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FORWARD

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INTRODUCTION

Streamflow response to rainfall is controlled by the net water budget and the routing of rainfall to stream channels. In rain-dominated forested regions, including the Oregon Coast Range, water budget aspects of forest hydrology are better quantified than water routing. As a result, most of the commonly used forest hydrology models have been designed to assist in the analysis of seasonal or annual runoff volumes (Goldstein, et al. 1974; Troendle and Leaf, 1980; Silvey and Rosgen, 1981). The effects of timber harvesting on runoff volumes are reflected primarily through reductions in evapotranspirational demand and the corresponding increases in soil water content.

An important management concern is the effect of timber harvesting on the size of large peak flows and the duration of flows greater than the bankfull discharge. There is a growing body of literature which suggests that it is not simply runoff volume, but flows which exceed the bankfull channel capacity which most strongly influence bed material transport, channel morphology and channel stability (Emmett, 1976; Dunne and Leopold, 1978; Jackson and Beschta, 1982; Andrews, 1983). Other discharge-related management issues for small forest streams include prediction of design flows for culvert and bridge design, floodplain delineation, and low flow analysis for water supply and fisheries.

There are no widely used, physically based storm-period runoff-routing models for the Oregon Coast Range. Thus, certain consequences of timber harvesting, such as roads and skid trails, which change the infiltration properties of forest soils and which may influence the routing of rainfall to streams, are difficult to analyze in terms of their hydrologic effects. Furthermore, the effects of timber harvest on other storm-period hydrologic variables such as individual peak flows and the duration of very high flows are not readily predicted. Altered hydrologic regimes in turn affect channel processes and influence instream biologic values.

The purposes of the report are to:

1) Describe the rainfall-runoff processes in the Oregon Coast Range, and review the results of studies in Western Oregon on the effects of timber harvesting on runoff.

2) Review and analyze available techniques for predicting the hydrologic effects of timber harvesting in rainfall-dominated regions.

3) Provide a review of the issue of sediment routing, channel stability, and the relationship between hydrology and channel morphology.

4) Provide additional analysis of the Alsea Watershed Study data (Harris, 1977) to better describe the effects of timber harvesting on high-flow durations and assess the potential for developing and validating simple storm-period rainfall-runoff prediction procedures.

We conclude that available techniques adequately predict the effect of timber harvest on annual and seasonal water yields. Techniques for estimating design flows for instream structures (e.g. culverts, bridges) are less well developed, but generally adequate for most forest hydraulic engineering purposes. Techniques for predicting the effects of timber harvest on peak flows or channel-modifying flows (defined as flows greater than the bankfull discharge) are not well developed. Furthermore, there are no subjective or deterministic procedures for quantifying changes in channel form or function given changes in channel-forming flows. It may be difficult to separate the effects of altered hydrologic regime from changes in sediment delivery and sediment transport when analyzing channel responses to timber harvest activities.

BACKGROUND LITERATURE REVIEW

Coast Range Hydrology

The hydrology of small, low-elevation forest streams in the Oregon Coast Range is dominated by low-to-moderate intensity, long-duration frontal rainstorms, and shallow subsurface transmission of water to stream channels. A descriptive model of the subsurface stormflow process was developed by Hewlett and Hibbert (1967) and further described by Dunne (1978) and Abdul and Gillham (1984). In that model, streamflow response to precipitation occurs rapidly through a process called transelatory flow. As rainfall over a watershed increases, the area directly contributing to streamflow, the source area, increases and the active channel network expands in length and width. The expansion and contraction of drainage networks, and stream discharge rates are controlled by precipitation characteristics, the amount of water stored in forest soils, and subsurface flow characteristics of the watershed. In general, as the amount of water stored in forest soils increases, the greater is the percentage of rainfall for any given storm which appears as stormflow runoff (quickflow). Streamflow response to precipitation is directly related to slope steepness and drainage density (Harr, 1983(a)).

About 75 percent of the annual precipitation and runoff in the Oregon Coast Range occurs between October 1 and April 1. Average annual precipitation is strongly influenced by elevation and ranges from 40 in. on the east edge of the Coast Range to over 100 in. at higher elevations. Snowfall does occur, but is generally transitory at lower elevations. Melting snowpack during rainfall has contributed to some of the largest peak flows in the region. Annual runoff is roughly 70 percent of annual precipitation. Most rainfall which does not appear as runoff is lost to interception (roughly 8-20 in./yr.) and evapotranspiration (roughly 6-20 in./yr.) (Harr, 1983(a)). Both interception and evapotranspiration losses increase with increased canopy cover and increased annual precipitation.

Harris, et al. (1979) classified the entire Oregon Coast Range as a single hydrologic region for purposes of flood-frequency analysis (Figure 1). The region has similar climatic and topographic characteristics. Geologic parent materials consist primarily of sedimentary rocks of the Tyee and, to a lesser extent, the Umpqua, Yamhill and Nestucca formations. There are also scattered areas of volcanic igneous rocks and, in the extreme southern portion of the range, considerable metamorphic geologic material. Soils are generally deep and range in texture from clay loams to sandy and gravelly loams. The Coast Range is sharply dissected and has high drainage densities, which result in rapid movement of water to stream channels.

First order stream channels are usually steep and incised to bedrock. Second and third order streams are often controlled structurally by large logs, which serve to trap fluvial sediments and dissipate stream energy in steps (Swanson and Lienkaemper, 1978). These channels are often adjustable and dynamic during periods of high runoff.

Campbell and Sidel (1984) analyzed steam gage records in the coast hydrologic region of Oregon (Fig. 1) as well as 5 other hydrologic regions and found the log-Pearson type III distribution with regional skew coefficients to be suitable for use in all regions tested. The following flood-frequency regression equations were developed for the Coast Region (Campbell and Sidel, 1984).

$$Q_{10} = 0.111A^{1.04}E^{.49} \qquad r^2 = 0.83 \qquad (1_a)$$
$$Q_{25} = 0.125A^{1.01}E^{.51} \qquad r^2 = 0.79 \qquad (1_b)$$
$$Q_{50} = 0.152A^{1.01}E^{.50} \qquad r^2 = 0.79 \qquad (1_c)$$
$$Q_{100} = 0.166A^{1.00}E^{.50} \qquad r^2 = 0.78 \qquad (1_d)$$

where

Q = flow, m^2/s
A = drainage area, km^2
E = elevation, m

Lystrom (1970) also developed multiple regression equations for different flood peaks for gaged rivers in Oregon based upon watershed characteristics (e.g., slope, length, storage, elevation, area, precipitation, precipitation intensity, etc.).

Complete reviews of the hydrology of western Oregon forested regions are provided in Harr (1976(b)) and Harr (1983(a)).

Effects of Logging Activities on Hydrology

Logging activities - timber harvesting, road building and skidding - may affect almost all aspects of the forest hydrologic cycle including net precipitation, infiltration, evaporation, transpiration and the delivery of rainfall to streams. Two research programs, the Alsea Watershed Studies and the H. J. Andrews Experimental Forest, have provided most of the quantified information on the hydrologic response to logging and road building in the Douglas Fir zone of western Oregon. The Alsea watersheds, located in the Coast Range near Newport, Oregon, provide data most representative of the Coastal hydrologic region.

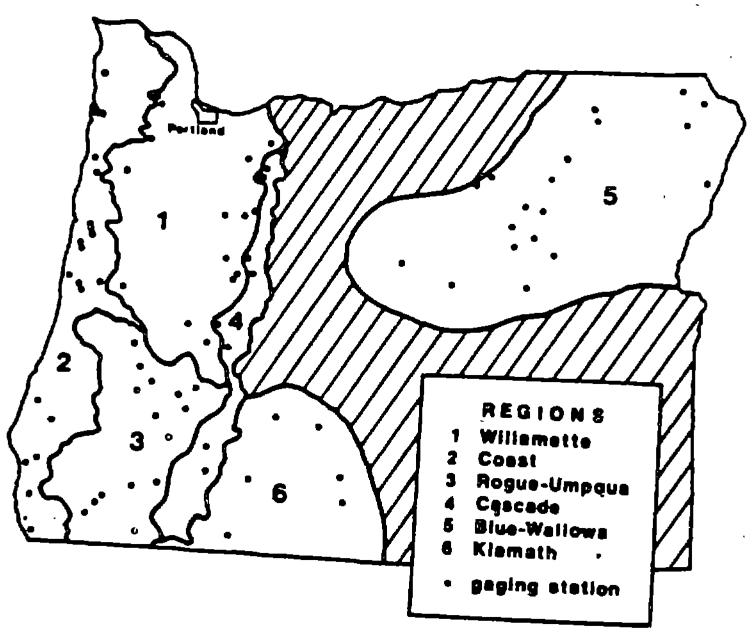

Figure 1. Flood Frequency Regions of Oregon: The Willamette, Coast, Cascade, and Rogue-Umpqua Regions are from Harris, *et al.* (1979). The Blue-Wallowa and Klamath regions were derived from topography. The cross-hatched area is left undefined because of a lack of suitable data base.

A brief summary is provided below on the effects of logging on specific aspects of the hydrologic cycle: interception, evapotranspiration, and infiltration. Streamflow response to these changes, including annual water yield, peak flows and low flows are then reviewed. The discussion is limited to the rainfall-runoff situation because of its predominance in low elevation coastal watersheds.

Interception

The amount of water intercepted by vegetation cover is a function of the form, density and surface texture of the vegetation. The amount of precipitation lost through interception is, in turn, equal to the amount of water evaporated from interception storage. Storm period interception loss increases with rainfall duration, but the proportion of rainfall lost to interception decreases as rainfall amounts increase over the storage capactiy of the vegetation. Annual interception loss increases with rainfall frequency and losses are greatest at the beginning of storms.

A common expression for storm period interception loss, I, is

$$I = c(1-e^{-p/c}) + VEt \qquad (2)$$

where:

- c = interception storage capacity
- p = precipitation amount
- V = number of vegetation surfaces
- E = evaporation rate
- t = time

In the Coast Range, c varies from less than .01 inches for deciduous trees without foliage to roughly .05 inches for mature Douglas fir. Clearcut logging reduces c to essentially zero. The recovery of c to mature forest levels is related to species composition and the rate of the reforestation process.

Increases in net precipitation following logging due to interception may be partially offset by reduced fog drip (Harr, 1982).

Evapotranspiration

Evapotranspiration is the process by which plants consume water from the soil and, ultimately, vaporize it to the atmosphere. The potential evapotranspirational rate, Es, is a maximum loss rate unlimited by soil or plant factors, but controlled by meterologic/energy factors. Actual transpiration, Ea, becomes less than Es when water availability in the soil becomes limiting. Leaf and Brink (1975) consider that Ea = Es when available soil water is more than 50 percent of field capability. The ratio Ea/Es then decreases linearly to zero as available soil water is reduced to the wilting point.

In the Alsea watersheds of the Coast Range, Harr (1983) estimated actual annual evapotranspiration was 12 to 18 inches per year. In old growth Douglas fir forest at the H. J. Andrews forest, Rothacher (1963) estimated annual evapotranspiration was roughly 16 to 18 inches per year. Clearcut logging reduces Ea to approximately zero the first year following logging (some bare soil evaporation will occur). The results of both the Alsea study and watershed studies at H. J. Andrews and elsewhere suggest that Ea recovers rapidly to pre-logging levels following reforestation. While recovery is a function of species type, site conditions and subsequent management, Ea may approach pre-logging levels in roughly 5 to 15 years in western Oregon Douglas fir forests.

Infiltration

Infiltration capacities in undisturbed Coast Range forests are high. Saturated hydraulic conductivities in excess of 118 inches per hour have been measured in Oregon Coast Range forests (Yee, 1975), explaining the general absence of overland runoff.

Road building and tractor skidding cause compacted soils, a loss of surface soil structure and reduction of macro pores. This results in greatly reduced infiltration capacities and increases the possibility for overland runoff (Johnson and Beschta, 1980). In some soils, the burning of slash following logging may also reduce infiltration capacities by creating a temporary hydrophobic condition. When soil disturbance results in surface runoff, surface erosion occurs and the translation of rainfall to streamflow may occur more quickly - possibly affecting instantaneous flow rates.

Surface runoff rates are usually high from road surfaces (Reid and Dunne, 1984). However, the runoff is susceptible to management through road placement, design, and drainage control; thus the effects on streamflow are very difficult to generalize.

The reduction in infiltration on skid trails is a function of the type of yarding (e. g., tractor, cable) - which affects the amount of compacting energy translated to the soil - soil texture, water content and organic matter content. Generally, tractor skidding on moist, fine-textured soils causes the greatest reduction in infiltration. Recovery to pre-logging infiltration capacities is dependent upon revegetation success and other soil-influencing processes such as freeze-thaw. Rehabilitation of infiltration capacities on skid trails is also subject to management through mechanical tillage or ripping, so effects on stream flow can be managed.

Annual Water Yield

The effects of logging on annual and seasonal water yield in the Douglas fir region of western Oregon have been well summarized by Rothacher (1971), Harr (1976(a)), Harr (1983(a)) and Harr (1983b), and are only briefly reviewed here.

Annual water yield increases at four clearcut watersheds at H. J. Andrews and Alsea watersheds ranged from about 4 inches per year to over 23 inches per year. Eighty-seven percent of variation in the annual increases, Y, at the H. J. Andrews studies was explained with a two parameter linear model incorporating number of years after logging, X_1, and annual precipitation, X_2.

$$Y = 31.41 - 2.08X_1 + 0.091\ X_2 \qquad (3)$$

In general, water yield increases are greatest in wetter years and in the fall and spring, and decrease over time following the recovery of vegetation. Reductions in interception and evapotranspiration losses, as described above, account for the observed increases in yield.

We developed a similar regression model for annual flow increases at Needle Branch in the Alsea Watersheds following logging:

$$Y = -0.32 - 0.21X_1 + 0.17\ X_2 \qquad (4)$$
$$r^2 = 0.42$$

(Std. error of the estimate of Y = 3.8 in.)

However, the post-logging period was only 7 years at Needle Branch which is probably inadequate for this type of analysis.

Increases in annual water yields have also been measured for partially logged watersheds (Harr, 1983b). Increases occurred in both patch-cut and shelterwood-cut watersheds, but were generally less than for clearcut watersheds. While increases in water yield can be expected to be related to the percent of the watershed logged, significant increases in annual water yields were not detected at one watershed at the H. J. Andrews forest until 40 percent of the watershed was clearcut. When a smaller percent of the watershed is logged, it is possible that reductions in Es are, in part, compensated for by increases in Ea in the remaining stand. Also, the location of a cut within a watershed may also influence its effect on runoff.

Peak Flows

The effects of logging on the size of peak flows - particularly large peak flows - are not easily generalized. Changes in seasonal or annual runoff volumes following timber harvest are well-documented and well-explained by net changes in the water budget. However, the timing, or rate, of runoff is another variable which influences the shape of the hydrograph and the size of peak flows. Therefore, to predict the effects of logging on peak (or instantaneous) flows, it is necessary to :

(1) Describe the effects of an altered water budget on the volume of storm-period runoff, and
(2) Describe the effects on the routing of rainfall to stream channels (i.e., the distribution of runoff over time).

The results of existing watershed studies strongly suggest that the effects of logging on peak flows in this region are to:

(1) Increase small peak flows in early fall and spring due to increased runoff volumes which are the result of higher soil water contents in logged areas than forested areas during those seasons, and

(2) Increase larger peak flows which result from rainfall on saturated soils in those situations where the routing of rainfall to streams has been accelerated (Harr, et al. 1975; Harr, 1976(a)).

In addition, there is a growing body of literature which suggests that the snow accummulation and melt process (including rainfall on snow) is sufficiently affected by logging to influence the size of peak flows in this region (Harr et al. 1982, Christner and Harr, 1982; Harr and Berris, 1983).

The principal mechanism for increasing the rate of delivery of rainfall to stream channels is to reduce infiltration to the point where overland runoff occurs - thus altering, in part, the routing mechanism from interflow to overland flow. This occurs primarily on road surfaces and severely compacted soils such as may occur on skid trails and landings.

In those cases where large peak flows have been significantly increased (e.g., Alsea study - Deer Creek 3), large proportions of the watershed (12%) were occupied by roads, skidtrails and landings (Harr, 1975; Harr, 1976(a); Harr, 1983(a)). In studies of the effects of cutting alone, when roads, trails and landings occupied a negligible percent of the watershed, large peak flows were generally unaffected, even though smaller peak flows (and the average of all peak flows) were increased (Rothacher, 1971; Harris, 1977; Harr et al., 1975; Harr, 1976(a); Harr, 1983(a)).

Whether roads, skidtrails and landings increase large peak flows and how large any increases will be is difficult to predict and is dependent upon quantification of both the pre- and post-logging hydrologic routing systems and percent of total watershed area these facilities occupy. While models of both subsurface storm flow routing (Sloan, et al. 1983) and surface runoff (e. g., Simons, et al. 1975) exist, they have not been combined and validated in this region. It is probable, however, that the percent area in roads, skid trails and landings is only one variable to be considered and that the location, design (including drainage design) and, if appropriate, reclamation of these features are also important factors influencing routing. All of these factors are subject to management. Thus, effects of timber harvest on peak flows may be very difficult to generalize or quantify.

Low Flows

The effects of logging on low flows are summarized in Harr, (1976(a)), Harr (1983(a)) and Harr (1983b). Increased low flows (or decreases in the number of low flow days) occurred following logging at the H. J. Andrews and the Alsea watersheds. Similar increases have been measured at other sites in western Oregon. While the increases in low flows were proportionally large (roughly 100%) between treated and controlled watersheds, they accounted for only a small amount of the total annual water yield increases.

Furthermore, the increase were short lived, with recovery to pre-logging conditions occurring in roughly five years.

Increases in low flows are generally explained by reduced evapotranspirational demand following logging. However, in at least one case at the Bull Run principal watershed near Portland, Oregon, low flows were decreased following patch cutting. This decrease has been attributed to reduced precipitation in the form of fog drip (Harr, 1980; Harr, 1982). Revegation of streamsides by phreatophytes may also reduce low flows compared to the prelogged condition until after they become overtopped by Douglas fir and Hemlock (Harr, 1983(b)).

Effects of Hydrology on Channel Processes

Flow, Qw, and sediment discharge, Qs, are the primary variables influencing the morphology of stream channels with adjustable beds and banks (Schumm, 1971). Alluvial channels are considered to be "graded" if their resulting hydraulic characteristics are the minimum required to transport the sediment delivered to the channel (Leopold and Maddock, 1953). Channel geometry and hydraulic characteristics will change when either flow, sediment discharge or both increase. Schumm (1971) proposes that a channel will widen when either Qw, Qs or both increase. Channel depth will increase with an increase in Qw but will decrease when Qs increases.

Based upon the above criteria, a stream channel should respond to an increased flow regime by becoming wider and deeper. However, increased flows following logging are almost always accompanied by increased sediment delivery to streams (Brown, 1980). Thus, in most documented changes in channel characteristics following logging, the effects of increased Qs predominate and channels tend to become wider and shallower (Lyons and Beschta, 1983; Beschta, 1984). Jackson and Beschta (1984) suggest that the initial response of a stream channel to increases in Qs will be to increase its hydraulic efficiency by reducing form roughness. Reduced form roughness is accomplished by reducing, or smoothing, the channel's pool-riffle pattern. This permits increased velocities and an increased capacity to transport water and sediment. Subsequent responses are reduced channel stability, increased channel width, and decreased channel depth. Reduction or elimination of pool-riffle patterns, reduced bed and bank stability and changes in particle size characteristics of substrate all influence the biologic quality of the channel (Moring and Lantz, 1975).

While the above-channel responses may occur following logging <u>if</u> there are sufficient increases in Qw and Qs, the hydraulic geometry equations cannot be solved explicitly - therefore the exact response is difficult to predict. It becomes even more difficult to separate out the effects of increased flows, alone. We do know, however, that not all flows affect channel processes, and therefore, channel morphology. There are flows below which there is very little, if any, movement of bed or bank material. When a channel has a sufficient supply of large bed material armoring occurs and flows required to initiate bed material movement (critical flows) may become quite large.

Several studies have provided data which suggest that critical flows for many alluvial channels approximate the bankfull discharge and have roughly a 1.5 to 2-year return period. Andrews (1983), in a study of 24 self-formed gravel-bed rivers in Colorado, determined the critical dimensionless shear stress required to entrain given particle sizes. He found that while smaller particles were frequently entrained, the larger (e.g., d_{90}) particle sizes were not entrained until roughly the bankfull discharge. At the Snake and Clearwater Rivers in Idaho, Emmett (1976) observed that bed-load transport rate was a different (and higher) function of unit stream power when the stream was capable of transporting all particle sizes than at lower stream powers where bed armoring progressively limited the bed material available for transport. Sands were transported at lower flows, whereas coarse gravels were primarily transported upon disruption of the armor.

Jackson and Beschta (1982) described bed material routing in a sand and gravel-bed channel at Flynn Creek in the Alsea watersheds. At discharges less than bankfull, bedload transport consisted primarily of sand-sized materials being routed over stable gravel riffles. At approximately the bankfull discharge, riffle materials were also transported downstream. However, because of nonuniform channel geometries, bedload transport rates were unsteady and several sequences of partial riffle scour and fill occurred. Upon the recession of high flows, a final riffle deposition and a natural sorting of particle sizes occurred. The bed material composition of the riffles was considerably more coarse than that of the bedload sediment in transport at the time riffle sediments are deposited. "Left-over" sand-sized bed material therefore deposited elsewhere in the stream channel at lower flows - generally in pools, channel edges and backwaters.

In summary, increases in Qs and Qw seem to be the principal factors effecting changes in channel morphology. While these two factors interact to influence channel geometry and channel processes, net channel response seems to reflect the effect of increased Qs over increased Qw. There is evidence that channels are most responsive to flows greater than approximately bank- full, when stream powers are sufficient to mobilize the entire range of sizes of bed and bank materials. Lower flows serve primarily to transport sand sized materials and redistribute sands to pools and other areas of relatively low shear stresses.

REVIEW AND ANALYSIS OF QUANTITATIVE TECHNIQUES FOR ASSESSING HYDROLOGIC EFFECTS OF LOGGING ACTIVITIES

Introduction

A comprehensive procedure for evaluating the effects of silvicultural activities on the hydrologic cycle is provided in WRENS (Troendle and Leaf, 1980). WRENS may be used to analyze changes in the seasonal or annual water budget for forested watersheds and can produce, as output, change in the 6-day flow duration curve for a watershed. The hydrology component of WRENS uses two existing models. The PROSPER model (Goldstein, et al., 1974) is used for rainfall-runoff situations and the WATBAL model (Leaf and Brink, 1973) is used for snowmelt-runoff situations. Other procedures, for example the HYSED, procedure developed for Region 2 of the Forest Service (Silvey and Rosgen, 1981) use modified versions of WRENS for their hydrologic components. Most procedures incorporating routing routines which are capable of analyzing individual storm events and instantaneous flows are either still under development, or are not validated for western Oregon. Sloan et al. (1983) evaluated several subsurface - storm-period models on steep- sloped forested watersheds in the eastern United States. While their models have not been validated in Western Oregon, they may provide a basis for eventual development of a physically based, storm-period model for this region. Overton and White (1977) developed a storm-period runoff model for two watersheds in the H. J. Andrews forest, but discontinued model refine- ment due to a lack of information on several ecosystem processes and a need for a snowmelt subroutine.

One method was reviewed which predicted changes in water yield and channel stability for the Kootenai National Forest (Galbraith, 1973). This method, however, assumes that water yield - not necessarily flows over some level - affect channel stability. This assumption is not well-supported in the literature.A more detailed review of the WRENS procedure, in particular the PROSPER model, is provided below. Additional reviews of the Overton and White model and the models reported in Sloan, et al. are also provided.

WRENS

The hydrology component of WRENS was developed to assist professionals in the analysis of the effects of logging - primarily vegetation removal - on seasonal and annual runoff volumes. Model parameters have been generalized for eight physiographic regions - one of which encompasses the coastal forests of northern California, Oregon and Washington. For rain-dominated regions below 3,000 - 4,000 feet in the Pacific Coast, the PROSPER model (Goldstein, et al., 1974) is applied. The model has been tested on Needle Branch of the Alsea Watersheds, with the objective of extrapolating local observations for regional use.

PROSPER is strictly a water balance model with no surface or subsurface routing components. PROSPER emphasizes the evapotranspiration component of the hydrologic cycle by integrating the interaction between available soil water, precipitation, and energy. Regional evapotranspiration data is used by the model. Leaf area index is used as the primary evapotranspiration modifier. In addition, the evapotranspiration function is affected by changes in available water caused by altering rooting depth. Water available for streamflow is determined by subtracting seasonal evapotranspiration from net precipitation. The model can handle several silvicultural prescriptions within a watershed.

A schematic of the PROSPER model as provided in the WRENS documentation (Troendle and Leaf, 1980) is reproduced in Appendix I.

The WRENS-PROSPER procedure provides a simple physically based procedure for calculating the effects of logging on seasonal and annual runoff. It has been validated at Needle Branch in the Oregon Coast Range. Because regional evapotranspiration potential is fairly uniform and not particularly sensitive to aspect and slope during the rain season, and because it can handle differences in watershed soils, WRENS-PROSPER should be fairly transferable to other watersheds in the Coast Range. Functions relating evapotranspiration to leaf area index were developed in the southeastern United States. They still need additional validation in the Oregon Coast Range.

The main drawback to WRENS-PROSPER is that the model output is not particularly useful in the analysis of many hydrology related management issues including design-flow hydrology, floodplain hydrology, or channel processes.

Overton-White

Overton and White (1978) developed several versions of a physically-based storm-period rainfall-runoff model for the H. J. Andrews experimental forest as part of a study of modelling strategy. Their objective was to develop a submodel of an ecosystem model and as such were less concerned with the accuracy of the hydrologic output than with the capability of the model to interact realistically with other ecosystem components. The model was developed for Watershed 10 of H. J. Andrews Forest, but some input data were generalized from Watershed 2 and other local data sets. As in PROSPER, the model is a water balance-type model, but unlike PROSPER it does attempt to represent both soil water flow and groundwater lateral flow. Also, by incorporating an infiltration routine, the model is capable of generating an overland flow component. By compartmentalizing the model into above-ground and below-ground components, it was shown that the above ground component - primarily evapotranspiration-controlled runoff volume, and the below ground component - which generally related to storage routing - controlled hydrograph timing and shape. Model output is daily discharge in m^3.

Additional descriptions of the models from Overton and White (1978) are provided in Appendix II.

Overton and White discontinued model refinement because they felt certain processes required more complete specification and because they felt a need for an improved snowmelt routine. No additional attempts have been found in the literature to further validate or refine the Overton-White models. Also, their ability to predict changes due to logging activities have not been tested. Expanding source areas are not well incorporated conceptually by the Overton-White models.

In addition to an improved snowmelt routine, additional work in modelling interflow and a capability to subdivide watersheds into "routing" or "response" units might improve the Overton-White models.

Some of these shortcomings have been addressed in other large, distributed parameter models - for example the USGS-PRMS modelling system (Leavensly et al. 1983). These models also have fairly primitive subsurface routing routines, fail to represent the concept of expanding source areas and would require calibration and validation in the Coast Range.

Most physically based distributed parameter storm-period models are data intensive and not conducive to quick "planning level" hydrologic analyses. However, as research tools and when applied to special situations, they offer potential for far more detailed insight into hydrologic systems - including instantaneous flows - than simple water budget models.

Sloan, Moore, Coltharp and Eizel

Sloan, et al. (1983) provide a thorough report on modelling surface and subsurface stormflow on steeply-sloping forested watersheds. Their objective is to identify and develop techniques for modelling rainfall-runoff consistent with the theories of variable source areas-interflow. They review the Variable Source Area Simulator Model (VSAS) (Troendle and Hewlett, 1979). They also review five subsurface flow models for homogeneous soils: two of the models were based on the Richards equation, on on the kinematic wave equation and two storage models - the Boussenesq and kinematic storage models. Finally the authors develop and test a simple continuous daily rainfall-runoff model based on the theories of interflow and variable source areas.

VSAS

The Variable Source Area Simulator Model (VSAS) was developed for small, forested watersheds and is based on the concept that instantaneous streamflow is the sum of subsurface flow, precipitation on channel and saturated zones and overland flow from impervious areas (Troendle and Hewlett, 1979). Subsurface flow is described by the Darcy and Richards equations, where unsaturated conductivity is related to water content by

$$K(O) = a\ e^{b} \qquad (5)$$

where a and b are constants (Green and Corey, 1971). The watershed is partitioned into segments and elements (including at least two soil layers) and flow goes from one element to another based upon the above laws of subsurface flow. A VSAS model required detailed descriptions of watershed soils and as such is data-intensive and probably best suited to research.

Sloan, et al.

A simple, daily rainfall-runoff model for forested watersheds was developed by Sloan, et al. (1983). The model consists of three water stores - an interception zone, a soil zone, and a groundwater zone. All net precipitation is assumed to enter the soil zone - except on saturated areas where it becomes runoff. The size of the saturated zone increases as the soil zone becomes wetter. Subsurface drainage from the soil zone increases exponentially as the water content increases. Evapotranspirational loss is limited either by potential evapotranspiration or available soil water. A model schematic, function descriptions and parameter descriptions are provided in Appendix I.

The model was calibrated and tested on the Little Millseat Watershed in eastern Kentucky using a split record technique. Daily streamflows were well predicted by the model.

Application of the model requires partitioning a watershed into a series of interconnected water stores, and as such is somewhat data intensive. However, because it is a conceptual lumped-parameter representation of the rainfall-runoff process, it has easier data requirements than more distributed parameter models such as Overton-White or PRMS. Its greatest strength may be in its conceptual handling of variable source areas. The model has not been tested in western Oregon but, because of its representation of expanding source area-interflow processes, it can be expected to apply to the Coast Range forested situation. Also, its ability to model the effects of logging has not been tested. By adding an infiltration routine (presently all net precipitation is infiltrated) the model should be suited to an analysis of the effects of roads and skidtrails on runoff because it allows the watershed to be subdivided into water "stores." Like Overton-White and PROSPER, the model needs a snowmelt routine and channel routing routine.

EFFECTS OF LOGGING ACTIVITIES ON FLOWS GREATER THAN BANKFULL:
A REVIEW OF THE ALSEA WATERSHED STUDY

Introduction

The key variable with respect to high flow influences on channel processes is the total force exerted on the channel bed and banks in excess of the critical force required for incipient bed or bank instability. If channel instability occurs at bankful or greater discharges, we can derive either average bed shear stress, τ_o, or unit stream power, VS, from discharge and channel geometry information and integrate for total force or stream power expended during bankfull or greater flows. If we assume the hydrograph peak can be approximated by a triangle, synthetic peak flow (greater than bank-full flow) hydrographs can be developed given estimates of peak flow, qp, and the duration of bankfull or greater flow, D_b. We can then solve for τ_o or VS given relationships between discharge and average flow velocity or discharge and hydraulic radius.

We hypothesized that q_p should be predictable from precipitation for a saturated watershed. To test our hypothesis, we tried to predict peak flows from precipitation and antecedent streamflow on the alien watersheds. We also hypothesized that D_b, would also be predictable. We attempted to test the effects of timber harvest on D_b by reviewing available data from the alien watersheds. Our analyses are exploratory and not complete. However, we wanted to determine if additional, more thorough analylses of the type might improve our ability to quickly estimate qp, and D_b, and eventually the effects of timber harvesting on channel modifying flows, without having to apply more cumbersome instantaneous flow models.

Review of Alsea Results

The effects of logging on high flows at two of the Alsea watersheds have been well analyzed by Harris (1977) and Harr, et al. (1975).

Harris analyzed all peak flows greater than 50 csm (roughly bankfull discharge) on Flynn Creek (the control) and concurrent peaks on Needle Branch and Deer Creek. The mean of these larger peaks was increased by 17.8 csm over the predicted mean of 91.2 csm on Needle Branch following 82 percent clearcutting and burning. This increase was not considered significant ($p = 0.95$). Five percent of the Needle Branch was in roads or skid trails. The mean large peak flow on Deer Creek was increased by 1.8 csm over the predicted mean of 78.5 csm following 25 percent patch cutting. Four percent of Deer Creek was in roads or skid trails. The increases in mean peak flow at Deer Creek also were not significant ($p = 0.95$).

The three-day runoff volumes associated with high instantaneous peak flows were analyzed. A significant increase of 1.21 in. was detected on Needle Branch (Figure 2). An increase of 0.10 in. for Deer Creek was not significant ($p = 0.95$).

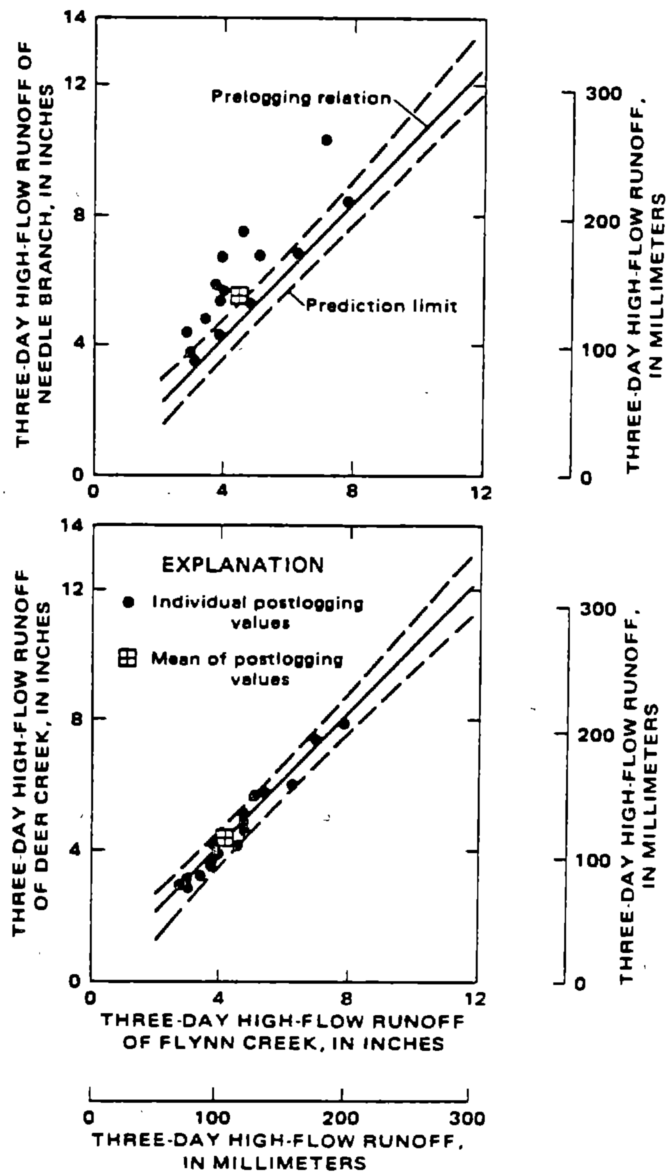

FIGURE 2 —Relation of 3-day high-flow runoff of Needle Branch and of Deer Creek to that of Flynn Creek. (Harris, 1977)

Harris' analysis suggests that while the average instantaneous peak flow for events greater than 50 csm did not increase significantly, runoff volumes associated with those flows did increase on Needle Branch. In effect, he is suggesting that the duration of high flows above some level was increased. Harr, et al. (1975) in their analysis of peak flows at the Alsea Watershed, included smaller peak flows than those used by Harris. Their average peak of 37 csm was roughly one-half that of Harris' peaks. What is significant, however, is that by including these smaller peaks, the average peak flow on Needle Branch increased significantly ($p = 0.99$) by 16. csm in the fall and

10 csm in the winter after the watershed was 82% clearcut. At Deer Creek (sub-watershed 3) increases of 3 csm in the fall and 12 csm in the winter were both highly significant. Deer Creek 3 was 65% clearcut and had 12% of the watershed in roads, landings and tractor skid trails. At Deer Creek 4, fall-period peak flows increased 27 csm. Fall peaks increased 16 csm and winter period peaks increased 10 csm following 90% clearcutting at Needle Branch. These increases were highly significant ($p = 0.99$).

Again, the implication of Harr et al. is that while increases in larger peaks were difficult to detect (and apparently related to road building and soil compaction), increases in lower instantaneous flows were significant ($p = 0.99$). The question becomes how much, if any, of the increases in instantaneous flows occurred as increases in the duration of channel-forming flows (flows over 50 csm bankfull discharge)? The next section provides some additional analysis of changes in the duration of flows greater than bankfull in the Alsea Watersheds following logging.

Analysis of High-Flow Duration

Bankfull discharge was estimated to be 50 csm for the Alsea watersheds. The daily flows associated with instantaneous peak flows of 50 csm or higher were determined by regression analysis for each of the three Alsea watersheds. These flows are shown below:

Stream	Bankfull Flow, cfs	Daily Flow Associated with a Bankfull Event, cfs
Flynn Creek	39	30 ($R^2 = 0.95$)
Needle Branch	13	9 ($R^2 = 0.81$)
Deer Creek	58	43 ($R^2 = 0.95$)

Days when flow exceeded the threshold flow were then tallied for each watershed and each year. These results appear in Table 1. The numbers represent the duration (days) of time that flow was likely to be at or above bankfull.

To test the effect of timber harvesting on this variable, we did covariance analyses of the high-flow durations of both Needle Branch and Deer Creek, using Flynn Creek as the control watershed.

Table 1. Duration of Flow at or Above Bankfull for Three Alsea Watersheds.

Water Year	Flynn Creek	Needle Branch	Deer Creek
		(Days)	
1959	2	4	4
1960	2	2	2
1961	7	7	9
1962	4	4	4
1963	2	3	3
1964	4	7	5
1965	11	12	12
1966	8	9	9
1967	4	11	3
1968	1	13	1
1969	7	15	6
1970	5	11	6
1971	5	12	10
1972	8	19	9
1973	3	8	3
Pre-treatment mean			
(1959-65)	5	6	6
Post-treatment mean			
(1967-73)	5	13	5

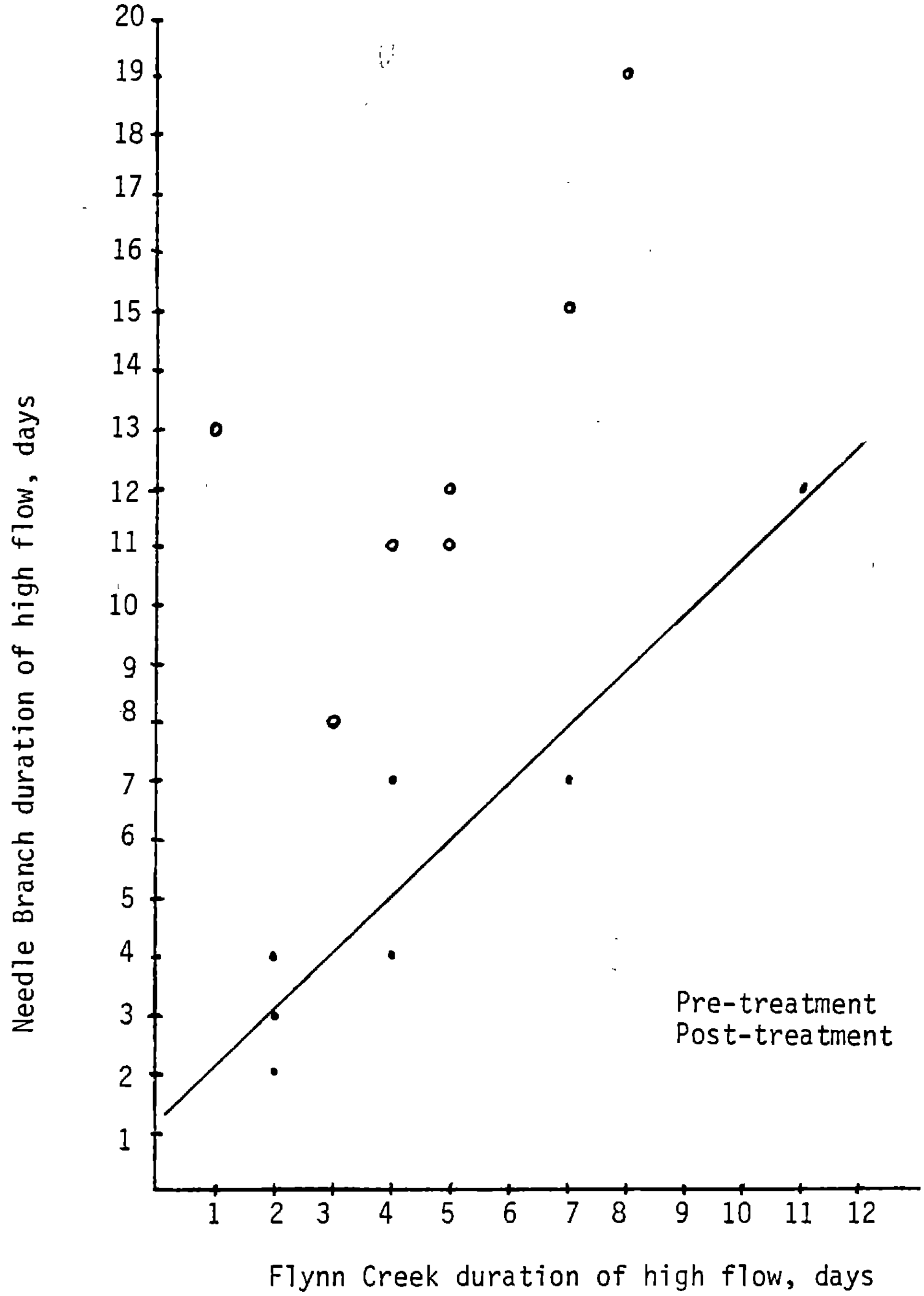

Figure 3. Regression of Needle Branch on Flynn Creek pre-harvest annual high flow duration. Post-harvest data points are also plotted for comparison. The pre-harvest relationship is (Needle Branch = 0.96 X [Flynn Creek] + 1.20). R^2 = 0.89.

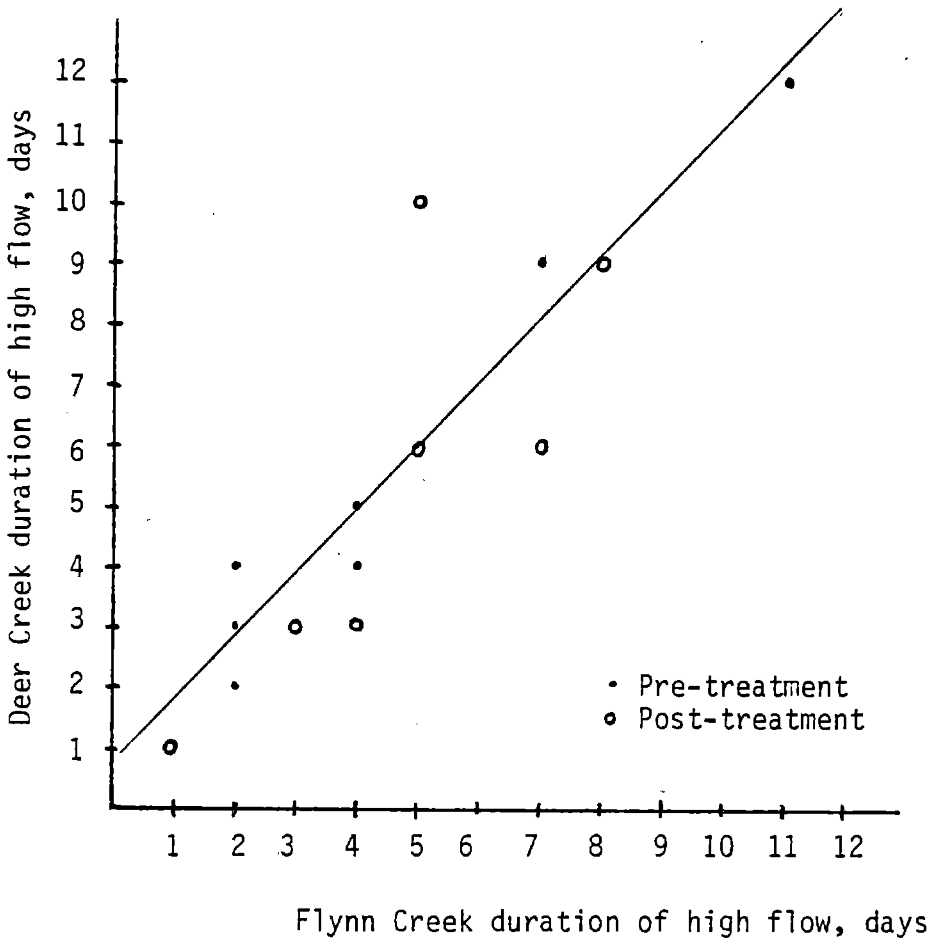

Figure 4. Regression of Deer Creek on Flynn Creek pre-harvest annual high flow duration. Post-harvest data points are also plotted for comparison. The pre-harvest relationship is (Deer Creek = 1.04 X [Flynn Creek] + 0.80). $R^2 = 0.95$.

Clearcutting of Needle Branch resulted in a 123% average increase in the duration of high flows ($p = .99$). All post-treatment years showed increases (Fig. 3). The analysis of Deer Creek, which was partial-cut, showed no difference in high flow duration as a result of the treatment ($p = .95$). One year, 1971, had a significant post-treatment response (Fig. 4).

Peak Flow Prediction: Regression Approach

Sloan, et al. (1983) report on investigations by Corbett (1979) on the quickflow response of a small forested watershed in Pennsylvania. The antecedent flow rate was found to have the greatest correlation with quickflow of all variables investigated. Antecedent flow was believed to be the best overall indicator of the extent of saturated source area within the watershed. The purpose of this section is to develop a two parameter linear model for peak flow prediction for the Alsea Watersheds of the form

$$Q_p = b_o + b_1 \, Qant + b_2 \, P_{24} \qquad (6)$$

where

Q_p = peak flow greater than 50 csm
$Qant$ = mean daily streamflow one day prior to the peak flow
P_{24} = 24 hour precipitation total for the day of peak flow
b_o, b_1, b_2 = regression coefficients

The objective was to see if (1) this model accurately predicted peak flows in the Alsea Watersheds, and (2) any changes in regression coefficient from the pre- to the post-logging period could be attributed to logging. Runoff data is from the USGS Water Supply Papers for Oregon. Precipitation data is from the National Weather Service 24-hour rain gage at Tidewater, Oregon. Ideally, one-hour precipitation data collected on the watersheds would have been available. Also, the Tidewater gage was read daily at 5:00 p.m., so the 24-hour amounts don't correspond precisely with the 24-hour periods for the runoff data.

Results of the stepwise multiple linear regression analysis are provided in Table 2. Means of variables are provided in Table 3 and correlation coefficients are provided in Table 4. Analyses were conducted both with all flows greater than 50 csm and with only those events which caused peaks greater than 50 csm on Flynn Creek (the "control" watershed). The ratio of the number of 50 csm events on Deer Creek and Needle Branch to those on Flynn Creek increased from 1.21:1 and 1.57:1 to 1.38:1 and 2.19:1 for Deer Creek and Needle Branch respectively in the post-logging period.

In general, the regression model proved to be a good predictor of instantaneous peak flow based upon the fairly high (.6 to .8) r^2 values. However, the entire correlation structure within the model changed in the post-logging period - even for the control watershed, so the model lacked sensitivity to changes caused by management.

Table 2a

Predictive Equations for Flows Over 50 csm in Flynn Creek

Stream	Model	Pre Logging Predictive Equation	r^2	Post Logging Predictive Equation	r^2
Flynn	1 variable	$Q_p = 21.3 + 16.0\ P_{24}$	.36	$Q_p = -15.1 + 29.9\ P_{24}$	.67
	2 variable	$Q_p = -9.4 + .8\ Qant + 17.6\ P_{24}$	.62	$Q_p = -71.0 + 1.4\ Qant + 37.6\ P_{24}$	.81
Deer	1 variable	$Q_p = 22.6 + 17.3\ P_{24}$	.53	$Q_p = .1 + 26.6\ P_{24}$	.65
	2 variable	$Q_p = 12.3 + .6\ Qant + 14.6\ P_{24}$	.77	$= -62.0 + 36.2\ P_{24} + 1.3\ Qant$	.80
Needle	1 variable	$Q_p = 19.7 + 22.2\ P_{24}$	.58	$Q_p = .6 + 35.9\ P_{24}$	.69
	2 variable	Qant not significant		Qant not significant	

Table 2b

Predictive Equations for All Flows Over 50 csm

Stream	Model	Pre Logging Predictive Equation	r^2	Post Logging Predictive Equation	r^2
Flynn	1 variable	$Q_p = 26.7 + 1.5$ Qant	.61	$Q_p = -15.1 + 29.9 P_{24}$	.67
	2 variable	$Q_p = -34.5 + 1.5$ Qant $+ 20.0 P_{24}$	.80	$Q_p = -71.0 + 1.4$ Qant $+ 37.6 P_{24}$	.81
Deer	1 variable	$Q_p = 42.4 + 1.2$ Qant	.78	$Q_p = -2.6 + 25.9 P_{24}$	.59
	2 variable	$Q_p = 15.9 + 1.1$ Qant $+ 9.9 P_{24}$	.85	$Q_p = -43.6 + 1.0$ Qant $+ 32.1 P_{24}$	.75
Needle	1 variable	$Q_p = 57.0 + 1.1$ Qant	.65	$Q_p = 10.5 + 30.9 P_{24}$	.58
	2 variable	$Q_p = 12.0 + .9$ Qant $+ 16.8 P_{24}$	.79	Qant not significant	

Table 3a

Means and Standard Deviations (S.D.) for Flows

Over 50 csm in Flynn Creek

		Pre Logging				Post Logging			
Stream		Q_p	Qant	P_{24}	Number of Events	Q_p	Qant	P_{24}	Number of Events
Flynn	mean	70.6	31.0	3.1	13	75.1	23.0	3.0	16
	s.d.	20.1	12.3	.8		36.9	11.1	1.0	
Deer	mean	75.8	28.8	3.1	13	80.3	25.3	3.0	16
	s.d.	18.0	14.2	.8		33.1	12.1	1.0	
Needle	mean	88.1	31.3	3.1	13	109.0	35.9	3.0	16
	s.d.	22.3	14.7	.8		43.7	16.9	1.0	

Table 3b

Means and Standard Deviations (S.D.)

for All Flows Over 50 csm

		Pre Logging				Post Logging			
Stream		Q_p	Qant	P_{24}	Number of Events	Q_p	Qant	P_{24}	Number of Events
Flynn	mean	78.1	34.6	3.1	14	75.1	23.0	3.0	16
	s.d.	34.2	18.0	0.7		36.9	11.1	1.0	
Deer	mean	77.8	29.9	3.0	17	74.1	21.4	3.0	22
	s.d.	30.0	22.4	0.8		29.8	12.5	0.9	
Needle	mean	88.5	29.8	3.0	18	93.5	27.2	2.7	35
	s.d.	32.4	24.6	0.8		35.3	18.0	0.9	

Table 4a

Correlations for Flows Over 50 csm

in Flynn Creek

	Pre Logging Correlations			Post Logging Correlations		
Stream	Q_p x Qant	Qp x P_{24}	Qant x P_{24}	Q_p x Qant	Qp x P_{24}	Qant x P_{24}
Flynn	.43	.60	-.12	-.07	.82	-.49
Deer	.64	.73	.22	-.19	.81	-.61
Needle	.39	.76	.17	-.26	.83	-.23

Table 4b

Correlations for All Flows Over 50 csm

	Pre Logging Correlations			Post Logging Correlations		
Stream	Q_p x Qant	Qp x P_{24}	Qant x P_{24}	Q_p x Qant	Qp x P_{24}	Qant x P_{24}
Flynn	.78	.46	.04	-.07	.82	-.49
Deer	.89	.51	.31	.03	.77	-.43
Needle	.80	.63	.33	.04	.76	-.11

In the pre-logging period, for those storms which caused peaks greater than 50 csm on Flynn Creek, antecedent flow was positively correlated to peak flow. Twenty-four hour precipitation, already the strongest predictor, became even more strongly positively correlated with post-logging peaks. The ratio of Qant/Qpeak decreased slightly from roughly 0.39:1 to 0.32:1 in the post-logging period for all three watersheds.

When the peak-flow regression analysis was originally tested, we hypothesized that a possible effect of logging might be to decrease the ratio of

Qant/Qp in the post-logging period, resulting in a higher peak flow for a given precipitation event. This result could also be reflected in an increase in the regression coefficient for P_{24} which would increase Q_p for a given P_{24}. While the coefficient for P_{24} increased in the post-logging period it did so on Flynn Creek too. Therefore, our analysis did not provide evidence in support of our initial hypotheses. This type of analysis might be more fruitful given shorter rainfall intensity data (e. g. 6 hr.), an on-site precipitation gage and a lower peak-flow threshold. Hewlett and Fortson (1984) found, however, that inclusion of maximum hourly rainfall intensity did not greatly improve peak flow predictions in 15 drainage basins when gross storm rainfall and pre-storm flow rate were included as independent variables.

RECOMMENDATIONS

The following are generalized recommendations, and should be applied only as one aid to professional judgement and after an understanding of their background and derivations.

Annual Flows

Annual flows are increased following logging, primarily because of reduced evapotranspiration. The increase is dependent upon vegetation type, annual precipitation, years since logging, and other factors. The increase can be estimated by equation (3) which was developed at the H. J. Andrews Forest. A more site-specific procedure recommended for estimating seasonal or annual runoff changes due to logging is provided in WRENS.

Design Peak Flows

In low elevation, rain dominated zones in the Oregon Coast Range, logging probably does not have a major effect on the 20 to 100-year return period flows generally used in culvert and bridge designs. To estimate large design flows a Log-Pearson III flood frequency analysis with regionalized skew coefficients is recommended on gaged basins. Any one of several empirical relationships, such as equations (1a-1d), can be used to estimate design flows on ungaged basins. We tried to validate the utility of the SCS Curve Number procedure for predicting design flows, and the effects (on Curve Number) of timber harvest. However, we could not identify a suitable data base for the analysis. Furthermore, 24-hour design storms are inadequate descriptors of major runoff-producing events.

Design flows may be affected somewhat when over 6-12% of a watershed is in roads or severely compacted skid trails, landings, etc. In this case, a small safety factor (15-20%) may need to be applied to design flows, and/or extra considerations may need to be given to location, design, drainage and rehabilitation measures for roads and compacted areas.

No consideration is given here to peak flows affected by snow accumulation and melt.

Bankfull Flows

Clearcut logging may increase the duration of flows over bankfull by roughly 120 percent. These increases will decay following revegetation and also will be proportionately smaller as smaller percentages of the watershed are logged. The effect of these larger (channel-forming) flows on channel stability and channel morphology will be highly dependent upon the corresponding increase in sediment delivery - particularly in the bed-material sizes. While channel stability may be reduced following logging by the increased duration of greater than bankfull flows, longer-lasting impacts on channel morphology can be expected when increased erosion (particularly mass erosion) results in increased delivery of bed load sediment to channels.

Daily or Peak Discharge Models

No daily or peak discharge models were identified which have been adequately validated for use in the Oregon Coast Range. The Overton-White Models (Overton and White, 1978) and the simple daily rainfall model developed by Sloan, et al. (1983) deserve additional testing, and validation in the Coast Range. Further investigation of the simple multiple linear regression approach as presented above (eqn. 6) for predicting storm-period runoff volume and instantaneous peak flow is also recommended - with special emphasis on relating model parameters to watershed variables.

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APPENDIX I

PROSPER

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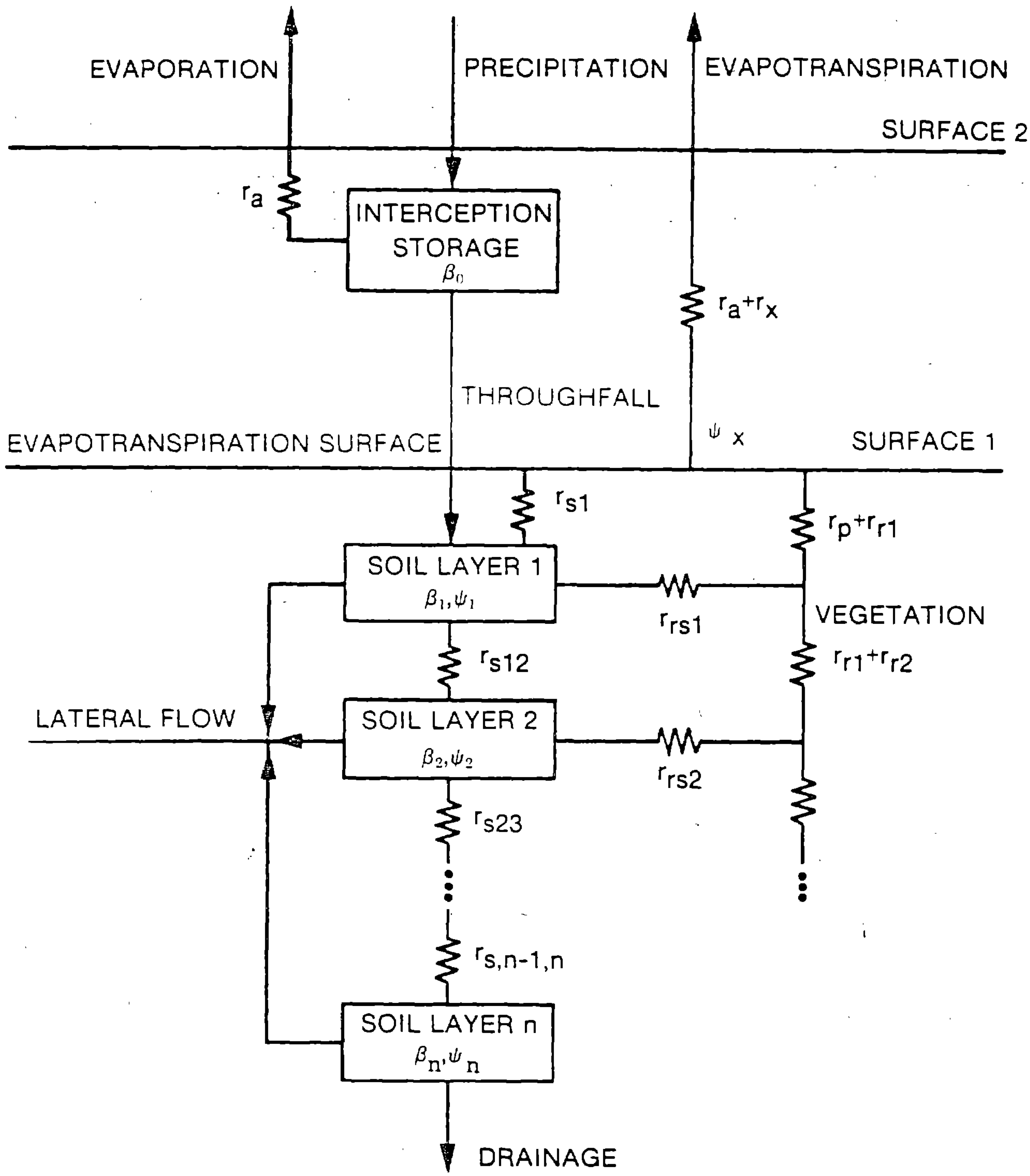

Figure III.B.2.—Schematic of PROSPER (from Goldstein and others 1974).

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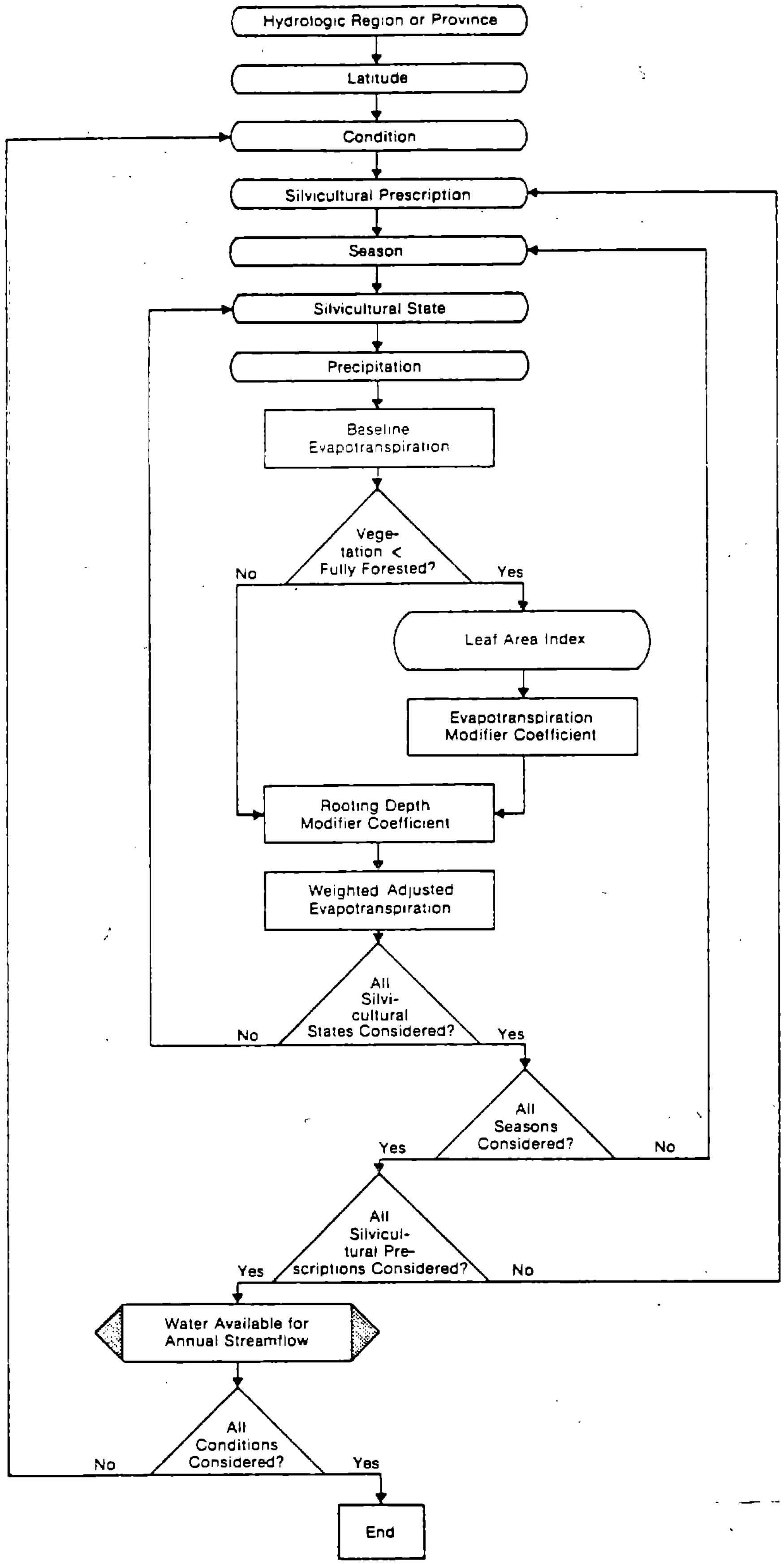

Figure III.9—Flow chart of methodology for determining evapotranspiration and water available for annual streamflow in rainfall dominated regions. (WRENS)

OVERTON-WHITE

FLEXFORM ABSTRACT Hydrology Model II
Resolution, one day; Date: 4/24/73
units, m^3/ha Revised: 7/11/73
(Figure 2–b)

Model II–a

State Variables

x_1 Canopy storage
x_2 Snow storage
x_3 Soil water
x_4 Ground water
x_5 Streamflow (daily)

Input Variables

z_1 Precipitation as rain
z_2 Precipitation as snow
$z_4 = b_9 z_1$ Thrufall, $(1-b_9)$ = Interception parameter
s_1 Potential Evapotranspiration (PET)
s_3 Average temperature (°F)

A. *Canopy Storage and Evapotranspiration (ET)*

$g_1 = (b_{10} - x_1)(1 - \exp(-b_{11} z_1))$ Input to x_1, where b_{10} = maximum storage capacity, $b_{11} = (1 - b_9)/b_{10}$.

$g_2 = \max\{[s_1 + 1] - [s_1 + 1]^{z_1/b_{12}}, 0\}$ Adjusted PET, see Figure 6.

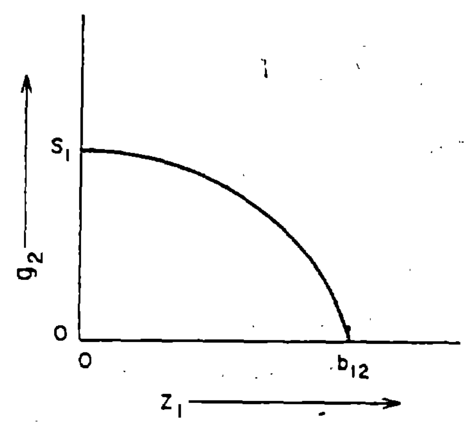

FIGURE 6 Adjusted atmospheric demand function for Model II.

$g_3 = \min\{x_1 + g_1, g_2(1 - \exp(-b_{13}(x_1 + g_1)))\}$ Canopy evaporation, where b_{13} = evaporation rate.

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g_4 = Transpiration, see Figure 7. b_1 = wilting point, b_2 = transpiration resistance point.

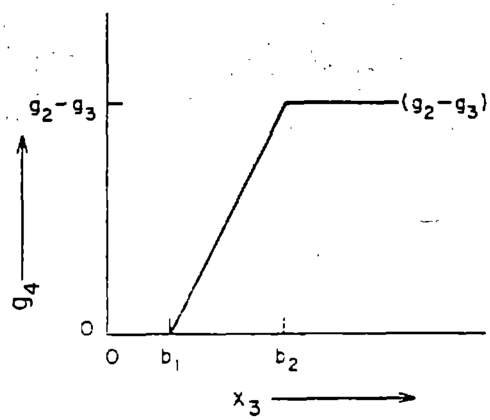

FIGURE 7 Actual transpiration function for Model II.

$g_5 = (1 - b_9)z_1 - g_1$ Drip to forest floor.

B. *Snow Melt*

$g_6 = \max\{0, (s_3 - 32)(254\,\mathrm{RAD} + 0.014(z_4 + g_5))\}$ Potential snow melt, where RAD is a monthly radiation index.

$g_7 = \min\{g_6, z_2 + x_2\}$ Actual snow melt.

C. *Infiltration*

$g_{12} = z_4 + g_5 + g_7$ Potential infiltration.

$g_{13} = \min\{b_7 - x_3, \ g_{12}\}$ Actual infiltration, where b_7 = maximum x_3.

$g_{14} = \max\{g_{12} - g_{13}, 0\}$ Surface runoff.

D. *Soil Water Flow*

$g_8 = \max\{\phi_1(x_3 - b_5), 0\}$ Total flow, where b_5 = retention capacity, b_3 = instantaneous flow rate, $\phi_1 = 1 - \exp(-b_3)$ = discrete flow rate.

$g_9 = \min\{b_8 - x_4, \ g_8\}$ Percolation, where b_8 = maximum x_4.

$g_{10} = g_8 - g_9$ Lateral flow.

E. *Ground Water Lateral Flow*

$g_{11} = \max\{\phi_2(x_4 - b_6), 0\}$ Total flow, where b_6 = retention capacity, b_4 = instantaneous flow rate, $\phi_2 = 1 - \exp(-b_4)$ = discrete flow rate.

F. *Stream Flow*

$g_{15} = g_{10} + g_{11} + g_{14}$ Stream flow.

Model II–b

Changed value for b_1 of function g_4 to see its effec on transpiration.

Model II–c

Readjusted parameter values and initial conditions, and used adjusted driving data for Watershed Also used this form with Watershed 10 driving dat This form may be directly compared with later models, while II–a and II–b may not

FLEXFORM ABSTRACT Hydrology Model II
Resolution, one day: units, m^3 Date: 5/4/7
(Figure 2–c, d)

Model III–a

State Variables

As in Model II, plus
x_{22} Lagged ground water

Input Variables

As in Model II, except
$z_4 = b_2 z_1$ Thrufall, where $(1 - b_2)$ = Interception parameter

A. *Canopy Storage and ET*

Same as Model II, except for units adjustment and notational changes.

B. *Snow Melt*

Same as Model II, except for units adjustment and notational changes.

C. *Infiltration*

$g_7 = z_4 + g_4 + g_6$ Potential infiltration, where g_4 = drip, g_6 = actual snow melt.

$g_{14} = \max\{0, (x_3 - b_{11}) + g_7 + (g_{11} - g_8) - g_{12} - g_{13}\}$ Surface runoff, where b_{11} = maximum x_3, g_{13} = transpiration; g_8, g_{11}, g_{12} see below.

$g_7 - g_{14}$ Actual infiltration.

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D. *Soil Water Flow*

$g_8 = \max\{0, \quad \phi_1((x_3 - b_9) + b_{15}g_7)\}$ Percolation, where b_9 = retention capacity, b_7 = instantaneous flow rate, $\phi_1 = 1 - \exp(-b_7)$ = discrete flow rate, b_{15} = "residence" parameter which is the proportion of g_7 which contributes to flow during the same day.

$$g_{12} = \begin{Bmatrix} g_{11} & , g_{11} \leqq g_8 \\ g_8 + \phi_1 b_{18}(g_{11} - g_8), & g_{11} > g_8 \end{Bmatrix}$$

Lateral flow, where b_{18} = "residence" parameter for $(g_{11} - g_8)$, ϕ_1 as in g_8.

No water moves laterally in soil until x_4 is filled.

E. *Ground Water Lateral Flow*

$g_9 = b_{17}(x_{22} + b_{16}g_8)$ Current lag effect, where b_{16} = "residence" parameter for g_8, b_{17} = lag factor.

$f_{22,22} = g_9 - x_{22}$ Implies that current g_9 becomes x_{22} for next time step.

$g_{10} = \max\{0, \quad \phi_2((x_4 - b_{10}) + b_{16}g_8 - g_9)\}$ Lateral flow, where b_{10} = retention capacity, b_8 = instantaneous flow rate, $\phi_2 = 1 - \exp(-b_8)$ = discrete flow rate.

$g_{11} = \max\{0, (x_4 - b_{12}) + g_8 - g_{10}\}$ Back flow of excess over storage capacity, where b_{12} = maximum x_4. Necessary device in stratified model since uphill strata ground water lateral flow can exceed storage capacity of downhill strata.

F. *Stream Flow*

$g_{15} = g_{10} + g_{12} + g_{14}$ Stream flow.

Model III–b

Incorporates a split ground water compartment, channelized (x_4) and non-channelized (x_{23}), imposed on Model III–a. Storage above retention capacity is split between the two with x_{23} receiving 30% ($b_{20} = 0.30$). Storage below retention capacity is assigned to x_4; assignment to x_{23} would not change behavior.

Percolation is split with x_{23} receiving 50% ($b_{19} = 0.50$). Excess over x_{23} storage capacity is sent to x_4 where it contributes to flow (or backflow) from there. Channelized lateral flow rate (ϕ_2) is left unchanged. Non-channelized lateral flow is considerably slower, but of the same form, and flows directly into the stream.

Model III–c

Identical to III–b except that a portion, $b_{22} = 0.03$, of calculated percolation passed directly to the stream. $b_{22} = 0$ in subsequent Model III forms, but was used again for Model V only.

Model III–d (with "dry sponge")

The retention capacity of x_3 was made a function of x_3, so that water entering a nearly dry compartment would pass through quickly.

$$g_{19} = \begin{Bmatrix} b_{14}, \; x_3 < b_{14} \\ x_3, \; b_{14} \leqq x_3 \leqq b_9 \\ b_9, \; x_3 > b_9 \end{Bmatrix}$$

Retention capacity of x_3, where b_9 = retention capacity when wet, b_{14} = transpiration resistance point.

$g_{16} = \max\{0, \phi_1((x_3 - g_{19}) + b_{15}g_7)\}$ Total percolation.

$g_8 = g_{16} - g_{17}$ Percolation into x_4, where g_{17} = percolation into x_{23}.

Model III–e

Model III–d with $z_2 = 0$, so that no snow occurs regardless of temperature.

FLEXFORM ABSTRACT Hydrology Model VII
Resolution, one day; Date: 12/20/73
units, m^3 Revised: 10/75
(Figure 2–h)

Model VII–a

State Variables

x_1	Foliage storage
x_2	Epiphyte storage
x_3	Snow pack storage
x_4	Potential infiltration
x_5	Channelized soil water
x_6	Non-channelized soil water
x_7	Channelized ground water
x_8	Non-channelized ground water
x_9	Stream flow (daily)
x_{29}	Lagged channelized ground water

Input Variables

z_1 Average temperature (°F)
z_2 Precipitation as rain
z_3 Precipitation as snow
z_5 Thrufall
z_6 PET

A. *Canopy Storage and ET*

$g_1 = (1-b_2)b_5/b_3b_4$ Rate of x_1 change, where $(1-b_2)$ = interception parameter. b_5 = proportion of interception taken by foliage. b_3 = maximum total canopy storage. b_4 = proportion of total in foliage storage.

$g_2 = (1-b_2)(1-b_5)/b_3(1-b_4)$ Rate of x_2 change.

$g_3 = (b_3b_4 - x_1)(1-\exp(-g_1z_2))$ Addition to x_1.

$g_4 = (b_3(1-b_4)-x_2)(1-\exp(-g_2z_2))$ Addition to x_2.

$g_5 = z_2 - z_5 - g_3 - g_4$ Canopy drip.

$g_6 = \max\{0, [z_6+1]-[z_6+1]^{z_2/b_6}\}$ Adjusted PET, where b_6 = rainfall needed to stop evaporation (graph similar to Figure 6).

$g_7 = \min\{x_1+g_3,\ g_6(1-\exp(-b_7(x_1+g_3)))\}$ Evaporation from x_1, where $b_7 = x_1$ evaporation rate.

$g_8 = \min\{b_8(g_6-g_7),\ x_2+g_4\}$ Evaporation from x_2, where b_8 = proportion of remaining ET demand satisfied by x_2.

g_{13} = Transpiration, see Figure 7, but with notational changes.

B. *Snow Melt*

Potential (g_9) and actual (g_{10}) snow melt as in Model III, with some notational changes.

C. *Infiltration*

$g_{11} = z_5 + g_5 + g_{10}$ Potential infiltration.

$g_{22} = \max\ \{0,\ x_5 - b_{13}(b_{17}-b_{15}) + (1-b_{21})g_{11} + g_{12} + g_{14} - g_{15} + g_{20} - g_{21}\}$ Surface runoff, where $b_{13}(b_{17}-b_{15})$ = amount of soil water storage above retention capacity which is channelized, $(1-b_{21})g_{11}$ = infiltration into x_5; g_{12}, g_{14}, g_{15}, g_{20}, g_{21} see below.

$g_{11} - g_{22}$ Actual infiltration.

D. and E. *Soil Water Flow and Ground Water Lateral Flow*

Percolation, lateral flow, and back flow (of excess over storage capacity) are all modelled as established in III–a; percolation to channelized ground water is lagged with regard to lateral flow. The order of flow operation is (see Figure 4):

g_{12} Lateral flow from x_6 to x_5;
g_{14} Surface flow from x_6 to x_5;
g_{15} Percolation from x_5;
g_{16} Percolation from x_5 to x_8 (remainder to x_7);
g_{17} Lateral flow from x_8 to x_7;
g_{18} Amount of water currently withheld from contributing to x_7 lateral flow;
g_{19} Lateral flow from x_7 to x_9;
g_{20} Back flow from x_7 to x_5;
g_{21} Lateral flow from x_5 to x_9.

F. *Stream Flow*

$g_{23} = g_{19} + g_{21} + g_{22}$ Stream flow.

Model VII–b

E. *Ground Water Lateral Flow*

The lag function was redefined so that it was as first conceptualized. This removed the multiplication by b_{28}, the "residence" parameter for percolation into x_7, during the updating of x_{29}.

For biographies and photographs of the authors, please see p. 137 of this issue.

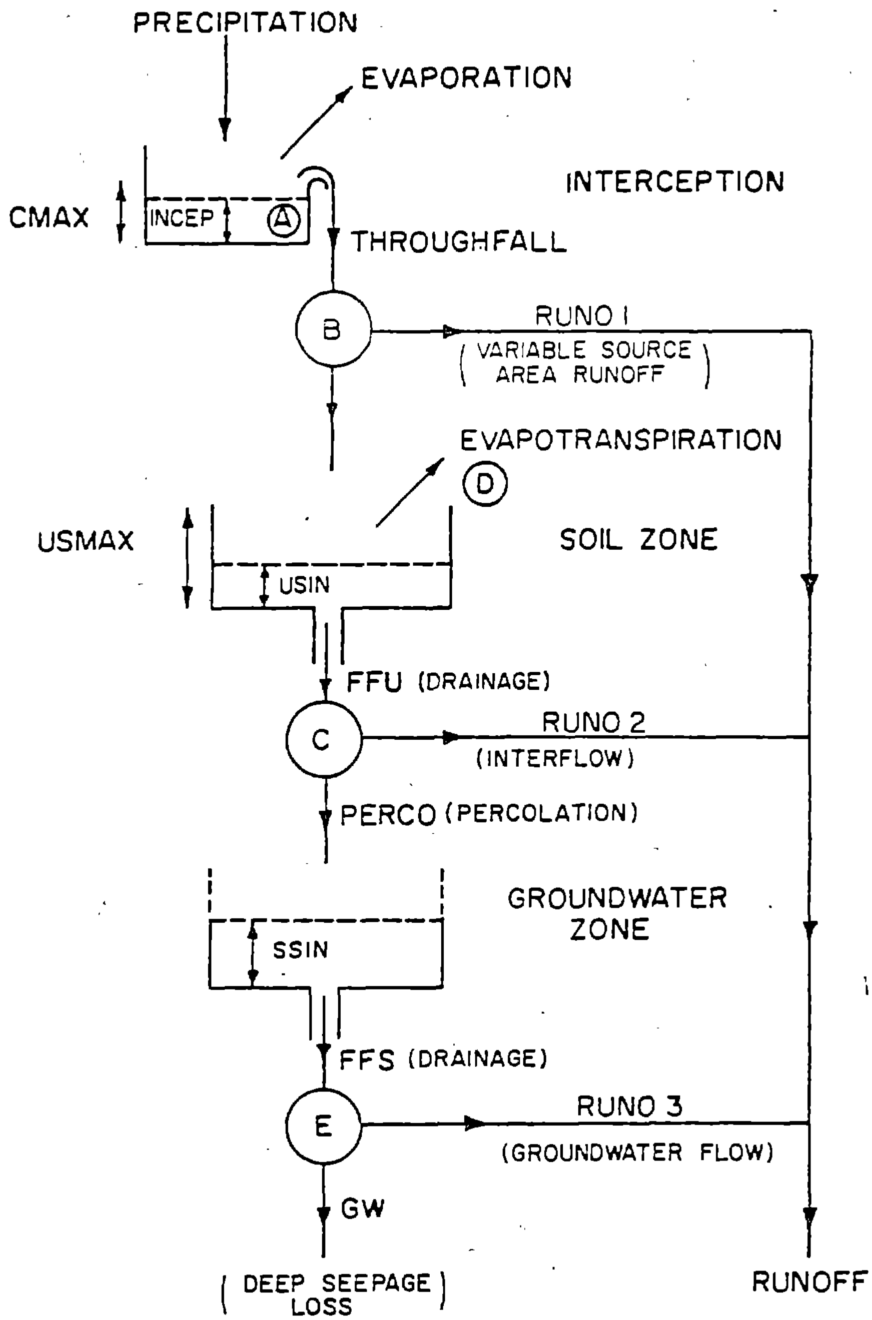


Figure 4.1 Schematic Flow Diagram of the Daily Watershed Model.

Table 4.1 Watershed Model Function Descriptions

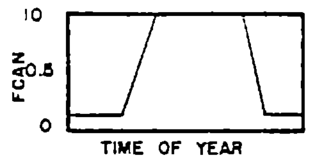
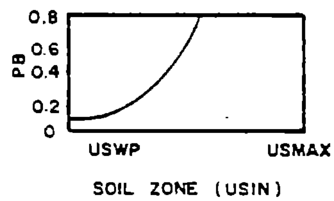
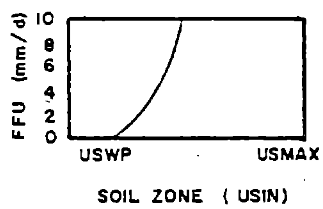
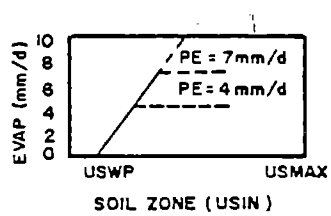
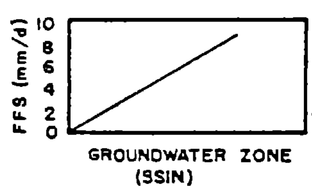
FUNCTION	EQUATION	FUNCTION SCHEMATIC	PROCESS
A	$CMAX = CEPMAX \ast FCAN$		INTERCEPTION
B	$RUNO1 = PB \ast PRECIP$ $INFIL = (1-PB) \ast PRECIP$ $PB = FSTP + PC\,e^{PAC \ast (USIN/USMAX)}$		VARIABLE SOURCE AREA RUNOFF
C	$RUNO2 = KI \ast FFU$ $PERCO = (1-KI) \ast FFU$ $FFU = FU \ast \left(\frac{USIN}{USMAX}\right)^{KU}$		SOIL ZONE DRAINAGE & INTERFLOW
D	$AEVAP = EVAP \; (EVAP < PE)$ $= PE \; (EVAP > PE)$ $EVAP = \frac{(USIN - USWP)}{ERATE}$		SOIL ZONE EVAPOTRANSPIRATI(
E	$RUNO3 = K2 \ast FFS$ $GW = (1-K2) \ast FFS$ $FFS = FS \ast (SSIN)^{KS}$		GROUNDWATER ZONE SEEPAGE & GROUNDWATEP FLOW

Table 4.2 Model Parameter Descriptions and Values

Process/Zone	Parameter	Definition	Parameter Value (Little Millseat Watershed)
Interception	CEPMAX	Maximum interception capacity (mm)	2.02
	FCAN	Canopy development function: modifies CEPMAX for time of year (i.e. canopy development)	See Table 4.1
Variable Source Area Runoff	FSTR	Fraction of watershed always contributing to direct runoff (i.e. area of stream)	0.05 (0.05)
	PAC	Source area exponent	39.295 (40)**
	PC	Source area coefficient	4.11×10^{-6} (4.1×10^{-6})**
Soil Zone	USMAX	Soil zone thickness (mm)	1087 (1070)
	KU	Soil water conductivity exponent (KU=2b+3, where -b is the slope of a log-log plot of the soil water retention curve)	11.810 (11.467)
	FU	Soil water conductivity coefficient	1.49×10^{7}
	K1	Fraction of Soil Zone drainage becoming interflow	1.0 (1.0)
Evapotranspiration	USWP	Wilting point water content (input as % by volume, used as mm of water in program)	124 (130) 11.44% (12.14%)
	ERATE	Evapotranspiration rate coefficient	27.4
Groundwater Zone	FS	Groundwater exponent (1 for linear groundwater recession)	-*
	KS	Groundwater recession constant	-*
	K2	Fraction of groundwater drainage becoming baseflow	-*
OTHER VARIABLES			
	CMAX	Actual interception capacity (mm)	
	USIN	Actual soil water volume (mm)	
	SSIN	Actual groundwater volume (mm)	
	PB	Fraction of water contributing to direct runoff	

* Groundwater Zone does not exist in the Little Millseat watershed.

Values in parentheses are the initial parameter estimates prior to optimization

** Values used in BROOK model (Federer and Lash, 1978) for Hubbard Brook Watershed