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LAND SURFACE AND LAKE STORAGE DURING SNOWMELT RUNOFF IN A SUBARCTIC DRAINAGE SYSTEM

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ABSTRACT

Snowmelt runoff in the Knob Lake basin at Schefferville, Quebec, passes through a complex drainage system composed of rough snow-covered slopes and a chain of lakes connected by poorly developed snow- and ice-filled channels. This study examines the hydrographs of inflow, outflow, and storage for this complex system and identifies the roles of each of the above elements in the snowmelt hydrology of this drainage system.

The land surface elements are the most important component of the drainage system. This is due to the very large storage capacity (5 to 15 cm) on the land surface and the time delay of the runoff from the land surfaces (3 to 4 d). Lake storage is also of importance but is volumetrically less than that on the land surfaces. The resultant discharge hydrograph is a smoothed seasonal hydrograph rather than daily response to the snowmelt.

INTRODUCTION

To date only a few hydrologic studies have been carried out in the Subarctic and Arctic (Findlay, 1966; Dingman, 1971; McCann and Cogley, 1972; Price et al., 1976; Woo, 1980). Of these studies, some (Dingman, 1971; McCann and Cogley, 1972) were concerned with channel-dominated drainage systems. The study by Price et al. (1976) was concerned with snowmelt runoff from small hillslope plots (1800 to 2000 m^2). The studies of Findlay (1966) and Woo (1980) included observations of discharge from the outfalls of lakes in subarctic and arctic drainage systems but did not include detailed study within the drainage basins. The studies of both Findlay (1966) and Woo (1980) showed only seasonal snowmelt flood responses while the other studies listed showed a daily hydrograph response. Woo and Sauriol (1980) observed another characteristic of channel-dominated systems; that is, the presence of ice and snow in the channel blocking the initial part of the snowmelt flood. The observations of Woo and Sauriol (1980) showed that this blockage results in an extremely steep (1 d to peak flow from zero flow) rising limb and high peak flow (6.5 m^2 · sec^{-1} · km^{-1}) for the seasonal snowmelt flood. In the present study snowmelt runoff from a complex drainage system (in which there are deep snowpacks, long gentle slopes and a chain of lakes connected by poorly developed, often snow- and ice-blocked channels) is examined. This basin is typical of the glacially deranged drainage systems in the Canadian Subarctic Lowlands.

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THE STUDY AREA

The study area is located at Schefferville, Quebec, in the Knob Lake research watershed (see Figure 1). The drainage system is composed of a series of lakes, each surrounded by a gently sloping contributing land surface. The lakes are connected by narrow channels with low gradients. Some of the lakes receive runoff from the surrounding land surface only (Easel, Phred, and Houston lakes). Others receive runoff from lakes upstream, as well as the surrounding land surface (Osprey, Malcolm, and Ares) (see Table 1).

The land surface in the study area has a rugged microtopography of small surface depressions and ridges with approximately 1 to 2 m of relief. The depressions are glacially eroded structural weaknesses, small kettle holes, and nivation hollows. Bedrock is ex-

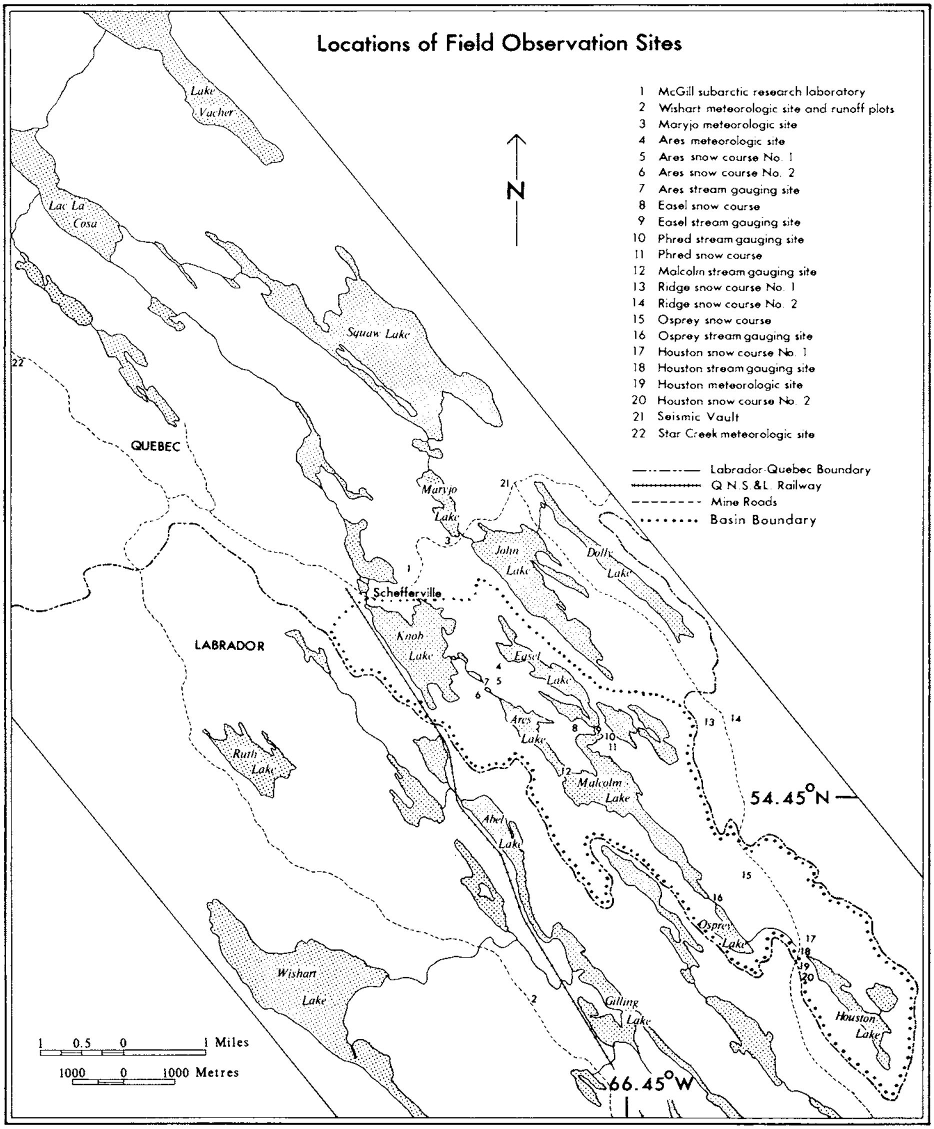

FIGURE 1. The study area.

posed in many places; overburden and soils are shallow. Much of the land is underlain by permafrost. By the end of winter, soils or bedrock are frozen to the permafrost or to depths of 2 or 3 m. Vegetation cover consists of tundra on ridge tops and lichen woodland on lower slopes and valley bottoms. Approximately 70% of the study area has been burned (see Figure 2). The ground is covered by a lichen mat which ranges in thickness from zero in recently burned areas to 20 cm in the lichen woodland.

Snowcover initially forms in late September or early October. Peak accumulation depths of 0.91 to 1.47 m are reached in late March (greater than average accumulation for the Canadian subarctic). The snowcover is most uniform in forested areas and more variable in the tundra and burned areas (FitzGibbon and Dunne, 1979). During the winter there is some redistribution of snow by wind which causes the microtopographic depressions to become filled, creating some very deep snowpacks. These deep snowpacks persist until late into the melt season and often become filled with meltwater. Exposed surfaces and ridge tops are wind swept, and frequently have a very thin snowcover (i.e., 0 to 30 cm).

Ice forms on the lakes in late October or early November and reaches peak thickness (75 to 125 cm) in late March. Ice also forms in the stream channels, constricting some and blocking others. The snowmelt begins in late April or early May and is usually complete by late May or early June. Flows generated by the snowmelt decline to summer flows by late June.

TABLE 1
Land and lake surface areas (km^2)

Basin	Total drainage area	Lake area	Contributing land surface area (CLSA)[a]	Lake area % of total upstream drainage area	CLSA % of total upstream drainage area
Ares	25.1	0.4	1.7	1.6	6.8
Malcolm	23.0	1.7	7.4	7.4	32.2
Easel	3.1	0.7	2.4	22.5	77.5
Phred	3.3	0.6	2.7	18.2	81.8
Osprey	7.4	0.3	1.4	4.0	18.9
Houston	5.7	0.5	5.2	8.7	91.2

[a]Contributing land surface area refers to the surface area contributing runoff directly to a particular lake.

FIGURE 2. The vegetation cover of the study area.

METHOD

The drainage system in the study area has two major components which influence the runoff of snowmelt water: the land surface element and the lakes. Examination of hydrographs of storage, inflow, and outflow for each element provides a means of evaluating their role.

Lake hydrographs were observed directly by measuring lake storage and discharge for each of the lakes in the study area. Discharge hydrographs for the land surface element were not directly observed but are determined through a water balance calculation. The equation is as follows:

$$Q_{ls} = Q_L - (G_L + P_L + E_L + I_{LU} + \Delta S_L) \quad (\mathrm{m^3\,d^{-1}}) \qquad (1)$$

where

Q_{ls} = land surface discharge[1] ($\mathrm{m^3\,d^{-1}}$)
Q_L = lake discharge ($\mathrm{m^3\,d^{-1}}$)
G_L = ground water inflow to the lake ($\mathrm{m^3\,d^{-1}}$)
P_L = rainfall on the lake ($\mathrm{m^3\,d^{-1}}$)
E_L = evaporation (−) or condensation (+) on the lake surface ($\mathrm{m^3\,d^{-1}}$)
I_{LU} = inflow from lakes upstream ($\mathrm{m^3\,d^{-1}}$)
ΔS_L = change in lake storage ($\mathrm{m^3\,d^{-1}}$)

All quantities in the above equation, except evaporation from (or condensation on) the lakes and ground water inflows to the lakes were measured directly. Evaporation (condensation) was calculated using a turbulent transfer equation which assumes a logarithmic wind profile. (For a complete explanation of the method see FitzGibbon, 1977.) Ground water inflows were estimated to be constant at $0.19\ \mathrm{m^3 \cdot km^{-2} \cdot s^{-1}}$. This figure was derived from base flow measurements made prior to the snowmelt season. These compared well with estimates of ground water obtained at mining sites near the study area (Pfleider, 1960).

A subsequent water balance equation was used to determine land surface storage hydrographs as follows:

$$\Delta S_{ls} = (M_{ls} + P_{ls} + E_{ls}) - Q_{ls}\ (\mathrm{m^3\,d^{-1}}) \qquad (2)$$

where

ΔS_{ls} = change in land surface storage[2] ($\mathrm{m^3\,d^{-1}}$)
M_{ls} = snowmelt on the land surface ($\mathrm{m^3\,d^{-1}}$)
P_{ls} = rainfall on the land surface ($\mathrm{m^3\,d^{-1}}$)
E_{ls} = evaporation (−) or condensation (+) on the land surface ($\mathrm{m^3\,d^{-1}}$)

Precipitation on the land surfaces was measured at four sites in and near the study area. The snowmelt was observed at 10 ten-stake snow courses located in the study area. The point measurements of melt were combined with estimates of areal extent of snowcover (determined by a method described by FitzGibbon and Dunne, 1979) to obtain the volumetric measure of snowmelt input. Evaporation (or condensation) for the land surface was estimated with a turbulent transfer equation similar to that used for the lakes. The land surface discharge determined from equation 1 was used as a component of equation 2.

The inflow hydrographs for the land surfaces and the lakes were obtained by summing the inputs to each element as indicated in the water balance equations (i.e., for the land surfaces inflow is equal to $M_{ls} + P_{ls} + E_{ls}$). The hydrographs were determined for the period starting with the first day on which there was a loss of water equivalent from the snowpack due to snowmelt.

RESULTS AND DISCUSSION

The hydrographs of inflow, outflow and accumulated storage for three selected lakes are shown in Figure 3. Summary parameters for all lake hydrographs are shown in Table 2. No diurnal response to snowmelt was observed in any of the basins. In general, storage in the lakes begins at a very low level, and rises to peak on the same day as the inflows and outflows. The exception to this is Malcolm Lake where inflow peaks one day before peak outflow.

In Phred and Houston basins, at the beginning of the melt season, the lake outlet was

[1]Land surface runoff is composed of surface runoff alone. The ground beneath the snowpack remains frozen and impermeable until the bulk of the snowmelt runoff has left the basin.

[2]Land surface storage is composed of surficial depression and detention storage.

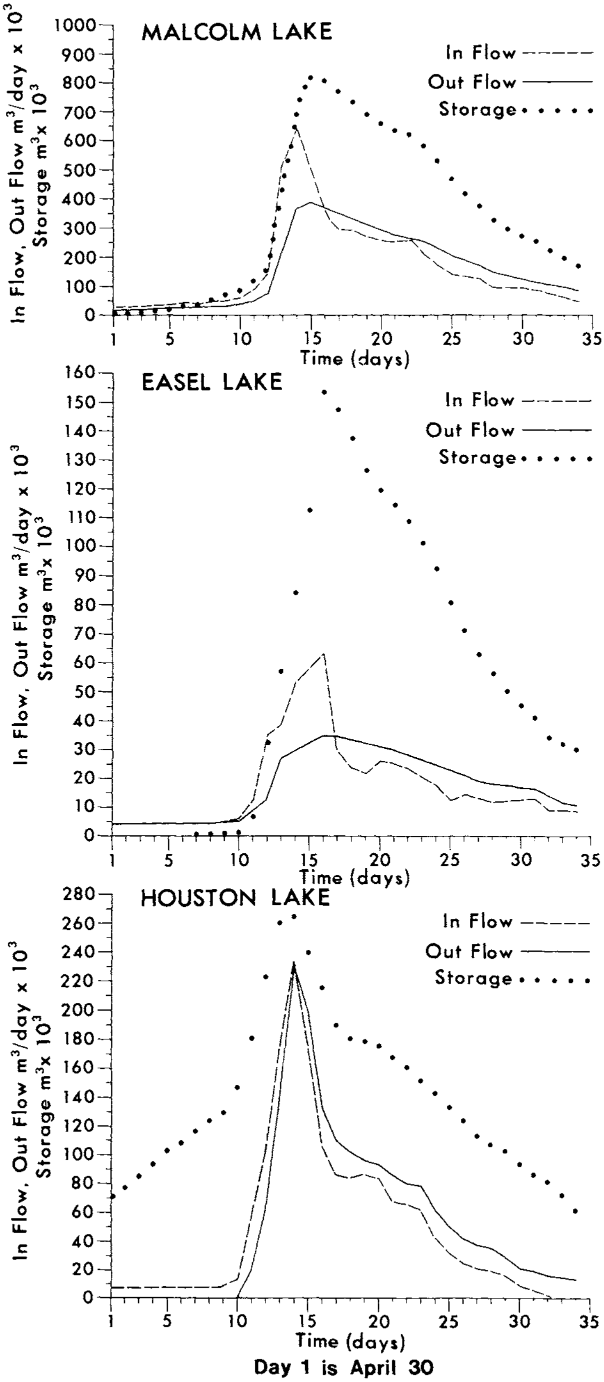

FIGURE 3. The lake hydrographs.

blocked by snow and ice. This held back discharges until 10 d after the beginning of melt, producing a steeper rising limb of the discharge hydrograph and higher peak discharge. The discharges from these lakes were observed to flood the 4- to 5-m deep snowpacks in the outlet channels and then clear the snow from the channels in a form of slush flow, the whole process occurring in the period of one day. A similar phenomenon was observed by Woo and Sauriol (1980). In spite of the ice dams, the peak storage, inflow and outflow occur at approximately the same time as in the lakes with open outlets.

Considering the lakes which were not blocked by ice and snow, it is notable that for the Ares and Osprey lakes, peak inflow and outflow are comparable. This is due to the fact that the contribution of runoff from the land surrounding the lakes is small compared to inflow from basins upstream. In addition, the lake area (in Ares and Osprey basins) represents less than 5% of the total drainage area, thus storage capacity of the lakes is small. The lag time (time difference between the center of mass of the inflow and outflow hydrograph, Linsley et al., 1949) is one day in Ares Lake and zero in Osprey Lake.

In the case of Malcolm and Easel lakes where the lake area is a larger proportion of the drainage area (7 and 22%, respectively) peak outflow is considerably reduced from peak inflow and peak storage volumes are much larger than either peak inflow or outflow. Further, the lag time is 2 d in the case of Malcolm Lake and 4 d in the case of Easel Lake.

The above observations indicate that the role of the lakes is primarily that of storagė. This is effective in smoothing and lagging the hydrographs only where the lakes are a significant portion of the drainage basin (i.e., greater than 5% of the basin area). Snow and ice blocks the outlets only in the smaller more exposed basins. Blockage results in a major modification of the form of the discharge and storage hydrographs but produces negligible effect on peak-flow timing.

TABLE 2
Lake hydrograph peak flow timing[a]

Lake	Peak inflow (day)	Peak outflow (day)	Peak storage (day)
Ares	15	15	15
Malcolm	14	15	15
Osprey	14	14	14
Easel	16	16	16
Phred	14	14	14
Houston	14	14	14

Magnitude of peak inflow, outflow and storage—lakes

Lake	Inflow ($m^3 d^{-1}$)	Outflow ($m^3 d^{-1}$)	Storage (m^3)	Storage cm
Ares	418,000	422,000	187,000	46.8
Malcolm	634,000	382,000	810,000	47.7
Easel	74,000	35,000	153,000	21.9
Phred	82,000	70,000	263,000	43.8
Osprey	254,000	262,000	152,000	50.7
Houston	233,000	230,000	263,000	52.6

Timing[a] of center of mass of lake inflows and outflows

Lake	Inflow (day)	Outflow (day)	Delay (day)
Ares	19	20	1
Malcolm	16	18	2
Easel	15	19	4
Phred	14	17	3
Osprey	18	18	0
Houston	14	18	4

[a]Days from beginning of the melt.

LAND SURFACES

In the Ares and Osprey basins the land surface area contributing runoff directly to the lakes is small relative to the total drainage area of the basin (see Table 1). Thus, land surface storage and discharge represent very small proportions of the water balance in these basins, and are probably subject to large errors. Errors are also probable late in the melt season when land surface contributions to runoff decline to near zero. During the postsnowmelt period, precipitation is largely accounted for by evapotranspiration (Mc-

Cann and Cogley, 1972; Woo, 1980). In the other basins, the land surface area contributing runoff directly to the lake is large, particularly in the Easel, Phred, and Houston basins (see Table 1). Thus the storage and discharge hydrographs are better defined. In view of the possibly large errors, the Ares and Osprey land surface hydrographs are not included in this discussion. Hydrographs of inflow, outflow and storage of selected lake basins are shown in Figure 4. A summary of important parameters for the land surface hydrographs is presented in Table 3.

In the Easel, Phred, Houston, and Malcolm basins peak land surface storage occurs 1 d after peak inflow and 2 to 3 d before peak outflow (see Table 3). The time difference between occurrence of peak inflows and peak outflows is 3 to 4 d. It is notable that the magnitude of the peak land surface storage in these basins ranges from 119 to 303% of the peak lake storage. In addition, the magnitude of the peak land surface storage represents a large proportion (from 44 to 74%) of the total runoff during the period of observation.

The very large land surface storage capacity (5 to 15 cm) is created by the numerous microtopographic depressions. In addition, the presence of deep snowpacks provides for storage during the melt, in excess of that created by the topography. The irreducible storage of water in the snowpack is estimated at 3% by volume (Gray, 1970). Further, water is stored in the saturated layer at the base of the snowpack. On a well-drained site the depth of this saturated zone may be as little as 0.5 cm, but in depressions water can saturate the entire thickness of the snowpack (which can be as much as 300 cm in depressions). Both the storage in the unsaturated zone and that in the saturated layer are temporary storage over and above that of the normal depression storage of the land surface with no snowcover.

The delay of peak outflow from peak inflow is in marked contrast to the absence of delay observed in the lakes' response. The land surface hydrograph delays suggest that it takes water 3 to 4 d to flow off the land surfaces. By dividing the average slope length by the time delay between the peak inflow and peak outflow the average velocity of flow at peak discharge is found to be 4.3 m h^{-1}. This compares with an average velocity of 5 m h^{-1} measured by Price (1975) on small plots with little depression storage.

TABLE 3

Land surface hydrographs' peak flow timing[a]

Contributing land surface	Inflow (day)	Outflow (day)	Storage (day)
Ares	13	14	13
Malcolm	11	14	12
Easel	11	15	12
Phred	11	14	12
Osprey	11	13	13
Houston	11	14	12

Peak discharges—Land surfaces

Contributing land surface	Inflow ($m^3 d^{-1}$)	Outflow ($m^3 d^{-1}$)	Storage (m^3)	Storage (cm)
Ares	39,000	23,000	137,000	8.1
Malcolm	263,000	239,000	1,157,000	15.6
Easel	71,000	69,000	367,000	15.3
Phred	99,000	77,500	313,000	11.6
Osprey	55,000	38,000	73,000	5.2
Houston	199,000	226,000	799,000	15.4

Timing[a] of centers of mass of land surface inflows and outflows

Contributing land surface	Inflow (day)	Outflow (day)	Delay (day)
Ares[b]			
Malcolm	11	14	3
Easel	11	15	4
Phred	11	14	3
Osprey[b]			
Houston	11	14	3

[a]Timing days from the beginning of the melt.
[b]Not determined.

It may also be noted that the delay between the center of mass of the inflow and outflow hydrograph is the same as the delay for peak flows. This suggests that there is no significant change in the velocity of flow from average to peak discharge levels. This would be expected since the flow is through the saturated snowpack and lichen mat. Under these conditions (Darcy flow) the factors governing the velocity of flow are the hydraulic gradient and the permeability of the saturated layer. Price et al. (1976) in modeling runoff from a snowpack used a constant snow permeability of the satu-

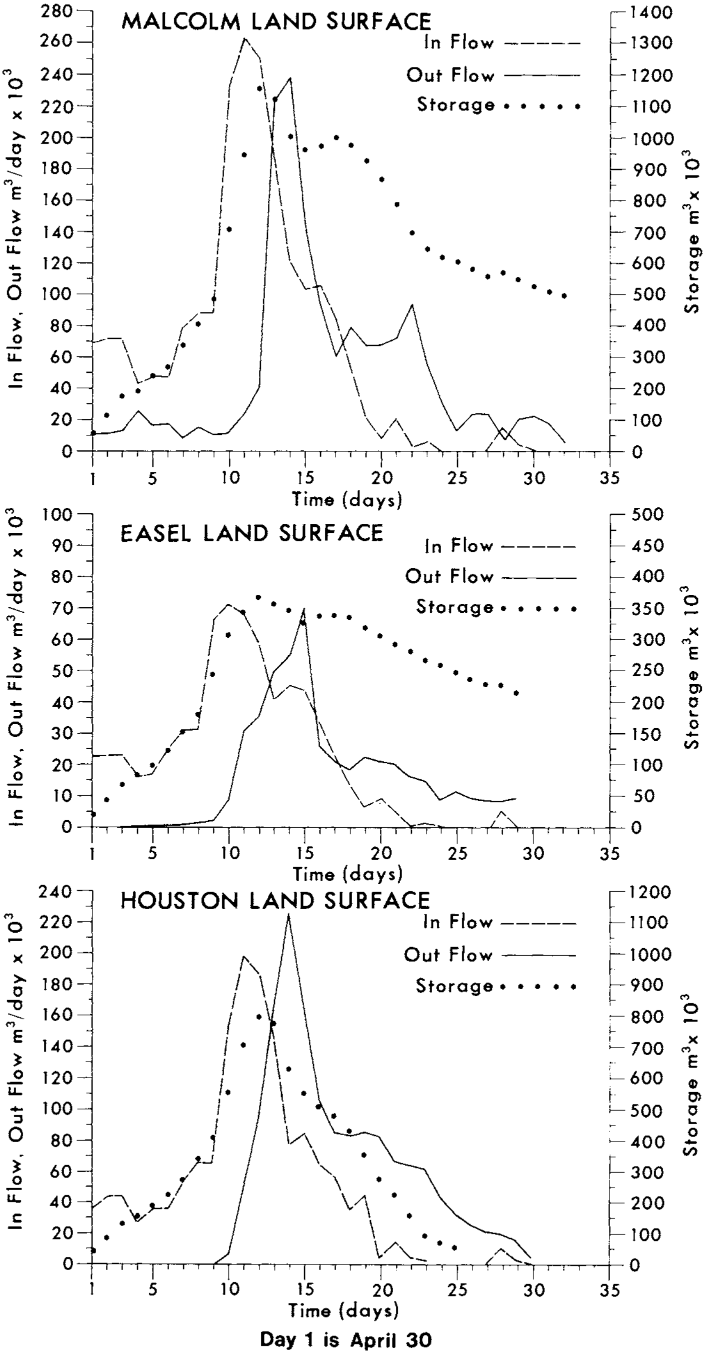

FIGURE 4. The land surface hydrographs.

rated layer with good success. In addition, the slope of the hydraulic gradient was taken to be the slope of the land surface. The lack of a difference between the hydrograph delays for peaks and centers of mass confirms that these approximations are reasonable.

CONCLUSIONS

The most notable discovery of this study is the identification of the importance of the land surface element of the drainage system. The land surfaces have considerable storage capacity which is enhanced by the presence of deep snowpacks. They cause smoothing and attenuation of the surface discharge hydrographs. The action of the lakes is primarily that of storage which causes further smoothing of the flood wave. The amount of smoothing caused by a lake is a function of the size of the lake relative to drainage area.

The role of channel ice and snow in altering the snowmelt flood has been noted. It is of particular importance in basins with small drainage areas where winter flows are insufficient to prevent the massive accumulation of snow and ice in the channel. The mechanisms by which snow is cleared from the channels of ice-blocked lakes is deserving of further investigation.

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