Debris-Covered Glaciers (Proceedings of a workshop held at Seattle, Washington, USA, September 2000).
IAHS Publ. no. 264, 2000.

Summer temperature profiles within supraglacial debris on Khumbu Glacier, Nepal

H. CONWAY & L. A. RASMUSSEN
Geophysics Program, University of Washington, Box 351650, Seattle, Washington 98195, USA
e-mail: conway@geophys.washington.edu; lar@geophys.washington.edu

Abstract Temperature measurements made during summer within supraglacial debris on Khumbu Glacier, Nepal show a strong diurnal signal that diffused downward into the debris with decreasing amplitude and increasing lag. Surface temperatures during the day were up to 35°C higher than the air temperature; energy transfer into the debris was dominated by the solar radiative flux. Temperature profiles through the debris indicate that heat flow deeper than about 0.2 m was primarily by conduction. The thermal conductivity k of the debris, estimated from a calculated thermal diffusivity and a representative volumetric heat capacity, was 0.85 ± 0.20 W m^{-1} K^{-1} at one site and 1.28 ± 0.15 W m^{-1} K^{-1} at another. At the first site the debris was 0.40 m thick and the average temperature gradient $\partial\overline{T}/\partial z$ = 19 K m^{-1}; the average flux of energy through the debris was sufficient to melt 4–6 mm of ice per day. The debris was thicker (estimated to be 2.5 m) and the temperature gradient lower (4.5 K m^{-1}) at the second site, and the calculated ice-melt was less than 2 mm day^{-1}.

INTRODUCTION

Khumbu Glacier, which flows from the western cwm below Mt Everest (Fig. 1), is typical of the debris-covered glaciers of Nepal. Debris thickness increases from zero just above Everest Base Camp to more than 2 m near the terminus (Nakawo *et al.*, 1986). Ablation rates of ~30 mm day^{-1} have been measured on clean ice near Base Camp, and numerous studies have shown that ice-melt is accelerated beneath a thin layer of supraglacial debris, but inhibited by a thick layer (Fujii, 1977; Inoue & Yoshida, 1980; Kayastha *et al.*, 2000). The critical thickness at which ablation is the same as for clean ice depends on the optical and thermal properties of the debris as well as the prevailing meteorological conditions (Nakawo & Takahashi, 1982; Conway *et al.*, 1996; Adhikary *et al.*, 1997). The lithology as well as the thickness of the debris can cause variations in ice-melt; for the same thickness, ablation beneath dark-coloured schist is generally higher than beneath lighter-coloured granite because of differences in absorbed solar radiation (Inoue & Yoshida, 1980).

The energy flux through a debris layer can be modelled by the one-dimensional diffusion equation:

$$\rho c\frac{\partial T}{\partial t}=\frac{\partial}{\partial z}\left[k\frac{\partial T}{\partial z}\right]+\frac{\partial Q_N}{\partial t} \quad (1)$$

where ρ is the bulk density of the debris, t is time and z is the vertical coordinate, positive downward. Here $k\partial T/\partial z$ is the downward conductive flux, and $\partial Q_N/\partial t$

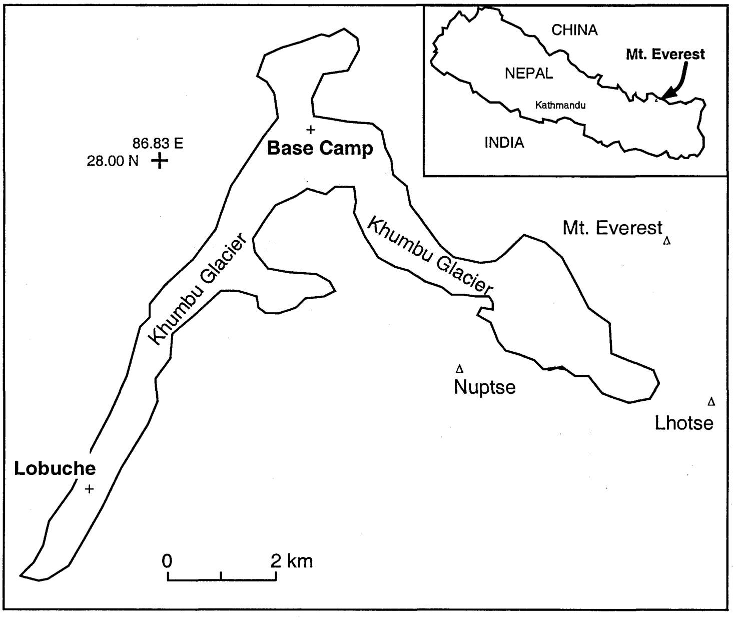

Fig. 1 Locations of study sites near Everest Base Camp (5360 m) and Lobuche (4960 m) on Khumbu Glacier, Nepal.

accounts for nonconductive processes such as convective or latent heat exchange. Thermal conductivity k and volumetric heat capacity ρc of the debris are functions of temperature T, porosity and moisture content (Hallet & Rasmussen, 1993). If the system is purely conductive ($\partial Q_N/\partial t = 0$) and k is constant with depth, then equation (1) can be written:

$$\dot{T} = \kappa T'' \tag{2}$$

in which $\kappa = k/\rho c$ is the thermal diffusivity and the derivatives are denoted by $\dot{T} = \partial T/\partial t$ and $T'' = \partial^2 T/\partial z^2$.

The terminus position of Khumbu Glacier has been relatively stable since the last major advance about 150 years ago but lower sections of the glacier have thinned by more than 70 m (Mayewski & Jeschke, 1979). About 50 m of ice still exists beneath the debris at Lobuche (Gades *et al.*, 2000) and the glacier there is stagnant (Kodama & Mae, 1976). Here we discuss temperature profiles measured within supraglacial debris at two sites on Khumbu Glacier from 19 May to 3 June 1999. Our study is motivated by the need to improve predictions of the response of debris-covered glaciers to changes in climate.

TEMPERATURE MEASUREMENTS

Hourly temperature measurements were recorded for six days at a site near Everest Base Camp where the debris was 0.40 m thick, and four days at another site near Lobuche where the debris was more than 2 m thick. Ten thermistors were used to measure vertical temperature profiles within the supraglacial debris, and a shielded thermistor positioned 1 m above the surface recorded air temperature. Surface temperature was measured by a thermistor inserted within a few millimetres of the surface of the debris. Other thermistors were arranged at 0.05 m spacing through the debris at Base Camp, which consisted of mainly angular, loosely packed, dark-coloured, schist cobbles up to 0.1 m size. The temperature at the debris/ice interface 0.40 m below the surface was constant at 0°C. Temperature measurements at other depths (Fig. 2) show a strong diurnal signal that diffused downward into the debris with decreasing amplitude and increasing lag. The surface forcing decreased in the afternoon of day 143 because clouds reduced the incoming solar flux.

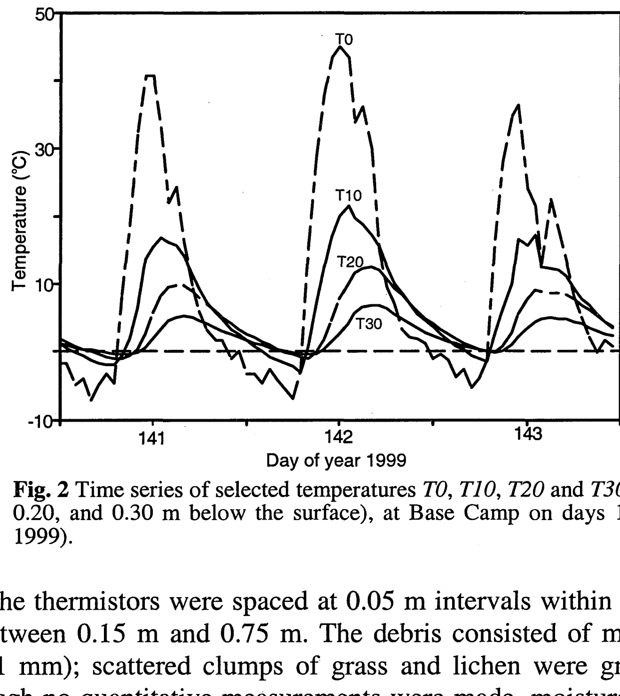

Fig. 2 Time series of selected temperatures *T0*, *T10*, *T20* and *T30* (at the surface, 0.10, 0.20, and 0.30 m below the surface), at Base Camp on days 141–143 (21–23 May 1999).

At Lobuche thermistors were spaced at 0.05 m intervals within the upper 0.15 m, and 0.1 m between 0.15 m and 0.75 m. The debris consisted of mainly coarse sand (grain size ~1 mm); scattered clumps of grass and lichen were growing within the debris. Although no quantitative measurements were made, moisture was observed in the upper 0.07 m of the profile during excavations on day 149. The excavations also indicated that the debris/ice interface was more than 1.75 m below the surface. Temperature profiles at this site also exhibit a strong diurnal signal that diffused downward into the debris with decreasing amplitude and increasing lag. The deepest thermistor (0.75 m below the surface) was above the debris/ice interface; its average temperature was 7.5°C.

Temperatures at the surface were often more than 35°C higher than the air during times of solar radiation. At night the surface was usually colder than the air, but the average surface temperature ($\overline{T0} = 10.1$°C) was warmer than the air ($\overline{T}_{air} = 2.3$°C). On average, sensible heat flowed from the debris to the air during the period of study; the flux of incoming energy was dominated by the solar component. The measurements show that high temporal resolution is necessary to fully capture the diurnal temperature cycle. Caution is needed when interpreting surface temperature from once-daily measurements.

TEMPERATURE PROFILES

The second derivative of temperature with respect to depth T'', scaled by the thermal conductivity, is the flux divergence (equation (2)). Negative curvature ($T'' < 0$) in a $T(z)$ profile represents cooling, while positive curvature represents warming. Four profiles from Base Camp (Fig. 3) show:

1. cooling curvature throughout;
2. warming curvature throughout;
3. warming at the top beginning to overtake weak cooling below;
4. cooling that has nearly overtaken weak warming at the bottom.

If the average temperature did not change over the 6-day observation period (i.e. $\dot{\overline{T}} = 0$), then for a purely conductive system with conductivity constant with depth, $\partial\overline{T}/\partial z$ should also be constant with depth (equation (1)). The measured $\partial\overline{T}/\partial z$

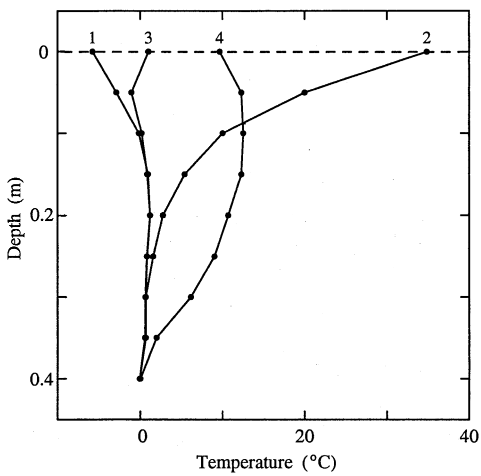

Fig. 3 Four representative temperature profiles at the Base Camp study site.

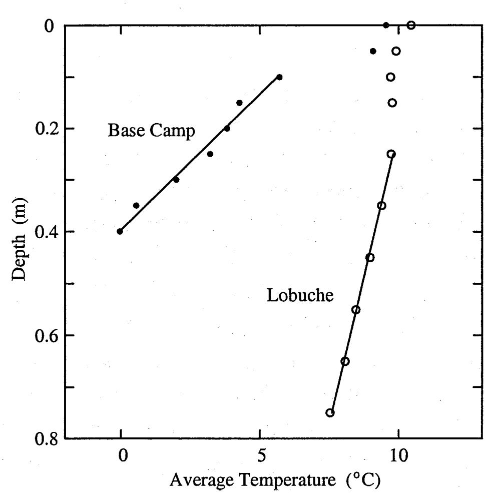

Fig. 4 Average temperature profiles at Base Camp and Lobuche.

(Fig. 4) show some departure from these conditions. Several factors could contribute to cause the deviations:

– $\dot{\overline{T}}$ may not have been zero over the period of measurements.
– The system may not be purely conductive. For example, convection is likely near the surface (Harris & Pedersen, 1998), while latent heat exchange might dominate near the debris/ice interface.
– Conductivity k may not be constant with depth. For example, k increases rapidly with moisture content (Farouki, 1981; Nakawo & Young, 1981) and any variation of moisture with depth would complicate the $T(z)$ profile.
– Measurement errors. In particular, the influence of uncertainties in the vertical position (5 mm) and thermistor calibration (0.05°C) depends on $\partial\overline{T}/\partial z$ and the effective measurement error may be up to 0.15°C (Appendix).

Nevertheless, results show $\partial\overline{T}/\partial z$ was remarkably constant at depths greater than 0.1 m at Base Camp, and 0.25 m at Lobuche (Fig. 4). Extrapolation of the Lobuche profile to 0°C indicates ice ~2.5 m below the surface, consistent with our excavations that reached 1.75 m without finding ice. The average temperature gradient at Base Camp (19 K m^{-1}) was much stronger than that at Lobuche (4.5 K m^{-1}), presumably caused in part by the difference in debris thickness. On average, heat flow deeper than about 0.2 m was apparently primarily by conduction during this dry, summer period (Fig. 4). Convective processes may be more important during winter when temperatures probably increase with depth, and buoyancy forces could induce air mixing through the debris layer (Harris & Pedersen, 1998).

THERMAL PROPERTIES OF THE DEBRIS

Calculating the flux of energy available for melting ice beneath debris requires knowledge of thermal conductivity k, as well as the temperature gradient $\partial\overline{T}/\partial z$ (equation (1)). We do not have direct measurements of k but instead we estimate the depth averaged thermal diffusivity $\overline{\kappa}$ using $\dot{T}$ and T'' (equation (2)). We estimate these temperature derivatives using standard centred finite-difference expressions and distributions of $\dot{T}$ *vs* T'' at selected depths are shown in Fig. 5; the slope of the best fitting line through the data gives an estimate of the average diffusivity at that depth. Surprisingly, the analysis is not affected by thermistor-position and calibration errors, provided the errors do not change with time or temperature (Appendix).

The scatter about the best fitting line at 0.20 m at Base Camp (Fig. 5(a)) indicates deviations from purely conductive behaviour were relatively small at that depth. In contrast, measurements at 0.35 m (Fig. 5(b)) show that T'' was both strongly positive and negative and yet $\dot{T}$ was near zero. Examination of equation (1) indicates that a nonconductive heat source/sink (i.e. $\partial Q_N/\partial t \neq 0$) would cause such conditions—a possibility is that phase changes and associated latent heat exchange occurred at this depth. A rough calculation indicates that latent heat would dominate if only 0.02% (by mass) of water changed phase per hour. The distribution of $\dot{T}$ *vs* T'' is well behaved at 0.25 m at Lobuche (Fig. 5(c)); apparently heat flow there is primarily by conduction and $\kappa = 0.9$ mm^2 s^{-1}.

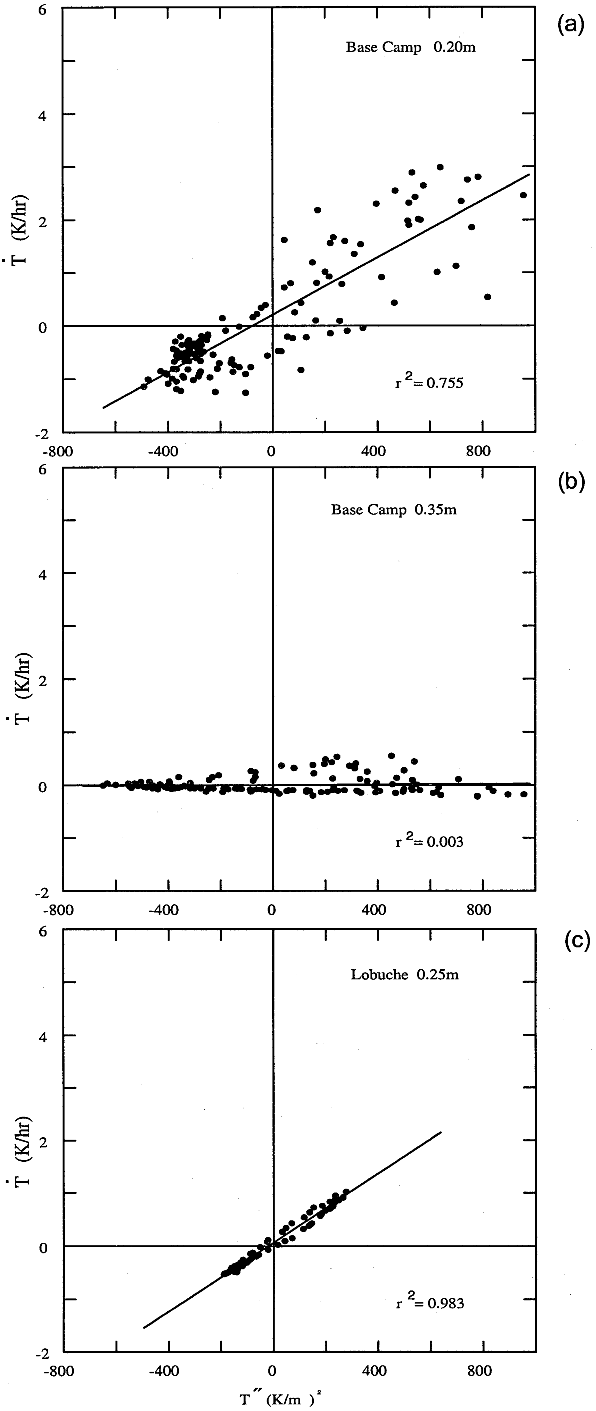

Fig. 5 Distributions of $\dot{T}$ *vs* T'' at selected depths within the supraglacial debris at study sites near Base Camp ((a) and (b)), and Lobuche (c). The slope of the best fitting line gives an estimate of κ (equation (2)).

Averaging thermal diffusivities through regions of the profile where heat flow is primarily conductive gives $\overline{\kappa} = 0.6 \pm 0.1$ mm^2 s^{-1} at Base Camp and 0.9 ± 0.1 mm^2 s^{-1} at Lobuche. Using typical values for the heat capacity (c = 750 J kg^{-1} K^{-1}) and density (ρ = 2700 kg m^{-3}) of rock (Clark, 1966), gives conductivity $k = 2.03\kappa(1 - \varphi)$ W m^{-1} K^{-1}, where φ is the porosity of the debris. The porosity of loose-packed debris is about 0.3, which yields k = 0.85 W m^{-1} K^{-1} at Base Camp, and 1.28 W m^{-1} K^{-1} at Lobuche. Assuming the estimate of $(1 - \varphi)\rho c$ has a 10% error, and that errors are uncorrelated (Bevington & Robinson, 1992), the uncertainty in k at Base Camp is 0.20 W m^{-1} K^{-1} and at Lobuche it is 0.15 W m^{-1} K^{-1}.

CALCULATED ICE-MELT

The average energy flux at the bottom of the debris layer:

$$\frac{\partial q}{\partial t} = k\frac{\partial \overline{T}}{\partial z} \tag{3}$$

yields a melt rate in mm day^{-1} of:

$$\frac{\partial h}{\partial t} = \frac{1}{\rho_i L}\frac{\partial q}{\partial t} = 0.29\, k\, \frac{\partial T}{\partial z} \tag{4}$$

Here L = 0.334 MJ kg^{-1} is the latent heat of fusion, and ice density ρ_i = 900 kg m^{-3}. At Base Camp, where the debris is 0.4 m thick, k = 0.85 ± 0.20 W m^{-1} K^{-1} and $\partial\overline{T}/\partial z$ = 19 K m^{-1}, the daily flux of energy at the debris/ice interface would melt 4–6 mm of ice. This is slightly lower than the melt measured beneath the same thickness of debris at a nearby site during the same period, which varied from 5 to 12 mm day^{-1} (Kayastha *et al.*, 2000). At Lobuche, where the debris is about 2.5 m thick, k = 1.28 ± 0.15 W m^{-1} K^{-1} and $\partial\overline{T}/\partial z$ = 4.5 K m^{-1}, the flux of energy through the debris would melt 1–2 mm day^{-1} of ice.

We expect that monsoon rains would contribute sensible heat to the system, and would also cause k to increase. Both effects would enhance ablation during the monsoon although they might be offset by increased cloudiness that would reduce the incoming solar flux. In any event, we suspect that the melt rate was close to a maximum at the time of our measurements (a few weeks before the summer solstice and before the monsoon). For a rough estimate, we assume that ice at Lobuche melts at about half this maximum rate for half the year (i.e. a periodic cycle), which yields an ablation rate of 100–200 mm year^{-1}. Complete melting of the ice at Lobuche (about 50 m, Gades *et al.*, 2000), will take 250–500 years—longer if the debris thickness increases.

For practical purposes it is appealing to estimate ice-melt beneath supraglacial debris from remote sensing of surface temperature Ts and an effective thermal resistance of the debris R. Provided the system is purely conductive and the temperature profile through the debris is linear, the rate of ablation can be written (Nakawo & Young , 1981):

$$\frac{\partial h}{\partial t} = \frac{1}{\rho_i L}\frac{T_s}{R} \tag{5}$$

where T_s is in °C, and the ice is at 0°C.

Results shown in Fig. 2, however, indicate that much caution is needed when estimating the surface temperature because the large diurnal variation makes it difficult to define the average surface temperature from once-daily measurements. Additional caution is needed because measurements shown in Fig. 4 indicate that the average temperature profile through the debris is typically not linear near the surface. Calculations are further complicated because deviations from linearity near the surface likely vary as convective processes become more or less active.

Deeper than about 0.2 m, however, the gradient $\partial\overline{T}/\partial z$ is more constant at both sites (Fig. 4) and heat transfer is primarily conductive. The thermal conductivity calculated through the lower portion of the debris at Base Camp (0.85 ± 0.20 W m^{-1} K^{-1}) was slightly lower than that at Lobuche (1.28 ± 0.15 W m^{-1} K^{-1}), but the average energy flux through the debris at Base Camp (~16 W m^{-2}) was much higher than at Lobuche (~5.8 W m^{-2}). Consequently, the ablation rate at Base Camp is expected to be about 3 times higher than at Lobuche.

Acknowledgements National Science Foundation supported this work through Grants ATM 9530691 and INT-9726704. We thank the Department of Hydrology and Meteorology of Nepal, and Ang Pasang Llama, Ang Pasang Sherpa, Man Bahadur, Pan Bahadur and Ang J.P. Llama who provided logistical support. We also thank L. Blank, A. Gades, T. Kadota, R. Kayastha, N. Naito, M. Nakawo, N. Nereson, C. Raymond and Y. Takeuchi for valuable contributions. Stimulating discussions with Bernard Hallet motivated this study.

REFERENCES

Adhikary, S., Seko, K., Nakawo, M., Ageta, Y. & Miyazaki, N. (1997) Effect of surface dust on snow melt. *Bull. Glacier Res.* **15**, 85–92.

Bevington, P. R. & Robinson, D. K. (1992) *Data Reduction and Error Analysis for the Physical Sciences*. McGraw-Hill, New York.

Clark, S. P. (1966) Thermal conductivity. In: *Handbook of Physical Constants: Memoir 97* (ed. by S. P. Clark), 459–482. Geological Society of America, New York.

Conway, H., Gades, A. & Raymond, C. F. (1996) Albedo of dirty snow during conditions of melt. *Wat. Resour. Res.* **32**, 1713–1718.

Farouki, O. T. (1981) Thermal properties of soils. *Tech. Report Mono, 91-1, US Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, USA.*

Fujii, Y. (1977) Field experiments on glacier ablation under a layer of debris cover. *J. Japan. Soc. Snow Ice (Seppyo)* **39,** 20–21.

Gades, A., Conway, H., Nereson, N., Naito, N. & Kadota, T. (2000) Radio echo-sounding through supraglacial debris on Lirung and Khumbu Glaciers, Nepal Himalayas. In: *Debris-Covered Glaciers* (ed. by M. Nakawo, C. F. Raymond & A. Fountain) (Proc. Seattle Workshop, September 2000). IAHS Publ. no. 264 (this volume).

Hallet, B. & Rasmussen, L. A. (1993) Calculation of the thermal conductivity of unsaturated frozen soil near the melting point. In: *Proc. of the Sixth Int. Conf. on Permafrost* (Beijing, China), vol. 1, 226–231. South China University of Technology Press, Wushan Guangzhou, China.

Harris, S. A. & Pedersen, D. E. (1998) Thermal regimes beneath coarse blocky materials. *Permafrost and Periglac. Processes* **9**, 107–120.

Inoue, J. & Yoshida, M. (1980) Ablation and heat exchange over the Khumbu Glacier. *J. Japan. Soc. Snow Ice (Seppyo)* **41**, 26–33.

Kayastha, R. B., Takeuchi, Y., Nakawo, M. & Ageta, Y. (2000) Practical prediction of ice melting beneath debris cover of various thickness on debris cover on Khumbu Glacier, Nepal, using a positive degree-day factor. In: *Debris-Covered Glaciers* (ed. by M. Nakawo, C. F. Raymond & A. Fountain) (Proc. Seattle Workshop, September 2000). IAHS Publ. no. 264 (this volume).

Kodama, H. & Mae, S. (1976) The flow of glaciers in the Khumbu region. *J. Japan. Soc. Snow Ice (Seppyo)* **38**, 31–36.

Mayewski, P. A. & Jeschke, P. A. (1979) Himalayan and trans-Himalayan glacier fluctuations since AD 1812. *Arctic Alpine Res.* **11**, 267–287.

Nakawo, M. & Takahashi, S. (1982) A simplified model for estimating glacier ablation under a debris layer. In: *Hydrological Aspects of Alpine and High Mountain Areas* (ed. by J. W. Glen) (Proc. Exeter Symp., July 1982), 289–296. IAHS Publ. no. 138.

Nakawo, M. & Young, G. J. (1981) Field experiments to determine the effect of a debris layer on ablation of glacier ice. *Ann. Glaciol.* **2**, 85–91.

Nakawo, M., Iwata, S., Watanabe, O. & Yoshida, M. (1986) Processes which distribute supraglacial debris on the Khumbu Glacier, Nepal Himalaya. *Ann. Glaciol.* **8**, 129–131.

APPENDIX

Measurement errors Analysis of a purely conductive, one-dimensional system with uniform diffusivity κ (equation (2)) is not hampered by measurement error δ that varies from sensor to sensor but that is constant in time t, and is therefore independent of temperature. If the measured temperature Θ differs from the true temperature T according to $T = \Theta + \delta$, when this is substituted into equation (2), it becomes:

$$\dot{\Theta} = \kappa\Theta'' + \kappa\delta'' \qquad \text{(A1)}$$

The error δ has no effect on the time derivative because it is constant in time, so $\dot{T} = \dot{\Theta}$. If Θ measurements at a particular level are analysed by finding the best-fitting line,

$$\dot{\Theta} = \alpha\Theta'' + \beta \qquad \text{(A2)}$$

and coefficients are equated, then $\kappa = \alpha$, and $\delta'' = \beta/\kappa = \beta/\alpha$. An example of a large δ'' detected by a linear fit of $\dot{\Theta}$ to Θ'' is at 0.20 m at Base Camp (Fig. 5(a)) where $\delta'' \sim 100$ K m^{-2}.

Although the best-fitting line does determine the value of κ at a particular level, it does not absolutely determine the sensor error δ there. The δ'' values at successive levels can be integrated twice with respect to z, however, to obtain calibration adjustments Δ that will produce values $T = \Theta + \Delta$ that satisfy equation (2), without need for a constant term, but the relation between Δ and the actual sensor error δ is not revealed.

Error ε in the depth z of a sensor corresponds to a mean error δ in its temperature that is proportional to the vertical gradient. Thus, where $\overline{T}$ is the mean temperature, the mean value is $\delta = \varepsilon\,\partial\overline{T}/\partial z$. A position error of 0.005 m at Base Camp, where $\partial\overline{T}/\partial z = 19$ K m^{-1}, corresponds to a temperature error of about 0.1°C. The same position error at Lobuche ($\partial\overline{T}/\partial z = 4.5$ K m^{-1}) corresponds to an error of only 0.02°C.