MODELING THE EFFECT OF RUNOFF PROCESSES ON SNOWMELT HYDROGRAPHS

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ABSTRACT

Much of the earlier research work on predicting snowmelt flood hydrographs has been concerned with improving models of the melt process. Field experience, however, would suggest that the processes of snowmelt runoff are more complex than normally considered in flood hydrograph models and that correctly predicting the processes of snowmelt runoff is at least as important as the prediction of melt rates. It is suggested that the actual flow path of melt water may make a difference of at least an order of magnitude in hydrograph peaks and lag times. This study focuses attention on the important threshold between melt that infiltrates into the soil surface and that which reaches the stream over the surface and through the snowpack itself. The role of frost in the surface soil horizon and the level of soil saturation are both important in governing the relative amounts of runoff going to fast surface and slow subsurface routes. The effects of runoff processes on snowmelt runoff are investigated under a variety of conditions by examining the sensitivity of a combined model of flow through snow and soil. The model can simulate dynamic contributing area and return flow effects that have been neglected in previous studies. The results of the model simulations show how the degree of surface frosting of the soil and slope convergence may affect the shape and magnitude of the predicted snowmelt hydrograph.

INTRODUCTION

Snowmelt runoff dominates the annual water balance over large areas of the nothern U.S. and Canada. Accurately predicting snowmelt hydrographs has considerable practical and economic significance. Previous research on snowmelt flood hydrographs has generally been concentrated on improving models of the melt process (Colbeck and Ray, 1979). However, there is now a considerable body of field experience to suggest that the paths followed by snowmelt may be equally, if not more, important than the melt process in affecting the magnitude and shape of snowmelt flood hydrographs (Dunne and Black, 1971; Dunne, 1978). Further evidence for this is provided by results from the current generation of conceptual simulation models of snowmelt runoff (See, for example, figure 4 of Anderson, 1979, that demonstrates the important effect of concrete frost on snowmelt hydrographs). Anderson states that 'no provision is currently included for the effects of frozen ground or other snow-soil interactions' (Anderson, 1979, page 338), and this is generally true of most simulation models.

It is precisely the interaction between snow and soil that is considered crucial in the present study. That interaction controls the important threshold between melt that infiltrates into the soil and that which reaches a stream channel over the surface and within the snowpack itself. The role of frost in the surface soil horizon and the level of saturation of the soil are both important in governing the relative amounts of meltwater following fast surface and slow subsurface flow routes. It is suggested that melt on a dynamic area of soil saturated to the surface by a rising water table may be the most common mechanism generating snowmelt hydrographs over large areas of the northern United States. Yet this process has been given only a cursory mention in the literature.

This study examines the extremes in snowmelt runoff hydrographs that might be expected for constant melt conditions, but with different antecedant conditions of soil frosting. This is achieved using a coupled model of both subsurface and snowpack flow processes. This model can simulate dynamic contributing area and return flow effects that have been neglected in previous studies.

THE COUPLED FLOW MODEL

The model used in this study combines a one-dimensional downslope representation of saturated flow through a snowpack with a two dimensional saturated/unsaturated model of flows through the soil horizons. The two components are coupled through the mechanism of infiltration with the pressure potential at the soil surface being defined by the depth of saturation in the snowpack when flow is occurring at the base of the snowpack. The combined model is similar to that used by Beven (1977) for the case of overland flow at the soil surface, but with modifications for treatment of flow through the snowpack as a porous media flow and for the case of restricted infiltration due to soil frosting.

Flow Through the Soil

The soil water flow component of the model assumes that the velocities of flow are adequately described by Darcy's law under both saturated and unsaturated conditions. The governing equation is then a form of the Richards' equation

$$BC(\psi)\ \frac{\delta\psi}{\delta t} = \frac{\delta}{\delta x}\ \ BK(\psi)\frac{\delta\psi}{\delta x}\ \ + \frac{\delta}{\delta z}\ BK(\psi)\frac{\delta\psi}{\delta z} + BK(\psi) \qquad (1)$$

where ψ is capillary potential, $K(\psi)$ is hydraulic conductivity, $C(\psi)$ is specific moisture capacity = $d\theta/d\psi$, θ is volumetric water content, t is time, B(x) is slope width, x is horizontal distance and z is height above some arbitrary datum.

In these preliminary simulations it is assumed that the soil is isotropic and that its hydraulic properties can be described by single valued relationships between ψ, $K(\psi)$ and $C(\psi)$. With reference to Figure 1 the boundary conditions used are:

along AE, $q = 0$ (2a)

along DE, $q = 0$ (2b)

along AB, $\psi = 0$ (2c)

along BC, $q = 0$ (2d)

along CD, $q = i,\ h = 0$ (2e)

where q is the flux rate normal to the boundary, h is thickness of saturation in the snowpack, and the infiltration rate i depends on the melt rate and degree of soil frosting. The boundary conditions along CD when $h > 0$ are considered below. It is assumed that the seepage face AB is not affected by flows in the channel to which it contributes water. The model incorporates a facility to allow slope width to vary in the x and z directions so that the effects of topographic convergence and divergence can be simulated in a quasi-three dimensional way.

Equation (1) is solved by a finite element technique based on the Galerkin method of weighted residuals. An implicit finite difference scheme is used to represent the time derivative and an iterative method of solution is used at each time step to cope with the non-linearities of the soil moisture characteristic curves in the unsaturated part of the flow domain. Calculation of discharge from the seepage face at the base of the slope is handled in a similar manner to that described by Neuman (1973) and Beven (1977).

Flow Through The Snowpack

Only saturated flow through the snowpack is considered, with vertical unsaturated flow being specified as part of the model input. In addition, further simplifying assumptions are made that the snowpack may be considered a homogeneous porous medium of constant hydraulic conductivity and effective porosity and that the hydraulic gradient within the saturated snow is equal to the slope of the soil surface. Thus, the effects of structural characteristics of the snowpack, such as ice lenses, are neglected. The structure of the pack may have an important effect on flow at the early stages of melt, particularly unsaturated flows. However, these assumptions should be reasonable for snow of high hydraulic conductivity and realistic melt rates. Under these assumptions, lateral flow through the snowpack may be described by an equation similar to that used by Colbeck (1972)

$$B\varepsilon \frac{\delta h}{\delta t} + \frac{\delta}{\delta x} (BK_{sw} h \sin \alpha) - i + f = 0 \qquad (3)$$

where h is the depth of saturation at the base of the snowpack, ε is the effective porosity of the snowpack, K_{sw} is the saturated hydraulic conductivity of the snow, α is slope angle, i is the melt rate reaching the saturated layer in the snow and f is the infiltration rate into the soil surface (negative for the case of return flow).

This is a kinematic wave equation that for the particular case of constant, K_{sw}, ε and (i - f) has a simple solution of a constant delay from any point of the slope independent of h. A numerical solution is used in the model since in general i and f may vary rapidly with distance and time.

Snow-Soil Interaction

Two types of snow-soil interaction may be distinguished: one in which there is a direct interaction between snow and unfrozen soil at the soil surface; and one in which there is an intermediate layer of frozen soil. In both cases, the link between the two components is made through the infiltration process. The link is an internal coupling because the infiltration capacity at the soil surface depends both on moisture conditions in the soil and the depth of saturation in the snow. Thus, the soil surface boundary conditions are specified by

$$f = i, \; h = 0 \qquad (4a)$$

$$\psi_s = h, \; h > 0 \qquad (4b)$$

where the subscript s refers to a value at the soil surface. The case of a layer of frozen soil of low hydraulic conductivity was handled by a simple approach that takes into account only saturated flow through the frozen layer. The boundary conditions for the subsurface flow solution are then given by

$$f = i, \ h = 0 \qquad (5a)$$

$$f = K_f \left[\cos\alpha + \frac{(h-\psi_b)}{d}\right], \ h > 0 \qquad (5b)$$

where ψ_b is soil capillary potential at the base of the frozen soil layer, K_f is the saturated hydraulic conductivity of the frozen soil and d is the thickness of the frozen layer.

MODEL SIMULATIONS

Data for the initial application for the model have been loosely based on conditions on three instrumented hillslope plots in the Sleepers River catchment, Vermont (see for example, Dunne and Black, 1971). The basic hillslope profile used for each simulation is shown in Figure 2. The initial conditions used (based on limited piezometer data), assumed that the water table ten days prior to snowmelt was at the junction between the two soil horizons with no further input before the onset of melt. The soil characteristic curves were based on experimental measurements on soil cores, with relationships of the form

$$\psi = \psi_s \left(\frac{\theta}{\theta_s}\right)^{-b} \qquad (6a)$$

$$K = K_s \left(\frac{\theta}{\theta_s}\right)^{2b+3} \qquad (6b)$$

fitted to the experimental points. Parametric data for the soils used in the simulations is given in Table 1. Representative melt rates for the site were based on energy budget calculations using data from a nearby meteorological station (Figure 3a). No attempt has been made in these preliminary simulations to match the observed slope discharges or changes in soil moisture status during a snowmelt period.

Four simulation runs of 100 hours duration have been made to demonstrate the effects of runoff processes on snowmelt hydrographs under different conditions at the same site.

Simulation 1.

The case represents an extreme for which infiltration at the soil surface is not limited at all by frosting. In this repsect, this case is similar to the type of conditions analyzed by Stephenson and Freeze (1974). For the particular conditions simulated in the present study, all the melt infiltrates into the soil and there is no lateral flow at the base of the snowpack. The simulated discharge from the slope is shown in Figure 3b. The daily snowmelt hydrograph peak is barely distinguishable, but is superimposed upon a very gradual increase in discharge as the slope wets up in this early part of the melt.

Table 1

Soil parameters used in the model simulations.

	θ_s	K_s (mh^{-1})	ψ_s (m)	b	K_f (mh^{-1})	d (m)	ε	K_{sw} (mh^{-1})
Silt loam subsoil	0.44	0.01	0.34	3.54	-	-		
Sandy Loam topsoil	0.58	0.36	0.34	3.54	-	-		
Frozen topsoil (Run 3)	-	-	-	-	0.0001	0.1		
Snow							0.6	100.0

Simulation 2.

In this case an impermeable frost layer at the soil surface was assumed. This is similar to the type of opposite extreme case analyzed by Colbeck (1974) in which the subsurface flow regime plays no part in producing the snowmelt hydrograph which is shown in Figure 3c. Peak discharge occurs within minutes of peak melt rate and discharge falls to zero at night.

Simulation 3.

This represents an intermediate case in which permeable concrete frost at the soil surface limits infiltration rates. Note from Table 1 that the hydraulic conductivity of the frosted layer was 3600 times less than that of the unfrosted topsoil. In this case there is a contribution to the snowmelt hydrograph from both subsurface flow and lateral flow through the snowpack (Figure 3d). Time to peak is again very short due to the high conductivity of the snow, but the magnitude of the peak is much reduced from the previous run. There is a slight rise in peak discharge as the slope wets up over the four day period simulated.

Simulation 4.

In the final simulation a convergent slope (3:1 linear convergence in plan) was subjected to the same melt conditions with an unfrozen topsoil. Again a period of ten days of drainage from the initial conditions was assumed prior to the onset of melt. The simulated hydrograph is shown in Figure 3e. The initial discharge was higher and the inital elevation of the water table at the base of the slope is higher when compared to Simulation 1. The convergent slope shows a more sensitive response to the melt inputs than the straight slope. Total volumes of discharge per unit width of slope base are higher than for the impermeable and permeable frosted slopes. In addition, the gradual increase in discharges during the melt period is more pronounced. The average peak delay was about four hours. The water table is maintained at a higher elevation than for the straight slope, but did not saturate the soil to the surface during this run.

DISCUSSION

The simulations reported in this paper have been used to explore the response of a single hillslope to snowmelt inputs under different degrees of soil frosting. It is known from field experiments that soil frosting can have a very important effect on the runoff hydrographs generated by snowmelt. The model results suggest that due to the generally low flux rates of melt water reaching the base of the snowpack, the hydraulic conductivity of the frozen layer must be very low before the frozen topsoil acts to restrict infiltration and generate significant amounts of lateral flow through the snowpack. The simulations also suggest that slope convergence may be at least as important as soil frosting in determining the magnitude of slope discharge. Field evidence suggest that the effects of slope convergence may be even more important where saturation of the soil profile occurs leading to rapid flow through the snowpack. No areas of soil saturation were generated in the simulations reported here, but this will be the subject of further study.

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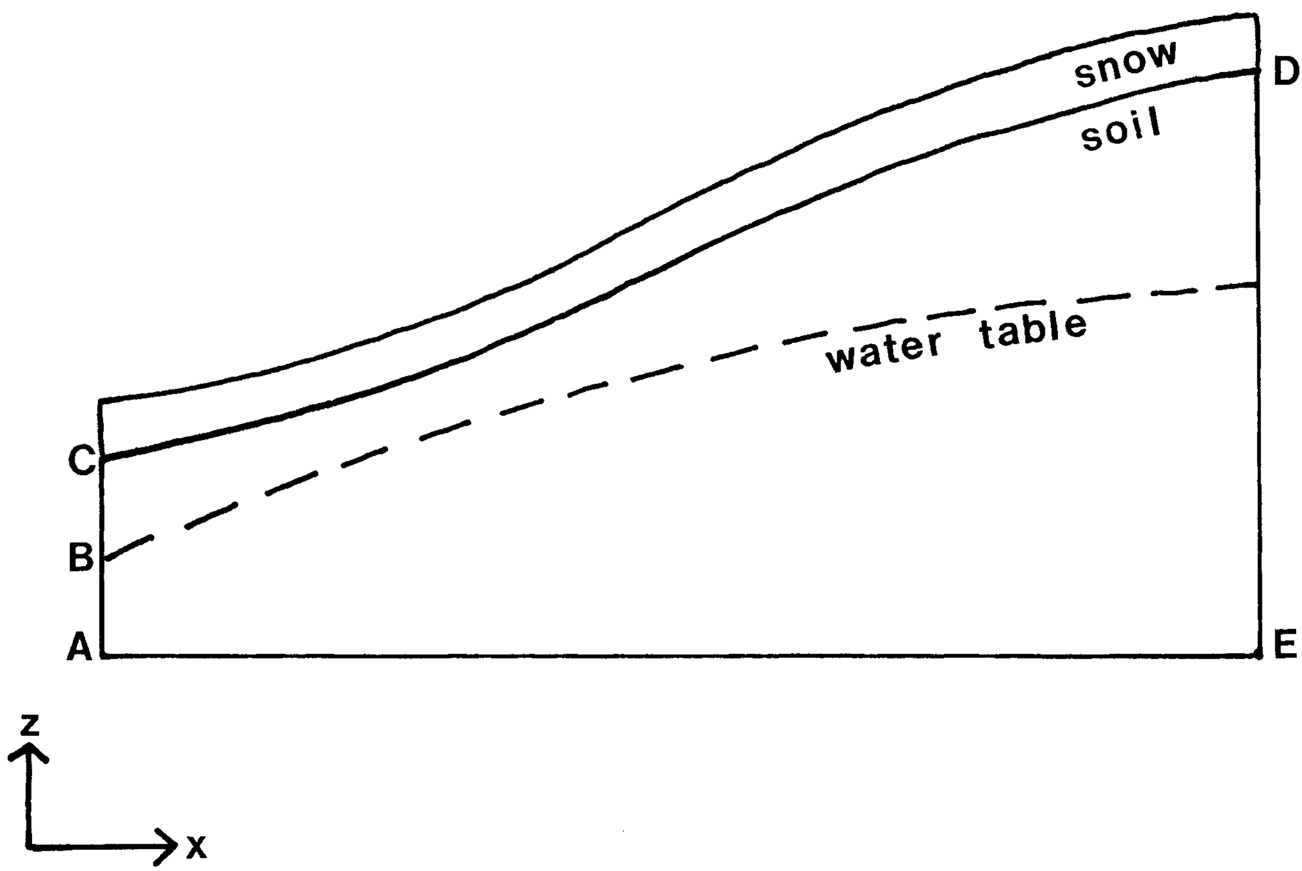
FIGURES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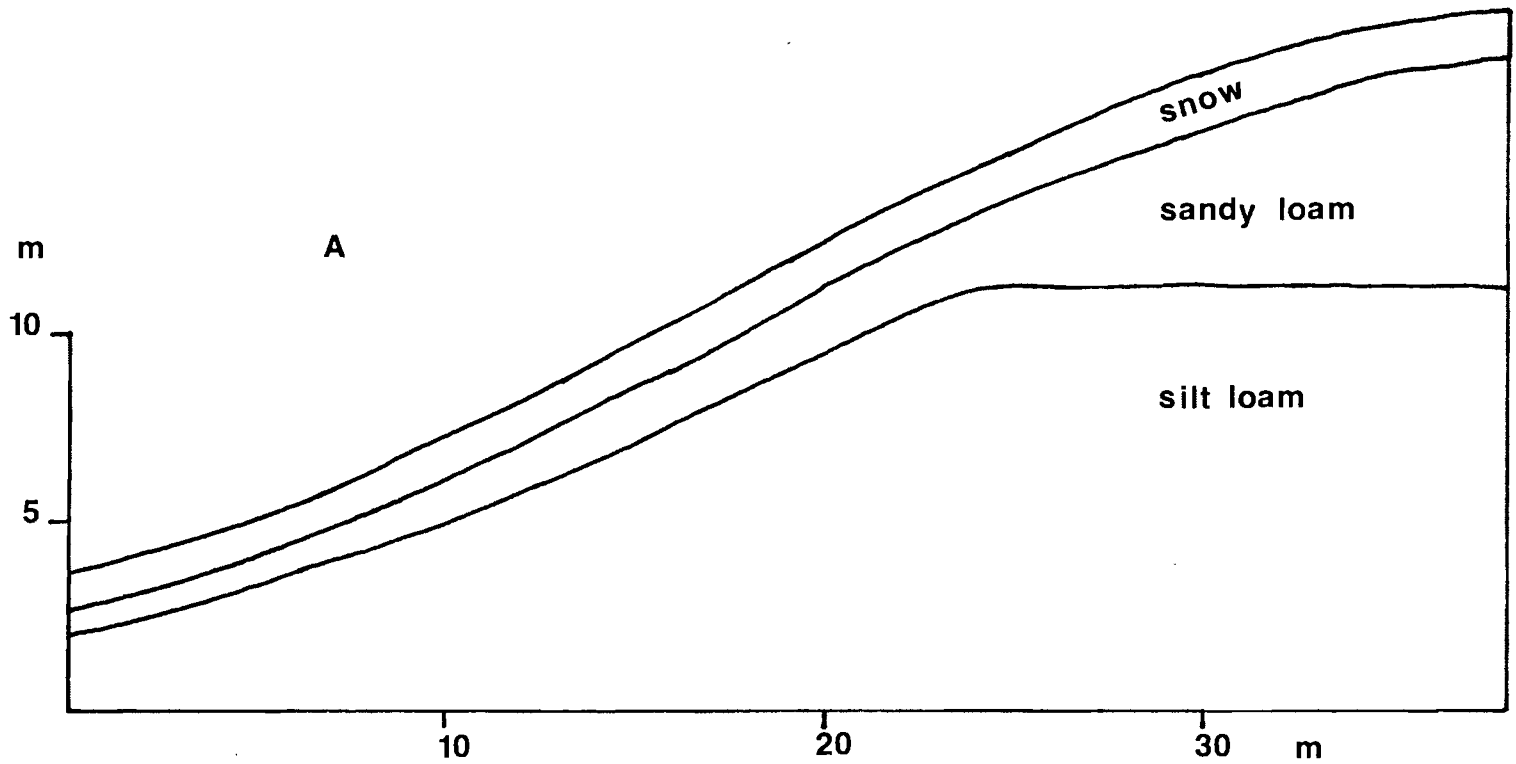
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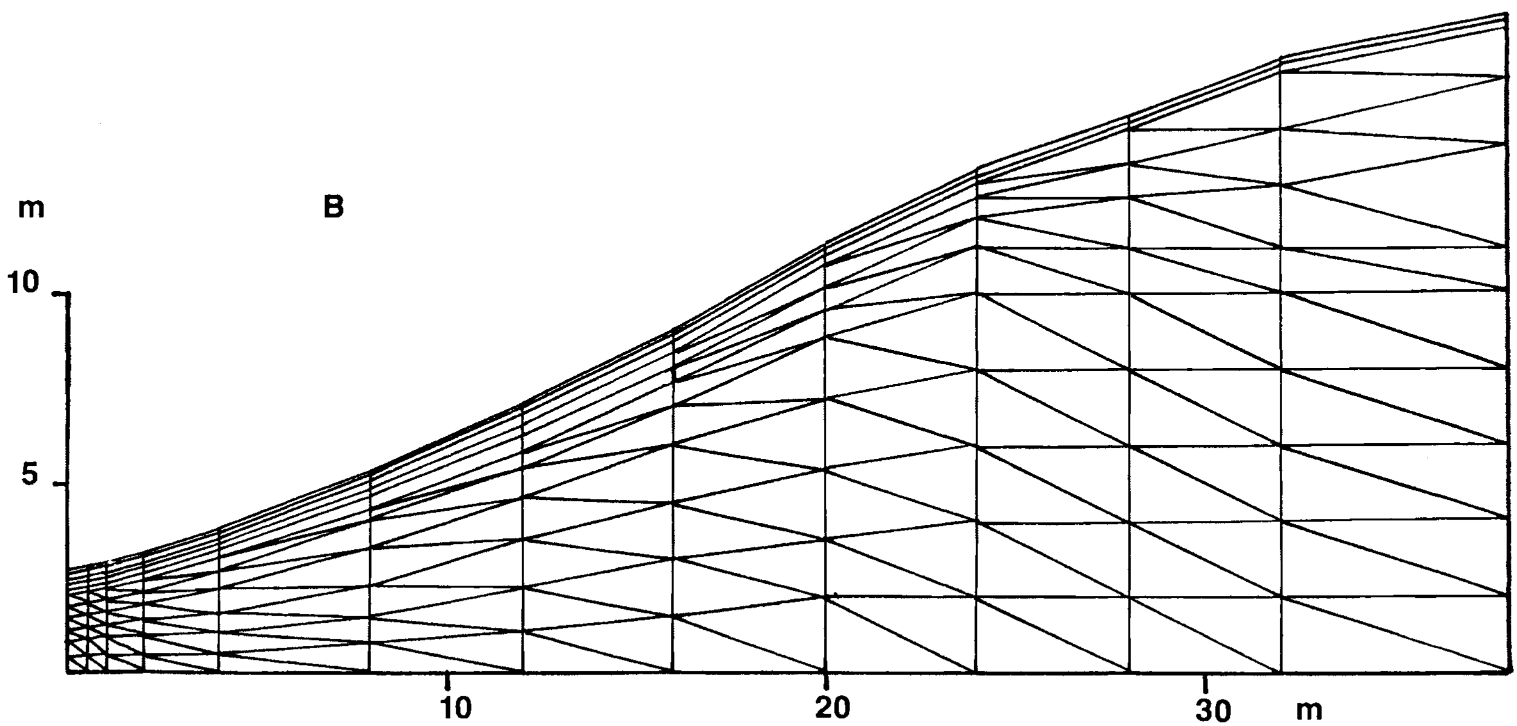
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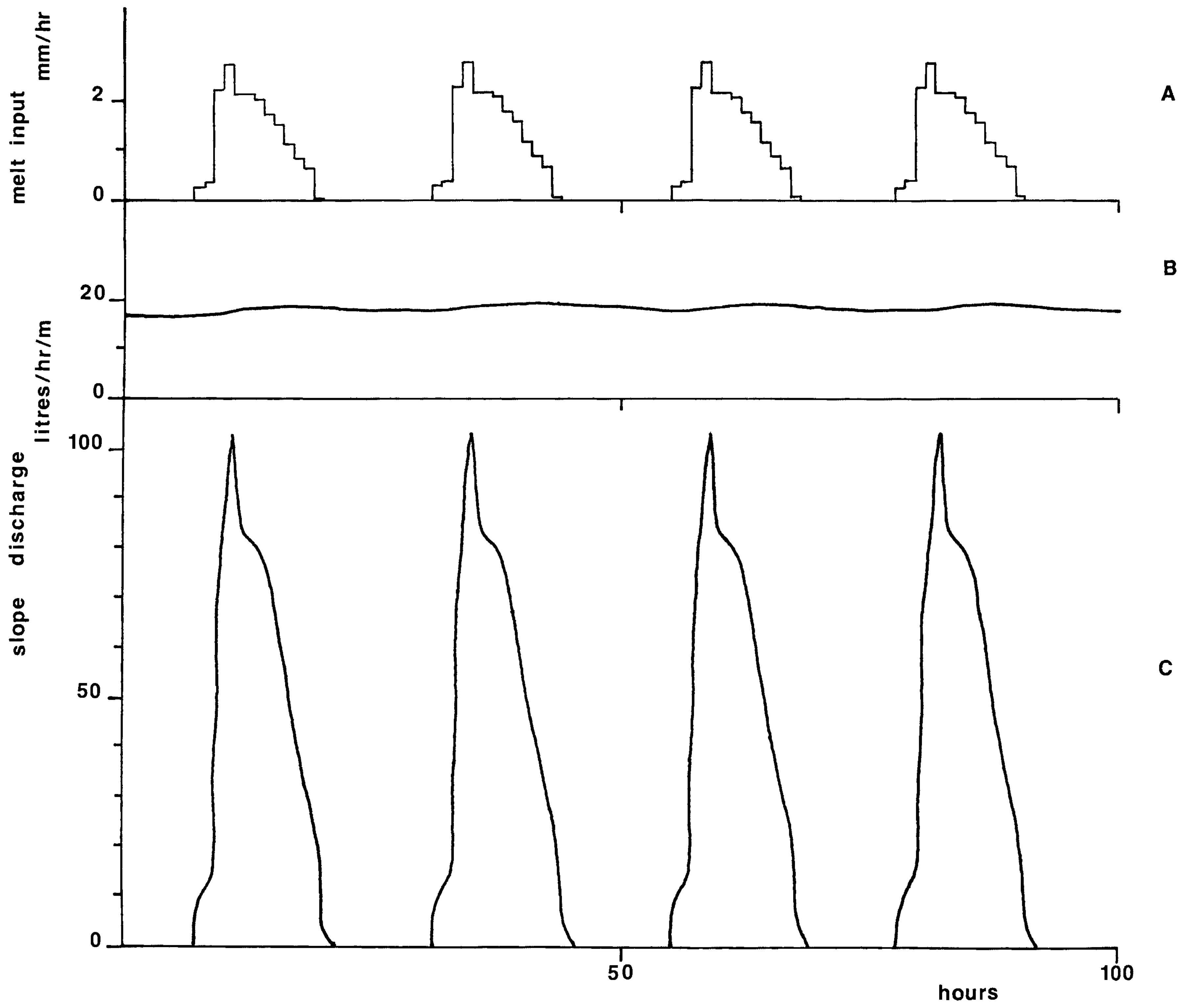

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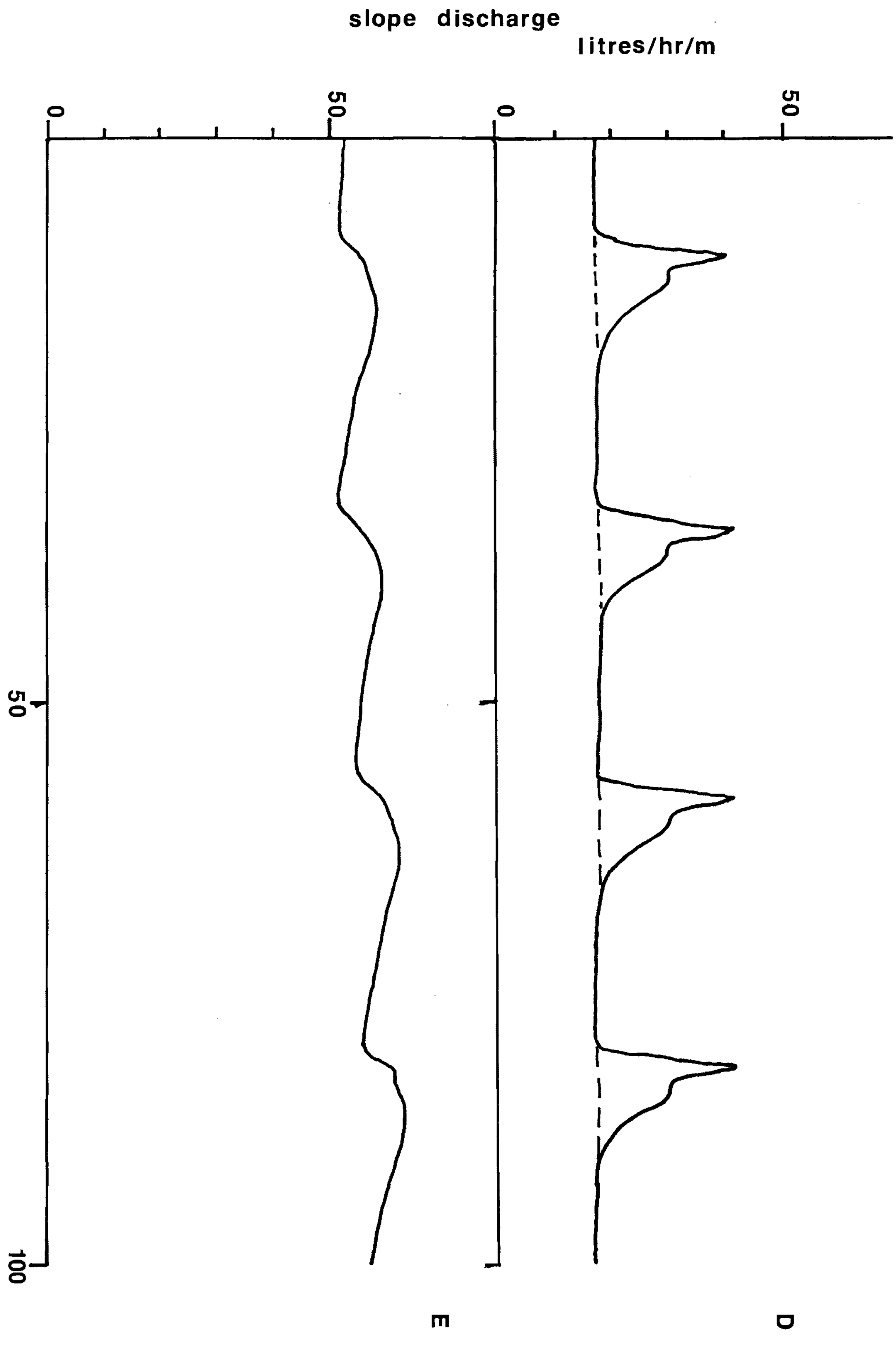
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