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Soil temperatures and stability of ice-cemented ground in the McMurdo Dry Valleys, Antarctica

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Abstract: Year-round temperature measurements at 1600 m elevation during 1994 in the Asgard Range Antarctica, indicate that the mean annual frost point of the ice-cemented ground, 25 cm below the surface, is -21.7 ± 0.2°C and the mean annual frost point of the atmosphere is -27.5 ± 1.0°C. The corresponding mean annual temperatures are -24.9°C and -23.3°C. These results imply that there is a net flux of water vapour from the ice to the atmosphere resulting in a recession of the ice-cemented ground by about 0.4–0.6 mm yr^{-1}. The level of the ice-cemented permafrost is about 12 cm below the level of dry permafrost. The summer air temperatures would have to increase about 7°C for thawing temperatures to just reach the top of the subsurface ice. Either subsurface ice at this location is evaporating over time or there are sporadic processes that recharge the ice and maintain equilibrium over long timescales.

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Introduction

The McMurdo Dry Valleys of Antarctica comprise a large (*c.* 4000 km^2) ice-free area located across McMurdo Sound from Ross Island and the Ross Ice Shelf. The valleys are not ice covered primarily because the Transantarctic Mountains block ice from the polar plateau from flowing through the valleys. In addition, on the valley floors, the potential evaporation greatly exceeds the annual snowfall of about 1 cm yr^{-1} (Bromley 1986), producing an extremely arid environment. The mean annual temperature on the valley floor is *c.* -20°C (Thompson *et al.* 1971a, Clow *et al.* 1988) and the relative humidity averages 50% (Clow *et al.* 1988). Below a few tens of centimetres beneath the surface the ground remains perennially frozen but is dry (not cemented with ice). At even deeper layers the ground can contain a large fraction of ice resulting in solid ice-cemented permafrost (eg. Bockheim 1995). At very high elevations (2000 m and above) snow accumulates providing the source material for alpine glaciers that flow down into the valleys.

The origin and stability of the ice-cemented permafrost is of interest in understanding the geological history of the dry valleys. Sugden *et al.* (1995) found ground ice beneath a layer of volcanic ash over eight million years old. They suggested that the ice predated the ash layer and so concluded that stable polar conditions have persisted in this region for at least eight million years. Campbell & Claridge (1987) consider that the depth to the ice-cemented permafrost appears to correlate with the age of the surface, leading them to suggest that the ice is gradually evaporating into the atmosphere. Campbell *et al.* (1993, 1994) have pointed out that in areas where human activity has removed the top layer of soil the newly exposed ice-cemented ground rapidly loses water to the atmosphere.

Measurements of ground temperatures in the dry valleys have been primarily conducted on the valley floors near deep permanent lakes (Thompson *et al.* 1971b) and the thermal perturbation of the lake makes these sites unrepresentative of the valleys in general and the higher elevations in particular. Thus, in order to obtain the detailed set of ground temperatures over an entire year needed to investigate the stability of ice-cemented permafrost a set of thermistors were placed in the ground at Linnaeus Terrace, 1600 m above the floor of Wright Valley. This is a location with ice-cemented permafrost 25 cm beneath the surface. In this paper we report our data and a preliminary analysis of the stability of the ice at Linnaeus Terrace.

Methods

Site description

Our site is located on Linnaeus Terrace (77°36'S, 161°05'E, 1600–1650 m elevation) in the Asgard Range, on the north slope of Mount Oliver facing Wright Valley (Friedmann 1982). Due to a rich collection of endolithic microorganisms at this site it has received considerable attention and has been the site of extended meteorological observations (Friedmann *et al.* 1987, McKay *et al.* 1993). Although well above the valley floors, Linnaeus Terrace is not a site of net accumulation of snow.

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At the site dry soil extends down to a depth of 25 cm. Below this level the ground is cemented with ice. The soil is a loose sandy material derived from the weathering of the Beacon sandstone which forms the local bedrock.

Instrumentation

Soil temperatures were measured with Campbell 107 soil thermistors. Thermistors were placed at 0, 17, 23 (just above the ice-cemented permafrost) and 40 cm depth. To emplace the sensors, first a soil pit was dug to the level of ice-cemented permafrost. Then a hammer drill was used to make a narrow, 15 cm deep, hole into the ice-cemented ground. The thermistor was placed into the drill hole. This procedure caused minimal disturbance to the ice-cemented ground. Air temperature and humidity were measured with a Campbell 207 air probe. The temperature on the surface of an outcrop of Beacon sandstone was measured with a T-type thermocouple. Sunlight was measured with a LiCor 200 pyranometer. The sensors were deployed during January 1993 but (due to equipment failure) data storage only began 6 January 1994 and was continued until 16 January 1995.

All temperature sensors were removed after the data collection interval and calibrated in the laboratory. Calibrations were applied using the Steinhart & Hart (1968) representations for the temperature dependence of the thermistor resistance. This was necessary since the instrumental calibration for the 107 thermistor provided by Campbell did not extend below -40°C. After correction, the error in the temperature measurement is estimated to be less than ± 0.2°C and error in the humidity measurement was <10%. However, an important caveat is that the Campbell 207 RH sensor has high errors for relative humidity values below 15% – tending to systematically overestimate values by more than 10% humidity. The error in the averaged light measurement is less than ± 10 W m^{-2}. All sensors were sampled once every 10 min and averages of three measurements were written to final memory every 30 min, corresponding to 48 recordings each day.

Soil properties

Development of quantitative models of vapour transport through the soil requires knowledge of the soil properties. Samples of the soil were collected from the top 10 cm of the site but any stratigraphy in the samples was not preserved. Samples were analysed by mercury porosimetry (Diamond 1970) and in laboratory experiments in which vapour diffusion was determined. Mercury porosimetry was performed by Micromeritics (Norcross GA).

The tortuosity of the soil was determined by placing samples in a chamber at reduced pressure (18 mbar) and measuring the accumulation of ice. Water vapour was supplied by a block of ice held at -7.5°C, while the soil had a temperature gradient across it. The surface of the soil sample (2.85 cm^2) in contact with the vapour was held at -5°C, while the other side of the 1.9 cm sample was held at -10°C. After nine days the amount of ice in the sample was determined by weighing the sample.

Results

Meteorological data

Figure 1 shows a plot of the daily average values of the meteorological sensors, and Fig. 2 shows the values for the surface and underground temperatures. (The complete data are available over FTP and Internet). Table I lists the minimum, maximum, and annual average values for each sensor and the frost points for the air and each depth. Table I also lists the annual average frost point. Physically the annual average frost point can be understood by considering a large isothermal reservoir of ice in contact with the atmosphere. If the ice is extremely cold it will gain mass due to condensation from the atmosphere. Conversely if the ice is warm it will lose mass due to evaporation into the atmosphere. There would exist a temperature for the ice such that the amount of evaporation in the summer would exactly balance the condensation in the winter and there would be no net transfer of ice from the reservoir to the atmosphere. This temperature is the annual average frost point for the atmosphere.

The frost point was determined in the following way. First, the absolute humidity – the water vapour density – at every recorded data interval (30 min) was determined and averaged over the year. The annual average frost point is the temperature at which pure ice would be in equilibrium with this average water vapour density. Because the water vapour density is a nonlinear function of temperature, the frost point is not, in general, equal to the average temperature even for constant humidity. The absolute humidity of the air depends on both the air temperature and relative humidity. The humidity of the ice-cemented layers was taken as 100% – no allowance was made for the possible presence of salts.

The air temperature never rose above 0°C during the measurement interval and the average air temperature was -23.3°C. This value is close to the average in the previous

Table I. Summary of meteorological data for first 365 day interval.

Parameter	average	maximum	minimum	frost point
sunlight (W m^{-2})	114 ± 10	1063 ± 10	0	-
relative humidity (%)	50 ± 5	96 ± 5	24 ± 10	-
temperature (°C) ± 0.2 unless noted				
air temperature	-23.3	-3.8	-42.6	-27.5 ± 1.0
soil surface	-25.1	11.5	-46.0	-18.7
17 cm	-24.9	-2.4	-40.2	-21.1
23 cm	-24.9	-5.8	-38.2	-21.6
40 cm	-24.9	-9.7	-36.5	-22.3
rock surface	-23.3	8.1	-43.0	-
Interpolated values for top of ground ice				
25 cm	-24.9	-6.3	-38.0	-21.7

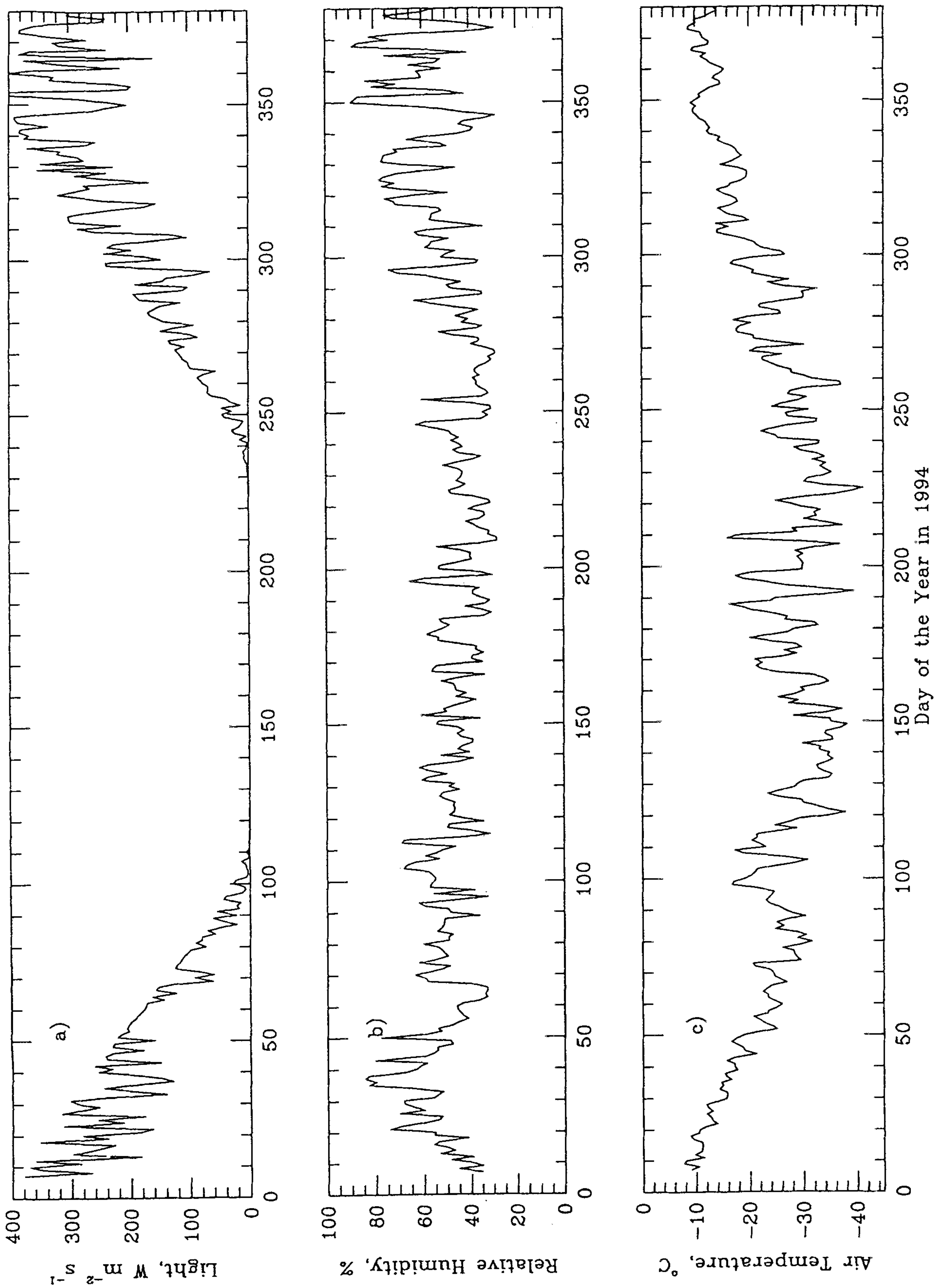

Fig. 1. Surface meteorology: **a.** daily average solar flux **b.** relative humidity, **c.** air temperature.

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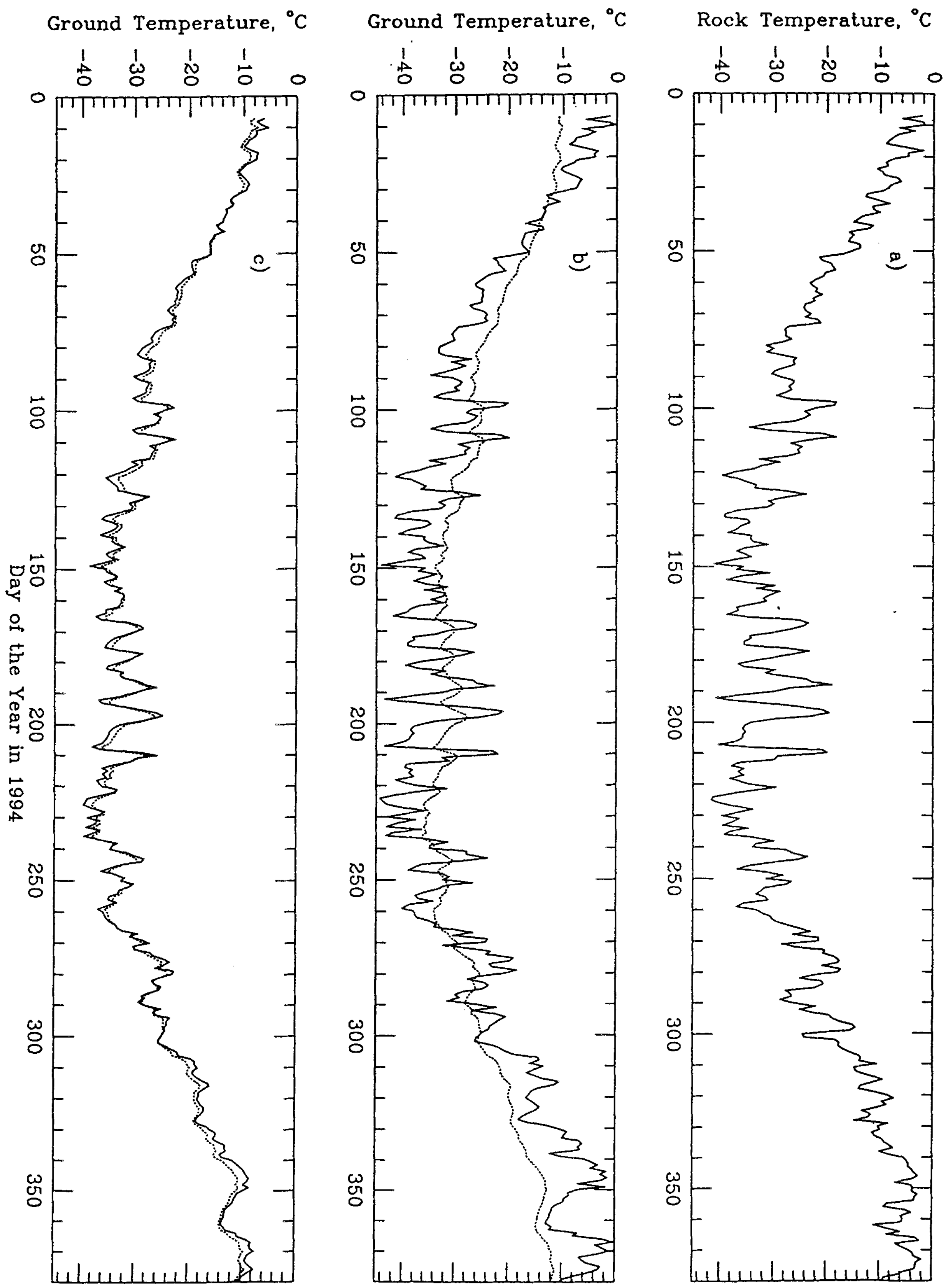

Fig. 2. Temperatures over the year: **a.** Temperature on the surface of a Beacon sandstone outcrop, **b.** on the surface of the ground (solid line), and 40 cm below the surface (dotted line), **c.** 17 cm (solid line), and 23 cm (dotted line). (The complete data are available over FTP and Internet).

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year, -23.2°C (Friedmann *et al.* 1994), and to averages reported earlier of -21 and -22°C for 1984 and 1985, respectively, using a different data logging systems (Friedmann *et al.* 1993). This suggests that 1994 was not a period of anomalous conditions. During the winter months the air temperature fluctuated over 20°C in response to downslope føhn winds from the polar plateau. A similar pattern is seen on the valley floors (Clow *et al.* 1988). The relative humidity is generally lower in the winter than in the summer, a trend also seen by Clow *et al.* (1988) on the valley floors. The minimum humidity recorded is 24.5% but the senor used has errors that are largest at the low values and so this number may be as low as 15%. There are considerable periods, during the summer, when the relative humidity is above 95%, consistent with snowing conditions. The most prominent case occurred for several days following day 350 during which the relative humidity was high and light levels were low consistent with snowfall. The surface temperature and the rock temperature can greatly exceed the air temperature (see Table I), by up to 15°C under full summer sunlight (Friedmann *et al.* 1987, Nienow *et al.* 1988, McKay *et al.* 1993). Beneath the surface the thermal variations are attenuated as expected. This is most clearly seen in the extreme temperatures reached at each depth. The maximum temperature decreased with depth by about 20°C between the surface and 40 cm while the minimum temperature increased by about 10°C over the same range. The extent of the active zone (the depth in which melting can occur since temperatures exceed 0°C) was determined to be about 12.5 cm by interpolating instantaneous temperatures between the surface and subsurface. Thus, at this site the top of the dry permafrost is about 12.5 cm above the top of the ice-cemented permafrost (25 cm).

Our error analysis is included in Table I. In general the measurements of temperature are accurate to ± 0.2°C. If this error represents a possible systematic offset it directly affects the annual averages and frost point calculations. Such an analysis represents a worst case since any random component of error should be greatly reduced in the averages. The error limits listed in Table I are likely to be overestimates of the expected error. A benefit of this conservative approach is that it makes clear that the frost point temperature of the air is significantly different from the frost point temperature in the ice-cemented permafrost. The determination of the frost point for subsurface layers depends only on temperature since the water activity is taken to be 100% at all times. For the atmosphere however the relative humidity is also used in the computation and the resulting error is therefore larger.

Soil properties

Pore size distribution (Fig. 3) was characteristically unimodal with most pores 40–160 μm in diameter. Figure 4 shows the results of the vapour diffusion experiments as well as a comparison to the expected values determined from theory (Mellon & Jakosky 1993) for a range of effective pore size and tortuosity. All other parameters in the numerical simulation matched experimental conditions. Considering these results, values of tortuosity between 1.5 and 2 can fit the data. Higher values of the tortuosity require unphysically large effective pore size.

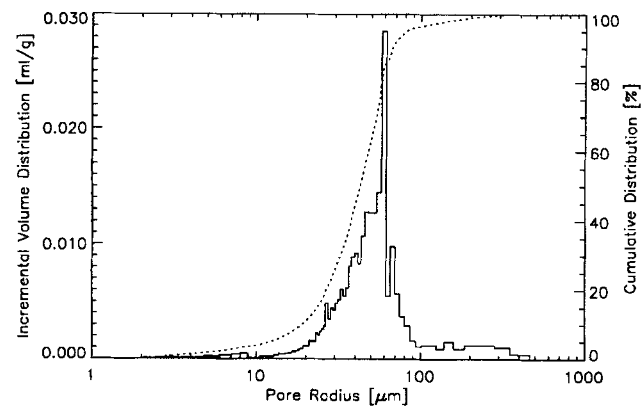

Fig. 3. Pore size distribution (solid line) and cumulative distribution (dotted) in soil determined by mercury porosimetry.

In addition, analysis of the subsurface temperature data using a standard thermal model allows the determination of the thermal diffusivity (thermal conductivity divided by the product of the bulk density and heat capacity). To match the attenuation and phase of the temperature variation with depth the thermal conductivity of the ice free region is found to be 0.6 ± 0.1 W m^{-1} K^{-1} while that of the ice-cemented ground is 2.5 ± 0.5 W m^{-1} K^{-1}.

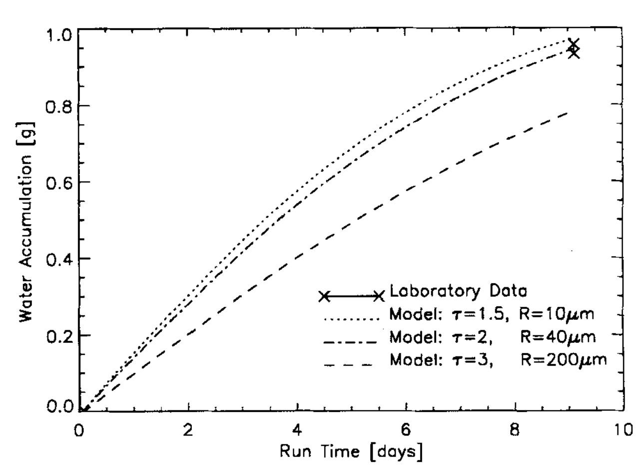

Fig. 4. Comparison of measured accumulation of ice in a soil sample (area: 2.85 cm² thickness: 1.9 cm; end temperatures held at -5 and -10°C) exposed to water vapour over ice at -7.5°C for nine days with theoretical calculations for different values of the effective pore sizes (*R*) and tortuosity (*t*).

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Mean frost point

The mean frost point as defined above is a useful indicator of the stability of subsurface ice against loss by vapour transport. The flux of vapour is proportional to the instantaneous gradient in the vapour concentration, n, with depth z,

$$F = D_e \frac{\partial n}{\partial z}, \tag{1}$$

where D_e is an effective diffusion coefficient. Considering the entire year, the net flux is approximately proportional to the difference between the annual average vapour densities, which determine the mean frost points,

$$\overline{F} = -D_e \frac{\partial \overline{n}}{\partial z}, \tag{2}$$

where the overline denotes an annual average. The relationship is exact if D_e is a constant. Thus the frost point is a direct measure of the stability of a ice reservoir. Vapour will flow from a location of high frost point (high vapour density) to a location of lower frost point (lower vapour density). The results in Table I clearly show that the subsurface ice is too warm to be stable against evaporation into the atmosphere.

Rate of vapour transport

If the ice-cemented ground is unstable with respect to the atmosphere, the rate at which ice is being lost by vapour diffusion alone, ignoring downward percolation of any liquid water, can be estimated from the diffusion equation. The basic process is molecular diffusion of atmospheric water vapour through the soil pores to and from the ice-cemented ground at 25 cm depth. The soil properties relevant to the transport calculation are summarized in Table II. At any instant the flux of vapour F will follow Fick's first law (e.g. Satterfield 1970),

$$F = -\left(\frac{\varepsilon D}{\tau}\right)\frac{\partial n}{\partial z}, \tag{3}$$

where D is the diffusion coefficient, ε is the porosity, τ is the tortuosity of the soil pores, $\delta n/\delta z$ is the gradient in the number density of water vapour. Between the atmosphere and the ice-cemented ground the flux can be roughly expressed as

$$F = -\left(\frac{\varepsilon D}{\tau}\right)\frac{n_{ice} - n_{atm}}{z_{ice}}, \tag{4}$$

where the subscripts *ice* and *atm* refer to the ice-cemented ground and the atmosphere, respectively. Just above the top of the ice-cemented ground the number density of water vapour will depend on temperature, while the atmospheric vapour density will depend on temperature and relative humidity. Since both temperature and humidity will vary diurnally and seasonally, the flux of water vapour will also vary both in magnitude and direction; water vapour will be transported toward regions where the vapour density is lowest. Subsurface ice is in equilibrium when, averaged over the year, the two densities are equal and the flux averages to zero.

Table II. Soil properties.

Parameter	Value	Note
bulk density, ρ	1.63 g cm^{-3}	mass/volume
intrinsic density, ρ_i	2.78 g cm^{-3}	particle density
porosity, ε	0.41	$(\rho_i - \rho)/\rho_i$
pore radius	95% > 12 µm	mercury porosimetry
(% of volume)	50% > 43 µm	
	5% > 83 µm	
tortuosity, τ	1.5–2	gas diffusion
thermal conductivity		thermal modelling to match data
ice-free region	0.6 W m^{-1} K^{-1}	
ice-cemented region	2.5 W m^{-1} K^{-1}	

Based on the measured data, the average atmospheric vapour density (averaging the product of relative humidity and saturation density as a function of air temperature) was found to be 1.46 x 10^{22} molecules m^{-3} (corresponding to a frost point of -27.5°C). Similarly, the subsurface vapour density at the top of the ice-cemented ground was found from the subsurface temperature data to be about 2.52 x 10^{22} molecules m^{-3} (corresponding to a frost point of -21.7°C). The higher density of water vapour at the top of the ice-cemented ground indicates that ice is out of equilibrium with the current climate and that a net loss of ice should be occurring unless other transport processes are operating.

To estimate the loss rate, the flux of water vapour between the ice-cemented ground and the atmosphere was calculated using Eq. (4) for each 30 min interval and these values were summed for the entire year. Normal diffusion coefficients for water vapour in air were determined from theory (see Mellon & Jakosky 1993); values of D range from 1.6–2.3x10^{-5} m^{-2} s^{-1} depending on the ground temperature. Knudsen diffusion was ignored because it becomes important only for pore radii smaller than 1 µm, inappropriate for the coarse grained soil at this location (see Table II and Fig. 3). The porosity was taken as ε of 41% and a tortuosity τ of 2 to 1.5, based on laboratory measurements of gas diffusion as described above. The resulting average loss rate of ice is 1.7–2.3x10^{17} water molecules m^{-2} s^{-1} respectively, and results in an annual rate of recession of the ice-cemented ground of about 0.44–0.59 mm yr^{-1}. If this were the only flux of water, it would take approximately 23 to 19 years, respectively, for the level of the ice-cemented ground (currently at 25 cm) to recede 1 cm. In simpler terms, assuming a constant diffusion coefficient of 2x10^{-5} m^{-2} s^{-1} consistent with theory and the mean values of atmospheric and subsurface water vapour densities described above, the same annual average results are obtained from Eq. (4). Note that the diffusion timescale, $\tau \simeq z^2/D_e$ for a water vapour molecule from the top of the ice-cemented ground to the atmosphere is of order several hours.

Discussion

To maintain equilibrium with the current Antarctic climate the ice-cemented ground would need to be resupplied by some mechanism other than molecular diffusion at this same rate. For example, occasional snow fall followed by melting and downward percolation might transport water directly to the level of the ice-cemented ground or near that level where molecular diffusion might more quickly complete the shortened path. The active zone currently extends to about 12 cm, requiring a depressed freezing point by the presence of salts or capillarity for liquid water to occur between 12–25 cm. The temperature in the ice-cemented ground never exceeded -6.3°C and the occurrence of snow at the surface (as a source of water) did not correspond with subsurface temperatures above freezing at any depth or temperatures in the ice-cemented ground above *c.* -12°C. At these lower temperatures liquid water transport is likely to be limited.

One interesting aspect of our data is the difference in depth between the bottom of the thawing zone (*c.* 12.5 cm) and the top of the subsurface ice (25 cm). However, the data shown here only represent one year and it is possible that the summer thaw does propagate to the top of the subsurface ice in exceptionally warm summers. In this case the absence of ice in the 12–25 cm range could be due to melting during such a warm event. The melt could then flow laterally, evaporate, or flow to greater depths if the there is pore volume available. To assess this, the temperature warming required to raise the maximum temperature at the top of the ice-cemented permafrost to 0°C has been computed. As indicated in Table I, the maximum temperature at the top of the subsurface ice (25 cm) is -6.3°C, and if the heat transport is linear then summer temperatures would have to increase by this amount to cause melting just above the ice. However, the transport of heat in ice-cemented ground is not exactly linear due to the change in conductivity with ice content. A detailed analysis of the maximum temperature as a function of the perturbation of the surface temperature is shown in Fig. 5. In this figure the abscissa, surface temperature perturbation, is a constant that has been added to the surface temperature for all values. The curves represent the maximum temperature reached at a particular depth. As can been seen in this figure an increase in surface temperature of *c.* 7°C is required for the top of ice-cemented ground to be coincident with the permafrost depth. Note that while a temperature perturbation has been added to the entire year of surface data the depth of thawing depends primarily on the summer soil temperatures which in turn are controlled by radiation.

Because the atmospheric frost point depends not just on temperature but on moisture as well, an increase in relative

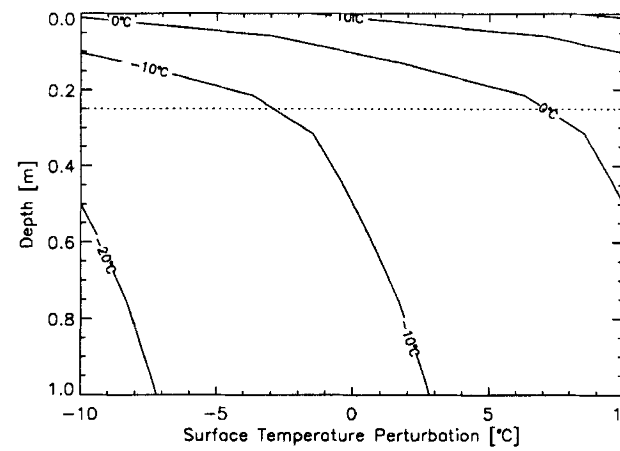

Fig. 5. Results of numerical computations of the maximum subsurface temperature reached from 0–1 m as a function of the perturbation of the surface temperature. The results are aligned with the data at 0° perturbation. The perturbation is a constant that is added to the value of the surface temperature for the entire year. The level of the ice-cemented ground is shown as a dotted line.

humidity can also stabilize the subsurface ice. Following this approach, the atmospheric frost point becomes equal to the ground frost point only when 40% is added to all humidity values (values greater than 100% are truncated). Thus, extremely moist conditions would have to prevail for the subsurface ice to be stable with respect to evaporation (i.e., evaporation and condensation balance over the year to maintain the level of ice-cemented ground at 25 cm).

Conclusions

At Linnaeus Terrace ice-cemented permafrost is located at 25 cm below the surface. Temperature data indicate that the ice-cemented ground is about 6°C too warm to be in equilibrium with the atmosphere. This implies that subsurface ice at this site, and by extension in the upper elevations of the dry valleys in general, is evaporating into the atmosphere. Furthermore numerical simulations indicate that the rate of evaporation over geological times through the loose sandstone soil at Linnaeus Terrace could have been over 0.4 mm yr^{-1} suggesting that the stability or resupply of subsurface ice has changed over timescales less than thousands of years. Extension of these results to other locations in the valleys must take into consideration differences in elevation as well as soil properties.

There are three possible explanations for our results:

1) subsurface ice is indeed evaporating over time (and the level of ice-cemented ground is receding) at this site and in the McMurdo Dry Valleys generally as suggested by Campbell & Claridge (1994).

2) There were periods of past climate when water was more plentiful and the air frost point temperature was warmer. The present subsurface ice is an evaporating relic of this past wetter epoch.

3) There are processes (such as liquid percolation into the ground at subzero temperatures) that act to recharge the ice during the present summers.

The first explanation mentioned above, that ice-cemented ground in the Dry Valleys is losing ice with time from an intial depositional period, is difficult to reconcile with the high evaporation rate and shallow depth of the ice at Linnaeus Terrace. The initial event would have been less than 500 years ago. This objection applies, albeit less stringently, to the second explanation listed above, that there was an anomalous wet epoch. It may be that there was a period of high enough air temperature and humidities within the past 500 years or so that the ice-cemented ground was recharged at that time. Given our analysis of the changes required to bring the subsurface ice in equilibrium with the atmosphere this second explanation implies significant climate change.

The third possible explanation is to assume the existence of non-vapour recharge processes under the present climate. There is in fact reason to believe that such transport processes operate in the McMurdo Dry Valleys, possibly related to salt concentration (Ugolini & Anderson 1973, Wilson 1979). The exploration of these non-vapour resupply mechanisms, however, will require a more sophisticated transport model than described here.

Note that recharging of the ice-cemented ground by an anomalous climate event as well as the percolation of liquid water though the dry upper surface both imply that the ice is of recent origin. Direct experimental tests of the age of the ice and hence its rate of exchange with the atmosphere may be possible using various tracers such as tritium.

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