RAIN ON SNOW: SHALLOW, TRANSIENT SNOWPACKS WITH FROZEN SOILS

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INTRODUCTION

Rain-on-snow events can contribute significantly to flood damage and may contribute significantly to annual water yield in some Pacific Northwest basins. Severe winter flooding occurs frequently and has produced some of the highest peak discharges recorded at some runoff stations (Johnson and McArthur, 1973). In the Columbia Plateau major land resource area (MLRA) B8 and elsewhere in the Pacific Northwest, these runoff events usually occur from December through March. MLRA B8 encompasses an area of approximately 42,730 km^2 (16,500 mi^2) of which more than 90% is in farms or ranches. Elevation ranges from 400 to 1,100 m (1312 to 3609 ft.). The topography ranges from nearly level to steeply sloping (USDA, 1981). Approximately 41 percent of the cropland has slopes in excess of 7 percent (Allmaras et al., 1979). The winter runoff that occurs on these steeply sloping croplands not only contributes to flooding, but also results in serious soil erosion and sediment production.

The study reported here was conducted in the southern portion of the Columbia Plateau (Figure 1), an area which is characterized by humid winters and dry summers with relatively low rainfall intensity, usually less than 4 mm (.15 in) per hour (Horner et al., 1957).

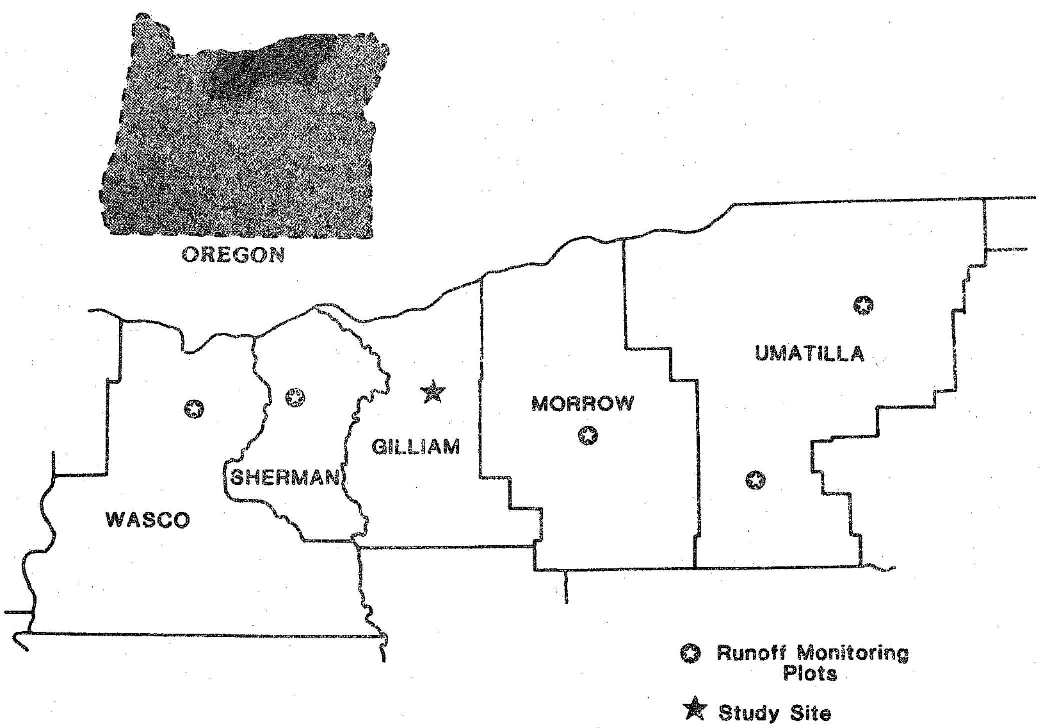

Figure 1. Location of runoff monitoring plots and study site in northeastern Oregon.

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Average annual precipitation is 327 mm (12.86 in) of which 65% occurs during the November through April period (average of 4 stations). Another feature of the winter climate is a shallow, transient snowpack which is subject to several accumulation and melt cycles each winter. However, this snowpack variation has not been described. Frozen soils are common and occur nearly every year. It has been estimated that 85 to 90 percent of the erosion hazard and hence runoff occurs during the winter period (Wischmeier and Smith, 1978; McCool et al., 1982). Because these runoff events have both short and long term economic consequences, more information about them is necessary. In this study we compare two snowmelt periods which occurred under different meteorological conditions. Antecedent soil and snow conditions for each of these two snowmelt periods were nearly identical.

INSTRUMENTATION AND DATA

Widely scattered sites in the dryland small grain producing area of northeastern Oregon (MLRA's B7, B8, and B9) have been instrumented each winter since 1980 to measure runoff and soil erosion. This study is designed in part to evaluate the climatic mechanisms and antecedent conditions responsible for the production of upland runoff and soil erosion. Instrumented sites are installed in newly seeded winter wheat (*Triticum aestivum* L.) after summer fallow in each of the northeastern Oregon counties shown in Figure 1. The sites are usually installed prior to 1 November and removed in April or May of the following year. Two holding tanks at each site are designed to collect all of the runoff from duplicate plots each 18.3 x 3.05 M (60 x 10 ft.). These holding tanks are equipped with continuously recording water level indicators. Each site is also instrumented to continuously record soil temperature at 5 cm (2 in), air temperature and relative humidity at the 1 m height, precipitation amount and intensity, and incident solar radiation (0.36 to 2.5 µm). Frost depth, snow depth, and snow water equivalent are measured weekly.

In four winter seasons 48 runoff events have been observed. Twenty of the 48 events (42%) were produced by either rain-on-snow or radiative melt; frozen soils and snowmelt were also involved in 17 of these 20 events. In the 1979-80 winter, snowmelt caused 12 out of the 14 runoff events, and frozen soils were present in 10 of these 12 events (Zuzel et al., 1982).

METHODS

From the extensive data base described in the preceeding section, we chose two snowmelt periods with nearly identical antecedent snow and soil conditions. The site characteristics and antecedent conditions are shown in Table 1. The test site is located in Gilliam County (Figure 1). To analyze these events we used an hourly snowmelt model of the form:

$$M = Q_N + Q_H + Q_E + Q_P \quad (1)$$

where: all terms are expressed in millimeters of melt; M is total melt, Q_N is radiative melt, Q_H is sensible heat melt, Q_E is latent heat melt, and Q_P is melt produced by rain water.

Table 1. Site characteristics and antecedent conditions at the Gwendolyn, Oregon site; 1980 and 1982.

Latitude: 45.35°N
Slope: 16.38°
Aspect: N28°E
Elevation: 915 m

Antecedent condition	12 January 1980	13 January 1982
Snow water (mm)	31.2	27.5
Frost depth (cm)	7.6	3.8
Soil temperature (°C)	-1.1	-0.3

The model assumes that during a rain-on-snow event, 1) direct solar radiation is insignificant, 2) snow surface temperature is 0°C (32°F), 3) rainwater temperature is equal to dewpoint temperature, and 4) the snowpack is isothermal at 0°C (32°F). Physical representations of the terms in Equation 1 were formulated from several sources (Corps of Engineers, 1956; Allen, 1974; Anderson, 1976; and Marks, 1978) and are listed in the appendix. During non-rain events, the model uses only the assumption that the snow surface is 0°C. Single-valued input variables are elevation, initial snow-water equivalent, albedo, and average windspeed during the event. Hourly values of air temperature, relative humidity, precipitation, and direct radiation are also required inputs. Calibration for the period is accomplished by varying the albedo and windspeed parameters. For 12 January 1982 we used a windspeed of 21.4 km/hr (13.3 mph) and for 13 January 1982 we used a windspeed of 4.3 km/hr (2.7 mph). These wind estimates are realistic because the National Weather Service station at Pendleton, Oregon reported winds in excess of 74 km/hr (46 mph) on 12 January 1980 while 13 January 1982 was a relatively calm day. Also, windspeeds measured at the Columbia Basin Agricultural Research Center averaged 21.4 and 4.4 km/hr (13.3 and 2.7 mph) for 12 January 1980 and 13 January 1982, respectively. Measured solar radiation values were corrected for slope and aspect using the algorithm described by Swift (1976).

RESULTS AND DISCUSSION

The results of the simulation of the 1980 and 1982 periods using Equation 1 are shown in Table 2. Both simulations covered an identical time period of 0500 to 2000 hrs. While

Table 2. Simulation results of snowmelt periods at Gwendolyn, Oregon; 12 January 1980 and 13 January 1982.

	12 January 1980	13 January 1982
Average site conditions		
Albedo (%)*	0.4	0.4
Windspeed (km/hr)*	21.4	4.3
Windspeed (km/hr)**	21.4	4.3
Air temp (°C)	4.0	4.3
Dewpoint (°C)	2.1	-1.5
Relative humidity (%)	88	66
Vapor pressure (mb)	7.1	5.5
Contribution to melt (mm)		
Net radiation (Q_N)	3.1	0
Sensible heat (Q_H)	16.9	0.5
Latent heat (Q_E)	9.3	-1.1
Precipitation (Q_P)	0.4	0
Total ablation (M)	29.6	1.6
Contribution to runoff (mm)		
Precipitation	10.7	0
Melt + precipitation	40.3	0.5
Measured runoff	40.1	0

* Calibration parameters
** Measured at Pendleton Experiment Station at 0.45 m

average air temperatures for both periods were nearly identical, there were major differences in average dewpoint temperature and windspeed. These two conditions caused differences in the sensible and latent heat components between the two snowmelt periods. Hourly air temperatures for both periods are shown in Figure 2; hourly dewpoint temperatures in Figure 3; and simulated hourly ablation for the two periods in Figure 4.

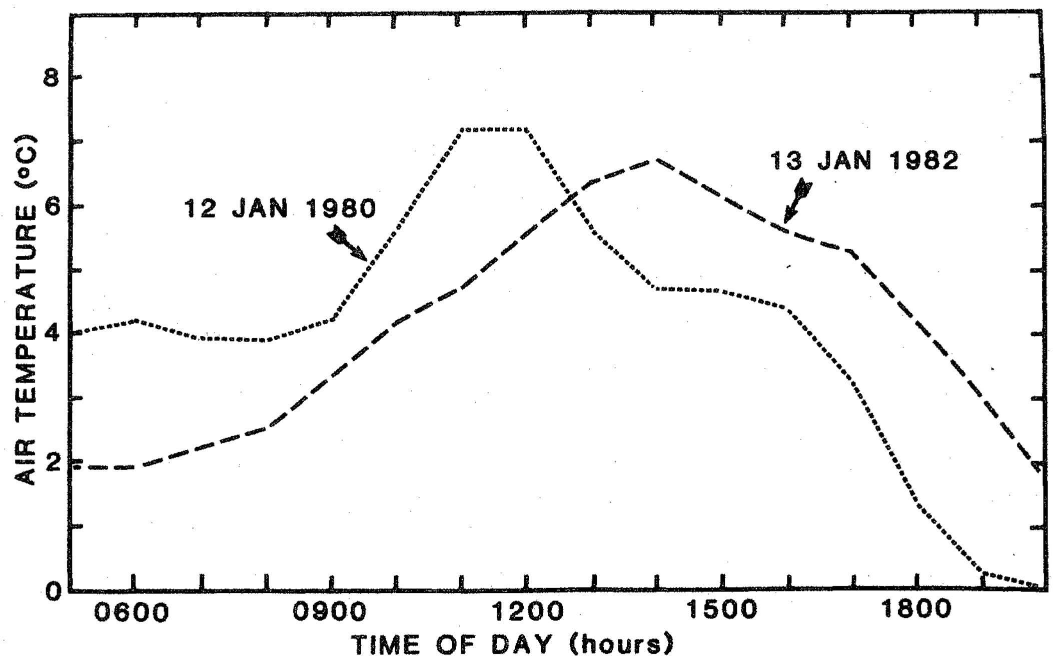

Figure 2. Air temperature regime during the snowmelt events of 12 January 1980 and 13 January 1983.

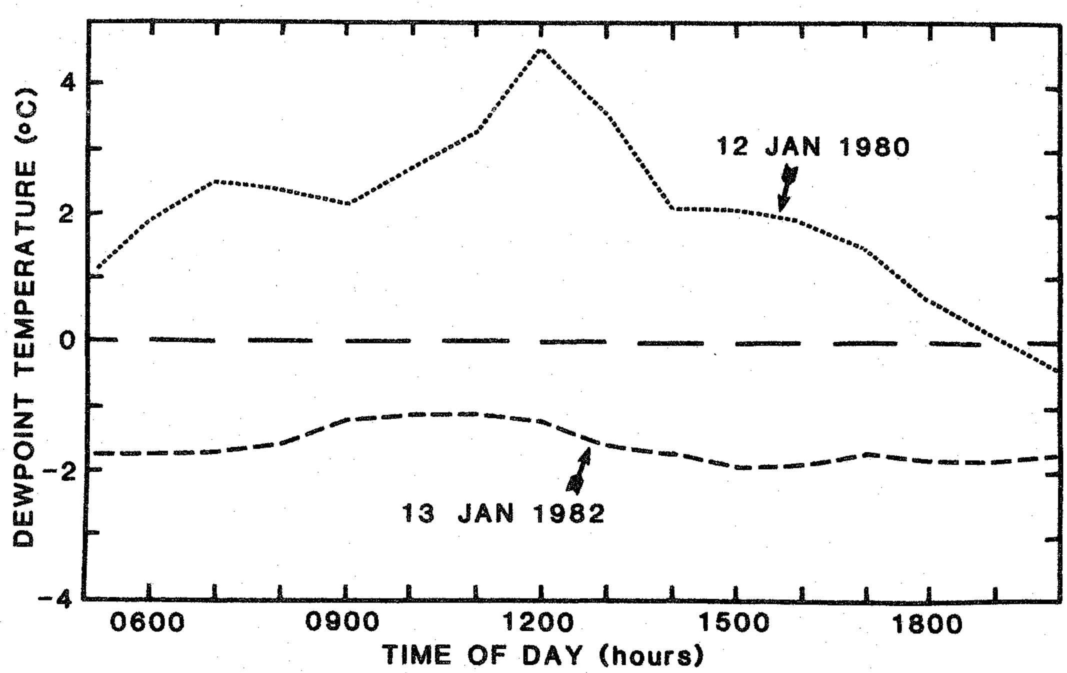

Figure 3. Dewpoint temperature regime during the snowmelt events of 12 January 1980 and 13 January 1982.

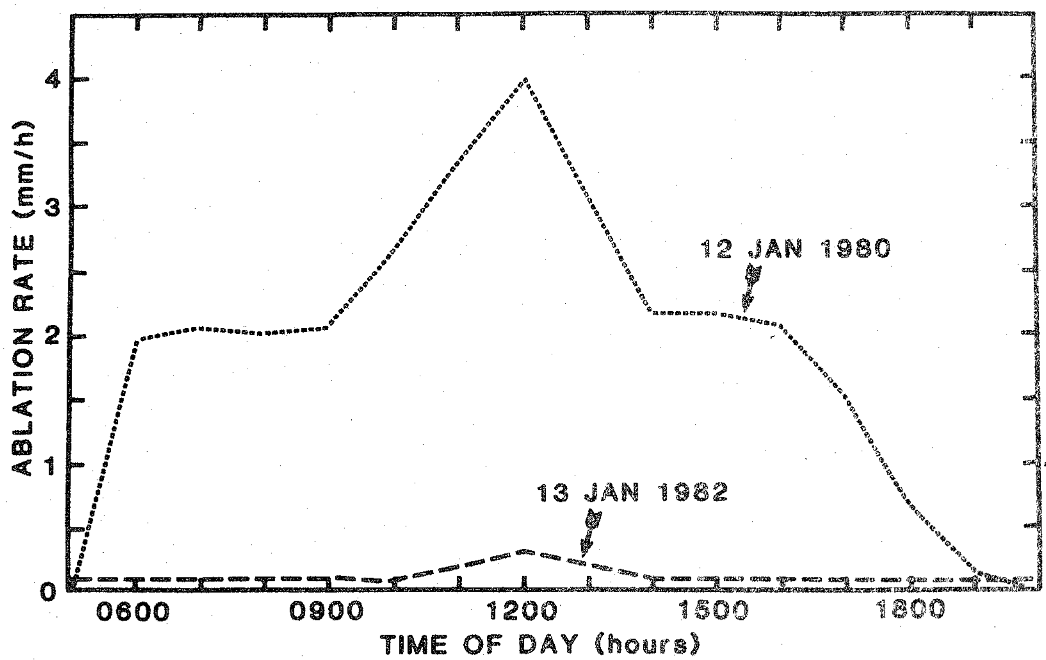

Figure 4. Ablation rate during snowmelt events of 12 January 1980 and 13 January 1982.

Meteorological and soil conditions preceeding the initial simulation time of 0500 hrs were also quite different for the two events. In 1980 the site had been snow covered; a snow water equivalent of 31.2 mm (1.22 in) had accumulated over a 3-day period. Soil temperature at 5 cm (2 in) was -1.1°C (30°F). Measured frost depth was 7.6 cm (3 in) with no evidence of surface thawing (Table 1). Average air temperature for the preceeding week was -5°C (23°F); daytime maximums remained below 0°C (32°F). A warm, moist Pacific air mass entered the area about 2300 on 11 January. Within one hour the air temperature rose from -4°C (25°F) to 4°C (39°F). Runoff began at 0700 hrs and continued until 1900 hrs on 12 January. This same weather sequence in combination with frozen soils also produced runoff at all other monitoring sites, at all of which the sum of snowmelt and rainfall nearly equaled runoff. Consequently, little if any water infiltrated as typified in Table 2.

For the 1982 period, the site had also been snow covered; a snow water equivalent of 27.5 mm (1.08 in) had accumulated during 3, 4, 5 January. Soil temperature at 5 cm (2 in) was -0.3°C (31°F). Measured frost depth was 3.8 cm (1.5 in), but the soil thermograph did show some evidence of surface thawing during daylight hours as expected by the antecedent soil temperature of -0.3°C (Table 1). Average air temperature during the preceeding week was -2.3°C (28°F). However daytime maximums of up to 6.1°C (43°F) were recorded. The continuous temperature trace indicated a very gradual warming trend. No surface runoff was observed for this period at any of the six monitoring sites.

We partitioned the snowmelt hydrographs shown in Figure 4 into the individual melt components. For 1980, the sensible heat component produced 58% of the total melt; latent heat, 31%; net radiation, 10%; and precipitation, 1%. For 1982, the sensible heat component produced all of the melt; sensible heat also provided the energy for evaporation. From the hourly energy fluxes shown in Figure 5, we can typify this period as a "cold" melt period even though the air temperature remained quite high. Because of slope steepness and orientation (Table 1), shortwave radiation input is very small (72 w/m^2 measured). The energy balance is largely negative and is dominated by longwave exchanges. Convective sensible heat (Q_H) was the only heat source for the melt produced from 1000 to 1400 hrs, although evaporation occurred throughout the event. Because the energy balance was negative for most of the time period, it is likely that any melt water produced at the surface either evaporated or was re-frozen within the snowpack. During the five-day period preceeding this event, meteorological conditions were very similar, but no runoff was observed.

Conversely, the hourly energy fluxes shown in Figure 6 typify a "warm" melt period, although the air temperature regime is very similar to that in 1982. In this 1980 period, shortwave radiation input was negligible because of cloud cover, and all terms in the energy balance were positive. Convective sensible heat was the dominant heat source; however the calculations show that condensation melt (Q_E, positive) accounted for up to 36% of the total hourly melt and averaged 31% for the entire event. The maximum melt rate was 4 mm/h (.16 in/h) with an average melt rate of 2.1 mm/h (.09 in/h). Because the soil was deeply frozen, the infiltration was severely curtailed, and nearly all of the available surface water ran off. Using Equation 1, we calculated a total ablation of 29.6 mm (1.17 in). Measured precipitation for the event was 10.7 mm (0.42 in), which when combined with ablation nearly equalled measured runoff for the period (Table 2).

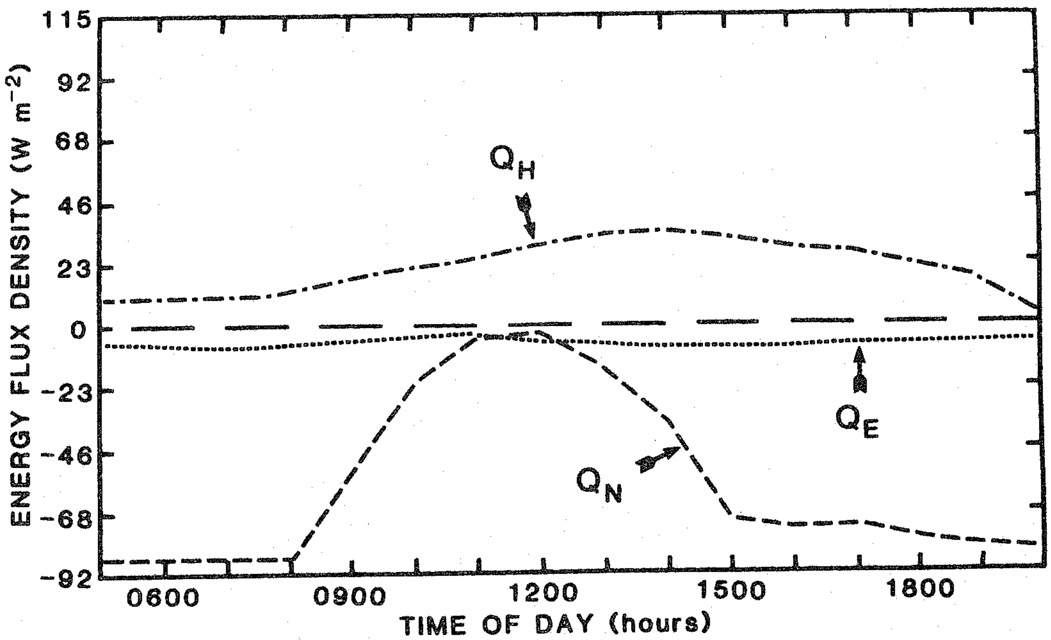

Figure 5. Calculated energy flux at the snow surface during the 13 January 1982 snowmelt event (energy flux components are described in Equation 1).

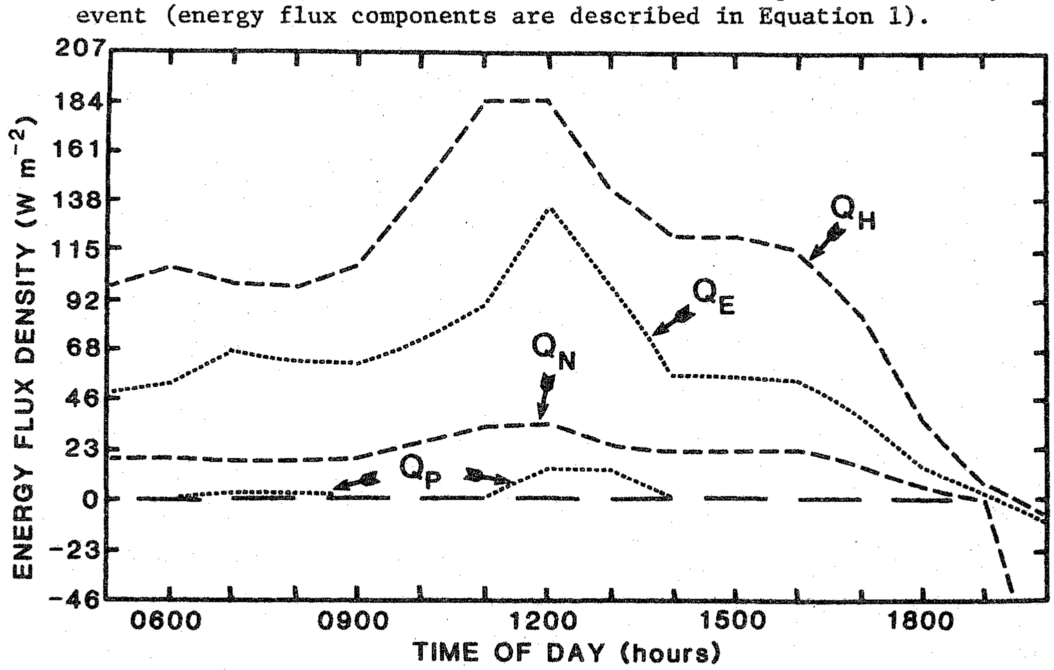

Figure 6. Calculated energy flux at the snow surface during the 12 January 1980 snowmelt event (energy flux components are described in Equation 1).

SUMMARY AND CONCLUSIONS

The data strongly suggest that rain on a shallow snowpack and a frozen soil are the two key factors for the production of accelerated upland runoff. Energy balance calculations show that condensation melt (Q_E, positive) is one of the major mechanisms responsible for accelerated snowmelt. High condensation melt rates associated with the intrusion of warm, moist Pacific air masses, accelerate the snowmelt rate beyond that expected from radiation, sensible heat and heat content of the rain. High windspeeds during these events provide the turbulent transfer mechanism necessary for both condensation and sensible heat melt.

The "cold" snowmelt situation, typified by the 1982 example is unimportant in the upland runoff process. Although air temperature is quite high the hourly energy balance is largely negative and only small amounts of meltwater are produced. Windspeeds are low, hence turbulent transfer rates are low. Furthermore, since the snowpack is cold, the melt is probably refrozen in the pack making evaporative melt (Q_E, negative) the major mechanism in ablation. Over a period of days, this could be a very significant loss. These data also suggest that degree-day or melt-factor predictive methods are not applicable to the shallow, transient snowpack, because any air temperature based method would tend to predict the same amount of melt for both types of events.

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APPENDIX

1. Physical representations used in Equation 1.

$$Q_N = Q_S(1.0-A)+(T+273)^4(6.12)(10^{-10})[1.24\ (\frac{e_a}{T} + 273.16)^{1/7}]-3.41$$

$$Q_H = PTW\ (1.41)(10^{-5})$$

$$Q_E = W\ (e_a-6.11)(2.8)(10^{-2})$$

$$Q_P = RT_d(1.25)(10^{-2})$$

2. List of symbols

Q_N Net radiative melt (mm)

Q_H Sensible heat melt (mm)

Q_E Latent heat melt (mm)

Q_P Precipitation melt (mm)

Q_S Incoming shortwave radiation ($cal/cm^2/hr$)

A Albedo (dimensionless fraction)

T Air temperature (^oC)

e_a Vapor pressure of the air (mb)

P Atmospheric pressure (mb)

W Average windspeed (km/hr)

R Rainfall amount (mm)

T_d Dewpoint temperature (^oC)