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Climate variability, snow, and physiographic controls on storm hydrographs in small forested basins, western Cascades, Oregon

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Abstract:

Large floods are often attributed to the melting of snow during a rain event. This study tested how climate variability, snowpack presence, and basin physiography were related to storm hydrograph shape in three small (<1 km^2) basins with old-growth forest in western Oregon. Relationships between hydrograph characteristics and precipitation were tested for approximately 800 storms over a nearly 30-year period. Analyses controlled for (1) snowpack presence/absence, (2) antecedent soil moisture, and (3) hillslope length and gradient. For small storms (<150 mm precipitation), controlling for precipitation, the presence of a snowpack on near-saturated soil increased the threshold of precipitation before hydrograph rise, extended the start lag, centroid lag, and duration of storm hydrographs, and increased the peak discharge. The presence of a snowpack on near-saturated soil sped up and steepened storm hydrographs in a basin with short steep slopes, but delayed storm hydrographs in basins with longer or more gentle slopes. Hydrographs of the largest events, which were extreme regional rain and rain-on-snow floods, were not sensitive to landform characteristics or snowpack presence/absence. Although the presence of a snowpack did not increase peak discharge in small, forested basins during large storms, it had contrasting effects on storm timing in small basins, potentially synchronizing small basin contributions to the larger basin hydrograph during large rain-on-snow events. By altering the relative timing of hydrographs, snowpack melting could produce extreme floods from precipitation events whose size is not extreme. Further work is needed to examine effects of canopy openings, snowpack, and climate warming on extreme rain-on-snow floods at the large basin scale.

KEY WORDS climate; flooding; rain-on-snow; hydrograph; Oregon

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INTRODUCTION

Rain-on-snow events, in which snowmelt may augment rain inputs, are responsible for extreme floods in the Pacific Northwest of the USA and in other regions, initiating landslides, structuring aquatic ecosystems and the riparian zone, and flooding downstream cities (Swanson *et al*., 1998, Wemple *et al*., 2001). Most snowpacks in the Pacific Northwest are 'warm' (near 0 °C) (Smith 1974), and snowmelt affects storm hydrograph shape, including response time, peak, and volume (USACE, 1956; Harr, 1981, 1986; Marks *et al*., 1998). Yet, despite the importance of rain-on-snow events in local and regional flooding, relatively little is known about how climate, snowpack dynamics, and basin physiography interact to determine hydrograph shape.

Climate factors (temperature, wind, precipitation timing and phase) determine antecedent soil moisture, input of moisture to a basin, and moisture storage as snowpack or in the soil. During a storm event, climate factors determine how much snowmelt occurs, and thus mitigate or amplify the role of the snowpack in storm hydrograph timing, peak magnitude, and volume. Given the right conditions (warm air temperatures, high wind speeds, and abundant rain), snowmelt is calculated to augment a flood peak by as much as 30% (Harr, 1981). Plot-scale measurements confirm that snowmelt contributed 40 to 50% of water delivered to soils in canopy gaps during an extreme flood (Marks *et al*., 1998). When snowmelt is rapid, storm hydrograph rise may be steeper than during rain-only events, deliver more volume, and have a higher peak discharge (Rothacher *et al*., 1967; Harr, 1981). Snowmelt is estimated to make a larger relative contribution to the hydrograph during small compared with large precipitation events (Harr, 1981).

Basin physiography (slope length, steepness) also control storm hydrographs by determining the effective hillslope length that contributes water during an event (Langbein, 1940; Rantz, 1971; Rodriguez-Iturbe *et al*., 1982). In steep forested basins, where water infiltrates vertically very rapidly, hydrographs may respond very rapidly to precipitation, because subsurface flowpaths control storm hydrograph response either by lateral flow in preferential flowpaths, or as vertical 'pressure waves' through the soil matrix (Torres *et al*., 1998; Montgomery and Dietrich, 2002b; Montgomery *et al*., 2002; Torres, 2002; McDonnell 2003; McGuire and McDonnell, 2006).

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By creating a saturated, but ephemeral, layer draped on the landscape, a warm snowpack may significantly alter hydrologic response. Prior studies at the H.J. Andrews Forest indicate that snowpack effects on the storm hydrograph appear to differ according to hillslope length, hillslope gradient, and basin aspect: although increased snowpack augmented peaks (Harr, 1986), logging had little effects on peaks in the seasonal snow zone (Harr *et al.*, 1982), but in basins with short, steep slopes, the presence of a snowpack delayed the hydrograph compared with rain events (Harr and McCorison, 1979).

Care is required when inferring the effect of snowpacks on whole-basin storm hydrographs in the Pacific Northwest. First, nearly all observations and models of snowpack dynamics have been conducted at the plot scale and in canopy gaps, rather than basin-wide and in forests. Energy balances under forest cover promote less snow accumulation and slower melt than in gaps (Coffin and Harr, 1992; Storck *et al.*, 2002); in the regional flood of 1996, snowmelt was five times greater in canopy gaps than under forest stands (Marks *et al.*, 1998). Second, snowpacks are heterogeneous (Anderton *et al.*, 2004) as well as very dynamic (Marks *et al.*, 1998); snow depth may vary by several orders of magnitude within small basins (10–100 ha). However, climate records and distributed hydrologic models can be used to simulate snowpack conditions retrospectively, at least in relative terms, and used to classify storm events.

In the H.J. Andrews Experimental Forest, which spans the rain, transient snow, and seasonal snow zones of the western Cascade Range of Oregon, long-term streamflow records provide an opportunity to examine snowpack effects on storm hydrographs in small, forested basins. Using hydrographs from nearly 800 storm events over a 30-year period from three small basins, we examined how measures of storm hydrograph shape—timing, peak, and volume—differ based on estimated snowpack and soil moisture conditions among basins with varying physiography, under mature and old-growth forest.

METHODS

Logic of the study

This study undertook a quantitative analysis of storm hydrographs produced under varying conditions of antecedent soil moisture, snowpack size, and basin physiography from three small, forested basins in low-, intermediate- and high-elevation portions of the Andrews Forest. Specifically, the work of Harr (1981, 1986) was extended to test (1) whether the presence of snow speeds the timing, increases the volume and the peak of storm hydrographs compared with rain-only conditions, (2) whether short, steep slopes produce faster hydrograph rise and higher peaks relative to long or gentle slopes, and (3) how antecedent soil moisture influences these relationships.

The analysis involved several steps: (1) basin-wide soil moisture and the presence/absence of a snowpack were retrospectively estimated using a distributed-parameter hydrologic model (MMS); (2) peak discharge events and associated precipitation were extracted from streamflow and climate records using an automated procedure (GetPQ); (3) peak discharge events were classified using the modelled antecedent soil moisture and snowpack data; and (4) storm hydrograph properties were analysed based on event type and basin type.

Study site description

The Pacific Northwest climate produces storm hydrographs associated with order-of-magnitude variations in antecedent soil moisture and snowpack conditions, and instrumented basins encompass two-fold variation in hillslope gradients and slope lengths. Study sites were three small control basins from three paired-basin experiments in the Andrews Forest, western Oregon (Figure 1). Watershed 9 (9 ha, WS 9) is a steep, first-order basin in the transient snow zone ranging in elevation from 425 to 700 m; the basin-wide average hillslope gradient exceeds 60%, and median slope length is 60 m. At Watershed 9, snow rarely persists longer than 1 to 2 weeks (Harr and McCorison, 1979; Perkins, 1997). Watershed 2 (60 ha, WS 2) is a steep, second-order basin with longer slopes in the transient snow zone ranging in elevation from 530 to 1070 m; the basin-wide average hillslope gradient exceeds 50%, and median slope length is 150 m. Watershed 8 (21 ha, WS 8) is a gentle-gradient, first-order basin with short hillslopes in the seasonal snow zone ranging in elevation from 960 to 1130 m; the basin-wide average hillslope gradient is 25%, and median slope length is 90 m. Snow depth may exceed 1·5 m and snow may persist for 6 months at WS 8 and in the upper elevations of WS 2 (Harr and McCorison, 1979).

Study site climate is Mediterranean with wet winters and dry summers. Mean annual precipitation is 2200 mm, and >80% of precipitation occurs from November to April. Precipitation is highest along the southern ridge of the Lookout Creek basin, which creates a rainshadow to the northeast, so that precipitation is almost identical at the high- and low-elevation basins. Mean daily minimum temperatures fall below freezing between early December and mid-February and vary little between the high- and low-elevation basins because of inversions and cold air drainage in clear periods and air mixing during stormy periods (Smith, 2002). Soil parent materials are weathered Tertiary or Quaternary volcanic rocks with some glacial deposits (Sherrod and Smith, 1989). Soil infiltration capacities of 20 cm h^{-1} are orders of magnitude higher than precipitation rates, and overland flow rarely occurs (Dyrness, 1969; US Forest Service, 1961). Vegetation consists of mature to old-growth conifer forest with leaf area indices exceeding 8·0 (Marshall and Waring, 1986). Forests are dominated by Douglas-fir (*Pseudotsuga menziesii*), western hemlock (*Tsuga heterophylla*), and western red cedar (*Thuja plicata*).

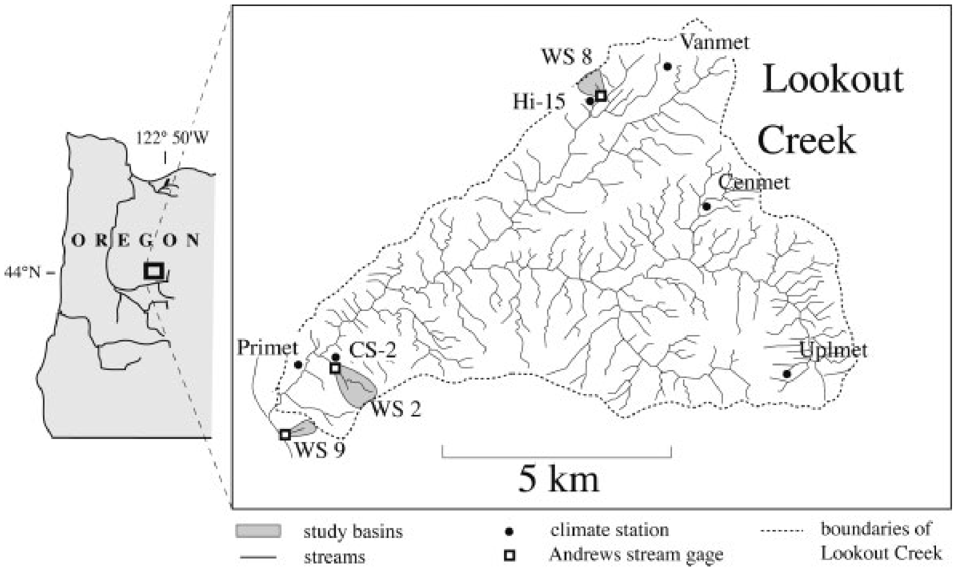

Figure 1. Study site location in the Lookout Creek basin, H.J. Andrews Experimental Forest, Oregon. The three study basins (Watershed 9, Watershed 2, and Watershed 8) their gauges, and nearby climate stations (CS-2, Hi-15, and VanMet) are shown

Data collection

Data from six climate stations (Primet, CS-2, Cenmet, Vanmet, Hi-15, upper Lookout) and three stream gauging stations (WS 2, 8, and 9) were used in this study (Figure 1). Because of data limitations, study periods ranged from November 1963 to December 1992 for analyses involving daily data at WS 2 and 8, from November 1985 to December 1992 for analyses involving continuous data at WS 8, and from November 1968 to December 1992 for analyses at WS 9. Streamflow has been measured continuously since 1952 at WS 2, since 1963 at WS 8, and since 1968 at WS 9. Today, discharge is measured in trapezoidal flumes at all basins, but H-flumes were used from 1968 to 1973 at WS 9, and from 1963 to 1996 at WS 8. At all basins stage height was recorded on Stevens A-35 strip charts (Stevens, Inc., Portland OR; details on instrumentation related to this study are available at http://www.fsl.orst.edu/lter/data/abstractdetail.cfm?dbcode=HF004&topnav=135), which were updated to digital pressure transducers in the 1990s. Small basin rating curves were obtained for the original flume installations, adjusted to account for the flume change at WS 9.

Precipitation and temperature have been measured at the CS2 climate station continuously since 1958, at Hi-15 daily since 1963 and continuously since 1984, at Cenmet daily since 1995, at Primet continuously since 1979, at Uplmet continuously since 1994, and at Vanmet continuously since 1990. Solar radiation has been measured at Primet since May 1972 with a thermopile sensor mounted on a platform 1·1 m above the ground (Bierlmaier and McKee, 1989). Snow water equivalent has been measured at Vanmet since 1987 in a 1 × 1 m Park Mechanical pressure pillow (details on instrumentation related to this study are available at http://www.fsl.orst.edu/lter/data/abstractdetail.cfm?dbcode=MS001&topnav=135) instrumented with a pressure transducer. Snowmelt has been measured since 1992 in a snowmelt lysimeter at the Hi-15 climate station. A tipping bucket rain gauge measures daily output from the lysimeter, which consists of a 2 × 2 m wood-frame collection box, 0·25 m deep, lined with black plastic and painted in colours of a forest floor.

Retrospective estimation of soil moisture and snow water equivalent

Neither soil moisture nor snow water equivalent measurements were available in study basins for model validation or calibration during the period of the study (1968 to 1992). Point measurements were collected starting in 1987 (snow) and 1998 (soil moisture) at climate stations (http//:www.fsl.orst.edu/lter), but these measurements are in canopy gaps, not under forests, of WS 2, 8, and 9. Therefore soil moisture and snow water equivalent were estimated retrospectively over the period from 1964 to 1992 for WS 2 and WS 8, and over the period 1968 to 1992 for WS 9. Estimates were made using the Modular Modeling System (MMS) (CADWES, 1994), a distributed-parameter hydrologic model in which a basin is divided into hydrologic response units (HRUs), areas of presumed homogenous hydrologic behaviour, called. HRUs were defined according to mapped soil series and vegetation (Rothacher *et al*., 1967; Dyrness and Hawk, 1972; Hawk and Dyrness, 1975). Parameters relating to soil moisture and snow water equivalent were obtained from A. Sikka (1994, unpublished data).

Daily precipitation, minimum and maximum air temperature, and solar radiation data were used as model input to MMS for each basin. Daily maximum and minimum temperature data were calculated as basin-wide, area-weighted averages of values in HRUs, whose temperatures were estimated using lapse rates and adjustment factors to account for temperature effects of hillslope orientation. Precipitation and temperature parameters were

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obtained from the CS2 climate station for WS 2 and 9, and the Hi-15 station for WS 8 (Figure 1). Solar radiation parameters were taken from Primet (Figure 1) starting May 1972. Daily solar radiation was estimated for the period from 1964 to May 1972 as potential solar radiation less daily sky cover following Duan (1996). Potential solar radiation was computed as a function of Julian day, latitude, slope, and aspect of the site following Frank and Lee (1966) and Swift (1976). Daily sky cover was calculated using an observed monthly relationship of sky cover to daily temperature range and daily maximum and minimum air temperatures (Duan, 1996).

The objective of soil moisture modelling was to determine whether basin-wide soil moisture was near-saturated or unsaturated on the day of a peak discharge, using area-weighted averages of soil moisture values from HRUs in each basin, modelled using MMS. Soil moisture stored in each HRU was estimated in MMS as the difference between water inputs (precipitation and modelled snowmelt) and outputs within the average rooting depth of the predominant vegetation (CADWES, 1994; Leavesley *et al*., 1983). Soil moisture storage capacities were set equal to the difference between measured field capacity and wilting point of each mapped soil series (Dyrness and Hawk, 1972; Hawk and Dyrness, 1975; Rothacher *et al*., 1967).

The goal of snow simulation was to determine whether or not a minimum snowpack was present in each study basin on the day of a peak discharge, using an area-weighted average of snow water equivalent from HRUs in the basin, modelled using MMS. Snow water equivalent for each HRU was estimated in MMS using an energy balance computed every 12 h (CADWES, 1994; Leavesley *et al*., 1983). Snowpack energy dynamics, and accumulation and depletion rates, were calculated following Obled and Rosse (1977).

Measure of hydrograph shape

Peak discharges were selected from the continuous discharge records at WS 2, WS 8 and WS 9 using a linear baseflow-quickflow separation technique incorporating a user-defined baseflow slope (GetPQ version 1·5; Dripchak and Hawkins, 1992) based on the work of Hewlett and Hibbert (1967). Applied to continuous streamflow records for the periods 1963 to 1992 for WS 2 and 8, and 1970 to 1992 for WS 9, this procedure resulted in initial files with varying numbers of peak discharge events ($n = 395$ for WS 9, $n = 284$ for WS 2, and $n = 239$ for WS 8). For each event, ten variables describing hydrograph shape were calculated from Get-PQ (Figure 2). Hyetograph shape was characterized by total event precipitation, precipitation volume before hydrograph rise, and maximum 15-min precipitation intensity. Hydrograph shape was characterized by timing (start lag, time to peak, duration, centroid lag) and magnitude (baseflow at the time of hydrograph rise, peak discharge, and total flow). The initiation time of each runoff event was determined when streamflow exceeded user-defined minima (0 mm precipitation, 2·03 mm streamflow depth, 0·23 mm h^{-1} discharge). GetPQ searched the continuous precipitation data backwards through time starting at the runoff event initiation

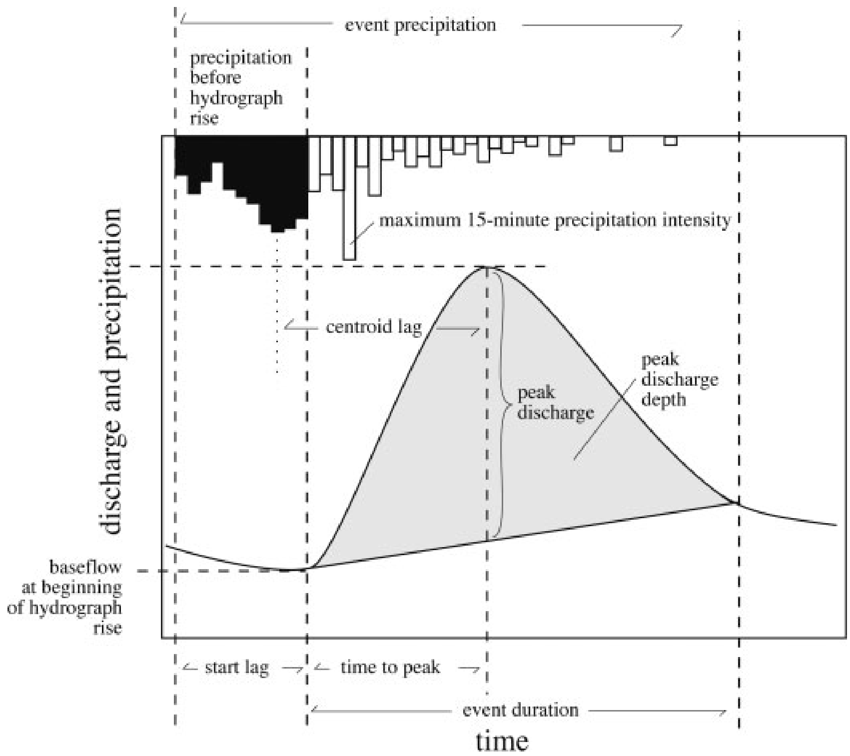

Figure 2. Variables determined from continuous hydrograph and hyetograph using get-PQ and used in analysis

Table I. Classification scheme for the most common peak discharge event types. Frequencies of these events are shown in Table II

	Precipitation[a]	Soil moisture[b]	Snow water[b]	Minimum temperature[c]
Rain on unsaturated soil	>0	≤90%	<0·25 cm	>0 °C
Rain on near-saturated soil	>0	>90%	<0·25 cm	>0 °C
Rain-on-snow on near-saturated soil	>0	>90%	>0·25 cm	>0 °C

[a] Observed daily data from nearest climate station.
[b] Estimated from MMS runs.
[c] Estimated for basin using lapse rates and aspect corrections applied to data from nearest climate station.

time until it encountered a user-defined duration of zero rainfall (set to 3 h in this study), which defined the beginning of the precipitation event (Figure 2). The complete set of GetPQ parameters was obtained for a reduced number of events ($n = 372$ at WS 9, $n = 212$ at WS 2, and $n = 44$ events at WS 8; this reduced sample is due to the short record of continuous precipitation at the Hi-15 climate station, starting in 1984).

Event classification

Peak discharge events from WS 2, 8 and 9 were classified to permit analysis of the effects of soil moisture status and snowpack presence upon properties of the peak discharge hydrograph (Table I). Over 80% of the sample of events in WS 2, 8 and 9 consisted of three event type classes (rain on unsaturated soil, rain on near-saturated soil, and rain-on-snow on near-saturated soil). 'Unsaturated rain' events had minimum daily temperatures >0 °C and estimated basin-wide, area-weighted average soil moisture was <90% of the estimated moisture storage capacity. Near-saturated rain' events had similar temperature to unsaturated rain events but estimated basin-wide soil moisture was >90% of capacity. 'Near-saturated rain-on-snow' events had similar soil moisture as near-saturated rain events, but were preceded by a period of snow (precipitation falling when the maximum daily air temperature was less than 0 °C) and the estimated basin-wide, area-weighted average value of snow water equivalent exceeded 2·5 mm (Table I).

Statistical analysis

Relationships between antecedent conditions, hyetograph, and hydrograph characteristics were tested using linear regression, principal components analysis, and multivariate analysis of variance, following methods in Johnson and Wichern (2001). All variables were tested for normality and equality of variance and were log-normally transformed before analysis. Analyses were conducted for the whole study period (1968–1992) and for the subset of the study period when data were available from all three basins (1985–1992).

RESULTS

Example storm hydrograph shapes

Storm hydrographs have small, variable peaks during unsaturated rain events, but large, similar peaks during extreme flood saturated-soil rain-on-snow events, which produce the highest discharge at the larger basin, Lookout Creek (Figure 3). Unsaturated rain events often occur early in the water year (October, November), such as the event of 3–6 November 2006 (Figure 3a). Despite 150 mm of precipitation from 1–3 November, which produced no peak, and 125 mm of precipitation from 3–6 November at maximum intensities exceeding 8 mm h^{-1}, unit area discharge remained low. The hydrograph at the low elevation basin with short, steep slopes (WS 9) had rapid, steep response, the highest unit area peak, and the shortest duration; the hydrograph at the intermediate elevation basin with long, steep slopes (WS 2) had a steep response, a peak less than half of that of WS 9, and a longer duration; and the hydrograph at the high-elevation basin with short, gentle slopes (WS 8) had a rapid but less steep response, a unit area peak only one-third of that at WS9, and the longest duration. All three hydrographs had similar, rapid times to peak. Runoff : rainfall ratios were very low (0·18, 0·10, 0·03, and 0·13 at WS9, WS2, WS8, and Lookout Creek) for this unsaturated-soil rain event.

Near-saturated rain events typically occur in midwinter (December to February), such as the event of 19–21 December 1974 (Figure 3b). After 200 mm of precipitation over the preceding 10 days, the storm delivered 160 mm of precipitation at maximum intensities exceeding 8 mm h^{-1}, and unit area discharges were higher than for the rain on unsaturated soil event (Figure 3a and 3b). The hydrograph at the low elevation basin with short, steep slopes (WS 9) had the earliest, steepest response, reached the highest peak, and had the shortest duration; the hydrograph at the intermediate elevation basin with long, steep slopes (WS 2) had a less rapid, less steep response, a peak 80% as high as WS 9, and a longer duration; and the hydrograph at the high-elevation basin with short, gentle slopes (WS8) also had a less rapid, less steep response than WS 9, a peak two-thirds of that at WS 9, and the longest duration. The centroid lag was 2 h at WS9 but 12 h at WS 2 and WS 8. Runoff : rainfall ratios were intermediate (0·54, 0·61, and 0·53 at WS9, WS2, WS8) for this near-saturated-soil rain event.

Near-saturated rain-on-snow events typically occur in winter and spring (December to April), such as the extreme flood event of 5–8 February 1996 (Figure 3c). Only 20 mm of precipitation fell over the preceding 5 days, but measured snowpacks contained 400 to 600 mm

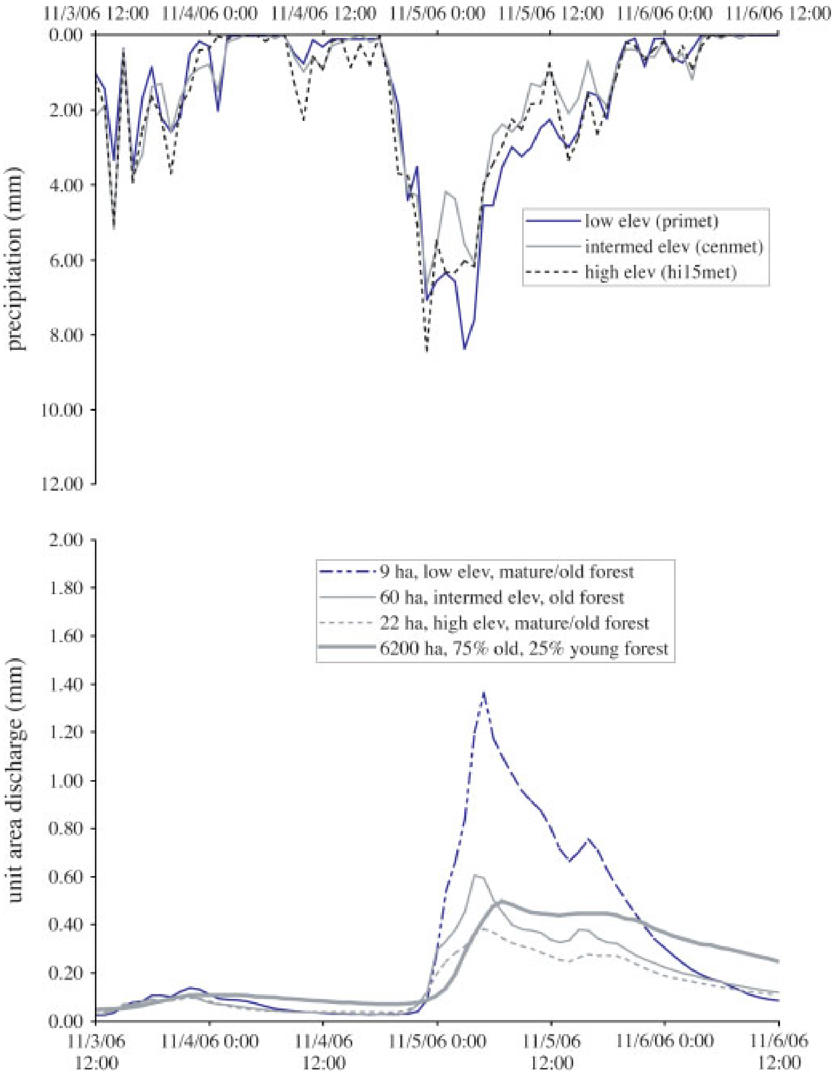

Figure 3. Hydrograph shapes at three forested watersheds during three storm event types. Note differing scales for unit area discharge in (a), (b), and (c). (a) rain event on unsaturated soils, 3–6 Nov. 2006; (b) rain event on near-saturated soils, 19–21 Dec. 1974; (c) extreme rain-on-snow event on near-saturated soils, 5–8 February 1996. Precipitation data are from low-elevation (dark solid line, CS-2 or Primet); mid-elevation (solid grey line, Cenmet or Hi-15; and high-elevation stations (dashed grey line, Vanmet or Uplmet)

of water equivalent at upper elevations. The storm delivered 250 mm of precipitation at maximum intensities exceeding 10 mm h^{-1}, producing the highest peak in the 50-year record. All three basins responded rapidly to precipitation. The low- and intermediate-elevation basins (WS 9 and WS 2) reached initial peaks within 2 h of bursts of precipitation on the evening of 6 February. Following the second burst of precipitation at 1–2 am on 7 February, the low- and intermediate-elevation basins (WS 9 and WS 2) sustained peak discharges from 2 to 11 am, and the high-elevation basin reached a peak at noon. Runoff : rainfall ratios were very high (0·79, 1·14, 0·84, and 1·20 at WS 9, WS 2, WS 8, and Lookout Creek for this near-saturated-soil rain-on-snow event.

Seasonal and interannual patterns of soil moisture and snow water

Snow water equivalents increase from October through February and then decline each year, but there is considerable within-season and interannual variability in snowpack water equivalent, and snow accumulation under forests (estimated using the model) is lower than in canopy gaps (where snow is measured) (Figure 4). Snow water equivalents estimated using MMS for the HRU containing Vanmet, assuming forest cover, were about eight-fold lower than values of snow water equivalent observed at Vanmet, in a canopy gap, for the period October 1987 to September 1992. These differences are similar to differences in snowpacks under forest cover versus canopy openings reported by Berris and Harr (1987), Marks *et al*. (1998), and Storck *et al*. (2002). Modelled and measured values reflect the ephemeral nature of snowpacks within a given season, even at high elevation in some years, especially 1991–1992 and 1993–1994, which were among the warmest and driest years in the 50-year climate record.

Retrospective estimates of soil moisture and snowpack water equivalent for the 1964–1965 water year at WS 2 correctly identified the extreme rain-on-snow floods of 22 December 1964 and 28–30 January1965 as near-saturated rain-on-snow events (shown by peaks

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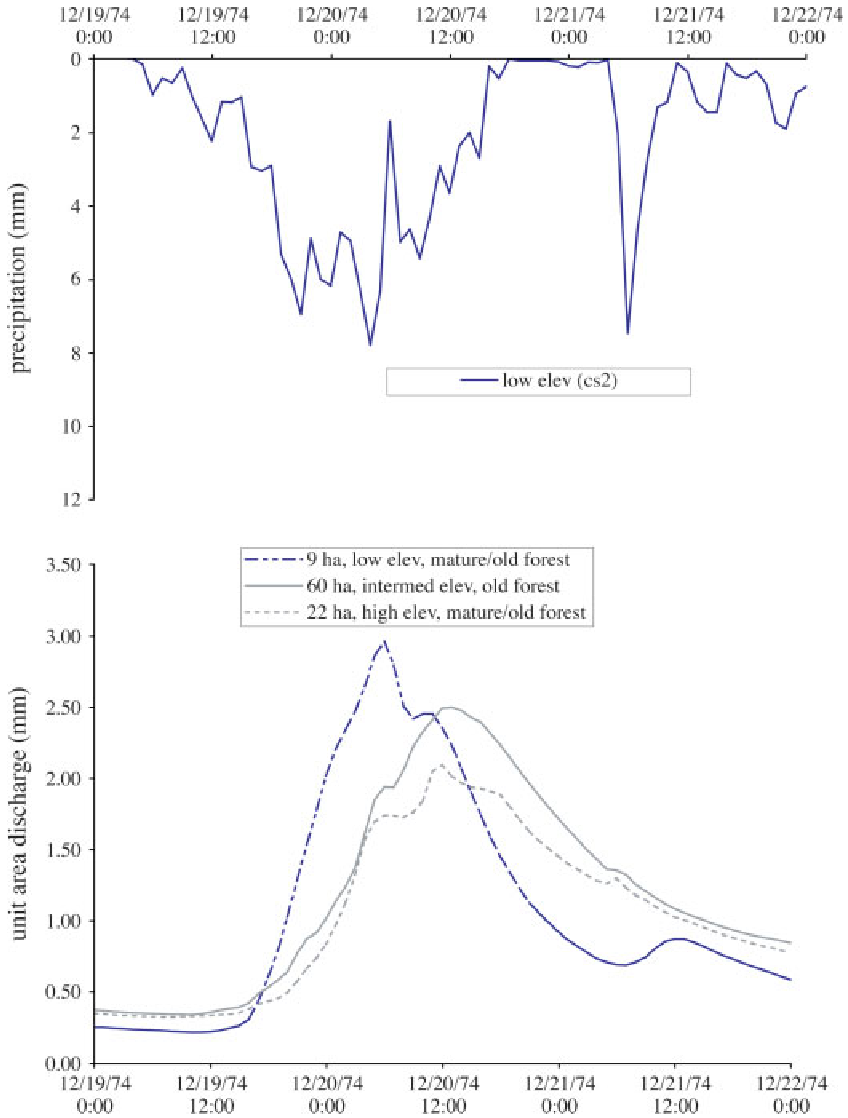

Figure 3. (*Continued*)

in precipitation, Figure 5a). Measured soil moisture and snowpack data can be used in an analogous fashion to retrospective modeling to classify days with near-saturated soils and rain on snow, as shown for the 1998–1999 water year (Figure 5b). Simulated soil moisture drained less rapidly than measured soil moisture (compare Figure 5a and b), but soil moisture at Primet (the only soil moisture records for 1998–1999) is measured in extremely well-drained soils developed on coarse cobbly river terrace deposits; subsequent soil moisture measurements at other sites display patterns more similar to simulated soil moisture. Both the simulated and observed snow water equivalents decrease during periods of precipitation on snow, indicating snowmelt during these rain-on-snow events, and the snowpack at 1200 m elevation is less responsive to rain than the simulated snowpack at 500–1000 m in WS 2 (Figure 5). Therefore, retrospectively estimated soil moisture and snowpack data were used to classify days during the study period (1968–1992), when no soil moisture or snowpack data were measured.

Some uncertainty exists in classifying snow conditions, as indicated by discrepancies between retrospectively modeled versus observed snow water equivalent (Figure 4 and Figure 5). Early in the water year (October and November) snow accumulations are more likely in canopy gaps than under forest, whereas near the end of the snowmelt period snow persists longer under forest than in the canopy gap (Figure 4). However, event classification was not very sensitive to small changes in the thresholds for rain, mixed, and snow conditions. Changing the temperature threshold for snow by a few degrees affected classifications only of event types excluded from this analysis (mixed rain and snow). Raising the threshold for an 'on-snow' event from 2·5 to 5 mm of snowpack decreased the number of 'on-snow days by about 3%, while lowering it to 0·1 mm increased the number of 'on-snow' days by about 2% (Figure 4). Despite these uncertainties, the event type classification appeared to capture the hydrologic conditions of snowmelt measured in the snow lysimeter at the Hi-15 climate station. Over the period 1 October 1992 to 30 September 1994, on

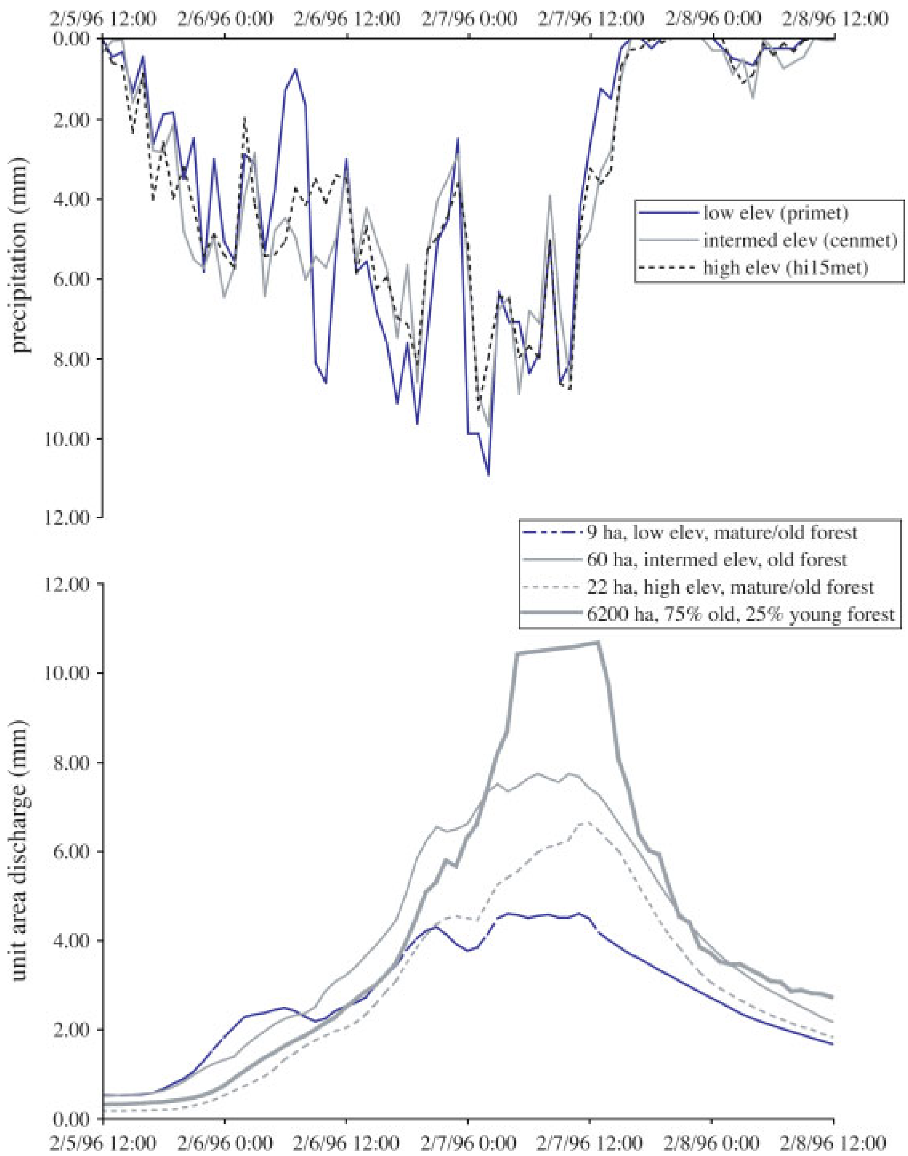

Figure 3. (*Continued*)

average, the lysimeter at the Hi-15 climate station yielded 130, 80, and 40% of precipitation on days retrospectively classified (using the same methods as above) as rain-on-snow, rain, and snow. Thus, the event-type classification appeared to capture the occurrence of snowmelt during rain-on-snow events, and the lack of snowmelt during rain and snow events.

Statistical properties of hydrograph shapes

Storm event types and peak discharge magnitudes were seasonally distributed, following patterns of precipitation, soil moisture, and temperature. From early March to mid-November, peak discharges were small ($<1{\cdot}5$ mm h^{-1}) and runoff ratios ratios (discharge depth/precipitation depth) were low ($<10\%$) (Figure 6). From mid-November through February, peak discharges exceeded 2 mm h^{-1} and runoff ratios exceeded 0·5. Storm event types also varied according to the elevation of the basin (Figure 6, Table II). Only 10% of all storm hydrographs were rain-on-snow events at the low-elevation

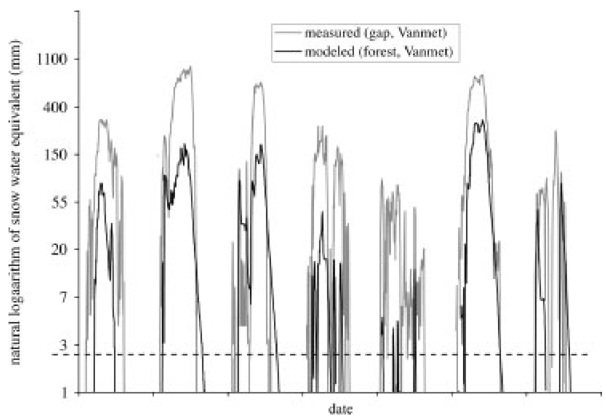

Figure 4. Measured and modelled snow water equivalents. Snow water equivalent was measured at a canopy gap containing the VanMet station (1100 m elevation; see Figure 1) and modelled for the HRU containing VanMet, but assuming that the area was forested. Patterns of measured and modelled snow demonstrate the order-of-magnitude differences in snow between canopy gaps (grey line, measured) and under forest (black line, modelled) and the ephemeral nature of snowpacks even at high elevation in some years. A snow water equivalent of 2·5 mm was used to define 'on snow' conditions (horizontal dashed line)

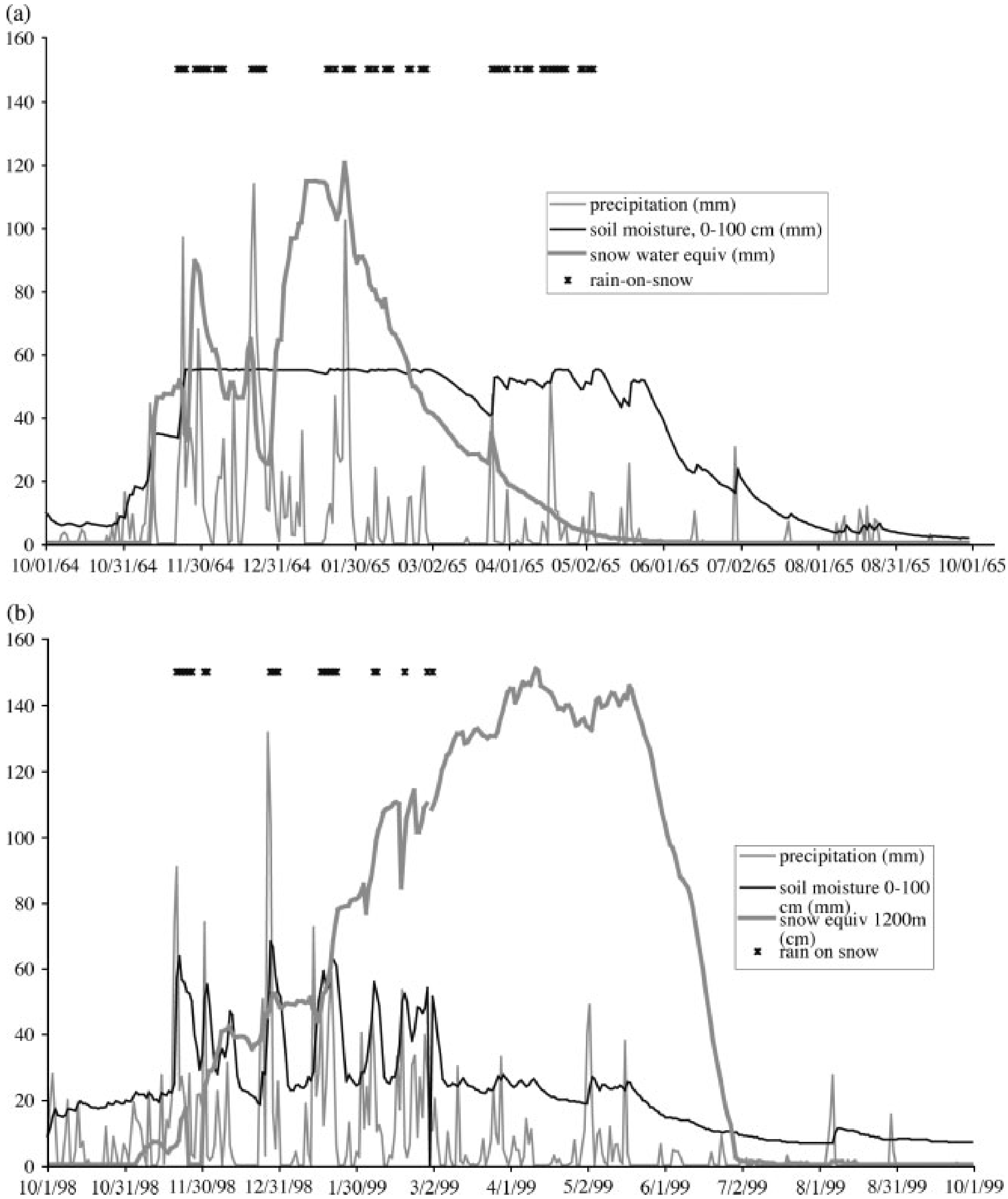

Figure 5. Comparison of storm event type classification using (a) retrospective modelling of soil moisture (0–100 cm) and snow water equivalent using MMS for 1964–1965 water year, and (b) measured soil moisture (0–100 cm, at Primet) and snow water equivalent (at upper Lookout) for the 1997–1998 water year. Asterisks indicate days with rain-on-snow (snow water equivalent >2·5 mm) on near-saturated soil (soil moisture >50 mm, or >90% of the soil moisture storage capacity)

basin (WS 9) where snowpacks are transient, but 40% of hydrographs were rain-on-snow events at the basins draining elevations (above 800 m) where seasonal snowpacks form (WS 2 and WS 8).

Three principal components, corresponding to total precipitation volume, hydrograph timing relative to precipitation, and initial baseflow, explain 75, 73, and 81% of variation in the ten hydrograph shape variables in WS 9, WS 2, and WS 8 (Table III). Five correlated variables expressing aspects of storm size explain 41 to 43% of variation in hydrograph shape variables at WS 9, 2, and 8 (Table III and IV). Three correlated variables express hydrograph timing relative to precipitation: the precipitation volume before the hydrograph begins to rise (pre-runoff precipitation volume); the lag between the initiation of precipitation and the time when the hydrograph begins to rise (start lag); and the lag between the median precipitation and the median storm discharge (centroid lag). These timing variables explain 22, 19, and 24% of variation in hydrograph shape variables at WS 9, 2, and 8 (Tables III and IV). The third principal component is primarily expressed by antecedent wetness (baseflow at the beginning of the storm) and explains an additional 11 to 15% of variation in the data.

Hydrograph shape differences between basins

Hydrograph size and timing vary among the basins. The basin with short steep slopes (WS 9) required less

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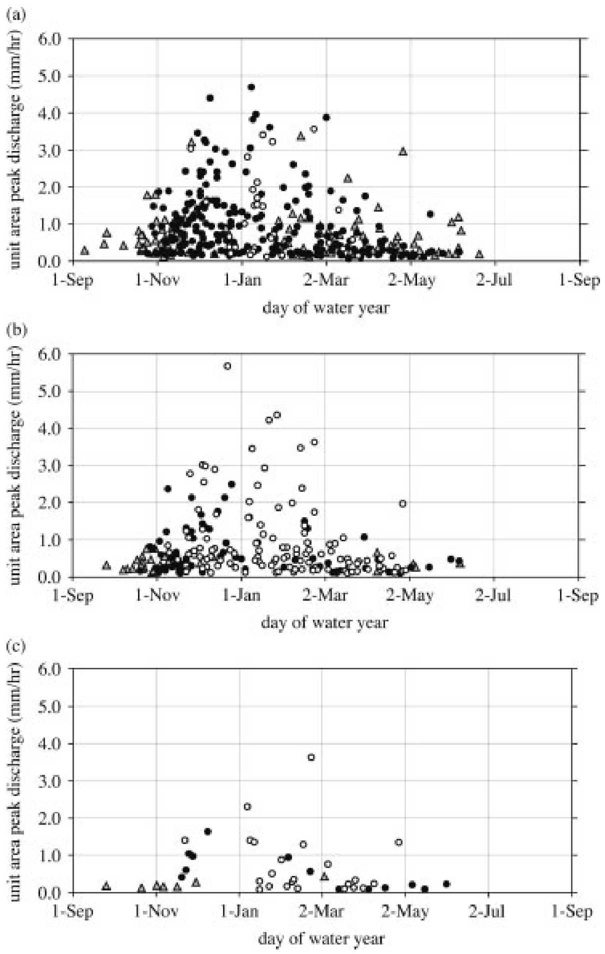

Figure 6. Unit area peak discharges (mm h^{-1}) over the water year (1 October to 30 September) for three event types: rain events on unsaturated soil (gray triangles), rain events on near-saturated soils (solid circles), and rain-on-snow events on near-saturated soil (open circles), from (a) Watershed 9, 1969–1992; (b) Watershed 2, 1964–1992; and (c) Watershed 8, 1985–1992

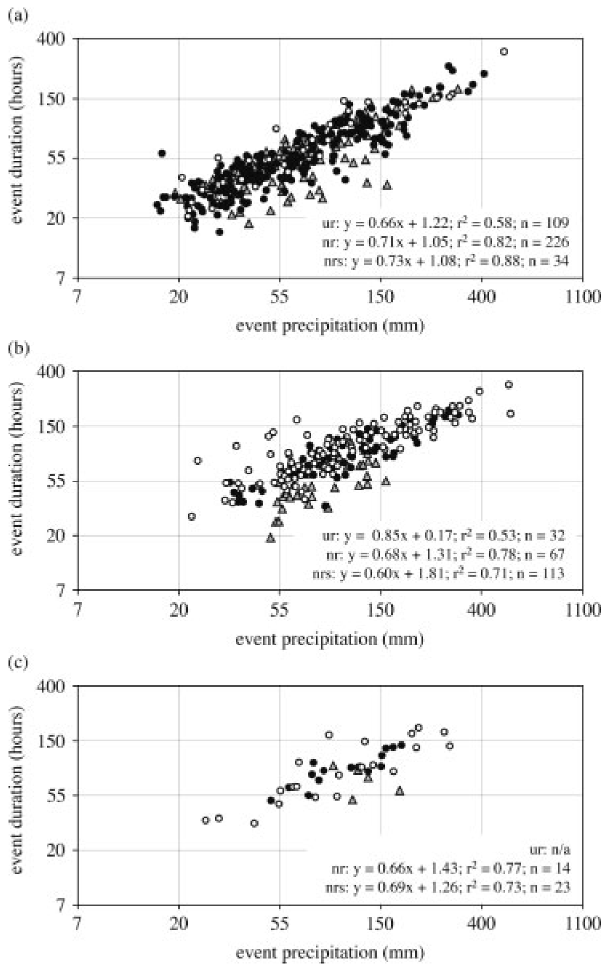

Figure 7. Relationships between event duration (h) and event precipitation (mm) for three event types in three watersheds: rain events on unsaturated soil ('ur', gray triangles), rain events on near-saturated soils ('*nr*', solid circles), and rain-on-snow events on near-saturated soil ('*nrs*', open circles), from (a) Watershed 9, 1969–1992; (b) Watershed 2, 1964–1992; and (c) Watershed 8, 1985–1992. Data were log-transformed for analysis and plotting

event precipitation, less pre-runoff precipitation, lower initial baseflow, and lower precipitation intensity to produce a storm hydrograph than the basin with long, steep slopes (WS 2), which in turn required lower values of these properties than the basin with short, gentle slopes (WS 8) (Figures 7, 8, 9 and 10). WS 9 produced many more short-duration storm hydrographs in response to small amounts of precipitation than WS 2 or WS 8 (Figure 7). The relationship between pre-runoff precipitation and start lag did not differ among basins (Figure 8). However, at WS 8, hydrologic response to pre-runoff precipitation changed markedly with event type: when pre-runoff precipitation was less than 20 mm, near-saturated rain-on-snow events started more slowly than near-saturated rain events, but they started more rapidly when pre-runoff precipitation exceeded 20 mm (Figure 8).

The three basins differed in their hydrograph response to initial baseflow and precipitation intensity. Only at WS 8 did hydrograph response time decrease with initial baseflow (Figure 9, Table IV). Storm precipitation intensity was positively related to storm size (peak discharge) at WS 9 and WS 8, but not at WS 2 (Table IV, Figure 10). At the basin with short steep slopes (WS 9) peak size increased faster with increasing precipitation intensity than at basins with longer (WS 2) or less steep (WS 8) slopes (Figure 10).

Hydrograph shape differences between event types

Hydrograph size and timing differed among the three event types (Table II). Hydrographs of unsaturated rain events were shorter and smaller than those of near-saturated rain events, which in turn were generally smaller and responded more rapidly than those of rain-on-snow events (Table II). For the period of overlapping records from all three basins (1985 to 1992), rain-on-snow events had longer durations and centroid lags than rain events, and they had higher runoff ratios than unsaturated rain events (Table V). Controlling for event precipitation, the shortest hydrographs were produced during rain on unsaturated soil events, and the longest during rain-on-snow on near-saturated soil events (Figure 7). Rain-on-snow events responded slightly faster than did

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Table II. Average hydrograph shape variables from unsaturated rain, near-saturated rain, and near-saturated rain-on-snow peak discharge events in Watersheds 9, 2, and 8 at the H.J. Andrews Experimental Forest over the period 1968–1992 (WS9 and WS2) and 1985–1992 (WS8)

	Watershed 9			Watershed 2			Watershed 8		
	ur[a]	*nr*[a]	*nrs*[a]	*ur*	*nr*	*nrs*	*ur*	*nr*	*nrs*
n[b]	109	226	34	32	67	113	30[b]	60[b]	94[b]
Precipitation (mm)	78	85	107	92	117	135	115	111	120
Maximum 15-min precipitation intensity (mm hr^{-1})	7	7	8	20	17	13	41	35	27
pre-runoff precipitation (mm)	11	8	9	10	6	12	17	11	26
Start lag (h)	14	10	12	13	10	15	21	13	23
Centroid lag (h)	13	11	12	12	16	22	12	20	26
Time to peak (h)	23	25	36	20	35	46	21	40	40
Duration (h)	59	66	85	56	93	113	67	105	106
Initial baseflow (mm hr^{-1})	0·07	0·18	0·19	0·03	0·14	0·20	0·05	0·17	0·20
Peak discharge (mm hr^{-1})	0·63	0·96	1·13	0·34	0·67	0·96	0·40	0·76	0·95
Storm event runoff ratio (mm mm)[c]	0·16	0·24	0·29	0·08	0·20	0·30	0·05	0·18	0·22

[a] *ur* = rain on unsaturated soil; *nr* = rain on near-saturated soil; *nrs* = rain-on-snow on near-saturated soil.
[b] Sample sizes from Watershed 8 are restricted for properties requiring continuous precipitation data, which was only available for the period 1985–1992. For these properties (precipitation, maximum 15-m precipitation intensity, pre-runoff precipitation, start lag, centroid lag, ratio of quickflow depth to precipitation), N-values for Watershed 8 were 7 for dry-soil rain events, 14 for for wet-soil rain events, and 23 for wet-soil rain-on-snow events.
[c] Storm discharge depth/event precipitation.

Table III. Coefficients of first and second principle components from multivariate analysis of ten hydrograph shape variables at Watersheds 9, 2 and 8. Coefficients above 0·30 and below −0·30 are underlined

Hydrograph shape variable	Watershed 9: Short steep slopes, low elevation			Watershed 2: Long steep slopes, intermediate elevation			Watershed 8: Short gentle slopes, high elevation		
	$\hat{e}_1$	$\hat{e}_2$	$\hat{e}_3$	$\hat{e}_1$	$\hat{e}_2$	$\hat{e}_3$	$\hat{e}_1$	$\hat{e}_2$	$\hat{e}_3$
Storm size									
Event precipitation (mm)	0·453	−0·043	−0·167	0·400	0·028	−0·436	0·422	−0·139	−0·229
Peak discharge (mm)	0·440	0·059	0·054	0·400	0·173	−0·244	0·439	0·168	−0·084
Time to peak (hrs)	0·353	−0·069	−0·078	0·372	0·054	0·015	0·373	0·006	0·248
Event duration (hrs)	0·437	−0·016	−0·079	0·445	0·124	−0·055	0·442	0·069	−0·053
Runoff : precipitation	0·391	0·094	0·316	0·406	0·243	0·183	0·373	0·358	0·129
Hydrograph timing relative to precipitation									
lprerunoff	0·051	−0·590	0·001	0·172	−0·623	−0·166	0·131	−0·568	0·136
lstartlag	0·052	−0·601	0·033	0·135	−0·652	−0·037	0·041	−0·537	0·137
lcentroidlag	0·088	−0·476	0·165	0·287	−0·221	0·330	0·283	−0·210	0·527
Antecedent wetness									
lqbase	0·045	−0·016	0·890	0·215	0·062	0·599	−0·046	0·405	0·451
Precipitation intensity									
lmaxprecip	0·344	0·130	−0·192	−0·058	0·162	−0·465	0·232	−0·029	−0·584
variance($\hat{\lambda}_i$)	4·252	2·127	1·138	4·198	1·868	1·217	4·132	2·376	1·556
Cumulative percentage of total variation	42·5	63·8	75·2	42·0	61·2	73·4	41·3	65·1	80·6

rain events when 20 mm or more precipitation fell before hydrograph rise (Figure 8). Initial baseflow was positively related to soil moisture (near-saturated compared with unsaturated events, Table V), but hydrograph timing was speeded by higher initial baseflow only at WS 8 (short gentle slopes), and not during rain-on-snow events (Figure 9).

Average unit area peak discharge did not differ significantly by event type, but the largest events at WS 2 and 8 were near-saturated rain-on-snow events (Figure 6). These hydrographs had large storm size and fast response (high scores in the first and low scores in the second principal components, Table III). The highest peak discharges for a given precipitation intensity occurred during rain-on-snow events at all three basins (Figure 10).

DISCUSSION

Analysis of storm hydrographs reveals how slope length and gradient interact with water flowpaths and a changing

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Table IV. Correlations between four major components of variation in properties of storm hydrographs in peak discharge events at three small forested watersheds in the H.J. Andrews Forest. The major components of variation are: storm size, hydrograph timing relative to precipitation, antecedent wetness, and precipitation intensity. Refer to Figure 2 for hydrograph properties

	WS9	WS2	WS8
Storm size			
peak discharge v. runoff ratio	0·84	0·73	0·80
storm precipitation v. event duration	0·91	0·84	0·75
peak discharge vs total precipitation	0·81	0·83	0·81
peak discharge vs time to peak	0·47	0·41	0·36
Hydrograph timing relative to precipitation			
start lag vs centroid lag	0·61	0·49	0·51
pre-runoff precipitation vs startlag	0·78	0·84	0·59
Antecedent wetness			
initial base flow vs centroid lag	−0·04	0·11	−0·17
Precipitation intensity			
maximum 15-min precipitation vs peak discharge	0·75	0·03	0·66

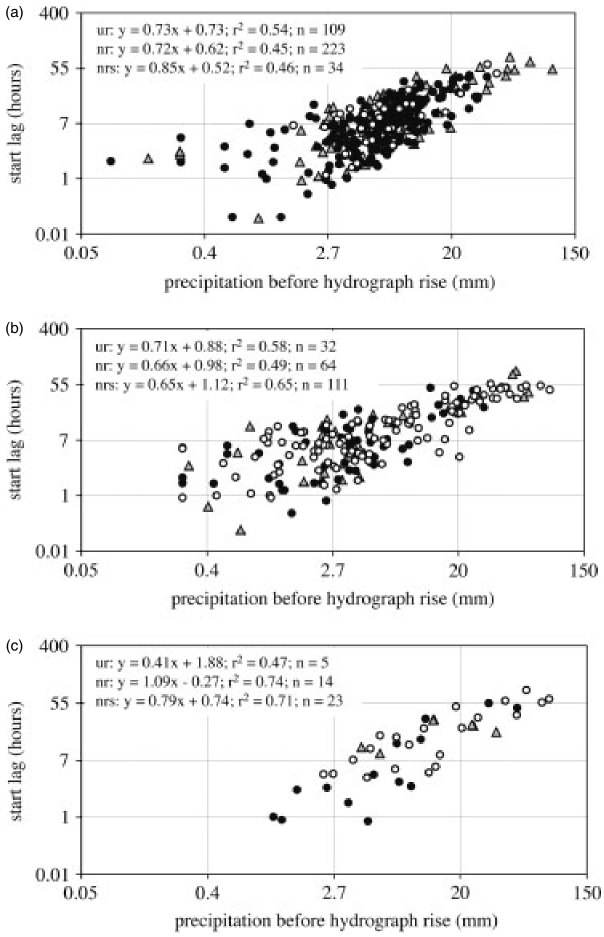

Figure 8. Relationships between precipitation before hydrograph rise (mm) and start lag (h) for three event types in three watersheds: rain events on unsaturated soil ('ur', gray triangles), rain events on near-saturated soils ('nr', solid circles), and rain-on-snow events on near-saturated soil ('nrs', open circles), from (a) Watershed 9, 1969–1992; (b) Watershed 2, 1964–1992; and (c) Watershed 8, 1985–1992. Data were log-transformed for analysis and plotting

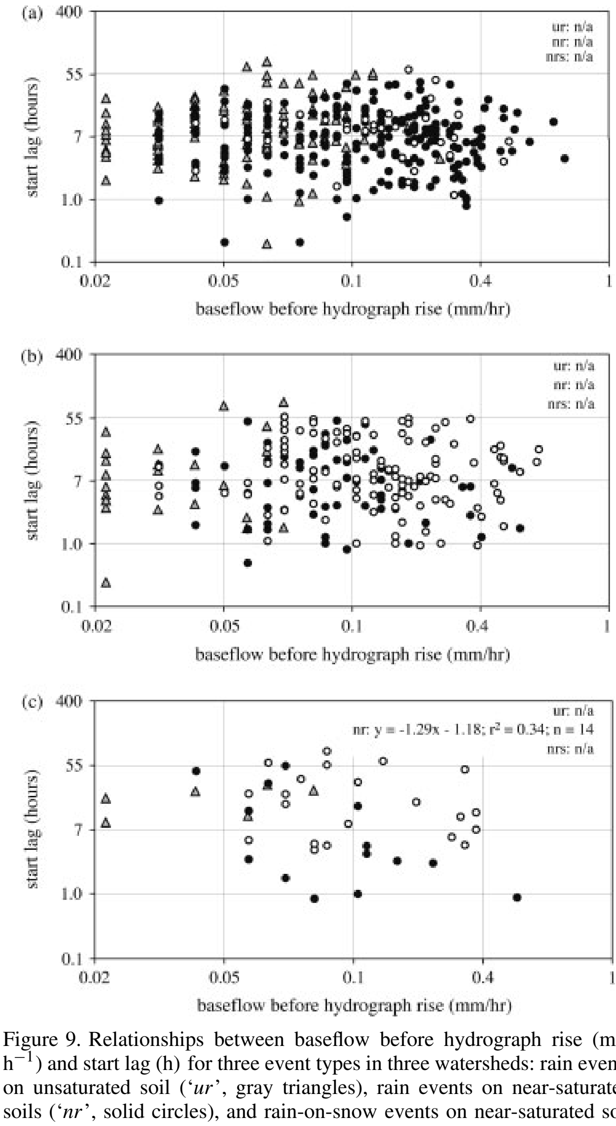

Figure 9. Relationships between baseflow before hydrograph rise (mm h^{-1}) and start lag (h) for three event types in three watersheds: rain events on unsaturated soil ('*ur*', gray triangles), rain events on near-saturated soils ('*nr*', solid circles), and rain-on-snow events on near-saturated soil ('*nrs*', open circles), from (a) Watershed 9, 1969–1992; (b) Watershed 2, 1964–1992; and (c) Watershed 8, 1985–1992. Data were log-transformed for analysis and plotting

contributing area during events with a range of antecedent conditions (Figure 11). Upper hillslopes may contribute water to the storm hydrograph if (1) local dispersion, e.g. along macropores, is high relative to slope length (Kirchner *et al.*, 2001); or (2) effective hydraulic conductivities are high, extending the effective hillslope contributing

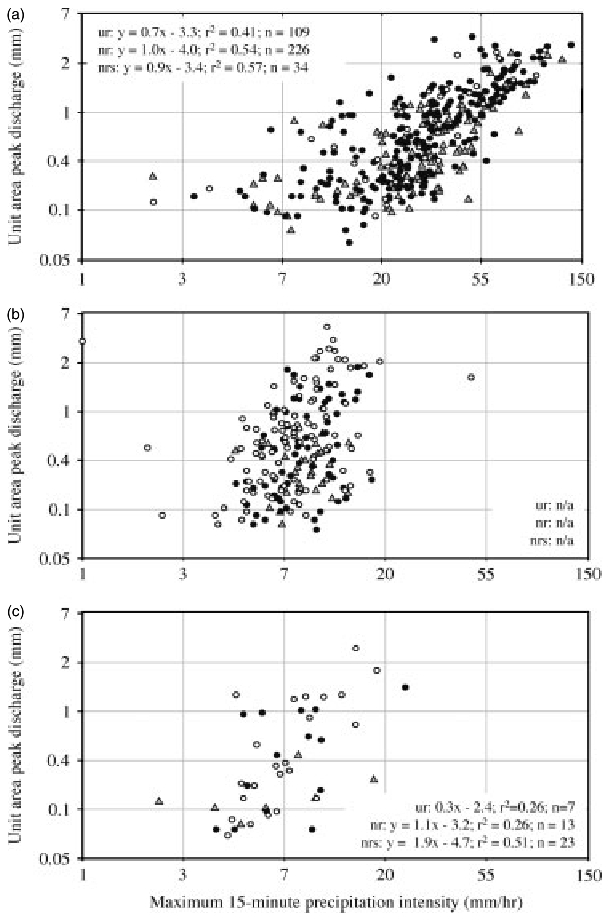

Figure 10. Relationships between maximum 15-min precipitation intensity (mm h^{-1}) and peak discharge (mm) for three event types in three watersheds: rain events on unsaturated soil ('*ur*', gray triangles), rain events on near-saturated soils ('*nr*', solid circles), and rain-on-snow events on near-saturated soil ('*nrs*', open circles), from (a) Watershed 9, 1969–1992; (b) Watershed 2, 1964–1992; and (c) Watershed 8, 1985–1992. Data were log-transformed for analysis and plotting

area upslope (Beven, 1982a, b). Shorter hillslope lengths and steeper gradients reduce the effective contributing area (Aryal *et al*., 2003). It is hypothesized that near-stream sources dominate during rain events on unsaturated soils; deep subsurface flowpaths become important during rain events on near-saturated soils; and both shallow and deep subsurface flowpaths are important during rain-on-snow events on near-saturated soils, when snowmelt water is moving into upper soil layers. Storm hydrograph peaks are sensitive to maximum precipitation intensities in basins with short slopes, in which pulses of precipitation are rapidly transmitted to channels, but not in basins with long slopes. At high elevations, where snowpacks are deep and slopes are short and gentle, storm hydrograph rise is delayed during near-saturated rain-on-snow events relative to rain events when initial baseflow—a measure of antecedent wetness—is high, because snowpacks absorb precipitation inputs, and may have slow percolation rates (Campbell *et al*., 2006). In contrast, initial baseflow is not related to hydrograph timing for any event type in basins with steep slopes, where hillslopes drain rapidly and the contributing area contracts between events. On short, gentle slopes, the presence of a snowpack delays hydrograph response when pre-runoff precipitation is low, but speeds hydrograph response when pre-runoff precipitation is high, because above some amount of pre-runoff precipitation, snowpacks transition from net sinks for precipitation inputs, to net sources of melt to underlying soils.

During a rain-on-snow event on steep slopes, the presence of a snowpack appears to result in melting and an expansion of the effective contributing area, delaying the storm hydrograph, but on short, gentle slopes, the presence of a snowpack speeds the start of the hydrograph if the snowpack has been 'primed' by receiving more than 20 mm of pre-runoff precipitation. Snowpack melting combined with infiltration of precipitation into the snowpack during the early part of a rain-on-snow event may saturate portions of the snowpack, reducing infiltration capacity. Short bursts of intense precipitation under these conditions could produce the higher peaks observed for rain-on-snow events. If snow meltwater requires hours to days to travel to the channel (Lundquist and Dettinger, 2005), snow meltwater may not reach the channel in time to contribute to the peak; instead snowpack melting may affect hydrograph shape by altering flowpaths and hydrograph timing.

Because during rain events the low-elevation, short-sloped, steep basins peak first, while the high-elevation, short and gently sloped basins peak last, a snowpack effect that delays hydrographs in steep, low-elevation basins, but speeds hydrographs in gently sloping, high elevation basins would have the effect of synchronizing the timing of the hydrographs from these three small basins. Synchronized peaks from all small basins are typical of extreme rain-on-snow floods such as those of December 1964 and February 1996 (Figure 3). By altering the relative timing of hydrographs, snowpack melting could produce extreme floods from precipitation events whose size is not extreme.

Creation of canopy gaps by clearcutting of national forests in the transient snow zone of Oregon (400 m to 1200 m) may have exacerbated the snowmelt effect on rain-on-snow floods since the 1950s (Coffin and Harr, 1992; Harr, 1986; Jones and Grant, 1996; Jones, 2000). Even extreme rain-on-snow floods in the Willamette Valley of Oregon (enumerated by Harr, 1981) date from periods in the mid- to late-1800s after large areas of forest in the transient snow zone had burned (Weisberg and Swanson, 2003). Since large areas of snowpack in the Pacific Northwest are 'at risk' from climate warming (Nolin and Daly, 2006), the magnitude of rain-on-snow floods may be sensitive to future snowpack effects.

CONCLUSIONS

In this study, long-term streamflow and climate records permitted the extraction of large sample sizes of events, allowing examination of hyetograph and hydrograph characteristics over a wide range of conditions. Retrospectively modelled snowpacks under forest cover were

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Table V. The 99% simultaneous confidence intervals for contrasting peak discharge event types from multivariate analysis of variance for the period 1985–1992 when data were available from all three watersheds. If the confidence interval includes zero, the two event types are not significantly different at 99%. A confidence interval with negative values means that the first event type in the comparison has significantly lower values of the hydrograph shape variable compared to the second event type. Significant differences at the 99% level are underlined. *N* values were 35 unsaturated rain events, 67 near-saturated rain events, and 16 near-saturated rain-on-snow events from WS 9; 8 unsaturated rain events, 19 near-saturated rain events, and 30 near-saturated rain-on-snow events from WS 2; and 5 unsaturated rain events, 14 near-saturated rain events, and 23 near-saturated rain-on-snow events from WS 8

Hydrograph shape variable	Unsaturated rain vs near-saturated rain events	Unsaturated rain vs near-saturated rain-on-snow events	Near-saturated rain vs near-saturated rain-on-snow events
Storm size			
Event precipitation (mm)	(−47·486, 38·126)	(−68·435, 23·363)	(−65·894, 11·462)
Peak discharge (mm)	(−0·293, 0·605)	(−0·717, 0·245)	(−0·486, 0·325)
Time to peak (h)	(−12·977, 23·821)	(−38·404, 1·053)	(−29·877, 3·372)
Event duration (h)	(−14·679, 38·458)	(−64·894,−7·917)	(−48·522,−0·509)
Runoff : precipitation	(−0·002, 0·145)	(−0·177,−0·020)	(−0·094, 0·039)
Hydrograph timing relative to precipitation			
Precipitation before hydrograph rise (mm)	(−18·642, 1·848)	(−9·600, 12·371)	(−16·269, 2·245)
Start lag (h)	(−16·666, 1·367)	(−8·066, 11·271)	(−14·195, 2·100)
Centroid lag (h)	(−6·568, 5·644)	(−14·974,−1·880)	(−14·406,−3·372)
Antecedent wetness			
Initial baseflow (mm)	(0·012, 0·131)	(−0·169,−0·041)	(−0·088, 0·020)
Precipitation intensity			
Maximum 15-min precipitation intensity	(−17·052, 12·131)	(−9·297, 21·995)	(−9·296, 17·072)

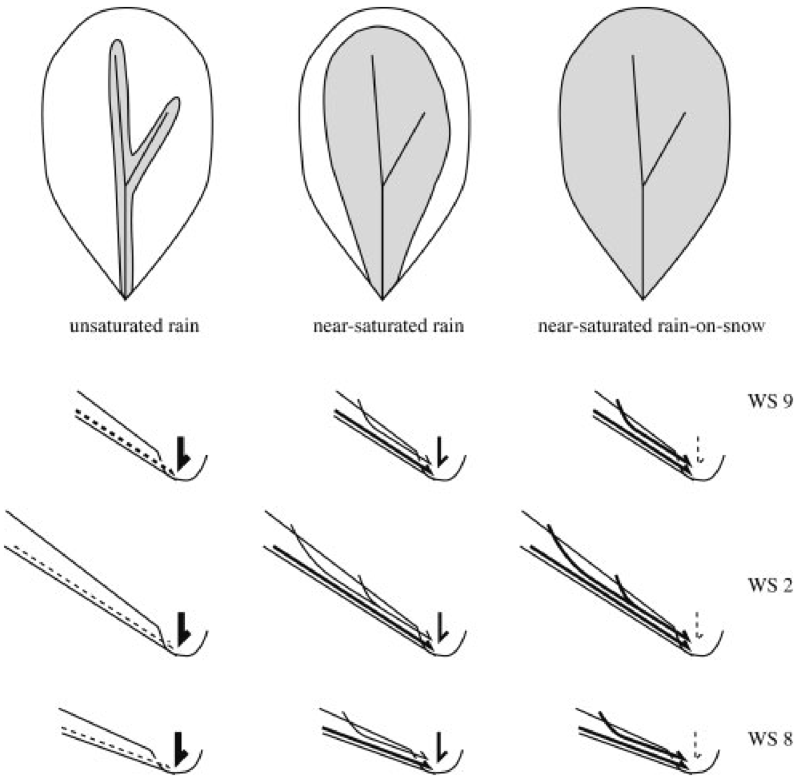

Figure 11. Planview diagrams depicting contributing area and cross-sectional diagrams depicting importance of various flowpaths for three event types in three watersheds: short, steep slopes (WS 9); long, steep slopes (WS 2); and short, gentle slopes (WS 8). Line thickness is proportional to the relative importance of deep and shallow hillslope flowpaths and channel interception during the event

smaller and developed more slowly, but could persist longer, than measured snowpacks in a canopy opening. Despite uncertainties in true snowpack and soil moisture, retrospective modelling of these properties greatly extends the periods over which relationships among climate factors and streamflow could be examined. Use of climate data and retrospective modelling to subdivide storm events into categories captured meaningful differences in soil moisture and snow water equivalent, and explained some of the variability in hydrograph size and timing, while revealing contrasting responses among basins based on hillslope gradient and length.

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In these small, forested basins, increasing soil moisture was associated with higher baseflow before hydrograph rise, higher storm runoff ratios, and longer lags between precipitation and peak runoff. The presence of a snowpack did not increase peak discharge for a given amount of precipitation, but it did increase the peak discharge for a given maximum precipitation intensity. The presence of a snowpack had contrasting effects on low-elevation, steep-sloped basins, where hydrographs were delayed, and high-elevation, gently sloped basins, where hydrographs were speeded, potentially synchronizing small basin contributions to the larger basin hydrograph during large rain-on-snow events. Further work is needed to understand how multiple small basins interact at the large basin scale during extreme floods, and how canopy openings, either natural meadows or openings created by clearcutting, may contribute to generating extreme rain-on-snow floods at the regional scale.

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