PERMANENT ICE COVERS OF THE MCMURDO DRY VALLEYS LAKES, ANTARCTICA: LIQUID WATER CONTENTS

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A novel method of analyzing ice temperature records is applied to several years of data from Lake Hoare and Lake Bonney (Antarctica) to estimate vertical distributions of liquid water in the perennial ice covers at the end of summer melting seasons. Three years of ice temperature data at Lake Bonney (1993-1995) show that the ice contained 20% liquid water located at 1 to 2.5 m below the ice surface near the end of all three melting seasons. Liquid water fractions at Lake Hoare were low (<15%) throughout the ice column in 1986. In 1987 and 1988 liquid water fractions increased to a maxima of 70% at depths between 1.5 and 2.5 m. Maxima in liquid water content for both lakes were coincident with layers of bubbles having arching morphologies which were predominantly associated with pockets of sedimentary material (silts, sand, and gravel). Interpreting these bubble morphologies in context of the ice energy budgets indicates that the majority (>90%) of the liquid water in the ice is generated when visible radiation is absorbed by lithogenic matter within the ice during the austral summer. Our work also demonstrates the potential use of this energy budget analysis for monitoring ecosystem-level responses to climate variability.

INTRODUCTION

Ice covers on lakes influence the exchange of momentum, thermal energy, and materials between the water column and the atmosphere. Thus the presence of an ice cover as well as the physical attributes of the ice cover influences a lake's ecosystem processes. The McMurdo Dry Valleys contain numerous lakes with thick (3 to 20 m) perennial ice covers. Because of their permanence, the ice covers control the lakes sedimentation processes [e.g., *Neddell et al.,* 1987; *Squyres et al.*, 1991; *Anderson et al.*, 1993; *Doran et al.*, 1994], the quantity and quality of light penetrating into the water column for photosynthesis [*Palmisano and Simmons*, 1987; *Lizotte and Priscu*, 1992; *McKay et al.*, 1994], as well as the transfer of gases into and out of the lakes [*Craig et al.*, 1992; *Wharton et al.*, 1993; *Priscu*, 1996].

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These ice covers exist as solid physical barriers during winter months. Each austral summer, however, the combination of relatively warm air temperatures (~0°C) and continual solar radiation leads to partial melting of the ice covers. At this time, it is presumed that the ice covers become permeable to materials [*Neddell et al.*, 1987; *Squyres et al.*, 1991; *Anderson et al.*, 1993] and gases [*Anderson et al.*, 1993].

Permeability is a physical characteristic of ice covers which is difficult to quantify directly because its complex nature involves several factors including porosity, connectedness, and tortuosity. However the permeability of liquid-containing composite materials is generally related to the amount of the liquid phase within the solid matrix. Therefore developing methods for determining the liquid water fractions of the permanent ice covers improves our ability to monitor the physical properties of the ice covers that directly relate to permeability and indirectly relate to the exchanges of matter and energy between the lakes and their surroundings.

An ice cover's liquid water fraction also is a quantity that can be used derive energy budgets over seasonal time scales. The ice covers on the McMurdo

Dry Valley lakes are believed to be sensitive indicators of climate change [*Wharton et al.* 1993]. Hence it is logical to assume that monitoring both annual and seasonal energy budgets will provide additional indices of climate variability that can be related to global climate change.

Internal melting in permanent ice covers has not been studied or quantified previously, despite its apparent importance in ecosystem processes and its potential for monitoring climate variability. However we have recently devised a novel method to quantify the liquid water contents of the ice covers at the end of the austral summer. The method consists of analyzing energy budgets of known depth intervals in the ice during the austral autumn when freezing fronts are propagating from the surface to the bottom of the ice covers. We apply this method to ice temperature records collected from Lakes Bonney and Hoare and provide the first quantitative estimates of the liquid water contents of the ice covers which are interpreted in context of their morphological features and seasonal climatological parameters. These analyses yield direct insights into the overriding mechanisms that control internal ice melting, ice permeability to materials, as well as ice-energy transfer processes.

METHODS

Site Description and Instrumentation

Lakes Hoare and Bonney are located in the Taylor Valley, Antarctica. General hydrology of the lakes and the physicochemical and biological characteristics of their water columns are described in detail by others [e.g. *Green et al.*, 1993; *Spigel and Priscu,* this volume]. The ice covers have surface topographies consisting of alternating ridges and troughs that are generally oriented parallel to the valley floor and prevailing winds. Previous evaluation of the annual mass balance of the Lake Hoare ice cover in the late 1980s indicated that about 35 cm yr^{-1} of ice sublimated or evaporated from the ice surface, while an additional 60 to 70 cm of the ice melted at the base during the austral summer [*Clow et al.*, 1988]. This annual loss was roughly balanced by new ice growth at the base of the ice in the austral autumn and winters. From these analyses it is apparent that the ice covers are physically dynamic features with a net upward movement of the ice at approximately 30 to 40 cm yr^{-1}.

Thermocouples (Campbell, Type T) were placed in the 4 m thick ice cover near the center of the east lobe of Lake Bonney in December of 1992 at 0.5, 1.0, 2.0, and 3.5 m from the ice surface. Thermocouples at Lake Hoare were placed in the ice (original depths 0.5, 1.0, 1.5, 2.0, 2.5, and 3.0 m) in December of 1985 and retrieved in the austral summer of 1988. Data were collected at hourly intervals in Lake Bonney using a Campbell 21X data logger powered by gel-cell batteries. This data logger concurrently recorded downwelling and upwelling photosynthetically active radiation (PAR) measured with LiCor 192S sensors set 1 m above the ice surface. Instantaneous readings, daily averages, maximum, and minimum daily temperatures in the Lake Hoare ice cover were recorded at local noon. Because ablation at the ice surface occurs throughout the year, the thermocouples experienced net movement relative to the ice surface throughout the seasons. However the array of thermocouples is presumed to have moved as one unit and the depths of the thermocouples relative to each other are assumed to have remained constant.

Liquid Water Contents Estimated from Autumnal Energy Budgets

The ice temperature records were analyzed during the periods when freezing fronts were propagating through the ice in the austral autumn. The energy budget for this period (conceptually illustrated in Figure 1) was used to calculate the changes in latent heat for depth intervals defined by the thermocouples. The change in latent heat was then used to estimate the amount of liquid water generated during the proceeding summer melt season.

The change of total internal energy within a layer of ice, E_T, is controlled by the fluxes of energy in and out of the layer. Within the ice column, energy is transferred primarily by short-wave radiation, E_r, and heat conduction, E_c, such that,

$$E_T = E_r + E_c.$$

Temperature gradients ($\delta T/\delta z$) above defined layers were used to calculate the time-integrated conductive heat loss out of the layers according to

$$E_c = \int k_i \frac{\partial T}{\partial z}\, dt$$

where k_i is the bulk thermal conductivity coefficient for the ice. Temperature gradients below the freezing fronts

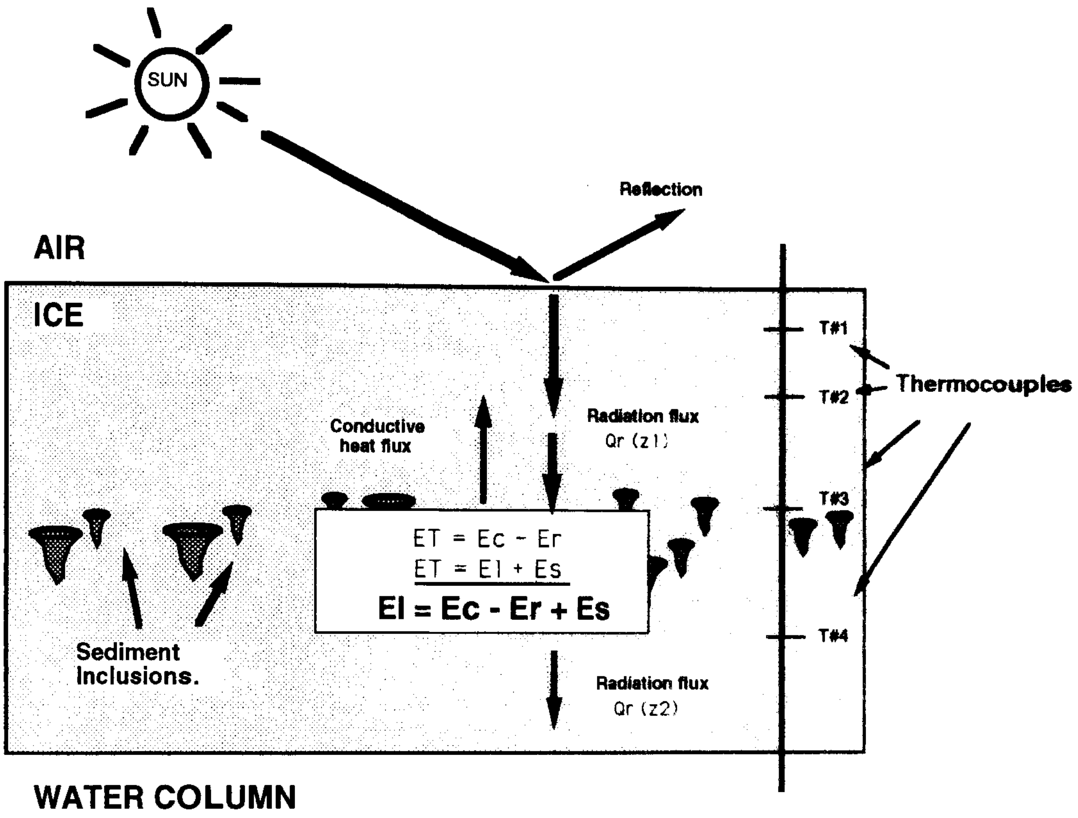

Fig. 1. Conceptual model of the energy budget used to derive the time and depth integrated flux of latent heat from layers within the permanent ice covers. The thermocouples and the sediment inclusions are illustrated schematically.

are isothermal at 0°C when freezing fronts are progressing through the defined layers. Therefore, no heat is transferred by conduction from below.

The coefficient for the bulk thermal conductivity of the ice is estimated as a volume-weighted function of the individual conductivity coefficients of the materials comprising the total ice volume such as pure ice (k_{pi}), air (k_a), and gravel (k_g), such that

$$k_i = k_{pi}\ V_{pi} + k_a\ V_a + k_g V_g$$

where V_{pi}, V_a, and V_g are the fractional volumes of the materials within a given ice volume.

The absorption of short-wave (penetrative) radiation in a layer of ice was estimated from a simple radiative-transfer model where down-welling irradiances above the ice surface, $I_d(0+,PAR)$, were propagated into the ice, $I_d(0-,PAR)$, after allowing for reflection, R_o, from the ice surface by the following:

$$I_d(0-,PAR) = I_d(0+,PAR)\ (1-R_o).$$

Downwelling irradiances at depth, $I_d(z,PAR)$, were then modeled assuming an exponential decrease with increasing depth in the ice, such that

$$\frac{\partial I_d(z,\ PAR)}{\partial z} = -K_d(z)\, I_d(0-,\ PAR)$$

where $K_d(z)$ is the diffuse downwelling attenuation coefficient of the medium, a composite apparent optical property [sensu *Kirk*, 1994]. During September 1995, spherical radiometers (LiCor) were frozen in the ice of the east lobe of Lake Bonney at 1 m intervals to monitor in-ice scalar irradiances, $I_o(z, PAR)$. These in-ice irradiances were then used to compute attenuation coefficients, $K_o(z, PAR)$. Values for $K_o(z, PAR)$ during the first several months of their deployment ranged from 1.2 to 1.8 m^{-1} in the upper 2.5 m. Assuming attenuation coefficients for scalar irradiances approximate those for downwelling irradiances, we utilize a value of 1.5 m^{-1} for K_d in our initial model. Analysis of the energy budget's sensitivity to a range of K_d values are addressed in a later section.

I_d(z, PAR) is a flux of photons over a broad band (400 to 700 nm) of electromagnetic radiation and is not an exact measurement of the energy flux. The energy flux at various depths in the ice was estimated by converting the modeled PAR values (μmol photons m^{-2} s^{-1}) to radiative energy fluxes, Q_r (W m-2), using a factor of 4.2 μmol photons J^{-1} [*Morel and Smith*, 1974]. Because this factor was derived from measurements in oceanic waters, an ice radiative transfer model [*Fritsen et al.*, 1992] was used to check the applicability of this relationship to in-ice irradiances of PAR. This comparison showed that the factor varied from 4.1 to 4.5 depending on optical depths and constituents within the ice

The depth and time-integrated absorption of short-wave energy, E_r, was then estimated as

$$Er = \int\int \frac{\partial Q_r(z)}{\partial z} dz\, dt$$

where the integrals are over the depth of the thermocouple spacing and the period of active freezing. Because the attenuation of downwelling irradiance is due to scattering plus absorption, the radiative transfer model is likely to yield an upper estimate of E_r. Again, sensitivity analysis, presented below, addresses uncertainties in the energy budget calculations associated with the assumptions of this radiative transfer model.

Net changes in total internal energy within a layer resulting from radiation and conductive heat exchanges are realized as changes in latent heat energy, E_l, and specific heat energy, E_s;

$$E_T = E_l + E_s$$

The change in specific heat was calculated as the product of the bulk specific heat capacity (c_i), bulk density (ρ_i), and the depth averaged change in ice temperature (T_i) within the layer. The bulk specific heat capacity and the bulk density were assumed to be proportional to the fractional volume of each constituent (air, ice, gravel) and its individual specific heat capacities. Values used for these individual coefficients are listed in Table 1. Solving the total energy budget in terms of changes in specific and latent heat caused by conductive and radiative energy transfer processes allows us to define the change in the latent heat within a layer of ice as,

$$E_l = E_c + E_s - E_r$$

Thereby, the change in latent energy within a defined depth interval can be evaluated once the specific heat change is known and the net radiation and conductive heat fluxes are evaluated. The change in the liquid water fraction (V_w) is then evaluated as,

$$V_w = E_l L_w^{-1} \rho_w^{-1} \Delta z^{-1}$$

where L_w and ρ_w are the latent heat and density of water, respectively, and Δz is the distance between the thermocouples for which E_l was derived.

RESULTS

East Lobe Lake Bonney

Ice temperatures during the austral summers on the east lobe of Lake Bonney were isothermal at 0°C for approximately 86 days in November to February of 1993-1994 and 66 days during December to February of 1994-1995 (Figure 2). In contrast, midwinter temperature gradients were typically 6°C m^{-1}. Freezing fronts passed the first (0.5 m) thermocouple between February 4 and February 12 in each of the three years, indicating a relatively narrow time-period when seasonal freezing of the liquid water within the ice was initiated.

The rate at which the freezing fronts propagated into the ice was similar in 1994 and 1995, with the rate of freezing front progression averaging 0.030 m d^{-1} in 1994 and 0.034 m d^{-1} in 1995. Conversely the 1993 rate of freezing front progression was 0.019 m d^{-1}, approximately 1.5 to 2 times slower than in 1994 or 1995. An analysis of the annual variability in the rate of freezing front progression in context of the overall energy budget shows that the liquid water content at the end of the austral summer in 1993 was two-fold greater than in 1994 and 1995 (Table 2).

The energy budgets also indicate that the liquid water was not distributed evenly in the ice. Liquid water content was low (5 to 9%) in the uppermost portion of the ice (0 to 1 m), whereas, the liquid water content was higher (27 to 32%) between the second and third thermocouples in each of the three years. The second and third thermocouples were located at 1.0 and

TABLE 1. Notation and Description of Variables.

Variable	Variable Description	Value(s)	Units
E_T	internal energy	variable	KJ m^{-2}
E_c	conductive heat loss	variable	KJ m^{-2}
E_l	change in latent heat	variable	KJ m^{-2}
E_r	absorbed short-wave radiation	variable	KJ m^{-2}
E_s	change in specific heat	variable	KJ m^{-2}
I_d	downwelling irradiance of PAR	variable	µmol photons m^{-2} s^{-1}
K_d	attenuation coefficient for downwelling irradiances	1.5	m^{-1}
K_o	attenuation coefficient for scalar irradiances	variable	m^{-1}
k_i	thermal conductivity of lake ice	variable	W m^{-1} °C^{-1}
k_g	thermal conductivity of vapor phase	variable	W m^{-1} °C^{-1}
k_{pi}	thermal conductivity of pure ice	variable	W m^{-1} °C^{-1}
k_s	thermal conductivity of sand	variable	W m^{-1} °C^{-1}
L_w	latent heat of fusion for water	335	KJ kg^{-1}
R_o	specular reflection of PAR	0.05	%
Q_r	shortwave-radiation	variable	W m^{-2}
V_g	fractional volume of gravel	variable	%
V_{pi}	fractional volume of pure ice	variable	%
Vs	fractional volume of sand	variable	%
Vw	fractional volume of water	variable	%
z	depth	variable	m
ß	PAR conversion factor	4.2	µmol photons J^{-1}
ρ_w	density of water	1000	kg m^{-3}
ρ_i	density of pure ice	917	kg m^{-3}

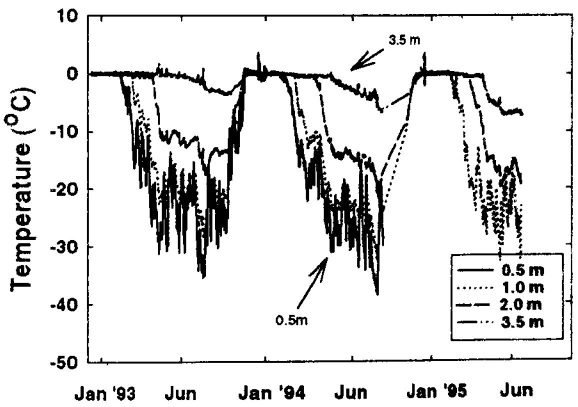

Fig. 2. Ice temperature records from 1992 to 1995 in the east lobe of Lake Bonney. Depths denote where the thermocouples were originally positioned in the ice. Note that the uppermost thermocouple was above the ice surface for a period in 1993 before being reset in the ice, and from September 1994 to the end of the record.

2.0 m in the 1993 and 1994 season; in 1995 they were estimated to be 50 cm higher in the ice at the time of the autumnal freezing. The vertical resolution of V_w (1–1.5 m) is relatively large and makes no distinction as to the dispersion of the liquid water throughout these depth intervals. For instance, when V_w is 20% over 1 m, we cannot make the distinction between a 20 cm layer of liquid water with 80 cm of solid ice or a 1 m layer of ice with 20% of the water dispersed throughout.

Vapor bubbles with arching morphologies and sediment pockets were found from 1.8 to 2.2 m in the ice on the east lobe of Lake Bonney [*Adams et al.*, this volume, also see color images on the CDROM that accompanies this volume]. *Adams et al.*[this volume] have shown these bubble morphologies to be indicative of refrozen pockets of liquid water. Bubbles indicative of pockets of liquid water were conspicuously absent both above and below these depths. The bubble distributions and morphologies [*Adams et al.*, this volume], together with our energy budgets, lead us to conclude that the values for V_w of 20–30% in the east lobe of Lake Bonney over the 1 to 2 m depth intervals are indicative of a 30 to 40 cm layer consisting of 80 to 90% liquid water at a depth of about 2 m from the ice surface. We further contend that the ice above and below this liquid water layer remains relatively solid (with liquid water comprising 5 to 10%). Values for

TABLE 2. Energy Budget Analysis Based on the Ice Temperature Records from the East Lobe of Lake Bonney for Years 1993 to 1995. Period of Freezing Front Passage is the Time during which the Freezing Front Passed from the Top to the Bottom of the Given Depth Interval. The Average Temperature Gradient ($\partial T/\partial z$) was the Average Temperature Gradient above the Freezing Front during the Time when the Freezing Front was Passing through the Depth Interval.

Year and Thermocouple Intervals	Depth Interval (m)	Period of Freezing Front Passage	Average $\partial T/\partial z$ (°C m^{-1})	Average ∂T (°C)	E_c	E_s	E_r	E_l	V_w (%)
<u>1993</u>									
T#1 – T#2	0.5 – 1.0	12 Feb – 9 Mar	7.7	3.1	25,800	5460	16,400	14,860	9.9
T#2 – T#3	1.0 – 2.0	9 Mar – 26 Apr	15.2	7.0	97,700	1240	4480	94,460	33.3
T#3 – T#4	2.0 – 3.5	26 Apr – 18 Jul	8.7	5.6	96,500	9900	0	106,400	23.8
<u>1994</u>									
T#1 – T#2	0.5 – 1.0	10 Feb – 27 Feb	12.3	7.0	28,300	4190	23,900	8590	5.7
T#2 – T#3	1.0 – 2.0	27 Feb – 9 Apr	13.5	9.1	74,500	16,200	9350	81,350	27.3
T#3 – T#4	2.0 – 3.5	9 Apr – 20 May	9.0	9.3	49,600	16,300	0	65,900	14.7
<u>1995</u>									
T#1 – T#2	in air – 0.25	N.D. – 12 Feb	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
T#2 – T#3	0.25 – 1.25	12 Feb – 25 Mar	13.4	10.0	73,800	17,700	18,800	72,700	24.4
T#3 – T#4	1.25 – 2.75	25 Mar – 26 Apr	14.1	4.9	60,500	8540	304	69,040	15.4

N.D. = not determined because thermocouple # 1 was out of the ice during the freezing period.
E_c, E_s, E_r, and E_l units are in KJ m^{-2}.

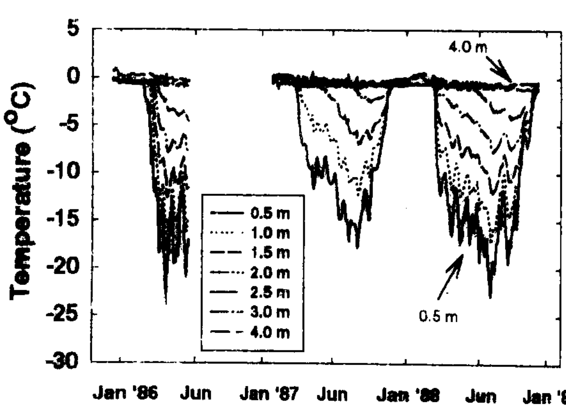

Fig. 3. Ice temperature records from 1985 to 1988 in Lake Hoare.

V_w of 24% and 15% in the lower depth intervals for years 1993 and 1995 are likely to have resulted from a portion of the liquid water pockets extending below the upper thermocouple defining the lower depth interval.

Lake Hoare

Ice temperatures in the austral summers at Lake Hoare were isothermal (at 0°C) from November 16, 1987 to February 7, 1986 (Figure 3) for a total of 111 days. Because the thermocouples were placed in the ice after the ice had become isothermal in 1985 and there was a data recording gap in 1986, the number of days of isothermy could not be determined for these years.

Timing of the freezing front progression past the first thermocouple differed each year. In 1986, the −1°C isotherm passed the first thermocouple (T#1) on February 24, whereas it took an extra month to reach T#1 in 1987 (March 25), and passed T#1 on March 8, 1988. Interpreting this year-to-year variation in the timing of the initial freezing front progression is confounded by the movement of the thermocouples relative to the surface of the ice over time.

Unlike the thermocouples at Lake Bonney, the thermocouple string at Lake Hoare was not repositioned during the second austral summer of the time-series and the records of the depths of the thermocouples are less complete. However, by analyzing the ice temperature gradients during times when the freezing fronts had reached the bottom of the ice (i.e., late winter to early spring), we were able to estimate the depth of the thermocouples when the thickness of the ice was known. Temperature gradients in the ice were computed from temperature profiles in late winter (Figure 4). The depth at which the ice temperature was 0°C (the y-intercept) was then assumed to be the thickness of the ice, Z_i. The resultant linear equation for ice temperatures,

$$Z_t = Z_i + \frac{dZ}{dT_i} T_i(Z_t)$$

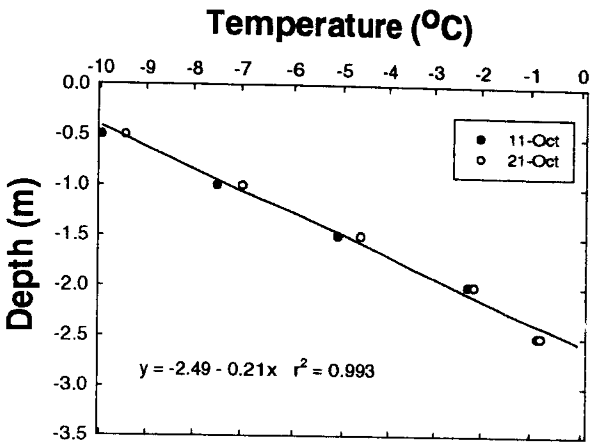

Fig. 4. Temperature profiles in the ice cover of Lake Hoare during October 1987. Linear profiles (r^2 = 0.993) are found during the late-winter/early-spring, which extend from the ice surface to the ice-water interface. Assuming the y-intercept is actually at the ice-water interface then these profiles in conjunction with known ice thickness allowed us to estimate the vertical position of the thermocouples in the ice.

was then solved for the actual depth of the thermocouples, Z_t, using the temperatures measured by the in-ice thermocouples, $T_i(Z_t)$.

The ice thickness, Z_i, at Lake Hoare was 3.2 to 3.3 m in October of 1987 [*Wharton et al.*, 1993]. The average daily temperature gradients at this time were 4.7 to 5.2°C m^{-1} and the daily average temperatures at T#1 [$T_i(Z_t)$] were –9.45 to –9.96°C. Using this information, T#1 was estimated to have moved downward from its original position of 0.5 m in 1985 to a position between 1.3 and 1.4 m in 1987. Therefore the delay in the timing of the -1°C isotherm passing to T#1 in 1987 can be explained, in part, by the thermocouples moving deeper into the ice despite of the continual upward movement of the ice (Table 3). However, the –1°C isotherm passed all the way to T#5 (2.5 m) by 29 March in 1986 and was between T#1 and T#2 (1.3 to 1.8 m) by this time in 1987, indicating that the year-to-year variation in freezing front progression did exist at Lake Hoare during the study period.

The detailed analysis of the autumnal energy budgets (Table 3) shows liquid water fractions were < 15% at all depths in the ice at the end of the 1985-1986 summer season (Table 3). In the following summers (1986-1987 and 1987-1988) liquid water contents reached 50 to 70% within the ice interior between the depths of 1.8 to 2.8 m. Therefore year-to-year variations in the freezing front progression at Lake Hoare were largely caused from variations in the liquid water fractions in the ice cover generated in the proceeding austral summers. Again, the energy budgets make no distinction as to the dispersion of the liquid water throughout the depth intervals bounded by the thermocouples.

Observations reported by *Squyres et al.* [1991] and those of *Adams et al.* [this volume] both indicate that sediments are present at depths of 1 to 2.5 m in the ice. Furthermore arching bubble morphologies indicative of pockets of liquid water were present at these depths during 1987 to 1988 [*Squyres et al.*, 1991] and in 1995 [*Adams et al.*, this volume]. These observations are similar to those of the ice at Lake Bonney and further suggest that the high liquid water content in the ice at approximately 2 m (70% by volume) is present as a concentrated layer of water pockets in association with sediments.

Sensitivity Analysis

Variance in estimates of V_w can arise from uncertainties in the following: (1) the ability of the radiative transfer model to accurately predict Q_r; (2) the true depths of the thermocouples; (3) the physical coefficients used to calculate thermal conductivities and specific heat capacities of the ice; and (4) the temperature measurements themselves. A sensitivity analysis was conducted during which coefficients and parameters in the energy budgets were individually varied over ranges that we believe encompass the natural variability in each coefficient or parameter. Table 4 lists these parameters, the range of values tested, and the change in estimates of V_w resulting from a change in each parameter. This sensitivity analysis was performed for two depth intervals on the Lake Bonney data from 1993 to illustrate the source of potential errors in the V_w calculations. The analysis was done at two depth intervals because the calculation's sensitivities are expected to differ with depth in the ice from changes in the relative proportion of E_r to the total energy budget as a function of depth and time of freezing (early autumn versus late autumn-early winter).

As expected, variations in parameters used to calculate Q_r (i.e., depth of the thermocouples, K_d (PAR), R_0 and β) produce the largest uncertainties in V_w values near the surface of the ice (0.27% to 14.6% from 0.5 to 1.0 m versus 0.3% to 2.5% from 1.0 to 2.0

TABLE 3. Energy Budget Analysis Based on the Ice Temperature Records from Lake Hoare for years 1986 to 1988. Period of Freezing Front Passage is the Time During which the Freezing Front Passed from the Top to the Bottom of the Given Depth Interval. The Average Temperature Gradient ($\partial T/\partial z$) was the Average Temperature Gradient above the Freezing Front during the Time When the Freezing Front was Passing through the Depth Interval.

Year and Thermocouple Intervals	Depth Interval (m)	Period of Freezing Front Passage	Average $\partial T/\partial z$ (°C m^{-1})	Average ∂T (C°)	E_c	E_s	E_r	E_l	V_w (%)
1986									
T # 1 – T # 2	0.5 – 1.0	23 Feb – 9 Mar	5.1	2.6	19,200	4600	1,490	22,310	14.7
T # 2 – T # 3	1.0 – 1.5	9 Mar – 14 Mar	5.8	1.4	7820	2420	168	10,072	6.8
T # 3 – T # 4	1.5 – 2.0	14 Mar – 20 Mar	3.5	2.1	5650	3630	73	9207	6.2
T # 4 – T # 5	2.0 – 2.5	20 Mar – 29 Mar	3.1	1.6	7460	2880	31	10,309	6.9
1987									
T # 1 – T # 2	1.3 – 1.8	25 Mar – 3 Apr	24.3	2.7	22,600	4700	46.7	27,253	18.3
T # 2 – T # 3	1.8 – 2.3	3 Apr – 30 May	4.8	2.5	72,900	4430	0	77,330	51.9
T # 3 – T # 4	2.3 – 2.8	30 May – 22 Aug	4.5	2.4	102,000	4350	0	106,350	71.3
T # 4 – T # 5	2.8 – 3.3	22 Aug – N.D.*	N.D.*	N.D.*	N.D.*	N.D.*	N.D.*	N.D.*	N.D.*
1988									
T # 1 – T # 2	0.4 – 0.9	7 Mar – 8 Mar	21.5	1.6	5780	2820	1380	7220	4.8
T # 2 – T # 3	0.9 – 1.4	8 Mar – 8 Mar	0.8	< 1	< 216	< 926	154	< 988	< 0.7
T # 3 – T # 4	1.4 – 1.9	8 Mar – 14 Mar	1.5	0.9	1980	1720	82	178	0.1
T # 4 – T # 5	1.9 – 2.4	14 Mar – 24 Jun	3.4	1.9	94,600	3350	88	97,862	65.6

N.D.* = not determined because the freezing front never reached thermocouple #5.
E_c, E_s, E_r, and E_l units are in KJ m^{-2}.

m, Table 4). This is because E_r comprises a larger fraction of the total energy budget during the early autumn when daily insulation is still substantial and radiation fluxes near the surface of the ice are larger. Variability in the parameters predicting Q_r have relatively little effect on the calculated values for V_w deeper in the ice (because Q_r deeper in the ice during the late autumn comprises a small fraction of the total energy budget). Other parameters creating notable variations in estimates of V_w are the changes in ice temperatures, and the temperature gradients used to predict conductive heat losses. Because the thermocouples offsets were approximately 0.1 to 0.7°C the temperature measurements are likely to have been accurate only on the order of 0.5°C. Therefore we tested the sensitivity at 1°C to estimate the higher limits to our uncertainties in V_w values. The range of V_w estimates based on this level of temperature precision ranged from 2.4% to 9.1% between 0.5 and 1 m and from 4.4% to 6.2% between 1.0 to 2.0 m. The range of uncertainties in V_w in the 0.5 to 1.0 m depth interval is relatively large when compared to the standard estimate of V_w, which was 9.9%. Whereas the range of estimates in the 1 to 2.0 m depth interval are small compared to the standard estimate of 32%.

Overall the sensitivity analysis shows that the estimates of liquid water contents are rather robust deep in the ice and are less certain in the upper regions. However the uncertainties are not large enough to obscure the trend showing that the majority of the liquid water in these ice covers was present deep in the ice at depths that coincided with the sedimentary layers and arching bubble patterns.

DISCUSSION

Depth of Liquid Water

Most of the data analyzed exhibited maximum V_w values at 2 m in the ice (Figure 5). This location coincides with the depths where sedimentary matter has been shown to be at its maximum [*Wing and Priscu*, 1993; *Adams et al.*, this volume]. The association of high liquid water fractions and arching bubble morphologies in direct association with pockets of sediments all support the notion that absorption of solar radiation by the lithogenic material is the primary process governing the generation of liquid water in these ice covers. If radiation absorption by the ice itself is the primary process generating liquid water, the

TABLE 4. Sensitivity Analysis on the Estimates of Liquid Water Fractions (V_w) in the East Lobe of Lake Bonney (1993) to a Range of Uncertainties in Variables used in the Analyses. Estimates of V_w using the Standard Values for the Variables were 9.9% at 0.5 to 1.0 m and 33.3% at 1.0 to 2.0 m.

	0.5–1.0 m			1.0–2.0 m		
Variable	Standard Values	Range of Values	Range of V_w%	Standard Values	Range of Values	Range of V_w%
Thermocouple depth	0.5 and 1.0	± 0.25	7.6 – 22.3**	1.0 and 2.0	± 0.25	31.9 – 34.3
K_d (PAR)	1.5	± 0.5	10.4 – 10.7	1.5	± 0.5	32.3 – 34.4
R_o	0.05	± 0.05	9.5 – 0.5	0.05	± 0.05	33.2 – 33.5
ß	4.2	± 0.3	9.1 – 10.7	4.2	± 0.3	33.1 – 33.6
k_s	0.2	± 0.2	9.9 – 9.9	0.2	± 0.2	33.3 – 33.3
k_a	0.024	± 0.02	9.9 – 9.9	0.024	± 0.02	33.3 – 33.4
c_s	820	± 50%	9.9 – 9.9	820	± 50%	33.3 – 33.3
c_a	1000	± 50%	9.9 – 10.0	1000	± 50%	33.3 – 33.4
ρ_s	2000	± 50%	9.9 – 9.9	2000	± 50%	33.3 – 33.3
ρ_a	1.2	± 50%	9.9 – 9.9	1.2	± 50%	33.3 – 33.3
V_g (above depth interval)	1	± 100	9.2 – 9.8	1	± 100	32.1 – 33.5
V_g (in depth interval)	1	± 100	9.9 – 10.1	21.6	± 100	33.3 – 33.5
I_d(0,PAR)	135	± 10	7.5 – 10.8	59	± 10	32.8 – 33.9
Days of Freezing	25	± 3	9.2 – 10.7	47	± 3	31.5 – 35.2
Average change in temperature	3.1	± 1	8.7 – 11.1	7.0	± 1	32.7 – 33.9
Average ΔT above depth interval	7.67	± 1	5.4 – 14.5**	7.56	±1	29.1 – 37.6

** = uncertainties greater than 50% of those estimated using the standard values

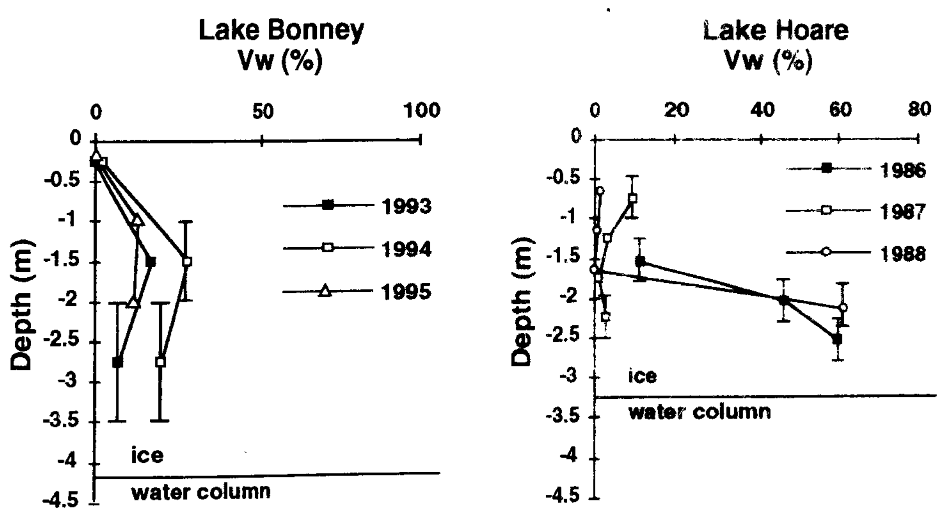

Fig. 5. Vertical profiles of liquid water content (V_w) in the ice from Lakes Bonney and Hoare. Whiskers demarcate the depth intervals overwhich the value of V_w is centered.

vertical distribution of liquid water would be expected to approximate the exponential decrease in short-wave radiation. Lake Hoare did show low V_w values exhibiting an exponential decrease during 1986 (Figure 5), implying that internal melting was not significant during this season. However in the other two years monitored at Lake Hoare and in all three years analyzed at Lake Bonney, internal melting predominated and liquid water contents were well in excess of 10% at depths associated with lithogenic material.

We have made the assertion that radiation absorption by lithogenic material is a primary process generating liquid water in the ice covers. The temperature records and the irradiance measurements before the advancement of the freezing fronts at Lake Bonney provide an independent method to assess whether or not radiation absorption by sediments in the ice can account for the high liquid water fractions at 2 m depth.

Using the previously described radiative transfer model, we have calculated the potential for liquid water generation in the ice during the summer months. For this analysis we assumed the following: (1) a simple exponential decay of I_d(PAR) to the depth of sediment inclusions; (2) that sediment inclusions absorb 100% of the energy impinging on their upper cross-sectional surface area; and (3) absorbed energy melts ice only during the period when the ice is isothermal.

Using the downwelling irradiance data at the surface of the ice at Lake Bonney [*Dana et al.*, this volume] as input for the model, we calculate that energy absorption would have been 1.2×10^8 J m^{-2} of sand cross-section during the 86 days of ice isothermy during the 1993-1994 season (Figure 2) (assuming the sediments were at 2 m). This first-order calculation translates into 0.39 m^3 H_20 (m^{-2} sand area) yr^{-1} (i.e., 39 cm yr^{-1}). Assuming the arching bubbles are remnant traces of liquid water pockets [*Adams et al.*, this volume], their dimensions are proxies for previous internal ice melt. In 1995 these arching bubbles extended 20 to 40 cm above sand inclusions [*Adams et al.*, this volume] and are assumed to represent a water pocket formed during the previous year with dimensions of 0.20 to 0.40 m^3 (m^{-2} sand area). These dimensions are on the same order as the 0.39 m^3 H_20 (m^{-2} sand area) predicted from the model. Therefore the solar radiation absorption in the ice interior by the sedimentary layers would have been sufficient to melt the pockets of liquid water within the ice interior.

Interannual Variations in Liquid Water Contents

Our previous calculations show that radiation absorption by lithogenic matter is a primary factor generating liquid water in the ice covers. It follows that annual variations in the liquid water contents should be related to factors effecting this net energy transfer. These factors include the optical properties of the ice and (if present) snow, the ice depth of the sediments, the flux of short-wave radiation, and the length of time flux of short-wave radiation, and the length of time over which absorbed radiation produces ice melt (i.e., the time over which the ice is isothermal). The length of time over which the ice is isothermal is influenced by all of the factors governing the total energy budget of the ice such as long-wave radiation, atmospheric temperatures, relative humidity, windspeed, in addition to short-wave radiation.

Differences in the liquid water contents of the Lake Hoare ice between 1986 and 1987 were likely to have been caused by interannual differences in the local weather and atmospheric conditions. During the austral summer of 1985 to 1986 the number of degree-days above freezing was 46 compared to 90 in 1986 to 1987 [*Clow et al.*, 1988]. Therefore the ice was isothermal for a longer period of time in the austral summer of 1986 to 1987, and the liquid water generation within the ice would have been greater, despite the fluxes of short-wave radiation being comparable [*Clow et al.*, 1988].

The high liquid water content at the end of the 1987 season at Lake Hoare indicates that the ice covers can reach advanced stages of deterioration. If the liquid water fractions in the upper 1.3 m were comparable to the 1986 season (i.e., 10 to 20%) we then calculate that the 3.3 m thick ice cover contained roughly 2 cubic meters of solid ice per square meter of ice cover. The ice cover had lost approximately one-third of its total mass. In order to completely melt, the ice cover would have to loose the additional two-thirds of the ice. This is unlikely to occur in one season given the current climate conditions. However two or three austral summers with a large number of degree days above freezing coupled with mild winters with lower than usual number of degree days below freezing could lead to the complete loss of the ice covers. These specific conditions which could lead to ice-out should be further explored through additional model constructions and analysis.

Ecological Considerations

Ice covers control the fluxes of energy and materials into and out of the lakes in the McMurdo Dry Valleys and, therefore, govern the energetics and biogeochemical cycles within the lakes. For example, the flux of photosynthetically active radiation through the ice is the primary path of new energy into the lakes; this flux of energy is primarily controlled by the thickness and optical properties of the ice covers [*Palmisano and Simmons*, 1987; *McKay et al.*, 1994; *Howard-Williams et al.*, this volume]. The optical properties of ice are functions of the amount and types of impurities [e.g., *Fritsen et al.*, 1992; *McKay et al.*, 1994] as well as the number, size, and shapes of vapor inclusions. Bubble morphologies and distributions are directly related to the liquid water generated in proximity to sediment inclusions [*Adams et al.*, this volume]. Liquid water may also fill bubbles (either entirely or partially) in the ice and refreeze during the winter season creating opaque disk-shaped fractures along the basal-plane of the ice crystals [*Adams et al.*, this volume]. Through these processes liquid water generation in the ice covers directly alters the distributions and morphologies of bubbles and the transparency of the ice.

The quantitative effects of the bubbles created by the process of liquid water generation and refreezing on ice optics has yet to be determined. However *McKay et al.* [1994] found it necessary to treat the ice covers as having several horizontal layers with distinct optical properties which varied independently throughout the year in order to achieve agreement between radiative transfer models and irradiance measurements. The analysis of *McKay et al.* [1994] further implies that the seasonal metamorphosis associated with melting is quantitatively important in determining variations in radiant energy fluxes through the ice covers.

The internal melting also creates liquid water habitats for microbial life co-occuring with the sediments in the ice [*Wing and Priscu*, 1993; *Fritsen and Priscu*, 1996; *Priscu and Fritsen*, 1996]. Once these microbes are exposed to liquid water they are capable of net photosynthesis and growth. New nutrients required for net growth are made available through the melting process. In 1995, inorganic nitrogen concentrations ranged from 0.61 to 4.8 mmol N m^{-3} in ice cores below the depths of the sediments in the east lobe of Lake Bonney. If one assumes that ice sediments melt approximately 0.3–0.4 m^3 m^{-2} yr^{-1} as they move downward (to balance the net annual upward ice movement) then the sediments and associated microbes encounter new nitrogenous nutrients in ice meltwater at rates of 0.18 to 1.92 mmol N m^{-2} yr^{-1} as they melt downward each summer season. The supply of nutrients to these microbes is likely to influence how fast these populations grow and accumulate.

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