University of Washington
Department of Civil and Environmental Engineering

FORECASTING SEASONAL SNOWMELT RUNOFF: A SUMMARY OF EXPERIENCE WITH TWO MODELS APPLIED TO THREE CASCADE MOUNTAIN, WASHINGTON DRAINAGES

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Water Resources Series
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by

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Project Completion Report: "Improving Reservoir Yield Through Forecasting Intraseasonal Snowmelt Runoff"

Project Period: October 1, 1976 to September 30, 1978

Principle Investigators: Stephen J. Burges, Associate Professor of Civil Engineering, University of Washington

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ABSTRACT

Two alternate approaches to forecasting seasonal runoff volumes from watersheds with substantial snow storage were investigated. In the first, a continuous simulation model developed by the National Weather Service River Forecast Center in Sacramento, California was applied to the Cedar River, Washington. Although the model has proved quite successful in forecasting runoff from several California Sierra Nevada streams, the model could not be successfully calibrated for use in forecasting Cedar River runoff. Problems were traced to the snowpack accumulation and ablation module of the model; adequate performance of the land module was observed as evidenced by successful simulation of runoff events during seasons of the year with minimum snow water storage. The principal difficulties with the snowpack module appeared to be (1) use of the saturated adiabatic lapse rate for temperature computation regardless of weather conditions, (2) an unrealistically low partition temperature for discrimination between precipitation falling as rain and snow, (3) the method of computation of cloud cover for modification of solar radiation reaching the snowpack, (4) unrealistically low wind movement required to calibrate the condensation melt portion of the snowpack module, and (5) possible underestimation of the free water retention in the snowpack. Some or all of these problems may be related to differences in the climatic regimes of the watersheds for which the model was developed and subsequently successfully applied, and the test watershed.

The second approach employed was a modification of Tangborn's (1977) basin storage accounting approach, in which parameters are estimated by regression. The modifications made to the model allowed several options for incorporating snow course measurements in the forecasts. The model was applied to three Washington streams: the Cedar, also used in testing of the continuous simulation

model, and the American and Stehekin Rivers. Inclusion of the snow course data substantially improved the accuracy of the Cedar River forecasts, substantially decreased the accuracy of the Stehekin River forecasts, and made little difference to the American River forecasts. The varying effect of inclusion of the snow course data appears to be related to the location and topography of the basins as well as to the location of the snow courses. The worth of runoff forecast period precipitation forecasts was also estimated; such forecasts had little value beyond about March 15 for all the basins and had negative value to the Stehekin River forecasts. Increased precipitation forecast worth was observed for early winter runoff forecasts.

CHAPTER 1 INTRODUCTION

Snowmelt runoff is one of the few natural phenomena for which relatively accurate long term forecasts can be made. Such forecasts have a long history. Peck (1972), for example, discusses snow cover measurements taken by the Chinese in the twelfth century A.D.. One can imagine that these measurements could have been used in subjective estimates of spring runoff. In this country, snow survey observations initiated in the early 1900's were first used to forecast the spring rise of Lake Tahoe in 1911 (Elliot, 1977). In 1935, the Bureau of Agricultural Engineering (the predecessor of the current Soil Conservation Service) took over the snow survey measurements and flow forecasting responsibilities in the eleven western states.

The techniques by which the early forecasts were made may be classified as runoff index methods; historic sequences of recorded runoff and index variable sequences (e.g., snow course measurements, soil moisture, etc.) were used to develop forecast equations. The early approaches generally utilized graphical techniques (see, for example, Linsley, et.al., 1958). Later approaches utilized multiple regression analysis, and additional refinements such as principal components (Marsden and Davis, 1968) and pattern search (Zuzel, et.al., 1975) have more recently been attempted.

Evolution of forecasting techniques has been spurred by the substantial economic worth of streamflow forecasts. Elliot (1977) has used a linear programming model in conjunction with agricultural water use to estimate the worth of summer runoff forecasts for basins in the eleven western states. The estimated value of forecast accuracy is linear with accuracy improvement, and although this somewhat oversimplifies the real situation, these results do provide an adequate first approximation of forecast worth to irrigated agriculture. The results of the Elliot study are summarized in Figure 1, where "current" forecast accuracy is that achieved by the multiple regression approach presently used by the SCS. The value of forecast accuracy for Washington is seen to be nearly the same as the westwide average (this estimate would, of course, be applicable primarily to the semiarid eastern part of the state). The incremental value of perfect forecast accuracy is about $2.50/acre. Of course, it is unlikely that perfect forecasts will ever be developed; model and precipitation forecast accuracy limitations discussed in this report suggest that a practical target for future accuracy improvements may be on the order of 50% of the current error levels. Elliot (1977) for instance, speculates that installation of telemetered snow course sensors could reduce mean absolute error to approximately 70-80% of the current level.

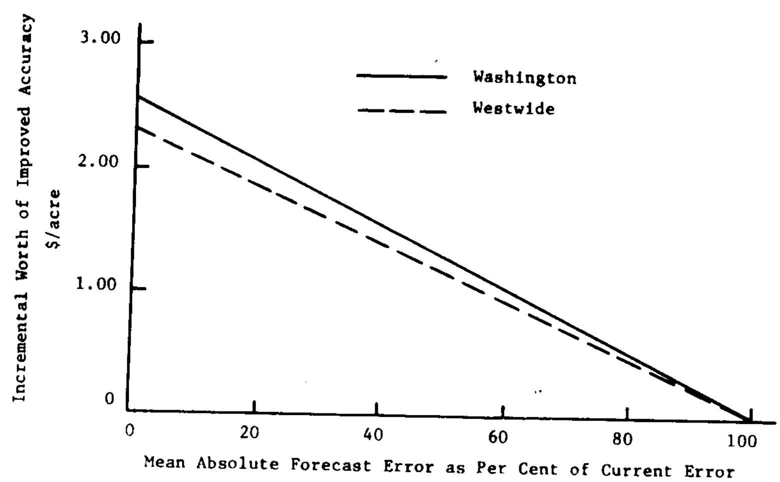

Figure 1. Estimated Worth of Summer Runoff Forecasts as a Function of Error Reduction (Data from Elliot, 1978).

Forecast improvements are of economic value not only to irrigated agriculture, but also to nonagricultural users, e.g., municipal and industrial, especially in the Southwest. The benefits here, although presently unquantified, may be even larger than those accruing to agricultural users, since the marginal value of water to nonagricultural users is often substantially higher (sometimes as much as an order of magnitude (Barr and Pingry, 1976)).

In any event, if a lower limit for potential incremental benefits of $1/acre is assumed, when applied to the approximately 20 million irrigated acres affected by snowmelt runoff in the western United States, the investment in research studies such as that reported here is not difficult to justify.

This report summarizes work performed as part of the second phase of an Office of Water Research and Technology (U.S. Department of the Interior) funded project, "Improving reservoir operation through forecasting intra-seasonal snowmelt runoff". The first phase of this work addressed the problem of determining the worth of forecasts in reservoir operation. This work is summarized in an accompanying report (Burges and Hoshi, 1978).

Initially, work in the second phase of the project was to have been directed towards exploration of deterministic watershed models couched in a state estimation framework for use in forecasting runoff volumes. This formulation was to be subsequently used to segregate sources of forecast errors according to forecast model errors and date input errors. As work in phase II progressed, it became apparent that, although state estimation techniques showed promise, this approach did not provide the best framework for forecast model assessment. One consideration which prompted a review of the approach was the development by Tangborn (1977) of a simple basin storage approach to forecasting snowmelt runoff. The Tangborn model, which appears to be more accurate in many applications than the best flow index approaches, served to provide a benchmark against which the performance of a more sophisticated conceptual model could be judged. Although the parameters of the Tangborn model are estimated by regression, the model differs from the simple flow index methods in that a physically realistic seasonal basin water storage balance is preserved.

The Tangborn model makes use of a test season forecast correction which is, in a sense, analogous to the measurement correction used in state estimation theory. The simplicity of the Tangborn model is also helpful in providing a basis on which to examine a further objective, which is to determine data requirements and model complexity as a function of forecast reliability. Consequently, the second phase of the project was approached as a comparison of performance of a modification of the Tangborn model (described in Chapter 4) and a conceptual simulation model.

The simulation model developed by Burnash and Baird (1975) and Burnash et.al. (1973) of the Sacramento River Forecast Center National Weather Service has proven quite successful in forecasting seasonal runoff for several California basins. This model was taken as typical of the current state of the art of conceptual simulation models. Both models were applied to runoff forecasts for the Cedar River, Washington as measured at USGS gage 12-1150 near Cedar Falls. The Tangborn model was also applied to the Stehekin (USGS gage 12-4510 at Stehekin) and the American River (USGS gage 12-4885 near Nile).

The remainder of this report describes experience with implementation of the water storage balance (modified Tangborn) and the Burnash/Baird (Sacramento) conceptual simulation models. Chapter 2 is devoted to a description of the Sacramento model. Chapter 3 describes experience with the attempted implementation of the Sacramento model to forecasting of the Cedar River runoff. Although it had been intended to generate actual forecasts using the Sacramento model, extensive difficulties with model calibration lead to the ultimate conclusion that the snow accumulation/ablation portion (module) of the model is not transferrable to the Cedar; some of the process descriptions which apparently perform well in simulation of Sierra Nevada snowpacks break down when applied to the lower elevation, western Cascade basin. In retrospect, the authors feel that the Cedar is perhaps one of the more difficult basins to model using the conceptual approach owing to the high sensitivity of areal snow accumulation and ablation to relatively small temperature changes and numerous rain-on-snow events which occur even at the highest elevations in the watershed.

Chapter 4 describes development of the storage accounting model, which is a modification of Tangborn's forecasting model to allow incorporation of snow course data. Chapter 5 summarizes results of implementation of three variations of the model: one is essentially identical to Tangborn's formulation and two use alternate methods of incorporating snow course data. Comparison of the three formulations is used to assess the worth of snow course data to this type of model. In addition, bounds on the value of summer season forecast period precipitation forecast accuracy were generated by incorporating fictitious "forecasts" using the recorded summer season precipitation.

Chapter 6 summarizes the experience with the storage accounting and conceptual simulation models. Advantages and limitations of the models are discussed, along with suggestions for future development effort.

CHAPTER 2 SACRAMENTO MODEL DESCRIPTION

Conceptual (continuous) simulation models are based on a description of the significant processes and interactions of the system being modeled. In a watershed model these processes are primarily physical: precipitation, percolation, interflow, etc. Such models attempt to combine physical theory, e.g., energy, momentum, and mass conservation with the available data base to obtain a description of system phenomena that can be compared with observations of some of the phenomena. Agreement between the simulated and observed records is taken as one indication of model validity.

Several workers have developed simulation models of watersheds with significant snowfall events (Anderson, 1973; U.S. Army Corps of Engineers, 1975; Burnash, et.al., 1973; Leavesley and Strittler, 1978). In addition to the inherent scientific desirability of the continuous simulation approach, the models theoretically have the ability to describe the magnitude and timing of a hydrograph. This is a potential advantage over statistical models, which can generally only be used in forecasting seasonal runoff (the Tangborn model discussed in Chapter 1 is an exception to this generalization). This allows the potential for short term forecasting of extreme events (e.g., spring snowmelt floods) based on weather forecasts as well as long term aggregate forecasts.

In snow-affected watersheds, the snowpack acts as a reservoir and lags the input of precipitation to the watershed. Thus a continuous simulation model that uses precipitation as the driving independent variable must account for the snow accumulation and ablation process, as well as the rainfall-runoff interaction in order to accurately simulate basin runoff hydrographs.

Simulation models provide the potential for hydrograph reproduction and thus the potential means to achieve more accurate forecasts of snowmelt and runoff events. In Chapter 3, a continuous simulation model is evaluated for potential use as a forecasting tool. Several issues need to be addressed in application of such models to forecasting and streamflow forecasting. Among them are data requirements, generality and transferability, and criteria for selection and use of a particular model. These issues provide the basis for this examination, which is addressed through application of a particular continuous simulation model developed by Burnash, et.al. (1973). A general discussion of these issues is contained in Section 3. However, a brief description of the mechanisms included in snowmelt simulation is in order before proceeding to these more general concerns.

2.1 Basic Concepts

Deterministic simulation of many physical processes can be accomplished by numerical solution of differential equations written to describe mass, energy, or momentum balances. An example is the exponential growth or decay function:

$$\frac{dX}{dt} = -kX \tag{2-1}$$

which has the closed form solution

$$X = X_o e^{-kt} \quad , \qquad -\infty < k < \infty \tag{2-2}$$

and can be solved iteratively by

$$\Delta X = kX_t \Delta t$$

$$X_{t+\Delta t} = X_t + \Delta X \tag{2-3}$$

The simulation model used in this study employs several coupled differential equations of the form:

$$\frac{\partial x}{\partial t} = f(a,b,j,k,p,q,x,y,t) \tag{2-4}$$

where a and b are calibration parameters used to scale the magnitude of the simulated processes; j and k are constants governing known physical relationships, i.e., the latent heat of fusion, etc.; p and q are environmental variables such as temperature, precipitation, etc.; and x and y are simulated variables such as snow pack water content, stream flow, etc. The value of x at any time t depends on the values of the other parameters and variables at time t. In some cases, however, when Δt is small the evaluation may be carried out at time $t - \Delta t$, allowing explicit solution of the difference equation. Description of a complex system requires several equations with multiple interacting variables. Solution of such systems of coupled differential equations is usually practically feasible only through use of numerical techniques.

2.1.1 Data Requirements

Data requirements comprise an important aspect of simulation models. Generally, data requirements to support an additional increment of model precision increase much faster than the additional model precision itself.

When modeling a physical phenomenon, theory may dictate collection of a large data set for accurate description. Resource and logistic constraints may, however, make it impossible to collect the desired information in any application other than in a pure research mode. Usually, historical data collection efforts are far from complete.

The combined effect of extensive data requirements for high resolution models, coupled with the likely inadequacy of historical data, is to limit the degree of resolution that can be practically included in an operational model. The model described in Chapter 3 attempts to maximize model resolution in light of a limited data base.

As sufficient data are seldom available at remote sites to measure all variables affecting a phenomenon, most models contain some assumptions or some approximate methods of deriving the necessary data. An excellent example of this problem as applied to snow pack energy balance and one approach to resolving it is contained in Anderson (1968).

Use of commonly available data imposes two additional conditions on a model. First, only precipitation, temperature, snow water content and streamflow records are usually available; consequently, the model must approximate or generate all other variables required to simulate the phenomenon of interest. This amounts to simulating what would be independent variables in a controlled or completely measured situation.

The data site is often different from the point of simulation. For instance, the commonly available records noted above are usually maintained at stations in or near inhabited areas, often in low-lying valleys. Thus, these recorded data represent conditions at locations which are displaced from the desired prediction point (e.g., for snowpack accumulation or precipitation volumes) in both distance and altitude. The model must be able to accept such displaced input data and adjust it to approximate the conditions at the point of simulation.

The foregoing discussion suggests five potential sources of model error:

1. Input data error,
2. Error in vertical or spatial adjustments of the data,
3. Error in estimation of unmeasured variable values,
4. Error in areal representation of predicted processes using the point model, and
5. Error attributable to incorrect assumptions or approximations in the theory on which the model is based.

Of these, only missing data and gross measurement errors are readily amenable to external corrections.

2.1.2 Model Transferability

The transferability issue was briefly touched upon above in regard to data requirements. A more fundamental aspect of model transferability deals with the generality of a model's structure.

Typically models are developed and first applied in data-rich situations, such as an experimental watershed. The danger always exists that some of the approximations and assumptions used may be unique to the development site. Conceptually, one may expect a model based on sound first principles to be easily transferable, since assumptions and approximations are less likely to be site-specific than in a "black box" approach. However, the assumptions required to describe the phenomena being modeled in a data-poor situation introduce additional potential for site-specific approximations, even when the physical phenomena are adequately represented. Of particular concern in this regard is the sensitivity of the phenomena to dominant processes. For example, if a phenomenon is governed by processes A, B and C, a model of the phenomenon may be developed at a location where process B consistently dominates. A model developed under such conditions may perform well if a good representation of process B is achieved, regardless of whether less reliable approximations are made in modeling either process A or C. When the model is applied at a different location, where there is a more even balance between A, B and C, however, the model may perform unsatisfactorily due to inaccurate representation of processes A and C. This hypothesis may only be tested by attempting to apply the model at a number of diverse locations.

2.1.3 Model Selection

James and Burges (1978) describe four watershed simulation model selection criteria:

1. The model must provide the kind of information needed,
2. The watershed characteristics represented by the model parameters

must in fact govern watershed response in the intended application,

3. The equations used must be correct in light of the state of the art, available data, and available computer facilities, and
4. The model must provide results which are suitable for the intended use and are of acceptable quality at a reasonable cost within the required time frame.

These criteria are listed in ascending order of assessment difficulty. Given reasonable model documentation, a competent hydrologist should be able to determine if a model meets the first three criteria in a straightforward manner. The suitability of the results may also be ascertained from documentation and sample output. However, the cost of implementing a model in terms of both resources and time is a difficult matter to judge in the absence of prior experience. The questions of model implementation as specifically related to required model complexity and model transferability are the focus of the case study of a particular watershed simulation model reported in Chapter 3.

2.2 Model Description

The model used in the case study was developed by the National Weather Service River Forecast Center Sacramento office (Burnash, et.al., 1973; Burnash and Baird, 1975). The rainfall-runoff portion (land module) of this model is also known as the Generalized Hydrological Model (GHM). A separate simulation of snowpack accumulation and ablation (snow module) is required to provide the appropriate timing of moisture delivery to the land module. This is accomplished by constructing a "pseudo-precipitation" record composed of snowmelt or, in the absence of a snowpack, direct precipitation. The snowmelt algorithm developed by Burnash and Baird (1975) is based on a snow ablation calculation procedure described by Winston (1965). The model was selected for this evaluation because it is applicable to a data-sparse situation (only daily precipitation and temperature and monthly evapotranspiration are required to drive the model) and because it appeared to be generally representative of the state of the art of snowmelt runoff simulation at the time the present project was initiated. This is not meant to imply any judgement of superiority for the Sacramento model; several other state of the art conceptual models exist. The choice was, rather, made on the basis of availability of the model and past evidence of its successful use in an operational forecast mode.

The model was applied to the Cedar River Basin, Washington above Chester Morse Lake. This basin was selected because snow melt contributes a significant amount of seasonal runoff, because the required weather and stream flow data were available in the proximity of the drainage and because the basin serves as a source of water supply for the City of Seattle, Washington, and therefore is of direct interest to the City of Seattle Water Department which provided partial project funding.

The basin is located approximately 50 miles southeast of Seattle, on the west slope of the Cascade Mountains (see Figure 7, Chapter 4). Altitudes range from 1500 feet above mean sea level (MSL) to more than 4500 feet MSL. The basin has an area of approximately 41 square miles distributed according to the area-altitude relationship shown in Figure 2.

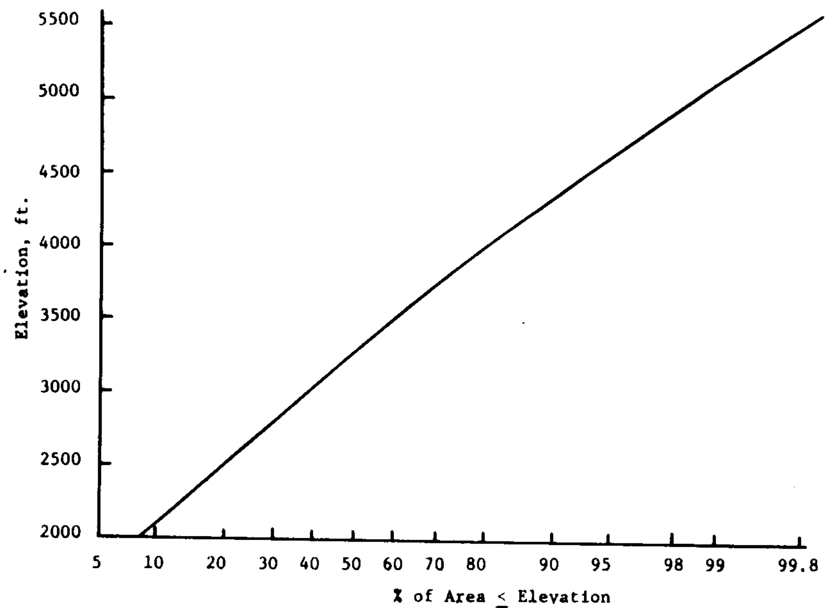

Figure 2. Hypsometric Profile of Cedar River Basin Above USGS Gaging Station No. 12-1150 Near Cedar Falls.

The basin lies somewhat west of the main body of the Cascade Mountains. As a result, it may be somewhat more subject to variations in temperature and precipitation resulting from maritime frontal storm patterns than some other watersheds in the Cascade Mountain range. Annual average basin precipitation is 120 inches and the basin experiences substantial snow cover in most years. The areal extent and amount of snow cover vary widely during the winter. Lower portions of the basin may be alternately covered with snow and bare several times during a typical winter.

2.2.1 Background

The streamflow simulation system described below consists of numerical solutions of several coupled differential equations designed to obtain maximum model precision from a limited data base. All analyses and simulation procedures are based on daily data sets which can normally be obtained from existing weather and streamflow records.

In watersheds where snowmelt contributes a significant portion of annual runoff it is necessary to include the snow accumulation and melt process in the overall modeling effort in order to accurately describe streamflow hydrographs. The snowpack simulation module used in this study was developed by Burnash and Baird (1975) based on a snow ablation calculation method described by Winston (1965). Its purpose is to describe the timing and magnitude of delivery of melt water to the hydrological regime for transport to the stream.

The land phase simulation algorithm used in this model is the General Hydrologic Model (GHM) developed by Burnash, et.al. (1973). It represents an attempt to include more realism in rainfall-runoff modeling than earlier models, such as the Stanford Watershed Model (SWM) developed by Linsley and Crawford (1966). In the SWM algorithm the partition of precipitation among surface, interflow and base flow runoff is based on a rather empirical calculation using a geometric approximation of watershed conditions. The partition and percolation of incoming precipitation in the GHM, however, is based on recent research into the nature of groundwater flow and appropriate methods of modeling it (Burnash et.al.,1973; Hauks, et.al., 1969; Green et. al., 1970).

Model implementation is initiated by calibrating the two modules to available records and verifying the calibration. The snowpack module is calibrated to snow course records for a site near or within the watershed. The rainfall-runoff module is calibrated to a streamflow gage record at the mouth of the basin. Verification is achieved by comparison of the errors obtained in the calibration process with the errors obtained using the calibrated parameter values to simulate streamflows over a different time period. The model is verified if several summary statistics of calibration and verification error indicate compatibility of the errors.

2.2.2 Snow Module

Winston (1965) developed a modified evergy balance model of a snowpack using hourly weather observations as the input data base. His procedure accounted for melt due to radiation, convection, and condensation; and additional changes in snowpack water content due to sublimation and rainfall. Burnash and Baird (1975) developed the daily snowpack algorithm described below as an adaptation of Winston's procedure. This snowpack model allows simulation of snow accumulation and ablation at a single point.

The algorithm follows the physical path of water through the snowpack in the order 1) precipitation, 2) pack accumulation, 3) melt and 4) delivery to the watershed. Daily precipitation and temperature records from stations near the watershed are used to determine the amount and form of precipitation. The observed records are weighted with either Thiessen or empirical weights to obtain a representative basin mean precipitation. The temperatures are lapsed using a saturated lapse rate of 3°F per 1000 feet to the elevation of the point simulation. The partition between rain and snow is made on the basis of the daily average temperature and consists of a sliding scale with 100% snow at 27°F and 100% rain above 32°F.

Radiation is calculated from a linear approximation of an annual sinusoidal function presented by Winston and is adjusted for cloud cover, elevation and latitude. The cloud cover adjustment is a function of the daily temperature spread using a relationship developed for central California radiation melt, and is calculated using:

RMELT = RADAT*RATIO*(1-Albedo)*(CST+PACK*ABSCF) 2-5 where: RADAT = solar radiation adjusted for day, elevation, latitude

RATIO = the fraction of solar radiation penetrating the cloud cover

Albedo = snow surface albedo calculated using an adaptation of Winston's empirical procedure

CST = a calibrated coefficient that converts radiation (Langleys) to melt (inches of water)

PACK = snowpack water content

ABSCF = a factor to consider the change in site shading with increased water content, i.e., deep snowpacks experience less shading from low lying ground cover.

The effects of condensation and convective heat transfer are accounted for in the convection-condensation melt term:

CCMLT = ((TD-32)*.0104+(TBAR-TD)*.00305*WS*ELTRM 2-6

where: TD = estimated dewpoint temperature from TD = T_{min} - 4°F

TBAR = daily mean temperature

WS = wind factor, daily nautical miles of movement at the pack surface (a calibration parameter)

ELTRM = elevation based air density adjustment.

The constants in this equation represent assumptions used to derive average daily conditions from the unit conversion in the hourly equations developed by Winston. In the model, convection-condensation melt can only occur if the daily mean temperature is above 32°F.

The last melt parameter included in the algorithm is melt due to rainfall, which utilizes Winston's formulation:

RFMLT = .007*R*(TD-32) 2-7

where: R = rainfall (inches).

As melt occurs and the pack ages the effects of density changes are accounted for in a compaction procedure. New snow is assumed to have a density of 0.25 inches of water per inch of depth. As the pack ripens, it approaches a density of 0.49. The compaction algorithm is included in Waddle (1978).

The sum of the water derived from the melt terms and precipitation reaching the ground surface are combined into a record of total daily liquid input to the soil surface. This pseudo-precipitation record is the primary output of the snow module. The pseudo-preciiptation record is taken as input to the watershed module and is discussed in section 2.2.3.

The validity of a model is dependent on assumptions and approximations used in constructing it. The following assumptions are included in the Burnash/Baird model:

1. The snowmelt model does not include estimations of sublimation, heat transfer within the snowpack (it is assumed to be isothermal), or heat transfer at the ground surface interface. Since these factors are not explicitly included, they are assumed to be accounted for in the values of the calibration parameters.
2. The model simulates the snowpack at a point and does not consider the changes in area covered as the snowpack is depleted. Thus, the pseudo-precipitation record is a point-generated record that must be averaged over the entire basin.
3. The lapse rate is taken as constant over all conditions and the TCR adjustment is assumed to best fit this fixed rate.

The snowpack model is coupled to the watershed model (GHM) by the pseudo-precipitation records. In the GHM these records are treated as independent input records since the model accepts recorded precipitation observations when applied to watersheds where snowmelt is not a factor in basin runoff.

Since the coupling of the models is through this synthesized data set, modifications to either of the models can be made independently of the other.

2.2.3 Land Module

The version of the Generalized Hydrologic Model used in this study represents level eight of the Generalized Streamflow Simulation System (Burnash, et.al., 1973). This model has been developed since the middle 1960's and has been adopted by the U.S. Weather Service as part of the River Forecast System standard rainfall-runoff model (Ferral, 1978; Anderson, 1978). The NWS River Forecast System does not, however, use the Barnash/Baird snow module, but rather incorporates a snow melt algorithm developed by Anderson (1973).

The processes incorporated in the watershed algorithm are described below. The model is a single point simulation of the mass balance of precipitation moisture in the watershed. The model accounts for losses due to evapotranspiration, unmeasured flow past the gage site, and importation and exportation of water from the basin if applicable. The basic watershed algorithm is summarized in Figure 3.

Direct runoff occurs from basin impervious area, which is a calibration parameter. Impervious area includes water bodies and a variable source area term to include saturation of stream banks, marsh areas, springs, etc. The remaining area of the watershed is considered to be permeable.

Evapotranspiration is supplied by two tension water zones. Tension water is considered to be that water that is bound to soil particles and remains in the soil after saturated conditions subside. Upper zone tension water (UZTW) represents the topsoil moisture deficiency that develops during dry periods. It must be filled before any water is available to percolate or produce lateral flow to a stream. Lower zone tension water (LZTW) represents

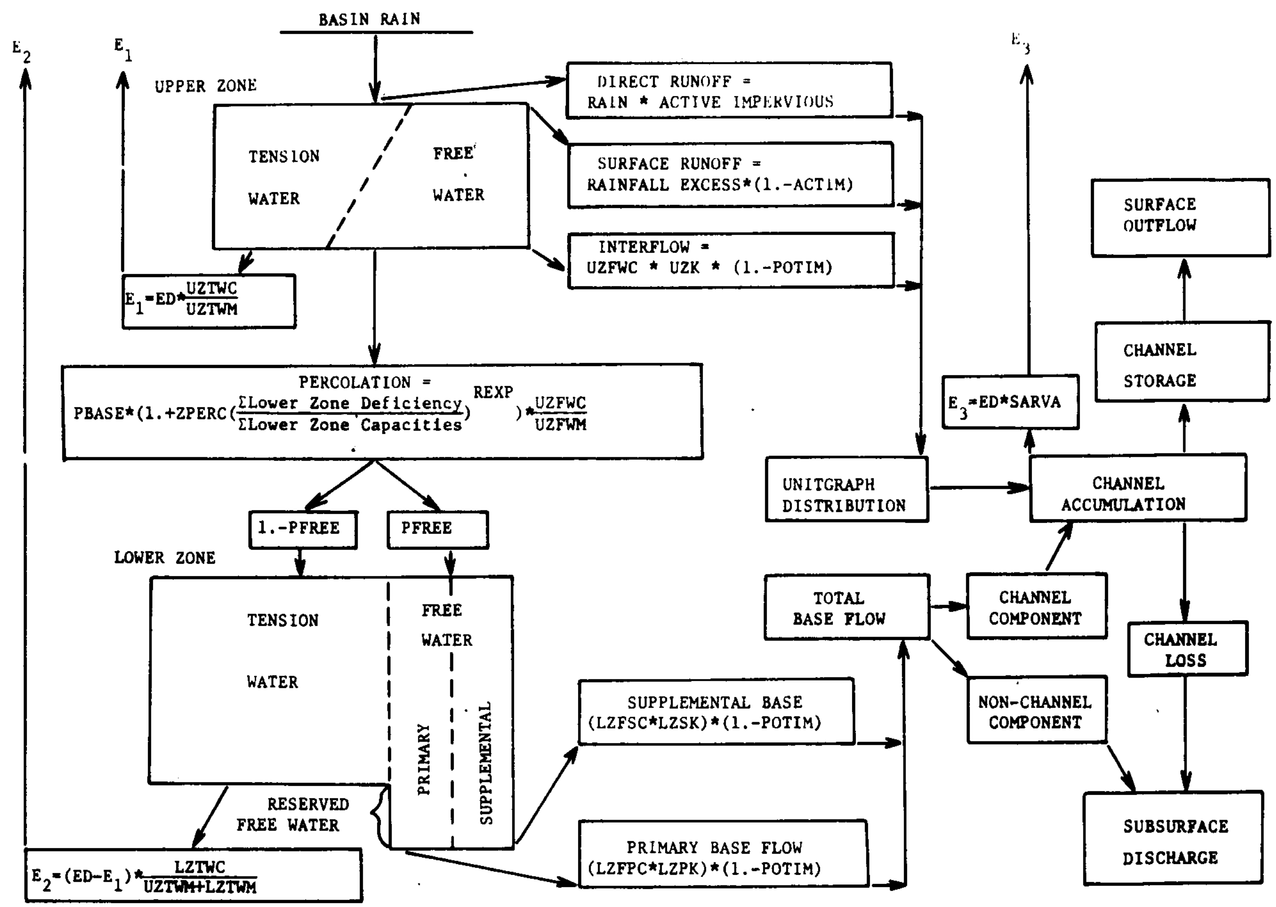

Figure 3. Schematization of Generalized Hydrologic Model (from Burnash, et.al., 1973).

the same particle-bound portion of water in the groundwater aquifers. Both tension water zones are considered accessable to plant root systems, hence they serve to supply evapotranspiration.

Interflow is supplied by a linear reservoir (e.g., outflow is a linear function of storage) called upper zone free water (UZFW). Water enters UZFW from UZTW after the tension water reservoir is filled. Interflow is consequently a first order exponential decay based on a calibrated interflow recession coefficient UZK.

The upper free water zone also supplies water to percolate to the lower zones. The percolation relationship is an exponential driving function that employs recently developed models of the percolation process (Hanks, et.al., 1969; Green, et.al., 1970) to relate the water deficiency of the upper and lower zones to the percolation rate.

Base flow is supplied by two linear reservoirs. The different recession coefficients of the two reservoirs allow additive combinations that better fit the non-linear characteristics of recession hydrographs.

To achieve final fitting of both models, a pattern search optimization technique is applied wherein parameters are sequentially perturbed so as to minimize the mean square of differences between simulated snowpack and observed snow water equivalent measurements. Use of the optimization routine requires preliminary adjustment of the model parameters to achieve the best "hands on" fit. The optimization is then allowed to fine tune parameter values to minimize the error criterion. The optimization routines are taken directly from Monro (1971).

2.2.4 Procedure for Model Use

Use of the snowpack module proceeds as follows. The available precipitation and temperature records are input as the driving independent variables.

The snowmelt module is calibrated against observed snow course measurements over a selected portion of the historical record used, in this case the ten years from 1952 to 1961.

There are ten calibration parameters of which six are precipitation station weights. The others are the radiation melt coefficient - CST, the snow depth shading modifier - ABSCF, an adjustment to air temperature at the pack location in addition to the lapse rate adjustment - TCR, and the wind speed factor - WS.

When the best fit to the observed snowpack records has been obtained, the module is used to generate up to five pseudo-precipitation records for input to the watershed module. These records are generated for locations roughly representative of equal divisions of the watershed's altitude range, hence they represent horizontal "slices" of the watershed. Since watersheds often do not have snow courses at several altitudes there is usually no opportunity to recalibrate the module for each altitude division. As a result, the calibrated parameter values remain fixed in generating the five pseudo-precipitation records; only adjustments in altitude and latitude are made. This procedure contains the inherent assumption that the calibration to one snow course is representative of the entire watershed.

The procedure for fitting the watershed model is similar to the snowpack fitting procedure. The major difference is in the number of parameters that must be adjusted. Both the size of the reservoirs shown in Figure 2 and their rate relationships must be calibrated, making this a more complex calibration than the snowpack. The model parameters must be modified until both reasonable replication of the streamflow hydrographs and minimal water balance errors are obtained. Again, a pattern search optimization routine is employed after the best "hands on" fit is obtained.

Once the two modules have been calibrated, an appropriate period, such as the subsequent 8-10 years of record, is selected for verification. Both models are run for the verification period. If the errors for the calibration period (summarized by mean square error, average absolute error, cumulative error, or other criterion) are similar to the error values obtained during the verification period, and the snow course ablation curves and flow hydrographs are matched consistently, the calibration is considered valid. If these conditions are not met, the calibration period is extended to include part of the verification period and a new period following the extended calibration period is used for verification. If verification still cannot be obtained, a complete recalibration must be considered, and ultimately the model validity may be questioned.

When both modules have been calibrated and the calibration verified, forecasts may be made using one of several approaches. Most snowmelt forecasts are made in the spring (February through May) for cumulative runoff through a "summer" end date usually ranging from the beginning of June to the end of September. When forecasting using a continuous simulation model, time series of the input data must be prepared to drive the model during the forecast period. These synthetic temperature and precipitation records may be scenarios selected from past records to bound the forecast by describing extreme high and low conditions and median conditions. Alternately, synthetic weather sequences generated by empirical or probabilistic means may be used.

The validity of a forecast is determined by comparing with observed values. That is, the model is run with real input up to a historic point (i.e., February 1965) and run with the synthetic data thereafter. The error in prediction can then be assessed. Repeating this process for many input scenarios over many years allows one to describe the accuracy of such forecasts under a variety of conditions. The procedure used is described in greater detail in Chapter 4 as applied to the modified Tangborn model.

CHAPTER 3 CONCEPTUAL SIMULATION MODEL EVALUATION

This chapter contains an evaluation of the Burnash/ Baird snowmelt and the GHM land modules based on application of the model to the Cedar River Basin. Specific emphasis is placed on model transferability, especially with regard to the snow module.

3.1 The Cedar River Experience, a Summary

The most extensive snow course record near the basin is the record at Stampede Pass, located near a U.S. Weather Service Station (see Figure 7, Chapter 4). Although this station is located approximately 5 miles south of the Cedar headwaters, the snow course is easily accessable (due to the proximity of the weather station) and is consequently read more frequently during the snow season than more remote snow courses. The greater frequency of measurement (eight to twelve and more measurements per year as opposed to two to three at more remote sites) provides more information on the overall rates of accumulation and melt. This additional information allows for more precise fitting of the snowpack module to observed conditions. For this reason, the Stampede Pass snow course was used in preference to several lower elevation snow courses located within the basin.

The snowpack module was calibrated to the Stampede Pass snow course for water years 1952-1961. A root mean square error (RMSE) of 5.02 inches of water content was achieved by the optimization routine. Calibration of the module consisted primarily of manipulating the temperature adjustment, radiation melt coefficient and wind speed factor to approximate the observed snow water content values. After initial calibration resulted in preservation of the approximate shape of the observed snow course time series, the optimization routine was employed to obtain the minimum error fit.

The module is designed to use up to five precipitation records from gages in the vicinity of the watershed and one temperature record. The following gages were chosen for this study due to their proximity to the study area and quality of record: Landsburg (45-4486), Snoqualmie Falls (45-7773), Cedar Lake (45-1233), Stampede Pass (45-8009) (see Figure 7, Chapter 4 for raingage location). Initial weights of 0.25, 0.25, 0.4, and 0.1, respectively were applied to the records due to their distance from the watershed and, in the case of Stampede Pass, to consideration of high potential for catch deficiency associated with high winds experienced in this area.

Figures 4a-c show the observed and simulated snowpack water content for water years 1955-57. Note that the minimum error (RMSE) criterion achieves a fit to the observed snowpack that roughly balances the cases where the simulated values exceed observed values against those cases where the reverse is true. The result for the entire ten year calibration period is that, while the minimum error is obtained, the timing of the spring melt is not always well matched. This is likely to lead to systematic within-year errors for subsequent runoff simulation, since accurate simulation of maximum snow accumulation is usually acknowledged to be one of the most critical variables in snowmelt runoff simulation (Carroll, 1978). Most of the years that were simulated show relatively large differences between the observed and simulated snow packs at some point in the year. Table 1 summarizes the quality of each calibration year's fit and Table 2 contains the optimized values of the calibration parameters.

Once calibrated, the snowmelt module was used to generate five pseudo-precipitation records. The records were generated for five altitudes representing the mid-points of five equal altitude divisions of the watershed altitude range to provide a basin-representative mix of rain-melt records for input to the watershed model.

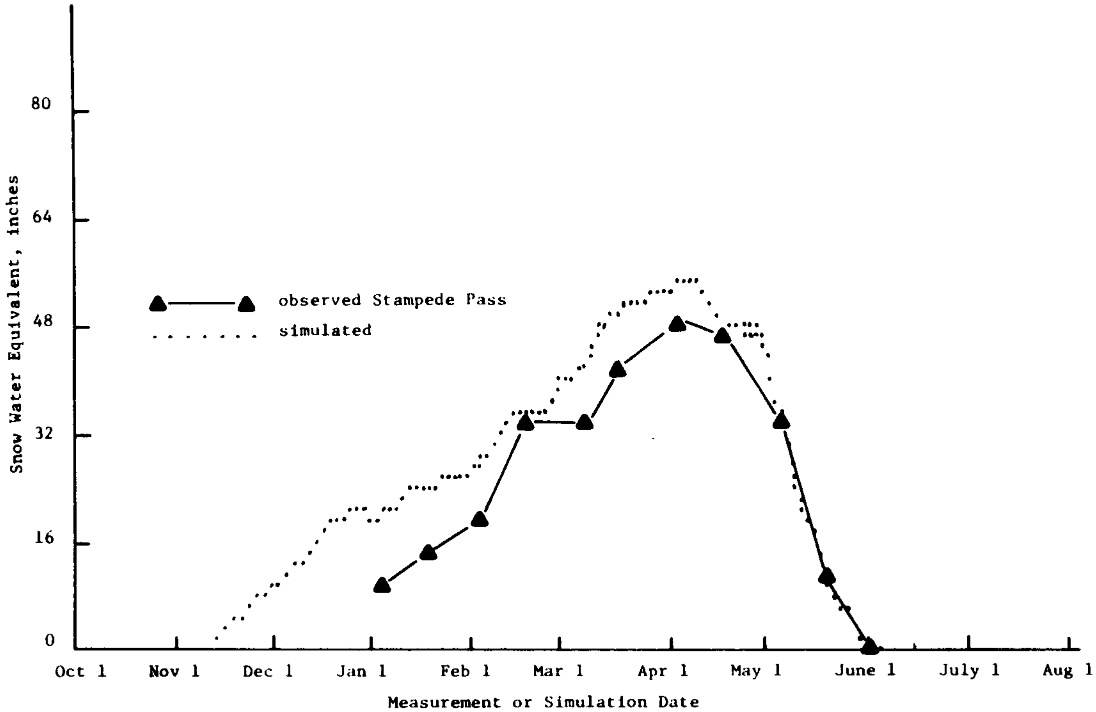

Figure 4a. Simulated and Recorded Snowpack Water Content at Stampede Pass, Water Year 1955.

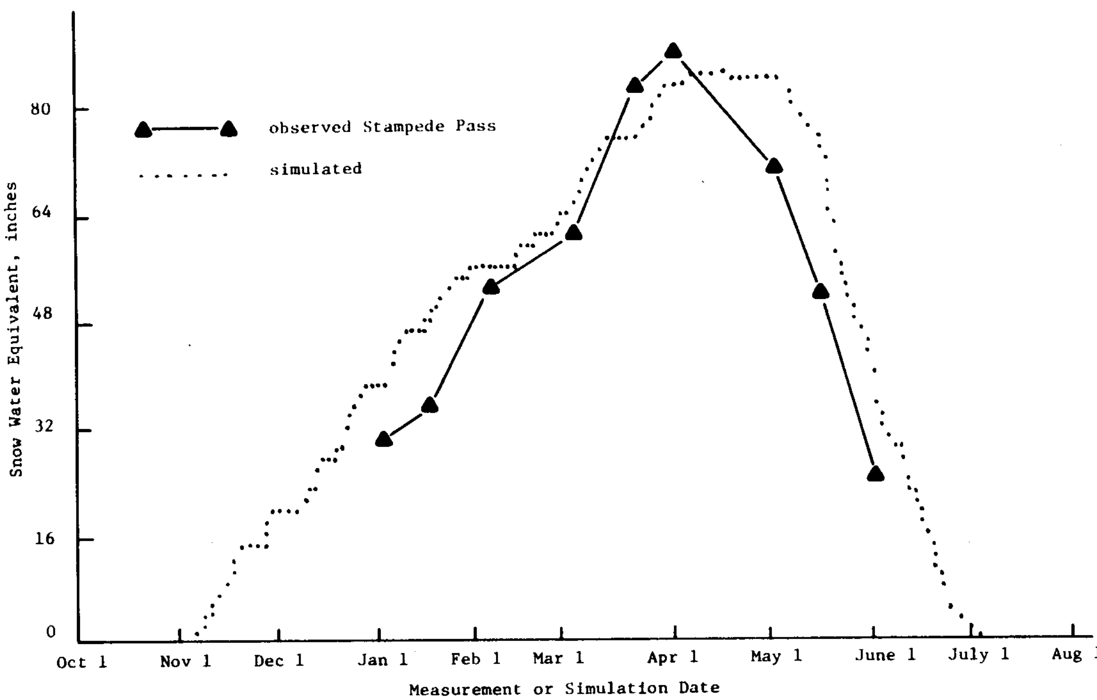

Figure 4b. Simulated and Recorded Snowpack Water Content at Stampede Pass, Water Year 1956.

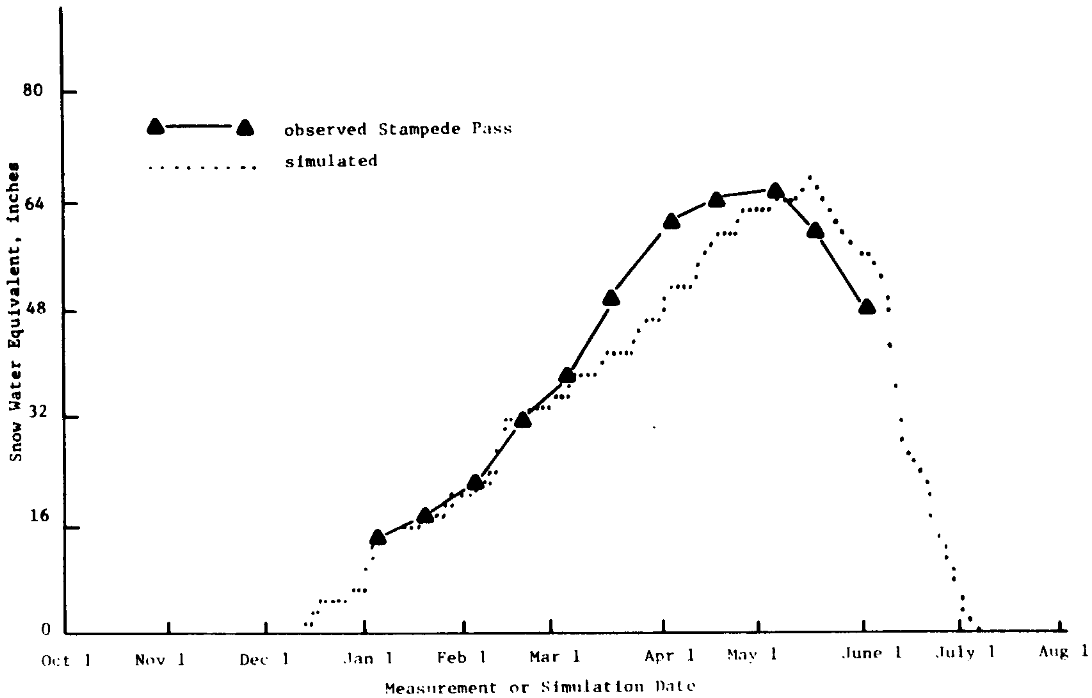

Figure 4c. Simulated and Recorded Snowpack Water Content at Stampede Pass, Water Year 1957.

Table 1. Summary of Snowpack Module Calibration Fit

Year	Description
1952	few observations, slightly undersimulated.
1953	variable - undersimulated January and February, oversimulated March and April, undersimulated May (melts too rapidly).
1954	consistently undersimulated, melts too rapidly.
1955	good fit in January and February, undersimulates March and April, oversimulates May and June (melts too late).
1956	consistently undersimulates except in late spring, melts too late.
1957	consistently oversimulates, but melt timing good.
1958	undersimulates, melts too rapidly.
1959	early season oversimulated, late season undersimulated, melts too rapidly.
1960	consistently oversimulates, melt timing approximately correct.
1961	generally oversimulated except late spring, melt timing approximately correct.

Table 2. Snowpack Module Parameter Estimates

Parameter	Description	Estimated Value
TCR	Temperature adjustment in addition to lapse rate	-3.5°F
WS	Daily wind factor	42.7 mi/day
CST	Radiation melt coefficient	.00215
ABSCF	Shading modification factor	3.07×10^{-6}
PWATE	Overall precipitation weight	1.312
PCP1	Individual precipitation weight Landsburg	.478
PCP2	Individual precipitation weight Cedar Lake	.203
PCP3	Individual precipitation weight Snoqualmie Falls	.207
PCP4	Individual precipitation weight Stampede Pass	.111

In fitting the watershed module to the observed hydrograph records, the need to distinguish between those activities more appropriately described as curve fitting and those that were based on hydrological considerations arose. This issue will be dealt with extensively in Section 3.4.2, but the following interim definition is necessary to clarify discussion of the watershed module calibration process. The term curve fitting will be used to refer to model parameter adjustments made to improve hydrograph fit that cannot be defended on sound physical principles. Adjustments that can be so defended do not fall under the category of curve fitting and are defensible as part of the calibration procedure.

The initial efforts to fit the watershed module included: 1) extracting the baseflow and interflow recession coefficients from a semi-log plot of the observed record, 2) estimating initial parameter values from comparison with an example model run contained in Burnash, et.al. (1973), and 3) assembling all basin physical characteristics needed to estimate other parameters. Base flow recessions on a semi-log hydrograph plot were extracted by comparing the slopes of approximately straight portions of the hydrograph during the summers of the lowest flow years of the calibration period. Projection of this slope back to the earlier parts of the recession provided the means to remove the interflow recession by difference.

Some parameters such as the percolation rate coefficient are only amenable to estimation by calibration of the model. Initial values for several parameters were selected from the example in Burnash, et.al. (1973) to achieve the correct order of magnitude and relative size without needless model iteration. Initial weights on the pseudo-precipitation records were taken directly from the hypsometric curve. The high altitude band of the watershed contained the least area and thus had the lowest weight; the other records were weighted accordingly.

The initial determination of the annual potential evapotranspiration (PET) curve was obtained using the following procedure. Twenty year averages of daily pan evaporation records maintained at the Agricultural Experiment Station in Puyallup, Washington (latitude 47°12' N, longitude 122°20' W) were compiled. Using suggestions contained in Burnash, et.al. (1973), monthly pan weights were developed that skewed the distribution of potential evapotranspiration so larger values of PET were observed in September, October and November than a simple scaling of the pan observations would suggest. The magnitudes of the weights were first adjusted to yield an annual PET at the mid-point elevation (3250 ft) of the watershed of 15 inches per year; approximately half of the annual pan evaporation at Puyallup (elevation 50 ft).

Initial and final values of the model parameters are included in Table 3. The adjustments made were based on the following considerations. The

Table 3. Initial and Final Values of Land Module Parameters

Parameter	Description	Initial Value	Final Value
RSW1	Pseudo-precipitation record 1 weight	.20	.21
RSW2	Pseudo-precipitation record 2 weight	.25	.22
PSW3	Pseudo-precipitation record 3 weight	.25	.25
RSW4	Pseudo-precipitation record 4 weight	.20	.23
RSW5	Pseudo-precipitation record 5 weight	.10	.14
UZTWM	Upper zone tension water maximum capacity	3.0	4.0
UZFWM	Upper zone free water maximum capacity	3.0	3.0
LZTWM	Lower zone tension water maximum capacity	5.0	6.0
LZFSM	Lower zone free water supplemental maximum capacity	3.0	2.0
LZFPM	Lower zone free water primary maximum capacity	8.0	8.0
UZK	Upper zone (interflow) recession constant	.12	.15
LZSK	Lower zone supplemental base flow recession constant	.05	.05
LZPK	Lower zone primary base flow recession constant	.0078	.0078
ZPERK	Percolation scaling factor	30	50
REXP	Exponent on percolation rate function	1.8	1.2
PCTIM	Percent impervious area	.01	.02
SARVA	Area of watershed containing phreatophytes and riparian vegetation	.018	.018
ADIMP	Added variable source area	.02	.10

first few runs of the model had consistently low hydrographs in the late melt months, June and July. An increase in the rain weights of the upper altitude zone pseudo-precipitation records was effected to cause more late melt to occur.

The upper and lower zone tension water capacities (UZTWN and LZTWM) were increased to provide more absorptive capacity in the first months of autumn. Early runs showed extremely high over-reaction to relatively small precipitation events in the autumn as the tension water zones were not able to absorb adequate moisture. This effect was combined with the low, 15-inch per year, PET which did not adequately remove moisture from UZTW.

Adjustments of upper zone free water capacity (UZFWM) produced the following results. Increasing UZFWM produced somewhat sluggish response and excessively prolonged interflow recessions followed by inadequate base flow. Decreasing UZFWM produced over-responsive behavior. It appears that larger values of UZFWM caused percolation to be reduced as the fraction UZFWC/UZFWM (contents/capacity) was forced to lower values. The smaller values of UZFWM tried produced more surface runoff. Though small incremental changes in UZFWM may be appropriate, the general magnitude appears correct.

The lower zone primary and supplemental capacities (LZPWM, LZSWM) were adjusted iteratively with ZPERC and REXP (as all four values are interactive) through the percolation function. Increases in the percolation rate to reduce over-reaction of the upper zone and adjust the lower portions of the major recessions were made by decreasing lower zone capacities so the ratio of lower zone deficiencies to capacities would become larger. Similarly, increasing ZPERC to increase percolation and reducing REXP to prolong the length of the exponential decay of the percolation function resulted in greater amounts of water percolated.

The lower zone recession coefficients were extracted from the observed hydrographs and appeared to adequately represent the observed base flows. The upper zone recession coefficient remains a matter of some concern. Conversations with R.L. Ferral (National Weather Service River Forecast Center, Sacramento) indicated that UZK may be as high as 0.35-0.5 in some cases. The interflow recession coefficient is the most difficult of the three recession coefficients to extract from the semi-log plot of recorded flows, and thus is most suspect when simulation errors occur. Use of UZK values approaching 0.3 produced very spiked, overreactive behavior in the simulation that could not be easily compensated for through adjustment of the other parameters. The incremental increase to 0.15 produced a small improvement in fitting some peaks and was thus retained.

Estimates of the percent impervious area (PCTIM), the fraction of the basin covered by lakes and streams and riparian plants (SARVA) and the impervious area added during storm events (ADIMP) were made from the USGS NAWDEX basin characteristics. These parameters were taken as fixed and were not adjusted in the calibration procedure.

Of the adjustments to model parameters described above, the changes in precipitation weights, the increased tension water capacities, and, to some extent, the adjustments to the percolation rate stem from reasonable deductions about the hydrologic system being modeled. The other manipulations fall more in the curve fitting category as they are less clearly supported by physical logic.

Figures 5a-c contain the simulated and observed hydrographs from the final calibration run for water years 1955-57. The following paragraphs discuss the characteristics of the simulation and possible reasons for the discrepancies. Discussion of the causes of various discrepancies will be

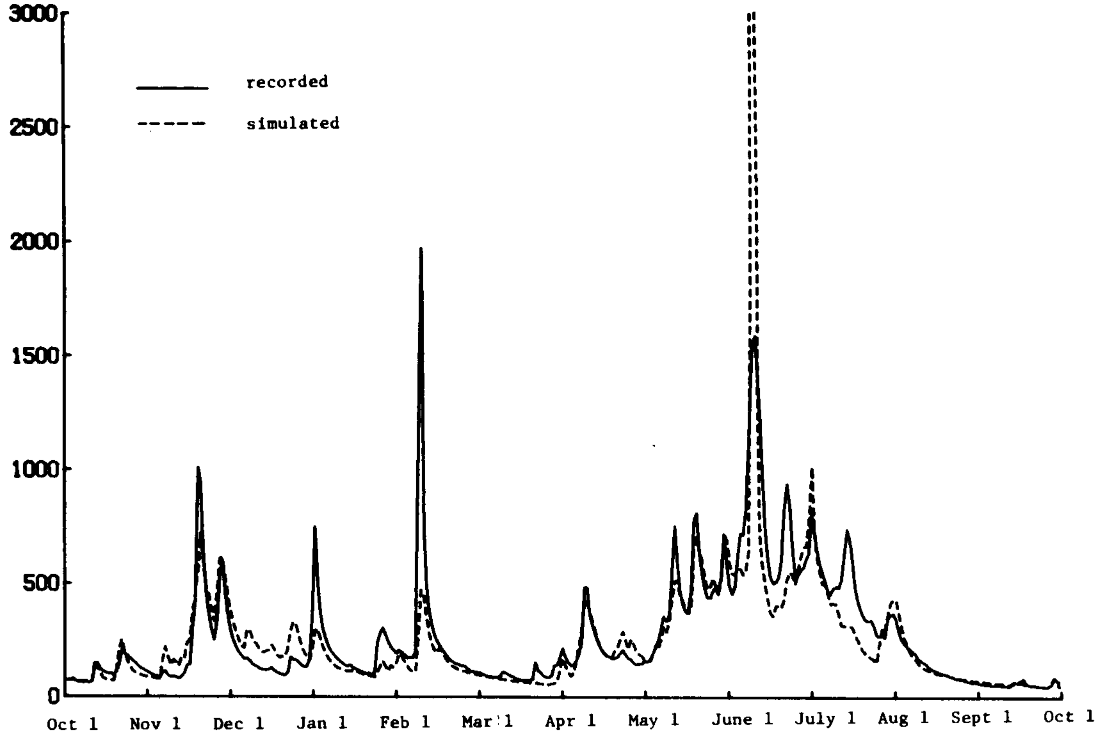

Figure 5a. Simulated and Recorded Runoff at USGS Gage Station No. 12-1150 (Cedar River Near Cedar Falls) for Water Year 1955.

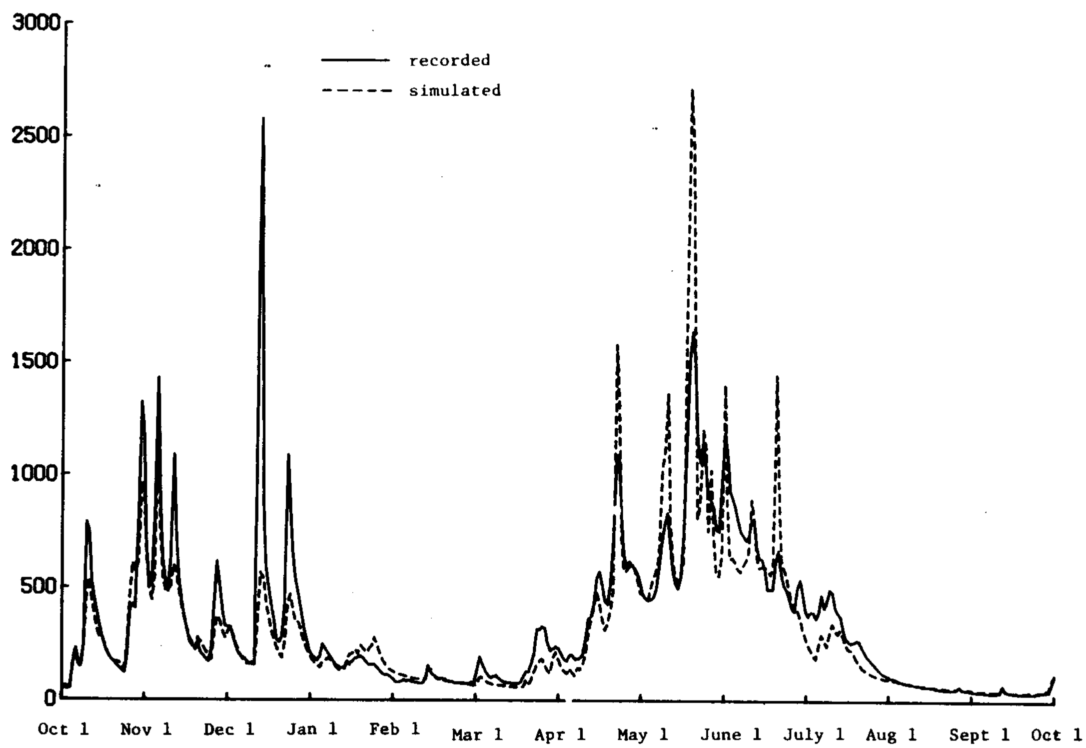

Figure 5b. Simulated and Recorded Runoff at USGS Gage Station No. 12-1150 (Cedar River Near Cedar Falls) for Water Year 1956.

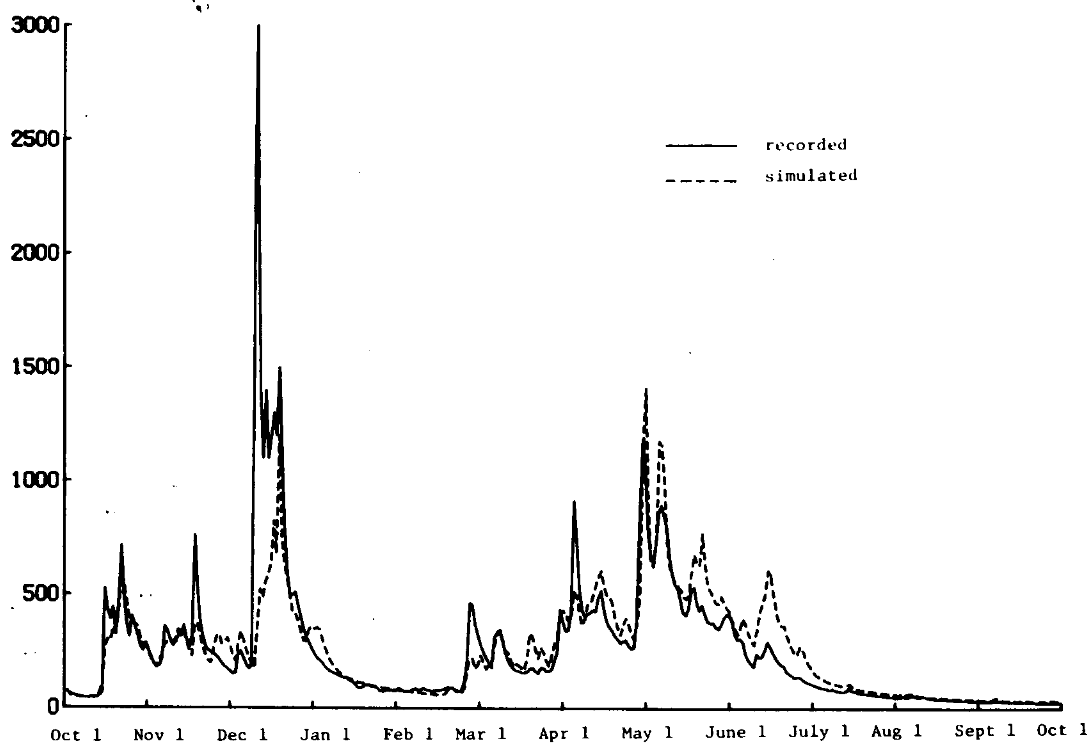

Figure 5c. Simulated and Recorded Runoff at USGS Gage Station No. 12-1150 (Cedar River Near Cedar Falls) for Water Year 1957.

confined to a description of interactions of existing model components. Underlying causes of discrepancies and evaluation of the model are deferred to Section 3.2.

First, noting the autumn months in water year 1955, there is some tendency to oversimulate basin runoff peaks until mid-November. This oversimulation may be due to inadequate removal of water from UZTW during the preceeding months. Similar effects occur in the 1953, 54 and 60 water years, indicating that increased PET may be required to fit these portions of the hydrograph. The runs shown, however, have an annual PET demand of 27.2 inches distributed as shown in Figure 6. This is nearly equal to the annual pan

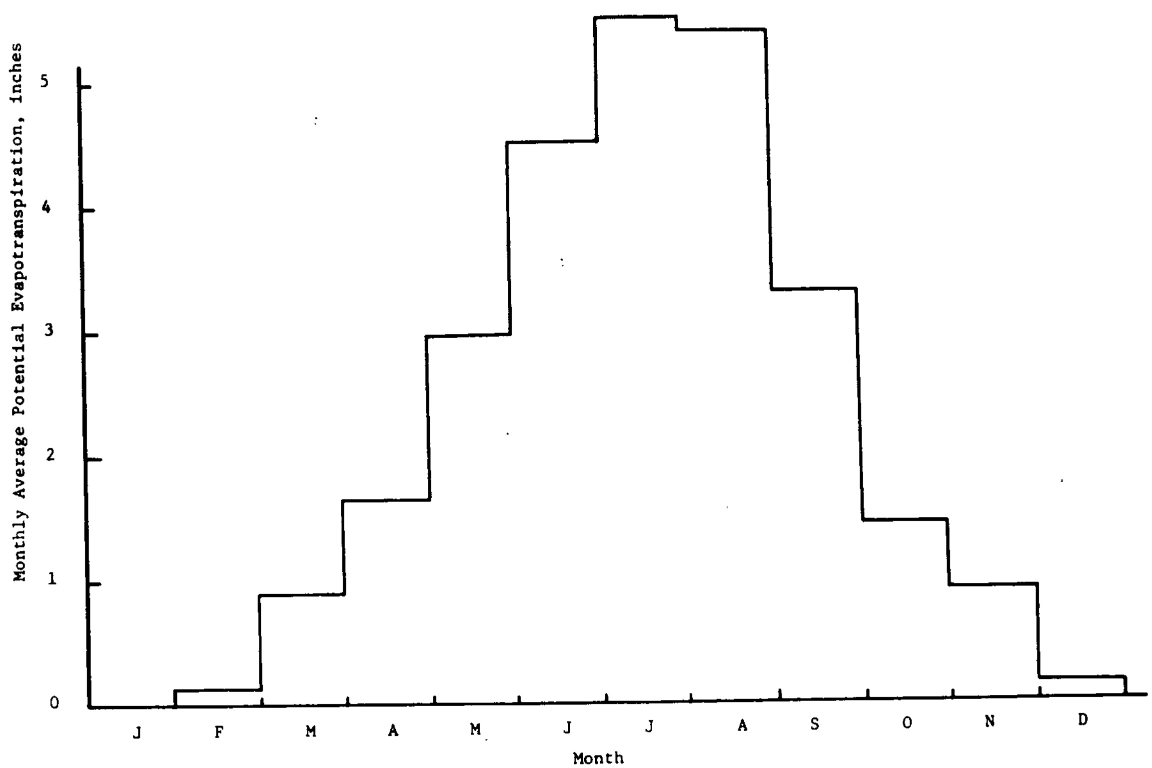

Figure 6. Potential Evaporation Demand at Basin Midpoint (elevation 3250 ft) Used in Model Calibration.

evaporation observed at Puyallup which is near sea level in an open plain, and is therefore very difficult to justify physically. Use of a PET demand of 27.2 inches did reduce the oversimulation phenomenon in most calibration years, however.

In all years the base flow recessions are fit with reasonable accuracy, indicating that the sizes of the lower zones and their recession coefficients are reasonably well estimated.

Several discrepancies in the hydrograph simulation appear to be associated with the precipitation input during the winter months. Examples of under-simulated cases are January and February of water year 1955, December of 1956, March of 1956, December of 1957 and March of 1957. Conversely, large overestimates of runoff are calculated for a storm in June of water year 1955, May and June of 1957, and March and June of 1957.

Throughout the 9-year calibration period several discrepancies between the simulated and observed hydrographs tended to be repeated. Late spring storms in several years tend to be grossly oversimulated. Examples of this phenomenon in the three sample years in Figures 5a-c are the first peak in June of 1955 and the second peak in May of 1956. Similar phenomena occur four times in the nine years of the calibration.

There are also several cases where early and mid-winter runoff events are drastically undersimulated. Examples are February of water year 1955, December of 1956, December of 1957 and February of 1957. Clearly, insufficient precipitation reaches the watershed at these times.

A third characteristic area of discrepancy occurs at the end of the spring runoff when a major recession occurs in June or July of most years as the river settles down to summer base flow conditions. In the years when This decline occurs in July, the hydrographs are consistently undersimulated indicating that late season snowmelt may occur after the simulated snowmelt ceases. In the years when this decline occurs in June, the hydrographs are generally oversimulated indicating that too much melt may be simulated in June.

The selected example years' hydrographs illustrate the effects described above. Examination of the data indicates no anamolous values for the days in question. The most extreme event occured on a warm June 9, 1955, when the weighted sum of observed basin precipitation was 0.22 inches and the daily mean temperature lapsed to the basin middle elevation at 3°F/1000 ft was 73°F. On this date the snowpack model simulated a melt of 14 inches of water content at Stampede Pass. When weighted as pseudo-precipitation input to the watershed model, this caused a flow in the watershed simulation of 134.5 cfs/mi^2, roughly twenty times the mean annual flow, and several times larger than the highest recorded spring melt peak.

On the other hand, the model appears to perform quite well in many cases. October of water year 1956 fits reasonably well as does the following November. As mentioned earlier, the summer recessions and most autumn months show a reasonable fit. For these summer events, the snowpack is depleted, however, and the snow module passes precipitation directly to the land module. The most pronounced errors seem to be associated with events involving the snowpack algorithm, particularly during the melt season.

Most discrepancies seem to arise from cases where the snowpack model either stored too much precipitation as snow or released too much snow storage to the pseudo-precipitation process. The timing and magnitude of precipitation events has a large influence on disposition of the water in both the model and the prototype system. The same total volume of monthly precipitation delivered in a 30-day drizzle or in a 2-day storm will generate widely varying runoff events. A larger portion of the slow drizzle will be available for evapotranspiration and less will be available for percolation to the ground-water aquifers because evapotranspiration will be able to remove water from the soil mantle nearly as fast as it accumulates. On the other hand, heavy precipitation and accompanying high runoff will cause positive water balance errors as the water that should percolate will run off due to saturation of the soil layers.

3.2 Model Assessment

At this point in the model implementation process it became apparent that accurate calibration, verification and use of the model as a forecasting tool required a better representation of the snow accumulation and ablation process than was possible using the existing form of the snowpack module. Implementation of the model in a forecast mode was not felt to be realistic owing to the substantial systematic errors encountered. Rather, the experience reported above was used to provide the basis for a critical evaluation of the two modules of the watershed model.

3.2.1 Model Evaluation-Snow Accumulation

From the description of the errors noted in the fit of the simulated hydrographs described above it is apparent that the timing and volume of snow-melt runoff are critical factors for simulating winter and spring hydrographs. The accuracy of the melt and rain time series input to the watershed module determines the ability of the model to generate flows matching those observed. Since the pseudo-precipitation record is generated by the snowpack simulation, any errors in snowpack algorithm effectively propagate through the entire model. The following paragraphs discuss the possible sources of error in the snowpack simulation. As in previous discussions of the algorithms, the logical order of water passage through the simulation will be followed.

The snowpack algorithm uses a fixed lapse rate calculation to determine the temperature at the snow course altitude from low elevation records. The lapse rate used is 3°F per 1000 feet. This corresponds to the saturated adiabatic lapse rate and contains the inherent assumption of high humidity such as would occur during a storm event. It would appear that this assumption could be the source of considerable error as the lapse rate should range from a minimum near 3°F/1000 ft for saturated conditions to 5.4°F per 1000 ft; the dry adiabatic lapse rate. There are a number of clear, dry days throughout the winter and spring that would be expected to experience greater lapse rates and thus cooler snow course temperatures. The calibrated temperature adjustment parameter may compensate for errors in the lapse rate calculation to some extent by acting to balance high estimates of the temperature at the snow course against low estimates. The degree of melt occuring in any one period determines the shape and magnitude of hydrograph peaks; if the basin temperature is inaccurately determined at least the timing and probably the magnitude of melt events will be incorrectly simulated, producing poor hydrograph fits and more importantly significant water balance errors in runoff simulation using the watershed module.

The partition between rain and snow used in the snowpack algorithm is probably unrealistic. This partition, taken from Winston (1965), suggests that when the ambient air temperature near the snowpack surface is 32°F, eighty per cent of the precipitation occurs as rain. Further, the air temperature must be below 27°F before all precipitation occurs as snow. Examination of Stampede Pass weather records, which include daily snow accumulation, indicate that this is simply not the case at Stampede Pass; almost all precipitation occuring at 32°F occurs as snow. Assuming that the air temperature at the snow course was 32°F and the assumed 3°F/1000 ft lapse rate prevailed, the air above the snow course would be expected to be at less than 32°F unless an inversion condition prevailed. Although this condition is observed as a temporary phenomenon in one or two storms per year, it does not occur as a permanent condition. It would seem more likely that precipitation reaching the snowpack surface would be in the solid phase if the air temperature was at 32°F, having fallen through cooler air. Anderson's (1968) results seem to support this. In Anderson's work a wet bulb temperature of 33°F is used to partition precipitation between rain and snow. The partition between rain and snow is extremely important in the Cedar basin, since small temperature changes have large effects on the basin area receiving snow in the majority of winter storms.

3.2.2 Model Evaluation-Radiation Melt

Burnash, et.al. (1975) report that cloud cover may be estimated from the daily temperature spread and season in central California. This relationship is used to determine the degree of cloudiness used to modify the radiation reaching the snowpack. Such a relationship may not hold in other areas and tends to reduce the general applicability of the model. The procedure used to derive this relationship should be included in the model documentation to allow evaluation of its applicability in other regions. Incorrect estimation of cloud cover may potentially cause a systematic bias in the amount of radiation reaching the snowpack surface. Such bias is not readily apparent in the model, however, due to influences of the albedo calculation and of the calibrated radiation melt conversion coefficient, both of which are discussed below.

The albedo of the modeled snowpack decreases with pack age. New snow has an albedo of 0.82, but this may decrease to about 0.6 as the pack ages. The albedo calculation is a modification of Winston's (1965) empirical procedure which was based on work by the U.S. Army Corps of Engineers (1956). The general range of albedo values used matches that used by Anderson (1968). The potential amount of error in the albedo calculation seems to be less significant than other aspects of the algorithm.

The radiation melt portion of the algorithm reduces the latitude-adjusted solar radiation using the cloud cover calculated as described above. The relationship between radiation reaching the pack surface and melt is determined by a simple linear multiplication. The coefficient that converts radiation to melt is a calibration parameter and, as such, may either compensate for (or alternately, obscure) the effects of inaccurate simulation of cloud cover, albedo, etc.

The radiation melt calculation does not include any consideration of back radiation from the atmosphere. Atmospheric radiation may not be a significant source of melt energy and it may not be estimable from the sparse data base in any case. The implicit assumption in using a calibrated radiation melt coefficient is that the calibrated value will include effects such as this, which are not explicitly included in the model.

3.2.3 Model Evaluation-Condensation and Convection Melt

The calculation of melt due to convection and condensation is a direct modification of Winston's (1965) method modified to deal with daily average temperatures and wind velocities. The condensation term assumes that dew point temperature is four degrees below the minimum temperature, or 32°F, whichever is higher. Condensation melt is calculated by converting the dew point-freezing point temperature difference into melt using a coefficient based on the heat of fusion. The convection term uses wind speed as a calibrated parameter to adjust the relationship between average temperature - dew point temperature spread and air density. Suggested values of the wind speed parameter were in the range from 1 to 10 nautical miles of wind movement per day. This parameter is the only portion of the convection-condensation melt equation that is subject to calibration. The apparent incompatibility of the values obtained for this parameter with observed wind velocities makes one suspect that the convection-condensation melt term may be a source of substantial error in simulation of the melt process. The condensation term, in particular, may be quite important in the Cedar Basin where many rain on snow events occur in the course of a winter, particularly at the lower elevations.

3.2.4 Model Evaluation-Snowpack Compaction

As the pack ripens it compacts according to the algorithm described in Appendix A. No calibrated adjustments of the compaction process are provided in the model. The inherent assumption in this case is that all snowpacks behave according to the mechanics of the algorithm. Work by Gerdel (1954) indicates the average free water retention capacity of a snowpack to be about 2%. The pack compaction routine effectively removes any free water from the pack in one or two days when no melt occurs. This squeezing effect may remove some water from the pack before it actually should be. The compaction routine would not appear to correspond to Gerdel's observation over a prolonged period.

3.2.5 Model Evaluation-Land Module

The most important part of the snowpack model is the pseudo-precipitation record it generates. Earlier it was shown that incorrectly predicted melt events can produce unusual runoff events which contribute to large water balance errors in runoff simulation. If possible it would be best to calibrate the model to produce the correct pseudo-precipitation record, rather than snowpack accumulation, but the information which would be required for such a calibration is not commonly available.

While the snowpack model appears to have some severe problems in describing the phenomena it was developed to simulate, the watershed model apparently adequately models the basic runoff processes. Conceptually, the watershed model accounts for the basic hydrologic phenomena, i.e., interflow, evapotranspiration, etc., and does so with relationships based on recent research. In the Cedar River application the watershed model performed well when the pseudo-precipitation record consisted mainly of rain with little snowmelt as in the summer and early autumn.

There is one area, however, where the realism of the land module must be questioned. The upper zone tension (UZTW) reservoir acts as a buffer to reduce the runoff effects of storms occuring after prolonged dry periods. UZTW is only depleted by evapotranspiration which is input to the model as monthly mean values. Consequently, as discussed in section 3.1, it may be determined that PET in excess of physically reasonable quantities is needed to calibrate the model.

3.3 Model Data Requirements

From the foregoing discussion of the quality of the snowmelt runoff simulation obtained in applying the combined snowmelt-runoff model it is apparent that the main source of error in runoff simulation lies in the magnitude and timing of snowmelt. A number of potential errors in the energy balance approximation used in the snowmelt model have been discussed in the preceeding sections. Testing of the model under a wide variety of conditions is perhaps the best method to disclose such problems.

As noted in the introductory chapter, assumptions and approximations made to allow modeling of a process using sparse data sets can be the deciding factor in general applicability of a model. Three categories of assumptions are those required to approximate portions of the process being simulated and those required to insure that the model provides spatially realistic output, despite the point simulating employed.

3.3.1 Data Adjustment

When attempting to continuously simulate a phenomenon at a remote point where continuous data collection is not performed or where the required data are not collected, a method must be derived for displacing the available data from the point of collection to the point of simulation. Areal average precipitation has been estimated using weighted averages such as the Thiessen method and isoheytal area weighted averages (Linsley et.al., 1975). Temperature changes with altitude can be estimated by assuming a lapse rate and potential evapotranspiration can be derived from assumed relationships to pan evaporation records.

To the maximum extent possible, these assumptions should be subject to modification by observed conditions. For example, the saturated adiabatic lapse rate could be assumed for stormy conditions and the dry adiabatic lapse rate could be assumed for clear weather conditions.

3.3.2 Data Estimation

When required data are simply unavailable, development of generally applicable relationships for generating such data is necessary. The relationship between season, daily temperature spread and cloud cover used by Burnash and Baird (1975) is an example of such an assumed relationship that has proven workable for a specific area. When general approximations of a particular phenomenon appear to be beyond the state of the art, the model should include a methodology for developing a locally applicable alternative that will perform with the same reliability as the original assumed relationship.

3.3.3 Areal Representation of Data

A final adjustment is required to make the model areally representative. Assignment of weights to input data (observed precipitation in the snowpack simulation, pseudo-precipitation in the watershed model) inherently assumes that the basin average conditions can be approximated as a weighted sum of observed conditions. Data limitations force such an assumption on the modeler in most situations. In the case of the Sacramento model, use of the optimization routine can identify the set of weighting coefficients that gives the best fit to observed data.

The weighting scheme used in the snow module, however, may be of questionable validity in one respect. The observed precipitation records are initially assigned weights that achieve the best fit of the simulated snowpack to observed snow course measurements. The pseudo-precipitation records generated from the snow module are then re-weighted when input to the watershed module. During the summer and early autumn, when no precipitation falls as snow and all (or nearly all) melt has occured, the precipitation weights used to estimate snowfall are directly applied to the watershed. That is, the precipitation weights that best described a process that only occured in winter (snowfall) are applied year-round by virtue of being incorporated in a single time series: the pseudo-precipitation records. Fortunately, the most significant precipitation events do not occur during the snow-free period so any errors that arise as a result of this phenomenon may be expected to be small.

Current research underway at the Hydrologic Research Laboratory of the National Weather Service aimed at generation of areal mean precipitation records from a single gage record could potentially aid greatly in reducing model errors currently attributable to spatial errors in input data. In addition, the necessity for inclusion of some of the multiple weighting coefficients as calibration parameters might be averted.

3.4 Model Evaluation-General Observations

A final general criticism of the model system's structure concerns the degree of sophistication of the two component modules. The snowpack module is a highly empirical model that uses a number of assumptions and approximations that do not appear to have been verified under a wide variety of conditions. The watershed module is a state of the art simulation of conceptual hydrologic processes that uses the results of recent research to describe watershed processes. Errors in the snowmelt simulation propagate through the watershed module making its output less reliable than desired. If possible it is desirable to couple modules of roughly equal sophistication so the accuracy of one is not diminished by the errors of the other.

3.4.1 Model Transferability

Model transferability includes two distinct issues, first, applicability of the model in regions other than the locale where it was developed, and second, technology transfer, the logistics of communication of the techniques of model use. A closely related topic is the question of physical reality or the degree of belief one has in the ability of a model to properly mimic component processes.

From the model evaluation discussion above one may conclude that the Burnash/Baird snowpack module does not successfully simulate the daily snow accumulation and melt phenomena. The description of sources of error in fitting the runoff hydrographs (section 3.2) provides the basis for this conclusion as related to modeling of the Cedar River basin. Some of the problems appear to be related to the region-specific basis of the snowpack module dynamics.

The land module, on the other hand, was generally able to describe the dynamic details of several types of watershed input and response events, including summer recessions and storm peaks of varying magnitudes. Adjustments to the watershed module parameters were largely made on the basis of hydrologically reasonable assumptions about basin repsonse. With the exception of the potential evapotranspiration issue (section 3.2.5), where more work is required to make a definitive statement, the watershed module appears able to represent the entire range of desired hydrologic phenomena and, on the basis of the Cedar River experience, transferability would be rated as good.

When considered as a hydrologic simulation system the combined snowpack and watershed model fails as a generalized, transferable model due to the errors in the snowpack module. Burnash et.al. (1973) report that the watershed module has been subject to extensive testing on several watersheds over a large range of sizes and weather conditions. Similar extensive testing of a snowmelt model over a wide range of climatic conditions would point out the types of problems encountered in this application and provide valuable input to the model development process. It should be noted that the National Weather Service Hydrologic Research Laboratory is currently carrying out such testing on a model which consists of a snow module described by Anderson (1973) and uses the GHM as the land module.

3.4.2 Physics Versus Art

The amount of expertise needed to use a model is a critical issue for a decision maker or organization considering implementation of a hydrologic model. James and Burges (1978) discuss the issues of model selection and implementation in detail. The following discussion briefly evaluates the amount of user knowledge required for economical application of the snowmelt-runoff modeling system described in Chapters 2 and 3 of this report.

The distinction between physically justifiable adjustments to model parameters and curve fitting was briefly discussed in section 3.1. Many of the processes contained in the snowpack and GHM modules involve calibration of synergistic rate-governing coefficients. Consequently, a fair amount of familiarity or "feel" for the various sensitivities of the model is required of the user. The degree to which this is true for any model may be an important selection criterion, depending on the expertise of the user. The question is one of how effectively a user inexperienced in use of a particular model (although presumably somewhat familiar with the real process dynamics) can implement the model. Several aspects of the calibration reported for the Cedar River are a result of adjustments made for hydrologically acceptable reasons. For example, increasing the relative weight of the upper altitude pseudo-precipitation records to account for late spring melt effects, estimation of an extra one per cent of basin impervious area to account for seeps and bogs in the valley bottom and rocky areas at higher elevations, and to some extent, increasing the overall basin potential evapotranspiration to account for the predominately dense conifer forest covers's ability to remove a large volume of water from both shallow and deep root zones, can be justified on a reasonable physical basis.

Other adjustments, particularly to the size of the various subsurface storage zones and the percolation rate factors appear to fall more clearly under the realm of modeler experience or "art", although they should not necessarily be considered as curve fitting. It may be expected that an experienced user of the model could achieve a somewhat better fit given the same data set. Although the value of user experience is undeniable, more widespread use of conceptual simulation models will require reduction in the amount of "art" or user expertise. For example, the National Weather

Service has plans to implement the NWS River Forecast System on some 2500 river basins throughout the country. The logistics of this enterprise dictate that parameter estimation will have to be reduced to a more scientific level, most likely through implementation of more efficient parameter optimization techniques (see, for example, Leavesly, 1978).

3.5 Summary

An (updated) modified version of the Generalized Hydrologic Model developed by Burnash, et.al. (1973) was applied to the Cedar River, Washington. The snowpack simulation model developed by Burnash and Baird (1975) was used to simulate snowpack accumulation and melt. The combined melt and rain time series produced by the snowpack model was input to the GHM as the driving precipitation function.

The resulting hydrograph simulation obtained, indicated that while the watershed model (GHM) appeared fully capable of simulating the hydrologic characteristics of the watershed, the snowpack simulation produced large errors in the timing and magnitude of runoff events due to inadequate accuracy in simulation of the snowpack mechanics. This apparent inability to adequately describe snowpack processes suggests that the combined snowmelt-runoff modeling system is not transferable to the Cedar River without modification of the snowmelt module. In fairness it must be pointed out that the Cedar basin presents an extremely severe test of any snowmelt model and that some modifications could, perhaps be expected to be required of a model developed in a vastly different climatic regime.

Because of difficulties encountered in model implementation, it was not felt to be feasible to employ the model in a forecast mode. Although conceptual simulation models have some potential advantages in forecasting seasonal flow volumes, a number of additional issues must be addressed before their general implementation can be recommended. First, is the question of how much data and what minimum quality of data are necessary to justify the additional effort needed to implement a conceptual simulation model as opposed to the simpler Tangborn style storage accounting model discussed in Chapters 4 and 5. Second is the question of model calibration and the objective function used to optimize parameters. The objective function most commonly used in parameter optimization is the mean square error of differences between simulated and recorded streamflow (or, in the case of the snow module, simulated and recorded snowpack observations). It is not clear that the resulting parameters will yield the most accurate streamflow forecasts; perhaps alternate calibration approaches dependent on intended use of a model might be entertained.

CHAPTER 4 WATER STORAGE ACCOUNTING MODEL DESCRIPTION

Difficulties in measurement of high altitude precipitation, particularly when occurring as snow, have long been recognized. The most significant source of measurement error in use of point precipitation as an indicator of mean areal precipitation appears to be catch deficiency caused by wind (Larson and Peck, 1974). Although snow water equivalent measurements are less affected by instantaneous wind effects, snow redistribution may have a substantial effect on these measures, which consequently may not provide a good indication of mean areal snow water storage. As a result, low altitude precipitation gages, which tend to be less affected by wind, often provide a better index of high altitude mean areal precipitation than do high altitude gages themselves.

Tangborn (1977) has proposed a hydrometeorological (HM) model for forecasting streamflow runoff volumes which utilizes only low altitude precipitation data to estimate a basin water balance from which summer runoff is forecasted. In several cases, this model has apparently achieved greater forecast accuracy than the more commonly used regression models, which use snow cover measurements, and in some instances, soil moisture and winter precipitation as index variables for summer runoff. However, the HM model does not incorporate winter snowpack measurements as an estimator of basin water storage, so the question remains as to whether an additional improvement in forecast accuracy might be achieved by using a basin water storage estimate based on both storage-accounting considerations and snow course observations. This chapter describes an extension of the storage balance model to incorporate snow course data.

4.1 Model Formulation

The HM model is based on a water balance for both a summer prediction and test season;

$$R_t^* \simeq C_w P_w - R_w + B_t \qquad 4\text{-}1$$

$$R_s^* \simeq C_{w+t} P_{w+t} - R_{w+t} + B_s \qquad 4\text{-}2$$

where the subscripts w, t, and s denote winter, test, and summer, respectively, and these seasons are contiguous. The subscript w+t denotes the combined winter and test season. The coefficients C and B are determined by regressing cumulative runoff to the end of the forecast (test or summer) season on winter or winter plus test season precipitation. A single precipitation sequence is used, however this sequence may be a composite of several observed station records. The summer prediciton R_s^* is corrected by a multiple k of the test season error E_t, where

$$E_t = R_t^* - R_t \qquad 4\text{-}3$$

$$E_s = R_s^* - R_s \qquad 4\text{-}4$$

from which $E_s \simeq kE_t$, $\quad k = \sum_{i=1}^{n} E_s E_t / \sum_{i=1}^{n} E_t^2$

with R_t and R_s denoting recorded test and summer season runoff over the n year calibration record.

The corrected forecast is then

$$R_s^{**} = R_s^* + kE_t \qquad 4\text{-}5$$

The success of the test season correction depends on the degree to which test season and summer forecast errors are correlated.

Several variations of the forecast given by equations 4-1 - 4-5 have been suggested by Tangborn (1977), for instance test season precipitation may be included in equation 4-1, and an error intercept as well as slope may be used in the test season correction. However, it is not apparent that these changes appreciably affect forecast accuracy and the form given by equations 4-1 - 4-5 was retained through this investigation.

The modifications made to the model to allow incorporation of snow course data were as follows. Initially, the coefficients C_w, C_{w+t}, B_t, and B_s were estimated for the unaltered HM model. N_s snow measurement dates (and M_s snow courses) were considered. The initial storage correction was computed as

$$S_{w+t}^{(1^-)} = C_{w+t}P_1 - R_1 \tag{4-6}$$

where the superscript 1^- denotes an estimate or measurement immediately before the snow course measurement and P_1 and R_1 are the cumulative precipitation and runoff to the snow measurement date. The updated storage estimate is

$$S_{w+t}^{(1)} = \sum_{i=1}^{M_s^{(1)}} W_i^{(1)} X_i^{(1)} + W_o^{(1)} \tag{4-6a}$$

where

$$\begin{aligned} X_1 &= S_{w+t}^{(1^-)} \\ X_2 &= Y_{S_1}^{(1)} \\ &\vdots \\ X_{ks+1} &= Y_{S_{M_s}}^{(1)} \end{aligned} \tag{4-7}$$

The $Y_{S_i}^{(1)}$ are the M_s snow water equivalent measurements at the first snow measurement date. X_i are the principal components of X, and the $W_i^{(1)}$ are determined by stepwise regression on the principal components. $M_s^{(1)}$ is the number of significant coefficients determined by the stepwise regression. The principal components transformation is performed to avoid problems of collinearity encountered because of the high correlations between the X_i.

The procedure continues through the N_s snow measurements, e.g.,

$$S_{w+t}^{(2^-)} = S_{w+t}^{(1)} + C_{w+t}P_{12} - R_{12} \tag{4-8}$$

where the subscript 12 denotes precipitation or runoff in the interval from snow measurement date 1 to snow measurement date 2. The updating continues through snow measurement N_s, from which the forecast storage estimate

$$S_{w+t}^{*} = S_{w+t}^{(N_s)} + C_{w+t}P_{N_sT_1} - R_{N_sT_1} \tag{4-9}$$

is made, where T_1 is the forecast date. An identical procedure is used to compute S_w^*, the storage estimate at the beginning of the test season. Forecasts are then made by

$$R_t' = S_w^* + B_t \qquad 4\text{-1a}$$

$$R_s' = S_{w+t}^* + B_s \qquad 4\text{-2a}$$

The test error coefficient is computed using equations 4-3, 4-4, and 4-5 with R_t' and R_s' substituted for R_t^* and R_s^*, respectively.

Initially, a principal components approach was also used to allow incorporation of multiple precipitation stations rather than a single composite record. However, it was found that although this procedure reduced calibration errors (as measured by the root mean square "forecast" error), forecast error increased and in several cases the root mean square error approached the unconditional forecast period standard deviation, indicating that the model provided no better a forecast than the process mean. The problem most likely is attributable to overfitting, i.e., the necessity to estimate an excessive number of model parameters. Similar problems are possible in use of the snow course data. Although the stepwise regression selects only significant variables, all snow courses enter into the principal components transformation, so even though not all the principal components may be used in estimating the storage update, all the original information (as obscured by measurement error) is incorporated. Consequently, it is important that economy be achieved in the number of stations M_s entering the model. Likewise, in order to obtain the most informative composite precipitation record, it is necessary to screen the candidate stations. The method used to perform the screening is described in the following section.

4.2 Screening Model

The screening model used was a simple stepwise regression of cumulative runoff (winter through summer) on winter precipitation or snow course observations.

Candidate precipitation and snow course stations were obtained by manual screening of all snow courses lying in or near the basin of interest and of all precipitation stations lying within approximately a 125 km radius of the river forecast station. Although a greater number of candidate stations were reviewed, only a few met the additional requirement that a fairly complete record for the period 1949-75 be available. Figure 7 shows the location of the precipitation, snow course, and river forecast stations for three Washington river basins used in the study. Also shown is the crest of the Cascade range, which has a substantial effect on precipitation patterns; generally basins with comparable mean topographic elevation lying west of the crest receive greater precipitation than east slope basins, while east slope basins receive a larger proportion of their precipitation as snow.

Missing observations for the candidate stations were estimated as follows. For precipitation stations, monthly station correlation matrices were estimated, and the station having the highest correlation with the base station was used to fill in missing observations. The missing data were then estimated as the (daily) observation at the estimating station scaled by the ratio of the cumulative monthly precipitation at the estimating station to the historic monthly mean at the base station. A similar procedure was used for snow course data. Although more refined procedures might be used, the number of missing data was generally small, and the large amount of data to be handled required use of a fairly simple procedure. A more detailed description of the data filling approach for precipitation and snow course data is given in Appendices B and C, respectively.

A very liberal entry significance level of 0.50 was used for all variables in order to allow a full review of variable significance as a function of forecast season. Estimated significance levels for precipitation are generally

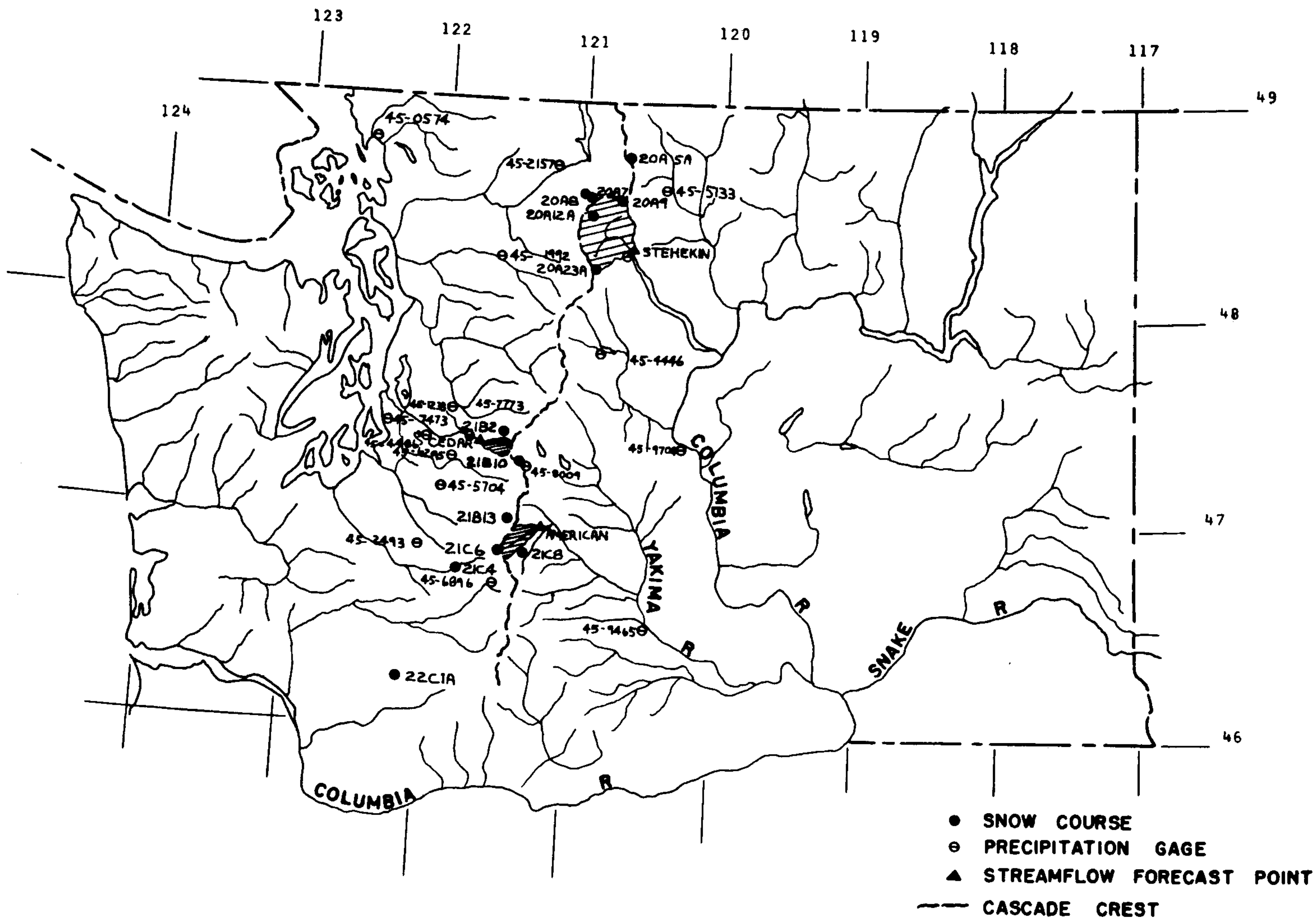

Figure 7. Location of Streamflow and Precipitation Gages and Snow Courses Used in Study.

lower than for snow course observations, apparently reflecting smaller measurement error and spatial variability in the data, which is also reflected in higher station correlations. Consequently, the precipitation data are more affected by collinearity than are the snow course measurements. As a result, precipitation stations were more rigorously screened than were snow courses.

4.3 Construction of Composite Precipitation and Snow Course Records

In cases where more than one precipitation or snow course record was selected by the screening model, an approach to incorporating the multiple records was needed. Although the records could be treated as separate inputs to the model, this approach requires estimation of an additional model parameter for each record, and as indicated in section 4.2, the result was generally a decrease in forecast accuracy when multiple precipitation stations were used. Similar results were subsequently found when multiple snow courses were used. Consequently an approach was needed to combine the information from the multiple records into a single composite record. Simple addition of observations for each date is undesirable because it ignores the station preference order determined by the screening model, and because substantial differences in station means can result in domination of the composite record by a single station.

Consequently, a nonparametric weighting scheme was devised. The weighting function is

$$W_j = (N-j+1) / \sum_{j=1}^{N} j \quad ,$$

where j is the preference order and N is the number of precipitation or snow course stations (M_p or M_s) passing the screening test. For three stations, for instance, the weights are 1/2, 1/3, and 1/6. In order to compensate for large variations in precipitation volumes a weighted precipitation annual mean was determined using station annual means weighted by the function given above. Daily precipitation volumes at each station were then scaled by the ratio of the weighted mean to the station mean. The composite record was subsequently formed by summing the adjusted precipitation at each station weighted by the W_j. The same procedure was used for snow course measurements, except that the station mean snow water equivalent on the given measurement date was used in place of annual precipitation volumes for weighting station means.

CHAPTER 5 STORAGE ACCOUNTING MODEL RESULTS AND DATA WORTH

The storage accounting model described in Chapter 4 is a simple approach which incorporates only the most rudimentary conceptual knowledge of the physical system. It is based in large part on the time-honored engineering criterion that "it works". It has the further advantage that it is easily implemented. Although conceptual simulation models such as the Sacramento model discussed in Chapters 2 and 3 are more physically realistic, transferability is clearly still a substantial problem with these models, and they are currently used for operational forecasting only on a limited basis. In fact, application of the Sacramento model to several Sierra Nevada basins represents one of the few current uses of such models in operational forecasting. The great majority of operational forecasts (for instance, the forecasts made throughout the West by the Soil Conservation Service) are made using flow index methods, which represent no physical knowledge of the system except that the amount of winter snowfall and/or valley precipitation affects subsequent summer runoff. Despite their simplicity, these methods have proven to be quite accurate in many cases. However, the storage accounting approach appears to offer an improvement in accuracy over these flow index methods while retaining their simplicity.

Although the potential forecast accuracy of conceptual models is unknown, even if it should exceed that of the storage balance approach it is quite likely that cases will remain where ease of model implementation will outweigh forecast accuracy improvements, particularly if they are modest. Insofar as the storage accounting approach is thought to represent the more accurate of the simple approaches, the worth of data input to the model should be of direct concern in choosing the form of the model to be used and in planning of data acquisition activities.

5.1 Worth of Snow Course Data

The alternate model formulation discussed above provides a basis for estimating the worth of snow course data to the storage balance accounting model in seasonal flow volume forecasting. Data worth is clearly dependent on the model formulation. The modified HM model probably represents about the maximum amount of refinement possible in a forecast model with parameters estimated by regression in the absence of much longer record lengths than are currently available.

Snow course data worth was assessed using a case study approach. Three variations of the storage accounting model described in Chapter 4 were applied to the Cedar, Stehekin, and American Rivers, Washington. The location of these basins and precipitation gages and snow courses entering the screening model are shown in Figure 7. Table 4 summarizes some physical parameters for the three basins.

For each of the three basins, forecasts were made using three variations of the HM model, each using identical composite precipitation and, where relevant, snow water equivalent input. The three models were 1) the unaltered HM model, which uses no snow course data; 2) the modified HM model described in section 4.1; and 3) the modified HM model as used in 2), with the exception that the vector X in equation 4-7 is a scalar containing only the (single) observation of the composite snow water equivalent record. October 1 was taken as the beginning of the winter season throughout the comparisons, and a 13-day test season was likewise used throughout.

5.1.1 Split Sample Approach for Assessing Forecast Accuracy

A procedure known as split sample testing is often used in calibration and verification of deterministic watershed models. A modification of this approach has been employed by Tangborn (1977) for assessment of the HM model.

Table 4. Summary of Physical Characteristics for Cedar, Stehekin, and American River Basins[a]

	Cedar River near Cedar Falls	Stehekin River at Stehekin	American River near Nile
USGS gage number	12-1150	12-4510	12-4885
gage elevation, ft	1560	1098	2700
drainage area, $mile^2$	40.7	344	78.9
average annual snowfall, inches	440	290	350
average annual precipitation, inches	120	99	74
average topographic slope, ft/mile	116	137	64
topographic mean elevation, ft	3230	5130	4860
forested area, per cent	77	83	91
average annual flow, cfs	273	1426	246

[a] information from U.S. Geological Survey Basin Characteristics File

In this approach model parameters are estimated for an initial calibration sequence of model inputs and outputs. Forecasts are then made using this set of parameters for a subsequent forecast sequence. The procedure is iterated in a stepwise manner; once the forecasts have been made for a given step, the calibration sequence is extended to include the forecast period of the previous step and forecasts are made for the next forecast sequence. The most realistic forecast sequence length is one year, since in practice information from all preceeding years of record is available for estimation of model parameters for the present year's forecasts. In this work, however, logistical constraints precluded the amount of computation necessary to perform yearly updating, and a compromise forecast sequence length of three years was used in making model comparisons.

5.1.2 Summary Statistics for Assessing Model Performance

For each model, forecasts were made using the split sample approach described in the introduction. From the eighteen forecasts obtained for each model, forecast data, and summer end date combination, a root mean square error was estimated as the square root of the average of the squares of differences between forecasted and observed runoff. A summary statistic which relates forecast accuracy to estimated variability of the time series of forecast period flow volumes is the coefficient of prediction;

$$C_p = 1 - \left[\frac{\hat{\sigma}_1}{\hat{\sigma}}\right]^2$$

where $\hat{\sigma}_1$ is the root mean square error and $\hat{\sigma}$ is the estimate of the standard deviation of recorded flow volumes during the forecast period. The coefficient of prediction nominally ranges from zero to one, although it is possible for a very poor forecast model to obtain small negative values as a result of sampling variability is estimation of $\hat{\sigma}_1$ and $\hat{\sigma}$. A coefficient of prediction of zero indicates that forecast accuracy is no better than that which would be obtained by forecasting with the sample mean, while a coefficient of prediction of one indicates a perfect forecasting technique. The nonlinearity of the index with $\hat{\sigma}_1$ served to inflate forecast accuracy above that which would be indicated by a linear function; for instance a model which obtains forecasts with a root mean square error equal to one-half of the estimated forecast period standard deviation yields a C_p of 0.75.

Clearly, other measures of forecast accuracy might be used, for instance forecast accuracy in low flow years might be of greater economic importance than accuracy in moderate or high flow years. However, the limited record lengths commonly available result in sufficient variability in estimation of an "average" statistic such as C_p to discourage attempts at estimating model performance indices related to tail behavior of the distribution of model errors.

In addition to the coefficient of prediction, model calibration quality was summarized by a coefficient of calibration, denoted C_c, which was estimated identically to the coefficient of prediction except that the root mean square of calibration errors for the final calibration period (1949-72, used in the 1973-75 forecasts) was used in place of the root mean square of forecast errors. As discussed earlier, statistics based on calibration errors are not valid measures of forecast accuracy. The ratio of the coefficient of prediction to the coefficient of calibration does, however, provide a measure of how well the model is calibrated. If the model were an exact description of the physical system and the parameters were known, C_c and C_p would be nearly identical. However, where model parameters must be estimated, C_c should exceed C_p. Increasing the number of model parameters will always increase C_c so long as an efficient parameter estimation procedure is used, but C_p will ultimately decrease with additional parameters as the model is "overfit".

The final model performance measure estimated was the correlation between test season and forecast season errors. The test season/forecast season error correlation gives an index to the value of the test season in correcting the forecasts. The correlation coefficient estimated here is based on the sequence of eighteen (1956-75) test and (uncorrected) forecast period errors, rather than the error sequences on which estimates of the correction coefficients were based, hence it is an aposteriori, rather than an operational measure of the potential success of the test season correction in improving forecast accuracy.

5.1.3 Screening Model Results for Cedar, Stehekin, and American Rivers

The screening model described in section 4.2 was used to determine which precipitation and snow course stations to use in the forecast model. The results obtained from the screening model are given in Tables 5 and 6. Station location is shown in Figure 7 for forecasts made from the indicated date through July 31. Screening was also performed for a summer season ending on September 30, however the results except in a few cases noted below, were quite similar to those obtained for the July 31 summer end so are not included here.

Generally, station preference was easier to establish for the precipitation stations than for the snow courses. The Cedar was, however, an exception, and the use of Snoqualmie Falls and Olallie Meadows as the sole precipitation station and snow course was an obvious choice. Three precipitation stations, in the preference order Darrington, Lake Wenatchee, and Mazama were selected for entry into the Stehekin forecast model. Use of Mazama was based largely on its relatively low estimated significance levels for September 30 summer end forecasts, not shown here. Selection of snow courses was more difficult, since no station was significant at even the 0.50 level for all three forecast dates. Harts Pass was selected as the single station, in part because its relatively

Table 5. Significance Levels Estimated by Screening Model for Precipitation Candidates

CEDAR RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
45-7773	Snoqualmie Falls	.001	.002	.001
45-1233	Cedar Lake	----	.221	.124
45-4486	Landsburg	.191	.491	.420
45-6295	Palmer	----	----	----
45-7473	Seattle-Tacoma Airport	----	----	----
45-8009	Stampede Pass	----	----	----

STEHEKIN RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
45-0574	Bellingham Airport	----	----	----
45-1992	Darrington Ranger Station	.002	.007	.020
45-2157	Diablo Dam	----	----	.366
45-4446	Lake Wenatchee	.087	.050	.061
45-5133	Mazama	.125	.243	.079
45-8059	Stehekin	.445	.337	----
45-9074	Wenatchee	.358	.313	----

AMERICAN RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
45-9074	Wenatchee	----	----	.237
45-2493	Electron	----	----	----
45-4704	Mud Mountain Dam	.302	.354	----
45-6896	Rainier Ohanapecosh	.000	.001	.000
45-9465	Yakima	.026	.018	.004

low significance level for the February 15 and April 15 forecasts cast doubt on the lack of significance for the interim forecast date. Although the preference between Harts Pass and Park Creek Ridge is not clear from Table 5, Harts Pass performed slightly better for the September 30 summer end forecasts. The selection of Rainier Ohanapecosh and Yakima for entry into the American River forecast model was fairly straightforward. Again, snow course selection was more difficult. Although Corral Pass was the preferred station for the March 15 and April 15 forecasts, no record existed for use with February 15 forecasts. Bumping Lake was a

Table 6. Significance Levels Estimated by Screening Model for Snow Course Candidates[1]

CEDAR RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
21B02	Olallie Meadows	.000	.000	.000
21B10	Stampede Pass	----	----	.096

STEHEKIN RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
20A05A	Harts Pass	.106	----	.001
20A09	Rainy Pass	----	.000	.249
20A23A	Lyman Lake	----	----	----
20A12A	Park Creek Ridge	.066	----	.068
20A08	Meadow Cabins	----	----	.409
20A07	Thunder Basin	----	.231	----

AMERICAN RIVER CANDIDATES

Station/Forecast Date		Feb. 15	Mar. 15	Apr. 15
21C06	Cayuse Pass	----	.193	.050
21C08	Bumping Lake	.068	.338	.004
22C01A	Plains of Abraham	----	.398	----
21C04	Ghost Forest	.007	----	.305
21B13	Corral Pass	----	.047	.084

[1]nominal snow course measurement date was fifteen days prior to forecast date.

fairly strong candidate, especially for the April 15 forecasts, consequently this station was used alone as the February 15 station, and Corral Pass and Bumping Lake were used to form a composite record for subsequent forecasts.

5.1.4 Estimated Snow Course Data Worth for Cedar, Stehekin, and American Rivers

Figures 8 and 9 show the estimated coefficients of prediction for the Cedar, Stehekin, and American River forecast models. Use of the snow course data substantially improves forecast accuracy for the Cedar forecasts. The dip in forecast accuracy for March 1, April 1, and May 1 forecasts is probably related to the timing of the snow course observations. For instance, the March 1 forecast uses a basin storage estimate based on a February 1 snow course observation updated to February 15 with a subsequent 13-day test season. However, low estimated test season/forecast error correlation for the models incorporating snow course observations (discussed below) indicates that better forecast accuracy could possibly be achieved without use of the test season, which would allow use of the March 1 snow course data in the March 1 forecast. The expected forecast accuracy improvement would probably also remove the jagged appearance of the two upper curves in Figures 8a. Figures 8b and 9b show a reverse of the relative accuracy for the three models from that observed for the Cedar. The highest accuracy estimates are achieved here by the unaltered HM model; the inclusion of the snow course data in the forecasts results in a reduction in forecast accuracy, especially for late winter and early spring forecasts. For the American River forecasts (Figures 8c and 9c) estimated forecast accuracy for the three models is nearly identical, except that the forecasts without snow data appear to have slightly improved accuracy for late winter forecasts.

Figures 10a-c show the estimated ratio of the coefficient of prediction to the coefficient of calibration for the Cedar, Stehekin, and American Rivers forecasts. For the Cedar (Figure 10a) the relative performance is similar to that observed in Figures 8a and 9a for the coefficient of prediction, except that the "valleys" associated with March 1, April 1, and May 1 forecasts are more evident. The general trend of improved forecast accuracy with improved calibration is also reflected in the performance of the Stehekin River forecast models (Figure 10b) where the model without snow approaches ideal calibration/forecast error compatibility for late season forecasts. The results for the American River (Figure 10c) however, demonstrate that calibration/forecast error compatibility is not necessarily an index to forecast accuracy. In this case, a clear preference for the no snow model exists on the basis of the C_p/C_c ratio (Figure 10 c), however less distinction is evident on the basis of C_p alone (Figures 8c and 9c). This may simply reflect the tendency for the coefficient of prediction to level off and the decline as non-informative parameters are added, while the coefficient of calibration continues to increase. As the coefficient of prediction approaches its maximum the coefficient of calibration is increasing at a much faster rate; the models with snow storage correction for the American River forecasts apparently represent a case where the additional information obtained from the snow course observations just compensates for the loss of accuracy resulting from increased parameter variability.

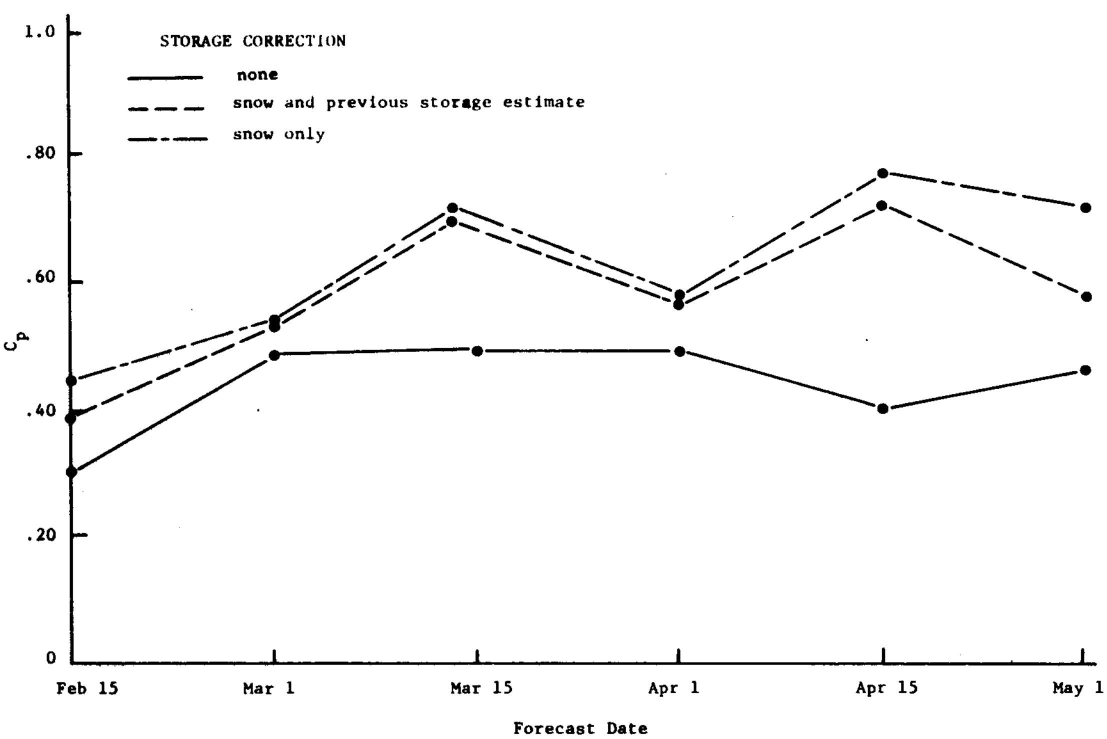

Figure 8a. Estimated Coefficient of Prediction for Cedar River Forecasts for Forecast Period End July 31.

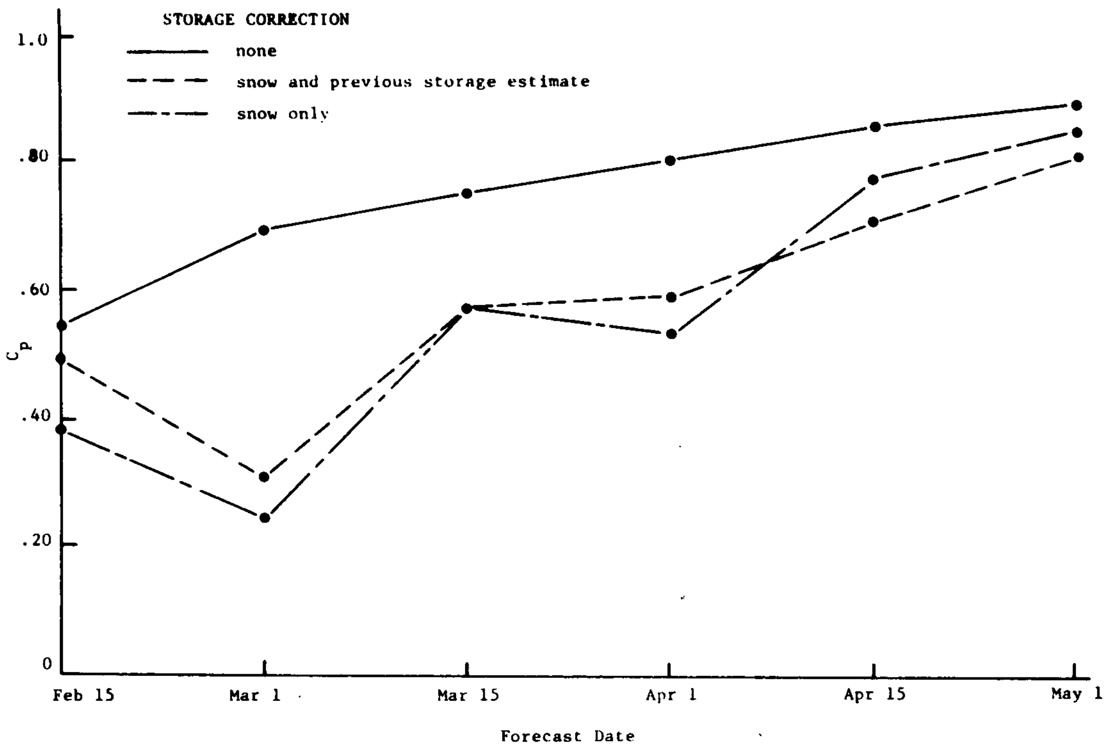

Figure 8b. Estimated Coefficient of Prediction for Stehekin River Forecasts for Forecast Period End July 31.

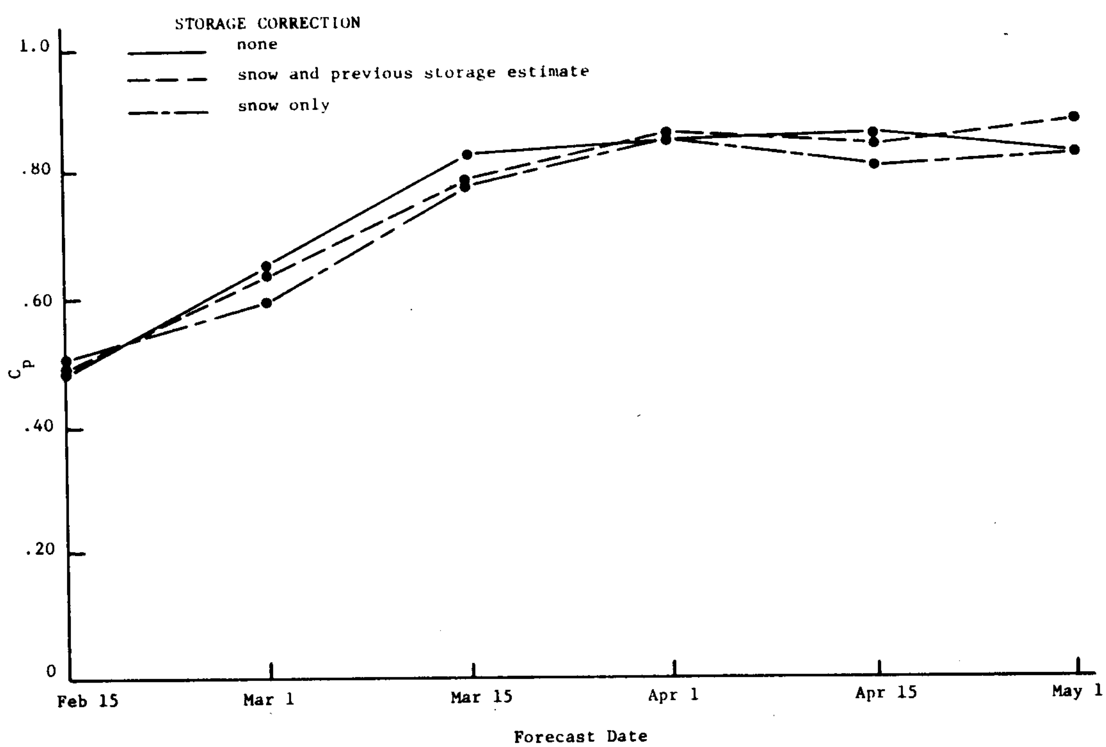

Figure 8c. Estimated Coefficient of Prediction for American River Forecasts for Forecast Period End July 31.

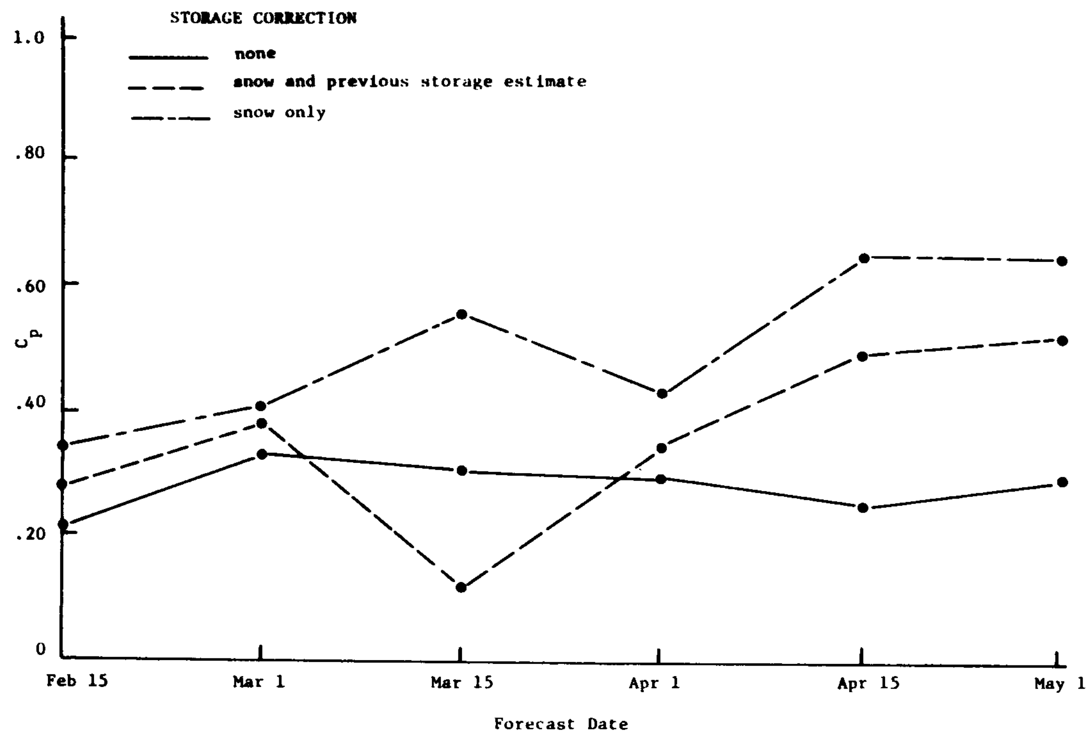

Figure 9a. Estimated Coefficient of Prediction for Cedar River Forecasts for Forecast Period End September 30.

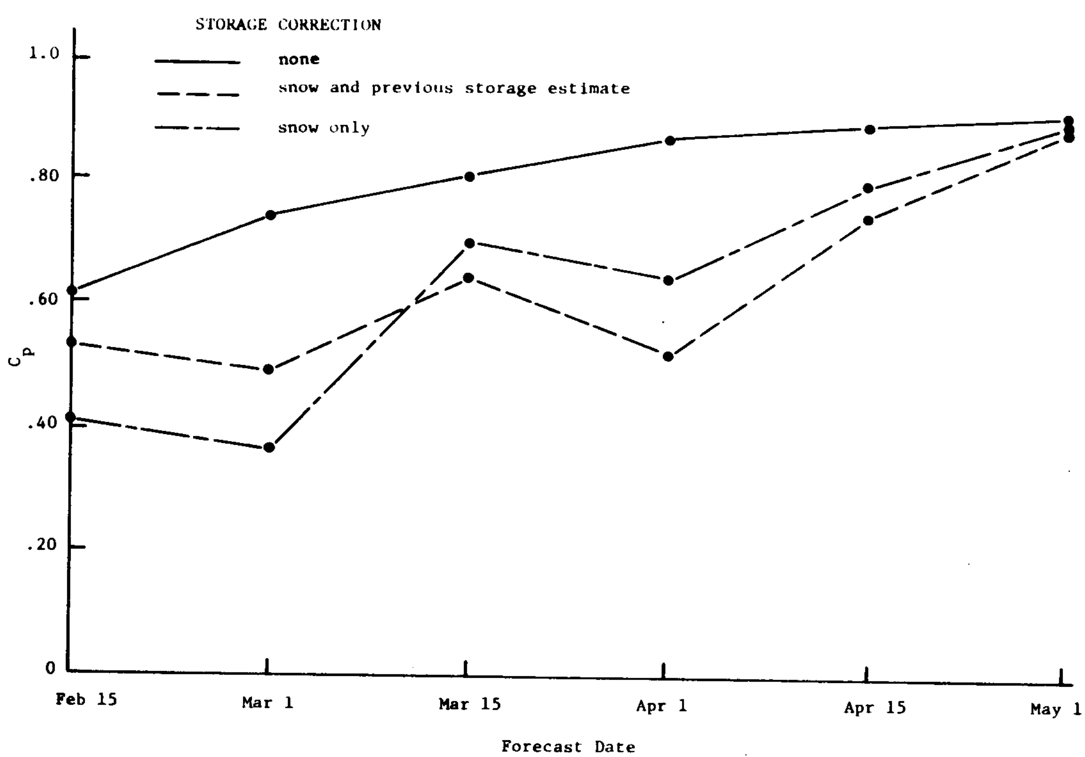

Figure 9b. Estimated Coefficient of Prediction for Stehekin River Forecasts for Forecast Period End September 30.

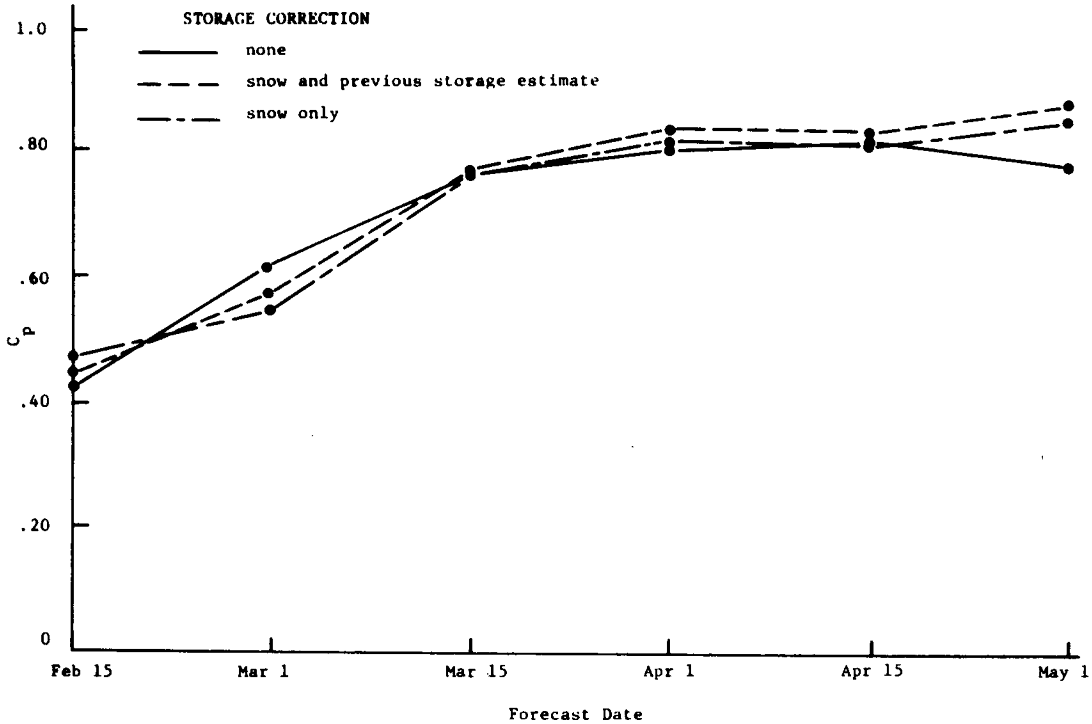

Figure 9c. Estimated Coefficient of Prediction for American River Forecasts for Forecast Period End September 30.

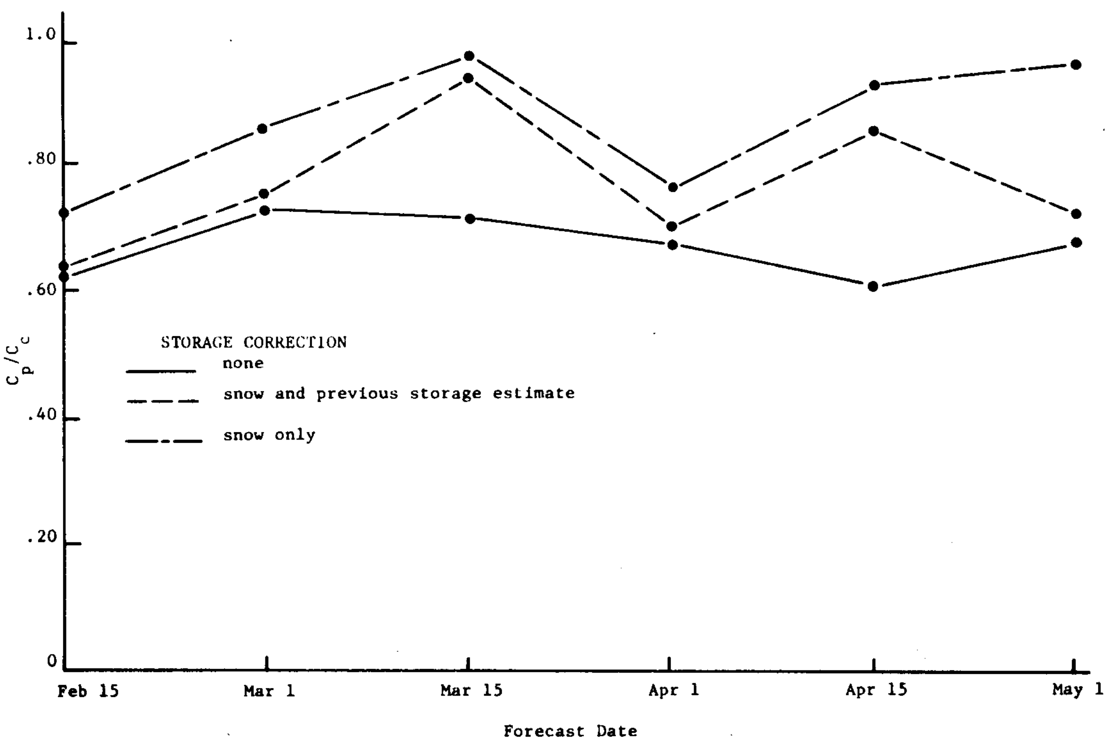

Figure 10a. Ratio of Estimated Coefficient of Prediction to Coefficient of Calibration for Cedar River Forecasts for Forecast Period End July 31.

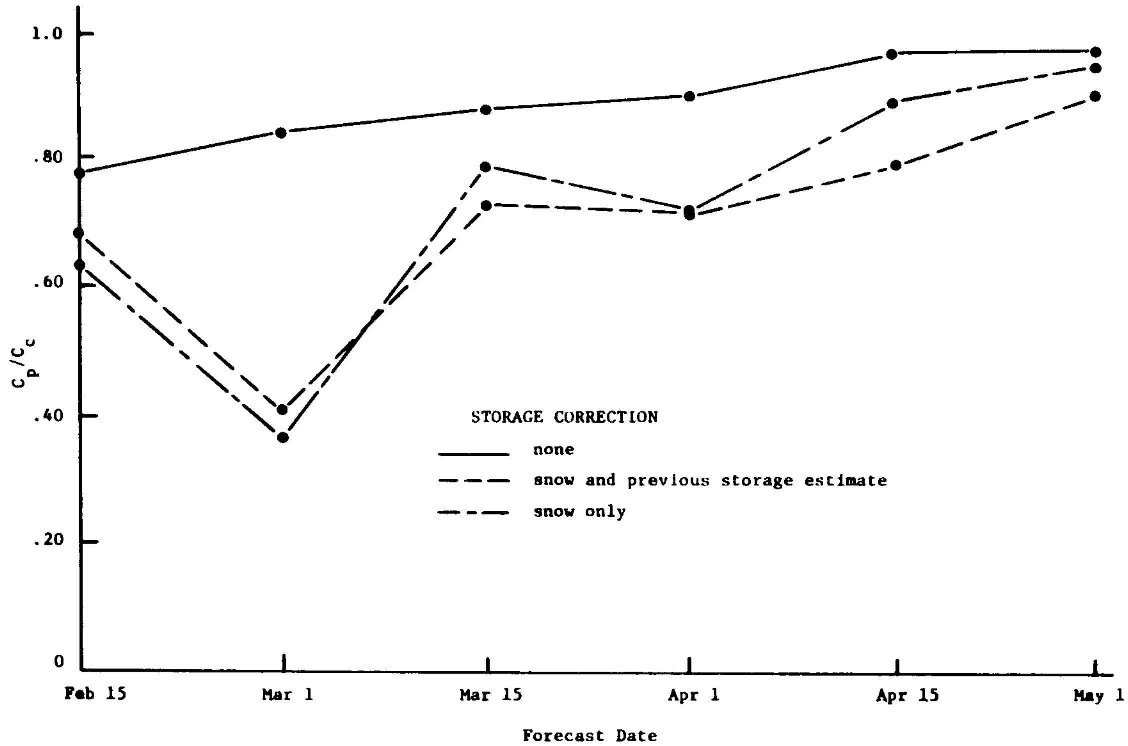

Figure 10b. Ratio of Estimated Coefficient of Prediction to Coefficient of Calibration for Stehekin River Forecasts for Forecast Period End July 31.

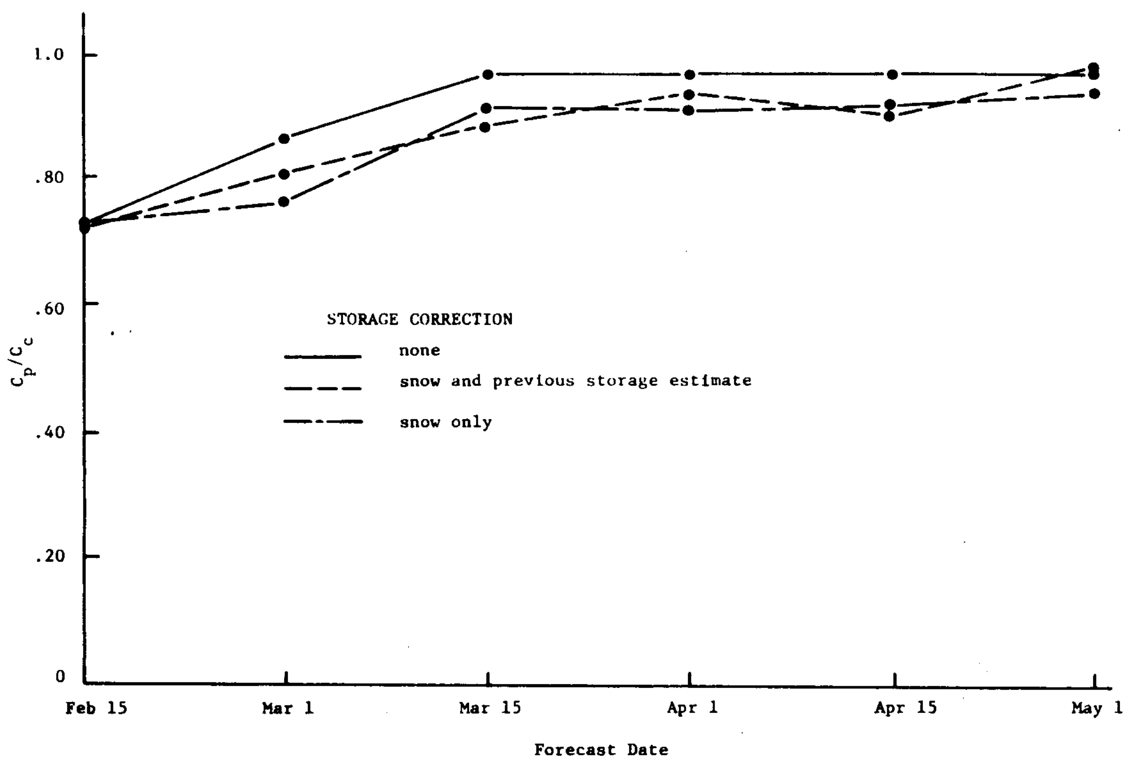

Figure 10c. Ratio of Estimated Coefficient of Prediction to Coefficient of Calibration for American River Forecasts for Forecast Period End July 31.

Figure 11 shows the estimated test error/forecast error correlation for the Cedar River forecasted runoff volumes through July 31. The unmodified HM model shows much higher correlations than either of the models which include snow course data. One interpretation of this result is that the models which incorporate snow course data are more successful in extracting information from the model input data than is the HM model without snow course data. The test error/forecast error correlations are so low for the models with snow course data that use of a test season appears unjustified. However the HM model (without snow) is somewhat sensitive to the length of the test season, which was not adjusted here. Thus, for comparison purposes the constant test season length used here may be justified. The general characteristics described above were also observed for the Stehekin and American River forecasts with summer period end July 31, and for all three river forecasts for summer season end September 30.

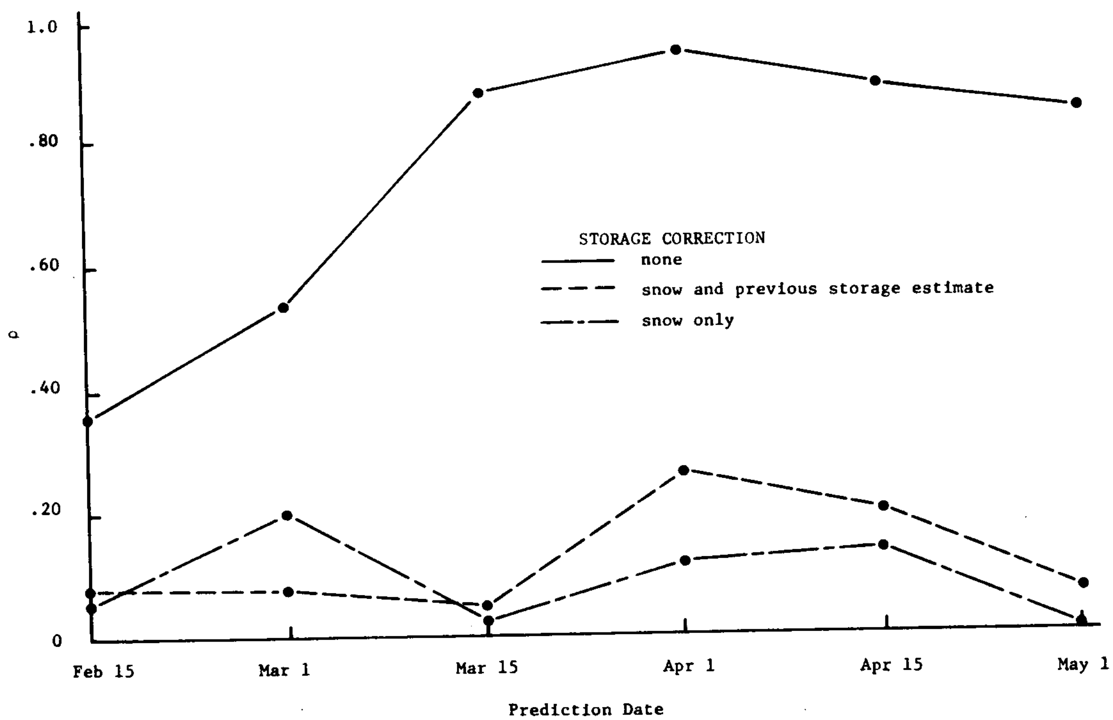

Figure 11. Test Season Error/Forecast Season Error for Cedar River Forecasts for Forecast Period End July 31.

It appears that the effects described may be somewhat general, unlike the behavior of C_p and C_c which appears to be unique for each basin.

5.2 Estimated Worth of Precipitation Forecasts

Although derivation of the worth of a precipitation forecast to seasonal flow volume forecast accuracy using the model described in Chapter 4 would, at best, be a tedius process if any general precipitation forecast accuracy were considered, the worth of a perfect precipitation forecast may be assessed relatively easily. This may be accomplished by simply substituting the cumulative winter and summer precipitation in equation 4-2 in place of the winter precipitation. Generally, the test season is of no help when this change is made, so R_s^* represents the forecast of runoff using perfect knowledge of the forecast period precipitation. Since the results given in section 5.1 effectively represent the case of no precipitation forecast, bounds may be determined representing the range from a forecast containing no information beyond that present in the historic record (e.g., a forecast with variance equal to the population variance) to a forecast with zero variance.

Estimates of the coefficient of prediction for both cases were obtained for the three streams used in the earlier experiments. In each case, the "no precipitation forecast" case was that yielding the most accurate operational forecast; for the Cedar River this represented the case where a storage correction utilizing snow cover data only was used, while for the Stehekin and American Rivers no storage correction was employed. In each case, the same model was used to obtain the "perfect precipitation forecast" seasonal runoff forecasts.

The results of these comparisons are shown in Figures 12a-c for runoff forecasts through July 31, and in Figures 13a-c for forecasts through September 30. As expected, the seasonal precipitation forecasts are most valuable for

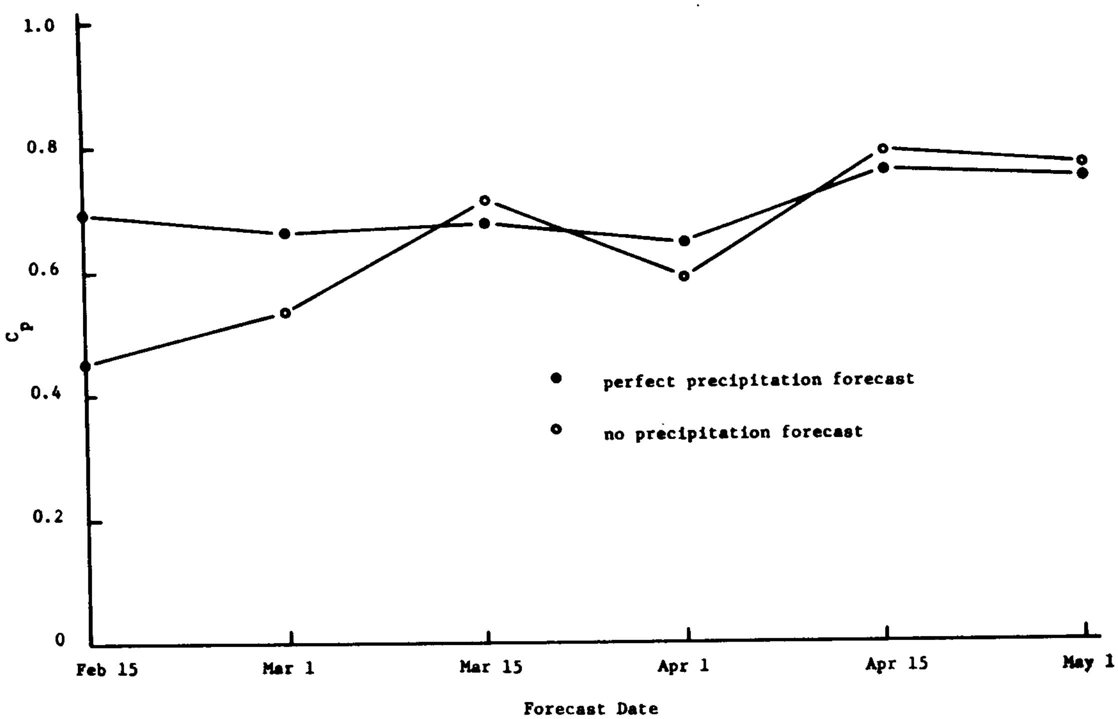

Figure 12a. Comparison of Accuracy of Best Cedar River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End July 31.

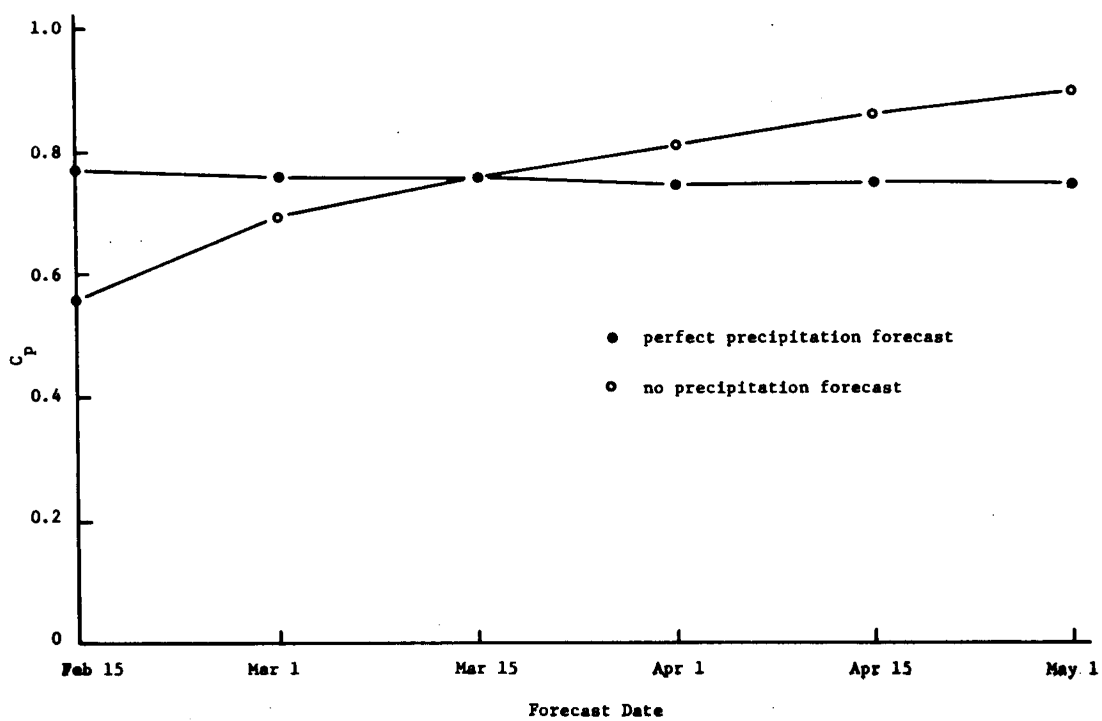

Figure 12b. Comparison of Accuracy of Best Stehekin River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End July 31.

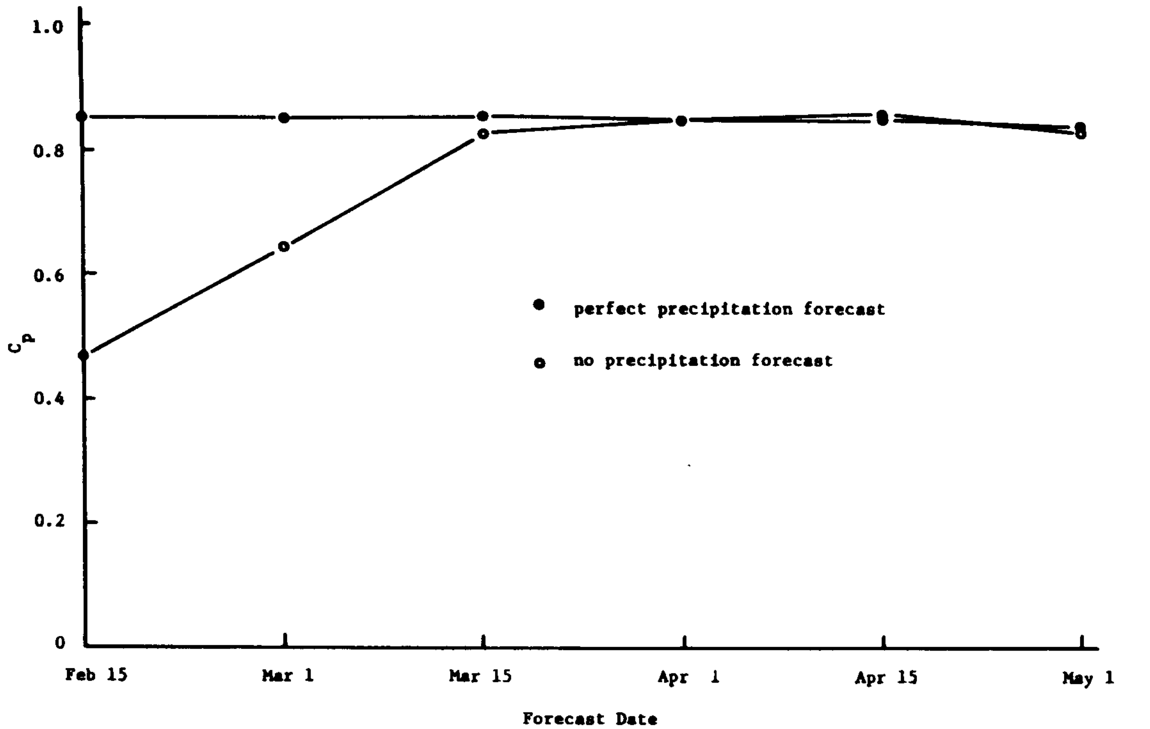

Figure 12c. Comparison of Accuracy of Best American River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End July 31.

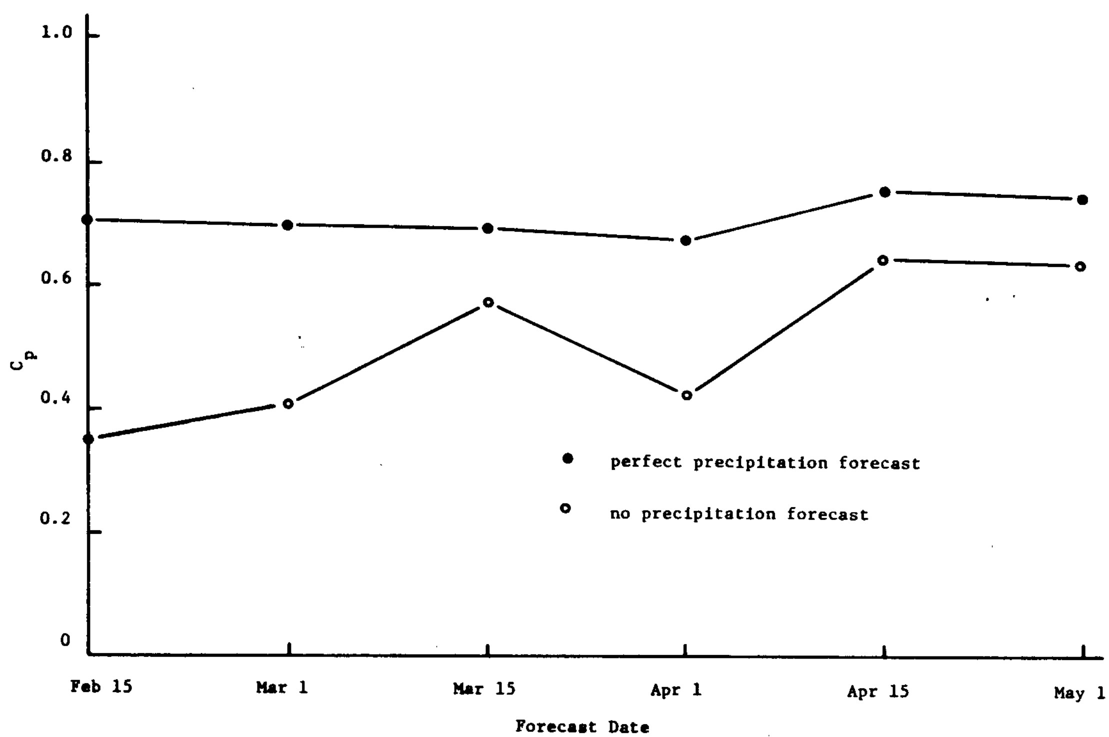

Figure 13a. Comparison of Accuracy of Best Cedar River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End September 30.

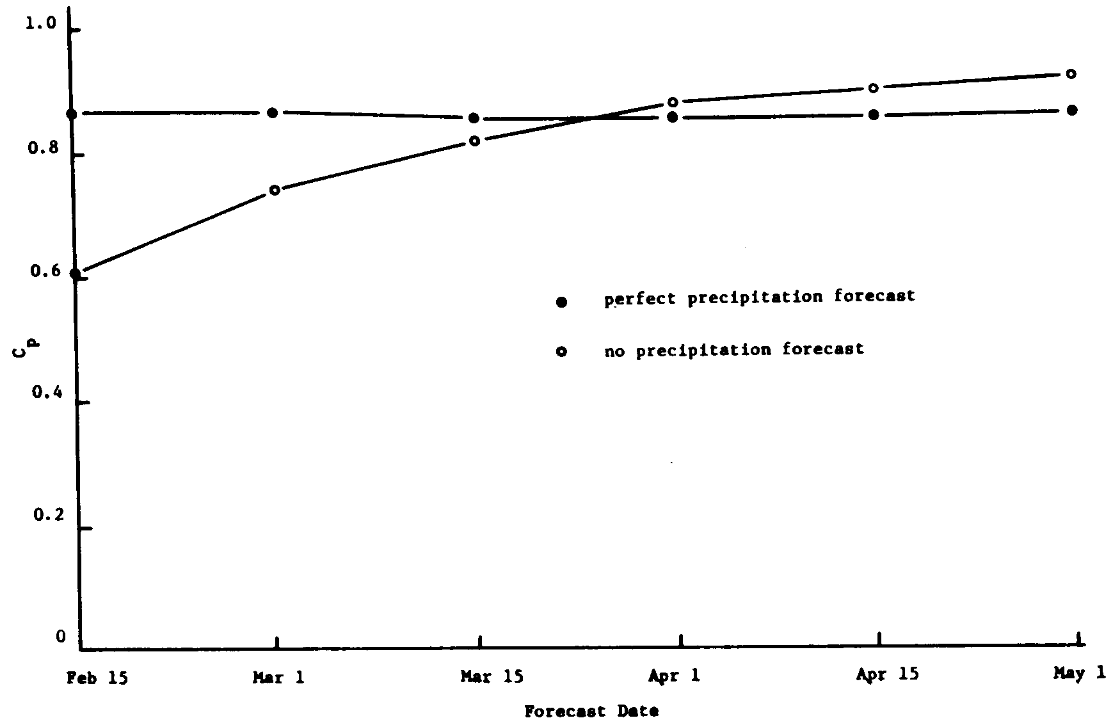

Figure 13b. Comparison of Accuracy of Best Stehekin River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End September 30.

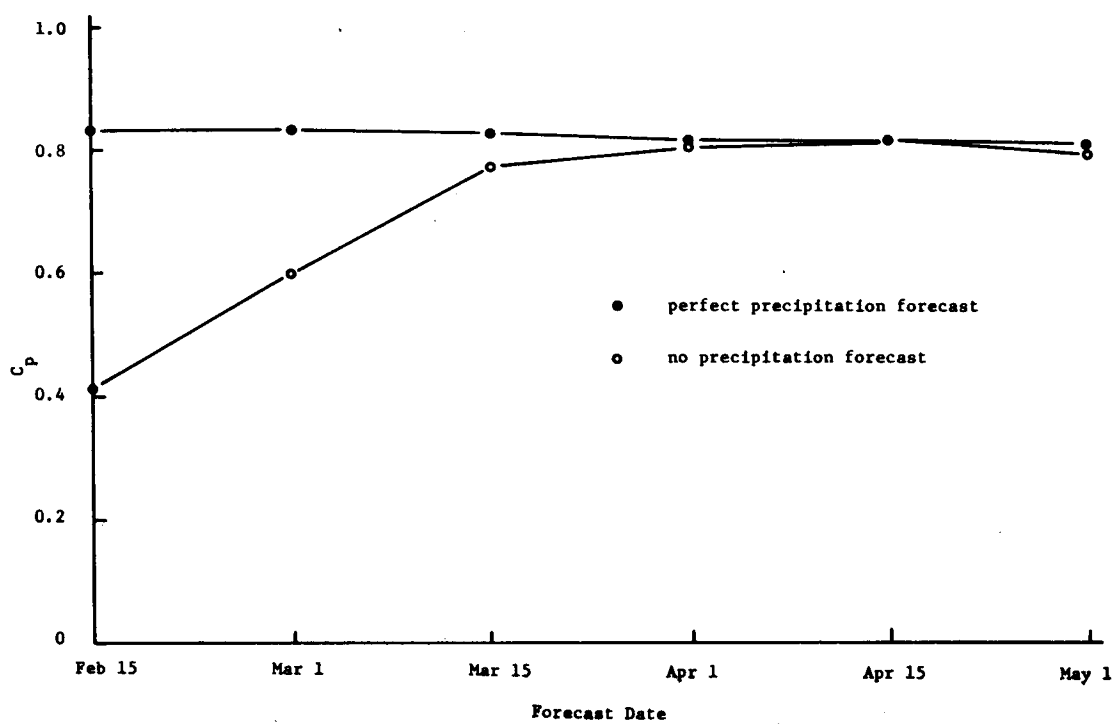

Figure 13c. Comparison of Accuracy of Best American River Operational Forecast and Forecast with Perfect Knowledge of Summer Precipitation for Forecast Period End September 30.

early season rnnoff forecasts, since a substantial part of the total winter snowpack accumulation has not yet occurred at the time these forecasts are made. For later runoff forecasts, the precipitation forecast effectively represents a forecast of spring and summer precipitation which is likely to occur as rain, and is of much smaller magnitude than winter precipitation.

Figures 12 and 13 indicate that there are some cases, particularly for the late season Stehekin forecasts, where the precipitation forecast has negative apparent worth to the runoff forecast. This may occur because spring and summer storm patterns differ from those experienced in the winter, and the low elevation valley precipitation stations may not provide as much information regarding basin precipitation as they do in the winter. This is likely to be true particularly in the case of West Cascade precipitation gages used in forecasting east side runoff; spring and summer storms tend to be more localized and precipitation on the west side is less likely to represent conditions on the opposite side of the Cascade crest.

For the two basins where snow course data are not used in the runoff forecasts, the "perfect precipitation forecast" coefficient of prediction is essentially flat with forecast date. This results because the root mean square forecast error is constant; only the forecast period standard deviation varies. This may be seen by examining the form of the forecast equation used (assuming a single precipitation record):

$$R_{s_1} + R_{w_1} \simeq C(P_{w_1} + P_{s_1}) + B$$

$$R^*_{s_1} = C(P_{w_1} + P_{s_1}) + B - R_{w_1}$$

$$R^*_{s_2} = C(P_{w_2} + P_{s_2}) + B - R_{w_2}$$

but $P_{w_1} + P_{s_1} = P_{w_2} + P_{s_2} = P$

$$E_1 = R^*_{s_1} - R_{s_1} = kP + B - (R_{w_1} + R_{s_1})$$

$$E_2 = R^*_{s_2} - R_{s_2} = KP + B - (R_{w_2} + R_{s_2})$$

but $R_{w_1} + R_{s_1} = R_{w_2} + R_{s_2} = R$

so $E_1 = E_2 = KP + B - R$,

i.e., the forecast error is independent of the forecast date. Within the range of forecast dates and forecast period end dates used, the standard deviation of the forecast period runoff, $\hat{\sigma}_1$ decreases with forecast date, so the coefficient of prediction for the "perfect precipitation forecast" case also decreases slightly. By contrast, in the case where a snow course storage correction is made, the forecast error for the "perfect precipitation forecast" runoff forecast does vary with forecast date.

With the exception of the Cedar River forecast for forecast period end September 30 (which is much more heavily affected by rain events than the other basins) even a perfect precipitation forecast appears to have relatively little value to runoff forecasts made after early March. This result is, of course specific to the forecast model used; in particular it is expected that a conceptual model which accounts for the timing of runoff would be more responsive to accurate precipitation forecasts. In any event, however, current precipitation forecasts have only marginal accuracy beyond five days; some techniques are claimed to yield some information on a longer term basis (e.g., for several months), (see, for example, Namias, 1975), however the usefullness of these methods is a question of much debate, and in any case their accuracy is far from perfect. Consequently, with the exception of very early seasonal runoff forecast (which undoubtably could have substantial economic value) the worth of precipitation forecasts appears marginal, and other factors such as limitations on estimation of mean areal precipitation and snow cover accumulation are likely to have greater effect on runoff forecast accuracy. In the case of early season runoff forecasts, forecasting of precipitation through the end of the snow accumulation period would be of greatest value in forecasting seasonal runoff volumes; while melt period precipitation (and other meteorological variables, most importantly temperature) affect the timing of runoff, their effect on runoff volume appears less important.

CHAPTER 6 SUMMARY AND CONCLUSIONS

Two approaches to forecasting seasonal streamflow runoff volumes were investigated. The first was to have utilized a conceptual (daily) simulation model, calibrated to historic streamflow and snow course observations, to forecast basin runoff. Forecasts made using this approach would be conditioned on preceeding rainfall patterns and subsequent storage accumulation in several subsurface storage zones. The model itself consisted of two modules developed by Burnash and Baird (1975) and Burnash, et.al. (1973); a snowpack module which simulated snowpack accumulation and ablation, from which a pseudo-precipitation record was constructed, and a land module driven by the pseudo-precipitation record, which modeled subsurface water storage and basin runoff. The model was applied to the Cedar River watershed, Washington above the USGS gaging station near Cedar Falls. This basin drains a 40.7 mile^2 area of the west slope of the Cascade Mountains; the basin receives approximately 120 in/yr of precipitation with annual snowfall of over 400 inches. These extreme climatic conditions differ substantially from those of the Sierra Nevada range, California where the model had previously been used successfully in a forecasting mode. Adequate calibration could not be achieved for the model in its existing form for the Cedar River, and the model was consequently not used in a forecast mode.

Particular difficulty was encountered in adequately simulating the timing and magnitude of winter and spring runoff peaks. The difficulties were attributed in part to five general problems related to transferability of the model from the California Sierra Nevada to a lower elevation, wetter Cascade west slope drainage. The principal suspected difficulties were:

1) The model uses the saturated adiabatic lapse rate for temperature computation regardless of weather conditions, thus may improperly estimate temperature as a function of elevation. Correct temperature

estimation is critical in the Cedar basin since the snow level tends to fluctuate substantially during the winter and is often located near the topographic mean elevation of the basin so that a substantial area of the basin is affected by relatively small variations in the snow level (i.e., a few hundred feet).

2) The temperature partition between rain- and snowfall is specified to be in the range of 27° - 32°F mean daily temperature, with all snow assumed to occur below 27°F and all rain above 32°F. In fact, rainfall is rarely observed in the Cascades at temperatures below 32°F.

3) Cloud cover is computed as a function of the daily temperature difference (maximum less minimum) on the basis of a relationship estimated for the California Sierra Nevada. Cloud cover is then used to reduce incipient solar radiation. Although this type of relationship has been found elsewhere to give results nearly as accurate as the more complex energy balance approach, and in any event is essential in the absence of solar radiation records in or near the basin, the particular form of this relationship may be specific to the Sierra Nevada basins on which the model has previously been tested and is suspect here.

4) The condensation melt term used in the model requires unrealistically low wind movement in order to achieve general model calibration, hence the realism of the entire condensation melt algorithm is questionable; and

5) Free water retention in the snowpack appears to be underestimated.

These problems all are related to performance of the snowpack module; the land module appeared to perform adequatly as indicated by reproduction of runoff hydrographs during periods of the year when snow accumulation was neglibible.

The second approach employed was a modification of Tangborn's (1977) storage accounting model, which uses a relationship between low elevation precipitation and accumulated basin runoff to estimate a lumped basin water storage, which is assumed to contribute to runoff during a subsequent forecast period. The model uses a direct approach to forecasting, rather than utilizing streamflow simulation as an intermediate step. The model was modified to allow several options for incorporation of snow course observations to correct basin storage estimates. The forecast model was applied to the Cedar as well as to the Stehekin River basin above the USGS gage at Stehekin, Washington and the American River above the USGS gage near Nile, Washington. Improved forecasts were obtained for the Cedar River when snow course data were included, however the forecasts for the Cedar were the least accurate of any of the basins with a maximum coefficient of prediction (defined as one minus the ratio of the mean square forecast error to the estimated forecast period variance) of about 0.8. Incorporation of the snow course data resulted in reduction of accuracy of the Stehekin River forecast, while the American River forecasts were relatively unaffected by use of the snow course data. The most accurate forecasts for the Stehekin and American Rivers had coefficients of prediction exceeding 0.9. The improvement or lack of improvement achieved through incorporation of the snow course data is thought to be related to the location of the basin and basin topography. The snow course used in the Cedar River forecasts is located near the Cascade crest, as is most of the area of the basin with the greatest snow storage; snow accumulation is relatively consistent in this area and is not highly affected by variations in storm patterns. On the other hand, the Stehekin and American Rivers are east Cascade drainages and so receive precipitation which is effectively filtered by the Cascade Crest; the magnitude of the filtering effect can vary substantially from storm to storm. In addition, the Stehekin basin is characterized by extreme topographic relief with several subranges located east of the Cascade crest but nearly perpendicular to the mean storm path providing additional variable filtering, consequently snow course measurements here represent a highly noise-corrupted measure of mean basin snow storage.

The value of seasonal precipitation forecasts to streamflow forecasts made using the storage accounting model was estimated by examining the performance of an idealized model which included a perfect precipitation forecast. Generally, the precipitation forecast value declined as basin storage increased during the winter, until essentially no improvement in runoff forecasts was obtained for precipitation forecasts beyond March 15. Although a decline in forecast accuracy with forecast date was expected, the magnitude was somewhat surprising since variation in summer precipitation should be one significant source of forecast error. The small improvement achieved using perfect knowledge of spring and summer precipitation (in the case of the Stehekin, use of the precipitation forecast beyond about March 15 resulted in decreased runoff forecast accuracy) may be related to differences in the ability of low elevation precipitation gages to represent basin precipitation during spring and summer months. In any event, it appears that for late winter and early spring forecasts, lack of knowledge of spring and summer precipitation is not a substantial contributor to runoff forecast error when the storage accounting model is used. On the other hand, early winter forecasts could, of course, be substantially improved if more accurate forecasts of precipitation for the subsequent winter months were available.

The results obtained for precipitation forecast value for the storage accounting model point up one of the major differences between the storage accounting approach and continuous simulaztion. A well calibrated continuous simulation model can generally achieve quite small seasonal accumulated runoff volume errors, which implies that if a perfect precipitation forecast were available, quite accurate runoff forecasts could be made. This is not the case, however, with the storage accounting model, where errors associated with the aggregation involved and the general (regression) approach to estimating model parameters assure that some residual error will remain even when precipitation inputs to the basin are perfectly known. Unfortunately, the inability of the continuous simulation model used to adequately reproduce observed events precluded its use in a forecasting mode, so it was not possible to compare forecast performance using the two approaches. The key question in such a comparison would, however, be how large an effect lack of knowledge of forecast period precipitation has on forecast accuracy for the simulation model. This question can be answered conclusively only through utilization of a conceptual model in a split sample testing program, similar to that used to assess the several forms of the storage balance model.

It is unfortunate that forecasts could not be made with the continuous simulation model, since this leaves no real yardstick against which to compare performance of the storage accounting model. This model did, however, appear to perform quite well, especially when applied to the east Cascade drainages which are less affected by rain events. The primary attraction of the storage accounting model is its simplicity. The model is self calibrating so that the user need only assemble the precipitation, runoff, and, if desired, snow course data and perform the screening required to select the best stations. No element of art, such as that involved in calibration of the continuous simulation model, is present, so implementation is much more straightforward and less costly in terms of user time. The primary shortcoming of this type of model as opposed to the continuous simulation approach is the lack of insight yielded into the mechanics of watershed response. Since the model uses only cumulative runoff and precipitation volumes and a lumped basin storage variable, identical runoff could be forecasted for substantially different rainfall and runoff (and consequently subsurface storage) histories. In particular, the model does not account for subsurface free water or soil moisture storage except as it is included in the lumped basin storage variable. Consequently, little insight can be gained into the runoff process under extreme conditions such as droughts. Likewise, given the relatively low worth of knowledge of forecast period precipitation, the model does not lend itself well to evaluation of "alternate scenarios" which are especially useful to water managers during droughts.

Clearly, the choice of the type of model to be used in forecasting is determined by several considerations, among which are ease of use and accuracy. The tradeoffs involved between the storage balance and conceptual simulation approaches seem to suggest that rather than advocating use of a universal forecasting model or modeling method, development of a range of approaches which might be tailored to the needs of individual users would be a more reasonable approach. The research results summarized in this report represent one step in such a process.

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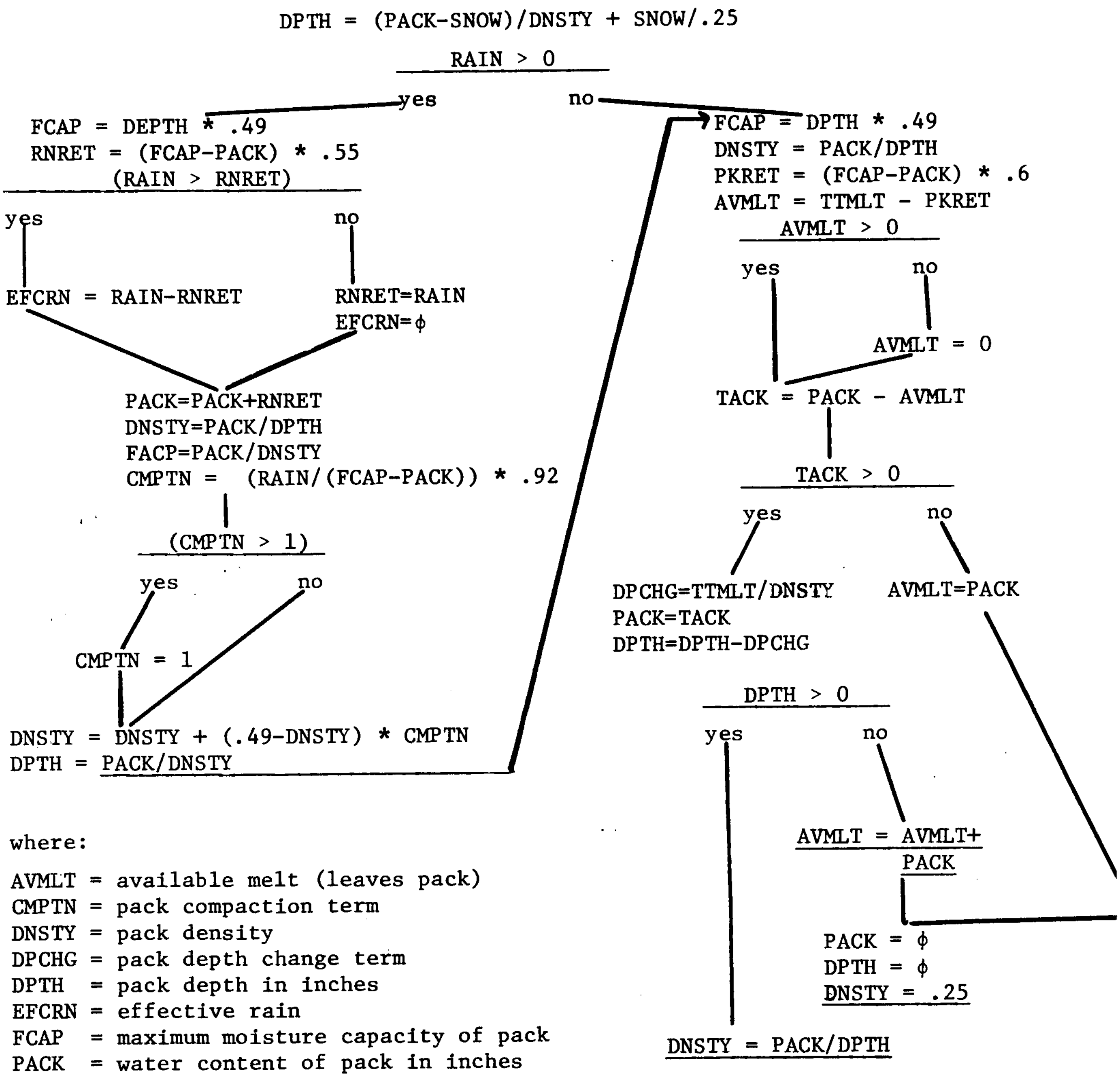
APPENDIX A: COMPACTION ALGORITHM

Figure A-1 summarizes the snowpack compaction algorithm. The algorithm contains two primary branches depending on whether rainfall has occurred during the day. If rain has occurred, a correction to snowpack density due to rainfall retention is made, then the normal compaction computation proceeds.

Figure A-1. Pack Depth Mechanics
(from Burnash, et.al., 1975)



where:

AVMLT = available melt (leaves pack)
CMPTN = pack compaction term
DNSTY = pack density
DPCHG = pack depth change term
DPTH = pack depth in inches
EFCRN = effective rain
FCAP = maximum moisture capacity of pack
PACK = water content of pack in inches
RAIN = liquid precip
RNRET = rainfall retained by pack
SNOW = water content of solid precip
TACK = temporary water content
TTMLT = total daily melt from all sources

Appendix B: Estimation of Missing Data for Precipitation Records

Two types of missing precipitation data were encountered in the records used. The first of these was data simply not collected due to equipment failure or other causes (recorded as 9999). The second was precipitation which was not recorded on the day it fell, but was accumulated and read with precipitation which fell on a later day or days (recorded as 9998).

Estimation of 9999's was accomplished by determining a "parent" station from monthly station correlation matrices. The stations were divided into geographic groups and the parent was selected from the same group. The parent station usually varied by month, so the actual choice was somewhat subjective. The station that had the highest correlation in more months than the others was usually chosen, with consideration also given to the quality of record at the nearby station (i.e., gaps not in the same place as those needed to be filled). Monthly correlations were usually in the neighborhood of 0.7 to 0.9, although they tended to be lower in summer months (0.3-0.6).

9999's were filled by multiplying the precipitation occurring at the nearby station by the ratio of average daily precipitation for the two stations for the appropriate month. 9998's were filled by distributing the precipitation evenly among the days, with all of the fractional parts caused by the division going to the last day (the day the accumulated precipitation was recorded).

Table B-1 summarizes the station groups, amount of missing data, and parent stations.

Table B-1. Summary of Precipitation Data

		Station	Number of 9998's	Number of 9999's	Station used for Replacing 9999's
STEHEKIN	west side	Bellingham 45-0564	5	6	Bellingham 45-0574
STEHEKIN	west side	Darrington 45-1992	761	101	Diablo Dam 45-2157
STEHEKIN	west side	Diablo Dam 45-2157	5	60	Darrington 45-1992
STEHEKIN	east side	Lake Wenatchee 45-4446	699	374	Stehekin 45-8059
STEHEKIN	east side	Mazama 45-5128	3	95	Stehekin 45-8059
STEHEKIN	east side	Stehekin 45-8059	30	494	Mazama 45-5128
STEHEKIN	east side	Wenatchee 45-9074	2	73	Chief Joseph Dam 45-1400
AMERICAN	west side	Electron 45-2493	13	6	Mud Mtn. Dam 45-5704
AMERICAN	west side	Mud Mountain Dam 45-5704	6	6	Electron 45-2493
AMERICAN	west side	Rainier Ohanapecosh 45-6896	312	168	Electron 45-2493
AMERICAN	east side	Yakima 45-9465	0	74	Before WY 1954 estimation (replaced by zeros). 54-75: Ellensburg 45-2505
CEDAR	a				

[a]The six Cedar River records had very few data omissions (generally less than five for the entire record) and these were estimated by visual inspection of the other records.

Appendix C: Corrections to Snow Course Data

The method used to fill in missing snow course data was essentially the same as that used for the precipitation data described in Appendix B. Correlations among snow courses for each month were calculated along with the average recorded snowpack for each month.

To fill the gaps in a snow course's record, it was intended to use the record of the nearby snow course with the highest correlation and multiply the value recorded there by the ratio of the mean snowpacks for the two stations for the appropriate month. However, gaps in the record of the most highly correlated (parent) station often occurred at the same time. In fact, almost all of the gaps in the stations occurred within the period 1949-1956. As a result, Stampede Pass was used for most of the corrections. This station has a fairly complete record (except for gaps in 1951-1953) and covered all months of interest. Bumping Lake also has an excellent record, but the correlation between it and the other stations was generally lower than for Stampede Pass. Occasionally a more highly correlated station was used when there was a period of several months that needed filling in which the nearby station had a complete record. During periods where only a few values could be filled from the most highly correlated station (due to gaps in this record as well), Stampede Pass was used to fill all of the gaps for the period. Stampede Pass had correlations usually of about 0.7 or greater for all stations and months, so the use of this station as the parent should be satisfactory in most cases.

Except for Olallie Meadows, no station on the west side had data for May 1951, all of 1952, and January 1953. Olallie had only March and April 1952 during this period. Consequently, subjective estimates had to be made for May 1951, February and May 1952, and January 1953. The estimates were based on other years in the record having similar March and April snowpacks. Several stations on the east side had data for this period, but it was felt that the correlations between east and west side stations would be too low to be useful.

The missing snow course observations and mthod of estimation are summarized in Table C-1.

Table C-1. Summary of Corrections to Snow Course Data

Basin	Snow Course	Months Available	Corrections		
			Month	Year	Source
STEHEKIN	Harts Pass 20A05A	Feb-May	Feb	1949	Stampede Pass
			Feb	1950	Stampede Pass
			Mar	1950	Stampede Pass
			Apr	1950	Stampede Pass
			Mar	1951	Stampede Pass
			Feb	1953	Stampede Pass
			Mar	1955	Stampede Pass
			Feb	1958	Stampede Pass
			Feb	1969	Rainy Pass
			Mar	1969	Rainy Pass
			Apr	1969	Rainy Pass
			May	1969	Rainy Pass
			Feb	1970	Rainy Pass
			Mar	1970	Rainy Pass
			Apr	1970	Rainy Pass
			May	1970	Rainy Pass
			Feb	1971	Rainy Pass
			Mar	1971	Rainy Pass
			Apr	1971	Rainy Pass
			May	1971	Rainy Pass
	Lyman Lake 20A23A	April		none	
	Meadow Cabins 20A08	Mar-May	Mar	1949	Stampede Pass
			Apr	1949	Stampede Pass
			May	1949	Stampede Pass
			Mar	1950	Stampede Pass
			Apr	1950	Stampede Pass
			May	1950	Stampede Pass
			Mar	1951	Stampede Pass
			May	1951	subjective
			Mar	1952	Olallie Meadows
			Apr	1952	Olallie Meadows
			May	1952	subjective
			Mar	1953	Stampede Pass
			May	1953	Stampede Pass

Basin	Snow Course	Months Available	Corrections Month	Year	Source
STEHEKIN	Park Creek Ridge 20A12A	Feb-Apr	Feb	1949	Stampede Pass
			Mar	1949	Stampede Pass
			Feb	1950	Stampede Pass
			Mar	1950	Stampede Pass
			Feb	1951	Stampede Pass
			Mar	1951	Stampede Pass
			Feb	1952	Rainy Pass
			Mar	1952	Rainy Pass
			Feb	1953	Stampede Pass
			Mar	1953	Stampede Pass
			Feb	1954	Stampede Pass
			Mar	1954	Stampede Pass
			Feb	1955	Stampede Pass
			Mar	1955	Stampede Pass
			Feb	1956	Stampede Pass
			Mar	1956	Stampede Pass
			Feb	1957	Stampede Pass
			Mar	1957	Stampede Pass
			Feb	1958	Stampede Pass
			Mar	1958	Stampede Pass
			Feb	1962	Stampede Pass
			Feb	1966	Stampede Pass
			Feb	1970	Stampede Pass
	Rainy Pass 20A09	Feb-May	Feb	1949	Stampede Pass
			Feb	1950	Stampede Pass
			Mar	1950	Stampede Pass
			Feb	1951	Stampede Pass
			Mar	1951	Stampede Pass
			May	1951	Bumping Lake
			Feb	1953	Stampede Pass
			May	1953	Stampede Pass
			Feb	1954	Stampede Pass
			Mar	1955	Stampede Pass
	Thunder Basin 20A07	Mar-May	Mar	1951	Stampede Pass
			May	1951	subjective
			Mar	1952	Olallie Meadows
			May	1952	subjective
			Mar	1953	Stampede Pass
			May	1953	Stampede Pass
AMERICAN	Bumping Lake 21C08	Jan-May	Jan	1969	Stampede Pass
	Cayuse Pass 21C06	Feb-Apr	Feb	1949	Stampede Pass
			Feb	1950	Stampede Pass
			Feb	1951	Stampede Pass
			Mar	1951	Stampede Pass
			Feb	1952	subjective

Basin	Snow Course	Months Available	Corrections		
			Month	Year	Source
AMERICAN	Cayuse Pass 21C06 (continued)	Feb-Apr	Feb	1953	Stampede Pass
			Feb	1954	Stampede Pass
			Feb	1955	Stampede Pass
			Feb	1956	Stampede Pass
			Mar	1956	Stampede Pass
	Corral Pass 21B13	Mar, Apr	Mar	1955	Olallie Meadows
			Mar	1956	Olallie Meadows
			Mar	1957	Ghost Forest
			Apr	1957	Ghost Forest
			Mar	1965	Ghost Forest
	Ghost Forest 21C04	Feb-Apr	Feb	1949	Stampede Pass
			Mar	1949	Stampede Pass
			Apr	1949	Stampede Pass
			Feb	1950	Stampede Pass
			Mar	1950	Stampede Pass
			Feb	1951	Stampede Pass
			Mar	1951	Stampede Pass
			Feb	1952	subjective
			Mar	1952	Olallie Meadows
			Feb	1953	Stampede Pass
			Mar	1953	Stampede Pass
			Feb	1954	Stampede Pass
			Mar	1954	Stampede Pass
			Feb	1955	Stampede Pass
			Mar	1955	Stampede Pass
			Feb	1956	Stampede Pass
			Mar	1956	Stampede Pass
	Plains of Abraham 22C01A	Jan-May	Jan	1949	Stampede Pass
			Feb	1949	Stampede Pass
			May	1949	Stampede Pass
			Jan	1950	Stampede Pass
			Feb	1950	Stampede Pass
			May	1950	Stampede Pass
			Jan	1951	Stampede Pass
			Feb	1951	Stampede Pass
			May	1951	subjective
			Jan	1952	subjective
			Feb	1952	subjective
			May	1952	subjective
			Jan	1953	Stampede Pass
			Feb	1953	Stampede Pass
			May	1953	Stampede Pass
			Jan	1954	Stampede Pass
			Feb	1954	Stampede Pass
			May	1954	Stampede Pass
			Jan	1955	Stampede Pass
			Feb	1955	Stampede Pass
			Mar	1955	Stampede Pass
			May	1955	Stampede Pass

Basin	Snow Course	Months Available	Corrections		
			Month	Year	Source
AMERICAN	Plains of Abraham 22C01A (continued)	Jan-May	Jan	1956	Stampede Pass
			Feb	1956	Stampede Pass
			Mar	1956	Stampede Pass
			Jan	1965	Stampede Pass
			Jan	1966	Stampede Pass
			Feb	1966	Stampede Pass
			Mar	1966	Stampede Pass
			Apr	1966	Stampede Pass
			May	1966	Stampede Pass
			Feb	1970	Stampede Pass