Parameterization of turbulent transfers between glaciers and the atmosphere

1 author:

E. M. Morris
University of Cambridge

Glaciers-Ocean-Atmosphere Interactions (Proceedings of the International Symposium held at St Petersburg, September 1990). IAHS Publ. no. 208, 1991.

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E.M. MORRIS
British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB3 0ET, UK
R.J. HARDING
Institute of Hydrology, Maclean Building, Crowmarsh Gifford, Wallingford, Oxfordshire OX10 8BB, UK

ABSTRACT Using the snowmelt routine of a distributed hydrological model, changes in snow pack depth can be simulated accurately from hourly meteorological data using scaling lengths $z_0 = z_E = z_H$ in the range 4-18 mm (Greenland), 3-6 mm (Finse), and 34-52 mm (Hintereis). The effective value of z_0 increases as the frequency of the input meteorological data is decreased but good simulations may still be obtained.

INTRODUCTION

Parameterization of the boundary conditions between the atmosphere and glaciers, ice sheets, and sea ice is an important part of the general problem of modelling the effects of climatic change. The physics of the boundary processes is fairly well known; the difficulty lies in choosing the appropriate space and time scales for modelling. This paper focuses on the problem of calculating turbulent transfers of mass, momentum and energy. Equations for these fluxes contain parameters z_H, z_E, and z_0, the so-called scaling lengths. It is normally assumed that their values are constants, determined by the characteristics of the snow or ice surface alone. However, in modelling practice, it is often found that the effective values of the parameters, i.e. those values which give the best simulations, are also influenced by the natural variability in the surface meteorological conditions.

THE IHDM SNOW ROUTINE

The Institute of Hydrology Distributed Model (IHDM) is a physically-based mathematical model for hydrological processes within a catchment. It contains a sub-model to calculate meteorological inputs for a given site from Automatic Weather Station (AWS) data and a snow routine to predict changes in the snow cover from the meteorological inputs, given the initial depth, density, and liquid water content of the snow.

The meteorological sub-model corrects air temperature and humidity for any difference in altitude between the snow site and the AWS (using standard lapse rates) and adjusts the radiation input according to the slope angle and aspect.

The snow routine is a development of the energy budget model described in Morris (1982). The energy input into the snow pack is used first to increase the average temperature T, then, if the maximum $T_{max} = 0^\circ$C has been reached, to increase the average liquid water content W and finally, if the maximum $W_{max} = 0.1$ has been reached, to melt snow at the upper surface and decrease the snow pack depth.

The energy input at the upper surface is the sum of net radiation, heat transported by precipitation, and sensible and latent heat gained by turbulent transfer from the atmosphere. These last two components, H and LE, are calculated from the equations

$$H = U\,k^2\,(T_a - T_s)/[\ln(Z/z_0)\ln(Z/z_T)]$$

$$= U\,k^2\,(T_a - T_s)/[\mathrm{l}\;(Z/z_H)]^2 \qquad (1)$$

$$E = U\,k^2\,(q_a - q_s)/[\ln(Z/z_0)\ln(Z/z_q)]$$

$$= U\,k^2\,(q_a - q_s)/[\ln Z/z_E)]^2 \qquad (2)$$

where z_0 is the aerodynamic roughness length and z_T, z_H, z_q, z_E, are the scaling lengths for temperature, sensible heat, humidity, and water vapour; k is von Karmans constant. U is the wind speed and T_a and q_a are the air temperature and humidity at height Z above the snow surface. The IHDM allows the use of different values for z_0, z_H and z_E if required, but in this paper it is assumed that they are equal.

The value of the snow surface temperature at time t+1 is estimated from the value at time t using the equation

$$T_s(t+1) = T_s(t) + 2\,E\,/(K\,\rho\,c_p\,\pi)^{1/2} \qquad (3)$$

where E is the energy input over the time step, K is the effective thermal conductivity (calculated from the density ρ) and c_p is the specific heat. The surface temperature is not, however, allowed to rise above 0°C. It is assumed that q_s is the saturated value at temperature T_s.

The efficiency of the model in simulating changes in snowpack depth is defined in terms of a function

$$Fz = (Z_p - Z_m)^2/(Z_m - \bar{Z}_m)^2 \qquad (4)$$

where Z_p are predicted and Z_m are measured depths. $\bar{Z}_m$ is the average value of Z_m. The effective value of z_o for a certain site is defined as that which produces the minimum Fz.

FIELD SITES

The study makes use of intensive micrometeorological data from three field sites: in the firn area of the Hintereisferner, Oetztal Alps (Austria), on a snow covered frozen lake near Finse, Hardangervidda (Norway), and in the SW coastal uplands of Greenland. In each case observations were made for a two to three week period during the main melt season (May in the case of Norway, July on the Hintereis and May/June in Greenland). At all sites measurements were made of snow depth, net all-wave radiation, evaporation, snow temperature and air temperature, humidity, and wind speed at least one height. The three data sets cover a wide range of snow melt environments with climates ranging from Arctic (Greenland) to Maritime (Norway) and Alpine (Austria).

The Finse site was on a snow covered lake (Finsevatten) to the north of the Hardanger Ice Sheet in southern Norway (60°42'N, 7°28'E, altitude 1220 m). At the start of the observations the pack was melting throughout its depth. The liquid water content was measured at 8% and must be assumed to be constant for the measurement period. Through the measurement period (11th May to 27th May) there was 238 mm of snow melt.

The Hintereis site was on a 5°NE facing slope in the firn area of the glacier (46°48'N, 10°56'E, altitude 2960 m) (Harding et al., 1989). As with the Finse site the pack was at 0°C throughout its depth at the start of the period and this continued throughout (apart from at night when some refreezing took place in the surface layers). From 11th to 22nd July the depth of the pack decreased from 128 to 101 cm and including the 50 mm water equivalent of snowfall between the 18th and 20th this constitutes 199 mm of melt.

The Greenland site was on a snow covered lake in an upland area in the SW coastal region, approximately 50 km from the open sea and 110 km from the edge of the Greenland Ice Sheet. Unlike the other two sites, the snow pack was well below freezing at the beginning of the study period (19th May). For the first nine days there was no change in the water equivalent of the pack and considerable energy went into raising its temperature and liquid water content.

For the three sites the daily melt rates were 15, 18, and 9.5 mm respectively. For the Norwegian and Austrian sites the net radiation provided 56% and 63% respectively of the total energy required for the melt (the bulk of the remainder being provided by sensible heat flux). For the Greenland site the net radiation exceeded the observed melt by 16%, the excess energy being used to produce evaporation. The average temperature and humidity were remarkably similar, considering the very different locations. This was probably the result of the control exerted by the large melting snow surfaces on the lower atmosphere (Harding, 1986). The variation of sensible heat between sites arose from variations in the wind speed and the roughness length. At the Finse and Greenland sites the wind speed was probably determined by the large scale flow but on the Hintereisferner by the glacier wind circulation.

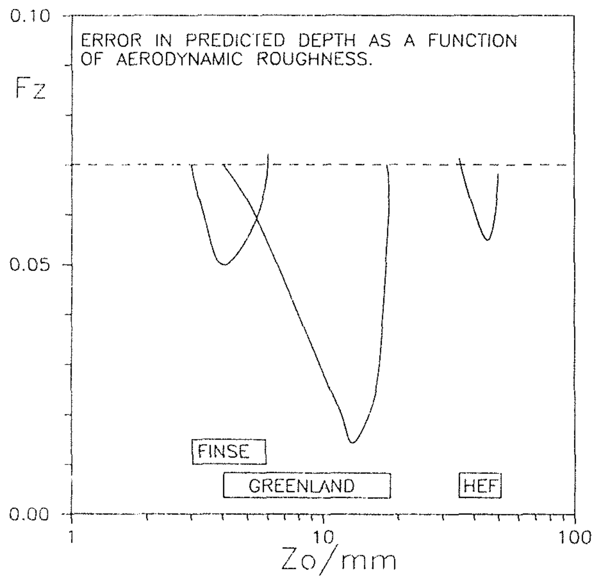

FIG. 1 Error in predicted depth as a function of aerodynamic roughness.

RESULTS OF THE IHDM SNOW MELT SIMULATIONS

At each site the IHDM snow routine was run using the hourly measured meteorological data and the snow depths predicted by the model were compared with observations. Measurements made during the first day of the simulation were used as estimates for the initial snow properties (depth, density and liquid water content). At each site the IHDM was run for a range of values of z_0. A well defined minimum of the error function Fz was found for each site (Fig. 1), although for the Greenland site Fz is not so sensitive to variations in z_0 since the heat flux and evaporation are of similar magnitude but opposite sign and thus the overall turbulent exchange makes only a small contribution to the energy available for melt. The effective values of z_0, i.e. the values at minimum Fz, differ at each site. Although those for the Norway and

Greenland sites (4.5 and 14 mm respectively) are probably not significantly different, that found on the Hintereis (45 mm) is considerably larger. Large changes in the effective value of z_0 can occur owing to errors in the input variables, (net radiation and snow density in particular), but the comparatively large value for the Hintereis cannot be explained by the expected errors in these quantities (Harding et al., 1989).

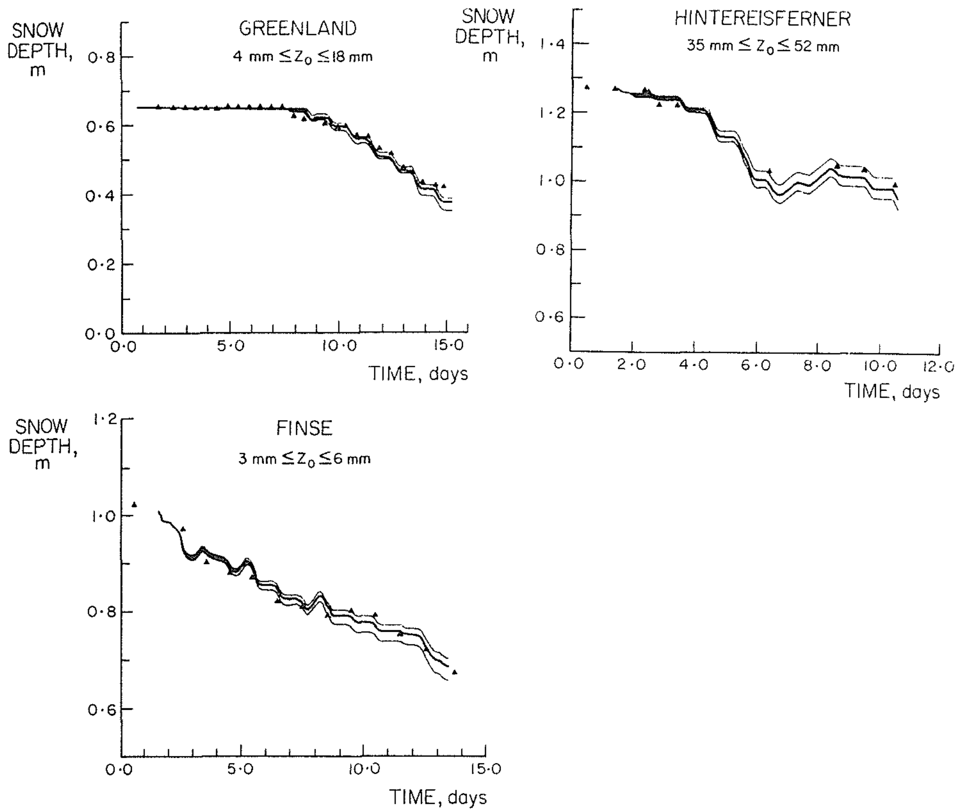

FIG. 2 Measured and predicted snow pack depths at Greenland, Finse and Hintereis.

Fig. 2 shows for each site the observed snow depth and the depths predicted from hourly meteorological data using the effective value of z_0 and the two values defined by Fz=0.07. Even with comparatively large departures from the optimum value of z_0, acceptable fits (with an error in the final depth of no more than ± 2 cm) can be achieved.

It is possible to obtain equally good simulations using less frequent meteorological data. Fig. 3 shows the results from model runs with the input data averaged over periods of 3, 6, 12, 24, and 48 hours and used as input to the IHDM which was, however, still run using an hourly time step for calculation. It was found that, for each site, the effective value of z_0 increased as the averaging period for the input data increased. Using daily data, for example, good simulations with Fz < 0.07 could be obtained with z_0 in the range 11-21 mm (Finse) and 57-66 mm (Hintereis). The goodness of fit, as evidenced by the width of the Fz = 0.07 confidence limits on Fig. 3, only decreases slowly with increasing averaging time. This is partly because the measurements of snow depth were made at 12 h (Greenland) or 24 h (Austria) intervals but also suggests that there was not a strong correlation between the two meteorological variables which go into the calculation of the sensible heat flux (i.e. temperature and wind speed).

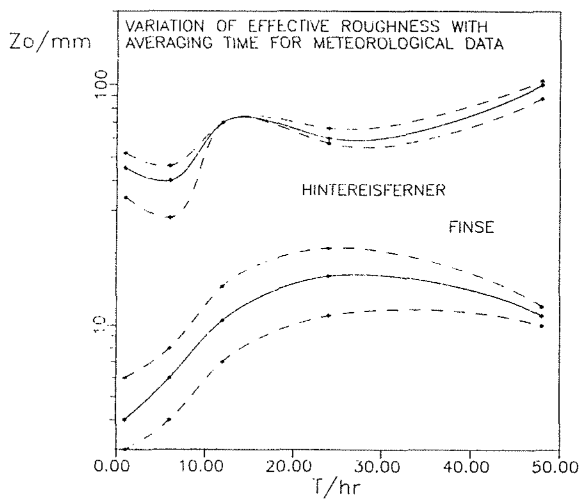

FIG. 3 Variation of effective roughness with averaging time for the meteorological data.

CONCLUSION

These results show that a physically-based model can predict snow melt at a point accurately, provided appropriate values of the turbulent transfer parameters are known. The effective values derived above can be used in a wide range of studies of the effects of climate change, for example, estimation of climate-induced changes in seasonal snow cover, the mass balance of ice sheets, and sea ice extent. It should be noted, however, that these values cannot be directly applied in the prediction of climate change per se. The effective value for z_0 over snow derived here by optimising predictions of snow pack depth is not necessarily that which would be obtained by optimising using a meteorological variable within a GCM.

One reason for this is the effect of the extrapolation of the point model to the GCM grid scale (typically 200 km). The variability of surface characteristics and processes on the sub-grid scale will be reflected in the effective value of z_0. Furthermore, the GCM allows feed-back between the snow and the surface layers of the atmosphere which has not been included in the snow model. This will affect the relation between z_0 and the frequency of the input meteorological data.

Present GCMs have extremely primitive parameterisations of snow melt processes. There should be advantages in the introduction of more physically-realistic models, such as the IHDM. However, use of parameters appropriate to the point scale may not lead to improved simulations. New effective parameters incorporating sub-grid scale variability and feed-back effects will be required.

ACKNOWLEDGEMENTS We should like to thank the many scientists who worked with us in the field, in particular: M. Kuhn, G. Kaser, H. Rott, H. Sogaard, A. Jenkins, I.R. Calder and R.C. Johnson.

REFERENCES

Harding, R.J. (1986) Exchanges of energy and mass associated with a melting snowpack. Proc. BUDAPEST Symposium, IASH Publ. 155, 3-15.

Harding, R.J., Entrasser, N., Escher-Vetter, H., Jenkins, A., Kaser, G., Kuhn, M., Morris, E.M. & Tanzer, G. (1989) Energy and mass balance studies in the firn area of the Hintereisferner. In: Glacier Fluctuations and Climatic Change (ed. by J. Oerlemans). Kluwer Academic Publishers, 325-341.

Morris, E.M. (1982) Sensitivity of the European Hydrological System snow models. In: Hydrological Aspects of Alpine and High Mountain Areas (Proceedings of the Exeter Symposium, July 1982). IAHS Publ. 138, 221-231.