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Movement of Meltwater in Small Basins

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A kinematic approach is applied to compute moisture flux and particle velocities in the unsaturated zone. The computations are compared with observations of tracer marked meltwater particles. Groundwater flow is computed accounting for storage and assuming constant particle velocity. It is found that meltwater particles that reach the groundwater not very close to the ground surface do not contribute to runoff until several months after the melt event. The meltwater constitutes a significant part of the melt runoff only when saturated overland flow occurs or when the percolated water flows along the rock beneath a thin soil layer.

Introduction

The response of small rivers on snowmelt depends on the soil and groundwater conditions. The quality of the meltwater changes depending on for how long time water particles reside in different soil zones. During and just after snowmelt, when the soil is wet and before the growing season has begun, the vertical flow in the soil is controlled by gravity. Capillary tension plays a minor role. Particle velocities and moisture flux can be estimated in a straight-forward way. In this paper tracer experiments are used for testing a simple kinematic approach of computing soilwater movements. From the soilwater computations and assuming simple groundwater particle translation, the travel time of meltwater particles from the bottom of the snowcover until they reach a stream is estimated.

Percolation

Percolating water is generally believed to flow preferentially along highly permeable pathways, but field evidence for rapid macropore flow is sparse. At low moisture content the water is stored in small pores. Small spatial heterogeneities cause water particles and pollutants to disperse during the downward movement in the soil. However, field experiments by Zimmerman *et al.* (1967) and Saxena (1984) have shown that in near homogeneous soils, the downward water movement can be described as piston flow, layer by layer. Not all the soilwater is mobile. Biggar and Nielsen (1962) pointed out the existence of domains of mobile and immobile water. The proportion of immobile water is high in soils of high clay content. For glacio-fluvial deposits and till soils in Sweden, Bengtsson *et al.* (1987) found that measured and computed flow rates agreed if the immobile water was 10-15% of the total soilwater content. Since the accuracy of the computation below for glacio-fluvial deposits is not within 10%, the portion of immobile water is not accounted for.

The vertical velocity of individual water particles in the unsaturated zone depends on soil characteristics and moisture conditions. Typically the velocity is a few metres per year. Soil moisture flux corresponds to the unsaturated hydraulic conductivity. A typical value is half a metre per year. Groundwater recharge is the soil moisture flux at groundwater level. Increased flux at root zone level is not immediately recognized as groundwater recharge. A disturbance as increased moisture content and thus increased flux propagates downwards with a propagation rate that depends on the moisture condition in the soil profile. If the soil is initially dry, the propagation rate is close to the particle velocity, but if the initial soil moisture conditions correspond closely to the new flux, the propagation rate is rapidly giving a fast groundwater response to infiltration. A typical speed of a propagating wave is a few metres per month.

The movement of a wetting front in unsaturated soil can be described with the same approach as used by Colbeck (1975) for describing the movement of the meltwater wetting front in a snowpack. The celerity of the front is

$$c \equiv \frac{q-q_i}{\Theta-\Theta_i} \tag{1}$$

where c = kinematic wave speed, q = moisture flux, Θ = moisture content and index i refers to initial conditions.

Since the moisture flux equals the unsaturated hydraulic conductivity which depends on the saturated hydraulic conductivity and on the moisture content, the kinematic wave speed is a function of moisture conditions only, and can be written as a function of moisture flux as

$$c = \frac{K_s^a}{\Theta_s} \frac{q-q_i}{q^a-q_i^a} \tag{2}$$

where K_s = saturated conducitivity, Θ_s = saturated porosity and the exponent a (about 0.1-0.3) relates the unsaturated hydraulic conductivity K and the degree of saturation

$$\left(\frac{K}{K_s}\right)^a \equiv \frac{\Theta}{\Theta_s} \tag{3}$$

The groundwater recharge r at depth h and time t equals the infiltration i at a previous time

$$r(t) = i\left(t - \frac{h}{c}\right) \tag{4}$$

but the wave speed in the vertical changes as higher moisture fluxes catch up lower fluxes, and the position of the groundwater surface may change.

Groundwater Flow

The travel time of water particles that have recharged the groundwater and travel horizontally towards a stream can be computed knowing the saturated hydraulic conductivity, the porosity and the hydraulic gradient. The hydraulic gradient can be roughly approximated by the hill slope. When computing the outflow to the stream, the groundwater storage has to be accounted for. In the present study a storage-discharge model with an entire hillslope segment as control volume was used, allowing as suggested by Sloan *et al.* (1983) the effective storage coefficient to change with time. Continuity requires

$$(\Theta_s - \Theta)\frac{\delta H}{\delta t} + \frac{K_s I H}{L} = r \tag{5}$$

where Θ = soil moisture in the unsaturated zone prior to saturation, I = hillslope, L = length of hillslope, H = groundwater thickness, r = recharge.

Observations

Measurements reported by Saxena (1987), who used oxygen-18 as tracer, and measurements by Dressie (1987), who used the artificial isotope tritium, revealed downward particle velocities of 2-8 mm/day in homogeneous glacio-fluvial deposits and particle velocities ranging from 0 to 40 mm/day where the soil consisted of moraine. Since the percolation was governed by gravity only, it was possible to relate moisture flux to soil moisture content. Having this relationship, it is possible to determine the moisture flux from measurements of soil moisture content. The saturated hydraulic conductivity of the glacio-fluvial deposits was about 5-10 m/d and the porosity 40-50%

The oxygen-18 of precipitation at a given locality varies in a seasonal way de-

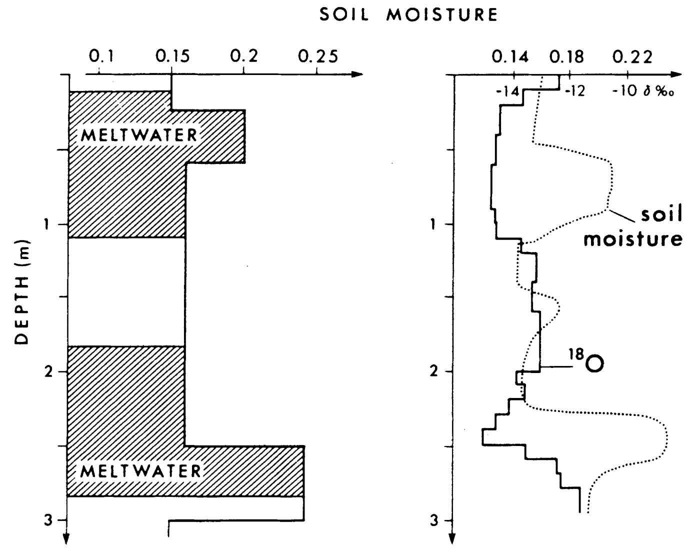

Fig. 1. Left: Computed soil moisture profile and position of meltwater. Right: Measured oxygen-18 (δ, where low δ indicates meltwater) and soil moisture in Uppsala Esker, May 1982.

pending on the condensation-precipitation history of the atmospheric water vapour. The difference between the isotopic composition of precipitation of different seasons is quite pronounced. In Uppsala, Sweden, winter and summer precipitation differ in their oxygen-18 content by about 6-8‰. Meltwater is of light isotopic composition compared with summer rain. When precipitation water penetrates below the upper part of the soil zone it is no longer affected by evaporation. ^{18}O becomes a characteristic property of the water, and was used to verify the kinematic solution of moisture flux in the unsaturated zone.

At a site near Uppsala the moisture flux was found to be related to moisture content as shown in Table 1. Using the kinematic wave approach, the moisture flux and the soilwater distribution over two consecutive years were computed. Since the meltwater was traced using the stable isotope oxygen-18, it was possible to compare the computed results with observed data. Computed moisture distributions and meltwater particle distributions are shown in Fig. 1 and compared with the observed profiles. The good agreement shows that a kinematic approach is sufficient for describing percolation in this type of soil (glacio-fluvial deposits).

Table 1 – Soil moisture flux (mm/d) for different soil moisture content for glacio-fluvial deposits near Uppsala.

Soil moisture	0.10	0.15	0.165	0.18	0.20	0.22
Moisture flux	-	0.5	1	2	3	5

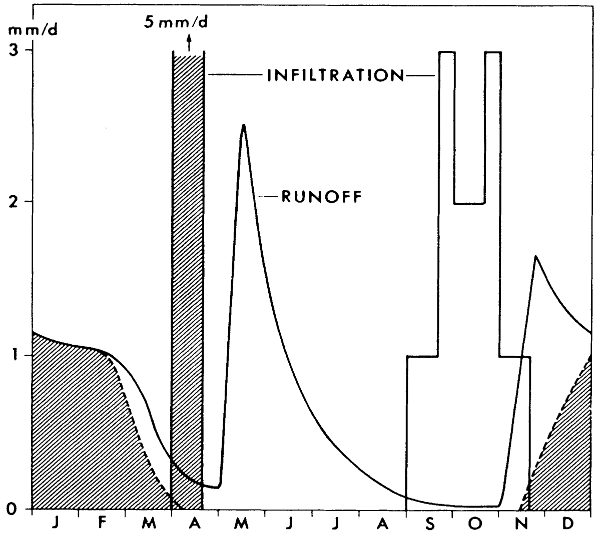

Fig. 2.
Computed subsurface runoff from a hillslope and distribution of meltwater particles (dashed area) in the runoff water.

Simulations

Using the flux-moisture content relationship in Table 1, the kinematic approach was used to compute the percolation resulting from different but typical Swedish snowmelt and autumn rainfall episodes and a groundwater level of 1.5 m below the ground. Following snowmelt the groundwater was computed to be recharged at a rate corresponding to the snowmelt for a period of 10-20 days starting more than a month after initiation of snowmelt. Even when the groundwater level is as high as 1.5 m, very large amounts of snow are required if the meltwater particles should reach the groundwater during or shortly after the snowmelt period. Most meltwater reaches the groundwater at 1.5 m in the autumn if the rainfall is considerable, but it may happen that melt water of one year does not reach the groundwater until during the snowmelt of the next year.

The groundwater outflow from a hillslope was computed by Eq. (5) with the recharge computed from the kinematic routing procedure. When performing the soil moisture computations, account was taken of the variation of the groundwater level. Below 1.5 m depth the soil was considered impermeable. A hillslope representative of glacio-fluvial deposits, where the porosity was 0.42, the hydraulic conductivity 10^{-4} m/s, the slope 0.05 and the length of the slope 50 m was chosen for the computations. The groundwater runoff, when the infiltration rate varied over the year between 0 and 5 mm/d, was computed to a maximum of 2.5 mm/day in the late melt period. The annual recharge was 240 mm. The computed runoff hydrograph is shown in Fig. 2.

The groundwater particle velocity for the given conditions is about 1 m/d. A particle was assumed to move horizontally at this speed when being below the groundwater surface, but when being above to move only vertically downwards. The hillslope was divided into 30,10 m long and 30 mm thick segments, the storage being equal in each vertical, so that all water particles in a segment could move into a downslope segment within a 10-day period. A lower segment in a vertical was assumed to be refilled with water from the upper segment in the same vertical and from a segment in the upslope vertical. In this way it was possible to estimate when recharged groundwater reached a stream at the foot of the hill. The computed distribution of meltwater particles in the subsurface runoff is shown in Fig. 2. The computed water level never reached above 1.3 m from the ground level and the infiltrated meltwater did not start to recharge the groundwater until in November. During the winter all the runoff was found to be meltwater from the previous spring.

Meltwater in the Runoff

By measuring the properties of stream water at the basin outlet it has been concluded in many studies that meltwater constitutes a minor but still a fraction of the runoff water from small basins during snowmelt. In Sweden, Rodhe (1987) has made such studies in several basins. An example of flow separation from Finland using oxygen-18 as tracer is shown in Fig. 3, Lepistö *et al.* (1988).

The kinematic approach and the plug flow assumption used in this paper have been applied to different hillslope conditions. It is found that a prerequisite for large portions of event water in the runoff is that either overland flow occurs or that the slopes are steep with only a thin soil cover above the bedrock. As discussed by Bengtsson (1982) by analyzing daily runoff fluctuations, meltwater can flow in cracks just beneath the frozen ground surface causing large diurnal flow variations. This kind of flow is similar to Hortonian overland flow and does not need to be

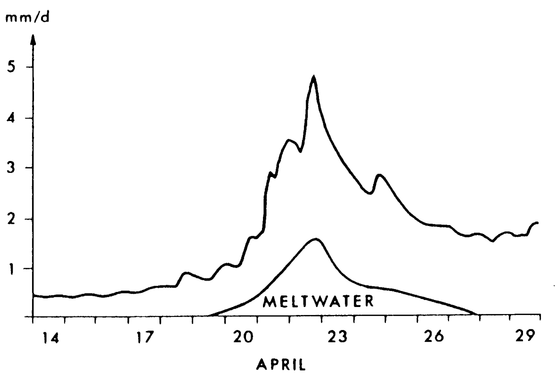

Fig. 3.
Runoff and separated meltwater contribution from the Teeressuonoja basin, April 1985, plotted from data presented by Lepistö *et al.* (1988).

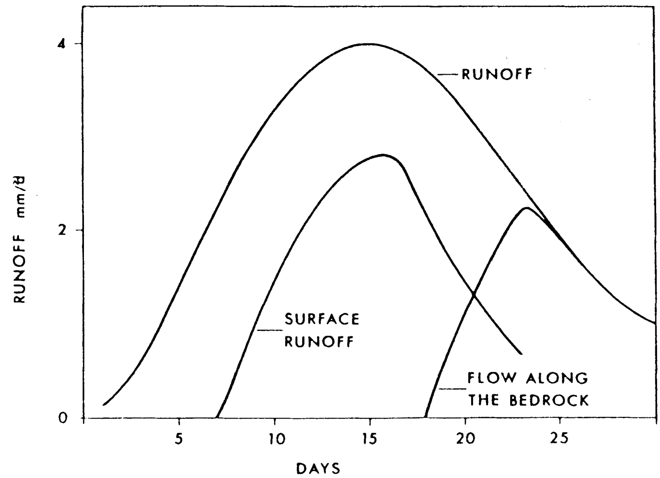

Fig. 4. Theoretical distributions of meltwater particles as compared to the runoff when saturated overland flow or flow along the bedrock takes place.

dependent on the groundwater flow, and therefore is not considered here. Further, it is found that the distribution of meltwater particles in the runoff water depends on the flow conditions. When saturated overland flow occurs the meltwater constitutes a fraction of the runoff and is as highest when the peak flow occurs. When subsurface flow along the bedrock takes place, the old subsurface water is pushed ahead and meltwater constitutes almost all of the runoff at the final phase of the melt period. The two principle distributions are shown in Fig. 4. Comparing with the distribution in Fig. 3, it is seen that overland flow is likely to have been taking place in the Finnish basin. Also Rodhe (1987) found that the extension of wet areas was large in basins where the portion of meltwater in the runoff during snowmelt was large.

Conclusions

The percolation of meltwater can be described by a kinematic approach using continuity and a moisture flux – moisture content relation. When the soil cover is thick and fairly permeable, meltwater particles do not leave a small basin until several months after the melt event. Prerequisites for large fractions of meltwater particles in runoff water during snowmelt are the occurrence of overland flow or of subsurface flow along the bedrock beneath a thin soil cover.

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