Climatological Observations and Predicted Sublimation Rates at Lake Hoare, Antarctica

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ABSTRACT

In December 1985, an automated meteorological station was established at Lake Hoare in the dry valley region of Antarctica. Here, we report on the first year-round observations available for any site in Taylor Valley. This dataset augments the year-round data obtained at Lake Vanda (Wright Valley) by winter-over crews during the late 1960s and early 1970s. The mean annual solar flux at Lake Hoare was 92 W m$^{-2}$ during 1986, the mean air temperature $-17.3^\circ$C, and the mean 3-m wind speed 3.3 m s$^{-1}$. The local climate is controlled by the wind regime during the 4-month sunless winter and by seasonal and diurnal variations in the incident solar flux during the remainder of the year. Temperature increases of 20$^\circ$–30$^\circ$C are frequently observed during the winter due to strong föhn winds descending from the Polar Plateau. A model incorporating nonsteady molecular diffusion into Kolmogorov-scale eddies in the interfacial layer and similarity-theory flux-profiles in the surface sublayer is used to determine the rate of ice sublimation from the acquired meteorological data. Despite the frequent occurrence of strong winter föhns, the bulk of the annual ablation occurs during the summer due to elevated temperatures and persistent moderate winds. The annual ablation from Lake Hoare is estimated to have been $35.0 \pm 6.3$ cm for 1986.

1. Introduction

Ice is currently so prevalent on the continent of Antarctica that the few ice-free areas have become known as “oases.” These oases are mostly in low coastal areas with the exception of the mountainous dry valleys of southern Victoria Land. (For a review of the Antarctic oases, see Pickard 1986.) Covering an area of $\sim$4000 km$^2$, the “dry valleys” are the largest and most renowned of the Antarctic oases (Fig. 1). Since their discovery by Scott (1905) and the establishment of United States and New Zealand research bases on nearby Ross Island in the 1950s, the dry valleys have been studied intensively. The valleys are ice-free primarily because glacial flow from the Polar Plateau into the valleys is presently obstructed by the Transantarctic Mountains.

In addition, the potential evaporation greatly exceeds the annual snowfall, producing an extremely arid environment.

One of the most intriguing features of the dry valleys is the presence of perennially ice-covered lakes on the valley floors. Given the strong influence that climate has in determining the character of these lakes, we have initiated a program to monitor the local climate and sublimation rates. In this paper, we report on the first 2 yr of meteorological observations acquired at a site in Taylor Valley.

The dry valley lakes are of considerable interest both physically and biologically. They provide a unique environment supporting numerous benthic and planktonic microorganisms despite the harsh Antarctic climate (Parker et al. 1982; Wharton et al. 1983). Blue-green algae living in the dry valley lakes are considered to be modern analogs of the organisms that once built deep-water stromatolites during the Precambrian era (Simmons et al. 1985). Physically, it is interesting that lakes can exist at all in a region where the mean annual
surface temperature is roughly $-20^\circ$C.\textsuperscript{1} Their existence can be explained through a combination of climatic factors that may be unique to the dry valleys of Antarctica. As noted by Wilson (1982), a perennially ice-covered lake requires 1) mean summer temperatures sufficiently low that complete thawing of the ice-cover does not occur, 2) ice sublimation rates high enough to match the rate of freezing at the base of the ice-cover, and 3) peak summer temperatures above 0$^\circ$C so that meltwater derived from local glaciers (or other sources) is available to flow into the lake through a temporary moat, supplying latent heat to the lake and replacing the water lost through ablation. McKay et al. (1985) developed a simple steady state model demonstrating the sensitivity of ice-cover thickness to the sublimation rate, the surface temperature, and the distribution and amount of absorbed solar radiation in the ice.

The ice covers of the dry valley lakes display a range of thicknesses, varying from 3 to 6 m. In addition, at least some of the ice covers are known to change with time; e.g., the ice on Lakes Chad, Hoare, and Fryxell have all thinned by $\sim$2 m during the last decade. Without adequate climatological data, the reason for these variations remains unknown. Possible causes include spatial and temporal variations in surface temperature, incident solar flux, and ice sublimation rates. Ice sublimation is anticipated to occur primarily during moderate to strong winds, whose frequency and duration are no doubt quite variable within the complex terrain of the dry valleys. On a larger scale, Parish and Bromwich (1987) demonstrated that the drainage pattern of cold air off the Polar Plateau is highly irregular.

\textsuperscript{1} Geothermal heating cannot account for the existence of the dry valley lakes since the local geothermal heat flux is estimated to be only 60–80 mW m$^{-2}$ (Decker and Bucher 1977). This flux is no greater than the planetary average geothermal flux.
Since the dry valleys do not lie in a confluence zone, they are exposed to episodic katabatic surges responding to the availability of cold air on the continental interior, rather than to the strong persistent katabatic winds occurring in some coastal areas. Observed ablation rates based on ice-cover morphology do vary considerably from lake to lake within the dry valleys, ranging from 15 to 50 cm yr\(^{-1}\) (Henderson et al. 1965; Bell 1967; Calkin and Bull 1967; Lyon 1979; Chinn 1982). The sublimation rate must be at least 10 cm yr\(^{-1}\) to prevent a meltwater-fed 30-m deep Antarctic lake from freezing solid (McKay et al. 1985).

The bulk of the meteorological observations within the dry valleys have historically been made near Lake Vanda in Taylor Valley. The first detailed studies from there were reported by Bull (1966), although they were restricted to summer conditions. His observations demonstrated the effects of the comparatively warm, dry downslope winds and the cool, damp easterly winds on the local climate. Complete year-round meteorological data were later obtained by winter-over teams at Vanda Station in 1969 and 1970 (Thompson et al. 1971; Riordan 1973). They established that the mean annual temperature at Lake Vanda is \(-20^\circ C\), and that temperatures are \(-7^\circ C\) higher in summer and \(\sim 6^\circ C\) lower in winter than the adjacent ice-covered areas. Total yearly precipitation is extremely low and highly variable, 8.4 cm of snow being recorded in 1969 and only 0.7 cm in 1970. The last year-round data collected at Lake Vanda were obtained in 1974, during which 11.5 cm of snow fell (Bromley 1985). Finally, Keys (1980) has provided a detailed climatological review of the air temperature, wind, precipitation, and humidity in the McMurdo region, including Lake Vanda.

Recently, automatic weather stations have been developed for use in Antarctica, allowing year-round data to be collected at more remote sites (Stearns 1982; Savage and Stearns 1985). From December 1979 through January 1983, one such station was deployed in Upper Wright Valley on the Linnaeus Terrace at an elevation of 1000 m. Two years of recorded data from Linnaeus Terrace have been published by Hervey (1984). This station was subsequently replaced by a year-round environmental monitoring system whose results are described in Friedmann et al. (1987). No other automatic meteorological stations were deployed within the dry valleys until late 1985. In an attempt to understand the climatic variations within the dry valleys and the details of the sublimation process there, we deployed a year-round meteorological station at Lake Hoare in Taylor Valley, 35 km southeast of Lake Vanda. This station has been successfully recording temperature, wind speed and direction, relative humidity, and solar quantum flux since December 1985 and now provides the most comprehensive meteorological dataset for Taylor Valley, Antarctica. All the data necessary to determine the ice sublimation rates are being routinely recorded by the meteorological station.

2. Site description and instrumentation

Lake Hoare (77°38'S, 162°53'E) lies in a narrow section of glacially carved Taylor Valley, approximately 16 km from McMurdo Sound. The elevation of the lake is 81 m (USGS 1984), while that of the Asgard Range immediately to the north is about 2 km, as is the Polar Plateau in the vicinity of the dry valleys. Westerly winds warm adiabatically by \(\sim 20^\circ C\) while descending into Taylor Valley from the Polar Plateau and consequently have a low relative humidity. Abutting the eastern margin of the lake is the Canada Glacier, whose terminus forms a 20-m high ice cliff (Chinn 1985). Poorly sorted glacial till covers most of the valley floor near Lake Hoare, producing an aerodynamically rough surface.

The meteorological station is situated on the edge of a small peninsular kame, 300 m west of the Canada Glacier (Fig. 2). This distance is sufficiently far to avoid the effects of the ice cliff on the wind field. Due to the dynamic nature of the lake's ice cover during the summer, it was deemed necessary to locate the station on terra firma rather than directly on lake ice. The selected site offers mechanical stability and is nearly surrounded by lake ice, so that the measured meteorological parameters are as representative of those over the lake as possible. Beyond the smooth perimeter ice, the lake's surface is quite rugged, consisting of a series of troughs and ridges generally aligned along the valley's axis. Topographic relief of the ridge/trough system is typically 1 m.

The meteorological station consists of a 21X Micrologger\(^2\) housed in a thermally insulated box, a 3-m aluminum mast, and various sensors. Sufficient power is supplied by two 12 volt, 40 amp-hr gel batteries\(^3\) so that recharging is unnecessary for 400 days of continuous operation. Temperatures are detected to within \(\pm 1.0^\circ C\) using copper-constantan thermocouple junctions enclosed in identical radiation shields on the lake ice, and at the 1.0 and 3.0 m heights on the meteorological mast. A 3-cup anemometer (Met-One model 014A)\(^4\) measures the wind speed at 3.0 m with an accuracy of \(\pm 1.5\%\) over its calibrated range (0–50 m s\(^{-1}\)). A simple vane-type wind direction sensor (Met-One model 024A) is mounted at the 1.0 m height. The manufacturer lists an accuracy of \(\pm 5^\circ\) for this sensor. Relative humidity is measured at 3.0 m using a temperature-corrected impedance-type probe (Campbell Scientific model 207). The accuracy of the probe's thermistor is \(\pm 1^\circ C\) for temperatures greater than \(-40^\circ C\) while that of the humidity sensing element is \(\pm 3\%\) (at 25°C) over the humidity range 12%-100%. For the humidity range 25%-94%, the accuracy of the humidity element is reported to be \(\pm 1\%\) at the cali-

\(^2\) Campbell Scientific, Logan, Utah.
\(^3\) Power Sonic, Redwood City, California.
\(^4\) Grants Pass, Oregon.
bration temperature, 25°C. A cosine-corrected silicon diode (LICOR model 190SA)\textsuperscript{3} is used to detect the number of solar photons (direct plus diffuse) in the 400–700 nm (PAR; photosynthetically active radiation) spectral band. This diode was calibrated under Antarctic atmospheric conditions using a precision double dome pyranometer,\textsuperscript{6} whose spectral response ranges from 285 to 2800 nm. The calibration tests were performed within 1°C of the mean-annual temperature at Lake Hoare to minimize the uncertainty associated with the temperature-dependent photoelectric response of silicon (\(\pm 0.15\%\) per °C relative to the calibration temperature). The resulting conversion factor for PAR to solar flux is 0.505 W m\(^{-2}\) per 1 μE m\(^{-2}\) s\(^{-1}\) at Lake Hoare, which is slightly higher than the value determined by Stefan et al. (1983) in Minnesota (0.488 W m\(^{-2}\) per 1 μE m\(^{-2}\) s\(^{-1}\)). We attribute this difference to a greater proportion of near-IR solar radiation in Antarctica due to reduced attenuation by atmospheric water vapor.

Each probe is sampled every 30 sec by the micrologger, which then computes the water vapor density at the 3-m height and the bulk Richardson number. The capacity of the solid state memory module (Campbell Scientific model SM64) and the logistical constraint that the module can only be retrieved once a year, currently restrict the storage of data to 6-h averages. Our averaging periods begin at 0000 LST each day. Mean solar time at Lake Hoare follows local standard time by nearly 1 h.

3. Meteorological observations

The principal climatic factors for Lake Hoare are summarized as monthly and annual averages in Tables 1 and 2 for the period December 1985 through November 1987, while the basic data for this 700-day period are presented in Figs. 3–7. It should be noted that interannual climatic variations in the dry valleys can be quite large. For example, at Lake Vanda the mean annual temperature changed by 3.6°C between 1969 and 1970, the mean solar flux by 13%, the mean wind speed by 26%, and the mean relative humidity by 20% (Thompson et al. 1971). Similar interannual variations

\textsuperscript{3} Lincoln, Nebraska.

\textsuperscript{6} Produced by the Eppley Laboratory, Newport, Rhode Island.
are expected at Lake Hoare, precluding a detailed comparison with other dry valley sites until year-round data are simultaneously collected at several locations. The last year-round meteorological experiment at Vanda was conducted in 1974.

The measured mean-annual quantum irradiance in the PAR waveband was 188 μE m⁻² s⁻¹ at Lake Hoare during 1986. Using the PAR/ pyranometer calibration data, the equivalent solar flux was 92 W m⁻². By comparison, the incident direct-beam solar flux at this site, neglecting atmospheric attenuation, would be 144.8 W m⁻² on an annual average. The reported solar flux at nearby Lake Vanda was 104 W m⁻² during 1970 (Thompson et al. 1971), which is consistent with the fact that Lake Vanda lies in a wider valley. Clouds are frequently present at Lake Hoare, as can be seen from the variability of daily-average solar flux (Fig. 3).

Two distinct regimes are apparent in the daily average temperature data (Fig. 4). As will be discussed further in subsequent sections, the summer climate in the dry valleys is dominated by the presence of solar radiation, while the winter regime is controlled by the wind. The mean-annual temperature at Lake Hoare was −17.3°C for 1986, 2.7°C warmer than the 1969–70 average at Lake Vanda (see Thompson et al. 1971). Summertime temperatures generally range from −10° to +5°C while those in the winter show much greater extremes, ranging from −40° to −8°C. The maximum and minimum instantaneous temperatures were +14.9°C (15 January 1987) and −43.1°C (9 August 1986), respectively. Wind speeds at Lake Hoare averaged 3.3 m s⁻¹ at the 3.0 m height. Extrapolation of these speeds to the height of the Lake Vanda wind measurements (~8 m), assuming the lower atmosphere is nearly isothermal during strong winds, yields a value of 4.5 m s⁻¹. This value is within the observed inter-

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* Incomplete month: mean-values based on 1–10 November.
annular variations at Lake Vanda (4.3–5.2 m s⁻¹; Thompson et al. 1971). The peak 6-h average wind speed during 1986–87 was 20 m s⁻¹ (Fig. 5). Wind direction constancy, defined by the magnitude of the resultant wind vector divided by the wind speed, is relatively high at Lake Hoare except during the spring and fall transition periods (Tables 1 and 2). This constancy reflects the channeling of winds by the east/west trending ridge-valley system. Six-hour average relative humidities vary between 30% and 80% throughout the year (Fig. 6) with an annual average of 55%. For comparison, the mean relative humidity was 64% at Vanda in 1969 and 44% in 1970. Finally, the transition between predominantly positive and negative atmospheric stability as indicated by the bulk Richardson number $B$, is surprisingly abrupt, occurring at a time when the sun is almost always above the horizon (Fig. 7). The transition in the thermal behavior at mast height (Fig. 4) coincides with the switch in atmospheric stability.

a. Winter

In the complete absence of sunlight, the climate at Lake Hoare is controlled by the wind. Two distinct wind patterns occur during the winter (Fig. 8a). The first is a local phenomenon, consisting of gentle (∼1 m s⁻¹) winds draining off the steep flank of the Asgard Range at azimuths between −45° and 25° relative to true north. These winds may originate in a fashion similar to the katabatic winds of the Polar Plateau. If so, they are due to the intense radiative cooling of the ground during the winter and subsequent downslope movement of cold near-surface air under the influence of gravity. Atmospheric stability on the valley floor ranges from slightly unstable to extremely stable during these periods. Relative humidities are typically 55%–80%. Westerly föhn winds rushing down Taylor Valley at speeds up to 20 m s⁻¹ on a 6-h average, constitute the second wind pattern. These winds heat adiabatically ∼20°C upon their descent from the Polar Plateau. During similar episodes at Lake Vanda (Wright Valley), relative humidities are frequently observed to fall to 6%–10% (Bromley 1985). However, air descending into Taylor Valley from the Polar Plateau must flow over an 85 km stretch of the Taylor Glacier. Rapid sublimation from the glacier’s surface prevents the relative humidity of the descending air from dropping below 30% during föhns (Fig. 9a). Wind directions typically become erratic at the onset and demise of a föhn wind,
Fig. 7. The micrologger calculates the bulk Richardson number every 30 sec. This plot displays the recorded 6-h average values. The transition from predominantly stable to unstable atmospheric conditions occurs when the sun is above the horizon for nearly 24 h per day.

accounting for the winds with azimuths beyond 25° in Fig. 8a. Some of the light westerly winds are also of a transitional nature. During periods of light winds,

the surface cools radiatively to temperatures between −35° and −40°C before another föhn causes rapid warming. Figure 10 clearly shows the strong correlations between wind speed, wind direction, temperature, and relative humidity for a typical 2-week winter period.

Föhn winds stronger than 5.0 m s⁻¹ occurred 33% of the time during the winter of 1986, whereas winds draining off the Asgard Range at speeds less than 1.5 m s⁻¹ occurred 44% of the time (Fig. 11). The föhn winds in Taylor Valley are probably instigated by katabatic surges from the Polar Plateau. It has been suggested that such surges result from quasi-periodic discharging and recharging of the cold air supply on the continental interior (Lettau and Schwerdtfeger 1967; Parish and Bromwich 1987). However, the autocorrelation of the wind-speed time-series at Lake Hoare fails to show any statistically significant cyclicities. This may be due to the complex nature of the surface wind patterns and may still be consistent with the recharge/discharge mechanism.
monthly air temperatures are typically $-1^\circ$ to $0^\circ$C during the warmest months, December and January (Table 1). Periods of diminished winds occasionally allow surface temperatures to exceed the melting point of water as the sensible heat flux from the surface to the atmosphere is reduced. The amount of meltwater derived each summer from local glaciers, which provide the inflow to the dry valley lakes, is very sensitive to the duration of these periods. Furthermore, our two summers of meteorological observations indicate that the conditions required for temperatures in excess of $0^\circ$C are sufficiently sporadic that large interannular variations occur. For the 1986/1987 austral summer, the number of degree-days above $0^\circ$C was 89.9, whereas it had been only 45.7 the previous summer. The degree of this variability is corroborated by the observation that large variations occur in the discharge rates of the glacially fed streams, such as the Onyx River (Chinn 1982).

Surface heating in the dry valleys during the summer reverses the wind regime to one of primarily upvalley flow (azimuths between $60^\circ$ and $100^\circ$; Fig. 8b). Summer winds intensify each afternoon (Fig. 12) and are almost exclusively from the east. Downvalley winds (westerlies), when they occur, tend to be a morning phenomena (Fig. 13), since the atmosphere is more stable due to reduced solar heating at this time. A similar diurnal wind pattern is observed at Lake Vanda (Riordan 1973). The generally unstable atmospheric conditions strongly suppress gravitationally driven downslope winds from the Asgard Range. Unlike in winter, the summer wind speeds have a unimodal distribution, with 90% of the winds occurring between speeds of 1.2 and 5.9 m s$^{-1}$ (Fig. 11).

Diurnal changes in the solar elevation angle are sufficient to cause a factor-of-3 variation in the incident solar flux even at summer solstice, driving diurnal temperature oscillations of about $4^\circ$C. Diurnal fluctuations are also apparent in nearly all the other observed meteorological parameters (Fig. 12).

4. Sublimation rates

To determine the ice sublimation rate from the observed meteorological quantities, we consider the integrated form of the vertical water vapor mass flux

$$E = \frac{\rho_v(0) - \rho_v(Z)}{I_v(0, Z)},$$  \hspace{1cm} (1)

where height $Z$ is sufficiently close to the surface that the flux is essentially the same as at $z = 0$, i.e., there are no sources or sinks of water vapor between $z = 0$ and $z = Z$. The function $\rho_v(z)$ is the water vapor density, $I_v$ is a resistance coefficient defined by

$$I_v(0, Z) = \int_0^Z dz \frac{z}{K_v},$$  \hspace{1cm} (2)

and $K_v$ is the generalized height-dependent water vapor...
diffusivity (eddy plus molecular). We break the resistance coefficient into two parts,

$$I_d(0, Z) = I_d(0, z_i) + I_d(z_i, Z), \quad (3)$$
evaluating across the interfacial layer in immediate contact with the surface and, separately, across the overlying segment within the fully turbulent surface sublayer. The transition height $z_i$ between these two layers depends on the nature of the surface and on the wind regime. Lake Hoare’s ice cover can be considered aerodynamically rough since its roughness Reynold’s number $z_0 = (U_0^2/v)$ is on the order of 5000, while the smooth to rough transition occurs at $z_0 \sim 1, z_0$ being the surface roughness length and $v$ being the kinematic viscosity. For such a surface, the top of the interfacial layer occurs at the elevation where the ratio of the wind velocity $U(z)$ to the friction velocity $U_0$ is 5 (Brutsaert 1975). This condition is met at $z_i \approx (5/3) h_0$, not far above the mean height $h_0$ of the surface’s roughness elements.

In the immediate vicinity of the surface, i.e., the interfacial layer, airflow is observed to occur through an intermittent bursting process with well-defined inrush and ejection phases (Grass 1971; Kim et al. 1971; Corino and Brodkey 1969). In accordance with these observations, Brutsaert (1965, 1975) has developed a model for heat and water vapor transfer within the interfacial layer involving nonsteady molecular diffusion into Kolmogorov-scale eddies penetrating into the interfacial layer from the surface sublayer. These eddies remain in contact with the surface for a random amount of time before being swept away by fresh eddies from above. Within the overlying surface sublayer, airflow is fully turbulent, so that eddy transfer processes dominate molecular diffusion. Here, the water vapor diffusivity can be defined in terms of the friction velocity and stability-dependent dimensionless gradients. Both components of the resistance coefficient $I_d(0, Z)$ are described in greater detail in the Appendix.

Summarizing our procedure for determining the sublimation rate, the water vapor density $\rho_w(Z)$ is calculated every 30 sec by the micrologger from the measured temperature and relative humidity at height $Z = 3 \text{ m}$ and stored as 6-h averages. At the ice surface, the relative humidity is assumed to be 100% so that the vapor density $\rho_w(0)$ can be found from the measured temperature alone. Unfortunately, the thermocouple on the ice surface developed an intermittent fault beginning on 6 November 1986. During these intermittent periods, $z = 1.0 \text{ m}$ temperatures are substituted for ice-surface temperatures to find $\rho_w(0)$. Next, the measured temperatures and wind speeds are used to establish the degree of atmospheric stability via the bulk Richardson number, the friction velocity is found, and the resistance coefficient $I_d$ calculated (see the Appendix). Finally, the sublimation rate $E$ is found from (1).

Figure 14 shows the calculated daily average sublimation at Lake Hoare and its rapid response to wind speed fluctuations. As expected, the majority of the ablation occurs during the summer due to the strong dependence of the saturation vapor pressure on temperature and the persistence of the winds. Nevertheless, 1–3 mm per day are calculated to ablate during the winter föhn winds, causing a total ablation of 9.6 cm
during the winter period, 1 April–15 October. Direct measurements of the level of the lake confirm that ~9 cm of ice sublimed from Lake Hoare during the 1986 austral winter. For a comparable period in 1974, 6.3 cm were observed to sublime from Lake Vanda with 0.8 mm d⁻¹ ablation during gale-force westerly winds (Bromley 1985). Present instrumentation makes it difficult to accurately measure relative humidities during the calm cold winter periods. However, this is of little importance in determining seasonal or annual ablation rates since the combination of light winds and low vapor pressures leads to very low sublimation rates during these times.

Integrating the estimated sublimation rates over the year, the annual ablation at Lake Hoare is calculated to be 35.0 cm for 1986. This figure is surprisingly robust. As has already been mentioned, the 1986/1987 austral summer was distinctly warmer than the previous summer, having twice as many degree-days above 0°C. Yet the integrated ablation over all 365-day windows within the period spanning both summers, ranges from 33.9 to 35.9 cm. Thus, unlike meltwater generation, the bulk of the annual ablation is associated with the seasonal rise in temperature during the summer rather than peak daily temperatures. The dominant sources of uncertainty in the calculated ablation are due to 1) a potential ±2.0°C uncertainty in the ice temperature (primarily due to the substitution of 1-m temperatures for ice temperatures when the ice thermocouple failed), leading to a 16.5% error in ablation, 2) a ±5% error in the relative humidity at Z = 3 m due to combined reference thermistor and sensing ele-

Fig. 12. Diurnal variations in the solar incidence angle during the summer are sufficient to cause corresponding diurnal changes in temperature, relative humidity, bulk Richardson number, and wind speed. This figure shows the data for a typical 15-day summer period.
FIG. 13. Wind run (per day) as a function of wind direction and time of day during the summer. Easterly winds intensify during the afternoon and early evening, while westerlies are more likely to occur in the morning.

FIG. 14. Calculated sublimation rate at Lake Hoare. The bulk of the ablation occurs during the summer despite the intensity and dryness of winter föhns.

ment errors (3.3% error in ablation), and 3) uncertainty in the surface roughness length z₀. Varying the ratio (h₀/z₀) from 5 to 20 changes the calculated sublimation by ±6.2%. Uncertainties introduced by wind speed errors and error in the measured temperature contrast along the meteorological mast are both less than 1.5%. These uncertainties lead to a 6.3 cm uncertainty in ablation.

5. Conclusions

Based on the limited data available thus far, the climate at Lake Hoare (Taylor Valley) appears to be similar to that of the most intensely studied site within the dry valleys, Lake Vanda (Wright Valley). The mean annual wind speed (3.3 m s⁻¹ at 3 m) and the relative humidity (54%) at Lake Hoare for 1986 are all within the interannual variations observed at Vanda Station during 1969, 1970, and 1974. Temperatures at the two sites are also quite similar except for the sunless winter, during which temperatures average ~8°C higher at Lake Hoare due to a higher frequency of föhn winds. Relative humidities in Taylor Valley do not drop as low as they do at Lake Vanda during föhns due to the mitigating influence of the Taylor Glacier. The mean annual solar flux at Lake Hoare for 1986 (92 W m⁻²) was 12 W m⁻² lower than the 1969 average at Lake Vanda. This is primarily due to more extensive shadowing by the surrounding ridges at Lake Hoare, although differences in cloud frequency may also contribute to the discrepancy. Again, interannual variations in the dry valleys are known to be quite large, precluding a detailed comparison between Lake Hoare and Lake Vanda until year-round measurements are made concurrently at both sites.

As at Lake Vanda, the winter climate at Lake Hoare is controlled by the wind, and cold calm conditions are repeatedly interrupted by strong föhn winds originating on the Polar Plateau. Föhn winds with wind speeds greater than 5.0 m s⁻¹ occur 33% of the time during the winter. Air temperatures rapidly increase 20°–30°C during these episodes while the relative humidity drops by 20%–30% over the lake. Due to the low albedo, the summer climate within the dry valleys is dominated by radiative heating and the associated instability of the atmospheric boundary layer. The wind regime changes to one of persistent moderate upvalley winds,
intensifying each afternoon. Westerly föhn winds are able to penetrate into Taylor Valley only during the morning, a time when atmospheric instability is reduced due to diminished radiative heating rates. Temperatures generally approach 0°C during December and January, allowing some melting of lake and glacial ice.

The annual ablation from Lake Hoare is estimated to have been 35.0 ± 6.3 cm for 1986. This meteorologically based figure is consistent with that determined from ice-cover morphology at nearby Lake Fryxell during the early 1960s (30–40 cm yr⁻¹; Henderson et al. 1965). Of the annual ablation, only 9.6 cm sublimated during the intense winter föhns at Lake Hoare during 1986. Despite the sensitivity of the instantaneous sublimation rate to wind speed and temperature fluctuations, the annual ablation rate is primarily determined by the mean summer temperature and wind speed. In contrast, the amount of meltwater derived from local glaciers is highly dependent on the number of degree days above 0°C, which is quite variable from year to year. Since glacial meltwater provides the only inflow and ablation the only outflow from a dry valley lake, and since mean summer temperatures and wind speeds are unlikely to fluctuate as much as the occurrence of melting temperatures, yearly changes in lake levels reflect variations in inflow rather than ablation. Given the current thickness of ice on Lake Hoare (3.3 m), the estimated sublimation rate implies that the turnover time-constant for the ice-cover is ~10 years.

Note: The Lake Hoare meteorological dataset is available from the authors on computer disks or tape at the cost of the magnetic medium.

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APPENDIX

Water Vapor Diffusion Resistance Coefficients

Within the surface sublayer, airflow is fully turbulent so that eddy transfer processes dominate molecular diffusion. In this case, the generalized water vapor diffusivity reduces to the eddy diffusivity,

\[ K_w = \frac{k u_* (z - d_0)}{\phi_w (\xi)} \]  

(A1)

where \( k \) is von Karman’s constant, \( u_* \) the friction velocity, \( \xi = (z - d_0)/L \) is the Monin-Obukhov stability parameter, and \( d_0 \) is the zero-plane displacement height, which is approximately \( 2/3 h_0 \) (Brutsaert 1982).

The function \( \phi_w \) is a dimensionless vapor density gradient defined analogously to the well-known dimensionless temperature gradient \( \phi_h \) used to describe similarity-theory based flux-profiles for this layer:

\[ \phi_w (\xi) = \frac{k (z - d_0) \partial \rho_w}{\rho_w} \frac{\partial z}{\partial z} \]  

(A2)

and

\[ \phi_h (\xi) = \frac{k (z - d_0) \partial \theta}{\theta} \frac{\partial z}{\partial z} \]  

(A3)

Here \( \rho_w = (-E/u_*) \) and \( \theta = -H/(\rho C_p u_*) \) define a scaling water vapor density and scaling temperature, respectively, with \( H \) representing the sensible heat flux, \( \theta \) the potential temperature, \( \rho \) the air density, and \( C_p \) the heat capacity of air. It is generally assumed that \( \phi_w \) equals \( \phi_h \) and some experimental data indeed support the notion that this is at least approximately true (see Crawford 1965; and Dyer 1967). The advantage of this strategy is that parameterizations that have been successfully developed for \( \phi_h \), such as

\[ \phi_h (\xi) = \begin{cases} \alpha + \beta \xi, & (\xi \geq 0) \\ \alpha (1 - \gamma_\lambda \xi)^{-1/2}, & (\xi \leq 0) \end{cases} \]  

(Businger et al. 1971; Dyer and Hicks 1970), can then be used to evaluate the eddy diffusivity \( (A1) \) and subsequently the surface sublayer resistance coefficient

\[ I_w (z_i, Z) = \begin{cases} \frac{1}{k u_*} [\alpha \ln (\xi_i/\xi) + \beta (\xi_i - \xi)], & (\xi \geq 0) \\ \frac{1}{k u_*} \alpha \ln (\xi_i/\xi) + 2\alpha \ln \left( \frac{\lambda_i + 1}{\lambda_i + 1} \right), & (\xi \leq 0) \end{cases} \]  

(A5)

(see Benoit 1977; and Wang et al. 1978). In this expression, subscripts \( Z \) and \( i \) refer to evaluations at heights \( z = Z \) and \( z = z_i \) respectively, and \( \lambda = (1 - \gamma_\lambda \xi)^{1/2} \).

Two quantities, the friction velocity \( u_* \) and the Monin-Obukhov stability parameter \( \xi \), remain to be specified to solve for \( I_w (z_i, Z) \). The friction velocity is a function of the measured wind speed at height \( Z \) and a function \( F_m \), which is essentially a resistance coefficient for momentum,

\[ u_* = \frac{U(Z)}{F_m (z_0, Z)}. \]  

(A6)

For unstable atmospheric conditions, we utilize the numerically well-behaved form of \( F_m \) developed by Benoit (1977), while the expression given by Wang et al. (1978) is used for stable conditions,
Here, $\xi = (1 - \gamma_m)^{1/4}$ and the subscripts $Z$ and 0 refer to evaluations at heights $Z$ and $z_0$, respectively. The surface roughness $z_0$ of Lake Hoare's ice cover is estimated to be 15 cm, based on the empirical relationship ($K_h/z_0$) $\approx$ 7.35 (Brutsaert 1975). We use the Businger et al. (1971) parameterization to establish values for the constants $\alpha$, $\beta$, $\gamma_h$, $\gamma_m$, and $k$ appearing in (A1) through (A7).

The Monin-Obukhov stability parameter $\zeta$ can be related to a quantity known as the bulk Richardson number $B$, which is based on the measured field parameters $\theta$ and $U$,

$$B = \frac{gZ(\theta_x - \theta_z)}{\theta U^2},$$

where $g$ is the acceleration of gravity. Sensor locations at the Lake Hoare site are such that $z_0 = 1.0$ m and $Z = 3.0$ m. Beginning with the expression

$$B = \frac{\zeta}{k_u} \left( \frac{1}{z_0} \right) \left( \frac{1}{K_m} \right),$$

it can be shown for stable atmospheric conditions that $\zeta$ is given by the negative root of the quadratic equation

$$[B\beta^2 \left( 1 - \frac{Z}{Z_0} \right)^2 - \beta \left( 1 - \frac{Z}{Z_0} \right)]^2 + \left[ 2B\beta^2 \left( 1 - \frac{Z}{Z_0} \right) - \alpha \eta \right] \frac{Z}{Z_0} + \beta \eta^2 = 0,$$

with $\psi = \ln(Z/z_0)$ and $\eta = \ln(Z/z_0)$. Here $K_u$ and $K_m$ are eddy diffusivities for heat and momentum, defined analogously to $K_w$ (A1). For unstable conditions, we use the series solution relating $\zeta$ and $B$ developed by Wang (1981).

Brutsaert (1975) developed a model describing evaporation from both smooth and rough surfaces using nonsteady molecular diffusion into Kolmogorov-scale eddies penetrating into the interfacial layer. Reformulating his expression for the aerodynamically rough case, the interfacial resistance coefficient may be written

$$I_w(0, z_0) = \frac{(v Z_0)^{1/4}}{C_R \kappa_w^{1/2} \bar{u}_w^{1/4}},$$

where $C_R$ is an empirical constant that evaluates to $(7.3)^{-1}$ and $\kappa_w$ is the temperature-dependent molecular diffusivity of water vapor in air,

$$\kappa_w = \kappa_w(\frac{T}{T_0})^2.$$

The value of $\kappa_w$ at the reference temperature $T_0 = 273.15$ K is $1.37 \times 10^{-5}$ m$^2$ s$^{-1}$ (Washburn 1929).

References


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