varied by factors of up to 17. At each location, the amounts of biogenic sediment and mud in the three traps are similar. Between locations, the amount of mud decreases with increasing distance from shore. The difference between sample DH2-A and the other samples from DH2 and between sample DH4-C and other samples from DH4 is the amount of sand and gravel (Table 1). This pronounced small-scale spatial inhomogeneity in sedimentation rate is clearly inconsistent with settling from a lake-wide suspension of sediment or from sediment-gravity flows originating upslope from the traps. It is consistent with fallout from highly localized sediment sources within the ice cover.

Lake Bottom Topography

Further evidence for a sedimentation rate that varies considerably over short distances is provided by the topography of the lake bottom. The bottom of Lake Hoare shows topographic relief over a range of scales. Over distances of tens of meters, the bottom shows the undulating topography characteristic of the glacial moraines that surround the lake above the water line. At smaller scales, however, the topography is clearly non-glacial. In many areas, the bottom topography at a lateral scale of a few meters is dominated by small mounds (Fig. 5). These were first observed in detail at DH2. At this location they are roughly equant in shape, up to $\sim$ 1 m in diameter, and 20–40 cm in height. Typical surface slopes are $\sim$ 20°. Excavation of the mounds shows that they are composed entirely of sand; no rocks were present in any of the mounds examined. When blanketing mat material is removed and sand on the flanks is disturbed, it does not readily flow downslope; evidently the flanks of the mounds at DH2 are not at angle of repose.

The mounds at GH1, where the sedimentation rate is much higher than at DH2, are quite different. First, they are significantly larger, reaching $\sim$ 1 m in height. Second, they are commonly conical rather than gently rounded. They have sharp peaks, and flanks that are at angle of repose. They have little or no surface mat, and when disturbed, the sand on the flanks avalanches readily.

During October 1987, a sand ridge was observed on the lake bottom at DH1. The ridge, which was about 30 cm wide and 35 m long, paralleled the crack in the ice cover (discussed above) from which divers' bubbles were observed escaping in January. A mound was also observed in detail at the same location. It was 15 cm high and about 50 cm diameter. Figure 6 is a photograph of an epoxy peel of a box-core sample from the flank of the mound. Microbial mat similar to the mat adjacent to the mound is found below the surface of the box-core sample. A wedge of moderately-sorted, medium-grained sand, 3 to 10 cm thick within the core, covers the mat. Indistinct stratification within the wedge is roughly parallel to the surface of the sediment.

We believe that the sand mounds in Lake Hoare are primary sedimentary structures that result from the unique nature of sedimentation in a perennially ice-covered lake. If sand that works downward through the ice cover falls into the lake from a highly localized source, it will accumulate in a discrete pile on the lake bottom. If the ice cover is moving laterally, the ability of this process to form mounds has some significant implications for the rate at which sedimentation takes place. Individual mounds may form very rapidly, perhaps even in essentially instantaneous events. To our knowledge, such structures have not been found elsewhere, either undergoing deposition or preserved in the sedimentary record. We will return to detailed consideration of their formation in the discussion of sedimentation mechanisms below.

Core Analyses

During the 1986/1987 field season, sediment cores were taken from the lake bottom to examine the stratigraphy of the lake-bottom sediment in water away from the influence of the shore or inflowing streams. The cores were taken through DH2 at a water depth of about 9.5 m. A relatively flat area was sampled in a 3 $\times$ 3 grid pattern, with a spacing of 1.5 m between each point. These cores were retrieved in December, 1986. The core tubes were made of plexiglass, and had a diameter of 3.8 cm. One end was beveled, and they were driven into the lake bottom with a hammer to the depth at which frictional resistance made further penetration impossible. The cores ranged in length from 9 to 40 cm.

The stratigraphy of the cores from the 3 × 3 grid at DH2 is shown in Fig. 7. The cores contain alternating layers of fine- to coarse-grained sand and organic-rich layers (which include decomposed microbial mat, diatom frustules, and $CaCO_3$). Samples taken from cores 2, 3, and 5 were analyzed for salt content, organic matter, $CaCO_3$, gravel, sand, and mud; the results are summarized in Table 2.

Sediment in the cores is similar to sediment observed elsewhere on the lake bottom and to that collected in the sediment traps. Most of the core material is moderately to poorly sorted, fine- to coarse-grained sand and pebbly sand. Most layers contain very little interstitial mud (Table 2). The sand is laminated to very thin-bedded. Most layers lack sharp boundaries, and stratification generally is indistinct. Both normally and inversely graded beds are common, but many beds and most laminae are ungraded. Rhythmic sequences or varves were not observed in any part of these cores, and ripples and other evidence of tractive currents are absent. Both graded and inversely graded beds could represent deposition by sediment-driven gravity flows, but definitive evidence, such as a complete Bouma sequence, has not been recognized.

Scattered layers of microbial mat occur throughout the cores. Most mat layers are thinner than 1 cm. Internally, these are thinly laminated with wavy, continuous to discontinuous layers of cohesive biogenic material interstratified with thin laminations of fine-grained sand. Calcite crystals are locally embedded in the mat, and carbonate cement is conspicuous in sand within about 1 cm of some mat layers.

Mat layers are tentatively correlated in Fig. 7. Most sand layers are difficult to correlate, even over the short distance of 1.5 m between adjacent cores. One sequence of layers of relatively fine-grained sand, at the top of cores 1, 2, and 3, probably is correlative and shows a change in thickness from 8 cm to 4 cm in 3 m. Other differences between adjacent cores appear to reflect abrupt changes in layer thickness over very short distances. These changes probably are produced by the same processes that caused the local variation in sedimentation rates observed in the sediment traps and in the mounds on the modern lake bottom, suggesting that these processes have dominated lake-bottom sedimentation throughout the time represented by the cores.

In the DH2 sediment traps (Table 1), the rate of accumulation of mud was seen to be relatively uniform over the area sampled, suggesting that mud is introduced in suspension and is not controlled by the processes that form the mounds. In the cores, mud is more abundant in layers of microbial mat than in layers of sand (Table 2). If the rate of accumulation of mud during deposition of the sediment in the cores is comparable to the rate represented by the 1982-85 sediment traps, then the amount of mud in the mat layers suggests that a mat layer 3 mm thick could represent on the order of 10 years of accumulation.

The biostratigraphy of organic-rich and sand-rich layers in the cores was determined by observing subsamples of each layer microscopically (Table 3). All sediment layers observed contained diatom frustules; however, the organic-rich sediment layers usually contained at least an order of magnitude more frustules than the sand-rich layers. There are no planktonic diatoms in Lake Hoare. Therefore, any diatoms observed in a sediment layer are derived either from the surficial microbial mat or from sediment deposition through the ice cover. As noted above, diatoms are abundant in the surficial mat. They are also common in pieces of microbial mat found in the sediment on the ice cover surface. The relatively abundant occurrence of diatom frustules observed in the organic-rich layers is consistent with these layers having formerly been surficial microbial mat layers that subsequently have been buried by sediment.

Cyanobacterial filament sheaths are found only within the organic-rich layers, which consist of the cyanobacterial sheaths and embedded diatom frustules and $CaCO_3$ crystals. The filamentous matrix was most likely a former surficial microbial mat that became buried by sediment deposition. Once a rapid sand depositional episode has concluded, microbial mat can be expected to recolonize the new sediment surface in a matter of just a few years.

Finally, rhombohedral calcite crystals were observed only within the organic-rich layers in the benthic sediment. Wharton (1982) and Wharton *et al.* (1983) suggested that these sharp-edged crystals were precipitated *in situ* within the microbial mat as a result of the metabolic activity of the microorganisms.

Radiocarbon (^{14}C) dates for sediment samples collected from Lake Hoare are shown in Table 4. The radiocarbon ages of these sediments range from just under 2,000 to nearly 6,000 years before 1950. Despite these generally reasonable dates, there are trends in the data that cause us to doubt the validity of their details. In particular, samples obtained from greater than 20 cm in some of the cores (*e.g.*, 4, 8, 10, 11) date the same age as or significantly younger than surface material. One possible problem is that the source carbon for these systems may be relatively old to begin with. In this case, modern organisms would be fixing and cycling this old carbon, giving a relatively old apparent age. Another point is that relatively long-term recycling of carbon within the lake could result in an average age of the inorganic carbon in the the lake that is much older than the carbonate delivered by the meltwater streams. The variations in the radiocarbon ages in the mat and sediment profile could represent changes with time in the rate at which carbon is being recycled or deposited in the sediments. These changes could be driven by changes in phytoplankton productivity caused by changing nutrient inputs or ice thickness. The only conclusion we feel comfortable drawing from these radiocarbon dates, then, is that they are crudely consistent with a mean deposition rate of several tens of cm in several thousand years (*i.e.*, $\sim$ 0.01 cm/yr).

Sedimentation Mechanisms

We can now summarize the evidence concerning the sedimentation mechanisms that have operated in Lake Hoare. Clearly, a certain amount of the fine sediment on the lake bottom has been carried into the lake by inflow streams and has settled from suspension. Sediment trap and core data show that this component of the total sediment flux is relatively uniform spatially and very small relative to the flux of coarser material. We concentrate here, then, on the mechanism for deposition of the sand and gravel that dominate the bottom of Lake Hoare.

It is clear that the bulk of the sediments reached the lake bottom by downward transport through the ice cover. The supply of sand on top of and within the ice cover is abundant. Moreover, the characteristics of this sand (grain size distribution, grain texture, mineralogy) closely match those of the dominant materials on the lake bottom (Nedell *et al.*, 1987).

The source of the sand on the ice surface is eolian transport from the surrounding terrain. The initial trapping of the sand on the ice surface appears to involve a feedback mechanism (Simmons *et al.*, 1986), in which sand is preferentially trapped in local depressions on the ice. The accumulated dark sand results in enhanced local melting and further deepening of the depressions, increasing the trapping efficiency of the topography.

The dominant mechanism for downward transport of sand through the ice appears to be simple gravitational settling through water-filled fractures and degradation of the integrity of the ice, making it permeable to gases (as shown by surface observation of divers' bubbles) and liquid (as shown by rapid filling of a partially melted dive hole). When a water-filled vertical conduit comes into contact with sand in or on the ice, the sand can be washed downward to deeper levels in the ice. Transport appears to be particularly effective in the lower $\sim$ 1.25 m, where no sand was observed and where filling of the dive hole suggests that the ice is permeable to water through a significant fraction of the summer season.

The distribution of fractures and voids in the ice is spatially inhomogeneous, and one would expect a corresponding spatial inhomogeneity in deposition rates on the lake bottom. This expectation is borne out by the sediment trap data, by the poor correlation among closely-spaced cores, and, most dramatically, by the sand mounds observed at some locations. All three indicators show that sedimentation rates can vary by an order of magnitude or more over spatial scales of no more than a meter.

A particularly interesting problem concerns the details of the origin of the mounds. They form when a discrete source of sand in the ice cover allows sand to accumulate in a tightly localized region on the lake bottom. Because the mounds are so localized, it is apparent that the source must not move laterally by an appreciable amount during the deposition of the mound. If there is any significant lateral motion of the ice over fairly short timescales, then the formation of a given mound must be rapid with respect to this motion. In fact, if a vertical conduit in the ice cover tapped into a large sand body on or in the ice, this sand could rapidly drain to the lake bottom, forming a mound essentially instantaneously. It should be noted that mounds are not ubiquitous on the lake bottom. Some regions instead have a subtle rolling topography that may be indicative of a more steady sand supply. Such was the case in the location at DH2 where our grid of cores was obtained, although there clearly was significant spatial inhomogeneity in deposition rate there as well.

Based on our understanding of the sedimentation process, it is possible to construct a simple model of mound formation. We consider sand grains to be released from a point at the underside of the the ice cover that does not migrate laterally during formation of the mound. As each grain moves downward through the water column, it will experience motions resulting from viscous drag forces exerted by the fluid. The lakes of the Dry Valleys are stably stratified and, largely because of their ice covers, remarkably free of currents. The horizontal motions experienced by the grain as it settles, then, will be dominated by the forces that result from irregular flow of water around the non-spherical grain. The settling grain will undergo a series of small lateral deflections in response to these forces, each essentially random in direction. The absolute magnitude of each of these deflections will, of course, depend in a complex manner on grain size and shape.

Mathematically, the motions of the grain in the horizontal (x and y) directions may be thought of as a symmetric random walk. Therefore, if the grain begins its descent at $x = 0$, $y = 0$, the probability density function of x and y after the grain descends some distance z will be a normal (*i.e.*, Gaussian) distribution with mean 0 and some standard deviation σ. The value of σ will be a linear function of z, and also a function of grain size and shape. The probability density function of x and y for all grains can rigorously only be determined by integrating over all grain sizes and shapes; we adopt a normal distribution with mean 0 and standard deviation $\sigma = \sigma' z$. So, for a given lake depth z, this Gaussian distribution gives the probability density function of the initial x and y settling positions of the grains forming the mound. The mound, at least initially, will itself have a Gaussian profile. Growth of the mound will continue in this manner until angle of repose is first reached on some segment of the slope. Grains settling on an angle-of-repose slope will not remain fixed in position, but will move downhill at least until reaching a position where the slope angle is less than angle of repose.

In Fig. 8, we show the results of a numerical simulation of this process. Grains are randomly dispersed in x with a normal distribution, and allowed to accumulate with the proviso that when addition of sand would cause the angle of repose to be exceeded locally, that sand moves downslope until it reaches a position where its addition does not cause angle of repose to be exceeded. Angle of repose is taken to be 35°. The upper part of Fig. 8 shows a calculation for a mound similar to those observed at DH2. The height is ~30 cm, and the diameter is ~1 m. The slope is everywhere below angle of repose, and a gently-rounded Gaussian shape like that observed at DH2 is seen. For this mound $\sigma' = 0.025$ and, because $z = 9.5$ m, $\sigma = \sigma' z \simeq 24$ cm. The lower part of Fig. 8 shows the results of a similar calculation for a mound at GH1. Again, $\sigma' = 0.025$, since σ' is a characteristic of the grains, and should not vary significantly from one location to another. However, there are two differences from the mound in the upper figure. First, since the water is approximately twice as deep at GH1 as it is at DH2, we take $\sigma' z \simeq 50$ cm. Second, this mound is taken to have a volume that is approximately twenty times that of the one in the upper figure. The factor of twenty was chosen because it is comparable to the difference in sand sedimentation rates measured at DH2 and GH1 (*e.g.*, DH2-A and GH1-A in Table 1). With a volume this large, the mound grows to a height of nearly 1 m. Moreover, while it begins with a broad Gaussian shape, it eventually reaches angle of repose over most of its slopes, producing a straight-sided, sharp-crested cone like the ones seen at GH1. This simple model for mound formation suggests, then, that the differences in mound morphology and size observed at DH2 and GH1 can be attributed simply to the differing sedimentation rates and water depths at the two sites.

Colonization of the lake bottom by microbial mat appears to be quite rapid after a deposition "event" has taken place. Only at GH1 is the sedimentation high enough that most mounds are not colonized by mat. As subsequent deposition takes place, mat material is buried, forming the alternating sand and mat layers found in the cores. The organosedimentary structures being formed are modern-day stromatolites, albeit of a rather different structure than is found in most marine or lacustrine depositional environments.

Discussion

Recognition of Deposits of Perennially Ice-Covered Lakes — Deposits in the geologic record of perennially ice-covered lakes should be recognizably different from those of other settings. Glaciolacustrine settings commonly are dominated by inflow of clastic material (Smith and Ashley, 1985; Drewry, 1986). Depending on the relative density of inflowing water and lake water, overflow, interflow, and underflow may be important in distributing sediment throughout the lake. Away from the lake shore, glaciolacustrine deposits commonly include rhythmic layers (varves), and clear evidence of tractive flow from underflow or turbidity currents (Smith and Ashley, 1985; Weirich, 1986).

As we have shown, a perennially ice-covered lake is different in several important ways from more typical glaciolacustrine environments, substantially affecting the sedimentation process there. The sedimentary structures we observe may be distinguished from those formed by other processes. For example, the mound illustrated in Fig. 6 consists of moderately sorted, medium-grained sand with very subtle internal layering. Larger mounds may show distinct internal lamination, either from variations in the texture of sediment delivered to the mound, or from avalanching on flanks that are at the angle of repose. These mounds are not internally cross-stratified, which will be helpful in distinguishing them from hummocky cross-stratified sand deposited during storms (Eyles and Clark, 1986).

Conical mounds up to 2 m high of distinctly to indistinctly stratified, poorly sorted sandy gravel and pebbly sand observed in Pleistocene glaciolacustrine deposits in Scotland were interpreted by Thomas and Connell (1985) as iceberg dump structures. These are coarser and more poorly sorted than the mounds we have studied. Also, the interstratified sediment studied by Thomas and Connell (1985) consists largely of rhythmically interbedded sand and fine-grained sediment, and contains turbidites. The texture of the mounds and the nature of the associated deposits should distinguish sand mounds formed in a perennially ice-covered lake from coarser sediment dumped by icebergs.

Ice over a continental shelf in a glaciomarine setting could trap and transmit sediment as it does in Lake Hoare. Modern and ancient glaciomarine sedimentation has been reviewed by Anderson (1983), Eyles *et al.* (1985), and Anderson and Molnia (1989). Glacial and ice-rafted marine sediments commonly are reworked by currents or redeposited by sediment-gravity flows, so the preservation potential in marine environments of mounds like the ones we have studied may be less than in ice-covered lakes. Associated sediments also are generally different. In addition to reworked deposits showing marine influence, many glaciomarine sediments are associated with carbonates that in Phanerozoic examples typically include skeletal fragments of marine animals (Anderson and Molnia, 1989). Therefore, glaciomarine facies probably are readily distinguishable from deposits of perennially ice-covered lakes.

In summary, sand mounds may form where sediment is transported through ice in many glaciolacustrine or glaciomarine settings. Deposits of a perennially ice-covered lake like Lake Hoare should be distinctive, however, showing a sand mounds of the sort we have described, an assemblage of deposits generally lacking in evidence of tractive currents, and perhaps microbial mat developed where the sedimentation rate is low.

Relationship of Lake Hoare Stromatolites to Others in the Geologic Record — Awramik *et al.* (1976) have reviewed the distribution of known Holocene stromatolites. Most modern stromatolites occur in (1) unusually warm and highly saline marine habitats, (2) temperate, tropical, and/or alkaline lakes, or (3) certain streams and hot springs. Parker *et al.* (1981) first reported the presence of modern stromatolites forming in antarctic Dry Valley lakes, and suggested that these structures are unique in the Holocene world. The microbial mats associated with these structures are adapted to extremely low light intensities, cold temperatures, fresh to saline water, as well as conditions ranging from anaerobic to supersaturated with oxygen. Parker *et al.* (1981) and Simmons *et al.* (1985) have gone so far as to suggest that the low light intensity, lack of burrowing and browsing organisms, and lack of turbulence in these lakes may mimic the Precambrian deep-water stromatolite environment. Wharton *et al.* (1989) suggest that it is a misconception that stromatolites (including modern forms) develop only in warm and/or saline environments. Several periods of glaciation may have occurred during the middle and late Precambrian — time periods when life was microbial and stromatolites were abundant (Frakes, 1979; Anderson, 1983; Walter and Bauld, 1983, Edwards, 1986). The modern stromatolites of Lake Hoare could be more closely related to Precambrian stromatolites than previously recognized, and particularly to stromatolites formed in Precambrian polar environments.

Conclusions

Our results from the 1986/1987 and 1987/1988 field seasons indicate that the primary mechanism of sedimentation in Lake Hoare is downward transport of sand through the lake's perennial ice cover. Transport takes place through vertical conduits in the ice formed by seasonal warming late in the summer. Long-term average sedimentation rates are of the order of 0.01 cm/yr for much of the lake, but may be locally much higher for very short periods when a sand deposition event takes place. Sedimentation from discrete sources in the ice cover results in considerable spatial inhomogeneity in deposition over lateral scales as small as a meter. Rapid colonization of fresh sand surfaces by microbial mat produces the vertically and laterally complex intercalation of organic and sedimentary materials observed in our cores. In some locations, rapid and highly localized sedimentation has built distinctive sand mounds on the lake floor. The size and morphology of the mounds appear to be controlled directly by sedimentation rate and water depth. They are primary sedimentary structures that to our knowledge have not been found elsewhere, and that appear unique to the perennially ice-covered lacustrine environment.

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Figure Captions

Figure 1: Bathymetric map of Lake Hoare, showing the locations of dive holes in the ice cover.

Figure 2: Schematic representation of the form, distribution and concentration of the major features found in the ice at DH2. Void features observed include large open spaces floored with sand, contiguous vertical bubble columns, vertical bubble columns composed of small spherical bubbles, vertical bubble columns with pinnate geometries, and tabular bubbles. Voids filled with sand are shown as dark.

Figure 3: Sand-filled vertical bubble columns exposed on the wall of DH1. The horizontal structures in the ice are produced by the melting process.

Figure 4: Microbial mat on the bottom of Lake Hoare. Vertical structures are caused by liftoff of mat from the lake bottom, induced by production of buoyant gas bubbles that form under supersaturated conditions. Scale across the photograph is approximately 2 m.

Figure 5: Large sand mound at GH1. The mound is about 1 m high; marks on the bottom to the left of the mound are diver's footprints. Sand in the mound is near angle of repose and is covered by a very thin layer of organic material.

Figure 6: Epoxy peel of box core of mound at DH1 sampled on 25 October 1987. Peel is 14 cm wide. Sand in mound forms a 3 to 10-cm thick wedge at top of peel overlying a later of microbial mat like that on the lake bottom surrounding the mound.

Figure 7: Stratigraphy observed in cores taken at DH2. Inset shows core layout. Inferred stratigraphic correlations across grid rows are indicated by roman numerals. Substantial variations among the cores indicate that there are large variations in sedimentation rate over lateral scales smaller than the core spacing (1.5 m).

Figure 8: Results of numerical models of mound formation. Top: Ten profiles during growth of a mound in 9.5 m of water at DH2. Result is a rounded, nearly Gaussian-shaped mound with a height of $\sim$30 cm and a diameter of $\sim$1 m. Bottom: Result for a mound at GH1. Compared to the calculation at top, the water depth is increased to 20 m, doubling the lateral grain dispersion, and the mound volume is scaled upward by the approximate difference in sedimentation rate measured at the two sites. The result here is a large conical mound with slopes at angle of repose.

Table 1
Summary of Sediment Trap Data: Deployed 1982, Retrieved 1985
All weights reported in grams

Trap ID	Total dry Mass	Organic Matter	Carbonates	Gravel	Sand	Mud	Other
DH1-A	17.52	0.81	6.61	0.00	0.25	9.85	0
DH1-B	24.38	0.63	13.79	0.00	0.43	9.53	0
DH1-C	16.95	1.10	6.04	0.00	0.61	9.20	0
DH2-A	44.25	3.11	2.46	3.88	32.65	2.15	0
DH2-B	4.58	0.21	1.52	0.00	0.00	2.85	0
DH2-C	4.98	0.48	1.41	0.64	0.10	2.35	0
DH4-A	2.58	0.10	0.22	0.00	2.13	0.13	0
DH4-B	2.08	0.13	0.18	0.00	1.67	0.10	0
DH4-C	36.35	0.32	0.89	2.58	32.39	0.17	0
GH1-A	633.00	39.69	18.38	0.00	572.42	1.51	1.00
GH2-B	544.60	4.30	15.30	0.00	522.74	2.24	0.02
GH3-C	856.20	7.53	30.74	0.00	813.97	2.94	1.02

Table 2
Cores from 1986-87 Field Season, Lake Hoare, Dive Hole 2

Sample	Dry Wt.	% IC[1]	%OC[2]	% Salt	% Gravel	% Sand	% Mud
2B	21.9881	1.9	0.1	0.0	0.3	97.5	0.2
2C	13.3451	2.5	0.1	0.1	4.9	91.4	1.0
2D	13.7311	2.3	0.5	0.3	0.5	95.7	0.7
2E	5.1587	3.4	0.3	0.2	0.0	93.0	3.1
2F	22.7844	1.6	0.1	0.1	0.5	97.6	0.1
2G	4.7932	3.7	1.0	0.2	1.2	91.8	2.1
2H	13.7731	1.9	0.0	0.2	0.9	96.7	0.3
2I	1.9907	40.8	13.1	*	0.0	40.2	5.9
2K	22.4789	2.1	0.1	0.1	0.3	97.0	0.4
3B	14.5965	1.9	0.1	0.1	0.5	97.2	0.2
3C	10.6028	1.7	0.0	0.0	1.8	96.3	0.2
3D	13.1691	2.4	0.1	0.2	0.0	96.5	0.8
3E	13.4224	2.1	0.2	0.3	0.6	96.4	0.4
3F	1.4695	19.2	2.7	*	0.0	75.0	3.1
3G	1.4176	38.9	9.0	*	0.0	49.0	3.1
3H	10.8840	1.5	0.1	0.0	0.0	98.3	0.1
5B	6.1408	3.7	0.1	0.2	0.6	94.0	1.4
5D	9.8468	2.6	0.2	0.3	1.5	93.4	2.0
5E	9.9516	2.0	0.2	0.0	1.9	95.7	0.2
5G	11.8841	1.7	0.2	0.0	0.0	97.9	0.2
5I	18.1378	2.3	0.2	0.2	0.3	96.2	0.8

1. Carbonates soluble in 3 N HCl.

2. Organic matter soluble in 20% H_2O_2.

* Not measured.

Table 3
Microscopic observations of sediment cores from 1986-87 field season, Dive Hole 2[1]

Sample	Type[2]	$CaCO_3$[3]	Frustules[4]	Filaments[5]
2A	OR	+	++	–
2B	SR	–	+	–
2C	SR	–	+	–
2E	OR	–	++	–
2F	SR	–	+	–
2G	OR	–	++	–
2I*	OR	+	++	+
2J	OR	+	++	–
2K	SR	–	++	–
3A	OR	+	++	+
3E	SR	–	+	–
3F*	OR	+	++	+
3G*	OR	+	++	+
5A	OR	+	++	+
5C	OR	–	++	+
5D	SR	–	+	–
5F	OR	+	++	+
5G	SR	–	+	–
5H*	OR	+	++	+
5I	SR	–	+	–

1. Several subsamples (< 1 g per slide) of each core layer were observed microscopically at 100, 400, and 1000 × magnification.

2. Type refers to visual observation of layers in split cores; OR = organic-rich, SR = sand-rich layer.

3. Carbonate crystals are composed of calcite and typically rhombohedral in shape (Wharton, 1982; and Wharton *et al.* 1983) + = calcite crystals present; – = crystals not observed.

4. Diatom frustules predominantly species of *Caloneis, Hantzchia, Navicula, Nitzschia*, and *Stauroneis*. ++ = abundant (> 100 frustules per slide); + = few (< 100 frustules per slide).

5. Filaments are observed in a "matrix" which consists of cyanobacterial filament sheaths (probably *Phormidium* sp.), diatom frustules, and $CaCO_3$ crystals. + = filaments present; - = filaments not observed.

* sample light-colored probably because of $CaCO_3$.

Table 4
^{14}C Dates for Sediment Samples in Lake Hoare, Antarctica,
Collected During 1985 and 1986 Field Seasons.

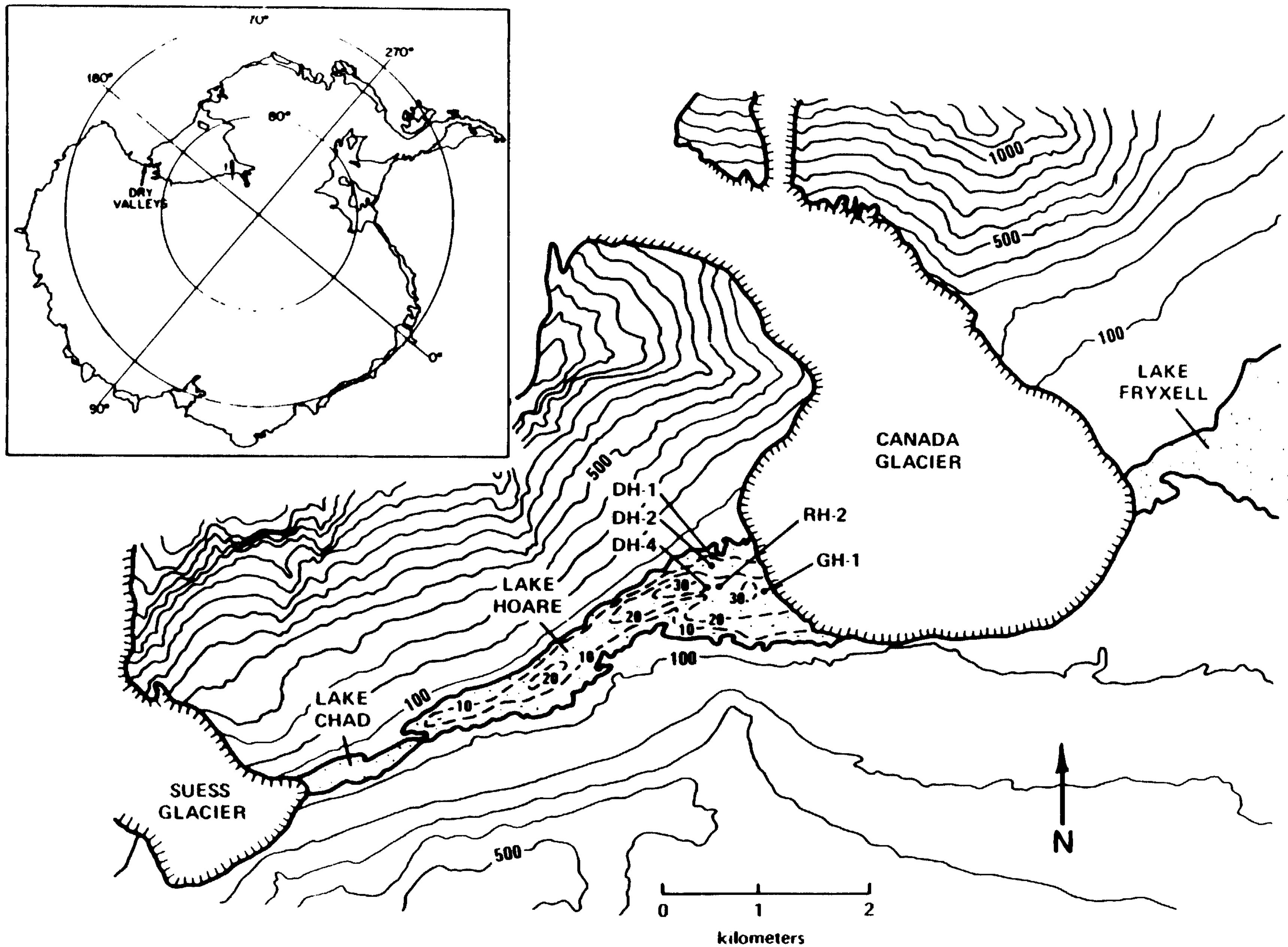
Sample[1]	^{14}C Age[2]	Collection depth, cm	% $CaCO_3$	% Organic matter
1. 85DH2S[3]	1935 ± 608	0	48.19 ± 3.71	18.87 ± 0.24
2. 85DH47A1	3975 ± 1205	0 - 1.5	3.95	4.72
3. 85DH47B	3435 ± 495	4 - 6	3.34	5.75
4. 85DH47F	4180 ± 475	20 - 21	2.59	4.72
5. 85RH1S	2780 ± 650	0	6.82	6.64
6. 85RH18E	2650 ± 1115	7.5 - 9.5	48.24	6.19
7. 86DH2S	4355 ± 305	0	*	*
8. 86DH1	3645 ± 250	15 - 22	*	*
9. 86DH2-8A	5935 ± 55	0	*	*
10. 86DH2-8B	4620 ± 80	34	*	*
11. 86DH2-9B	4295 ± 95	35	*	*

1. Samples 1-8 were analyzed by Kruger Enterprises, Inc., Geochron Laboratories Division, Cambridge, Mass.; samples 9-11 were analyzed by the NSF Accelerator for Radioisotope Analysis, University of Arizona, Tucson, Ariz. To the extent possible, $CaCO_3$ was removed from all samples before dating. All samples were corrected for ^{13}C.

2. Years before 1950.

3. Sample collection notes: **1.** *85DH2S* Surface microbial mat from DH2 collected by grab sample 11-85. **2.** *85DH47A1* Surface microbial mat layer obtained from core from DH4 collected 11-85; black flocculent material; H_2S smell. **3.** *85DH47B* Organic and sand-rich layers obtained from core from DH4 collected 11-85. **4.** *85DH47F* Organic-rich layers obtained from core from DH4 collected 11-85. **5.** *85RH1S* Surface mat from RH1 collected by grab sample. **6.** *85RH18E* Organic-rich layers obtained from core from RH1 collected 11-85. **7.** *86DH2S* Surface mat from DH2 collected 1-87 by grab sample. **8.** *86DH1* Layers of organic-rich material obtained from core from DH1 collected 1-87. Layers dark-green, leaf-like, very competent and tissue-like. **9.** *86DH2-8A* Surface mat from DH2 collected 11-86. **10.** *86DH2-8B* Layer of organic-rich material obtained from core from DH2 collected 11-86. **11.** *86DH2-9B* Layer of organic-rich material obtained from core from DH2 collected 11-86.

* Not measured



70°
180°
270°
80°
DRY VALLEYS
0°
90°
1000
500
100
LAKE FRYXELL
CANADA GLACIER
DH-1
DH-2
DH-4
RH-2
GH-1
30
20
10
LAKE HOARE
LAKE CHAD
SUESS GLACIER
N
0
1
2
kilometers

Figure 1

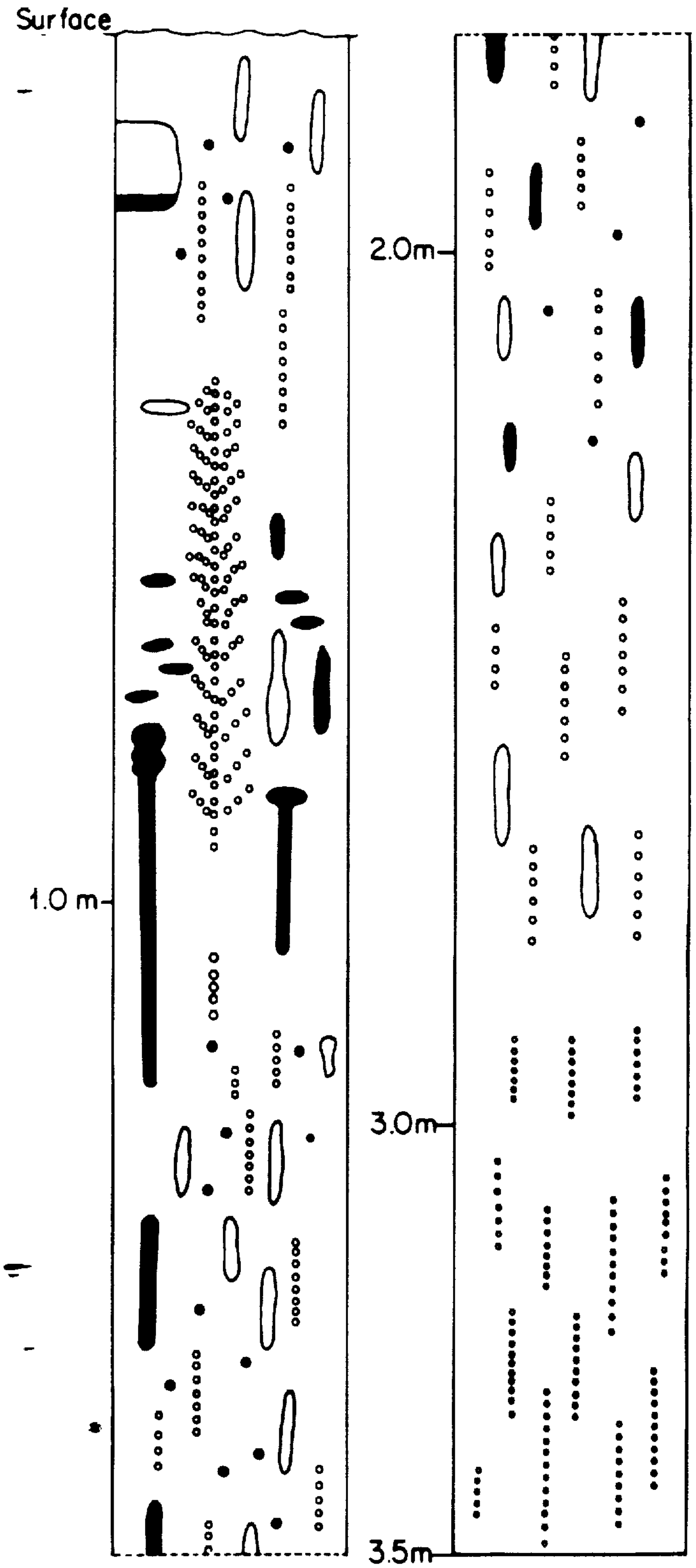


Figure 2

Figure 3

Figure 4

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Figure E

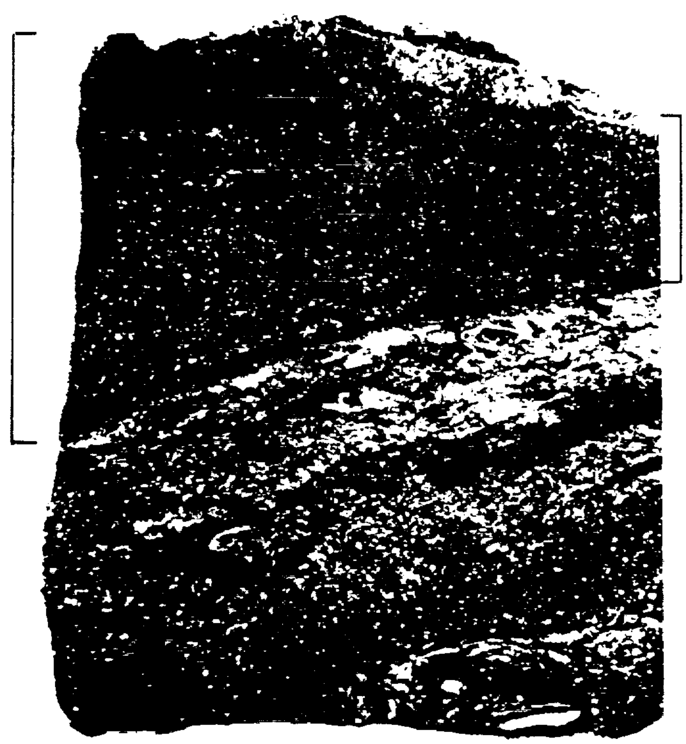

Figure 6

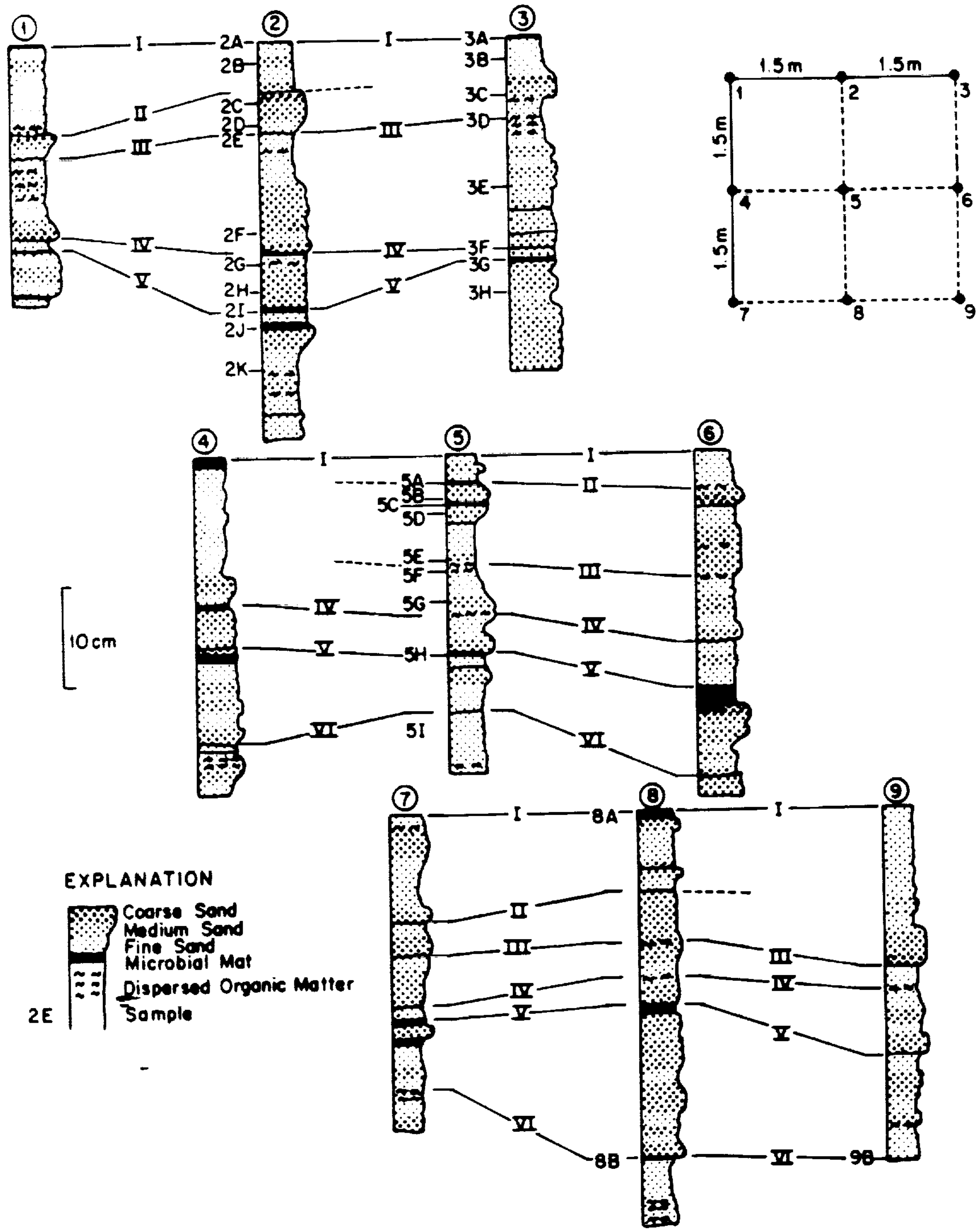
1.5m
1.5m
1.5m
1.5m
10cm
EXPLANATION
Coarse Sand
Medium Sand
Fine Sand
Microbial Mat
Dispersed Organic Matter
Sample
2E

Figure 7

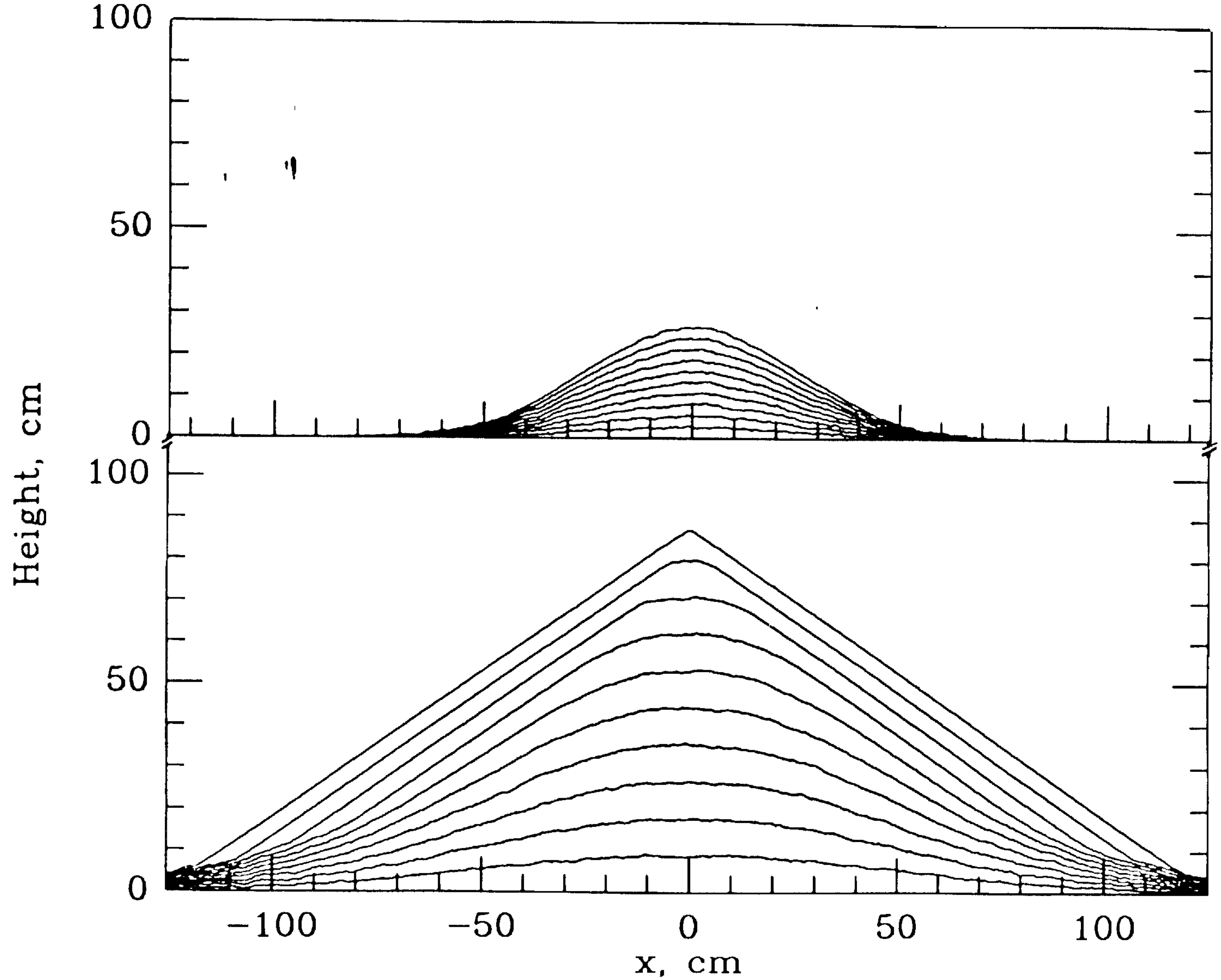

Figure 8

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AN ANTARCTIC RESEARCH OUTPOST AS A MODEL FOR PLANETARY EXPLORATION

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During the next 50 years, human civilization may well begin expanding into the solar system. This colonization of extraterrestrial bodies will most likely begin with the establishment of small research outposts on the Moon and/or Mars. In all probability these facilities, designed primarily for conducting exploration and basic science, will have international participation in their crews, logistical support and funding. High fidelity Earth-based simulations of planetary exploration could help prepare for these expensive and complex operations. Antarctica provides one possible venue for such a simulation. The hostile and remote dry valleys of southern Victoria Land offer a valid analog to the Martian environment but are sufficiently accessible to allow routine logistical support and to assure the relative safety of their inhabitants. An Antarctic research outpost designed as a planetary exploration simulation facility would have great potential as a testbed and training site for the operation of future Mars bases and represents a near-term, relatively low-cost alternative to other precursor activities. Antarctica already enjoys an international dimension, an aspect that is more than symbolically appropriate to an international endeavor of unprecedented scientific and social significance — planetary exploration by humans. Potential uses of such a facility include:

1) studying human factors in an isolated environment (including long-term interactions among an international crew);

2) testing emerging technologies (e.g., advanced life support facilities such as a partial bioregenerative life support system, advanced analytical and sample acquisition instrumentation and equipment, etc.); and

3) conducting basic scientific research similar to the research that will be conducted on Mars, while contributing to the planning for human exploration. (Research of this type is already ongoing in Antarctica).

1. INTRODUCTION

Throughout the history of space exploration, simulation facilities have played an important role in defining and designing space missions. The complex nature of the challenge and the many options that will be available as humans embark on exploration missions beyond Earth orbit will require that, in the early stages, simulation facilities be established on Earth. Indeed, a full range of simulation facilities may be required to enable us to understand the complexities involved in exploration missions that transport humans to the Moon once again and then outward to the planet Mars. These facilities may range from small scale environmental simulations and/or computer models that will aid in the development of new materials to full scale mockups of spacecraft and planetary habitats. It may be useful to place a large scale simulation facility such as a planetary habitat designed for the Martian surface in an Earthly environment that duplicates (to as great a degree as possible) the conditions in which it will be used by future occupants.

Antarctica's potential as an analog environment for planetary exploration was recognized by space flight pioneers Ernst Stuhlinger and Wernher von Braun as early as 1966 [1]. They suggested that "the basic problem was how to provide a group of scientists in a remote Antarctic outpost with the necessary support which would permit them to live and work under extremely hostile conditions. This problem, which is easy to formulate but very hard to solve, is encountered in a very similar form by those preparing the astronauts' flight to the Moon and later to the planets."

Recently, there has been renewed interest in Antarctica as an analog for space environments and exploration [2,3,4,5,6]. Both the United States and the Soviet Union have initiated planning activities and research directed at missions and possibly settlements on the Moon and Mars. Such missions may be preceded by timely investigations of the effects of long-term isolation on human behavior and performance and to

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this end, Antarctica can be used as an experimental analog for much of the research in space sciences. Studies on the psychological, physiological and sociological aspects of long-term isolation will provide insight into many issues having an impact on the achievement of mission goals through maximization of crew performance, efficiency and effectiveness. Research activities similar to the scientific exploration activities of a crew on a planetary surface could also be conducted on an Antarctic outpost. Additionally, an Antarctic simulation facility could be used to develop and test critical technological systems and concepts that may be required for future exploration missions.

It is suggested that a planetary exploration simulation facility in Antarctica could provide an immediate, economical analog (relative to the cost of a mission to Mars) for the development of a program directed at human exploration of Mars. The essential elements of a planetary exploration simulation facility are discussed and locations on the Antarctic continent where such a facility might be located are considered. The political and programmatic aspects of such an endeavor are also considered.

2. ELEMENTS OF AN ANTARCTIC PLANETARY EXPLORATION SIMULATION FACILITY

A round-trip, human mission to Mars is currently anticipated to take up to three years, including a stay time of about one year on the Martian surface [7,8,9,10]. Given the importance of such a mission and the high costs likely to be associated with the effort, it is crucial to learn as much as possible prior to sending the first crew to Mars. The need to understand the problems associated with future missions to Mars argues for a simulation facility capable of providing mission planners and astronauts with an environment that resembles planetary conditions with the greatest possible degree of fidelity. The facility must also be capable of providing preparatory work in the range of scientific disciplines likely to be included on a Mars mission. A simulation facility might include three principal elements:

— Human factors research;

— Testing of critical technologies; and

— Research in scientific disciplines relevant to the exploration of other planetary bodies.

Human factors research would comprise such areas as crew selection; training; psycho-social interaction; habitat design and architecture; human-machine interactions; and psychological, behavioral and physiological studies of humans in remote, isolated and potentially hazardous environments.

The testing of critical technologies would include the use and evaluation of advanced life support facilities such as a partial bioregenerative life support system. Continuous recycling of fresh water and the harvest of fresh vegetables would not only provide greater comfort to the inhabitants of such a remote field research facility but would significantly increase their self sufficiency while providing engineering tests of the equipment. Portable life support systems for planetary extravehicular activity could also be tested at this simulation facility. While there are significant differences in atmospheric pressure between Mars and Earth, (7.0 mb vs 1.0 bar respectively) [11] temperature regimes are not that dissimilar between Mars and the dry valleys of Antarctica. The need for comfortable, low bulk environmental protection in both of these hostile environments is a necessity for activities outside of the habitat. The test and evaluation of relevant technologies should include the design of teleoperated rovers and portable analytical instrumentation as well as techniques for sampling and *in situ* analysis during egress activities. Teleoperated vehicles could augment the capabilities of scientists at remote field locations by providing them with the ability to explore inaccessible or hazardous areas. At the same time, the use of such vehicles at the simulation facility would provide an opportunity to test, evaluate and develop telepresence technology for use on Mars.

Scientific investigations at the simulation facility would include field studies in geology, biology and astronomy relevant to the science to be carried out on future planetary missions. This would not only provide a broader base for developing a scientific rationale for the exploration of planetary bodies but would give the occupants of the simulation facility meaningful tasks in which they have a vested scientific interest. Such "real work" would be useful to the planning of scientific studies on the exploration missions and would, at the same time, allow more valid human factors research to be conducted.

It is envisioned that a small outpost, capable of supporting four to six individuals, would be placed in the Antarctic to accomplish the aforementioned simulation activities. The facility would be capable of housing this group for up to one year (or longer). Initial designs for an Antarctic habitat may differ significantly from that which will eventually be emplaced on the surface of Mars (i.e., Antarctic designs would not have to account for the lower atmospheric pressure of Mars), although the facility could and indeed, should evolve towards the actual design of the habitat to be used on Mars.

3. WHERE IN ANTARCTICA: DRY VALLEYS OR POLAR PLATEAU?

The continent of Antarctica has been isolated from other land masses for some 60 million years and is now separated from its closest neighbor, South America, by the 1,000 km Drake Passage. The continent lies almost entirely within the Antarctic Circle and has the highest average elevation in the world, about 3,000 m. At 14.2 million km^2, Antarctica has an area approximately the size of the United States and Mexico combined. It has a challenging environment, with temperatures on the polar plateau averaging -60° C and an annual snowfall rate of < 5 cm (water equivalent). Most of the continent is covered by an ice cap several km thick; only a small percentage, mainly near the coast, remains ice-free. The dry valleys of southern Victoria Land, the largest of these ice-free regions, are approximately 4,000 km^2 in area [12].

Fig. 1 shows the orientation of Antarctica with respect to the other land masses as well as the location of the dry valleys of southern Victoria Land.

Generally speaking, there are two locations on the continent of Antarctica suitable for the emplacement of a simulation facility(ies) - on the polar plateau or in an ice-free dry valley. The unique attributes of each site lend them to very different functions as simulation facilities.

It is suggested that the Antarctic dry valleys are potentially better analogs to planetary conditions (with an emphasis on field exploration and science) than sites on the polar plateau. Simulation facilities in the southern Victoria Land dry valleys would be close to the main U.S. base, McMurdo Station (fig. 2) and New Zealand's research facility, Scott Base (fig. 3), making them logistically more convenient and less expensive to maintain than stations high on the polar plateau.

Fig. 1 Map of the dry valley region within southern Victoria Land, Antarctica. The Polar Plateau is to the west.

Fig. 2 McMurdo Station, the main U.S. station in Antarctica, as seen during the austral summer from nearby Observation Hill. McMurdo Station is located on the southeast end of Ross Island.

Fig. 3 Scott Base, the main New Zealand station, located 3 km south of McMurdo on Ross Island. Photo courtesy of National Science Foundation, 1981.

The dry valleys of southern Victoria Land, occurring between 160° and 164° E longitude and 76°30' and 78°30' S latitude, are the largest and best known of the ice-free "oases" located around the Antarctic continent. The dry valleys are free of ice primarily because glacial flow from the polar plateau is obstructed by the Trans-antarctic Mountains. The potential evaporation greatly exceeds the annual snowfall, producing an extremely arid (desert) environment. The dry valleys receive about four months each of sunlight, twilight and darkness. The mean annual temperature is about -20° C. During the winter months, strong *föhn* winds descend from the polar plateau and buffet the valleys. Year-round temperature, light and wind conditions for the dry valleys are illustrated in fig. 4.

Fig. 4 Climate data recorded during 1986 from Taylor Valley, Antarctica. Shown are daily averaged temperature and light levels. Winds are the daily maximum of 6 hour averages. Data are abstracted from the results of Clow et al (1988).

Fig. 5 Remote field camp on the shores of Lake Hoare, Taylor Valley, Antarctica. The camp accommodates four to six scientists during the austral summer. The Canada Glacier is in the background.

Because of the extreme cold and arid conditions, the dry valleys form what may be the best terrestrial analog of the surface conditions existing on Mars [5,13,14,15] . In fact, several scientists have recognized the dry valleys as an area where life has adapted to extreme conditions with little available liquid water and have conducted biological investigations there in preparation for the Viking exploration of Mars [16]. Research relevant to planetary science (exobiology and geology) is ongoing in the dry valleys. This research, jointly funded by the National Science Foundation (NSF) and the National Aeronautics and Space Administration (NASA), involves a number of scientists from the United States and abroad. As a result of the systematic study of the physical and biological processes occurring in the dry valleys, a better understanding of conditions on Mars has been gained and a scientific rationale for future exobiological and geological investigations of that planet is being developed.

Sites on the polar plateau, such as the Soviet Vostok station or the U.S. South Pole station, might be more valuable for the simulation of spacecraft (versus planetary base) science for long duration space flights. In contrast to the dry valley regions, very little field research is conducted on the plateau. Research efforts at these facilities are oriented toward observational science (such as upper atmospheric physics and solar astronomy) which can be conducted primarily from the confines of a structure placed on the plateau.

We suggest a planetary exploration simulation facility could be located in the dry valleys of southern Victoria Land near current planetary scientific research activities, such as those in Taylor Valley (fig. 5) or Wright Valley (fig. 6). A habitat in the dry valleys could support a small group of scientists and engineers (4-6) during the austral summer (and eventually over the winter months as well) and could duplicate a long sojourn on the Martian (or lunar) surface.

Fig. 6 The New Zealand Station on the shore of Lake Vanda in Wright Valley, Antarctica. The station is the base-camp for summer field operations in the Valley. Vanda station has accommodated two winter over parties.

4. WHY ANTARCTICA?

Antarctica has a number of important characteristics that warrant serious consideration for its use as a site for a planetary exploration simulation facility:

— It is an environment of real danger and isolation, yet fatalities are uncommon during research or training

During the forty-one years between 1946 and 1987, the U. S. Antarctic program has experienced 29 incidents resulting in 52 deaths [17]. Despite the fact that the continent of Antarctica is remote and quite hostile to those who go, with proper logistical support and safety awareness, science can be conducted with an acceptable level of risk.

— There is a logistics infrastructure primarily directed toward science support already in place that could sustain operations in the dry valleys or on the polar plateau

The United States Antarctic Program has developed a remarkably efficient logistics infrastructure over the course of the last three decades. Ships and aircraft bring major supply items and personnel to the station at McMurdo and ski equipped LC-130 aircraft and helicopters then transport personnel and needed supplies to the interior of the continent. This logistics infrastructure has had the benefit of being developed around the needs of scientists, since research is the primary activity of the U.S. program.

— Ongoing scientific research opportunities in Antarctica are relevant to planetary sciences

The dry valleys of southern Victoria Land are probably the best terrestrial analog to the Martian environment. The valleys, being cold, dry deserts, are natural laboratories for studying life in extreme environments. Additionally, the perennially ice covered dry valley lakes are currently being used as models of ice covered lakes that may have existed on the surface of Mars during a warmer, more clement epoch. The very nature of this work makes it suitable as part of a Mars research outpost simulation. The inhabitants of a dry valley habitat would have relevant scientific research to conduct during their long simulation sojourn.

— There is a history of human factors research in the Antarctic, recognizing it as an analog to space flight and planetary exploration

Because of Antarctica's geographical isolation, the continent has been used as a natural laboratory for studying small populations over long periods of time. During the winter months, these small groups of people are completely shut off from the rest of the world except for radio communications and perhaps a single airdrop resupply during mid-winter. A number of investigators have compared these experiences with those that may one day be encountered during long duration space flight or on a remote planetary base.

— The Antarctic Treaty (Section 5.) provides a proven and workable framework for international cooperative exploration and scientific efforts

This remarkable treaty has provided the international science community with a means to explore a continent unfettered by political barriers familiar to the rest of the world. For more than thirty years the Antarctic Treaty has maintained peace on that continent and a spirit of cooperation found nowhere else. With this powerful tool at hand, the southern continent is the ideal location for a simulation of a multinational space science project.

Many aspects of planning for planetary exploration can and will be carried out by other modes of simulation. Studies of some aspects of human factors may be performed more effectively in a laboratory setting that provides for greater control and monitoring capability but at the expense of fidelity and realism. Underwater habitats continue to be successfully used as analogs to long duration space flight [18] but they cannot simulate crew conditions or field research activities similar to those to be conducted on a planetary surface. In contrast, a habitat in Antarctica would involve researchers doing meaningful work relevant to their counterpart lunar or Martian tasks, without strict supervision. Planetary simulation facilities might be established in temperate desert regions and could be useful in developing and evaluating newly designed equipment. In addition, computer simulations may be of some value in simulating planetary environments. There are also other areas on Earth, mostly in the Arctic, that are similar to Antarctica with respect to isolation and cold desert environmental conditions. Nonetheless, Antarctica has important and unique characteristics that make a compelling case for its use as a site for a planetary exploration simulation facility.

5. ANTARCTIC POLITICS

The exploration of Antarctica officially assumed an international character in the late 1950's, when more than 60 research bases were established there by 12 nations as part of the 1957-1958 International Geophysical Year. The participation of the international community was codified in the language of the Antarctic Treaty, signed in 1959 by the 12 nations. As of mid-1988, 38 nations had acceded to the Treaty [19]. The Treaty specifies that Antarctica shall be used for peaceful purposes only and that participating nations should take an active role in conducting science on the continent. Scientific results and observations are to be freely exchanged and the treaty contains a provision for free access and inspection of all bases by any nation that is an active participant within the framework of the Treaty. The Treaty has provided for international cooperation and is one of the few long-standing international arenas in which United States and Soviet positions coincide. It is important to note that this spirit of cooperation between these two world superpowers has survived the last 30 years in the face of previous cold war tensions.

Because of its success, the Antarctic Treaty has been suggested as a model for regulating future international activity on the Moon and planets [20,21]. It is possible that future manned Mars or lunar programs will be international endeavors involving the United States, the Soviet Union, European nations, Japan and other countries. The desirability of international cooperation may have an overriding influence on site selection for a simulation facility. The continent of Antarctica is, by the very nature of the Antarctic Treaty,

uniquely suited for the combined efforts of international partners and would provide a logical location for such a simulation facility.

6. PROGRAMMATIC CONSIDERATIONS

The implementation of a planetary exploration simulation facility in an analog environment such as the Antarctic will require cooperation between government agencies and the private sector within the United States and between the United States and other interested countries. We suggest an orderly progression of events that will lead to the establishment of an international planetary exploration simulation facility located in the dry valleys of Antarctica. Eventually this facility could be used by a number of nations in preparation for the establishment of scientific outposts on the Moon or Mars.

The current United States program in Antarctica is implemented by the Division of Polar Programs within the NSF. U.S. policy is established by Presidential memorandum. Within the United States, the first logical step toward an Antarctic simulation facility would be the establishment of a joint NSF-NASA working group to draft a memorandum of agreement for conducting joint research in Antarctica. This step would be followed by the identification of interested private sector businesses and perhaps the establishment of an international *ad hoc* study group to assess the potential for an international planetary exploration simulation facility in Antarctica. Initial contact between NSF and NASA on this subject has already occurred. Within the European Space Agency (ESA) there is an ongoing study directed at defining the use of Antarctic bases as models of space flight. The conclusion of a preliminary report recognizes the value of Antarctic facilities to ESA long term program planning and their broad potential scope [22]. The other major spacefaring nations, the Soviet Union and Japan, also have active Antarctic research programs and may be already considering the use of Antarctica as a space analog.

If a planetary simulation facility is considered for Antarctica, environmental concerns must be addressed. For example, the dry valleys are unique and fragile ecosystems which may not recover rapidly from environmental damage [23]. Clearly, any plan for a planetary exploration simulation facility must be predicated on a requirement of acceptable levels of environmental impact. However, the placement and operation of a planetary exploration simulation facility need not be detrimental to the dry valley environment. In fact, if properly planned, the advanced technologies associated with such a test facility could reduce the impact of the existing research activities through more complete recycling of wastes, increased use of remote and automatic data acquisition systems and reduction in the number of required support personnel.

Future planetary exploration missions are feasible if spacefaring nations of the world begin now to develop the capabilities to work cooperatively and with maximum efficiency. The development and utilization of a planetary simulation facility in the dry valleys of Antarctica would provide program managers, scientists, and engineers with a realistic simulation of humans living and working on the planet Mars. The experience gained at an Antarctic planetary testbed will help facilitate the peaceful expansion of humans in the solar system and eventually lead to the establishment of lunar bases and Martian outposts that will make the best use of human talents.

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